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The Tropical Cyclone Response to Structural and Temporal Variability in the Environmental Wind Profile

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A dissertation submitted in partial fulfillment of the requirements for the degree of Doctor of Philosophy

THE TROPICAL CYCLONE RESPONSE TO STRUCTURAL AND TEMPORAL VARIABILITY IN THE ENVIRONMENTAL WIND PROFILE

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The aim of this dissertation is to attain a better understanding of how tropical cyclones (TCs) respond to variations in the three-dimensional environmental wind field. Much attention has been given to the impact of environmental wind shear in the 850 – 200 hPa layer on tropical cyclones. However, even with the same magnitude of shear, helicity in this layer can vary significantly. A new parameter is presented, the tropical cyclone-relative environmental helicity (TCREH). Positive TCREH leads to a tilted storm that enhances local storm scale helicity in regions of convection within the TC. Initially we proposed that this enhanced local scale helicity may allow for more robust and longer lasting convection which is more effective at generating latent heat and subsequent TC intensification. Further investigation shows that this is a secondary influence on TC intensity and that variations in the azimuthal and radial position of convection in the TC play a stronger role. Vertical tilt of the vortex is often attributed to wind shear. Different values of helicity modulate this tilt and certain tilt configurations are more favorable for development or intensification than others, suggesting that mean positive environmental helicity is more favorable for development and intensification
than mean negative helicity. Idealized modeling simulations demonstrate the impact of environmental helicity on TC development and intensification. Results show that wind profiles with the same 850-200 hPa wind shear but different values of helicity lead to different rates of development. TCREH also is computed from Era-Interim reanalysis (1979 – 2011) and GFS analyses (2004 – 2011) to determine if a significant signal exists between TCREH and TC intensification. Mean annular helicity is averaged over various time periods and correlated with the TC intensity change during those periods. Results suggest a weak but statistically significant correlation between environmental helicity and TC intensity change with positive helicity being more favorable for intensification.

Another goal of this dissertation is to identify the mechanisms that lead to the observed variations in intensification rate. Results suggest that the difference in intensification rate between TCs embedded in positive versus negative TCREH primarily results from the position of convection and associated latent heat fluxes relative to the wind shear vector. When TCREH is positive, convection is more readily advected upshear and air parcels that experience larger fluxes are more frequently ingested into the TC core. Trajectories computed from high resolution simulations demonstrate the recovery of equivalent potential temperature downwind of convection, latent heat flux near the TC core, and parcel routes through updrafts in convection. Trajectory characteristics show that low-level unstable air is lofted into deep convection near the radius of maximum winds more frequently when TCREH is positive. Contoured frequency-by-altitude diagrams (CFADs) show that convection is distributed differently around TCs embedded in environments characterized by positive versus negative
TCREH. They also show that the nature of the most intense convection differs only slightly between cases of positive and negative TCREH.

Finally, the implications of time-varying environments around TCs are examined. Until now, idealized numerical simulations of the tropical cyclone (TC) response to time-varying wind shear have applied instantaneous changes in the TC environment. A new modeling framework allows for smoothly transitioning environmental wind states: time-varying point-downscaling (TVPDS). TVPDS is an enhancement of the point-downscaling technique (Nolan 2011) developed for the Weather Research and Forecast (WRF) model. It uses analysis nudging to smoothly transition between different environmental vertical wind (and/or temperature and moisture) profiles while coordinating the point-downscaling method such that the environment remains in balance. Using this new framework, results from previous studies are reexamined to test whether the instantaneous ‘shock’ to the environment has implications for TC intensity evolution. Results suggest that instantaneous changes to the TC environment indeed do lead to an unrealistic response to an increase in shear. TVPDS simulations of quasi-steady state, moderately intense (~50 ms⁻¹) TCs show that the response to increasing wind shear is a steady reduction in intensity without a recovery to the pre-shear intensity. TVPDS simulations also show that the rate at which the TC weakens depends on how rapidly the environment transitions from low to high shear. Analyses of surface fluxes and regions of convection are presented to determine how the time-varying shear affects the TC.
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Chapter 1. Introduction

1.1. Shear vs. Helicity: The Environmental Wind Profile

Numerous studies (e.g., Frank and Ritchie 2001; Wang et al. 2004; Riemer et al. 2010; Reasor and Eastin 2012) have demonstrated the importance of deep layer (850 – 200 hPa) wind shear in regard to modulating the development, intensity, and structure of tropical cyclones (TCs). Zehr (1992) identified 850 – 200 hPa wind shear of 12 ms\(^{-1}\) as the maximum threshold for TC formation. The studies listed above investigated how both vertical wind shear magnitude and duration affect TCs. Observational research supports the general finding that vertical wind shear typically inhibits TC intensification (Gray 1968; Merrill 1988; DeMaria 1996) though there are notable exceptions (e.g., Molinari et al. 2004, 2006, Chen and Gopalakrishnan 2015). This weakening (or lack of intensification) may result directly from shear effects (decoupling of the low and upper-level portions of the TC) or indirectly through alterations to the convection within the TC. For example, the strongest convection may occur at larger radii and be less effective at releasing latent heat within the radius of maximum winds (RMW). Modeling studies have shown that the general response of the TC to vertical wind shear is to develop a wavenumber one asymmetry in the eyewall structure (e.g., Bender 1997; Frank and Ritchie 1999; Reasor et al. 2000). Observational studies are in agreement and show that convection in vertically sheared TCs is maximized in the downshear left quadrant (Corbosiero and Molinari 2002; Marks et al. 1992; Molinari et al. 2006; Franklin et al. 1993; Reasor et al. 2013).
Several studies have examined the effects of shear in a simplified dynamic framework. Jones (1995, 2000) and DeMaria (1996) showed that the balanced response to a tilting of the vortex wind field leads to an increase in static stability at low levels. DeMaria suggested the path to weakening then was a suppression of convection in the eyewall region due to increased static stability. Jones, however, showed that the balanced response can be asymmetric and lead to anomalously weak static stability in downshear regions, thus potentially increasing convection in these areas. The vigor, duration, and placement of convection relative to the RMW are very important for intensity change in TCs. Nolan et al. (2007) and Vigh and Schubert (2009) showed that latent heating inside the radius of maximum wind in a TC is most efficient for developing or strengthening the warm core and intensifying the wind field. An environmental wind profile that leads to a TC tilt with increased convection inside the RMW (when compared to a TC in an environment characterized by zero vertical wind shear) can be favorable for TC intensification.

While the effects of vertical wind shear on TCs have been heavily researched, much less attention has been given to the shape of the environmental wind profile. Nolan (2011) performed idealized simulations in the Weather Research and Forecasting (WRF) model to demonstrate that very large differences in TC evolution are possible even with identical values for 850 – 200 hPa wind shear. In the simulations of Nolan (2011), the environmental vertical wind profile outside the TC was held nearly constant using the point-downscaling (PDS) technique. This technique allows for wind shear across the modeling domain without the typically required temperature and pressure gradients to balance the shear. PDS adds a forcing term to the $u$ and $v$ momentum equations which is
equal to the pressure gradient force that would be present if temperature gradients were to exist in the environment. This modification to the momentum equations was previously used in a number of idealized studies of mid-latitude convection (e.g., Skamarock et al. 1994; Davis and Weisman 1994; Weisman and Trapp 2003). Nolan (2011) found that an environment characterized by positive (clockwise hodograph) environmental helicity is more favorable for TC intensification than an environment with negative helicity. Formally, helicity is defined as the vector product of the wind vector and the vorticity vector. Often “storm relative” environmental helicity is used to diagnose the likelihood of supercells (Davies-Jones et al. 1990). Storm relative environmental helicity (SREH) is defined as:

\[ SREH = \int_0^h \left[ (\mathbf{v} - \mathbf{c}) \cdot \left( \mathbf{k} \times \frac{\partial \mathbf{v}}{\partial z} \right) \right] dz \]  

(1.1)

where \( h \) often is taken to be 1 or 3 km, \( \mathbf{v} \) is the horizontal component of the wind field, and \( \mathbf{c} \) is the motion vector of individual thunderstorms. Prior to Nolan (2011), limited research was conducted with regard to the effect of helicity in TCs. Molinari and Vollaro (2008, 2010) investigated the distribution of helicity around TCs and found that it is typically much larger in downshear quadrants. Their studies investigated the effects of helicity on convection within the TC and focused on radii of less than 400 km within which supercells often occur in TCs. The presence of supercells or other persistent convection near and within the radius of maximum wind can aid in intensification of the parent TC. Hendricks et al. (2004) and Montgomery et al. (2006) introduced the concept of vortical hot towers (VHTs) which are deep rotating thunderstorms within TCs. Persistent VHTs near the TC center release large amounts of latent heat. Nolan et al.
(2007) and Vigh and Schubert (2009) showed that latent heating inside the radius of maximum wind in a TC is most efficient for developing or strengthening the warm core and intensifying the wind field. Hogsett and Stewart (2014) proposed a mechanism by which deep updrafts near the TC radius of maximum wind may propagate inward toward the TC center. They described a “left-moving” storm cell which propagates to the right of the vertical wind shear vector (corresponding to leftward of the storm cell motion vector and hence radially inward).

A natural question to ask then is what environmental factors around the TC (radius > 500 km) lead to a local environment within the TC that is most favorable for persistent VHTs near the radius of maximum wind. In this dissertation, the sensitivity of TC intensity to environmental helicity both in idealized WRF simulations and in reanalysis data is analyzed. Numerical simulations show similar results to Nolan (2011) with environments with positive helicity favoring intensification. A new parameter is identified: the tropical cyclone relative environmental helicity (TCREH) which is formulated identically to eq. (1) except that $e$ is replaced by the TC motion rather than some form of thunderstorm cell motion. Results from reanalysis data support the numerical simulations and show a weak but statistically significant correlation between TC intensity change and environmental helicity. A key finding is that the correlation between TCREH and TC intensity change is similar in magnitude to the correlation between 850 – 200 hPa wind shear and TC intensity change, particularly when time periods longer than 72 h are considered. This suggests that the environmental wind profile modulates the tilt of the TC, which then affects the distribution of convection within the TC. Simulations show that positive TCREH leads to anomalously large values
of local SREH in the regions of maximum convection within the TC. The opposite is true for negative TCREH when anomalously smaller values of SREH occurred in the region of maximum convection. In addition to the simulations, TCREH in annuli around best track points is computed using reanalysis data. These results are in agreement with the simulations and showed that environments characterized by positive TCREH were more favorable for TC intensification than those characterized by negative TCREH, with correlations on the order of $R = -0.35$. Interestingly, the correlation between 850 – 200 hPa wind shear and TC intensity in the same annuli was of similar magnitude ($R = 0.40$). These results suggest that the 850 – 200 hPa vertical wind shear vector alone is not optimal for diagnosing the favorability of a TC environment and that the shape of the environmental wind profile also should be considered.

1.2. Time-Varying Tropical Cyclone Environments

To this point, idealized simulations of tropical cyclones in time-varying vertical wind shear have applied sudden changes in the environmental flow. One example is the simulations of Frank and Ritchie (2001) in which idealized TCs developed for 48 h under conditions without wind shear. At this point the simulation was paused, environmental shear was imposed, the temperature and pressure fields were recomputed to balance the imposed shear, and the simulation was resumed. Depending on the magnitude of the imposed shear, the TC either reached a quasi-steady asymmetric state (e.g., shear $\sim 5$ ms$^{-1}$ case) or dissipated completely (shear $= 15$ ms$^{-1}$ case). The authors noted that the rapid weakening of the TC in the 15 ms$^{-1}$ case may have been due partially to the shock from the sudden imposition of vertical wind shear. However, they pointed out that in most
cases, after an initial adjustment period, the TCs once again attained balance with their environments, though this balance meant the TC was well below its maximum potential intensity (MPI). Frank and Ritchie (2001) proposed a sequence of events that result from the imposed shear and lead to weakening of the TC. First, the shear creates a wavenumber one asymmetry in the eyewall region at all altitudes through the TC circulation. Next, the asymmetry in the upper levels becomes strong enough to prevent parcels with high potential vorticity and equivalent potential temperature from entering the eye. This weakens the warm core of the TC and raises the central pressure. Finally, the authors pointed out that the shear advects asymmetric features, leading to a vortex that is tilted downshear. They suggested that a TC in vertical wind shear weakens from the top down. Frank and Ritchie (2001) also noted that the shape of the vertical shear should be explored for its impact on TC intensity change.

Another study by Wang et al. (2004) showed that diabatic processes are important when considering the ability of a mature TC to resist the imposition of vertical wind shear. They found that dry vortices could resist vertical wind shear to some extent, however, when full physics simulations including moist processes were considered, TC resiliency increased. Their simulations allowed a TC to develop over a 96 h period to an intense steady state with a minimum central pressure of approximately 925 hPa. From this point, vertical wind shear of varying magnitude was imposed (instantaneously). Their simulations showed that shear stronger than that imposed by Frank and Ritchie (2001) was required to fully weaken the TC. In their 4 simulations with different values of environmental vertical wind shear (4.25, 12.75, 17.00, 21.25 ms$^{-1}$), only the simulation with shear of 21.25 ms$^{-1}$ caused the initial TC vortex to weaken to a depression state and
never restrengthen. Interestingly, the simulations with shear of 12.75 and 17.00 ms\(^{-1}\) weakened for approximately 24 h after shear imposition, but then reintensified to almost their initial intensity (925 hPa).

Riemer et al. (2010) performed similar numerical simulations to Frank and Ritchie (2001) and Wang (2004) in which a mature TC developed and then an environment with vertical wind shear was imposed. They suggested that lower equivalent potential temperature (\(\Theta_e\)) air enters the inflow layer of the core of a TC and that this process is a result of the balanced response to vertical shear forcing. They argued for a close linkage between environmental vertical wind shear and the thermodynamics of the boundary layer near the TC core. They performed simulations using the Regional Atmospheric Modeling System (RAMS) in which TCs spin up over an 84 h period. Following the methodology of Frank and Ritchie (2001), Riemer et al. (2010) imposed varying shear environments 48 h into their simulations. As in the Frank and Ritchie (2001) simulations, Riemer et al. (2010) noted an imbalance when the shear environment was imposed, however, they noted that the brief adjustment to this imbalance did not affect the interpretation of their results. Much like the simulations discussed previously, the results of Riemer et al. (2010) showed a period of weakening when shear was imposed at 48 h followed by a “recovery” period during which the TCs resumed intensification. Even when shear reached 20 ms\(^{-1}\), the TC recovered and intensified to a strong hurricane (maximum sustained wind \(~85\) ms\(^{-1}\)) by 72 h after shear imposition. The authors pointed out that shear in their simulations was maintained only through forcing by the boundary conditions and that the strong TC was able to modify its environment. They speculated that strong TCs significantly reduce shear in the near-core region by axisymmetrizing the
storm relative flow. Riemer et al. also pointed out that the intensity (and corresponding impacts) of cooler equivalent potential temperature downdrafts may be sensitive to the prescribed background sounding and the microphysics parameterizations. They suggested future numerical experiments to test this sensitivity. Reimer et al. (2013) expanded upon their previous results with additional time-varying shear simulations characterized by more realistic TC environments and peak intensities. The authors found that magnitude of the $\Theta_e$ depression underneath the eyewall correlated with the magnitude of weakening of the TC after shear was imposed. As was the case in their 2010 study, the Reimer (2013) simulations showed an initial weakening of the TC after shear was imposed followed by a subsequent recovery, even for vertical wind shear magnitude of 15 ms$^{-1}$.

In this dissertation, a new modeling framework is developed that allows for smoothly transitioning environmental wind states: time-varying point-downscaling (TVPDS). TVPDS is an enhancement of the point-downscaling technique (Nolan 2011) developed for the WRF model. The new method uses analysis nudging on the outermost domain to smoothly transition between environmental vertical wind (and/or temperature and moisture) profiles while coordinating the point-downscaling method such that the environment remains in balance. A detailed description of the TVPDS method and implementation in the WRF model is provided in Chapter 4. Using this new framework, methodologies from previous studies are emulated. The TC response to environments that transition very rapidly between wind profiles is compared to the response when a more gradual transition is simulated in order to test whether the instantaneous ‘shock’ to the environment has implications for TC intensity evolution. Results suggest that
instantaneous (or in the case of the TVPDS simulations: near-instantaneous) changes to the TC environment indeed do lead to an unrealistic response to an increase in shear. TVPDS simulations of quasi-steady state, moderately intense (~50 ms\(^{-1}\)) TCs show that the response to increasing wind shear is a steady reduction in intensity without a recovery to the pre-shear intensity. And, if the imposed shear is of sufficiently small magnitude, the TC will reach a new quasi-steady lower intensity. This result is particularly different from the Reimer et al. (2010, 2013) studies that showed a recovery to higher intensity after the weakening that occurred immediately after shear imposition. TVPDS simulations in the present study also show that the rate at which the TC weakens depends on how rapidly the environment transitions from low to high shear. Analyses of surface fluxes and regions of convection are presented to determine how the time-varying shear affects the TC. Finally, TVPDS simulations of TCs transitioning from environments characterized by large shear into low-shear environments are performed to determine if a similar but opposite response occurs. These results, presented in Chapter 4 of this dissertation, demonstrate the utility of the TVPDS method in terms of investigating time-varying TC environments.
Chapter 2. Tropical Cyclone-Relative Environmental Helicity in Idealized Simulations and Reanalysis Data

2.1. Introductory Remarks

Environmental factors around tropical cyclones (TCs) such as vertical wind shear and tropospheric relative humidity are known to affect the rate of intensification (or weakening) of TCs. While much research has focused on the impacts of environmental wind shear on TCs (see detailed reviews by Reasor and Eastin 2012 and Nolan and McGauley 2012), much less attention has been given to helicity in the large-scale environments around TCs. Molinari and Vollaro (2008, 2010) showed that helicity can be large at relatively small radii (<400 km) within TCs and they noted that storm-relative helicity was largest in the downshear quadrants. Other studies have pointed out the importance of persistent strong convection near the radius of maximum winds and how it can accelerate the intensification of TCs. Hendricks et al. (2004) and Montgomery et al. (2006) identified vortical hot towers as a mechanism for releasing copious amounts of latent heat at small radii within TCs. Nolan et al. (2007) and Vigh and Schubert (2009) showed that latent heat that is released inside the RMW is most effective when it comes to generating or intensifying the warm core of a TC.

Given the findings of previous studies, it is important to determine what environmental factors around the TC (radius > 500 km) lead to a local environment within the TC that is most favorable for persistent vigorous convection and associated latent heat release near the radius of maximum wind. Nolan (2011) performed simulations in which the kinematic environment around TCs was controlled and
compared the impacts of positive and negative environmental helicity. Nolan (2011) suggested that positive environmental helicity leads to a juxtaposition of convection in downshear quadrants and large values of local cell-relative helicity. It is important to point out the large difference in magnitude when comparing environmental helicity around TCs to local helicity at smaller radii (< 400 km). The simulations of Nolan (2011) used an environment characterized by 0 – 6 km storm-relative helicity of only 4.3 m²s⁻² whereas local 0 – 6 km storm-relative helicity near the location of maximum convection inside the TC was 262 m²s⁻². The simulations suggest that for identical values of 850 – 200 hPa environmental vector wind shear but different environmental helicities, drastically different TC intensity evolution is possible. Therefore, applying the reasoning of Hendricks et al. (2004) and Montgomery et al. (2006) to the Nolan (2011) simulations, it is possible that environmental helicity variations modulate the frequency, vigor, and duration of VHTs and thus the efficiency of intensification in TCs.

In this chapter, connections between the shape of the environmental wind profile and TC intensity are identified. These connections are shown both in numerical simulations and in observational estimates (reanalyses). Finally, a mechanism by which environmental helicity modulates convection with TCs is discussed. The chapter is organized as follows. Section 2.2 describes the numerical simulations that are performed. It explains the modeling techniques, results, and provides some interpretation. Section 2.3 explains the methods and results from reanalysis data. Section 2.4 provides a brief discussion comparing the simulations to the results derived from the reanalysis data. Finally, section 2.5 gives a summary of the primary findings.
2.2. Numerical Simulations

2.2.a Modeling Techniques

To study the effects of environmental helicity on TC intensity and evolution, the Weather Research and Forecasting Model (WRF; Version 3.4.1) is used with two unique configuration techniques. Point-downscaling (PDS; Nolan 2011) is used to control the environment around the TCs in the simulations performed as part of this study. For controlled simulations, it is desirable to have a background environment around the simulated TCs that is very consistent in time and space. By homogenizing the environmental state, it becomes simpler to attribute changes in TC structure and intensity to that background environment. PDS adds a forcing term to the $u$ and $v$ momentum equations which is equal to the pressure gradient force that would be present if temperature gradients were to exist in the environment. In this way a TC can be simulated in wind shear, while the absence of temperature advection means that the wind shear will remain nearly constant in time. This modification to the momentum equations was previously used in a number of idealized studies of mid-latitude convection (e.g., Skamarock et al. 1994; Davis and Weisman 1994; Weisman and Trapp 2003). PDS can be thought of as the Coriolis force acting only on the perturbation winds and it allows for vertical profiles of temperature, humidity, and winds which are nearly constant in time in the simulated environment. This also allows for doubly periodic boundary conditions used on the outermost domain. In addition to the PDS technique, an idealized version of analysis nudging (FDDA; Stauffer and Seaman 1990, 1991) was used to control the TC environment even further. In this idealized version of FDDA, the environment is nudged
toward the initial vertical profile to work in concert with PDS to maintain the nearly constant background state. A nudging timescale of 12 h (relaxation coefficient $\epsilon = 2.31 \times 10^{-5}$ s$^{-1}$) is used.

The simulations in this chapter use 3 grids with 18, 6, and 2 km horizontal resolution and 240×240, 120×120, and 240×240 grid points respectively. 40 vertical levels are used, equally spaced in the WRF sigma coordinate system between the surface and an altitude of approximately 20 km. FDDA was applied only to the 18 km grid. This restricts changes in the TC environment while still allowing the nested domains to freely calculate the higher resolution dynamics within the TC itself. Microphysical processes are simulated with the WRF 6 class microphysics scheme, which includes graupel (WSM6, Hong and Lim 2006). Surface fluxes, friction, and vertical mixing in the planetary boundary layer (PBL) are parameterized using the Yonsei University PBL scheme (YSU, Noh et al. 2003; Hong et al. 2006). The parameterizations for surface fluxes of heat, moisture and momentum for fluxes at high wind speeds follow Dudhia et al. (2008). Longwave and shortwave radiation are not active. The Coriolis parameter is set to $5.0 \times 10^{-5}$ s$^{-1}$ across the domain and sea surface temperature (SST) was held constant throughout the simulations at 29°C.

Utilizing the PDS and FDDA techniques, 120 h simulations are performed with prescribed vertical profiles of height, temperature, humidity, $u$, and $v$. Temperature and specific humidity in the environment are derived from the Dunion (2011) moist tropical sounding. The TC is initialized as a modified Rankine vortex with maximum tangential velocity of 20 ms$^{-1}$, a radius of maximum winds (RMW) of 90 km, and a far field decay parameter of $a = 0.4$. An additional exponential decay function forces the tangential
vortex wind field to zero beyond a radius of 1150 km. The environmental sounding then is held nearly constant by the PDS and FDDA techniques throughout the 120 h simulations. Figure 2.1 shows the environmental $u$ and $v$ profiles as a function of height. Variations in the magnitudes of the cosine function for the meridional wind lead to the different hodographs (Fig. 2.2) used in the 11 simulations.

One concern when attempting to control the environment in a numerical model is that gradual changes to environmental parameters still may occur. Because mean annular tropical cyclone relative environmental helicity (TCREH) is calculated from reanalyses (see section 2.3.a), it is also computed for the simulations and time plots are created (Fig. 2.3) to check for changes during the 120 h simulations. TCREH is formulated similarly to storm-relative environmental helicity in the mid-latitudes (eq. 1) except that the movement of the TC is used rather than some form of storm cell motion. To calculate mean TCREH in an annulus around the TC, a mean vertical profile of wind is calculated in the annulus centered on the TC. This mean wind is defined simply by the average of the $u$ and $v$ components at all available model grid points inside the annulus. The annulus dimensions are 500 – 1500 km (though other sizes were tested; see section 2.3.a). TCREH then is computed as the TC-relative helicity in the mean wind profile. TCREH is computed for several different vertical layers including 0-1, 0-3, 0-6, and 0-12 km. Figure 2.3 shows a time series of 0-3 km TCREH for the simulation corresponding to the wind profiles in Fig. 2.1. 0-3 km TCREH is nearly constant throughout the simulations, remaining within 2.5 m$^2$s$^{-2}$ of its initial value (11.5 ± 1.25, for example, in the simulation corresponding to Fig. 2.1) through each of the entire 120 h simulations. The oscillations in TCREH seen in Fig. 2.3 are due to inertial oscillations triggered by the adjustment
process associated with the formation of an Ekman layer in the boundary layer. These oscillations are most apparent in TCREH computed for low layers.

2.2.b Modeling Results

Figure 2.4 shows the evolution of minimum surface pressure for the 11 120 h simulations along with their corresponding hodographs. Clockwise (positive TCREH) wind profiles are represented with the blue colors while counterclockwise (negative TCREH) wind profiles are represented by green and yellow colors. Clearly, the positive helicity is more favorable for development. When negative environmental helicity falls below approximately -10.5 m²s⁻² (orange line), the TC does not develop. Interestingly, while negative TCREH seems to delay intensification, several of these cases rapidly develop in the 72 – 110 h time range suggesting that during this period these storms overcome their less favorable environment.

TC vortex vertical tilt also is computed for each case. Tilt is defined as the horizontal distance between the vortex centers at 850 hPa and 300 hPa. The vortex center is identified as the centroid of vorticity on the innermost (2 km) nest. Because the simulations are initialized from a weak, precursor vortex, tilt is often large (~200 km) until the TC has strengthened. Figure 2.5 shows a tilt and minimum central pressure diagram for several of the simulations. For cases that develop, tilt decreases and is small by the time development occurs. The delay in tilt reduction is longer for cases with larger negative values of TCREH. It is interesting to compare the upper right and lower left panels of Fig. 2.5, for which the magnitudes of TCREH are equal but of opposite sign. Though 850 – 200 hPa wind shear is identical in the environments of both
simulations, tilt remains large considerably longer for the case with negative helicity (lower left panel). For the case with 0-3 km TCREH of -11.5 m²s⁻², development does not occur and the weak vortex remains strongly tilted throughout the 5 day simulation.

2.2.c Idealized Vortex Tilt

In addition to computing tilt from the simulations, tilt also can be considered in an idealized framework. Both observational studies and numerical simulations (e.g., Huntley and Diercks 1981; Reasor et al. 2004, 2013; Braun and Wu 2007; Rappin and Nolan 2012; Rogers et al. 2013) demonstrate that wind shear affects the vertical alignment of TCs. Figure 2.6 provides a simplified illustration of how a TC is tilted in different background flows: one with positive helicity and one with negative helicity. By shifting the wind fields to mimic a tilted vortex, the effects on local storm scale helicity become evident. Figure 2.6 demonstrates how the environment around a TC can lead to a tilted TC which then modulates the local storm scale helicity. Row a shows how a vertically aligned vortex becomes tilted downshear in a way that increases the amount of positive local scale helicity in the downshear quadrants of the vortex. These quadrants typically are where convection is the most common and most intense (Corbosiero and Molinari 2002). In the presence of negative TCREH (row b of Fig. 2.6), negative local scale helicity (or at least a reduction in the magnitude of positive helicity) is introduced. Even though the hodographs are quite different in Fig. 2.6, the deep layer shear vectors are identical and convection is preferred in the downshear quadrants in both cases. It should be noted that the idealized shifting shown in Fig. 2.6 does not account for boundary layer effects which turn low-level winds inward, leading to an increase in local scale helicity in all quadrants. To compare to the idealized tilting structures in Fig. 2.6,
“tilt-hodograph” plots showing the vortex center at different altitudes are generated from the WRF simulations. Figure 2.7 shows the time-averaged vortex centers for moderately positive and moderately negative values of TCREH. The vortex centers are defined by the vorticity centroid. Fig. 2.7a shows how the vortex center is shifted in a positive TCREH environment while Fig. 2.7b shows the shifting for a negative TCREH case.

One way to consider the effects of TCREH on TC tilt is to horizontally shift the wind fields of a vertically aligned barotropic modified Rankine vortex. In this case we used a maximum tangential wind speed of 30 ms\(^{-1}\) and a radius of maximum winds of 90 km. The wind fields at different heights from this vortex then are shifted as they would be if environmental wind shear were present. For the purposes of illustration, the wind fields at each vertical level are shifted arbitrarily based on the magnitude of the prescribed \(u\) and \(v\) environmental wind components. For example, if the environmental flow at model level 12 is prescribed to values of \(u = 10 \text{ ms}^{-1}\) and \(v = -5 \text{ ms}^{-1}\), the wind fields at that level are shifted 10 grid points (20 km) eastward and 5 grid points (10 km) southward. After this process, helicity in each grid column is calculated to show what regions have the largest local helicity and how these correspond to regions of enhanced convection. Increased local helicity in regions of enhanced convection may lead to organization of vorticity and amplified TC intensification. Figure 2.8 compares the 850 – 200 hPa local helicity within the TC for shifted wind fields in a clockwise (top), straight hodograph (middle), and a counterclockwise (bottom) sense. Local 850 – 200 hPa helicity is largest in the north-northeast region of the TC when positive TCREH is imposed (Fig. 2.8a). Local helicity is largest in the south-southeast region when negative TCREH is present (Fig. 2.8c). However, the maximum value is smaller than in the north-northeast region of
Fig. 2.8a (+260 m²s⁻² vs. +370 m²s⁻²). The case with zero TCREH yields a local helicity maximum directly downshear in the eastern quadrant with weak magnitude around 75 m²s⁻² (Fig. 2.8b). In each case, an anomalously negative region of local helicity exists 180 degrees azimuthally away from the maximum region.

Local helicity generated by idealized shifting of the wind fields in a Rankine vortex can be compared to local helicity from the WRF simulations. Figure 2.9 shows mean hodographs (averaged within the black box in each row) from the region of maximum convection (left column) and helicity contoured over simulated reflectivity from the 700 hPa level (right column) at t = 24 h. The region of maximum convection is defined as a box centered on the maximum value of reflectivity after a smoother is applied. (Actual reflectivity is shown in Fig. 2.9.) Row a of Fig. 2.9 shows results for a simulation with 850 – 200 hPa TCREH of +76 m²s⁻² (0-3 km TCREH of +11.5 m²s⁻²), row b shows results for TCREH of 0 m²s⁻², and row c shows results for 850 – 200 hPa TCREH of -76 m²s⁻² (0-3 km TCREH of -11.5 m²s⁻²). These environments yield mean local 850 – 200 hPa helicity values (helicity of the mean vertical profile of u and v) in the region of maximum convection of 193 m²s⁻² (Fig. 2.9a), 145 m²s⁻² (Fig. 2.9b), and 134 m²s⁻² (Fig. 2.9c). It is evident from the figures that TCREH influences the distribution of convection, with negative TCREH producing convection downshear-right of the shear vector. While there is no clear threshold for environmental helicity when diagnosing the likelihood of supercells, positive TCREH leads to a local environment within the TC that is more often supportive of rotating thunderstorms. In other words, larger positive values of TCREH lead to larger local helicity in the regions of maximum convection. While this
can potentially have implications for TC intensification, these regions often are located in
downshear quadrants outside the RMW which limits latent heat release near the TC core.

2.3. TCREH in Reanalysis Data

2.3.a Reanalysis Methods

The modeling cases of section 2.2 motivate an investigation to determine to what
degree environmental helicity as represented in analysis and reanalysis data can be
correlated to intensity change in TCs. Two data sets are chosen to analyze environmental
helicity around TCs. The first is the ERA-Interim data from the European Centre for
Medium-Range Weather Forecasts (ECMWF; Dee et al. 2011) and the second is Global
Forecast System (GFS) analyses. The ERA-Interim data period is from 1979-2011
(10,162 Best Track 6 h periods greater than 100 km from land) while the GFS analyses
data covers 2004-2011 (2278 Best Track 6 h periods greater than 100 km from land).
Both data sets are global in coverage. However, only tropical depressions, tropical
storms, and hurricanes in the Atlantic are considered (i.e., no tropical waves, sub-tropical
systems, etc.). The ERA-Interim data are 0.75 by 0.75 degrees in horizontal resolution
while the GFS analyses are 0.5 by 0.5 degrees.

For each Best Track time, TCREH is calculated in an annulus centered on the TC
location for 4 different layers (0-1, 0-3, 0-6, and 0-12 km). Centered differencing is used
to compute a TC motion vector from the Best Track data and this vector is used when
computing TCREH. Multiple annuli were tested including a 200-800 km annulus, 500-
1000 km annulus, and a 500-1500 km annulus. TCREH calculated from the 500-1500
km annulus correlates most strongly with TC intensity change and primarily is used for
these analyses. In addition to varying the annulus size, different regional subsets of the Atlantic basin are tested for correlation between TCREH and intensity change. TCs north of 25°N often are more sheared and less tropical in nature and thus are not included in the statistical analysis. It is desirable to have a TC environment that changes as little as possible during the analysis periods so that the impacts of environmental helicity can more easily be evaluated. The effects of land are also reduced by considering only Best Track points greater than 100 km from land. This prevents rapidly weakening storms from skewing the results.

When computing correlations between TCREH and intensity change, different time periods are considered. Mean annular TCREH is computed around each Best Track point (6 h intervals), averaged over different lengths of time, and then compared to the change in TC intensity during the same period. For example, TCREH can be calculated for each Best Track point for a single TC over a 96 h period and its average value then is compared to the change in intensity which is defined as the intensity at $t = 96$ h minus the intensity at $t = 0$ h. The longer the time period becomes, the fewer cases that are available, since it is required that Best Track data is available at all 6 h periods within the correlation period being considered. Correlations for all time periods from 12 h to 168 h are computed to see for which time period TCREH best correlates with TC intensity change. One thing to consider when correlating these factors is whether or not the sign of TCREH changes during the correlation period. For example, it is quite uncommon for TCREH to remain entirely positive or entirely negative for a continuous 96 h period. TCREH remains constantly signed for only 10.2% (168 of 1646 available periods) of the
96 h periods in the ERA-Interim data, and for only 11.5% (64 of 557 available periods) of the periods in the GFS data.

By calculating TCREH from both model simulations and reanalysis with the same methodology, comparisons can be drawn. How important is environmental helicity over time periods of several days? Can the impact of TCREH compare to that of 850-200 hPa shear? Is there a meaningful signal between TCREH and intensity in reanalysis as shown in the simulations? Answers to these questions are provided in the following sections.

2.3.b Reanalysis Results

Results from reanalyses show a weak but statistically significant correlation between time-averaged TCREH and TC intensity change. 1646 concurrent 96 h periods are available from the 8324 best track times for TCs originating south of 25° N and east of 65° W in the Era-Interim data. This geographic criterion is chosen to minimize the effects of land and extratropical transition. Correlating the mean value of 0-3 km TCREH during these 96 h periods to the intensity change during the same periods yields a correlation coefficient (R-value) of -0.2844 ($R^2 = 0.0809$). While this correlation is quite weak, the large number of 96 h periods available provides results which are significant beyond the 99% confidence level. Results from the GFS analyses are similar with an R-value of -0.3607 ($R^2 = 0.1301$). Figure 2.10 shows scatter plots of 96 h mean annular 0-3 km TCREH vs. intensity change for both the ERA-Interim and the GFS data. The red best-fit lines demonstrate that larger values of TCREH correspond most to intensifying TCs while larger negative values correspond to weakening storms (note that negative values of pressure change on the y-axis implies a strengthening TC). Clearly
there are many cases in which the magnitude of 0-3 km TCREH is quite small (<5 m²s⁻²).
The relationship between intensity change and TCREH becomes stronger when these cases were removed and correlations were recomputed. When considering only cases with 96 h mean 0-3 km TCREH absolute value > 5 m²s⁻² in the ERA-Interim data (not shown), the R-value improved to -0.4172 (R² = 0.1741) and the result remained statistically significant. These results suggest that for larger positive (negative) values of annular mean TCREH strengthening (weakening) is more likely. Figure 2.11 shows the same analysis as Fig. 2.11 except for 48 and 72 h time periods. The correlation between 0-3 km TCREH and intensity change is smaller for shorter time periods (R = -0.0772 for 48 h and R = -0.1733 for 72 h).

The R-values computed for 96 h TCREH are quite similar to correlation coefficients computed for 850 – 200 hPa wind shear. For example, DeMaria (1996) reports correlation coefficients on the order of 0.3 at forecast ranges of 24 – 72 h. The similar R-values suggest that 0-3 km TCREH may have nearly as much predictive capability as the value of 850 – 200 hPa shear. DeMaria defined TC intensity change using maximum surface wind speed. Figure 2.12 shows a scatter plot of 96 h mean annular 0-3 km TCREH vs. intensity change as defined by the change in maximum wind speed during the period. When correlating 0-3 km 96 h TCREH with change in maximum surface wind speed, the R-value magnitude increases slightly (R = 0.3108, compared to R = -0.2844 when using minimum central pressure).

To show an equivalent analysis for wind shear, Figure 2.13 shows a scatter plot of 96 h mean 850 – 200 hPa wind shear (computed with the same 500 – 1500 km annulus used for TCREH) vs. intensity change. The magnitude of the R-value (0.3993) is similar to
that of TCREH (-0.2844). Both plots show large scatter, particularly when TCREH and shear magnitude are small.

0-3 km TCREH is computed for different time periods and Fig. 2.14 shows correlation coefficients vs. time period. Figure 2.14 here is quite similar to Fig. 8 of DeMaria (1996) which shows a similar analysis for 850 – 200 hPa wind shear. While DeMaria showed that wind shear correlates better on shorter timescales (0 – 72 h), our results suggest that TCREH and wind shear are useful as predictors of TC intensity change at longer forecast ranges (96 – 168 h). One potential reason for this delay in impact may be that a two-step process is required for the TCREH to influence the convection inside the TC. First, the TCREH must tilt the TC, which may require persistence of the TCREH. Then, this tilt leads to a response in the convection at smaller radii inside the TC. The response of TC intensity to differences in convection again may require additional time. It is important to note that fewer cases with concurrent Best Track points are available as the time period lengthens. However, the results shown in Fig. 2.14 are statistically significant beyond the 99% level even when considering 168 h forecast periods. The blue and cyan lines of Fig. 2.14 show correlation coefficient vs. time period for 850 – 200 hPa wind shear. While wind shear generally correlates slightly more strongly with TC intensity change when compared to TCREH, a 24 h period from 72 to 96 h exists in which TCREH (cases for which magnitude greater than ±5 m²s⁻² are not considered) correlates more strongly. Figure 2.15 shows a scatter plot of 0-3 km 96 h TCREH vs. zonal wind shear. It could be possible that helicity is simply correlated with wind shear. However, the correlation between TCREH and 850 – 200 hPa zonal wind shear is modest, with R = -0.4111. There is considerable scatter with a preference for
westerly shear when TCREH is strongly negative. This modest inter-correlation suggests that wind shear and TCREH both are explaining variance in TC intensity change and that the combination of these factors explains more variance than does just one or the other by itself.

Along with different time periods other than 96 h, different annulus sizes also were tested for their correlation with intensity change. Annuli of 200 – 800 km and 500 – 1000 km were tested but yielded correlations between TCREH and TC intensity change that were smaller than those derived from the 500 – 1500 km annulus. Correlation between TC intensity and TCREH also were tested for subsets of the data based on TC intensity. Correlation drops to $R = -0.1105$ when TCs at or above hurricane intensity ($> 33 \text{ ms}^{-1}$) are not considered. This suggests that weakening hurricanes embedded in negative TCREH likely play a strong role in determining the correlation between TCREH and TC intensity.

2.4. Discussion

When comparing the modeling results to the results from reanalysis some inferences can be made. Reanalysis data supports the idealized simulations that suggest that increased positive environmental helicity is correlated with the intensification of TCs. The reanalysis data also suggest that the relationship becomes stronger over longer time periods (particularly those longer than 72 h). In the idealized simulations, TCs with positive TCREH tend to intensify most during the 36 – 72 h time frame. This may be a function of the initial precursor vortex state but this has not been tested. Reanalysis data suggests that TCREH is playing a role on longer time scales but does agree with the
general result that positive helicity favors intensification. One of the primary differences between the idealized simulations and the TCs in reanalysis is the strict control and steadiness of the environment that is present in the simulations. Never in nature is a TC environment truly steady for any meaningful amount of time, let alone for five days. This, perhaps, is why TCs in the simulations respond much more quickly to the TCREH in their environments. While it takes longer in nature, TCREH does seem to play an important role with correlations at 96 – 156 h that are nearly as great as those with 850 – 200 hPa shear.

2.5. Summary

WRF simulations of tropical cyclones in environments of varying helicity suggest a relationship exists between TC intensification and TC-motion relative environmental helicity (TCREH). 11 simulations with 0-3 km TCREH varying from -11.5 m$^2$s$^{-2}$ to +11.5 m$^2$s$^{-2}$ show that development occurs sooner for cases of positive TCREH, later for cases with negative helicity, and not at all for values of TCREH less than -10.5 m$^2$s$^{-2}$. These simulations all are performed with identical values of environmental 850 – 200 hPa wind shear showing that the shape of the vertical wind profile can be as important as the magnitude of the shear vector. Correlations derived from 33 years of reanalysis data suggest that this signal is apparent, though primarily at longer time scales than shown in the modeling simulations. Era-Interim reanalysis data shows increasing correlation between TCREH in an annulus around TCs and TC intensity change through 156 h. Previous studies have demonstrated the strong relationship between 850 – 200 hPa wind shear and intensity change through 72 h. TCREH appears to provide useful information
about intensity at longer ranges out to 156 h and perhaps even longer. It is important to point out that these relationships are derived from reanalyses and may not exist when comparing long range forecasts of TCREH to verifying intensity change.

One reason that positive environmental TC-motion relative helicity seemingly is more favorable for intensification is tied to the distribution of convection. Idealized horizontal shifting (section 2.2.c) of the wind fields in a TC embedded in an environment of no flow (thus no shear) demonstrate how convection and positive local-scale (i.e., thunderstorm scale) helicity coexist when environmental helicity is positive. This overlap of positive helicity and convection potentially leads to longer lasting and more vigorous thunderstorms that can more efficiently generate latent heating. This more persistent heating then can lead to intensification of the TC. However, this overlap of large local helicity and convection often occurs outside the radius of maximum winds and thus releases little or no latent heat near the TC core. When TCREH is negative, regions of convection typically are displaced from regions of positive local-scale helicity. This unfavorable alignment combined with the fact that convection is occurring at large radii slows intensification and can even lead to weakening, as suggested by the reanalysis data.

In the following chapter it will be shown that the azimuthal and radial position of convection and associated surface latent heat fluxes is related to the sign of TCREH. It is this relationship that plays the key role in demonstrating why positive TCREH favors intensification over negative TCREH.
Fig. 2.1. Vertical profiles of $u$ (blue) and $v$ (red) wind components (ms$^{-1}$) for a simulation with positive helicity (left panel) and the corresponding hodograph (right panel).
Fig. 2.2. The 11 hodographs (850 – 200 hPa) used for the prescribed environmental flow in the WRF simulations.
**Fig. 2.3.** Time series of 0-3 km TCREH (m$^2$s$^{-2}$) for a simulation with an idealized clockwise turning hodograph. This simulation was run with the environmental flow set to the wind profile shown in Fig. 2.1.
Fig. 2.4. Evolution of minimum surface pressure (left) for the 11 120 h simulations with varying TC-relative environmental helicity and their corresponding 850 – 200 hPa hodographs (right).
Fig. 2.5. 850 – 300 hPa tilt (km, 9 h running mean) in black and minimum central pressure (hPa) in red vs. time (h) for 4 simulations with different environmental TCREH: +11.5 m²s⁻² (a), +6 m²s⁻² (b), -6 m²s⁻² (c), -11.5 m²s⁻² (d). Tilt is defined as the horizontal distance between the vorticity centroids at 850 and 300 hPa.
Fig. 2.6. Idealized vortex tilting. The green circles (and corresponding lines and arrows) represent the vortex near the surface. The yellow circles and lines represent the vortex at approximately 1500 m and the red circles and lines represent the vortex at approximately 3000 m. In both the top and bottom sections, the left panel shows a vertically aligned vortex. The middle panel shows the imposed environmental wind profile. The right panel shows the resulting vortex alignment. The top diagram shows the effects of positive TCREH and the bottom diagram shows the effects of negative TCREH. The purple curves demonstrate the orientation of the resulting local hodograph (clockwise for positive TCREH and counterclockwise for negative TCREH) in the downshear left quadrant.
Fig. 2.7. Tilt-hodographs for a positive TCREH (a) and negative TCREH (b) case. The blue circles indicate the position of the vorticity centroid of the vortex relative to the centroid at the surface for 900, 850, 700, and 500 hPa. The centroids positions are averaged from $t = 48$ h to $t = 66$ h which corresponds to the period during which significant development occurs (see Fig. 2.4).
Fig. 2.8. Comparison of local scale 850 – 200 hPa helicity for idealized shifted wind fields in a clockwise (top), straight hodograph (middle), and a counterclockwise (bottom) sense. Black (red) contours indicate positive (negative) local helicity. Shear is purely zonal (westerly) in each row.
Fig. 2.9. Hodographs in the region of maximum convection (left column) and contoured local helicity (m$^2$s$^{-2}$) overlaid on 700 hPa simulated reflectivity (dBZ; right column). Each row corresponds to a simulation with 850 – 200 hPa TCREH equal to +76 m$^2$s$^{-2}$ (row a), 0 m$^2$s$^{-2}$ (row b), and -76 m$^2$s$^{-2}$ (row c). The hodographs represent the mean vertical wind profile from within the black box on the reflectivity plots.
Fig. 2.10. Scatter plots of 0-3 km 96 h annular mean TCREH (m²s⁻²) vs. intensity change (hPa) for 1646 96 h periods in ERA-Interim reanalysis (top) and GFS analysis (557 periods; bottom) data. The red lines are the best-fit lines with R = -0.2844 (ERA-Interim) and R = -0.3607 (GFS). The green lines are for only cases with TCREH absolute values > 5 m²s⁻² and R = -0.4172 (ERA-Interim) and R = -0.2686 (GFS). An annulus of 500 – 1500 km centered on the TC was used.
Fig. 2.11. Scatter plots of 0-3 km 48 (left) and 72 (right) h annular mean TCREH (m²s⁻²) vs. intensity change (hPa) in ERA-Interim reanalysis data. The red line is the best-fit line with of $R = -0.0772$ (48 h) and $R = -0.1733$ (72 h). An annulus of 500 – 1500 km centered on the TC was used.
Fig. 2.12. Scatter plot of 0-3 km annular mean TCREH (m²s⁻²) vs. intensity change (as defined by change in maximum wind speed; kt) for 1646 96 h periods in ERA-Interim reanalysis data. The red line is the best-fit line with $R = 0.3108$. The green line is the best fit line when considering only cases with TCREH absolute values $> 5$ m²s⁻² and $R = 0.4744$. An annulus of 500 – 1500 km centered on the TC was used.
Fig. 2.13. Scatter plot of 850 – 200 hPa annular mean wind shear (ms\(^{-1}\)) vs. intensity change (hPa) for 1646 96 h periods in ERA-Interim reanalysis data. The red line is the best-fit line with R = 0.3993. An annulus of 500 – 1500 km centered on the TC was used.
Fig. 2.14. Absolute value of correlation coefficient (between 0-3 km annular TCREH and TC intensity change and between 850 – 200 hPa annular wind shear and TC intensity change) vs. time period length. The red line is for cases where absolute values of TCREH of less than 5 m²s⁻² are not considered while the blue line considers all magnitudes of TCREH. The green line represents 850 – 200 hPa wind shear correlation vs. time period and the cyan line is for cases where shear magnitudes of less than 5 ms⁻¹ are not considered. All dots are on 12 h intervals.
Fig. 2.15. Scatter plot of 0-3 km annular mean TCREH (m²s⁻²) vs. zonal 850 – 200 hPa wind shear (ms⁻¹) for 1646 96 h periods in ERA-Interim reanalysis data. An annulus of 500 – 1500 km centered on the TC was used.
Chapter 3. Tropical Cyclone-Relative Environmental Helicity and the Pathways to Intensification in Shear

3.1. Introductory Remarks

While vertical wind shear generally limits the intensification of TCs, the specifics of the vertical shear profile can modify the intensification rate. Rappin and Nolan (2012) used numerical simulations to show that vertical wind shear profiles with equal shear magnitudes but opposite direction relative to TC motion can lead to a different rate of TC development. Nolan and McGauley (2012) expanded upon this result demonstrating that westerly shear was more favorable for genesis than easterly shear due to the differences in surface flux positioning around the TC that arise between the two shear environments. Other studies have noted the dependence of TC-intensification rate on the translation rate of the TC or the specific layer under consideration when computing vertical wind shear. For example, Zeng et. al (2010) found that the intensification rate of fast moving TCs (≥ 6 ms⁻¹) was more correlated with wind shear in the 600 – 200 hPa layer than in the more typically used 850 – 200 hPa layer, suggesting that TCs are strongly affected by shear in the middle to upper troposphere when they are moving quickly.

In Chapter 2 it was shown that the shape of the vertical wind profile can have significant impact on TC development rates even when the 850 – 200 hPa wind shear vector is held constant. Those results showed that positive values of tropical cyclone-relative environmental helicity (TCREH) promoted faster rates of intensification, using both numerical simulations and reanalysis data. Recall that TCREH is a measure of how
the wind vector changes with height and is defined as the vector product of the TC motion-relative environmental wind and horizontal vorticity vectors:

$$TCREH = \int_0^h \left[ (v - c) \cdot \left( k \times \frac{\partial v}{\partial z} \right) \right] dz \quad (3.1)$$

where $h$ is the depth over which $TCREH$ is computed, $v$ is the horizontal component of the wind field, and $c$ is the motion vector of the TC. It is important to distinguish $TCREH$ (radius $> \sim 500$ km; magnitudes typically $< 100$ m$^2$s$^{-2}$) from TC-scale helicity (radius $< \sim 500$ km; magnitudes often $> 300$ m$^2$s$^{-2}$). In Chapter 2, two hypotheses were presented to explain why $TCREH$ modulates TC intensification rate. The first is that the characteristics of convection in the TC are affected by the TC response to the environmental helicity. Results presented in Chapter 2 suggested that positive $TCREH$ leads to larger magnitudes of TC-scale helicity, thus allowing more organized and vigorous convection. Secondly, those results suggested that positive TC-scale helicity and the location of maximum convection (in downshear quadrants) were more collocated when $TCREH$ is positive. In this chapter, these ideas are more rigorously tested and an alternative explanation is presented in which the azimuthal position of convection and surface latent heat flux is more conducive to TC-intensification when $TCREH$ is positive.

Recent studies have related the azimuthal position of convection to the timing of TC intensification because this position influences the feedback between diabatic heating and the TC primary circulation (Tao and Zhang 2014). Additional studies have discussed how the precession of the middle or upper-level circulation around the low-level TC center affects the timing to genesis or rate of intensification (Reasor and Montgomery 2001; Rappin and Nolan 2012; Zhang and Tao 2013). Zhang and Tao (2013), Stevenson
et al. (2014), and Chen and Gopalakrishnan (2015) noted that intensification ensued soon after the upper-level center of the vortex precessed beyond 90 degrees to the left of the vertical wind shear vector. In this chapter, it will be shown that environments characterized by positive TCREH favor a faster precession of convection and associated surface fluxes into the upshear quadrants.

3.2. Azimuthal Positioning of Surface Fluxes and Convection

3.2.a Methods

32 simulations using 2 km grid spacing were performed in which 850 – 200 hPa TCREH and wind shear were held fixed but various constant background flows were added to the wind field. These simulations were designed to test whether certain background flows promoted faster advancement of convection and enhanced surface fluxes into upshear quadrants. The Weather Research and Forecast Model (WRF) version 3.4.1 was used with point-downscaling (Nolan 2011) and analysis nudging (Stauffer and Seaman 1990, 1991). The model configuration, TC initial intensity, radius of maximum winds (RMW), sea surface temperature (SST), and other characteristics follow exactly the configuration described in Chapter 2. Specifically, all TC simulations described in this chapter were initialized as modified Rankine vortices with initial maximum tangential velocity of 20 ms$^{-1}$ on an $f$-plane at 20° N with an SST of 29° C.

In addition, two very-high resolution (667 m) simulations were performed to analyze the importance of the azimuthal position of features such as surface fluxes, diabatic heating rates, and convection. The 667 m simulations follow the configuration from Chapter 2 except with the addition of a 240 × 240 km fourth nest (360 × 360 grid points),
centered on the TC, embedded within the 2 km nest. The environmental 850 – 200 hPa wind profiles are shown in a hodograph diagram in Fig. 3.1 and are characterized by westerly shear of 10 ms\(^{-1}\). It is proposed that when air characterized by larger fluxes, increased moisture, and larger diabatic heating rates is advected cyclonically into upshear quadrants near the RMW, then more of this air is ingested into the TC core which leads to faster intensification. Three methodologies were used to test this idea. The first method used time-averaged, TC-center-relative composites of surface fluxes, simulated reflectivity, and diabatic heating rates. Results from these composites are discussed in the following section.

3.2.b Results

The azimuthal positions as well as the characteristics of surface fluxes, simulated reflectivity, and diabatic heating rates are compared for simulations with positive and negative TCREH. Comparisons are made primarily during periods just prior to when TC intensity diverged between simulations with positive and negative TCREH. TC intensity diverges for the two 667 m simulations beginning at approximately \(t = 36\) h (Fig. 3.2). This period is chosen in an attempt to relate differences in azimuthal positioning to the differences in subsequent TC intensification rate. Theses simulations show that the location of larger values of latent heat flux advances into upshear quadrants near the RMW more quickly when TCREH is positive. Figure 3.3 shows surface latent heat fluxes which are averaged over 4 h subsets of the 12 h period prior to divergence of minimum central pressure between the simulations with positive and negative TCREH. The top row (panels a – c) shows how fluxes of larger magnitude proceed into the upshear quadrants sooner than they do in the case with negative TCREH (bottom row:
panels d – f). The result of larger latent heat fluxes in the upshear-left quadrant near the RMW is a more efficient ingestion of high-\(\Theta_e\) surface air into the TC core. Note that the 850 – 200 hPa vertical shear vector is identical for both simulations and is directed due eastward with a magnitude of 10 ms\(^{-1}\). Figure 3.4 shows time-averaged plots of convection, as visualized by 700 hPa simulated reflectivity, which show the faster upshear advection in the positive TCREH simulation (top row). While the magnitude of time-mean reflectivity increases with time for the negative TCREH case (bottom row of Fig. 3.4), the azimuthal position of the maxima makes essentially no progress towards the upshear quadrants. Time-mean reflectivity for the positive TCREH case (top row of Fig. 3.4) shows a steady advancement of convection into upshear quadrants. It is this advancement of convection and surface fluxes (shown further in the trajectory analysis in section 3.3.b) that leads to the discrepancy in intensification rate.

The azimuthal location of convection and surface fluxes are related to TC-vortex tilt and the precession of the mid-level vortex around the low-level vortex. Tilt is calculated as the vector difference between the surface and 500 hPa circulation centers. The time at which this vector first points towards the upshear-left quadrant is noted (Fig. 3.2) since this has been shown to be related to the onset of rapid intensification (Rappin and Nolan 2012; Tao and Zhang 2014). By comparing the precession of the circulation center to the azimuthal location of convection and fluxes, inferences can be made about the relative importance of both factors. Figure 3.5a shows time series of the storm-relative azimuth angle from the TC surface center to the location of maximum 700 hPa convection for the 667 m simulations with positive (red) and negative (blue) TCREH. Azimuth angles for the mid-level circulation center also were computed (not shown) and were generally
correlated with the position of the maximum 700 hPa convection. Figure 3.5b shows the radial distance from the TC surface center to the location of maximum 700 hPa convection (positive TCREH: red, negative TCREH: blue). The time series demonstrate that convection precesses faster and reaches smaller radii sooner when TCREH is positive. Interestingly, in both the positive and negative TCREH simulations, the advancement of convection and surface fluxes occurs at the same time or slightly before the precession of the mid-level circulation center (not shown). This suggests that surface fluxes and ensuing deep convection may play an important role in determining how easily the mid-level vortex can precess cyclonically around the low-level TC-center (perhaps by preconditioning the environment in the upshear quadrants), or instead that convection is relocating the center by generating vorticity. This result stands somewhat in contrast to the results of Davis et al. (2008) who showed that convection to the left of the vortex tilt vector can inhibit the precession rate by reducing the cyclonic advection of the low-level center by the upper-level potential vorticity maximum.

Because TC motion is slightly different in the positive and negative TCREH cases described above, it was proposed that the position of surface fluxes and convection relative to the TC motion vector may play an important role in determining how efficiently buoyant surface parcels are ingested into the TC core. To investigate this issue, 32 simulations similar to the 667 m cases described above were performed at coarser grid spacing (finest domain uses 2 km rather than 667 m) and with the addition of spatially constant background flow. The background flow is of constant magnitude at all altitudes such that the shape of the environmental hodograph does not change, however its position on the diagram moves around depending on the direction of the background
flow. Figure 3.6 shows an example of how the environmental hodograph is shifted by the addition of southwesterly background flow of 4 ms\(^{-1}\). Four simulations are performed for each background wind direction in 45 degree increments (i.e., northerly, northeasterly, easterly, southeasterly, southerly, southwesterly, westerly, and northwesterly). For each direction, two simulations with negative TCREH and the addition of background flow are performed: one with magnitude 4 ms\(^{-1}\) and one with 8 ms\(^{-1}\). Two simulations with additional background flow of 4 ms\(^{-1}\) and 8 ms\(^{-1}\) but with positive TCREH also are performed for each direction. Significant differences in the timing of intensification result simply from varying the direction of the background flow.

To illustrate the differences in intensification rate between the westerly and easterly background flow cases, Fig. 3.7 shows minimum central pressure evolution from a mini ensemble containing 6 positive-TCREH simulations (3 with westerly and 3 with easterly background flow). This mini ensemble is generated by introducing small random perturbations (< 0.5 ms\(^{-1}\)) to the initial TC vortex wind and is used to demonstrate how TCREH and vertical wind shear remain positive and approximately constant for all 6 simulations in which a uniform background flow with magnitude 8.0 ms\(^{-1}\) is added to the three-dimensional wind field. The main differences between the westerly and easterly experiments are the direction from which this background flow is imposed and the resultant increased (decreased) storm-relative shear in the easterly (westerly) case. Storm-relative shear has been shown to modulate the distribution of convection in TCs (e.g., Rogers et al. 2003) and is defined as the difference between the wind shear and storm motion vectors. There is a fairly large range of intensification rate when only the direction of the background flow is varied. For example, a difference of approximately
22 hPa exists at $t = 60$ h between the westerly background flow cases (Fig. 3.7: blue lines) and the easterly background flow cases (Fig. 3.7: red lines). The different intensification rate is smaller (~10 hPa) but consistent in terms of easterly flow being more favorable when a weaker background flow of 4 ms$^{-1}$ is added. The simulations also demonstrate a steady increase in development rate as the additional background flow transitions from westerly to easterly (Fig. 3.8; from westerly to southwesterly, to southerly, to southeasterly, to easterly).

To determine what causes the different rate of development in these cases, time-mean plots of 700 hPa simulated radar reflectivity and surface latent heat flux from one westerly and one easterly additional background flow case are analyzed during the 12 h period just prior to when TC intensity diverges. Figure 3.9 shows how stronger convection remains near and just upshear of the TC core when the additional background flow is easterly (row 1: panels a – c) versus westerly (row 2: panels d – f). The convection also progresses more rapidly around the TC center when the additional background flow is easterly. By time $t = 24 – 28$ h, a more symmetric appearance is noted in the easterly background flow case when compared to the westerly case (panel c vs. panel f).

Corresponding to the more organized convection in the easterly background flow case, a broad area of relatively large surface latent heat fluxes persists near the RMW in the upshear-left quadrant. Contours of horizontal divergence on the lowest model level (Fig. 3.9; rows 3 – 4: panels g – l) show that there is much more overlap of larger surface latent heat flux and low-level convergence when the background flow is easterly. The overlap of surface latent heat flux and convergence occurs despite the fact that storm-
relative shear is increased when the background flow is easterly. Storm-relative shear is reduced when the background flow is westerly, however, decreased overlap between convection and low-level convergence disallows a faster TC-intensification rate. The combination of increased convection and fluxes collocated with low-level convergence in upshear quadrants most likely is why the case with easterly additional background flow develops more rapidly than the westerly case.

We propose that the reason for the differences between these cases has to do with the TC-scale advection of air experiencing large fluxes relative to TC storm motion. For both the easterly and westerly cases, convection is initially maximized north (left of shear) of the TC center. In the easterly background flow case (Fig. 3.9 rows 1 and 3: panels a – c and g – i), the path from the surface flux maximum to the storm motion vector (which is directed toward approximately 280 degrees) is much shorter than the same path in the westerly case. In addition to the shorter path, near-surface easterly winds to the north of the TC in regions of convection are stronger when the additional background flow is easterly due to the projection of TC motion onto the vortex winds. These stronger surface winds lead to larger fluxes over a broader region in the upshear-left quadrant. Surface latent heat flux averaged over a 360 × 360 km box centered on the storm are 376 Wm$^{-2}$ when the background flow is easterly (Fig. 3.9 row 3: panels g – i) compared to 331 Wm$^{-2}$ when the background flow is westerly (Fig. 3.9 row 4: panels j – l). This difference increases when considering a 180 × 180 km box in the upshear-left quadrant where average surface latent heat fluxes are 497 Wm$^{-2}$ vs. 370 Wm$^{-2}$. Another way to conceptualize the difference is by comparing the motion vector of the near surface air north of the TC (and its associated fluxes) to the TC motion vector. In the easterly
case these vectors are approximately 10 degrees apart while in the westerly case they are nearly 180 degrees apart. This difference and the resultant slower rate of TC development in the westerly case implies that the ingestion of surface air experiencing large fluxes into the TC-core becomes less efficient when this angle is larger. This idea is more rigorously tested in the following section.

Because varying the additional background flow appears to lead to large differences in intensity (~22 hPa difference at $t = 60 \text{ h}$ between easterly and westerly background flow cases), the results from the simulations with positive TCREH and background flow (Fig. 3.7) were compared to simulations with background flow and negative TCREH. Figure 3.10 shows 4 simulations with additional background flow of 8 ms$^{-1}$. The solid lines show westerly (blue) vs. easterly (red) background flow for simulations with positive TCREH. The dashed lines show westerly (blue) vs. easterly (red) background flow for simulations with negative TCREH. What is apparent from Fig. 3.10 is that both the direction of the background flow and the sign of TCREH are important in dictating intensification rate. For example, the intensity at $t = 48 \text{ h}$ is approximately 972 hPa for the simulation with easterly flow and positive TCREH compared to 992 hPa for the simulation with easterly flow and negative TCREH. The difference here is of equal magnitude to that when varying just the background flow. However, this difference increases to 28 hPa when TCREH is negative and the background flow is westerly. This result indicates that both TCREH and the orientation of the background flow are important in dictating intensification rate.
3.3. Trajectory Analysis

3.3.a Methods

The next analysis method was to compute trajectories (backward and forward) to see the paths traveled by parcels near the TC core and in regions of large fluxes or diabatic heating. In order to compute accurate trajectories it was required that very high resolution be used in both space and time. Output from the innermost domain of the 667 m simulations was saved every one minute. The methods used for the trajectory computation followed a predictor-corrector technique. An example for the case of forward trajectories follows. First, wind components are interpolated to the initial parcel location. These components then are used to make a prediction for where the parcel will be advected over the next 30 seconds. This location serves as the predicted midpoint. Wind components then are interpolated spatially and temporally to this predicted midpoint. These interpolated wind components then are used to advect the parcel for the full one minute time step. This example describes just one predictor-corrector step per model output interval. Alternatively, each one minute interval can be broken into numerous predictor-corrector steps. For the trajectories shown in section 3.3.b, each one-minute interval is broken up into 30 two-second predictor-corrector steps. This two-second time stepping was chosen because additional time resolution did not substantially improve trajectory accuracy (i.e., trajectories were nearly identical when using a one-second predictor-corrector step instead of a two-second step). Air mass properties such as diabatic heating rate, water vapor mixing ratio, equivalent potential temperature ($\Theta_e$), etc. were stored at one-minute intervals along the computed trajectories. Trajectories
were seeded in locations such as maxima in surface fluxes, within thunderstorms, or near the surface downwind of convection to compare how air in these locations arrives to or proceeds from these locations. Comparisons were typically made during periods just prior to when TC intensity diverged between the positive and negative TCREH simulations. Zhang et al. (2013) showed that boundary layer $\Theta_e$ recovery rates are related to the distribution of convection around TCs. To determine if boundary layer recovery occurs more rapidly when TCREH is positive, $\Theta_e$ recovery rates in the boundary layer downwind of convection were examined.

3.3.b Trajectory Results

Results from forward trajectories suggest that near-surface parcels experiencing large latent heat flux are advected around the TC center and lofted into new convection more efficiently when TCREH is positive. Figure 3.11 shows 6 h forward trajectories for the 667 m positive-TCREH simulation from time $t = 30 – 36$ h and Fig. 3.12 shows forward trajectories for the 667 m negative-TCREH simulation for the same time period. In both cases, the trajectories originate from the lowest model level in regions of large surface latent heat flux. It is apparent when comparing Fig. 3.11a to Fig. 3.12a that parcels wrap around the TC core more efficiently when TCREH is positive. It is also apparent is that parcels tend to be lofted more frequently in the positive TCREH simulation (Fig. 3.11b vs. Fig. 3.12b). Parcel $\Theta_e$ remains larger and parcels experience more diabatic heating during the 6 h period for the positive TCREH simulation than in the negative simulation (Fig. 3.11c, d vs. Fig. 3.12c, d). All of these factors contribute to the faster rate of TC development observed in the positive TCREH simulation.
In addition to the 6 h trajectories shown in Figs. 3.11 and 3.12, forward trajectories are computed for surface (lowest model level) parcels originating in a $33 \times 33$ km box in the upshear left quadrant for both the positive and negative TCREH simulations. This location is chosen because it is just downwind of convection during this time period. Trajectories are spawned from each model grid point in this box so that 2500 trajectories are computed for both simulations. Of the 2500 parcels originating on the lowest model level in this location, 18% are lofted above 2000 m when TCREH is positive vs. only 1.5% when TCREH is negative. Figure 3.13 shows the evolution of parcel altitude vs. time. 5% (0%) of parcels are lofted above 5000 m and 2.5% (0%) are lofted above 10,000 m when TCREH is positive (negative). This again suggests that the process of advecting $\Theta_e$-rich air and relatively large latent heating rate into the TC circulation is more efficient when the environment outside the TC is characterized by positive TCREH.

Another way to consider the impact of TCREH on parcel trajectories is to analyze the parcel paths through convective updrafts. Figures 3.14 and 3.15 show parcel trajectories through updrafts located in the upshear-left quadrants in the 667 m simulations with positive (Fig. 3.14) and negative (Fig. 3.15) TCREH. In the positive TCREH case, the trajectories shown include 3 h of backward trajectories and 3 h of forward trajectories. For the negative TCREH case, the trajectories include 0.5 h of backward trajectories and 5.5 h of forward trajectories. These intervals for forward and backward trajectories were chosen in the negative TCREH case so that a convective feature in the upshear-left quadrant could be examined. Convection in this quadrant was very rare during this time period in the negative TCREH simulation. Both Figs. 3.14 and 3.15 represent the 6 h period from $t = 30 - 36$ h. The first clear difference is that the parcels are lifted higher
(Fig. 3.14b vs. Fig. 3.15b) and curve cyclonically about the TC center (Fig. 3.14a vs. Fig. 3.15a) much more when TCREH is positive. It is also clear that $\Theta_e$ remains larger for the trajectories in the positive TCREH simulation (Fig. 3.14c vs. Fig. 3.15c). $\Theta_e$ in parcels originating on the lowest model level drops from approximately ~$346 \text{ K}$ to ~$344 \text{ K}$ for trajectories in the positive TCREH case while $\Theta_e$ falls from $345 \text{ K}$ to approximately $337 \text{ K}$ for the trajectories in the negative TCREH simulation. This drop likely is related to the entrainment of mid-level air characterized by relatively lower $\Theta_e$. Also, as was the case for surface parcels described earlier, diabatic heating (Figs. 3.14d, 3.15d) is larger for the trajectories in the positive TCREH simulation. An apparent difference between these two sets of trajectories is that the thunderstorm is much deeper in the positive TCREH simulation. Parcels traveling through this storm reach altitudes approaching 10,000 m whereas parcels traveling through the storm in the negative TCREH simulation reach altitudes only around 3800 m. This shallower convective feature in the negative TCREH case was selected because it was one of only two storms in the upshear-left quadrant during this time period and it was closer to the RMW. This is consistent with the fact that convection is much more common and more vigorous in the upshear-left quadrant when TCREH is positive.

The final way in which trajectories are analyzed is by considering the boundary layer $\Theta_e$ recovery rate downwind of convection. We propose that faster boundary layer recovery rates in upshear quadrants promote fresh convective development and faster deepening of the parent TC. Figures 3.16 (positive TCREH simulation) and 3.17 (negative TCREH simulation) show 2 h backward trajectories from locations near the surface (lowest model level) downwind of convection. For both experiments the
trajectories are released at time $t = 38$ h (time $t = 120$ m in Figs. 3.16 and 3.17) and go back to time $t = 36$ h (time $t = 0$ m in Figs. 3.16 and 3.17). In both cases, parcels travel through convection at larger radii en route to their locations in the boundary layer downwind of convection at smaller radii. The primary difference between the two cases is the rate at which $\Theta_e$ recovers during the final stages of this journey. While the recovery rates shown in Figs. 3.16c and 3.17c look similar, the mean rate in the positive TCREH case is $6.4 \text{ Kh}^{-1}$ compared to $4.1 \text{ Kh}^{-1}$ in the negative TCREH simulation. These recovery rates are different even though the initial values of $\Theta_e$ for both sets of trajectories are quite similar. The faster recovery rate likely is due to the fact that wind speeds and surface fluxes are larger in regions downwind of convection when TCREH is positive (Fig. 3.11a vs. Fig. 3.12a). This faster recovery also allows surface parcels to be lifted into new convection sooner and more easily due to increased buoyancy, as was seen above (18% of parcels rising into convection vs. only 1.5% in the negative TCREH case).

### 3.4. Contoured Frequency by Altitude Diagrams

#### 3.4.a Methods

The third method used to analyze the spatial characteristics of fields like simulated reflectivity, diabatic heating rate, and vertical velocity was to compute contoured frequency by altitude diagrams (CFADs: Yuter and Houze 1995). CFADs were generated to determine if the frequency at which convection reached large altitudes differed between simulations with positive or negative TCREH. CFADs also can highlight differences in the convection at a specific altitude. The CFAD results described
below in section 3.4.b were computed over two spatial areas. The first area was a 75 × 75 km box centered on the region of maximum 700 hPa simulated reflectivity. To determine this location, the simulated reflectivity at the 700 hPa level was processed 500 times through a 9-point smoother and the centroid of the smoothed reflectivity was identified. The second spatial location chosen for CFAD computation was in an annulus centered on the TC and with radial range between 25 and 225 km. This radial range was chosen such that the areas of relatively strong convection were encompassed. For both spatial methods, CFADs are shown for simulated reflectivity and vertical velocity.

3.4.b Results

Results from contoured frequency by altitude diagrams (CFADs) show that the characteristic differences between convection (and associated vertical velocities and heat fluxes) in the positive and negative TCREH simulations are not as distinguishing as expected. The primary difference is where the convection is located azimuthally. Figure 3.18 shows CFADs of simulated reflectivity (dBZ) vs. altitude (m) for time \( t = 30 – 36 \) h for the positive TCREH simulation (a, d) and the negative TCREH simulation (b, e). The CFADs are computed using a 75 × 75 km box centered on the region of maximum convection. This region is chosen so that the characteristics of the strongest convection can be directly compared between the two cases. Figure 3.18c shows the difference in reflectivity frequency vs. altitude (positive TCREH case minus negative TCREH case). Figures 3.18a and 3.18b show that the overall characteristics of the convection between the two cases are quite similar. The main difference is that stronger convection (reflectivity > 35 dBZ) reaches higher altitudes when TCREH is \textit{negative}. A key result here is that even though the strongest convection may be deeper in the negative TCREH
simulation, the rate at which minimum central pressure falls remains slower than when TCREH is positive. This highlights the importance of the azimuthal and radial position of the convection relative to the storm motion. Because convection was located farther upshear when TCREH is positive, the associated diabatic heating and larger surface fluxes were closer to the path of the TC allowing for a faster rate of intensification. The results from CFADs of vertical velocity (Fig. 3.18d-f) show that the characteristics of updrafts and downdrafts are similar for the two simulations. Figure 3.18f shows that, except for small magnitudes of vertical velocity (-3 < w < 3 ms\(^{-1}\)), the differences in frequency are less than 1%. Total vertical mass flux in the region of maximum convection (not shown) is larger for the negative TCREH case. However, since this flux occurs at larger radii and downshear-right of the TC, it is less effective at reducing the minimum central pressure.

A second way to compute CFADs is to choose an annulus with radii that encompass all the areas of relatively strong convection. Figure 3.19 shows this annulus within radii 25 and 225 km. As before, the time period of consideration is t = 30 – 36 h which is just prior to when intensities diverge. The primary difference when this annulus is used is that stronger convection is more frequent at all altitudes when TCREH is positive. This highlights the fact that the strongest convection may be slightly stronger in the negative TCREH (Fig. 3.18c) case but overall convective coverage is larger when TCREH is positive (Fig. 3.20c). As was the case when the data was centered on the region of maximum reflectivity, the characteristics of vertical velocity are similar for the two cases outside the range of ±2 ms\(^{-1}\), likely due to the fact that strong updrafts are highly
localized. Total convective mass flux (not shown) was similar for the positive and negative TCREH cases for the annular region.

### 3.5. Sinusoidal Wind Profile Simulations

#### 3.5.a Methods

The primary objective of these simulations was to determine if low-level TCREH was dictating TC-intensification rate or if larger magnitudes of deep-layer TCREH (similar to the simulations described in Chapter 2) are necessary to cause different rates of intensification. Four idealized simulations were performed to assess whether or not environmental helicity in certain vertical layers was more important for TC-intensification and to what extent the deep layer (850 – 200 hPa) TCREH magnitude plays a role. Figure 3.21 shows the vertical profiles of $u$ and $v$ along with the corresponding hodographs for the two simulations where low-level TCREH was positive and upper-level TCREH is negative. The other two simulations were performed with negative TCREH in low levels and positive TCREH in upper levels such that the values of $v$ are equal in magnitude but of opposite sign to the profiles shown in Fig. 3.21. Figure 3.22 compares wind profiles from one positive and one negative low-level TCREH simulation. These simulations can be interpreted, in the green curves on Fig. 3.21 for example, as northeasterly low-level shear and southwesterly upper-level shear. This shear profile might not be very different from those commonly observed in the Tropical North Atlantic. The total value of 850 – 200 hPa TCREH was small in these simulations (4.2 m$^2$s$^{-2}$ (Fig. 3.21a), 7.0 m$^2$s$^{-2}$ (Fig. 3.21b)) due to cancellation between the upper and lower levels. However, 0 – 6 km TCREH magnitude is larger (19.9 m$^2$s$^{-2}$ (Fig. 3.21a),
33.1 m$^2$s$^{-2}$ (Fig. 3.21b)). Note that in all cases the 850 – 200 hPa wind shear vector is identical.

### 3.5.b Results

Results from the four simulations performed with sinusoidal $v$-wind profiles show that TC intensification rate depends on the sign of TCREH only when the magnitude of upper-level TCREH becomes sufficiently large. Figure 3.23 shows a diagram of minimum central pressure and hodographs for the 4 simulations. Table 1 shows intensification rate from $t = 48 – 96$ h for the four simulations. The cases with positive low-level TCREH (negative upper-level TCREH; i.e., the red and black curves in Fig. 3.23) show that development and subsequent intensification occurs only when upper-level TCREH is not too largely negative. While both cases in which low-level TCREH is negative develop into TCs, the case with weaker low-level TCREH develops sooner and more rapidly. These results suggest that TC development becomes less probable as the mid-level meridional shear becomes more extreme, even if the 850 – 200 hPa meridional shear equals zero. Figure 3.24 shows simulated reflectivity at the 700 hPa level for the cases with positive (b and d) and negative (a and c) 850 – 200 hPa (i.e., deep layer) TCREH. It should be noted that a large portion of the mid and upper-levels contains positive TCREH for panels a and c despite the deep layer TCREH being negative (as shown by the green curve in the hodograph in Fig. 3.22). These results suggest that TCREH in the 3500 – 7000 m layer can modulate TC intensification rates. This is somewhat different than the result from reanalysis data found in Chapter 2.3.b where 0 – 3 km TCREH correlated most with TC intensification. This discrepancy may be due to
the very steady nature of the environment in idealized simulations as opposed to the constantly evolving environments around observed TCs.

3.6. Conclusions

TC intensification rate is modulated by the shape of the environmental wind profile which can be measured by TCREH. Time composites of surface latent heat flux and simulated reflectivity show how the maxima of these features rotate cyclonically into upshear quadrants more rapidly when TCREH is positive. Simulations including the addition of uniform background flow demonstrate that certain storm motions relative to the vertical wind shear vector allow for more rapid intensification. These simulations suggest that intensification is most efficient when the TC is moving towards the regions experiencing larger surface latent heat flux or increased diabatic heating. In these cases, there is considerably more overlap of boundary layer convergence and latent heat flux. Trajectories show that the process of advecting buoyant, $\theta_e$-enhanced air into the TC is more efficient and a much higher percentage of parcels are lofted into convection in the upshear quadrants near the RMW when TCREH is positive. They also show that boundary layer recovery is faster downwind of convection when TCREH is positive.

CFAD analyses illustrate the fact that differences in characteristics of convection such as depth of reflectivity between TCs embedded in positive vs. negative TCREH are small. CFADs computed from an annulus centered on the TC, however, show that total convective coverage is slightly larger when TCREH is positive. In Chapter 2, both simulations and reanalysis data showed that positive TCREH favors faster TC intensification. In that chapter it was suggested that the nature of convection in a TC
embedded in positive TCREH was more organized and robust. In the current chapter, further evaluation of TC structure and intensification in environments characterized by varying TCREH suggests that discrepancies in intensification rate primarily result from the ability of convection and associated surface latent heat fluxes to rotate cyclonically into the upshear quadrants. This precession has been shown in previous studies to relate to the onset of rapid intensification and it occurs more rapidly when TCREH is positive.

Results from simulations with sinusoidal vertical wind profiles show that variability in TC intensification rate is possible even when the deep-layer value of TCREH is small due to cancellation. When shear becomes relatively large in the middle to upper levels (3500 – 7000 m) then the sign of TCREH in that layer becomes important with positive TCREH in this layer favoring intensification.
Fig. 3.1. Hodographs for the 667 m simulations with positive (red) and negative (blue) TCREH.
**Fig. 3.2.** Evolution of minimum central pressure for the 667 m simulations with positive (red) and negative (blue) TCREH. The green lines encompass the time period during which much of the composite analysis of section 3.1 occurred. The red (blue) arrow denotes when the mid-level circulation center first advances into the upshear-left quadrant when TCREH is positive (negative).
Fig. 3.3. Time-averaged surface latent heat flux (Wm$^{-2}$) for a simulation with positive TCREH equal to 43 m$^2$s$^{-2}$ (top row: panels a – c) and negative TCREH equal to -43 m$^2$s$^{-2}$ (bottom row: panels d – f). Column 1 (panels a and d) shows surface latent heat flux averaged from 24 – 28 h, column 2 (panels b and e) shows 28 – 32 h, and column 3 (panels c and f) shows 32 – 36 h. These three 4 h time periods correspond to the 12 h period just prior to the divergence of minimum central pressure between the two simulations.
Fig. 3.4. Time-averaged 700 hPa simulated radar reflectivity (dBZ) for a simulation with positive TCREH equal to 43 m²s⁻² (top row: panels a – c) and negative TCREH equal to -43 m²s⁻² (bottom row: panels d – f). Column 1 (panels a and d) shows reflectivity averaged from 24 – 28 h, column 2 (panels b and e) shows 28 – 32 h, and column 3 (panels c and f) shows 32 – 36 h. These three 4 h time periods correspond to the 12 h period just prior to the divergence of minimum central pressure between the two simulations.
Fig. 3.5. Time series of storm-relative azimuth angle of location of maximum convection (a) and radial distance (km) from the TC center to the location of maximum convection (b) for the 667 m simulations with positive (red) and negative (blue) TCREH. Zero degrees is defined as parallel to and in the same direction as TC motion such that angles between 0 and 180 degrees denote a convective maximum to the right of TC motion.
Fig. 3.6. Hodographs for a simulation with no additional background flow added (a) and for a simulation with southwesterly background flow of 4 ms$^{-1}$ (b) added. The black arrow on panel b denotes the additional southwesterly flow and the green arrows denote the resultant storm motion for both simulations.
**Fig. 3.7.** Mini ensemble showing minimum central pressure evolution for 6 positive TCREH simulations with the addition of a vertically constant background flow with magnitude equal to 8 ms\(^{-1}\). The direction of the additional background flow is westerly (blue lines) and easterly (red lines). The 850 – 200 hPa vertical wind shear vector and TCREH magnitude are identical for all 6 cases. Small perturbations to the initial vortex wind field create the differences for the mini ensemble. The green lines encompass the time period during which the composite analysis occurred for the additional background flow discussion.
Fig. 3.8. Minimum central pressure vs. time for simulations with westerly (blue), southwesterly (green), southerly (red), southeasterly (cyan), and easterly (black) additional vertically uniform background flow of 8 ms\(^{-1}\).
Fig. 3.9. Time-averaged 700 hPa simulated reflectivity (dBZ; rows 1 – 2: panels a – f), surface latent heat flux (shaded; Wm^{-2}; rows 3 – 4: panels g – l) and divergence on the lowest model level (contoured, negative only; s^{-1}; rows 3 – 4: panels g – l) for two simulations with positive TCREH and additional background flow with magnitude 8 ms^{-1}: one with easterly background flow (rows 1 and 3) and one with westerly background flow (rows 2 and 4). Column 1 is averaged from $t = 16 – 20\ h$, column 2 is averaged from $t = 20 – 24\ h$, and column 3 is averaged from $t = 24 – 28\ h$. The vectors in the lower left corner of panels a and d represent mean storm motion during the period $t = 16 – 28\ h$ (red) and 850 – 200 hPa environmental wind shear (10 ms^{-1}; blue).
Fig. 3.10. Minimum central pressure (hPa) vs. time (h) for four simulations with additional background flow of 8 ms$^{-1}$. The four simulations are denoted: positive TCREH and westerly flow (solid blue), positive TCREH and easterly flow (solid red), negative TCREH and westerly flow (dashed blue), negative TCREH and easterly flow (dashed red).
Fig. 3.11. 6 h forward trajectories for parcels originating on the lowest model level in a region of large latent heat flux overlaid on 6 h time-averaged surface latent heat flux (Wm$^{-2}$) from $t = 30 – 36$ h (a). Parcel altitude (m) vs. time (b). Parcel equivalent potential temperature (K) vs. time (c). Parcel diabatic heating rate (Kh$^{-1}$) vs. time (d). These trajectories are from the 667 m grid spacing simulation with positive TCREH. The green boxes denote the beginning of the forward trajectories.
Fig. 3.12. 6 h forward trajectories for parcels originating on the lowest model level in a region of large latent heat flux overlaid on 6 h time-averaged surface latent heat flux (Wm$^{-2}$) from $t = 30 – 36$ h (a). Parcel altitude (m) vs. time (b). Parcel equivalent potential temperature (K) vs. time (c). Parcel diabatic heating rate (Kh$^{-1}$) vs. time (d). These trajectories are from the 667 m grid spacing simulation with negative TCREH. The green boxes denote the beginning of the forward trajectories.
Fig. 3.13. 2 h forward trajectories showing altitude (m) vs. time (min) for parcels originating on the lowest model level in the upshear-left quadrant downwind of convection for a simulation with positive (a) and negative (b) TCREH. The green boxes denote the beginning of the forward trajectories.
Fig. 3.14. 6 h trajectories (3 h backward, 3 h forward) for parcels traveling through an upshear-left thunderstorm updraft overlaid on 6 h time-averaged surface latent heat flux (shading: Wm$^{-2}$) and surface pressure (hPa: magenta contours) from $t = 30 \ldots 36$ h (a). Parcel altitude (m) vs. time (b). Parcel equivalent potential temperature (K) vs. time (c). Parcel diabatic heating rate (Kh$^{-1}$) vs. time (d). These trajectories are from the 667 m grid spacing simulation with positive TCREH. The green boxes denote the beginning of the trajectories and the brown boxes denote the endpoints.
**Fig. 3.15.** 6 h trajectories (0.5 h backward, 5.5 h forward) for parcels traveling through an upshear-left thunderstorm updraft overlaid on 6 h time-averaged surface latent heat flux (shading: Wm$^{-2}$) and surface pressure (hPa: magenta contours) from $t = 30 – 36$ h (a). Parcel altitude (m) vs. time (b). Parcel equivalent potential temperature (K) vs. time (c). Parcel diabatic heating rate (Kh$^{-1}$) vs. time (d). These trajectories are from the 667 m grid spacing simulation with negative TCREH. The green boxes denote the beginning of the trajectories and the brown boxes denote the endpoints.
Fig. 3.16. 2 h backward trajectories for parcels originating on the lowest model level downwind of convection at time $t = 38$ h overlaid on 700 hPa simulated reflectivity valid at $t = 38$ h (a). Parcel altitude (m) vs. time (b). Parcel equivalent potential temperature (K) vs. time (c). These trajectories are from the 667 m grid spacing simulation with positive TCREH. The brown boxes denote the ending locations of the backward trajectories.
Fig. 3.17. 2 h backward trajectories for parcels originating on the lowest model level downwind of convection at time $t = 38$ h overlaid on 700 hPa simulated reflectivity valid at $t = 38$ h (a). Parcel altitude (m) vs. time (b). Parcel equivalent potential temperature (K) vs. time (c). These trajectories are from the 667 m grid spacing simulation with negative TCREH. The brown boxes denote the ending locations of the backward trajectories.
Fig. 3.18. Contoured frequency by altitude diagrams (CFADs) of simulated reflectivity (dBZ; panels a – c; x-axis) and vertical velocity (ms$^{-1}$; panels d – f; x-axis) vs. altitude (km; y-axis) for a simulation with positive TCREH (a, d), negative TCREH (b, e), and the difference between the two (c, f). The CFADs are computed for a 75 × 75 km box centered on the region of maximum convection for the time period $t = 30 – 36$ h. Panels a, b, d, and e are contoured with a logarithmic scale such that a frequency of 100% corresponds with a value of 2. Panels c and f show the difference between the frequencies for the two cases (positive TCREH case minus negative TCREH case) on a non-logarithmic scale. The thick black contour on panels c and f denotes a zero percent difference.
Fig. 3.19. Example of the annular area for CFAD computations. The black contour denotes the region in which the CFAD is computed, the magenta contours show smoothed surface pressure, and simulated 700 hPa reflectivity is shaded.
Fig. 3.20. Contoured frequency by altitude diagrams (CFADs) of simulated reflectivity (dBZ; panels a – c; x-axis) and vertical velocity (ms\(^{-1}\); panels d – f; x-axis) vs. altitude (km; y-axis) for a simulation with positive TCREH (a, d), negative TCREH (b, e), and the difference between the two (c, f). The CFADs are computed for an annulus centered on the TC between the radii of 25 and 225 km for the time period \(t = 30 – 36\) h. Panels a, b, d, and e are contoured with a logarithmic scale such that a frequency of 100% corresponds with a value of 2. Panels c and f show the difference between the frequencies for the two cases (positive TCREH case minus negative TCREH case) on a non-logarithmic scale. The thick black contour on panels c and f denotes a zero percent difference.
Fig. 3.21. Vertical profiles of $u$ (blue) and $v$ (red) wind components (ms$^{-1}$) for simulations with positive 850 – 200 hPa environmental helicity: 4.2 m$^2$s$^{-2}$ (a), 7.0 m$^2$s$^{-2}$ (b), and the corresponding hodographs (right panels).
Fig. 3.22. Vertical profiles of $u$ (blue) and $v$ (red = positive low-level TCREH, green = negative low-level TCREH) wind components (ms$^{-1}$) for simulations with weakly positive and negative environmental helicity ($\pm 4.2$ m$^2$s$^{-2}$) (a), and the corresponding hodographs (b).
Fig. 3.23. Evolution of minimum surface pressure (a) for the 4 120 h simulations with varying TCREH and their corresponding 850 – 200 hPa environmental wind hodographs (b).
Fig. 3.24. 700 hPa simulated reflectivity (dBZ) for simulations with TCREH = -7.0 m²s⁻² (a and c) and +7.0 m²s⁻² (b and d) at \( t = 72 \) h (a and b) and \( t = 84 \) h (c and d). The dashed lines denote surface pressure with a contour interval of 4 hPa.
Chapter 4. The Tropical Cyclone Response to Changing Shear Using the Method of Time-Varying Point-Downscaling

4.1. Introductory Remarks

To our knowledge, all previous studies using idealized simulations with time-varying environmental vertical wind shear have imposed an instantaneous change of the background wind field. This typically was accomplished by allowing a TC to spin up in a low-shear or quiescent environment, then the simulation was paused, the mass field rebalanced for the new vertical wind shear, and the simulation was resumed. This method presents the TC with an unrealistic shock when it suddenly encounters shear in all quadrants. Previous studies (Frank and Ritchie 2001; Wang et al. 2004; Riemer et al. 2010, 2013) have used this technique and some (Frank and Ritchie 2001; Riemer et al. 2010) suggested that the instantaneous switch to a new shear regime did not have significant impact on their results. One key result of the Riemer et al. (2010) study was that their simulated TCs recovered to very high intensity even under large-shear (20 ms\(^{-1}\)) conditions. They suggested that this dynamic response played a larger role in TC resiliency to shear than environmental moisture. The results of Riemer et al. (2010, 2013) play a role in motivating the research in this chapter. A new modeling framework is developed in which changes to the environmental wind shear are possible without the instantaneous change of regimes. This method is time-varying point-downscaling (TVPDS).
4.2. The Time-Varying Point-Downscaling Method

In order to allow for smoothly transitioning background states in TC simulations, additional modifications to the WRF source code beyond those made for the PDS technique were required. Idealized analysis nudging is the method by which the background state is varied. A routine was developed to create 3D nudging fields that change in time. Specifically, these 3D environmental states were defined at time intervals (every 6 h, for example) in a four-dimensional input data file read by the WRF model. To create the changing environment, the background state switches during one or more of the time intervals in the input nudging data. Typically the entire change in the environment is defined between two time intervals, however a shift defined smoothly over multiple intervals in the nudging data also achieves a changing background state. These nudging fields then were applied in the WRF model to smoothly transition between background states of wind, temperature, and moisture. In the default distribution of WRF, point-downscaling may be activated via the “pert_coriolis” option. When pert_coriolis is activated, WRF stores the vertical profiles of $u$ and $v$ from grid point (1,1) on the coarsest domain (d01) and this wind profile is defined as the mean state which is subtracted off when applying the Coriolis term in the momentum equations (see Chapter 1 for a more details on the PDS technique). Because of this default behavior, imbalances arise when nudging toward a state that does not match the pert_coriolis vertical wind profile. These imbalances then often lead to undesirable inertial oscillations in the environmental wind field. To overcome this, code was added in the WRF model to update the pert_coriolis profile at each time step to approximately match the profile that results due to the nudging. Nudging in the WRF model follows the
methodology of Stauffer and Seaman (1990, 1991) in which nudging of the environment is defined such that:

\[
\frac{du(z)}{dt} = -\frac{1}{\tau} (u(z) - u_n(z, t)) \tag{4.1}
\]

where \( u(z) \) defines the zonal wind at level \( z \), \( u_n(z, t) \) defines the prescribed nudging value for the zonal wind at level \( z \) for time \( t \), and \( \tau \) is the nudging time scale. A similar equation is applied to compute the time evolution of \( v(z) \). TVPDS applies this formula to update the base environmental profile on which the pert_coriolis feature operates. With this addition, smoothly transitioning background states of temperature, moisture, and wind with doubly periodic boundary conditions are possible. The primary caveat of the TVPDS technique is that, compared to a typical implementation of the WRF model, mass and momentum are conserved to a lesser extent. This limitation is reduced somewhat because all simulations in this dissertation occur on an \( f \)-plane at 20° S where the horizontal temperature gradient required to balance shear of 10 ms\(^{-1}\) over the diameter of a TC is small (\(< 0.5^\circ \text{C}\)). Disallowing the environmental temperature gradient, therefore, compromises conservation less than it would if the simulations were performed at higher latitudes.

The rate at which the environment transitions between environmental wind states depends both on the length of the interval over which the nudging fields change and the value of \( \tau \), the nudging time scale. Figure 4.1 demonstrates how the environment transitions more gradually as \( \tau \) gets larger. The nudging field (blue line) transitions over a 24 h period but the actual simulated response transition period length depends on the
value of $\tau$. By carefully choosing the nudging fields and the value of $\tau$, any shear-transition period can be created.

### 4.3. Results for Various Shear Values and Transition Periods

In an effort to determine whether the instantaneous switch to shear is important in terms of the TC response to changing environmental wind shear, TVPDS simulations intended to mimic those of Riemer et al. (2010, 2013) were performed. These TVPDS simulations initially are run in a very similar framework to those described in Chapters 2 and 3. Tropical cyclones are initialized as moderately intense modified-Rankine vortices with a peak tangential wind speed of 30 ms$^{-1}$. The SST for these simulations is fixed at 27°C. The 850 – 200 hPa environmental vertical wind shear is held constant at 5 ms$^{-1}$ for 264 h until the TC reaches a quasi-steady intensity around 55 ms$^{-1}$. This long period of light shear is chosen so that the TC can reach a steady state before being subjected to changes in environmental shear. At time $t = 264$ h, shear begins a transition to larger values. Four simulations are performed which transition from the initial shear of 5 ms$^{-1}$ to magnitudes of 7.5, 10, 15, and 20 ms$^{-1}$. These values are chosen so that inferences can be made about the TC response to changes in shear magnitude and so that comparisons to Riemer et al. (2010, 2013) can be made. Figure 4.2 shows environmental 850 – 200 hPa wind shear vs. time for the four simulations with smoothly transitioning shear. The time series show that the transitional period is approximately 48 h long although 90% of the transition occurs during the first 24 h of this transition period. Additional simulations were performed with $\tau = 1$ h and these cases transition much faster, requiring only approximately 9 h to complete the transition to larger vertical wind shear (not shown).
Results from the TVPDS simulations described above show that the TC response to increasing shear is a general weakening without a recovery to the pre-change intensity. Figure 4.3 shows minimum central pressure vs. time for the 4 simulations performed with a nudging time scale $\tau = 6$ h. As the final-state environmental shear value becomes larger, the TC weakens both sooner and faster. The TC also reaches a final intensity that is weaker when the final-state shear is larger. Weakening occurs at a steadily faster rate as the shear transition becomes larger. This is clear when comparing the 5 ms$^{-1}$ baseline experiment (Fig. 4.3 blue) to the simulations transitioning to 7.5 and 10 ms$^{-1}$ (cyan and green respectively). However, when the intended shear reaches 15 and 20 ms$^{-1}$, rapid weakening occurs during the first 48 h after shear imposition and the TCs dissipate by the end of the simulation. Figure 4.4 shows simulations with shear transitioning to the same values as in Fig. 4.3 except with a nudging time scale $\tau = 1$ h. The results are very similar despite the much more rapid transition to shear. As in the case when $\tau = 6$ h, there is no recovery after shear is imposed, although the case that transitions to 7.5 ms$^{-1}$ (they cyan line in Fig. 4.4) does demonstrate some modest recovery or a steadying of pressure near the end of the simulation. Comparing the simulations that transition to 10 ms$^{-1}$ (green lines; Figs. 4.3, 4.4) shows that TC weakening occurs more quickly when the nudging time scale decreases. The simulation transitioning to shear of 10 ms$^{-1}$ with $\tau = 6$ h reaches a final intensity of approximately 982 hPa while the corresponding simulation with $\tau = 1$ h weakens further to a final intensity of 992 hPa. This implies that TCs encountering environments characterized by moderate shear are indeed sensitive to the rate at which the transition occurs. This result stands in contrast to the results of Riemer et al. (2010, 2013) which showed a pause in intensification or a slight weakening after the
imposition of shear followed by a recovery to intensity larger than the pre-shear intensity. This was the case in their results even when shear increased to a magnitude of 20 ms\(^{-1}\). Even though the TVPDS simulations above did not impose shear instantaneously as was the case in the Riemer et al. simulations, sensitivity to the transition rate is evident. The lack of recovery also is consistent with observational studies that show a strong negative correlation between environmental shear and TC intensity.

In addition to the simulations described above, numerous simulations (not shown) were performed with higher SST (up to 30° C) and with various environmental lapse rates. Because the environmental temperature and moisture was set to the Dunion moist tropical sounding (Dunion 2011), it was proposed that steeper lapse rates could lead to more TC resiliency in environments of increasing shear. Both of these factors proved, however, not to have a significant influence on TC intensity response to changing vertical wind shear. There were, however, some structural differences (see Chapter 4.4). Final-state intensity was higher when SST was larger, but recovery still did not occur. The rate at which TCs weakened in varying lapse rate environments and with higher SSTs also was very similar to the simulations performed with SST = 27° C. Figure 4.5 shows a time series of minimum central pressure vs. time during the 96 h period after shear imposition for two simulations with environmental vertical shear transitioning from 5 to 15 ms\(^{-1}\): one with SST = 27° C (blue) and one with SST = 30° C (red). The slopes of the two time series are very similar despite the fact that the minimum pressures at shear imposition time are different by 40 hPa. The primary difference in the simulations performed with SST = 30° C is that convection at larger radii in rain bands and outside of the TC itself (not shown) is much more common. This additional convection at larger
radii appears to play a minimal role in dictating changes in intensity as was the case in the simulations discussed in Chapters 2 and 3.

The one exception to this lack of recovery occurs when nudging is turned off after the shear transition has occurred. In this case, a temporary recovery occurs followed by an eventual weakening. Figure 4.6 shows a comparison of two simulations performed with an SST of 30° C and shear transitioning from 5 to 13.5 ms⁻¹. This peak value of environmental shear (13.5 ms⁻¹) was chosen because any values larger than this prevented TC recovery altogether. In the simulation in which nudging is turned off after the shear transition is complete (Fig. 4.6 red), a significant recovery occurs over the 72 h period after nudging ceases. However, the TC does then return to a weakening state and is rapidly weakening at the end of the simulation. When nudging remains active throughout the simulation (Fig. 4.6 blue), a much more stable response to shear occurs and no major weakening or strengthening occurs after the TC reaches a new quasi-steady intensity around 955 hPa. These results suggest that when nudging is deactivated, the TC is able to modify the local environment just outside the TC primary circulation enough to overcome the environmental shear and temporarily reintensify. Though nudging is deactivated after the shear transition is complete, environmental 850 – 200 hPa shear in the large-scale environment does remain steady between 13 and 14 ms⁻¹ because the PDS technique remains active.

The differences between the two simulations shown in Fig 4.6 highlight one of the key aspects of the TVPDS method. The TVPDS technique nudges the environment in a way such that the TC experiences a nearly constant environmental 850- 200 hPa wind shear, with little ability to modify the near-TC environment. This effectively is as if the
TC is experiencing an environment which is being forced to a given shear state. In this situation, when the large-scale is forcing vertical wind shear to be present, the TC circulation may not have a large effect on its environment. It is possible that the recoveries observed in the simulations of Riemer et al. (2010, 2013) were related to the ability of the TC to modify the near-TC environment.

While some of the differences between the TVPDS simulations of this section and the simulations of Riemer et al. (2010, 2013) likely are due to the instantaneous (vs. non-instantaneous) nature of the shear imposition, another cause of the differences may be due to differences in model resolution and physics. The simulations of Riemer et al. were performed using the Regional Atmospheric Modeling System (RAMS), developed at Colorado State University (Pielke et al., 1992; Cotton et al., 2003). Their simulations were performed on an $f$-plane at 15° N with horizontal grid spacing of 5 km. The primary difference between their 2010 and 2013 simulation were several upgrades to physics and surface exchange coefficients. The TVPDS simulations performed as part of this dissertation used a horizontal grid spacing of 2 km. It is likely that the combination of higher resolution, state-of-the-art parameterizations, and a smooth transition between shear regimes leads to a more realistic TC response to changes in vertical wind shear. Because of modeling framework differences it remains very difficult to determine the primary physical cause of the differences between the TVPDS results and those of Riemer et al. (2010, 2013).
4.4. TC Convection and Surface Fluxes in Time-Varying Environments

In order to examine the TC structural response to time-variant shear, time composites of convection, surface fluxes, and low-level equivalent potential temperature ($\theta_e$) are computed during and after the period when shear is imposed. Previous studies noted the importance of diabatic processes in terms of terms of TC resiliency (e.g. Wang et al. 2004). Riemer et al. (2013) discusses the effects that dry mid-level air which is brought into the boundary layer near the TC core has on intensification. They described a process in which persistent shear-induced downdrafts deliver relatively colder air to the boundary layer near the TC core and that surface fluxes are insufficient at raising $\theta_e$ back to previous values. The more stable boundary layer then reduces the potential intensity of the TC. This process is different from the one described by previous studies which identified the mixing of mid-level low-entropy air into the TC as a means by which shear reduces TC intensity (Simpson and Riehl 1958; Cram et al. 2007; Marin et al. 2009; Tang and Emanuel 2012). Figure 4.7 shows time-averaged simulated reflectivity at the 700 hPa level during the 24 h period after shear is imposed for a TVPDS simulation that transitions from 5 to 15 ms$^{-1}$ (corresponding to the red line on Fig. 4.3). The nudging time scale $\tau = 6$ h for this simulation, the SST is 27° C, and the environmental temperature and moisture profiles follow the Dunion moist tropical sounding (Dunion 2011). Immediately evident in Fig. 4.7 is the steady degradation of the symmetry of the TC. By 12 – 16 h after shear is initially imposed (Fig. 4.7b), 700 hPa convection in the southeast quadrant of the simulated TC has been significantly reduced. Interestingly,
despite that fact that convective symmetry is quickly reduced over the first 18 hours after shear imposition, the magnitude of the simulated reflectivity maximum remains approximately constant. During the time period corresponding to the panels of Fig. 4.7, TC minimum central pressure rises from approximately 963 to 978 hPa. This result emphasizes one of the key findings in Chapter 3 which was that the position of the convection relative to TC motion and the deep layer shear vector is more critical than the intensity of the convection. Another interesting result is that the location of the maximum in convection shifts very little in the tangential direction while shear increases. This highlights the fact that with less convective symmetry, there is less total latent heating near the TC core. Figure 4.8 shows time-averaged surface latent heat flux for the same periods shown in Fig. 4.7 ($t = 8 – 32$ h after shear imposition) along with time-averaged convergence on the lowest model level. While the symmetry of the latent heating and convergence are reduced, the peak magnitudes change very little over the 24 h period. The region of stronger surface convergence does appear to shift downshear (eastward) when shear increases. The reduction in symmetry of surface latent heat flux and low-level convergence is similar to what is seen with 700 hPa simulated reflectivity. This supports the idea that symmetric heating within the RMW is critical to maintaining a TC circulation.

To determine if azimuthal mean diabatic heating magnitude changes as environmental shear increases and as the symmetry of convection decreases, radius-height cross sections of time-composited diabatic heating are computed for the same time periods shown in Figs. 4.7 and 4.8. Figure 4.9 shows these time-composites in 4 h time intervals over the 24 period after shear is imposed. There is a steady reduction in azimuthal mean diabatic
heating during this period corresponding to the reduction in the symmetry of convection as visualized by 700 hPa simulated reflectivity (Fig. 4.7). Diabatic heating is particularly reduced above 500 hPa after time $t = 20$ h after shear imposition (Fig 4.7e-f). These time-composites illustrate how TC intensity is closely linked to diabatic heating at or inside the RMW.

Another way to analyze the effects of increasing shear on TCs is to examine low-level equivalent potential temperature ($\theta_e$). Previous studies have identified vertical wind shear induced low-$\theta_e$ downdrafts as a mechanism for weakening TCs (Molinari et al. 2013; Riemer et al. 2010; Tang and Emanuel 2012). Riemer et al. (2013) referred to these cold downdrafts as “anti-fuel” for the tropical cyclone. Figure 4.10 shows a time series of time-averaged equivalent potential temperature on the lowest model level for the same simulation discussed above during the same time periods. Unlike the simulated reflectivity, surface latent heat flux, and low-level convergence, $\theta_e$ on the lowest model level does show relatively larger changes during the period when shear increases and the TC begins significant weakening. Initially, in the 8 – 12 h time range after shear is imposed, lower values of low-level $\theta_e$ are present near the RMW, however, much cooler $\theta_e$ air is flushed into the boundary layer near the TC core in the following periods. This is particularly true by the 16 – 20 h period when a large region of low-$\theta_e$ air arrives in the boundary layer in the upshear quadrants just outside the TC eye near the RMW. This cooler and less buoyant near-surface air restricts updrafts downstream in the southern and eastern quadrants. This effect is evident in the simulated reflectivity fields in Fig. 4.7 which show a steady reduction in convection in these quadrants that are immediately downstream of the lower-$\theta_e$ air which enters the boundary layer. A steady reduction in
eye surface $\theta_e$ also occurs during this period, likely predominantly related to the rising pressure of the TC and this cooling in the eye continues beyond the time periods shown in Fig. 4.10.

To further investigate the role of relatively low equivalent potential temperature boundary layer intrusions, radius-time Hovmöller diagrams of $\theta_e$ on the lowest model level are compared to Hovmöller diagrams of mid level (400 hPa) relative humidity and diabatic heating rate. These levels were chosen in order to compare the relative impacts of mid-level drying and low-level low-$\theta_e$ intrusions since both mechanisms have been proposed to reduce TC intensity when vertical wind shear is present (i.e., mid-level ventilation mechanism described by Simpson and Riehl 1958; Cram et al. 2007; Marin et al. 2009; Tang and Emanuel 2012 vs. the low-level low-$\theta_e$ intrusion (“anti-fuel”) mechanism described by Riemer et al. 2010). Hovmöller diagrams were chosen to determine if one mechanism precedes the other. Figure 4.11 shows azimuthal mean 400 hPa relative humidity (%: a) and diabatic heating rate (Kh$^{-1}$: b) for the 165 h period encompassing the shear transition period. Relatively low-$\theta_e$ intrusions on the lowest model level (Fig. 4.11: black contours) are overlaid to show where these features are located radially and when they occur relative to drying in the mid-levels. Approximately 12 h after shear is imposed ($t = 62$ h in Fig. 4.11), low-$\theta_e$ intrusions in the boundary layer appear near the RMW. Later, after the environmental shear transition is completed, very dry 400 hPa air is present at the TC center (Fig. 4.11a) and subsequent low-$\theta_e$ downdrafts in the boundary layer appear at the RMW and emanate radially outward in time. Periodic switches between dry and moist air in the TC core at 400 hPa then occur, each associated with a period of relatively stronger diabatic heating (Fig. 4.11b) and low-$\theta_e$ downdrafts in
the boundary layer. It remains difficult to assess which factor drives the other in terms of low-level $\theta_e$ and mid-level dry air. While low-level low-$\theta_e$ intrusions appear first right before the TC weakening process begins (Fig. 4.11a: $t = 62$ h), future intrusions of low-$\theta_e$ appear to be directly tied to the drying events in the 400 hPa relative humidity field near the TC core. The relative contribution of each of these factors remains a subject for future research.

One unique process in this simulated TC is the fact that simulated reflectivity and surface latent heat flux maxima remain in the upshear quadrants despite the fact that shear increases to 15 ms$^{-1}$. Numerous studies have shown the preference for convection to proceed to or remain in the downshear quadrants (particularly the downshear left quadrant) when a TC encounters shear. The fact that this simulated TC weakens even while convection and flux maxima remain in the upshear quadrants supports the theory that relatively cooler $\theta_e$ air flushed into the boundary layer is a leading cause of weakening in sheared TCs. This also highlights the importance of sophisticated microphysical schemes for cloud and precipitation processes that play an important role in driving changes in equivalent potential temperature. This may be one of the key differences between the TVPDS simulations performed as part of this dissertation and the simulations of Riemer et al. (2010, 2013).

### 4.5. Time-Varying Shear and TC Tilt

In addition to affecting the distribution of convection and boundary layer equivalent potential temperature, vertical wind shear effects the vertical alignment of TCs. Numerous previous studies (e.g., Marks et al. 1992; Franklin et al. 1993; Jones 1995,
2000; DeMaria 1996; Corbosiero and Molinari 2002; Molinari et al. 2006; Reasor et al. 2013) outlined the impact of shear on TCs in terms of tilt and convective distribution. These studies identified potential ways in which tilt reduced TC symmetry and intensity. They noted that TCs tilt downshear and favor convection in the downshear-left quadrant. These studies also pointed out how stability near the TC core can be reduced by tilting. Reasor and Montgomery (2001) noted how TC tilt and subsequent precession can modify the rate at which TCs intensify or weaken. To examine the extent to which tilt and precession effect TC resiliency and the rate of weakening in the TVPDS simulations, time series of tilt were generated for the simulation transitioning from 5 to 15 ms\(^{-1}\) (Fig 4.3 red). Figure 4.12 shows the tilt of the vorticity centroid vs. time for the 144 h period after shear is imposed. Smoothed vorticity centroids are computed at 850 and 400 hPa and then the horizontal distance between the centroids is identified hourly. Initially, TC tilt is near zero. Tilt then steadily increases to a value of approximately 50 km by time \(t = 54\) h after shear imposition. At this point a notable precession pattern develops and continues through the end of the simulation. Interestingly, this precession corresponds with minor fluctuations in the minimum central pressure as shown by the red line of Fig. 4.3. Whenever the upper-level (400 hPa) vorticity centroid precesses near or upshear of the 850 hPa centroid the TC minimum pressure stabilizes or decreases slightly, however, the overall trend of weakening continues. As the TC becomes weaker, the general trend is for the tilt to continue to increase through 96 h when tilt reaches 235 km. By this time the TC has weakened significantly to a minimum central pressure of approximately 1000 hPa. While it does appear that the precession process acts to reduce the impact of the vertical shear (15 ms\(^{-1}\)), the large magnitude of shear disallows any significant recovery.
Figure 4.13 shows time series of the 850 – 400 hPa vorticity centroid tilt vs. time for all 4 of the TVPDS simulations transitioning into larger shear regimes with a constant SST = 27° C (from 5 to 7.5 ms⁻¹: cyan, to 10 ms⁻¹: green, to 15 ms⁻¹: red, to 20 ms⁻¹ : black). The time series show that this large tilt and precession process occurs only when shear increases to greater than 10 ms⁻¹. When shear increases to only 10 ms⁻¹ or less, the TC resists tilting and precession with 850 – 400 hPa tilt remaining less than 15 km for the duration of the simulation. Despite the lack of larger tilt, minimum pressure does rise in both of these cases (Fig. 4.3). This again suggests that the asymmetry of convection and the flushing of lower equivalent potential temperature air into the boundary layer near the RMW are the primary reasons for the weakening of the TC. When shear increases to 20 ms⁻¹, both tilt and precession begin sooner, however, the precession frequency (i.e., revolutions per day of the upper-level vorticity centroid about the low-level centroid) is very similar to the case where shear increases to 15 ms⁻¹. The precession rates in the TVPDS simulations are approximately \(8.28 \times 10^{-4}\)s⁻¹ (precession period = 23.0 h) compared to the results of Reasor and Montgomery (2015) who found slightly faster precession rates between 5.5 and \(6.4 \times 10^{-4}\)s⁻¹ (precession period = 15.2 – 17.8 h) from observation-based experiments. The simulation transitioning to 20 ms⁻¹ also experiences fluctuations in minimum central pressure corresponding to the precession process as was the case in the 15 ms⁻¹ simulation (Fig. 4.3).

Another factor reducing tilt during shear imposition is SST. 850 – 400 hPa tilt computed from a simulation performed with SST = 30° C and shear transitioning from 5 to 15 ms⁻¹ shows that the TC can resist tilting. This was not the case when SST = 27° C. Despite the fact that tilt is resisted, minimum central pressure steadily rises throughout
the simulation. Figure 4.14 shows 850 – 400 hPa vorticity centroid tilt vs. time for three simulations that transition from shear of 5 to 15 ms\(^{-1}\) (SST = 27° C red, SST = 28.5° C blue, SST = 30° C black). Comparing the three time series, it is evident that tilt increases with decreasing SST. Precession is quite evident when the SST is 27° C, somewhat evident when SST is 28.5° C, and essentially absent when SST is 30° C. Despite these significant differences in tilt and precession, the general weakening rates in terms of minimum pressure is similar for the three cases. Equivalent potential temperature cooling rate on the lowest model level inside a radius of 40 km is computed for these 3 simulations over the first 24 h after shear imposition. This cooling rate equals 0.05 Kh\(^{-1}\), 0.04 Kh\(^{-1}\), and 0.02 Kh\(^{-1}\) for the simulations with SST = 27° C, 28.5° C, and 30° C respectively. When these rates are computed over the radial range 30 – 60 km, \(\theta_e\) actually warms slightly as shear increases when SST = 28.5 and 30° C, possibly due to the significant reduction in tangential wind speed and the associated reduction in evaporation. The lack of a clear relationship between boundary layer \(\theta_e\) cooling rate near the RMW and TC intensity change suggests that additional factors (i.e., reduced diabatic heating and relative humidity in the mid-levels) play an important role in dictating TC weakening rate.

### 4.6. Transitioning into Varying TCREH Environments

To connect the concepts of Chapters 2 and 3 with the TVPDS simulations of the present chapter, simulations were performed to investigate the effect of TC environments that transition to varying states of tropical cyclone-relative environmental helicity (TCREH). Six simulations were performed that transition from quiescent (no flow) wind
environments to the background states that were used to simulate TCs in varying TCREH environments in Chapter 2 (Fig. 2.4). Figure 4.15 shows the first set of three simulations which were performed with a nudging time scale $\tau = 12$ h with a transition interval of 24 h to allow for a gradual transition into the varying TCREH environments (one with TCREH = 10.5 m$^2$s$^{-2}$: blue, one with TCREH = 0 m$^2$s$^{-2}$: yellow, and one with TCREH = -10.5 m$^2$s$^{-2}$: red). The hodographs in the bottom row of Fig. 4.15 show the final-state 850 – 200 hPa environmental wind (clockwise curving, positive TCREH: blue; straight, zero TCREH: yellow; and counterclockwise curving, negative TCREH: red). The time series demonstrate that gradually transitioning into different states of TCREH has a limited effect on the final-state TC intensity, with positive or zero TCREH being slightly more favorable for TC resiliency.

This result changes when the transition to different values of TCREH occurs more rapidly. Figure 4.16 shows three simulations in which TCREH transitions to the same final states except that the nudging time scale is much shorter ($\tau = 2$ h). In these cases, there is more delineation in terms of TC intensity. When $\tau = 2$ h, the TC in the simulation transitioning to a negative TCREH environment gradually dissipates while the cases transitioning to zero or positive TCREH weaken at a slower rate. In the case when the nudging time scale is larger ($\tau = 12$ h), the transition occurs over a 48 h period compared to approximately 30 h when the nudging time scale is shorter ($\tau = 2$ h). In both of the sets of three simulations the 850 – 200 hPa environmental wind shear vector (as opposed to TCREH) at any given time is approximately equal for all three cases.

Two primary conclusions can be drawn from these experiments transitioning to varying TCREH background states. First, the rate at which the environmental wind state
changes plays a role in dictating future TC intensity evolution. This was shown to be the case in the simulations of section 4.3 in which the weakening rate was slightly faster when the nudging time scale was shorter. This result again implies that the instantaneous transition to shear which was imposed by previous studies was at least partially responsible for the unrealistic TC response. The second conclusion is that, as was shown in Chaps. 2 and 3, positive (or non-negative) TCREH is more favorable for TC intensification than negative TCREH. This remains the case even when TCREH evolves in a transitory nature, as shown in Fig. 4.15 and 4.16.

4.7. The TC Response to the Reduction of Environmental Wind Shear

To determine if the opposite effect occurs when environmental wind shear is reduced, simulations were performed in which a quasi-steady state TC in relatively strong wind shear is transitioned smoothly into an environment of less shear. In this simulation, the TC is allowed to develop for 234 hours until the quasi-steady state is reached. Figure 4.17 shows a time series for a simulation with shear transitioning from 15 to 5 ms\(^{-1}\). The blue line represents minimum central pressure for the control simulation in which environmental shear remains at 15 ms\(^{-1}\) and the red line shows minimum pressure for the simulation with shear reduced to 5 ms\(^{-1}\) over a 36 h period. As in the experiments described in previous sections, approximately 90% of the transition in shear occurs during the first 24 h after the shear transition is initiated. The transition of TC intensity takes longer, approximately 60 h. For these simulations the SST = 28.5° C and the nudging time scale \(\tau = 6\) h. The slightly higher SST (28.5 ° C as opposed to 27° C) is
chosen so that a moderately intense quasi-steady state TC (category 1 hurricane: minimum central pressure = 990 hPa and maximum sustained 10 m wind speed = 40 ms\(^{-1}\)) is possible despite environmental shear of 15 ms\(^{-1}\). In the control simulation, minimum central pressure fluctuates but remains in the 985 – 995 hPa range. In the case in which shear is reduced, intensity steadily increases reaching a value of 943 hPa at the end of the simulation.

Similarly to the cases described in which shear increases, the convective symmetry, boundary layer equivalent potential temperature, and TC tilt provide insight into the reasons why the TC intensifies as shear decreases. Figure 4.18 shows a sequence of time-averaged 700 hPa simulated reflectivity for 6 h periods during the 36 h period after shear begins to decrease. Figure 4.18a shows a very asymmetric system in the presence of environmental wind shear of nearly 15 ms\(^{-1}\). Despite the asymmetry and strong shear, the region of maximum convection in the downshear left quadrant is relatively intense and widespread compared to the subsequent 5 periods. However, the asymmetric nature and radial distance of the convection from the center makes it less effective at intensifying the TC via latent heat release. Over the following time-averaged periods symmetry steadily increases, first encircling the eye in the \(t = 18 – 24\) h period after shear transition begins. By the final period shown in Fig. 4.18d \((t = 30 – 36\) h), strong convection fully encompasses the eye and the TC has acquired a much more symmetric appearance with slightly more convection upshear (west) of the center. Despite the symmetry, the magnitude of the strongest convection is no more intense than when shear is initially imposed. This result is similar to those from the time-averaged convection analysis and contoured frequency by altitude diagrams (CFADS) shown in Chap. 3 which
showed that the distribution and tangential location was much more critical to intensification than was the magnitude.

Time-composites of azimuthal mean diabatic heating rate show a steady increase in diabatic heating as shear decreases. Figure 4.19c shows the first period \((t = 12 – 18 \, \text{h after shear imposition})\) in which diabatic heating in the mid-levels near the RMW significantly increases. After a brief reduction in heating in the subsequent 6 h period, diabatic heating again increases significantly in the TC eyewall during the final two periods (Fig. 4.19e-f). The relationship between 700 hPa convective symmetry (Fig. 4.18) and azimuthal mean diabatic heating (Fig. 4.19) is similar to that seen when shear increases. In the case of decreasing (increasing) shear, the symmetry of simulated reflectivity and the azimuthal mean diabatic heating both increase (decrease).

Time-averaged plots of equivalent potential temperature \((\theta_e)\) on the lowest model level show that relatively more stable air near the RMW becomes steadily less widespread as shear decreases. Figure 4.20 shows time-averaged plots of near-surface \(\theta_e\) for 6 h intervals during the 36 h period after shear begins to decrease. Initially, in the first 6 h when shear still is near 15 ms\(^{-1}\), a large region of low-\(\theta_e\) air in the boundary layer exists in the upshear quadrants, maximized in the upshear-left quadrant. Throughout the following periods (Fig. 4.20 b-f) there are bouts of relatively cooler \(\theta_e\) entering the boundary layer but the magnitude of these intrusions is reduced and their radial distance from the TC core increases. The combination of these factors is to reduce the effect of these cool \(\theta_e\) intrusions on TC intensification and to allow more buoyant air to be present at the RMW. \(\theta_e\) in the TC eye steadily increases during this 36 h period as the TC responds to the reduction in shear and intensifies.
Radius-time Hovmöller diagrams are computed for azimuthal mean $\theta_e$ on the lowest model level and are compared to Hovmöller diagrams of azimuthal mean mid level (400 hPa) relative humidity and diabatic heating rate. Figure 4.21 shows the diagrams for the 115 h period encompassing the shear transition period. Low-$\theta_e$ intrusions are contoured in black in 1 K intervals from 340 – 343 K. The relationship between low-$\theta_e$ intrusions into the boundary layer and 400 hPa relative humidity (Fig. 4.21a) is less clear than in the case when shear increases (Fig. 4.11), however, the frequency of low-$\theta_e$ intrusions is reduced and mid-level relative humidity increases as shear decreases. In a similar fashion, 400 hPa diabatic heating rate near the RMW increases when shear is reduced (Fig. 4.21b).

TC 850 – 400 hPa vorticity centroid tilt decreases when shear is reduced, with tilt dropping to less than 5 km within 30 h of the time at which shear initially begins to decrease (Fig. 4.22). Due to the higher SST of 28.5° C, even the strongly sheared storm is tilted only by an average of approximately 25 km throughout the control simulation. There are more high-frequency fluctuations in tilt when shear remains at 15 ms$^{-1}$, however the magnitude of tilt never exceeds 45 km. The smooth transition of the TC to a vertical alignment corresponds to the symmetrization of the 700 hPa simulated reflectivity and the reduction in cool $\theta_e$ intrusions into the boundary layer near the TC core. No precession is apparent in either the control case or the simulation in which shear decreases.

A final set of simulations were performed in which shear decreases from 15 to 5 ms$^{-1}$ at different rates. Figure 4.23 shows the magnitude of vertical wind shear (a) and the minimum central pressure (b) vs. time for these simulations. The red, green, and blue
lines represent simulations in which a transition interval of 3 h is combined with a nudging time scale $\tau = 12$ h, 6 h, and 1 h respectively. The black line represents the simulation with a transition interval of 12 h and $\tau = 6$ h. The effect of the shear transition rate has a limited effect for these cases when shear decreases. During the period $t = 12 – 36$ h while shear is undergoing transition, the minimum central pressure is progressively lower for cases that transition to smaller magnitudes of shear more quickly. For example, minimum pressure falls to approximately 968 hPa by $t = 24$ h after shear begins decreasing when $\tau = 1$ h (Fig. 4.23 blue) compared to 983 hPa at the same time for the simulation with $\tau = 12$ h (Fig. 4.23 red). The differences are modest after time $t = 42$ h when most or all of the shear transition is complete for each simulation. Interestingly, there is some divergence of minimum pressure near the end of the 96 h simulations, with TCs that encountered a faster shear transition (e.g., $\tau = 1$ h) ending up in a weaker state than those experiencing a slower transition (e.g., $\tau = 12$ h). Because this trend is reversed when comparing the two cases with $\tau = 6$ h and nudging intervals of 3 h (green) and 12 h (black), it is difficult to determine if the difference in end-state intensities is significant.

### 4.8. TVPDS Summary and Proposed Future Experiments

A new method has been developed to allow smoothly transitioning environments of vertical wind shear. Analysis nudging and point-downscaling work together to accomplish the transition between shear regimes. The time-varying point downscaling simulations in this chapter demonstrate that smoothly transitioning environmental wind shear produces a more realistic TC response than does an instantaneously changing shear environment. The simulations show that TC resiliency increases when the transition to
larger shear occurs more gradually. TC resiliency also is a function of SST, with higher SST allowing the TC to resist tilt and remain more intense. While surface latent heat flux distribution and symmetry may play a role in dictating the TC response to increasing shear, 700 hPa convective asymmetry and relatively low boundary layer equivalent potential temperature intrusions near the RMW play an important role in reducing TC intensity. Time-composites of these features demonstrate a clear transition from symmetrical convection without significantly low-\(\theta_e\) air in the boundary layer to an asymmetric convective distribution and low-\(\theta_e\) intrusion near the RMW as shear increases. Time-composites of azimuthally averaged diabatic heating rate show that the heating is significantly reduced as shear increases and as convective symmetry is reduced. Radius-time Hovmöller diagrams show that low-\(\theta_e\) intrusions increase just prior to the beginning of the reduction in TC minimum central pressure. These diagrams also show significant mid-level drying events in the TC core after shear increases and associated low-\(\theta_e\) downdrafts in the boundary layer.

TC tilt demonstrates a clear increase with increasing environmental wind shear, similar to the results of previous studies. When sufficiently low SST and large shear are combined, a notable precession occurs with corresponding fluctuations in minimum central pressure. However, the general trend for TCs to weaken in shear greater than 10 ms\(^{-1}\) prevails. In addition to the final-state shear magnitude in TVPDS simulations, the shape of the final-state vertical wind profile also is important in dictating the TC response. Simulations of TCs transitioning into environments characterized by positive or zero TCREH show that these regimes are more favorable for TC resiliency than those transitioning into negative TCREH. The rate at which the transition occurs in these cases
is important with slower transition leading to less delineation between final-state intensities of TCs embedded in positive vs. negative TCREH.

Finally, TCs that transition into environments characterized by a reduced magnitude of vertical wind shear show a steady increase in organization and intensity. Time-composites of diabatic heating rate show an increase in heating as shear is reduced. Radius-time Hovmöller diagrams show a reduction in low-$\theta_e$ intrusions into the boundary layer and an increase in mid-level relative humidity when shear decreases. Each metric that showed a reduction in symmetry or cooler $\theta_e$ in the near-core boundary layer when shear increased shows the opposite trend when shear is reduced.

The results of the simulations described in this chapter motivate potential additional experiments in which other environmental factors are transitioned from one state to another. For example, previous studies have pointed out the complex TC response to interactions between environmental wind shear and moisture (e.g., Ge et al. 2013). Simulations transitioning between shear states in different values of environmental relative humidity could illuminate these interactions. The TVPDS technique also allows for simulations with constant shear but smoothly transitioning values of environmental moisture, or even time-evolving environmental wind and moisture simultaneously. There is potential to gain further insight into the TC intensification process using the TVPDS modeling framework.
Fig 4.1. Response of environmental $u$ wind component to different values of $\tau$, the nudging time scale, in a TVPDS simulation with shear transitioning from 5 to 15 ms$^{-1}$. The nudging interval in each case is 24 h. The blue line represents the nudging field toward which the simulated $u$ wind field is nudged.
Fig. 4.2. Environmental 850 – 200 hPa wind shear (ms⁻¹) vs. time (h) for simulations with shear transitioning from 5 ms⁻¹ to 7.5 (cyan), 10 (green), 15 (red), and 20 (black) ms⁻¹. $\tau$ for these simulations is 6 h.
Fig. 4.3. Minimum central pressure (hPa) vs. time (h) for TVPDS simulations with environmental wind shear transitioning from 5 ms\(^{-1}\) (blue) to 7.5 ms\(^{-1}\) (cyan), 10 ms\(^{-1}\) (green), 15 ms\(^{-1}\) (red), and 20 ms\(^{-1}\) (black). SST for these simulations is 27\(^\circ\) C and \(\tau = 6\) h.
Fig. 4.4. Minimum central pressure (hPa) vs. time (h) for TVPDS simulations with environmental wind shear transitioning from 5 ms$^{-1}$ (blue) to 7.5 ms$^{-1}$ (cyan), 10 ms$^{-1}$ (green), 15 ms$^{-1}$ (red), and 20 ms$^{-1}$ (black). SST for these simulations is 27° C and $\tau = 1$ h.
Fig. 4.5. Minimum central pressure (hPa) vs. time (h) for TVPDS simulations with environmental wind shear transitioning from 5 m\(s^{-1}\) to 15 m\(s^{-1}\) with a SST = 27° C (blue) and a SST = 30° C (red). The value of \(\tau = 6\) h for both simulations.
Fig. 4.6. Minimum central pressure vs. time (h) for two simulations performed with shear increasing from 5 to 15 ms$^{-1}$ over a 24 h period ($t = 0 – 24$ h). The red line represents a simulation in which nudging is turned off after $t = 24$ h and the blue line represents the case when nudging remains on for the duration of the simulation. In both cases, when nudging is active, the nudging time scale $\tau = 6$ h.
Fig. 4.7. Time-averaged 700 hPa simulated radar reflectivity (dBZ) for a simulation with shear increasing from 5 to 15 ms\(^{-1}\) over the periods displayed. Row 1 shows reflectivity averaged from 08 – 12 h (a), 12 – 16 h (b), and 16 – 20 h (c) after shear begins increasing. Row 2 shows time-mean reflectivity from 20 – 24 h (d), 24 – 28 h (e), and 28 – 32 h (f) after shear imposition.
Fig. 4.8. Time-averaged surface latent heat flux (color-filled; Wm\(^{-2}\)) and divergence (black contours, negative only; s\(^{-1}\)) for a simulation with shear increasing from 5 to 15 ms\(^{-1}\) over the periods displayed. Row 1 shows latent heat flux and convergence averaged from 08 – 12 h (a), 12 – 16 h (b), and 16 – 20 h (c) after shear begins increasing. Row 2 shows time-mean latent heat flux and convergence from 20 – 24 h (d), 24 – 28 h (e), and 28 – 32 h (f) after shear imposition.
**Fig. 4.9.** Time-composites of azimuthal mean diabatic heating rate (K h⁻¹) vs. pressure (hPa) for a simulation with shear increasing from 5 to 15 ms⁻¹ over the periods displayed. Row 1 shows diabatic heating rate averaged from 08 – 12 h (a), 12 – 16 h (b), and 16 – 20 h (c) after shear begins increasing. Row 2 shows the heating rate from 20 – 24 h (d), 24 – 28 h (e), and 28 – 32 h (f) after shear imposition.
Fig. 4.10. Time-averaged equivalent potential temperature ($\theta_e$; K) on the lowest model level for a simulation with shear increasing from 5 to 15 ms$^{-1}$ over the periods displayed. Row 1 shows $\theta_e$ averaged from 08 – 12 h (a), 12 – 16 h (b), and 16 – 20 h (c) after shear begins increasing. Row 2 shows time-mean $\theta_e$ from 20 – 24 h (d), 24 – 28 h (e), and 28 – 32 h (f) after shear imposition. For each period, the minimum $\theta_e$ value is displayed in the upper right corner of the panel.
Fig. 4.11. Radius-time Hovmöller diagram of 400 hPa relative humidity (a: %) and diabatic heating rate (b: K h$^{-1}$) for the 165 h time period encompassing the shear transition period for a simulation with shear transitioning from 5 to 15 ms$^{-1}$. The black contours represent regions of relatively low equivalent potential temperature on the lowest model level (K; contoured at 1 K intervals from 342 to 345 K). The black arrows indicate the beginning time and approximate end time of the shear transition period.
Fig. 4.12. 850 – 400 hPa TC tilt (km) as measured by the relative vorticity centroids for a TVPDS simulation transitioning from 5 to 15 ms$^{-1}$. This simulation corresponds with the red line of Fig. 4.3.
Fig. 4.13. 850 – 400 hPa TC tilt (km) as measured by the relative vorticity centroids for TVPDS simulations transitioning from 5 to 7.5 ms$^{-1}$ (cyan), 5 to 10 ms$^{-1}$ (green), 5 to 15 ms$^{-1}$ (red), and 5 to 20 ms$^{-1}$ (black). These simulations correspond to the simulations shown in Fig. 4.3.
Fig. 4.14. 850 – 400 hPa TC tilt (km) as measured by the relative vorticity centroids for TVPDS simulations with environmental shear transitioning from 5 to 15 ms$^{-1}$ with SST = 27° C (red), SST = 28.5° C (blue), and SST = 30° C (black). The nudging time scale $\tau = 6$ h for each simulation. These red line corresponds to the red line shown in Figs. 4.3 and 4.11.
Fig. 4.15. Evolution of minimum surface pressure (top), evolution of the magnitude of 850 – 200 hPa vertical wind (middle), and “End State” hodographs (bottom) for 3 simulations performed with time-varying wind shear. A nudging time scale of 12 h was used with a transition interval of 24 h. The blue curve on each figure corresponds to a simulation that transitions from no background flow to one characterized by positive environmental helicity (TCREH = 10.5 m²s⁻²). The yellow curve transitions to a state with zero environmental helicity (TCREH = 0 m²s⁻²), and the red curve transitions to a state with negative environmental helicity (TCREH = -10.5 m²s⁻²).
Fig. 4.16. Evolution of minimum surface pressure (top), evolution of the magnitude of 850 – 200 hPa vertical wind (middle), and “End State” hodographs (bottom) for 3 simulations performed with time-varying wind shear. A nudging time scale of 2 h was used with a transition interval of 24 h. The blue curve on each figure corresponds to a simulation that transitions from no background flow to one characterized by positive environmental helicity (TCREH = 10.5 m²s⁻²). The yellow curve transitions to a state with zero environmental helicity (TCREH = 0 m²s⁻²), and the red curve transitions to a state with negative environmental helicity (TCREH = -10.5 m²s⁻²).
Fig. 4.17. Minimum central pressure (hPa) vs. time (h) for a TVPDS simulation with environmental wind shear transitioning from 15 to 5 ms\(^{-1}\) and a control experiment in which shear remains at 15 ms\(^{-1}\) throughout. SST for these simulations is 28.5° C and \(\tau = 6\) h.
Fig. 4.18. Time-averaged 700 hPa simulated radar reflectivity (dBZ) for a simulation with shear decreasing from 15 to 5 ms$^{-1}$ over the periods displayed. Row 1 shows reflectivity averaged from 0 – 6 h (a), 6 – 12 h (b), and 12 – 18 h (c) after shear begins decreasing. Row 2 shows time-mean reflectivity from 18 – 24 h (d), 24 – 30 h (e), and 30 – 36 h (f) after the shear begins decreasing.
Fig. 4.19. Time-composites of azimuthal mean diabatic heating rate (K$h^{-1}$) vs. pressure (hPa) for a simulation with shear decreasing from 15 to 5 ms$^{-1}$ over the periods displayed. Row 1 shows diabatic heating rate averaged from 0 – 6 h (a), 6 – 12 h (b), and 12 – 18 h (c) after shear begins decreasing. Row 2 shows the heating rate from 18 – 24 h (d), 24 – 30 h (e), and 30 – 36 h (f) after shear imposition.
**Fig. 4.20.** Time-averaged equivalent potential temperature ($\theta_e$; K) on the lowest model level for a simulation with shear decreasing from 15 to 5 ms$^{-1}$ over the periods displayed. Row 1 shows $\theta_e$ averaged from 0 – 6 h (a), 6 – 12 h (b), and 12 – 18 h (c) after shear begins decreasing. Row 2 shows time-mean $\theta_e$ from 18 – 24 h (d), 24 – 30 h (e), and 30 – 36 h (f) after the shear begins decreasing. For each period, the minimum $\theta_e$ value is displayed in the upper right corner of the panel.
Fig. 4.21. Radius-time Hovmöller diagram of 400 hPa relative humidity (a: %) and diabatic heating rate (b: Kh⁻¹) for the 115 h time period encompassing the shear transition period for a simulation with shear transitioning from 15 to 5 ms⁻¹. The black contours represent regions of relatively low equivalent potential temperature on the lowest model level (K; contoured at 1 K intervals from 340 to 343 K). The black arrows indicate the beginning time and approximate end time of the shear transition period.
Fig. 4.22. 850 – 400 hPa TC tilt (km) as measured by the relative vorticity centroids for a TVPDS simulation with environmental shear transitioning from 15 to 5 ms$^{-1}$ (red) and a control experiment with environmental shear of 15 ms$^{-1}$ throughout (blue). SST = 28.5° C and the nudging time scale $\tau = 6$ h for both simulations.
Fig. 4.23. 850 – 200 hPa vertical wind shear (a: ms$^{-1}$) and TC minimum central pressure (b: hPa) vs. time (h) for simulations transitioning from shear of 15 to 5 ms$^{-1}$. The red, green, and blue lines represent simulations in which a transition interval of 3 h is combined with a nudging time scale $\tau = 12$ h, 6 h, and 1 h respectively. The black line represents the simulation with a transition interval of 12 h and $\tau = 6$ h. SST = 28.5 °C for these simulations.
Chapter 5. Summary and Future Work

While environmental wind shear modulates the intensity of tropical cyclones, specifics about the vertical wind profile also dictate TC structural and intensity changes. Deep layer (850 – 200 hPa) vertical wind shear tends to tilt the TC in the direction of the shear and causes asymmetries in the convection and surface fluxes, typically with a preference for maximum convection to occur in the downshear-left quadrant. This general rule may not accurately describe the TC response to variations in the vertical wind profile, however, particularly when the magnitude of environmental helicity becomes large. In this dissertation a new quantity was identified which describes the degree to which there is turning with height of the environmental wind field with respect to tropical cyclone motion: tropical cyclone-relative environmental helicity (TCREH). The sign of TCREH proves to be a useful predictor of TC intensity change and details about TC tilt and structure also are related to TCREH. A set of simulations were performed in the Weather Research and Forecasting Model (WRF) using the point-downscaling method in combination with analysis nudging. This modeling framework allowed for nearly constant environmental states of temperature, moisture, and wind shear with doubly periodic boundary conditions. The simulations identified a clear signal between the shape of the environmental hodograph and subsequent TC intensification rate with positive TCREH favoring faster intensification and negative TCREH leading to slower intensification or weakening.

TCREH computed in annuli around TCs in the Tropical Atlantic from reanalysis data from the period 1971 – 2010 also showed a correlation between TCREH and TC
intensification rate. Correlations were modest ($r \sim 0.35$) with positive TCREH favoring intensification and negative TCREH favoring steady-state TC intensity or weakening. Reanalysis data suggest that TCREH affects TC intensity when comparisons are made over longer time scales of 96 – 156 h. This difference from the rapid and larger changes in the model simulations is due to the nearly constant state of the environment which was imposed by the point-downscaling and analysis nudging techniques. In nature, the environment around a TC is always changing and thus relationships between TCREH and intensity only appear when data are averaged over longer time periods. For example, it was quite rare for TCREH to remain of constant sign for periods longer than 72 h. However, despite the sign of TCREH changing in most of the longer time periods, correlation still exists between the average value during the period and intensity change during the corresponding period. The results of Chapter 2 suggest that the magnitude of the 850 – 200 hPa wind shear alone is not optimal for diagnosing the favorability of the kinematic environment around a TC, and that the shape of the vertical profile of wind as measured by TCREH also should be considered.

After identifying a relationship between TCREH and TC intensity, pathways were identified through which environmental helicity modulates TC structure and intensity. While subtle differences in the nature of the strongest convection existed between positive and negative TCREH cases, the local convective environment within the TC that results from the sign of TCREH did not play a critical role in dictating intensification rate. Local-scale helicity near the regions of maximum convection was larger when TCREH was positive, however, these regions typically were at radii outside the RMW where latent heat release is less effective at reducing the TC minimum central pressure.
The contoured frequency by altitude diagrams (CFADs) of Chapter 3 showed the very limited nature of the differences in regions of strong convection between TCs embedded in positive and negative TCREH.

A new pathway through which TCREH modulates intensity was presented in Chapter 3 which is the azimuthal distribution of convection and surface latent heat fluxes. Time-composited sequences of these fields were plotted over the periods just prior to when intensity diverged between TCs embedded in positive vs. negative TCREH. The time-composited plots showed that regions of maximum convection and surface latent heat flux make steady progress into the upshear quadrants when TCREH is positive. In the cases when TCREH was negative, these maxima advanced much more slowly and required more time to reach the upshear quadrants and this delayed TC intensification. When TCREH reached a sufficiently large negative magnitude (< 11 m²s⁻²) this advancement process failed to occur and the TC never intensified.

Simulations including the addition of vertically uniform background flow showed that the motion of the TC relative to the vertical wind shear vector is important in determining intensification rate. These simulations demonstrated that TC motion can modulate the azimuthal distribution of convection as well as surface flux distribution and magnitude. In the presence of westerly shear, easterly background flow proved to be more favorable for intensification than westerly flow, particularly when TCREH was positive. Both the storm motion relative to the shear vector and the sign of TCREH played important roles in dictating TC intensification rate.
Trajectories were computed from very high resolution (667 m) simulations to track air parcels traveling through regions of convection and large surface flux near the RMW. Results from these trajectories showed that parcels were lofted into deep convection near the RMW in the upshear left quadrant much more frequently when TCREH was positive. The trajectories also showed that equivalent potential temperature of low-level parcels recently entering the boundary layer through downdrafts recovered faster in regions of strong surface fluxes in the upshear quadrants when TCREH was positive. Convection and regions of large surface latent heat flux were very uncommon in the upshear-left quadrant during the first 48 h of the TC simulations when TCREH was negative. Trajectories through the limited convection that did exist at these locations showed the convection was shallower and that less latent heat was released near the RMW. The combined analysis of the trajectories and time-averaged sequences of convection and surface latent heat flux showed that buoyant boundary layer air was ingested into the TC core in the upshear quadrants more effectively when TCREH was positive. This led to a subsequent symmetrization of convection and an acceleration of the TC intensification rate.

A new modeling framework was developed to simulate the TC response to wind shear environments that change smoothly in time. This method, time-varying point-downscaling (TVPDS), allows for smoothly transitioning vertical wind shear states without an instantaneous modification of the TC environment as was imposed by previous studies investigating time-varying shear. Results from the simulations described in Chapter 4 suggested that the instantaneous switch between environmental states in previous studies may indeed have unrealistic consequences in terms of the TC intensity
response. TVPDS simulations showed a more uniform weakening response to increasing shear, without a recovery to pre-shear intensity. This weakening occurred, in general, regardless of SST or environmental lapse rate. When the SST was warmer (e.g., 30° C vs. 27° C), the simulated TCs reached a much stronger intensity before additional shear was imposed, however, the rate at which the TCs weaken did not appear to be strongly related to SST, provided the shear was increased to 15 ms\(^{-1}\). Time-composited sequences of mid-level simulated reflectivity showed that the symmetry of TCs subjected to increasing vertical wind shear is steadily degraded. Similar sequences of boundary layer equivalent potential temperature (\(\theta_e\)) showed that intrusions of relatively stable air entered the boundary layer near the RMW much more frequently when shear increased. These low \(\theta_e\) intrusions led to voids in deep convection downstream related to the asymmetries seen in the mid-level reflectivity. The combination of these two processes greatly reduced the symmetric diabatic heating near and within the RMW and thus led to the simulated reduction in minimum central pressure. Time composites of azimuthal mean diabatic heating rate confirmed the reduction in heating when shear increased and convective symmetry was reduced. Radius-time Hovmöller diagrams of azimuthal mean low-level \(\theta_e\), mid-level relative humidity, and mid-level diabatic heating rate showed that low-\(\theta_e\) intrusions into the boundary layer preceded TC weakening. These diagrams also showed significant mid-level drying events and corresponding downdrafts into the boundary layer.

Simulations of the reverse process, quasi-steady state TCs transitioning from strong shear to weaker shear, demonstrated that the opposite sequence of events occurs as shear is reduced. In these cases, intrusions of low equivalent potential temperature into the
boundary layer near the RMW steadily became less common as shear was reduced. Meanwhile, mid-level convective symmetry and azimuthally averaged diabatic heating increased. These combined effects led to a strong recovery of the TC with significant reductions in minimum central pressure. The simulations of Chapter 4 confirm previous studies (e.g., Molinari et al. 2013; Riemer et al. 2010; Tang and Emanuel 2012) which noted that downdrafts characterized by low equivalent potential temperature play an important role in the TC response to shear. Radius-time Hovmöller diagrams showed, however, that ventilation of dry air into the TC core at mid-levels also may play a role in dictating the rate at which TCs weaken. The relative impact of these factors as well as determining which precedes the other will be a topic for future study.

Future research considering TCREH could help to solidify understanding of how variations in the vertical wind profile affect TC intensification. Expanding the statistical analysis described in Chapter 2 to additional ocean basins beyond the Atlantic could identify regions where TCREH plays a stronger role. TCREH also could be considered for statistical hurricane prediction schemes. The new TVPDS modeling framework offers expansive opportunity for idealized TC simulation and research. Due to the highly specifiable environment, experiments can systematically vary factors like vertical wind shear, temperature, and moisture in the TC environment. Simulations of time-varying moisture at constant shear, time-varying shear at constant moisture, or the combination of both are possible. Additional simulations in which the environmental lapse-rate is smoothly modified in time could provide insight into the TC response to changes in stability. Because any background state may be defined prior to and after transition
periods, the TVPDS method presents the opportunity to simulate the TC response to numerous changes in environmental conditions.
References


