ABSTRACT

CHURCH, DAVID EARL. Long-lived African Easterly Waves. (Under the direction of Dr. Anantha R. Aiyyer.)

The first part of this study uses an observational approach to study long-lived African easterly waves (AEWs). Long-lived AEWs are defined to be waves that exit the coast of Africa and propagate to 60°W before either developing into a tropical cyclone (TC) or dissipating. These waves show resilience in crossing the tropical Atlantic without either developing into a TC or dissipating. The NHC Tropical Cyclone reports were used to identify long-lived, developing AEW cases, while the ERA-Interim reanalysis and TRMM 3B42 data sets were used to confirm and track these cases. In all, 29 long-lived, developing cases were identified within the 2000 to 2010 hurricane seasons, accounting for about 15% of all hurricanes during that period. Two case studies are examined and they reveal an association between convection within the wave and a strengthening wave, while a lack of convection was associated with a weakening wave. Expanding the analysis to all 29 cases also suggested that convection within the wave may be critical to maintaining the wave. Convectively active times in the 29 cases were linked to increasing wave-level potential vorticity (PV), while convectively inactive times were linked to decreasing wave-level PV. While the observational portion of the study suggests that convection is important to maintaining or strengthening the long-lived waves, it does not give information about the physical processes linking individual thunderstorm clusters and the generation of wave scale vorticity in the middle and lower troposphere.

The second part of the study uses a high-resolution numerical modeling simulation of the pre-Ernesto (2006) long-lived AEW. The simulation is noteworthy, as it closely matches the evolution of the observed wave over a 9-day simulation period. Detailed analysis of the simulation shows that a wave pouch (Dunkerton et al. 2009) appears in the middle to lower troposphere a week before TC genesis. Dunkerton et al. (2009) show the wave pouch to be an area of enhanced vorticity that can protect the convection within. Our analysis confirms these findings and further implicates the wave pouch as an important feature in maintaining the wave. A PV budget and a vorticity budget are calculated from the WRF simulation. The budgets reveal that the convection within the wave is key to maintaining or strengthening the wave. The convective heating profile within the wave pouch is conducive to convergence in the lower to mid-troposphere and divergence aloft. This pattern of low-level convergence and strong vertical motion within the wave pouch is the main contributor to PV and vorticity generation through convergence and stretching. Our study supports the bottom-up school of thought for TC genesis and extends that philosophy to wave maintenance. We also examine the effects of stratiform versus convective precipitation by separating their individual contributions to the
vorticity budget. The results show the convective portion of the precipitation acts to maintain and strengthen the wave. We conclude that there are multiple pathways to achieving a long-lived AEW since the wave can be sustained through a balance of processes that support and inhibit convection. Future work will examine the sensitivities of an AEW to conditions that inhibit convection in order to better quantify the conditions needed to dissipate an AEW.
Long-lived African Easterly Waves

by
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DEDICATION

I would like to dedicate this work to my family and friends, who have always been supportive of my endeavors.
I grew up in Virginia Beach, VA, where I spent my childhood fascinated by the weather in the region. I watched eagerly as thunderstorms built on hot summer afternoons, I would wait in suspense for the much anticipated snow storms in the winter (which disappointed more often than not), and I tracked the progress of the hurricanes in Atlantic and experienced the wrath of a few of those storms. I attended the Mathematics and Science Academy at Ocean Lakes High School, where I began building a solid foundation for my understanding and passion for science. With my strong background as a student, my passion for meteorology and encouragement from my family I attended North Carolina State University for my undergraduate degree in Meteorology. I graduated summa cum laude and among the top of my meteorology class in May 2009. I decided to continue on with my education in Meteorology by starting my work as a graduate student under the advisement of Dr. Anantha Aiyyer in August 2009.

While meteorology is my passion, I enjoy being a well rounded person. In my spare time I enjoy being active, whether through running, biking, swimming or playing sports. I am also an avid sports fan and love watching NC State football and basketball, Carolina Hurricanes hockey among other sporting events. I also enjoy traveling and spending time with friends and family.
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Chapter 1

Introduction

1.1 Introduction

African Easterly Waves (AEWs) are the primary precursory disturbances for tropical cyclones (TCs) in the Atlantic and East Pacific basins (Pasch et al. 1998). Of interest in this study are long-lived AEWs, which can traverse the Atlantic basin before developing into a tropical cyclone in the Caribbean Sea or Gulf of Mexico. This study considers a long-lived AEW as an AEW that leaves the west coast of Africa and propagates past 60°W before either developing into a tropical cyclone, dissipating or becoming absorbed. Long-lived AEWs show resilience in crossing the Atlantic basin without dissipating, being absorbed by mid-latitude or other tropical disturbances or developing into tropical cyclones. The track across the tropical Atlantic typically takes the waves about a week, but they can persist up to two weeks by the time they reach the western Gulf of Mexico. The nature of long-lived AEWs suggests a balance between processes that strengthen the wave and processes that weaken the wave. In order to understand these processes and how they interact, we must first understand more about how an AEW is sustained over the open Atlantic.

Many current studies focus on the genesis of AEWs and their structure over Africa and many focus on the transition from wave to tropical cyclone, but few studies examine the structure and maintenance of these waves over the tropical Atlantic. Kiladis et al. (2006) thoroughly discuss the structure and formation of AEWs over Africa from both an observational and modeling framework. Over the African continent, the waves are observed to grow on and interact with the African Easterly Jet (AEJ). The AEJ is a result of the temperature gradient reversal between the wet and cooler Sahel to the south and the hot and dry Sahara to the north. The AEJ also denotes a region of sign reversal in the potential vorticity, which is one of the necessary conditions for instability in the background flow (Charney and Stern 1962). Kiladis
et al. (2006) show AEWs to have a tilted structure over the continent with the lower level vorticity centered north of the AEJ and the mid-level vorticity centered south of the AEJ. The vertically titled structure allows for the wave to grow due to baroclinic energy conversions, using the background temperature gradient. The waves have also been shown to grow through barotropic energy conversions, related to the horizontal wind shear in the region. Hsieh and Cook (2007) conclude that both the barotropic and baroclinic energy conversion are important to the development of AEWs and that the two processes may have a positive feedback resulting in non-linear wave growth.

Once the AEW moves out into the Atlantic basin, it moves away from the AEJ and the potential vorticity (PV) gradient is no longer reversed, as the temperature gradient over the tropical Atlantic is very weak. The AEW is also observed to shift from a vertically titled structure to a vertically stacked structure over the Atlantic (e.g., Pytharoulis and Thorncroft 1999, Thorncroft and Hodges 2001, Vizy and Cook 2009). The transition yields a wave that is no longer extracting energy from the background state and the transition to a vertically stacked system is likely a symptom that the wave is no longer baroclinic. Molinari et al. (1997) suggests an AEW will be dispersive since it is characterized as a Rossby wave. This dispersive nature will cause an AEW to slowly weaken as it crosses the Atlantic basin and ultimately dissipate, unless the wave can extract energy from another source. Molinari et al. (1997) suggest a possible sign reversal in PV over the Caribbean as one potential source of instability that could strengthen the waves and also mention convective coupling as another potential source for maintaining an AEW. This study will focus on the presence of convection as a potential mechanism for maintaining the wave.

It is also important to have a strong understanding of the transition from an AEW to a tropical cyclone since many long-lived AEWs cases result from failed tropical cyclogenesis events. The two main schools of thought for tropical cyclogenesis are the top-down view (Bister and Emanuel 1997 and Simpson et al. 1997) and the bottom-up view (Hendricks et al. 2004, Tory et al. 2006a, Kieu and Zhang 2009 and Dunkerton et al. 2009). The top-down view, as described by Bister and Emanuel (1997), suggests that the mid-level vorticity found in the stratiform regions of an AEW merge to create a large area of cyclonic vorticity. The mid-level vorticity is then brought to the surface through the cold downdrafts in the stratiform regions. However, Tory et al. (2006a) find that tropical cyclogenesis is a bottom-up development process where convergence into deep convective regions is necessary to increase the low-level vorticity. They find the low-level vorticity is increased through the convergence and stretching of vorticity, which is then vertically advected to create a deep vorticity tower.
Part 2 of the paper (Tory et al. 2006b) expands upon the bottom-up development process by explaining that the PV generated by individual clusters of thunderstorms merges with existing PV clusters. The merging creates an upscale aggregation of PV that results in a centralized, upright PV tower. They also find that the convective heating profile and secondary circulation (convergence in the lower to mid-troposphere and divergence aloft) play an important role in supporting the PV aggregation process. The third part of the paper (Tory et al. 2007) examines some of the non-developing cases (failed tropical cyclogenesis) and finds that the main processes inhibiting genesis are vertical wind shear and an insufficient large scale cyclonic environment. The wind shear disrupts the PV towers and may tear them apart, thus inhibiting upscale aggregation. The lack of a sufficient cyclonic environment may not confine the convective heating environment enough to create a focal point for TC genesis.

Dunkerton et al. (2009) introduce the concept of viewing a tropical wave in the Lagrangian framework, by subtracting the motion of the wave from the mean flow. The wave relative flow is more important for understanding the dynamics of the wave and the pathway to tropical cyclogenesis than the standard Eulerian view. Dunkerton et al. (2009) introduce the concept of the cat’s eye, or wave pouch, which forms at the intersection of the wave’s trough axis and the critical latitude. The critical latitude can be understood as the latitude at which the background flow matches the phase speed of the wave. Therefore, when subtracting the phase speed of the wave from the total flow, the wave pouch will appear as closed streamlines centered about the intersection of the trough axis with the critical latitude. Figure 3.5 illustrates the results of a thought experiment, where panel (a) illustrates the typical Eulerian view of an AEW while panel (b) illustrates the Lagrangian view of an AEW, where the wave speed has been subtracted from the total flow. Here it is easy to see that prior to a closed circulation appearing in the typical Eulerian view, a closed circulation may already exist within the wave.

Dunkerton et al. (2009) cite the development of a wave pouch as an important feature in the process leading to tropical cyclogenesis, where virtually all 55 of the named tropical storms they studied showed a wave pouch as far out as 3 days prior to genesis. The wave pouch provides an area within the wave that is protected from outside intrusions of dry air and wind shear. This makes the pouch favorable for supporting convection, thus allowing for a primarily convective heating profile. Later analysis of the PV budget and vorticity budget suggest this convective heating profile may be key in supporting the AEW. The pouch also provides a region of enhanced cyclonic vorticity and weak deformation, which is favorable for vortex aggregation up the vorticity gradient. Thus vorticity generated by random convection within the wave pouch are likely to aggregate toward the wave pouch center while the negative vorticity generated is likely to be ejected from the wave pouch. Dunkerton et al. (2009) discuss
three hypothesis:

- the development of the pouch is key to tropical cyclogenesis, providing a mechanism for bottom-up development

- the pouch provides a favored region for convection by allowing the air within to be repeatedly moistened and by providing protection from dry air intrusions

- the parent wave may be maintained and even enhanced by the eddies within the wave

While they discuss the first two items in depth, the third hypothesis is left open ended and is the primary point of concern for this study. The third hypothesis provides a pathway for the random convection within the wave to generate vorticity at the mesoscale level and cascade the energy to the wave scale, thus maintaining the parent wave.

Montgomery et al. (2010) further analyze the Marsupial Paradigm theory using idealized modeling and find evidence of upscale aggregation of vorticity. The high-resolution simulation of a TC genesis case shows that vortical hot towers (VHTs) play a critical role in the aggregation of vorticity and also cause strong boundary layer convergence. It is suggested that the boundary layer convergence is key to spinning up low-level, storm-scale vorticity. Wang et al. (2010a) examine the genesis of Hurricane Felix (2007) using a high resolution WRF simulation. The findings support the Marsupial Paradigm, where a wave pouch is present to help organize and aggregate vorticity maxima. The study also uses a vorticity budget to show that the low-level vorticity is generated by convergence and stretching, which favors the bottom-up school of thought over the top-down theory.

Based on the current understanding of AEWs and tropical cyclogenesis, this study looks to fill in a gap in understanding of how AEWs can traverse the tropical Atlantic without decaying or developing. As an AEW moves off the African coast, it moves away from the AEJ and PV sign reversal, which are important in growing the wave. Once over the Atlantic, the wave becomes a stacked system (in the absence of vertical wind shear) and must be maintained through an alternate mechanism. Without an alternate maintenance mechanism, the Rossby wave nature of the AEW should cause it to slowly decay and dissipate. From a forecasting perspective, it is generally accepted that waves with strong convection are likely to persist or develop into a tropical cyclone, but the exact processes that describe this positive feedback are not understood. By understanding the process that are responsible for maintaining or strengthening the wave, it may be easier to understand what competing processes may be detrimental to the wave.
The focus of this study is to better understand the structure and maintenance mechanisms for an AEW as it moves over the tropical Atlantic. In order to accomplish this goal, this study will use a two-pronged approach. The first step is an observational study, which will use the ERA-Interim reanalysis data set and TRMM data set. The observational study is important for understanding the climatology of long-lived AEWs and obtaining a generalized understanding of the important processes maintaining an AEW. The second step is a modeling study, which will use the Weather Research and Forecasting (WRF) model to produce a high-resolution simulation of a long-lived AEW. The modeling study’s importance is to provide a high-resolution, temporally consistent data set that can be used to understand the structure of the AEW and to effectively quantify the important physical processes in maintaining the wave. The modeling study is noteworthy since it simulates the long-lived pre-Ernesto (2006) AEW with remarkable similarity to observation, over a 9-day period.
Chapter 2

Observational Study

2.1 Data

2.1.1 ERA-Interim

The observational portion of this study uses the European Center for Medium-Range Weather Forecasts (ECMWF) Reanalysis - Interim (ERA-Interim) data set, which is available on a 6-hourly, 1.5° latitude-longitude grid with 37 pressure levels (Dee et al. 2011). Several other studies have successfully used ECMWF data for studying AEWs (e.g., Reed et al. 1988, Liebmann and Hendon 1990, and Boer 1995), citing the model’s skill in predicting AEWs over Africa, the tropical Atlantic and Caribbean, its ability to accurately represent the structure of the waves and its ability to sustain the wave in data sparse regions, due to the model’s 4D data assimilation. However, Molinari et al. (1997) cautions that while the model does well with synoptic features in the tropics, any detailed calculations should be avoided as the structural details are sensitive to the presence of data. This reemphasizes the need for the high-resolution WRF simulation in the second part of this study.

2.1.2 TRMM 3B-42

The Tropical Rainfall Measuring Mission (TRMM) 3B-42 data set is used to supplement the reanalysis data with observed precipitation. The TRMM 3B-42 data is a multi-sensor data set that incorporates data from SSMI, AMSR and AMSU precipitation estimates as well as geostationary infrared (IR) cloud top temperatures when the former are not available over a region. The IR rainfall estimates are calibrated to match the other high quality estimates; and the combination of these sources results in a 0.25°, 3-hourly precipitation estimate data set over the tropics.
2.2 Observational Wave Tracking and Identification

2.2.1 African Easterly Wave Tracking Method

The method used to track AEWs in this study used an algorithm (Tyner and Aiyyer (2011)) that identifies coherent structures using potential vorticity on the 315 K isentropic surface. This algorithm does not automatically yield nice AEW track information due to occasional failures to identify an AEW at every time. Therefore, the track information from the algorithm was used as the basis for the AEW tracks and the waves were tracked by manually piecing together the automatically identified wave centers. For analysis times where the tracking algorithm missed the wave center, the track points were linearly interpolated between the last identified track point and the next identified track point.

2.2.2 Identification of Long-lived African Easterly Waves

Long-lived AEWs were broken down into two classifications depending on whether they eventually developed into tropical cyclones. The AEWs that eventually developed into tropical cyclones were identified through the National Hurricane Center’s Tropical Cyclone Reports for the 2000 to 2010 hurricane seasons. If the report indicated that the storm originated as an AEW, and the longitude of tropical cyclone formation was west of 60°W then the wave was considered a long-lived case. This process was completed for both the Atlantic and east Pacific basins, generating a database of all long-lived, developing tropical cyclones during those 11 seasons. The tracks for these waves were then identified using the method described above and were cross referenced with GOES satellite and TRMM data to ensure their accuracy. Any tropical cyclones that the NHC claimed formed from an AEW but the track of that wave could not be verified through the ERA-Interim reanalysis or through satellite analysis was removed from the long-lived, developing AEW case list.

Long-lived AEWs that did not develop were more difficult to identify since a database of AEW tracks is not maintained. In order to estimate the number of AEWs that are long-lived and dissipate, all AEWs were tracked between July and October 2005 using the semi-objective method described above. During the tracking process, the fate of each wave was categorized as: developed into a tropical cyclone, dissipated, split, or absorbed by another weather system. From the database of all AEWs between July and October 2005, dissipating, long-lived waves were identified by those waves that passed 60°W and had a fate of either dissipated or absorbed.
For this study, only the long-lived, Atlantic basin, developing AEWs are considered in the observational study. Tropical cyclones that developed in the east Pacific that were attributed to long-lived AEWs were generally much harder to track and verify using the ERA-Interim dataset. The overall difficulty in accurately tracking AEWs that lead to east Pacific TCs would undermine the confidence in the observational portion of the study. Also the dissipating waves were not considered in the observational study due to the tedious nature of the semi-objective tracking technique. Since this study assumes that any long-lived AEW is characterized by processes that maintain the wave and processes that are detrimental to the wave, the ultimate fate of the wave is irrelevant in understanding processes that make the wave long-lived.

2.3 Climatology

Between 2000 and 2010, the Atlantic Basin had 194 tropical cyclones, of which 29 formed from long-lived AEWs. Table 2.1 lists all 29 long-lived developing cases. Figure 2.1 shows the yearly percentage of tropical cyclones forming from long-lived waves. The 2005, 2008 and 2010 hurricane seasons experienced an above average number of tropical cyclones forming from long-lived AEW, at about 23%, while the 2002, 2004, 2006 and 2007 seasons each only had 1 tropical cyclone that formed from a long-lived AEW.

When examining the tracks of TCs that formed from long-lived waves (Figure 2.2), it is apparent that these storms are more likely to pose a threat to land. These storms are forming at a closer proximity to land making it more difficult to re-curve without making landfall. The statistics for the 11 hurricane seasons from 2000 to 2010 show 53% of all Atlantic tropical cyclones made landfall and by removing the long-lived AEW cases this number drops to 41%; however, of the 29 tropical cyclones that formed from long-lived waves, 79% made landfall.

Despite many of these TCs forming within close proximity to land, figure 2.3 shows there is no preference for TCs forming from long-lived AEWs to be stronger or weaker than TCs that do not form from long-lived AEWs. A t-test showed no statistical difference between the distribution of intensities for storms forming from long-lived AEWs and those that did not.

The climatology suggests that it is important to understand long-lived AEWs since these wave contribute to about 15% of all TCs in the Atlantic basin. TCs that form from long-lived AEWs are more likely to develop just prior to landfall and showed no trend toward being stronger or weaker than TCs that did not form from long-lived AEWs. This combination cre-
ates a problem from a forecasting perspective as well as an emergency manager and public awareness perspective.

2.4 Observational Study

The observational study suggests there are likely two main mechanisms for maintaining AEWs. The first is through a PV flux into the wave from either mid-latitude sources or from other tropical disturbances and second is through convective generation of PV middle troposphere, due to the convective heating profile within the wave. However, the merger of a long-lived AEW with a mid-latitude or tropical PV source was rare, accounting for only 3 of the 29 cases. Instead, convection within the wave is linked to maintaining or intensifying a wave, while a lack of convection is linked to a decaying wave.

2.4.1 PV Basics

PV is used to track and describe the evolution of the waves in the study because it is a conserved quantity for insentropic, adiabatic processes. PV describes the combination of vorticity and stability as is given by the following equation:

\[
PV = \frac{1}{\rho} \zeta^a \cdot \nabla \theta \tag{2.1}
\]

where \(\rho\) is the density of the fluid, \(\zeta^a\) is the absolute vorticity, and \(\nabla \theta\) is the gradient of potential temperature (Haynes and McIntyre 1987)

2.4.2 PV Flux into the Wave

The three examples of waves merging with mid-latitude disturbances were pre-Erin (2007) and pre-Kyle (2008) and pre-Omar (2008). These waves are characterized by weakening 600 hPa PV prior to interacting with a mid-latitude disturbance that became caught in the tropical easterly flow. These cases were identified by visually inspecting the track of each long-lived AEW case and by viewing the Hovmoller diagrams for each wave. These 3 AEWs showed clear evidence of two distinct PV maxima merging in the associated Hovmoller diagram and in the visual inspection of the plan view PV field. Figure 2.4 shows a representative Hovmoller diagram of the pre-Kyle (2008) case, where the PV associated with the wave is weakening prior to
the mid-latitude PV source catching up to, and merging with the AEW. Although part of the mid-latitude source continues to propagate past the AEW, the merger contributes to increasing PV within the wave. Four days after the merger the TC develops, denoted by the "X" on the Hovmoller.

The interaction between the mid-latitude PV source and the wave is favorable, and leads to a rejuvenation of the wave and to a tropical cyclone several days later. However, in all three cases the merger with the mid-latitude PV source did not occur until 90°W, 60°W, and 50°W, respectively. Both pre-Erin and pre-Kyle had already met the criteria for a long-lived AEW, while pre-Omar was just 10° short of making it to the 60°W cutoff. These cases were already long-lived (or just about in the case of pre-Omar) even prior to the PV merger. While the mergers were important in reinvigorating the wave and likely helped lead to tropical cyclogenesis, the mergers were not the main reason for supporting the wave as it crossed the tropical Atlantic. This suggests that the PV flux into the wave may not be the most common mechanism for maintaining long-lived AEWs.

It is also important to note difficulty in performing any PV budget calculation on the ERA-Interim data set, especially for a moving box surrounding the wave. The track of the waves in the reanalysis can be inconsistent in time and have sudden shifts in structure from one time to the next. Thus, quantifying the net flux of PV into or out of the wave-following box may not achieve the desired result, and may describe the variability in the reanalysis itself. Therefore, this study does not attempt to determine the contribution of the PV flux for each wave case, although in most cases it is assumed to be small. Visual inspection of the 29 cases only found 3 cases where the PV flux into the wave was an important part of the wave's lifespan.

2.4.3 PV Generation from Convection: Case Studies

The convective generation of PV in the middle troposphere due to the convective heating profile within the wave is found to be the most important contributor to maintaining the long-lived AEWs. Examination of two long-lived AEW case studies shows an association between convective activity within the wave and an increasing or steady PV anomaly associated with the wave. Furthermore, weak or non-existent convective activity within the wave corresponds with a weakening PV anomaly associated with the wave.

An example of the relationship between convection and increasing PV is seen in two case studies, pre-Ernesto (2006) and pre-Stan (2005). In figure 2.5 the PV maxima associated with
the pre-Stan (2005) AEW is steady for the first 2 days at about 0.4 PVU before decreasing to 0.2 PVU by 23 September 2005. During this time of steady to decreasing PV anomaly, both the TRMM rainfall rate and the reanalysis vertical velocity are very small, indicating weak to no convection within the wave during this time. However, from 23 September until TC genesis on the 1 September, the PV anomaly associated with the AEW increases from about 0.2 PVU to about 0.55 PVU. During this time, the TRMM rainfall rate indicates bursts of convection within the wave. The ERA-Interim’s vertical velocity field represents this increase in convection as well.

Also in examining figure 2.6 for the pre-Ernesto (2006) AEW, a similar pattern is observed. Initially the PV associated with the wave is steady to decreasing from 18 August to 21 August. During this time, the TRMM rainfall rates and reanalysis omega field indicate weak and decreasing precipitation associated with the wave. However, on 20 August, both the TRMM rainfall rates and the reanalysis omega field indicate an increase in precipitation associated with the wave. The increase in precipitation rate precedes the increasing PV on 21 August.

Both of these examples show a relationship between convection associated with the wave and PV associated with the wave. The pre-Ernesto case points to the increasing convection occurring prior to the increase in PV since there is a delay from when the convection begins to when the PV begins to increase. The Stan case (figure 2.5) also shows the convection returning to the wave on on 23 September when the PV field is at a local minimum. It is only after the convection returns to the wave that the PV begins to recover. This observation is consistent with the second hypothesis, in that convective coupling with the wave may play a significant role in maintaining or strengthening the PV anomaly within the wave.

The two case studies described above point to a link between convection and a strengthening wave. The arrow of time in these case studies points to convection occurring first, followed by a strengthening of the mid-level PV anomaly associated with the AEW. While the two case studies qualitatively point to convection being important for wave maintenance, it is important to quantitatively check the robustness of these results with the remaining 27 cases. If the qualitative analysis is correct in assuming that the convection plays a role in strengthening the wave scale PV, then - over all the long-lived cases - we expect waves with convection to be strengthening and waves without convection to be weakening.
2.4.4 PV Generation from Convection: Extension to all Long-lived Developing Cases

To study the differences between convectively active time periods and convectively inactive time periods for all waves, the TRMM precipitation rate was averaged in a 6° x 6° box around the wave center at every 6 hour reanalysis time for every wave. This procedure yields a time series of the precipitation rate for each wave. However, if all the times from all the waves are combined, then a distribution of the precipitation rates for all the cases is generated. From this distribution, threshold values for when convection is active versus when convection is inactive can be determined. Convectively inactive times were considered to be in the lower quartile of the distribution, and have precipitation rates of less than 0.14 mm day$^{-1}$, while convectively active times were considered to be in the upper quartile, and have precipitation rates greater than 0.73 mm day$^{-1}$. The cut off values were visually checked with plan view plots of the TRMM precipitation and it was determined that the values meet the expectation for low convective activity and high convective activity, respectively. Once every convectively active and convectively inactive time was determined for all 29 cases, other variables were also area-averaged over the 6° x 6° box surrounding every wave at every time. Using the precipitation rate information, the convectively active and inactive times were used to separate the other variables to determine if there is a difference in the variables between the convectively active times and inactive times.

A summary of the results can be seen in table 2.2. A t-test of each variable showed a statistically significant difference between the convectively active and inactive times. First, the relative humidity is, on average, higher at both 850 hPa and 600 hPa for the convectively active cases. Figure 2.7 (a) and (b) show the distributions for the relative humidity at 850 hPa and 600 hPa, respectively. The convectively active time periods (hatched) have a RH distribution that is bunched to the right of the graph, while the convectively inactive time periods (solid) have a distribution that has a longer tail toward lower RH values. This result shows that the ERA-Interim matches well with the TRMM precipitation fields since the reanalysis should show higher RH at both levels when convection is observed to be active. This gives confidence in using the TRMM rainfall rate information to categorize convectively inactive and active times and separate the remaining variables.

Convectively active time periods are also described by larger values of mid-level PV, as can be seen in table 2.2 and by the shift in the distribution in figure 2.7(c). This has some relevance in showing that convective activity is linked to the strength of the wave. However, the PV tendency is more important for describing whether convection contributes to increasing PV and thus a strengthened wave. The analysis shows that the average PV tendency for a
convectively inactive wave is -0.107 PVU day$^{-1}$, while the average tendency for a convectively active wave is 0.529 PVU day$^{-1}$. Figure 2.7(d) shows the distribution of PV tendencies for the convectively active times is shifted to the positive PV tendencies while the distribution for the convectively inactive times is shifted to the negative PV tendencies. This statistic shows that over all 29 long-lived AEWs, the convectively active times are associated with increasing mid-level PV while the convectively inactive times are associated with decreasing PV.

The wind shear is counter-intuitive, with the convectively active time periods experiencing, on average, stronger wind shear than the convectively inactive time periods. It is hard to understand this result without further analysis, but the increased shear could be a result of the convection itself. The results suggest that for convection within waves, a small amount of wind shear may not be detrimental, even though it may not help with the organization of the wave. Dunkerton et al. (2009) even suggest that in the early stages of convective organization within the wave, the shear may play an important role in organizing the convection and making it longer lived. It is only when strong shear interacts with a mature vortex that it can be detrimental to the overall structure of the storm.

2.4.5 Case Study: Ernesto 2006

The purpose for examining the pre-Ernesto (2006) long-lived AEW is two fold. First, the case is studied extensively in the modeling study portion of the paper and so it is important to understand the evolution of the wave. Second, for the observational section of the paper, we examine the presence of a wave pouch as a possible mechanism for organizing convection within the wave and providing a pathway for convection to lead to wave maintenance.

Background

Atlantic Hurricane Ernesto (2006) formed from a long-lived AEW that emerged from the west coast of Africa on August 18th. The wave propagated across the Atlantic ocean for 7 days and spanned about 5000 km before becoming classified by the NHC as a tropical depression at 18Z on August 24th. Convection remained active within the wave for much of the time it crossed the Atlantic with only a few periods of weakened convection.

The pre-Ernesto wave was affected by high wind shear and subsidence related to the Saharan air layer (SAL) outbreak that occurred as the wave moved off the African coast (Vizy and Cook 2009 and Zawislak and Zipser 2010). Vizy and Cook (2009) compared the pre-Ernesto
wave to the wave behind it that developed in TC Debby upon entering the Atlantic. Their study found that the Debby case exhibited stronger surface convergence, stronger upper-level divergence and higher relative humidity values than the pre-Ernesto case. All three of these parameters are very important in the bottom-up theory of tropical cyclogenesis and explain why the pre-Ernesto wave was not able to develop immediately upon entering the Atlantic. The increased wind shear associated with the SAL also kept the pre-Ernesto wave from becoming a stacked system for about 3-4 days after it moved off the African coast, while the pre-Debby wave was stacked almost immediately upon exiting the coast (Vizy and Cook 2009 and Zawislak and Zipser 2010). While the environmental conditions were not favorable for TC genesis for several days as the pre-Ernesto wave moved into the tropical Atlantic, the wave and its associated convection survived until the environmental conditions were favorable for TC genesis at 18Z August 24.

Wave Pouch Analysis

Examining the 600hPa relative flow along with the PV field in figure 2.8 suggests that the wave pouch may be a transient feature that is present only some of the time during the waves life span. The 18Z 18 August, 18Z 20 August and 18Z 24 August show a wave pouch while the remaining times show open flow through the wave. However, in examining each reanalysis time (not shown) the pouch appears and disappears at irregular intervals. The wave pouch should evolved smoothly as it depends on the wave speed and background flow speed which also evolve smoothly with time. The erratic nature of the pouch evolution with time in the ERA-Interim suggests that the reanalysis is not consistent in time with the low-amplitude, pre-Ernesto wave.

One key to understanding why the pouch does not vary smoothly in time comes in examining the track of the wave (figure 2.9). The ERA-Interim had a hard time with the wave track as seen by the bunching of the track points for August 21 and August 22, indicating the wave was nearly stationary for a 24 hour period. Problems with the track speed in the reanalysis have a direct impact of the wave relative flow analysis as the wave pouch is sensitive to both the background wind speed and the translation speed of the wave. Since the track of the pre-Ernesto wave is not smoothly varying in time, it is no surprise that the wave pouch appears to come and go with time.

This case study suggests that the wave pouch can be present long before TC genesis as is shown by the presence of the pouch in figure 2.8(a), 6 days prior to TC genesis. Although the wave pouch does not appear at all times in the reanalysis, this may be an artifact of the ERA-Interim poorly resolving the wave structure and its translation speed. The presence of a
wave pouch suggests a mechanism for organizing convection within the wave and a pathway for upscale aggregation of vorticity. The difficulty in resolving the wave pouch for the pre-Ernesto case stresses the need for the high-resolution modeling study, which will more accurately describe this sub-synoptic scale structure.

2.4.6 Conclusions from the Observational Study

The observational data available between the TRMM and ERA-Interim reanalysis suggest that the coupling of the AEW to convection may be key in maintaining or strengthening the wave. The population of long-lived AEW cases suggest that waves with strong convection are strengthened while waves with weak to no convection are weakened. However, the limitations of the ERA-Interim data set, in horizontal resolution, temporal resolution, and temporal consistency in wave structure prevent detailed calculations to better describe the important dynamical and thermodynamical processes.

The analysis of the pre-Ernesto wave shows a pouch structure may be present long before TC genesis. The favorable environment within the wave pouch, as described by Dunkerton et al. (2009), may be key to providing an area where convection is favored and can help sustain the parent wave. However, it is hard to have confidence in the exact structure of the pouch as described by the ERA-Interim reanalysis since this feature can be on a smaller scale than the wave itself. Also, variability in the ERA-Interim representation of the wave between reanalysis times gives little confidence in the temporal variability of the structure.

Both the observational analysis of the convection supporting the parent wave and the wave pouch structure are not very satisfying, since many questions are left unanswered due to the lack of detail in ERA-Interim. This highlights the need for a high resolution model simulation of a long-lived AEW, so the important physical processes can be better understood. The modeling study will allow the mesoscale processes, which appear to be very important, to be resolved and will provide a temporally consistent data.
### Table 2.1: Long-lived, Developing AEW cases, 2000-2010

<table>
<thead>
<tr>
<th>Year</th>
<th>Name</th>
<th>Max Cat</th>
<th>Wave Date</th>
<th>TC Genesis Date</th>
<th>Genesis Lat</th>
<th>Genesis Lon</th>
</tr>
</thead>
<tbody>
<tr>
<td>2000</td>
<td>Beryl</td>
<td>TS</td>
<td>08/03</td>
<td>08/13</td>
<td>22.5</td>
<td>93.5</td>
</tr>
<tr>
<td></td>
<td>Gordon</td>
<td>1</td>
<td>09/04</td>
<td>09/14</td>
<td>19.8</td>
<td>87.3</td>
</tr>
<tr>
<td>2001</td>
<td>Barry</td>
<td>TS</td>
<td>07/24</td>
<td>08/02</td>
<td>25.7</td>
<td>84.8</td>
</tr>
<tr>
<td></td>
<td>Dean</td>
<td>TS</td>
<td>08/16</td>
<td>08/22</td>
<td>17.6</td>
<td>64.3</td>
</tr>
<tr>
<td>2002</td>
<td>Isadore</td>
<td>3</td>
<td>09/09</td>
<td>09/14</td>
<td>10.0</td>
<td>60.5</td>
</tr>
<tr>
<td>2003</td>
<td>Grace</td>
<td>TS</td>
<td>08/19</td>
<td>08/30</td>
<td>24.3</td>
<td>92.4</td>
</tr>
<tr>
<td></td>
<td>Henri</td>
<td>TS</td>
<td>08/22</td>
<td>09/03</td>
<td>27.4</td>
<td>87.7</td>
</tr>
<tr>
<td></td>
<td>Larry</td>
<td>TS</td>
<td>09/17</td>
<td>10/01</td>
<td>21.0</td>
<td>93.2</td>
</tr>
<tr>
<td>2004</td>
<td>Jeanne</td>
<td>3</td>
<td>09/07</td>
<td>09/13</td>
<td>15.9</td>
<td>60.1</td>
</tr>
<tr>
<td>2005</td>
<td>Franklin</td>
<td>TS</td>
<td>07/10</td>
<td>07/21</td>
<td>25.0</td>
<td>75.0</td>
</tr>
<tr>
<td></td>
<td>Gert</td>
<td>TS</td>
<td>07/10</td>
<td>07/23</td>
<td>19.3</td>
<td>92.2</td>
</tr>
<tr>
<td></td>
<td>Harvey</td>
<td>TS</td>
<td>07/22</td>
<td>08/02</td>
<td>28.2</td>
<td>68.8</td>
</tr>
<tr>
<td></td>
<td>Katrina</td>
<td>5</td>
<td>08/11</td>
<td>08/23</td>
<td>23.1</td>
<td>75.1</td>
</tr>
<tr>
<td></td>
<td>Stan</td>
<td>1</td>
<td>09/17</td>
<td>10/01</td>
<td>18.9</td>
<td>85.6</td>
</tr>
<tr>
<td></td>
<td>Tammy</td>
<td>TS</td>
<td>09/24</td>
<td>10/05</td>
<td>27.3</td>
<td>79.7</td>
</tr>
<tr>
<td></td>
<td>Alpha</td>
<td>TS</td>
<td>10/15</td>
<td>10/22</td>
<td>15.8</td>
<td>67.5</td>
</tr>
<tr>
<td>2006</td>
<td>Ernesto</td>
<td>1</td>
<td>08/18</td>
<td>08/24</td>
<td>12.7</td>
<td>61.6</td>
</tr>
<tr>
<td>2007</td>
<td>Erin</td>
<td>TS</td>
<td>08/03</td>
<td>08/15</td>
<td>23.7</td>
<td>90.7</td>
</tr>
<tr>
<td>2008</td>
<td>Dolly</td>
<td>2</td>
<td>07/11</td>
<td>07/20</td>
<td>17.8</td>
<td>83.6</td>
</tr>
<tr>
<td></td>
<td>Fay</td>
<td>TS</td>
<td>08/06</td>
<td>08/15</td>
<td>18.4</td>
<td>67.4</td>
</tr>
<tr>
<td></td>
<td>Kyle</td>
<td>1</td>
<td>09/12</td>
<td>09/25</td>
<td>21.5</td>
<td>70.0</td>
</tr>
<tr>
<td></td>
<td>Omar</td>
<td>4</td>
<td>09/30</td>
<td>10/13</td>
<td>15.4</td>
<td>69.0</td>
</tr>
<tr>
<td>2009</td>
<td>Claudette</td>
<td>TS</td>
<td>08/07</td>
<td>08/16</td>
<td>27.0</td>
<td>83.4</td>
</tr>
<tr>
<td></td>
<td>Danny</td>
<td>TS</td>
<td>08/18</td>
<td>08/26</td>
<td>24.3</td>
<td>69.6</td>
</tr>
<tr>
<td>2010</td>
<td>TD-2</td>
<td>TD</td>
<td>06/24</td>
<td>07/08</td>
<td>23.7</td>
<td>93.7</td>
</tr>
<tr>
<td></td>
<td>Bonnie</td>
<td>TS</td>
<td>07/13</td>
<td>07/22</td>
<td>21.5</td>
<td>73.8</td>
</tr>
<tr>
<td></td>
<td>Karl</td>
<td>3</td>
<td>09/01</td>
<td>09/14</td>
<td>18.3</td>
<td>84.2</td>
</tr>
<tr>
<td></td>
<td>Matthew</td>
<td>TS</td>
<td>09/11</td>
<td>09/23</td>
<td>13.7</td>
<td>74.8</td>
</tr>
<tr>
<td></td>
<td>Otto</td>
<td>1</td>
<td>09/26</td>
<td>10/06</td>
<td>22.0</td>
<td>67.2</td>
</tr>
</tbody>
</table>

### Table 2.2: Convectively Active vs Convectively Inactive

<table>
<thead>
<tr>
<th>Variable</th>
<th>Convectively Inactive</th>
<th>Convectively Active</th>
<th>T-Test</th>
</tr>
</thead>
<tbody>
<tr>
<td>RH 850 hPa</td>
<td>75.52 %</td>
<td>82.92 %</td>
<td>0 - Significant</td>
</tr>
<tr>
<td>RH 600 hPa</td>
<td>58.59 %</td>
<td>73.92 %</td>
<td>0 - Significant</td>
</tr>
<tr>
<td>PV 600 hPa</td>
<td>0.271 PVU</td>
<td>0.361 PVU</td>
<td>0 - Significant</td>
</tr>
<tr>
<td>PV Tendency</td>
<td>-0.107 PVU day⁻¹</td>
<td>0.529 PVU day⁻¹</td>
<td>0 - Significant</td>
</tr>
<tr>
<td>Shear</td>
<td>9.63 m s⁻¹</td>
<td>11.48 m s⁻¹</td>
<td>0 - Significant</td>
</tr>
</tbody>
</table>
Figure 2.1: A histogram of the percentage of tropical cyclones that formed from long-lived AEWs each year between 2000 and 2010
Figure 2.2: Tracks of all 29 long-lived, developing AEW cases between 2000 and 2010. Blue indicates a wave, green indicates a tropical depression, orange indicates a tropical storm and red indicates a hurricane.
Figure 2.3: Comparison of the maximum storm classifications for tropical cyclones forming from long-lived AEWs (red) and those forming from non-lived disturbances (blue).
Figure 2.4: A Hovmoller diagram of 600 hPa PV from the pre-Kyle (2008) AEW and its merger with a mid-latitude PV source prior to tropical cyclogenesis. The "*" symbol illustrates the point of merger and the "X" symbol illustrates the point of TC genesis as denoted by the NHC.
Figure 2.5: A time series of the area-average (6” x 6” box) of the 600 hPa PV (top), TRMM rainfall rate (middle) and 600 hPa omega (bottom) for the pre-Stan (2005) AEW.
Figure 2.6: A time series of the area-average (6° x 6° box) of the 600 hPa PV (top), TRMM rainfall rate (middle) and 600 hPa omega (bottom) for the pre-Ernesto (2006) AEW.
Figure 2.7: Histograms that compare the non-convective (shaded) to the convective (hatched) reanalysis times for the 29 long-lived, developing AEW cases.
Figure 2.8: 600 hPa wave-relative streamlines and PV for the pre-Ernesto AEW from the ERA-Interim reanalysis.
Figure 2.9: The wave track for the pre-Ernesto AEW overlaid on the time averaged PV field from 18 August 2006 to 24 August 2006.
Chapter 3

Modeling Study

3.1 Modeling Study

3.1.1 High Resolution Model Simulation

To obtain a high resolution data set that is consistent with time, a WRF-ARW model simulation was run for the pre-Ernesto case (2006). The model has 28, terrain-following, vertical levels with the model top set at 50 hPa. Figure 3.1 shows the 2-domain configuration. A 12 km horizontal resolution outer domain encompasses the tropical Atlantic and spans the longitudes transversed by the long-lived pre-Ernesto AEW; and a 4 km moving nest that is initialized within domain 1, centered near 15°W, 10°N. The inner domain is defined to be about 20° latitude by 20° longitude to fully encompass the AEW and was prescribed preset moves to keep the domain centered on the pre-Ernesto AEW. The preset moves for the inner domain were originally defined by the anticipated translation speed of the wave from the ERA-Interim reanalysis. A test run was conducted to determine the approximate track of the wave in the 4km simulation and the preset moves were adjusted to keep the wave centered within the 4km domain.

The model was initialized from the ERA-Interim reanalysis on 0000 UTC 18 August 2010. The reanalysis data set was also used to update the boundary conditions on the 12 km grid every 6 hours. Sea surface temperature data was retrieved from the NCEP/MMAB real-time, global, sea surface temperature analysis (RTG SST). The RTG SST is a daily, half degree resolution data set that is derived from buoy, ship and satellite measurements. The daily sea surface temperatures were linearly interpolated to every 6 hours and were updated within the WRF model at that frequency.

Testing was conducted to determine the combination of parametrizations that would most
closely reproduce the observed behavior of the pre-Ernesto wave. The observed case remained convectively active while it traversed the tropical Atlantic, before developing into a tropical depression by 1800 UTC 24 August 2006, as it entered the Caribbean Sea. The physics options combination that most closely replicated the observed behavior was: the WRF Single-Moment 6-class microphysics scheme, the Rapid Radiation Transfer Model longwave radiation scheme, the Dudhia shortwave radiation scheme, the MM5 Similarity surface layer scheme, the NCEP/NCAR/AFWA Noah land surface model, the Yonsei University PBL scheme, and the Kain-Fritsch cumulus scheme on the 12 km domain with no cumulus parametrization on the 4 km domain. Also, the Garratt formulation for hurricane-specific surface fluxes (isftcflx 2) was used.

The goal of this study is not to determine the sensitivity of tropical cyclogenesis to microphysics schemes, simply to generate a WRF simulation that closely reflects the observed case. This high resolution simulation will serve as a high quality data source to better understand the important processes in a long-lived AEW. However, it is intriguing that the microphysics scheme used in a simulation has such a profound impact on whether or not a wave develops into a tropical cyclone and how quickly it develops.

3.1.2 Model Verification

The simulation with the parametrizations described above performed remarkably well considering the length of the simulation. The simulation was started at 0000 UTC on 18 August 2006 and was concluded at 0000 UTC on August 27 2006. The simulation is essentially a 9 day forecast, except the boundary conditions on the outer domain were updated by the ERA-Interim reanalysis, which provided information about the correct, large scale features to the outer model domain, and the SST was updated with the observed values.

Track Verification

By comparing the track of the simulated pre-Ernesto wave with the ERA-Interim track and NHC best track in figure 3.2, we see that the WRF simulation closely reflects the track of the wave. Remarkably, the track error at the end of the 9 day forecast is very small, as the WRF simulation is about a degree farther west than where the NHC best track information places the hurricane. The WRF simulation better describes the propagation speed of the wave than the ERA-Interim. The ERA-Interim wave track shows inconsistencies in the wave speed, which can be seen by the small distance traveled between 20 August and 21 August and the large
distances traveled between 19 August and 20 August and 23 August and 24 August; however, the distances traveled each day in the WRF simulation are very consistent and any changes in speed are slow and continuous.

**Intensity Verification**

Also, by analyzing the mean sea level pressure with time (figure 3.3), the chosen simulation (line d) closely matches the overall intensity of the wave and the transition to a tropical cyclone. Two of the other WRF simulations developed sooner than observed (lines b and c), while the other simulation did not result in tropical cyclogenesis and the wave begins to dissipate by the end of the simulation (line a). From the 18th to the 22nd, the chosen WRF simulation closely matches the ERA-Interim (line e) semi-diurnal pressure fluctuations. On 22 August, a weak low level circulation develops in the WRF simulation and can be seen as a lowering of the pressure by about 5 hPa from the ERA-Interim reanalysis. The development of the low-level circulation, which is not captured by the ERA-Interim, indicates a regime change from a true AEW, which has a circulation concentrated in the mid-levels, to a pre-Wind Induced Surface Heat Exchange (WISHE, Emanuel 1986) disturbance. From 25 August 2006 to 27 August 2006 a continual drop in sea level pressure is observed in the WRF simulation, which is associated with tropical cyclogenesis, and closely reflects the observed drop in sea level pressure measured by the NHC (line f). Not surprisingly, the ERA-Interim reanalysis fails to capture the dropping central pressure between 25 August and 27 August and is likely because of the course 1.5° resolution failing to represent the newly forming tropical cyclone.

**Precipitation Verification**

Figure 3.4 shows a comparison between the precipitation fields as simulated by WRF (a) and as observed in the TRMM 3B42 data set (b), averaged over a two-day period from 0000 UTC 19 August 2006 to 0000 UTC 21 August 2006. The TRMM-observed precipitation field in figure 3.4(b) shows an observed precipitation band west of 20°W centered between 5°N and 10°N. The WRF simulation accurately places the observed precipitation band, but it is slightly stronger in intensity than observed, as indicated by the presence of more shading over 64 mm/day. West of 35°W the WRF simulated precipitation associated with the ITCZ is displaced north of the TRMM-observed precipitation.

Figure 3.4 (c) and (d) shows the WRF simulation and TRMM-observed precipitation rate (mm/day) averaged between 0000 UTC 22 August 2006 and 0000 UTC 24 August 2006. Dur-
ing the averaging time, the wave travels between about 37°W and 55°W and shows an overall northward shift in precipitation from the previous averaging time. The observed precipitation is centered near 10°N during the averaging period and transitions from weaker rain rates in the east, about 4 to 8 mm/day, to stronger rain rates in the west, greater than 32 mm/day in some locations. The weaker rain rates observed near 37°W indicates decreased convection with the wave around 0000 UTC 22 August 2006, before the convective activity increases in the following two days. The WRF simulation reflects the observed structure with the weakest precipitation intensity near 37°W, increasing toward the west. The main difference from the observed precipitation is the lack of precipitation over South America near 55°W in the WRF simulation. In the WRF simulation, the precipitation over the continent is displaced west of 60°W (not shown) with similar intensity to what is observed by the TRMM. This is of little concern for our study as the precipitation pattern directly related to the pre-Ernesto AEW appears consistent with observation.

Figure 3.4 (e) and (f) shows the WRF simulation and TRMM-observed precipitation rate (mm/day) averaged between 0000 UTC 25 August 2006 and 0000 UTC 27 August 2006, days 8 and 9 of the simulation. Remarkably, the WRF simulated precipitation pattern (figure 3.4(e)) closely matches the observed precipitation pattern (figure 3.4(f)). The primary area of interest is the precipitation over the Caribbean, which is associated with the developing tropical cyclone in both the observed and simulated cases. The WRF simulation tracks the wave slightly faster than observed, spanning from about 62°W to 75°W, while the observed precipitation is shifted eastward, spanning from about 60°W to 73°W. This agrees with the track information from figure 3.2, which shows the WRF simulation placing Ernesto about a degree farther west than the NHC best track data. The WRF simulation is slightly wetter than observed but captures the structure well, even capturing the northward track of the storm on day 9 of the simulation.

**Verification Conclusions**

The track, intensity and precipitation field closely reflects the observed long-lived pre-Ernesto AEW. While the simulation is not a perfect match, that is not the goal of this study. The WRF simulation provides a high resolution data set of a long-lived AEW, that is remarkably similar to observation, which will allow for more in-depth analysis that was not previously possible with the reanalysis data.
3.2 Modeling Results

3.2.1 Wave Evolution and Wave Pouch Analysis

Wave Pouch Evolution

Following Dunkerton et al. (2009), the wave is analyzed from a Lagrangian frame of reference, with the wave motion subtracted from the total flow. The wave speed is estimated from the Hovmoller diagram in figure 3.6, where the 850 hPa v-wind is averaged between 7°N and 17°N. The analysis shows an initial wave speed of about 8.6 ms\(^{-1}\) until 20 August, when it speeds up to about 11 ms\(^{-1}\) before slowing to about 7.11 ms\(^{-1}\) on 24 August. These wave speed estimates are then subtracted from the total u-wind component, resulting in a wave relative frame of reference.

Figure 3.7 shows the evolution of the wave pouch at several levels throughout the simulation. The wave pouch is well defined above 700 hPa by 00Z August 20 (48 hours), but shows a northward tilt at 850 hPa and no closed pouch at 925 hPa. This structure mates well with the observation in previous studies (Vizy and Cook 2009 and Zawislak and Zipser 2010), where the equatorward tilt with height persisted for a few days over the Atlantic due to the vertical shear associated with the SAL. By 00Z August 22 (96 hours) the wave pouch is well defined at all levels, which corresponds well to the time when the surface circulation develops. Shortly after this time, the wave encountered vertical wind shear, subsidence and drying in the upper levels. The effects of the synoptic scale influence can be seen at the 350 hPa level, where the pouch has given way to streamlines that flow directly through the wave. However, the pouch remains stacked and disturbed below 500 hPa. By 00Z August 26 the vorticity in lower troposphere has become concentrated in the center of the wave pouch, showing a contraction in scale from earlier time frames. The contraction in scale shows a merger of smaller scale vorticity maxima with time as the system moves toward TC genesis, consistent with the upscale aggregation of PV shown in the bottom-up pathway for TC genesis.

Figure 3.8 shows the evolution of the wave pouch at 600 hPa, in the wave relative frame, and the simulated reflectivity over the 8 day period. Convection is concentrated near the center of the wave pouch, which is consistent with the findings in Dunkerton et al. (2009) and Wang et al. (2010a). The pouch feature allows for a focal point for the convection, which allows for a confined area where the vorticity generated by the convection can more easily aggregate upscale and feed back to the parent wave.
Wave Pouch and Relative Humidity

Figure 3.9 shows the evolution of the wave pouch and relative humidity field at 700 hPa. For the first 72 hours the wave pouch is surrounded by fairly moist air (generally great than 60% humidity). However, by 00Z 23 August (figure 3.9(e)) very dry air, near 30% RH, is in place on the north side of the wave. However, by examining the wave-relative streamlines, we can see that wave pouch protects the air inside from the dry air. The air near the center of the wave pouch is repeatedly recirculated and not exposed to the outside dry air, allowing it to remain more moist than the environmental air. By 00Z 24 August (figure 3.9) some of the dry air from the northern side of the wave has been advected around the south side of the wave, but the air within the center of the wave pouch remains moist.

The evolution of the RH field at the 350 hPa level (figure 3.10) is different from the 700 hPa evolution. A pouch structure develops at 350 hPa by 00Z August (48 hours), indicating that the wave has developed some vertical extent into the upper troposphere and that the mean wind speed at that level closely matches the translation speed of the wave. The pouch structure at this level helps to protect the center of the wave from dry air intrusions, since the very dry air (less than 10% RH) to the north and west is directed around the center of the wave. However, by 00Z 23 August (120 hours) the pouch breaks down as the upper level background wind speed increases and shears the wave. The dry air intrusion in the upper levels can also be seen in the height-time plot (figure 3.11(c)). The dry air is associated with synoptic subsidence that inhibits the convection on the western half of the wave (as can be see in figure 3.8(e)).

Height-time Analysis of the Wave Pouch

Height-time cross sections have also been developed following the wave pouch. Figure 3.11 shows the evolution of the parameters within a 440 km x 440 km box centered on the wave pouch. The size of the box was chosen to include only the wave pouch and thus the area-averaged quantities are representative of the environment within the wave pouch. For the first 72 hours, the PV and absolute vorticity are maximum in the mid-troposphere and consistent with an AEW. Between 72 hours and 96 hours both fields suggest a descending maximum, reminiscent of the top-down school of thought for TC genesis (although later analysis will show this as a bottom-up process). During the first 96 hours, the relative humidity in the wave pouch remains above 70% up to about 400 hPa. The divergence profile is consistent with a deep convective profile, with the strongest convergence located in the boundary layer and strongest divergence located between 200 hPa and 100 hPa. The diabatic heating profile is also consistent with deep convection with strongest heating concentrated in the mid-troposphere. Between 48 and 72 hours there is a
minor disruption in the strong convective divergence profile, with a weakening of the surface convergence and a weakening of the strong divergence aloft. This disruption occurs concurrently with a stabilization (neither increasing or decreasing) in the mid-level vorticity and PV maxima. Between 72 hours and 96 hours, the deep convective divergence profile returns, along with the deep convective diabatic heating profile. With a return of the deep convective heating profile, the PV and vorticity fields begin to strengthen and descend from the mid-levels again, showing that the convection is key to supporting the parent wave.

Between 96 hours and 126 hours, the strongest vorticity and PV signature is located near the top of the boundary layer, and much lower than what is expected for a pure AEW. The development of vorticity near the surface is consistent with a low-level circulation that develops about 96 hours into the simulation, indicating that the wave has begun to transition to a pre-WISHE disturbance. Not long after the surface circulation develops, about 102 hours into the simulation, the mid-level vorticity begins to weaken. The weakening of the mid-level vorticity coincides with very dry air intruding on the wave pouch between 400 hPa and 250 hPa (3.11(c)). Also during this time, the divergence profile (3.11(d)) is no longer representative of deep convection as the level of maximum divergence is located in the mid-troposphere. The disruption in the convective heating profile can also be seen in the diabatic heating (3.11(e)) as the level of maximum heating is confined below 500 hPa. This shows that the large scale subsidence suppresses the convection within the wave, which leads to a weakening of the parent wave, which is seen in the weakening vorticity and PV fields.

After 126 hours, the divergence and diabatic heating transition back to a deep convective profile. The low level vorticity and PV stop decreasing and stabilize through about 180 hours. The relative humidity begins to slowly increase above 500 hPa after the deep convective profile returns to the wave pouch, thus implying the return of deep convection within the wave helps to moisten the upper levels of the troposphere. The convective heating profile appears to be essential in maintaining or strengthening the wave; however, the PV and vorticity budgets in the following sections help reveal the pathway through which this mechanism works to strengthen the wave.

It is not until a strong convective burst occurs near hour 204, that tropical cyclogenesis occurs. The diabatic heating field reveals the strongest heating rates of the simulation centered in the mid-troposphere along with a deep layer of convergence from the surface to near 300 hPa and strong divergence from 300 hPa to 100 hPa. The higher relative humidity values begin to increase in height up to upper troposphere during this time. The absolute vorticity profile also begins to increase in height with time, which in combination with the relative humidity field
suggests a vertical development of the vortex into the upper troposphere. This sequence seems to fit the bottom-up theory for tropical cyclogenesis, where the low level vorticity is stretched and advected upward, creating a deep, coherent vortex that becomes the tropical cyclone.

**Wave Pouch Analysis Conclusions**

The height-time cross sections show an association between convection within the wave and a maintaining or strengthening wave. When the convection becomes suppressed within the wave, the wave scale vorticity and PV decreases. The convection within the wave pouch promotes diabatic heating in the middle troposphere, convergence in the lower to mid-troposphere and divergence in the upper troposphere. This supports the findings of Dunkerton et al. (2009), where the wave pouch is characterized by a convective heating profile, enhanced RH and enhanced vorticity.

The wave pouch analysis also shows the presence of a pouch between 500 hPa and 700 hPa from 24 hours to 216 hours (the end of the simulation). A deep pouch (500 hPa to 925 hPa) develops by 96 hours or about 4 days prior to TC genesis. The presence of this pouch in the lower troposphere likely helps to ensure the integrity of the wave when it experiences the detrimental effects of synoptic scale subsidence. After the interaction with the upper level dry air and subsidence, the wave pouch in the lower troposphere still provides a favored region for convection and aggregation of vorticity. This time series also suggests that convective coupling may be key to maintaining or strengthening the mid-level and low-level vorticity; however the direction of causality is hard to infer from these plots and will be examined more closely in the PV and vorticity budget sections that follow.

Dunkerton et al. (2009) only discuss the importance of the pouch within a few days prior to TC genesis, while this case study shows this structure may exist long before TC genesis begins. The wave pouch persists for 7 days prior to tropical cyclogenesis and may be an important mechanism for protecting the convection within the wave. The pouch may also help cascade energy from the mesoscale storm features to the synoptic scale wave due to the enhanced cyclonic vorticity and weak deformation.

### 3.2.2 Potential Vorticity Budget

To examine and quantify the processes responsible for maintaining the pre-Ernesto AEW, this study uses the potential vorticity budget originally described in Haynes and McIntyre (1987)
and adapted to a storm-following form by Kieu and Zhang (2009). Kieu and Zhang (2009) formulate the PV budget equation from Haynes and McIntyre (1987) into a volume averaged form, which allows for a time series analysis of the budget in a storm following format; and is written as:

\[
\frac{d}{dt} \left( \int_{V(t)} q dV \right) = \int_{V(t)} (q \nabla \cdot u) dV + \int_{V(t)} \frac{\omega \cdot \nabla H}{\rho} dV \\
+ \int_{V(t)} \frac{\nabla \cdot (F \times \nabla \theta)}{\rho} dV + \int_{S(t)} q (U - u) \cdot n dS
\]  \tag{3.1}

where \( q \) is the potential vorticity, \( u \) is the full 3D wind field, \( \omega \) is the 3D absolute vorticity, \( H \) is the diabatic heating rate, \( F \) is the 3D frictional force, \( U \) is the speed of the boundaries and \( \nabla \) is the 3D gradient operator. This formulation of the PV budget states that the volume averaged time rate of change of PV is described by the terms on the right hand side. Those terms are, from left to right, the 3D divergence of PV or stretching term, the diabatic PV production term, the frictional generation/destruction term and the across boundary PV flux. To make the computation easier, it should be noted that the across boundary PV flux term can be rewritten using Gauss theorem, which states that a surface flux can be written in terms of the volume integral of the divergence within the enclosed region. Rewriting the last term yields:

\[
\frac{d}{dt} \left( \int_{V(t)} q dV \right) = \int_{V(t)} (q \nabla \cdot u) dV + \int_{V(t)} \frac{\omega \cdot \nabla H}{\rho} dV \\
+ \int_{V(t)} \frac{\nabla \cdot (F \times \nabla \theta)}{\rho} dV + \int_{V(t)} \nabla \cdot (qU) - \nabla \cdot (qu) dV
\]  \tag{3.2}

While this form of the PV equation may provide useful information in a storm total frame, information about the vertical distribution of the budget terms will be lost, which will be especially useful when analyzing an AEW, which show little signal above 500 hPa. To better understand the vertical structure of the system we need to rewrite the PV budget equation in an area average format. Thus, the area average form of the equation is written as:
\[
\frac{d}{dt} \left( \int_{S(t)} q dS \right) = \int_{S(t)} (q \nabla \cdot u) dS + \int_{S(t)} \frac{\omega \cdot \nabla H}{\rho} dS \\
+ \int_{S(t)} \frac{\nabla \cdot (F \times \nabla \theta)}{\rho} dS \\
+ \int_{S(t)} (\nabla \cdot (q U) - q \nabla \cdot u - u_h \cdot \nabla q) dS \\
+ \int_{S(t)} -\left( \frac{dwq}{dz} \right) dS
\] (3.3)

where \( u_h \) is the horizontal components of the wind, \( \nabla_h \) is the 2D gradient operator and \( w \) is the z-direction wind speed. This formulation of the PV budget shows that the time rate of change of area average PV on a height surface can be explained by the terms on the right hand side. The first three terms on the right hand side of the equation are the same as in equation (1) except that instead of volume averaging the terms are averaged at each height surface, thus retaining the budget information in the vertical. For clarity, the boundary term is broken into two terms, the first describes the net flux of PV across the lateral boundaries of the height surface, while the second term is the area average vertical flux of PV. It is important to separate the vertical flux from the boundary term in the area average context so that it is possible to distinguish between PV being fluxed into or out of the area of interest and PV that is within the area of interest but is moving between height levels due to vertical motion.

The PV budget is calculated using the output from the 4-km model domain at 12 minute intervals. Before plotting the height-time cross sections of the PV budget a 1-hour running average is applied to the fields to help eliminate some of the high-frequency noise in the computation. The friction term is ignored in this study since the focus is on levels above the boundary layer, specifically in the middle troposphere. The center of the wave pouch is recorded every 6 hours and is interpolated to a 12-minute position for centering the PV budget box for area or volume averaging. The speed of the lateral boundaries (U) is calculated from the centered difference of the 12 minute positions.

The diabatic heating rate was originally read in directly from the WRF output files as a combination of the heating due to the microphysics, planetary boundary layer and radiation schemes. However, initial PV budget analysis resulted in a residual term with distinct over- and under-estimates around the melting level. The same pattern was noticeable in the diabatic PV generation term and so an alternate method for obtaining the diabatic heating was implemented. The diabatic heating term used in the PV budget analysis was obtained by computing
the total derivative of theta \( \frac{D\theta}{Dt} \) at each time and subtracting the 3D theta advection \( -u \cdot \nabla \theta \). In doing so, it is assumed that any changes in theta not accounted for by advection are diabatic heating tendencies. In theory, this method should be more accurate as it automatically includes numerical diffusion, mixing and frictional effects, which cannot be read from the WRF output. To check that this method accurately describes the diabatic heating tendency, it was compared to the diabatic heating tendency read directly from the WRF file. The comparison showed the two methods to be almost identical, except near the melting level where some minor differences appeared. Calculating the PV budget with the calculated diabatic heating term eliminated the residual problems centered on the melting level, and improved the overall accuracy of the PV budget calculation.

### 3.2.3 Potential Vorticity Budget Analysis

Figure 3.12 shows the results from the PV budget in height-time format for easy comparison with figure 3.11. All terms are area averaged at each height level over a 440 km x 440 km box centered on the wave pouch. To better understand how the mid-troposphere potential vorticity anomaly is maintained, evolves and descends to the surface, each term of the potential vorticity budget must be carefully considered at all phases in the evolution of the wave. First a general overview of the budget will be discussed with the aid of the height-time plots, followed by a more careful analysis of each phase in the evolution using time-averaged profiles of the budget and linking that information back to the meteorological conditions during each phase.

#### Divergence Term Overview

Throughout the course of the simulation, the divergence term (figure 3.12(b)) consistently contributes a positive PV tendency throughout much of the depth of the troposphere. This term can be related to the mass exchange across the boundaries of the box, which leads to either a concentration or dilution of PV within the box. The times of strongest positive tendency in the divergence term can be related to the times of strongest diabatic heating (figure 3.11(e)) as well as the times of the strongest deep convection divergence profile (figure 3.11(d)). The strong diabatic heating produced by the convection lowers the density and pressure of the air in the box. This results in a net mass loss from the box, as shown in Kieu and Zhang (2009). Since PV is scaled by the mass in a given volume, this acts to increase PV. This suggests that the presents of convection within the wave is very valuable to maintaining the wave itself and also suggests a pathway from sustained convection to reinforcing the wave-scale potential vorticity maximum.
Diabatic + Vertical Flux Term Overview

The diabatic tendency term and vertical flux term have been added together for presentation in the height-time format (figure 3.12(c)) but will be displayed independently in the next section of time-average vertical profiles. It is easier to understand the net effect of the diabatic term and vertical flux term when they are combined since the terms are generally equal and opposite in sign. The diabatic tendency is typically positive in the lower troposphere and negative in the upper troposphere due to the deep convective diabatic heating profile. Since the maximum level of heating is in the mid-troposphere, this acts to add static stability to the lower portion of the atmosphere and decrease the static stability of the upper portion of the atmosphere. By definition, PV can be increased by increasing the static stability, thus the PV tendency due to the diabatic heating profile should be positive below the level of maximum heating.

Conversely, the vertical flux term is largely negative in the lower troposphere and positive in the upper troposphere, opposite to the diabatic term. This pattern is explained by the vertical flux divergence of PV in the low-levels, where it is being diabatically generated, and flux convergence in the upper levels. Therefore, the vertical flux of PV and the diabatic generation of PV are nearly in a steady state. Thus, since the patterns largely cancel, more information can be gathered from the net effect of the two, offsetting processes. With these two processes combined, we find that the net effect of the vertical flux and diabatic heating is generally a negative PV tendency in the mid-troposphere, generally acting in opposition to the divergence term.

Boundary Term Overview

The boundary term (figure 3.12(d)) is generally much smaller than the preceding terms. This should be the case since the averaging box is designed to contain the bulk of the PV anomaly associated with the wave, and thus only very small fluxes should occur through the edges of the box. However, if the fluxes through the side of the box are larger than their typical near-zero values it can be interpreted as a flux into or out of the wave itself due to the careful placement and size of the averaging box. Throughout the simulation, the boundary term is largely near-zero except for a few time frames where some stronger negative tendencies are seen in the mid-levels.

Residual Term Overview

The residual term (figure 3.12(e)) is fairly small and random compared to the leading terms, with the exception of a strong negative tendency near the surface, which can be directly at-
tributed to frictional affects. The negative tendency due to friction is especially more noticeable once the low-level circulation forms just after 96 hours. Also, once the convection becomes increasingly active in the wave, after 120 hours, there is a net positive tendency that appears in the residual term in the mid-troposphere, but this is still smaller than the leading terms of divergence, diabatic heating and vertical flux. Overall, the residual term does not discredit the usefulness of the PV budget terms in finding some understanding about dynamics and thermodynamics of the AEW and its transition to a tropical cyclone.

**PV Budget Evolution**

Between hours 24 and 96 the mid-level PV associated with the wave increases from about 0.5 PVU to 0.9 PVU. The PV is also descending with time and the low-level PV continues to increase.

- The divergence term (figure 3.12(b)) is mainly positive throughout this period of PV maintenance and strengthening. The divergence term is positive throughout much of the troposphere suggesting that diabatic heating is working to decrease the mass within the volume and contributing to the overall concentration of PV, leading to a positive PV tendency.

- The diabatic / vertical flux term (figure 3.12(c)) is primarily negative throughout the depth of the troposphere, but especially in the low to mid levels.

- The boundary term (figure 3.12(e)) is mainly small and noisy compared to the preceding terms, with the exception of shortly after 72 hours in the mid and upper troposphere. Shortly after 72 hours, only the boundary term can account for the negative PV tendency in the mid to upper levels, which suggests a flux of higher PV air out of the wave and lower PV air into the wave.

The time-averaged PV budget profile (figure 3.13(a)) for hours 24 to 96 confirms the findings of the height-time plot analysis. The main contributor to the positive PV tendency in the middle and lower troposphere is the divergence term. Here we can see the opposing effects of the diabatic and vertical flux terms, and when compared to the height-time plot it is easy to see how the terms add up to be overall negative throughout much of the troposphere. Also from this time-average perspective, the boundary term is indeed much smaller than the leading terms.

Hours 96 to 126 of the simulation show the disruption of the deep convective heating profile. This stage of the simulation is characterized by decreasing PV in the mid and upper level, with
the mid level PV decreasing from about 0.9 PVU to about 0.6 PVU.

- The strong mid-level diabatic heating tendency (figure 3.11(e)) is weakened and becomes concentrated in the lower-levels. The divergence profile (figure 3.11(d)) shows the disruption of the deep convective heating profile by the loss of strong divergence in the upper troposphere, and the change from weak convergence to weak divergence in the mid levels.

- In the mid-levels, the positive tendency of the divergence term becomes very weak compared to the negative tendency of the diabatic / vertical flux term and the boundary term, and so an overall decrease in the PV is observed.

- The boundary term is very important in decreasing the mid and upper level PV as low PV air enters the wave while interacting with the large scale subsidence.

- Since the divergence term is still largest in the boundary layer and the diabatic / vertical flux term becomes positive in the boundary layer, the low-level PV is maintained.

The time-averaged PV budget profile (figure 3.13(b)), for hours 96 to 126, reflects the analysis of the height-time plot.

- The divergence term is very weak but still positive and the diabatic / vertical flux term is weak but still negative. The boundary term describes the decrease in mid-level PV well as the wave interacts with large scale subsidence during this time period.

- The suppression of the deep convective heating profile inhibits concentration of PV, thus limiting the only term that can provide a significant PV source for the wave.

Hours 126 to 192 of the simulation illustrate a rejuvenation of the deep convective heating profile (figure 3.11(e)) along with a strengthening and maintenance of the PV associated with the pre-depression.

- The divergence term is the dominate positive tendency term throughout the depth of the troposphere, while the diabatic / vertical flux term is the dominate negative tendency term.

- Figure 3.13(c) shows that even the boundary term contributes negatively to the mid-level PV during this period.

- The only term responsible for the strengthening and maintenance of the wave is the divergence term, which relies on the deep convective heating profile.

During the time of tropical cyclogenesis, between hours 198 and 210, the PV budget shows a continued positive divergence term, while the diabatic / vertical flux term becomes less negative
and the boundary term becomes less negative. Figure 3.13(d) shows that the divergence term is the leading positive contributor to the total PV especially near the surface, and best explains the increase of PV throughout the depth of the troposphere. Since the divergence term did not strengthen from prior times, tropical cyclogenesis may have been the result of the boundary term and the diabatic/vertical flux becoming less negative.

This confirms that long-lived waves may be a result of processes that both support the wave and inhibit the wave. Here, the wave is maintained through convection within the wave pouch, but upper-level shear and subsidence are enough to prevent tropical cyclogenesis. The balance between these competing processes allows the wave to be maintained without developing.

### 3.2.4 Vorticity Budget

To further support the PV budget analysis and to provide more insight into the dynamics of the AEW, a vorticity budget is calculated following Haynes and McIntyre (1987), Kieu and Zhang (2009), and Wang et al. (2010a). The flux form of the vorticity equation in isobaric coordinates is written as:

\[
\frac{\partial \eta}{\partial t} = -\nabla \cdot (u' \eta) - \nabla \cdot \left( -\omega k \times \frac{du'}{dp} \right) + \text{Solenoid} + \text{Friction}
\]  

(3.4)

where \( \eta \) is the 2D absolute vorticity calculated using the wave relative winds, \( u' \), and \( \omega \) is the pressure coordinate vertical velocity. Consistent with the previous studies, the solenoid and friction terms are assumed to be small when compared to the leading terms and are thus not included in the calculation. The solenoid term will only be active where there are large density and temperature differences which are not typically found in the tropics and the friction term will be small except in the boundary layer, which is of little concern to understanding vorticity tendencies in the middle troposphere. The first term on the right hand side is the flux divergence of vorticity, which combines the horizontal advection into or out of the area averaging box and stretching in the vertical. The second term is a combination of tilting horizontal vorticity into the vertical and the vertical advection of vorticity. The tilting portion of second term is considered to be small due to generally weak wind shear in the tropics and thus the vertical advection will dominate this second term.

The vorticity budget is calculated similarly to the PV budget, in that, it is calculated on the 4-km model domain at 12 minute intervals. A 1-hour running average is computed on the 12-minute calculations to help eliminate any high frequency noise in the calculation.
center of the wave pouch is recorded every 6 hours and is interpolated to a 12-minute position for centering the vorticity budget box for area or volume averaging. The speed of the lateral boundaries (U) is calculated from the centered difference of the 12 minute positions.

### 3.2.5 Vorticity Budget Analysis

The vorticity budget is used to gain further understanding of the dynamical processes responsible for maintaining the long-lived AEW and its transition to a tropical cyclone. It is also helpful to further verify the findings of the PV budget. The results from the vorticity budget will be presented the same as the PV budget, first with a general overview in the height-time format (figure 3.14) and then it will be followed up with a more specific discussion using time-averaged profiles from specific events within the wave’s life-span.

**Horizontal Advection / Stretching Term Overview**

Figure 3.14(b) illustrates the horizontal advection and stretching tendency terms of the vorticity budget. The horizontal advection term includes the boundary fluxes into or out of the wave and the convergence of vorticity into the wave pouch center. The stretching term is also important within the wave as strong convective updrafts stretch the already present vorticity. This term is primarily active in the middle and lower troposphere throughout the wave life span. This pattern is expected given the deep convective heating profile within the wave pouch. The lower troposphere convergence leads to a convergence of planetary vorticity into the wave center within the boundary layer. The positive tendency of the horizontal advection / stretching term within the boundary layer matches well with the strong convergence seen in the boundary layer in figure 3.11(d). The positive tendency also generally extends upward into the middle troposphere, especially when the mid-level vorticity is strengthening. This pattern also matches well with the deep convective heating profile, since the upward vertical velocities accelerating to the middle and upper troposphere act to stretch the vorticity.

**Tilting / Vertical Advection Term Overview**

The tilting and vertical advection term in figure 3.14(c) generally has a positive tendency in the middle and upper troposphere and is generally negative in the lower troposphere. The tilting term is expected to be generally small, especially in the tropics where the vertical wind shear is typically small. In this term, the vertical advection is dominate and the pattern seen in the tendencies reflects that. Since the vorticity is generally maximized in the middle troposphere
during the wave lifespan, the positive vertical velocity associated with the deep convective profile will lead to negative tendencies in the lower troposphere and positive tendencies aloft.

**Residual Term Overview**

The residual term (figure 3.14(d)) is small and noisy when compared to the leading terms. The primary coherent feature in the residual term is the negative vorticity tendency in the boundary layer due to the unaccounted frictional spin-down. Also, when the vorticity tendency is positive the residual term is typically small and negative indicating the positive tendencies are accounted for by the combination of the leading terms. Thus the residual term, while not zero, does not detract from the interpretation of the leading terms.

**Vorticity Budget Evolution**

Hours 24 to 96 of the simulation represent a classic AEW with vorticity maximized in the mid-troposphere (figure 3.14(a)). During this period the vorticity is strengthening at the mid-levels from about $7.5 \times 10^{-5}$ s$^{-1}$ to about $13.5 \times 10^{-5}$ s$^{-1}$. The vorticity pattern also shows a downward building vorticity maxima.

- Visually inspecting figure 3.14(b) and (c), during this 3 day period, suggests that the mid-level increase in vorticity is a combination of the horizontal advection / stretching term and the tilting / vertical advection term since both terms appear to be positive during this period.

- The horizontal advection / stretching term is the only term that is positive near the surface, in the boundary layer. This further supports the idea that the horizontal convergence near the surface, due to the low level convergence of the deep convective heating profile, may be key to spinning up low-level vorticity.

- The findings here support the bottom-up camp of TC genesis, in that the convection is vital to producing surface convergence that concentrates low-level vorticity near the wave center.

Figure 3.15(a) shows the vorticity budget tendencies profile, time-averaged over the 3 day period.

- The main contributor to the low-level vorticity is the horizontal advection and stretching term. It is balanced by negative tendencies due to the vertical advection term and by
surface friction; however, the surface convergence is strong enough to prevail over the other terms.

- Between 800 hPa and 600 hPa, both the horizontal convergence/stretching and tilting/vertical advection terms contribute positively to the mid-level vorticity tendency.

- Above 600 hPa the vertical advection term becomes negative due to the vertical gradient reversal in vorticity above the mid-troposphere, while the horizontal advection/stretching term remains positive. The combination of these terms above the 600 hPa level results in a decreasing vorticity tendency with height, although it remains positive because of the large positive horizontal advection/stretching term.

- During this 3-day strengthening phase, the primary contributor to the low-level vorticity is through the convergence and stretching of vorticity, while the wave-level vorticity is strengthened through the combination of vertically advecting the low level vorticity and stretching.

Between hours 96 and 126 of the simulation, the mid-level vorticity decreases from greater than 13.5 $\times 10^{-5}$ s$^{-1}$ to just less than 12 $\times 10^{-5}$ s$^{-1}$ while the low-level vorticity remains fairly constant (figure 3.14(a)). During this time period, the deep convective heating profile is disrupted by subsidence in the upper levels that leads to drying above 500 hPa (figure 3.11(c)).

- Figure 3.11(e) shows the disruption of the deep convective heating profile, with the magnitude of heating weakening and the level of maximum heating lowering. This subsidence pattern can also be seen in the disruption of the deep convective divergence pattern (figure 3.11(d)), where the strong upper level divergence signature is disrupted and the mid-level weak convergence becomes mid-level weak divergence.

- The horizontal advection/stretching term remains positive in the boundary layer but becomes mostly negative in the mid-levels.

- The tilting/vertical advection term remains positive in the upper troposphere, but becomes weak and slightly negative in the mid-levels and slightly positive in the boundary layer.

Figure 3.15(b) shows the time-averaged vorticity budget tendency terms between 96 and 144 hours.

- With the low-level convergence surviving through this period of weakened convection, the horizontal advection/stretching terms remains positive in the boundary layer.
• The development of a low-level circulation near 96 hours likely contributes convergence and allows a mechanism for maintaining positive vorticity tendencies near the surface despite the suppression of deep convection.

• In the mid-levels, both the horizontal advection/stretching term and the tilting/vertical advection terms contribute to the negative vorticity tendency. The weakened convective heating profile suggests a shallower layer of ascending air which limits the stretching term. The divergence in the mid-levels will also act to spin down vorticity and to advect higher vorticity air in the mid-levels away from the wave pouch.

• This advection out of the wave pouch can be confirmed through the boundary term of the PV budget (figure 3.12(e)), where a mid-level negative vorticity tendency is due to the flux of high PV air out of the wave and low PV air into the wave.

• The weakened convection during this period significantly impacts the deep convective heating profile that appears to be optimal in maintaining and strengthening the wave level vorticity.

Between hours 126 and 192 the deep convective profile returns to the wave as the synoptic subsidence is no longer affecting the system. This portion of the wave lifespan is inherently different than hours 24 to 96 as the system has developed a low-level circulation and is no longer a pure AEW, but instead a pre-depression disturbance. The vorticity budget reflects this inherent change in the structure of the system as it displays some distinct differences from the 24 to 96 hour period. Figure 3.15(c) illustrates that the horizontal advection/stretching term is still the dominate positive contributor in the middle and lower troposphere. However, the tilting/vertical advection term becomes negative in the middle troposphere and positive within the boundary layer and continues to be positive in the upper troposphere. Despite the low-level convergence profile seen in figure 3.11(d) the horizontal advection/stretching term is small within the boundary layer, suggesting that the convergence of low-level vorticity may no longer be enough to increase the total vorticity since a surface vortex is present. However, during the times of greatest mid-level increases in vorticity, the horizontal advection/stretching term is very active in the mid-troposphere and this is confirmed in figure 3.15(c) as the only positive tendency source in the mid-levels. This suggests that once a deep layer vortex is established from the surface through the mid-troposphere, the deep convective heating profile supports strong stretching of the vorticity and leads to a spin-up of vorticity throughout the vortex depth. The tilting/vertical advection term is negative in the mid-troposphere suggesting that the strong vertical velocities associated with the deep convection is working against the stretching term by advecting lower vorticity values from near the surface upward into the mid-troposphere.
At the time of tropical cyclogenesis, around 205 hours, the vorticity budget shows the dominate positive tendency term in the lower to mid-troposphere to be horizontal advection/stretching term. Given the strong vertical velocities associated with the deep convective profile and the abundance of vorticity within the center of the wave pouch, it is not surprising that the stretching term should dominate.

Conclusions from the Vorticity Budget

The vorticity budget provides support for the findings of the PV budget. Through 96 hours the wave is primarily maintained and strengthened through the presence of the deep convective heating profile, where low-level vorticity is concentrated by convergent winds in the lower troposphere. The vorticity is then advected vertically and stretched leading to an increase in vorticity in the mid-levels feeding back to the wave scale vorticity. Once a deep vortex is formed, from the surface to the mid-troposphere, stretching becomes the dominate mechanism for maintaining the mid-level vorticity. This process closely reflects the bottom-up school of thought for tropical cyclogenesis. Here, we also see that the important factors for making this case a long-lived AEW is having a balance between processes that maintain the wave and processes that are detrimental to the wave.

3.3 Effects of Stratiform and Convective Precipitation

The precipitation is classified as convective or stratiform following Tao et al. (1993), Braun et al. (2010) and Wang et al. (2010b). The procedure for separating the convective regions from the stratiform regions is described in detail in Wang et al. (2010b). The precipitation type is determined using the data on the 4 km grid, since it is a convection resolving grid. The purpose of separating the convective and stratiform regions is to examine the vertical structure of these regions and their relative roles in the vorticity tendencies within the wave. Figure 3.16 shows a snapshot at hour 207 of the split between convective precipitation and stratiform precipitation, where figure 3.16a shows the total rain rate, figure 3.16b shows the convective portion of the total rain rate and figure 3.16c shows the stratiform portion of the total rain rate. This shows the convective portion of the precipitation is smaller in spatial scale than the stratiform regions, however the convective portions produce the strongest rain rates. Wang et al. (2010b) found that while the convective region is smaller in spatial scale, it contributes at least as much total precipitation as the stratiform regions. They also found that the convective regions are concentrated closer to the center of the wave pouch, while the stratiform regions are broad and
extend farther from the center.

The diabatic heating profiles match expectations for the convective and stratiform regions (figure 3.17a and 3.17b). The convective heating profile (3.17a) shows diabatic heating from the surface to near 250 hPa, centered in the mid-troposphere. The stratiform precipitation heating profile (3.17b) illustrates the level of maximum heating in the mid-troposphere with cooling in the lower troposphere. The relative magnitude of the heating from the stratiform regions is much less than that contributed from the convective regions.

Figure 3.17c and 3.17d show the divergence profiles for the convective and stratiform regions, respectively. The divergence profiles match the diabatic heating profiles and expectations. The divergence profile for the convective region is characterized by convergence below 700 hPa and divergence aloft, while the stratiform region shows convergence centered between 600 hPa and 500 hPa, with divergence above and below. Our findings match those in Wang et al. (2010b), where the near surface convergence from the convection is stronger than the near surface divergence from the stratiform regions. In the mid-levels, the stratiform convergence is stronger than the convective divergence creating a net convergent profile. Strong divergence is created in the upper-troposphere by the combination of the divergence from the convection and stratiform regions. The net effect is deep convergence from the surface to near 300 hPa with divergence above 300 hPa.

Wang et al. (2010b) find the convective heating profile to become dominate, especially approaching the time of TC genesis, despite the smaller spatial coverage. This suggests that the convective heating profile is most important for developing a low-level circulation and creating a deep vortex for tropical cyclogenesis. However, at the times prior to TC genesis Wang et al. (2010b) suggest that the convective and stratiform contributions are fairly equivalent.

3.3.1 Vorticity Budget for Convective vs Stratiform Regions

To better understand how each precipitation type contributes to maintaining or strengthening the AEW, the vorticity budget has been divided by precipitation type. The vorticity budget is calculated at all grid points and is then divided by precipitation type prior to area averaging for the height-time plots. This allows examination of the vertical structure of vorticity tendencies that are generated in each precipitation type region.

The vorticity tendency produced by the convective regions is primarily positive in the middle
and lower troposphere (figure 3.18a), where it is most important for maintaining and strengthening the wave. However, in examining the vorticity tendency produced in the stratiform regions, it is primarily negative in the middle to lower troposphere (figure 3.19a). Especially in the maintenance period between 144 and 192 hours, we see the convection is responsible for producing positive vorticity tendencies while the stratiform and the non-precipitating regions (figure 3.20a) are responsible for negative vorticity tendencies.

The processes responsible for the positive vorticity tendencies in the convective regions are the horizontal advection, stretching and vertical advection terms.

- The horizontal advection term (figure 3.18b) can be thought of as the vorticity flux into or out of the convective regions. The horizontal advection term in the stratiform regions (figure 3.19b) is roughly opposite in pattern and magnitude to the convective regions. This suggests that vorticity is readily traded from the stratiform regions to the convective regions within wave. If these two terms are added (not shown), the horizontal advection term is very small, showing there is large cancellation within the box from vorticity being moved between these regions.

- The stretching term in the convective regions (figure 3.18c) is large and positive in the lower troposphere and extends into the middle troposphere toward TC genesis time. The vertical advection term in the convective regions (figure 3.18e) is also large and positive, but in the middle and upper troposphere. The combination of these terms suggests the convection acts to converge and stretch low-level vorticity and then advect it vertically.

- The tilting term in the convective regions (figure 3.18d) is mainly negative, but not large enough in magnitude to offset the three other positive vorticity generation processes.

3.3.2 Conclusions about Convective vs Stratiform Precipitation on the Wave

The analysis shows the convective precipitation regions within the wave to be the primary mechanism for maintaining and strengthening the wave. The stratiform precipitation regions actually tend to produce negative vorticity tendencies. The vertical diabatic heating profile and vertical divergence profile of the convective regions are shown to best contribute to the positive vorticity tendencies needed to maintain or strengthen the wave by acting to produce vorticity in the lower to middle troposphere.
Figure 3.1: Model domains as initialized on 0000 UTC 18 September 2006. The outer domain is a 12 km fixed domain while the inner domain (d02) is a 4 km moving nest that remains centered on the pre-Ernesto wave.
Figure 3.2: Comparison between the WRF simulated AEW track (black line with white dots and labels) and the ERA-Interim track (blue line) and NHC best track data (red line).
Figure 3.3: Comparison between the WRF simulations, ERA-Interim and NHC mean sea level pressure between 00Z 18 August 2006 and 00Z 27 August 2006. Line (a/blue) is the WRF simulation with WSM-3 microphysics, line (b/red) is the WRF simulation with LIN microphysics, line (c/orange) is the WRF simulation with WSM-6 microphysics and standard surface fluxes (ISFTCFLX=0), line (d/green) is the WRF simulation with WSM-6 microphysics and hurricane specific surface fluxes using the Garratt formulation (ISFTCFLX=2), line (e/black) is the ERA-Interim reanalysis, line (f/purple) is the NHC best track.
Figure 3.4: Composite precipitation (mm/day) for the pre-Ernesto AEW. The WRF simulation (a) and the TRMM 3B42 (b) averaged between 0000 UTC 19 August 2006 and 0000 UTC 21 August 2006. (c),(d) as in (a),(b) except averaged between 0000 UTC 22 August 2006 and 0000 UTC 24 August 2006. (e),(f) as in (a),(b) except averaged between 0000 UTC 25 August 2006 and 0000 UTC 27 August 2006.
Figure 3.5: The result of a thought experiment to compare the typical Eulerian view to the Lagrangian view of an AEW like vortex. Panel (a) shows a Rankine vortex embedded in easterly flow to represent a typical easterly wave. Panel (b) shows the same vortex with the same background flow, but subtracts a typical AEW phase speed from the total flow. The solid vertical lines indicate the wave trough, while the solid horizontal lines indicate the critical latitude for this case. The bold, solid contour in panel (b) indicates the boundary of the wave pouch, where the air inside this streamfunction line is recirculated and protected from outside intrusion.
Figure 3.6: Hovmoller plot of the 850 hPa v-wind averaged between 7°N and 17°N for the pre-Ernesto AEW (2006) from the WRF simulation. The solid line denotes the trough axis. The three estimated phase speeds of the wave can be seen by the change in slope of the solid line with time.
Figure 3.7: PV and wave-relative streamlines in a 720 km x 720 km box, center on the wave pouch. Plot levels increasing with height from bottom to top and time increasing from left to right. This figure illustrates the evolution of the wave pouch with height throughout the course of the simulation.
Figure 3.8: Simulated radar reflectivity (DBZ) overlaid on 600 hPa wave-relative streamlines in a 720 km x 720 km box, center on the wave pouch.
Figure 3.9: Relative humidity (%) and wave-relative streamlines at 700 hPa in a 1200 km x 1200 km box, centered on the wave pouch.
Figure 3.10: Relative humidity (%) and wave-relative streamlines at 350 hPa in a 1200 km x 1200 km box, centered on the wave pouch.
Figure 3.11: Height-time cross sections averaged over a 440 km x 440 km box, centered on the wave pouch for the 9-day simulation. Potential vorticity in 10*PVU (a), absolute vorticity in $10^{-5}$ s$^{-1}$ (b), relative humidity in percent (c), divergence in $10^{-5}$ s$^{-1}$ (d) and diabatic heating in K s$^{-1}$.
Figure 3.12: Height-time cross section of the PV budget averaged over a 440 km x 440 km box. (a) PV (contoured from 0 to 1.5 PVU by 0.1) and PV tendency (shaded), (b) divergence term, (c) diabatic term, (d) vertical advection, (e) boundary flux, (f) residual. All tendency terms in $10^{-5}$ PVU s$^{-1}$. The y-axis is the height meters and the x-axis is hours since 00Z 18 August 2006.
Figure 3.13: Time-average PV budget terms. All tendency terms are in $10^{-5}$ PVU s$^{-1}$. 

(a) Budget Terms - Time Average 24-96 hours
(b) Budget Terms - Time Average 96-126 hours
(c) Budget Terms - Time Average 126-192 hours
(d) Budget Terms - Time Average 198-210 hours

Divergence
Vertical Advection
Diabatic
Boundary
Figure 3.14: Height-time cross section of the vorticity budget averaged over a 440 km x 440 km box. (a) absolute vorticity (contoured from 0 to 15 $10^{-5}$ s$^{-1}$ by 1.5) and absolute vorticity tendency ($10^{-9}$ s$^{-2}$, shaded), (b) Horizontal Advection and Stretching ($10^{-9}$ s$^{-2}$), (c) Tilting and Vertical Advection ($10^{-9}$ s$^{-2}$), (d) Residual ($10^{-9}$ s$^{-2}$). The y-axis is the height in pressure coordinates (hPa) and the x-axis is hours since 00Z 18 August 2006.
Figure 3.15: Time-average vorticity budget terms. Units are $10^{-9} \text{ s}^{-2}$. 
Figure 3.16: (a) Rain rate, (b) convective precipitation rain rate and (c) stratiform precipitation rain rate at hour 207. Units are mm/hr
Figure 3.17: Height time plots of (a) convective regions heating profile in K s\(^{-1}\), (b) stratiform regions heating profile in K s\(^{-1}\), (c) convective regions divergence profile in 10\(^{-5}\)s\(^{-1}\), and (d) stratiform regions divergence profile in 10\(^{-5}\)s\(^{-1}\).
Figure 3.18: Vorticity budget tendency terms for the convective regions only. Units are $10^{-9}$ s$^{-2}$.
Figure 3.19: Vorticity budget tendency terms for the stratiform regions only. Units are $10^9 \text{s}^{-2}$. 
Figure 3.20: Vorticity budget tendency terms for the non-precipitating regions only. Units are $10^{-9}$ s$^{-2}$. 
Chapter 4

Conclusions

4.1 Conclusions

This study attempts to fill a gap in understanding the evolution of AEWs as they move across the tropical Atlantic. Specifically, we examine long-lived AEWs, which must make it past 60°W before either developing or dissipating. As the AEWs move away from the AEJ and into the tropical Atlantic, they lose their main mechanism for growth. Therefore, for waves to maintain over the tropical Atlantic there must another mechanism responsible for maintaining or strengthening the AEWs.

We hypothesize that waves can be maintained or by two mechanisms:

- A flux of PV into the wave from mid-latitude or tropical sources
- PV generation from convection within the wave that cascades energy upscale to support the parent wave

The observational study suggests that flux of PV into the wave may be rare and that convection within the wave is the primary contributor to wave maintenance.

4.1.1 Observational Study

The observational study was conducted using the ERA-Interim reanalysis and the TRMM 3b42 data sets. Long-lived, developing AEWs were identified from the NHC Tropical Cyclone Reports for each hurricane season between 2000 and 2010, with 29 confirmed cases. The results of the observational portion of the study are:
Two case studies were examined, one that remained convectively active and one that lost its convection before becoming reinvigorated in the Caribbean. These case studies suggested that the presence of strong convection within the wave may contribute to positive PV tendencies within the wave.

A lag is seen between increased convective activity and the increasing PV anomaly in the timeseries plots. By the arrow of time, this suggests that increased convection within the wave is leading to an increasing PV anomaly, therefore strengthening the wave.

Analysis is extended to all 29 cases to test the robustness of the convective versus non-convective modes. By separating the convectively active times from the convectively inactive times, we examined the RH, PV and wind shear distributions for those times. The analysis showed:

- Convectively active times were described higher RH values at both 600 hPa and 850 hPa than convectively inactive times, suggesting the ERA-Interim compares well to the TRMM data.
- Convectively active times were described by a positive PV tendency while convectively inactive times were described by a negative PV tendency.
- The result is statistically significant as determined by a t-test on the distributions, showing this is a robust result.
- The analysis of all 29 cases quantitatively supports the qualitative findings of the case studies, where convective activity is linked to a positive PV tendency within the wave.

The case study of the pre-Ernesto (2006) long-lived AEW suggests that a wave pouch may exist long before the time of tropical cyclogenesis as suggested in Dunkerton et al. (2009).

- The suggested presence of the wave pouch long before tropical cyclogenesis supports the theory in Dunkerton et al. (2009), that the wave pouch may act to help organize convection within the wave and provide a pathway for the convection to support the parent wave.
- The analysis is hampered by the lack of horizontal and temporal resolution as well as the lack of temporal consistency in the ERA-Interim, which highlights the need for a high resolution modeling study.
4.1.2 Modeling Study

The most novel part of this study is the high-resolution modeling study examined the pre-Ernesto (2006), long-lived AEW using the WRF-ARW version 3.2.1. The model was initialized on 00Z 18 August 2006 from the ERA-Interim reanalysis and was run through 00Z 27 August 2006, making for a 9-day simulation. A simulation of this length - that accurately represents an AEW and its transition to a TC - is unprecedented. We showed that the simulation closely reflected the observed Ernesto case with respect to track, intensity, time of tropical cyclogenesis and precipitation fields. The result of the simulation produced a temporally consistent, high resolution data set of a long-lived AEW that allows for detailed analysis and computation of vorticity and PV budgets.

Wave Pouch Analysis

The high-resolution simulation confirms the presence of a wave pouch that was suggested in the ERA-Interim analysis. In the modeling study reveals:

- The analysis of the high resolution simulation shows that a wave pouch develops between 500 hPa and 700 hPa just 24 hours into the simulation, about 7 days prior to TC genesis. The presence of the wave pouch indicates that the wave is providing a favored region for convection. The convection was observed within the center of the wave pouch consistent with Dunkerton et al. (2009), where the center of the pouch provides the deterministic point for eventual tropical cyclogenesis.

- The relative humidity field confirms that pouch in the lower troposphere protects the convection within the pouch from intrusions of very dry environmental air. The dry air is diverted around the outside of the pouch and the moist air within the pouch is recirculated. The maintenance of the moist pouch in the lower troposphere likely aids in rejuvenating the convection after the large scale subsidence subsides, helping to maintain the wave.

- A deep wave pouch from the mid-troposphere to the surface may not be enough to guarantee tropical cyclogenesis. We note a deep wave pouch that develops about 4 days prior to TC genesis.
  
  - TC genesis was disrupted by synoptic scale subsidence on the western half of the wave. The subsidence inhibited convection on the western side of the wave and disrupted the deep convective heating profile within the center of the wave pouch.
• The height-time analysis confirms the findings of Dunkerton et al. (2009), that the wave pouch is characterized as a convective heating profile dominate region, with the level of maximum diabatic heating the middle troposphere. The dominate convective heating profile favors convergence in the middle and lower troposphere, maximized in the boundary layer, and divergence in the upper troposphere.

**PV and Vorticity Budget Analysis**

The PV budget showed the divergence term to be the dominate positive PV tendency term throughout the depth of the troposphere throughout the simulation. The combination of the vertical advection term and the diabatic term generated a net negative PV tendency throughout the troposphere. The PV budget shows that the wave is maintained through the convergence and stretching of vorticity that is present in the wave pouch due to the convective heating profile.

The vorticity budget further supports the findings of the PV budget, in showing that the convergence and stretching terms are the dominate terms in the middle and lower troposphere. Also in the wave stage, the vertical advection of vorticity generated in the lower troposphere contributes to the positive vorticity tendency in the middle troposphere, helping the support the wave level vorticity.

The combination of these budgets shows the presence of convection within the wave is critical to the maintenance or strengthening of the wave. When the convection becomes suppressed, the divergence term in the PV budget becomes significantly weakened and can no longer offset the negative tendencies of the vertical advection / diabatic term and the boundary term. If the convection is suppressed, the convergence and stretching term in the vorticity budget no longer contributes to the maintenance of the AEW and a net weakening of the wave is observed when the convection is suppressed.

The results from the observational study help to extend the results of the modeling study to all long-lived AEWs. This shows that on average, over all long-lived cases the most important mechanism to maintaining the wave is through convection within the wave. The modeling study helps to explain the pathway through which the convection acts to strengthen the mid-level PV associated with the wave.
4.1.3 Final Thoughts

The overall result of the study shows that convection within the wave is key for maintaining or strengthening the vorticity and PV associated with the wave.

- The convective heating profile supports convergence in the lower troposphere that is critical to converging vorticity and the deep convective updrafts are key to stretching that vorticity. This describes a bottom-up process for both wave maintenance and for a transition to a tropical cyclone.

- The key to sustaining an AEW as it crosses the tropical Atlantic - without it developing into a tropical cyclone or dissipating - is a balance between processes that support convection and processes that suppress convection.

  - This also suggests that there is more than one pathway for an AEW to achieve a long-lived status in crossing the tropical Atlantic. In the case of the pre-Ernesto (2006) wave, convection remained active throughout the life span of the wave, thus allowing the wave to stay strong. However, interaction with synoptic scale subsidence and vertical shear disrupted the convective heating profile of the wave pouch enough to prevent the wave from developing into a tropical cyclone earlier.

To better understand these long-lived variety of AEWs it would be important to perform sensitivity tests. Since the presence of convection has been deemed important to maintaining the AEW, future studies could examine the robustness of the wave by experimentally eliminating the convection to examine how long it would take the wave to dissipate. It would also be important to examine waves that don’t make it to 60° to examine what environmental conditions are enough to fully dissipate an AEW.
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