ABSTRACT

MARCIANO, CHRISTOPHER GERALD. Climate Change and Wintertime East Coast Cyclones. (Under the direction of Dr. Gary Lackmann).

Previous studies investigating the impacts of climate change on extratropical cyclones have primarily focused on changes in the frequency, intensity and distribution of these events. Fewer studies have directly investigated changes in the storm scale dynamics of individual cyclones. Precipitation associated with these events is projected to increase with warming due to increased atmospheric water vapor content. This presents the potential for enhancement of cyclone intensity through lower-tropospheric diabatic potential vorticity (PV) generation. Here it is hypothesized that changes in future extratropical cyclone intensity are partly attributable to enhanced condensational heating.

The Weather Research and Forecasting model is used to simulate individual wintertime Miller-A and Miller-B cyclone events along the Eastern United States in present and future thermodynamic environments. Present and future cyclone composites are generated from these simulations in order to identify common changes among these events. A trend toward stronger storms with lower central pressures is found. Surface cyclones also exhibit a more eastward track with warming. Future cyclones exhibit increases in total precipitation and decreases in snowfall overall. Eastward track shifts are attributable to increases in upper-tropospheric westerly winds. Increases in intensity are attributed to changes in both the maximum Eady growth rate and storm scale Ertel PV. Limited statistical
significance is found in this study which may be due to small sample size. Sensitivity studies indicate that the model configuration used here may produce overly conservative results.
DEDICATION

This work is dedicated to my brother, Dylan, who has been a source of inspiration through much of the research and writing process.
BIOGRAPHY

Chris was born in Atlantic City, NJ and grew up along the Jersey Shore in the nearby town of Northfield, NJ. Some of Chris’s earliest memories of the weather include two high impact weather events in 1996. The first of those events was the Blizzard of ’96 which dumped over a foot of snow in his hometown and canceled school for a week. The second, and more interesting, event was Hurricane Bertha which his family inadvertently camped through in September 1996. With those two events firmly engraved in his memory Chris became a weather enthusiast and decided to pursue his undergraduate studies in meteorology. He graduated Summa Cum Laude from Rutgers University in 2011 with his Bachelor of Science in Meteorology. Chris enrolled in the Atmospheric Science graduate program at North Carolina State University in August 2011 where he has been working under the guidance of Dr. Gary Lackmann. Upon finishing his Master’s degree, Chris will be pursuing his PhD at North Carolina State University.
ACKNOWLEDGMENTS

First, I would like to thank my advisor, Dr. Gary Lackmann, for giving me the opportunity to work on this project. He has been a great mentor and friend over the past two and a half years and has provided an immeasurable amount of guidance and support. I would also like to thank my thesis committee members, Dr. Walter Robinson and Dr. Anantha Aiyyer, for their valuable suggestions and insight which greatly contributed to the completion of this work.

I thank the U.S. National Science Foundation (NSF) for supporting this research through Grant #AGS-1007606. I would also like to express my gratitude to the National Center for Atmospheric Research (NCAR) for making the Weather Research and Forecasting (WRF) model available to the public as WRF was used extensively in this work. The National Climatic Data Center (NCDC) also deserves thanks for use of the North American Regional Reanalysis (NARR) data. I thank the Program for Climate Model Diagnosis and Intercomparison (PCMDI) for making the CMIP3 model output available as well.

I would be remiss if I didn’t acknowledge my labmates, Jordan Dale, Allison Michaelis, Jeff Willison, Tiffany Gardner, Michelle Cipullo, Michael Graves, Xiayou Long, Jennifer Tate, and Nate Farrington, who have been a source of great friendship and advice over the years. I would also like to thank my fellow graduate students and friends within the department, particularly the MEAS soccer contingent.

Finally, I would like to thank my family and friends who have been there for me every step of the way. Specifically, I want to thank my aunt and uncle, Denise and John DuBois, for making me feel at home when I first moved to North Carolina. I also want to
thank my mother, Joanne, my father, Jerry, and my brother, Dylan, who have all provided me with unwavering love and support over the years. Your encouragement to follow my dreams has driven me to great success and without you I couldn’t possibly be where I am today.
# TABLE OF CONTENTS

LIST OF TABLES ........................................................................................................................................ viii

LIST OF FIGURES ......................................................................................................................................... ix

1. Introduction ............................................................................................................................................... 1
   1.1 Motivation ................................................................................................................................................ 1
   1.2 Literature Review .................................................................................................................................. 3
      1.2.1 Categorization of Extratropical Cyclones ...................................................................................... 3
      1.2.2 Extratropical Cyclogenesis in the Potential Vorticity Framework .................................................. 6
      1.2.3 Future Climate Projections ............................................................................................................ 9
         1.2.3.1 Projected Precipitation Changes and Impacts on Extratropical Cyclones ............................... 9
         1.2.3.2 Projected Changes in North Atlantic Extratropical Cyclones ............................................... 13
      1.2.4 Cyclone Compositing: Benefits and Limitations ........................................................................... 16
   1.3 Science Questions .................................................................................................................................. 19

2. Methods .................................................................................................................................................... 24
   2.1 Categorization of Coastal Extratropical Cyclone Events ..................................................................... 24
   2.2 Configuration of the Weather Research and Forecasting Model ......................................................... 26
   2.3 Cyclone Compositing .......................................................................................................................... 28
      2.3.1 WRF Composites .......................................................................................................................... 29
      2.3.2 NARR Composites ......................................................................................................................... 30
   2.4 Future Thermodynamic Changes, Composites and Statistical Significance ........................................ 31
   2.5 Potential Vorticity Analysis .................................................................................................................. 35
3. WRF Composite Analysis ............................................................................................................43

3.1 Assessment of Present WRF Simulations ..............................................................................43

3.2 Synoptic Overview of WRF Composites ..............................................................................44

3.2.1 Synoptic Overview of Miller-A WRF Composites ..........................................................45

3.2.2 Synoptic Overview of Miller-B WRF Composites ..........................................................46

3.3 Analysis of Present to Future WRF Composite Changes .....................................................47

3.3.1 Changes in Intensity and Track .......................................................................................48

3.3.2 Changes in Precipitation: Total Precipitation and Snowfall ..........................................52

3.3.3 Changes in Lower and Upper-Level Storm Dynamics ......................................................56

3.3.4 Eady Growth Rate and Potential Vorticity Analysis .......................................................65

4. Sensitivity Studies and Additional Experiments ......................................................................125

4.1 Physics Ensemble Study .......................................................................................................126

4.2 GCM Ensemble Study .........................................................................................................128

4.3 Sensitivity to Domain Size ...................................................................................................130

4.4 Sensitivity to Resolution .......................................................................................................132

4.5 Sensitivity to Compositing Technique ...............................................................................133

5. Summary and Conclusions ....................................................................................................154

REFERENCES ................................................................................................................................161
LIST OF TABLES

Table 2.1: Starting and ending times for each case within the Miller-A and Miller-B composites................................................................................................................................38

Table 3.1: Root mean square (RMS) error is presented for each Miller-A and Miller-B simulation in terms of minimum sea-level pressure (MinSLP) and track. RMS error for minimum sea-level pressure and track are expressed in units of millibars (mb) and kilometers (km), respectively ....................................................................................................................72

Table 3.2: Factors that lead to stronger and weaker Miller-A and Miller-B cyclones in the future........................................................................................................................................73

Table 4.1: Configurations for each of the six physics ensemble members: EMB5, EMB6, MYK5, MYK6, QQK5 and QQK6. The EMB5 configuration uses the ETA similarity surface layer scheme, Mellor-Yamada-Janic PBL scheme, Betts-Miller-Janjic CP scheme and WRF Single-Moment 5-Class (WSM5) microphysics scheme. The MYK5 configuration uses the MM5 surface layer scheme, Yonsei University (YSU) PBL scheme, Kain-Fritsch (KF) CP scheme and WSM5 microphysics scheme. The EMB5 configuration uses the Quasi-Normal Scale Elimination (QNSE) surface layer scheme, QNSE PBL scheme, KF CP scheme and WSM5 microphysics scheme. The EMB5 configuration uses the Quasi-Normal Scale Elimination (QNSE) surface layer scheme, QNSE PBL scheme, KF CP scheme and WSM5 microphysics scheme. The EMB6/MYK6/QQK6 configurations are the same as the EMB5/MYK5/QQK5 configurations except that the WRF Single-Moment 6-Class (WSM6) microphysics scheme is used instead of WSM5. ...................................................................136
LIST OF FIGURES

Figure 1.1: Taken from Plant et al. (2003). (a) The surface cyclonic circulation induced by a positive upper-level potential vorticity anomaly during the early stages of an extratropical cyclone is shown. (b) The impacts of latent heating on the surface circulation and upper-level PV anomaly are shown. Negative diabatic PV tendencies aloft erode the upper-level positive PV anomaly while positive diabatic PV tendencies at low-levels enhance the surface circulation. .................................................................21

Figure 1.2: Taken from O’Handley and Bosart (1996). The process by which a surface cyclone track changes as it encounters the Appalachian Mountains is shown. The thick solid lines with arrows correspond to the circulation associated with the upper level cyclone which tracks to the east as shown by the dashed arrows. As the upper level cyclone moves from location 1 to location 2, vortex stretching to the north causes the surface cyclone to deflect northward (dot-dashed line). Similarly, as the upper level cyclone moves from location 2 to location 3 vortex stretching results in the redevelopment of the surface cyclone to the south of the Appalachians. The upper level cyclone and surface cyclone then come back into phase as they move away from the mountains.................................................................22

Figure 1.3: Taken from Lackmann (2011) and adapted from Hoskins (1990). The mutual amplification of upper-level and low-level positive potential vorticity anomalies is shown. A cold upper-level trough and warm low-level trough is represented with isentrope contours (green and orange) at each level. The positive potential vorticity anomaly associated with the upper-level and low-level troughs are represented as a blue cross and red cross, respectively. The cross thermal advection from each anomaly onto the other are shown via dashed arrows .................................................................................................................23

Figure 2.1: The domains for the North American Regional Reanalysis (NARR) dataset and WRF model simulations are shown in purple and pink, respectively ......................................................39

Figure 2.2: Flow charts showing how the (a) WRF cyclone composites and (b) pseudo-idealized WRF simulations are generated. To generate the WRF cyclone composite, NARR data for each cyclone event is first fed into the WRF pre-processing software (WPS) which prepares the data to be read by WRF. The pre-processed data is then used by WRF to simulate individual cyclone events. Fields of interest from each of these simulated events are then averaged together to generate a WRF cyclone composite. To generate the pseudo-idealized WRF simulation, fields of interest for each individual cyclone event are averaged together using NARR data first. This generates a NARR composite. This NARR composite cyclone is then fed into WPS to pre-process the data. The pre-processed NARR composite data is then used by WRF to generate a pseudo-idealized simulation......................................................40

Figure 2.3: Flow charts showing how the (a) future WRF cyclone composites and (b) future pseudo-idealized WRF simulations are generated. The process by the future WRF cyclone
composite is generated is exactly the same as that shown in Figure 2.2 except with one additional step. Prior to feeding the NARR data for each cyclone event into WPS, thermodynamic changes are applied. A similar step is also added for the generation of future pseudo-idealized WRF simulations as well. Prior to feeding the NARR cyclone composite into WPS, thermodynamic changes are also applied ...............................................................41

Figure 2.4: Future minus present change in 2-m (a) temperature (C) and (b) temperature gradient (C/10^5 m) along eastern North America at model initialization (i.e. hour 00) ........42

Figure 3.1: 500-mb (a,b) and surface (c,d) analysis of the present (a,c) and future (b,d) Miller-A WRF composite at hour 00. Geopotential height (dam) [contours] and absolute vorticity (10^{-5} s^{-1}) [color fill] are plotted at the 500-mb level. Mean sea-level pressure (mb) and simulated radar reflectivity (dBZ) are plotted at the surface ............................................74

Figure 3.2: As in Figure 3.1 except for hour 24.......................................................................75

Figure 3.3: As in Figure 3.1 except for hour 48.......................................................................76

Figure 3.4: As in Figure 3.1 except for hour 72.......................................................................77

Figure 3.5: 500-mb (a,b) and surface (c,d) analysis of the present (a,c) and future (b,d) Miller-B WRF composite at hour 00. Geopotential height (dam) [contours] and absolute vorticity (10^{-5} s^{-1}) [color fill] are plotted at the 500-mb level. Mean sea-level pressure (mb) and simulated radar reflectivity (dBZ) are plotted at the surface ............................................78

Figure 3.6: As in Figure 3.5 except for hour 24.......................................................................79

Figure 3.7: As in Figure 3.5 except for hour 48.......................................................................80

Figure 3.8: As in Figure 3.5 except for hour 72.......................................................................81

Figure 3.9: Location of the surface cyclones, determined by minimum sea-level pressure, that make up each composite [red X]. The (a) present Miller-A and (b) future Miller-A simulations are analyzed up to hour 63. The (c) present Miller-B and (d) future Miller-B simulations are analyzed up to hour 69.................................................................82

Figure 3.10: Average minimum sea-level pressure (mb) evolution of the present [blue] and future [red] Miller-A simulations. Error bars denote the one standard deviation range........83

Figure 3.11: Minimum sea-level pressure difference (mb) between the present and future Miller-A simulations. Difference is computed as the future minimum sea-level pressure minus the present minimum sea-level pressure. The average difference is plotted as the thick
black line with error bars denoting the one standard deviation range. The difference for each case [see above legend for colors] is also plotted for context.................................84

Figure 3.12: Average minimum sea-level pressure (mb) evolution of the present [blue] and future [red] Miller-B simulations. Error bars denote the one standard deviation range ........85

Figure 3.13: Minimum sea-level pressure difference (mb) between the present and future Miller-B simulations. Difference is computed as the future minimum sea-level pressure minus the present minimum sea-level pressure. The average difference is plotted as the thick black line with error bars denoting the one standard deviation range. The difference for each case [see above legend for colors] is also plotted for context..................................................86

Figure 3.14: Track of the average (a) present [solid blue] and (b) future [solid red] Miller-A surface cyclones for hours 18-63. The one standard deviation (i.e. 1-sigma) cone for the average present and average future track is shown with blue dashed and red dashed lines, respectively ..............................................................................................................................87

Figure 3.15: Tracks of the individual (a) present and (b) future Miller-A surface cyclones for each case. Each track is shown for hours 18-63 of its respective simulation. The track for each case is labeled with a different color as shown in the legends above..............................88

Figure 3.16: Track of the average (a) present [solid blue] and (b) future [solid red] Miller-B surface cyclones for hours 18-69. The one standard deviation (i.e. 1-sigma) cone for the average present and average future track is shown with blue dashed and red dashed lines, respectively ..............................................................................................................................89

Figure 3.17: Tracks of the individual (a) present and (b) future Miller-B surface cyclones for each case. Each track is shown for hours 18-69 of its respective simulation. The track for each case is labeled with a different color as shown in the legends above..............................90

Figure 3.18: Composite total precipitation for Miller-A and Miller-B cyclones. The (a) present and (b) future Miller-A composite total precipitations are for hours 00 through 63. The (c) present and (d) future Miller-B composite total precipitations are for hours 00 through 69. Precipitation is plotted in units of millimeters (mm) ...........................................91

Figure 3.19: Composite total precipitation change for Miller-A and Miller-B cyclones. The (a) Miller-A composite total precipitation change is for hours 00 through 63. The (b) Miller-B composite total precipitation change is for hours 00 through 69. Precipitation is plotted in units of millimeters (mm) .................................................................92

Figure 3.20: Composite (a,c) total precipitation change [contours], (b,d) snowfall change [contours] and statistical significance [color fill] for Miller-A and Miller-B cyclones. The (a,b) Miller-A composite changes are for hours 00 through 63. The (c,d) Miller-B composite
changes are for hours 00 through 69. Precipitation is plotted in units of millimeters (mm). A 20-point Gaussian weighted smoothing function is applied to the fields prior to computing the difference and statistical significance. Statistical significance is plotted for the 90, 95 and 99% confidence levels using a 2-sample, 2-tailed Student’s t-test for equal sample sizes and variances .........................................................93

Figure 3.21: Histograms of total precipitation (mm) values greater than zero over all Miller-A and Miller-B simulations. The (a) present, (b) future and (c) difference histograms for Miller-A simulations are over 63 hours. The (d) present, (e) future and (f) difference histograms for Miller-B simulations are over 69 hours. Precipitation bins are every 1 mm .........................94

Figure 3.22: Histograms of composite reflectivity (dBZ) values greater than zero over all Miller-A and Miller-B simulations. The (a) present, (b) future and (c) difference histograms for Miller-A simulations are over 63 hours. The (d) present, (e) future and (f) difference histograms for Miller-B simulations are over 69 hours. Composite reflectivity bins are every 1 dBZ .................................................................95

Figure 3.23: Composite snowfall for Miller-A and Miller-B cyclones. The (a) present and (b) future Miller-A composite snowfall totals are for hours 00 through 63. The (c) present and (d) future Miller-B composite snowfall totals are for hours 00 through 69. Snowfall is plotted in units of kilograms per meter squared (km/m²) .................................................................96

Figure 3.24: Composite snowfall change for Miller-A and Miller-B cyclones. The (a) Miller-A composite snowfall change is for hours 00 through 63. The (b) Miller-B composite snowfall change is for hours 00 through 69. Snowfall is plotted in units of kilograms per meter squared (km/m²) .................................................................97

Figure 3.25: Histograms of total snowfall (kg/m²) values greater than zero over all Miller-A and Miller-B simulations. The (a) present, (b) future and (c) difference histograms for Miller-A simulations are over 63 hours. The (d) present, (e) future and (f) difference histograms for Miller-B simulations are over 69 hours. Snowfall bins are every 1 kg/m² .........................98

Figure 3.26: Composite 10-m wind speed (m/s) [fill] and wind barbs at hour 54 for (a) present Miller-A, (b) future Miller-A, (c) present Miller-B, and (d) future Miller-B cyclones ........................................................................................................99

Figure 3.27: Maximum 10-m wind speed (m/s) at each grid point over all times for all (a) present Miller-A, (b) future Miller-A, (c) present Miller-B, and (d) future Miller-B simulations .................................................................100

Figure 3.28: Change in maximum 10-m wind speed (m/s) at each grid point over all times for all (a) Miller-A and (b) Miller-B simulations .................................................................................101
Figure 3.29: Histograms of 10-m wind speed (m/s) exceeding 0.0 m/s within the area bound from 25° to 49°N and 90° to 45°W over all Miller-A and Miller-B simulations. Data for the (a) present, (b) future and (c) difference histograms of Miller-A simulations are output from 00 to 63 hours. Data for the (d) present, (e) future and (f) difference histograms of Miller-B simulations are output from 00 to 69 hours. Wind speed bins are every 1 m/s.

Figure 3.30: Histogram difference (future minus present) of 10-m wind speed (m/s) exceeding 0.0 m/s over land within the area bound from 25° to 49°N and 90° to 45°W over all times for all (a) Miller-A and (b) Miller-B simulations. Miller-A values are output from hours 00 to 63. Miller-B values are output from hours 00 to 69.

Figure 3.31: Histogram difference (future minus present) of 10-m wind speed (m/s) exceeding 0.0 m/s over water within the area bound from 25° to 49°N and 90° to 45°W over all times for all (a) Miller-A and (b) Miller-B simulations. Miller-A values are output from hours 00 to 63. Miller-B values are output from hours 00 to 69.

Figure 3.32: Composite 850-mb wind speed exceeding 25 m/s [contour] (m/s) and specific humidity [fill] (kg water vapor/kg moist air) at hour 57 for (a) present Miller-A, (b) future Miller-A, (c) present Miller-B, and (d) future Miller-B cyclones.

Figure 3.33: Composite change in meridional wind component [contours] (m/s) and statistical significance [color fill] at hour 57 for (a) Miller-A and (c) Miller-B cyclones. Wind vectors for present [blue] and future [red] (b) Miller-A and (d) Miller-B composites are also shown. A 20-point Gaussian weighted smoothing function is applied to the fields prior to computing the difference and statistical significance. Statistical significance is plotted for the 90, 95 and 99% confidence levels using a 2-sample, 2-tailed Student’s t-test for equal sample sizes and variances.

Figure 3.34: Histograms of 850-mb wind speed (m/s) greater than or equal to 25 m/s within the area bound from 25° to 49°N and 90° to 45°W for all Miller-A and Miller-B simulations. Data for the (a) present, (b) future and (c) difference histograms of Miller-A simulations are output from 00 to 63 hours. Data for the (d) present, (e) future and (f) difference histograms of Miller-B simulations are output from 00-69 hours. Wind speed bins are every 1 m/s.

Figure 3.35: Vertical cross section of temperature change (K) and change in meridional temperature gradient (10⁻⁶ K/m) at hour 48 for the (a, b) Miller-A and (c, d) Miller-B composites. Cross section is computed from along the 90°W meridian from 25° to 45°N.

Figure 3.36: Change in the 250-mb composite zonal wind component [fill] (m/s) at hour 48 for the (a) Miller-A, and (b) Miller-B composite. 250-mb geopotential height (dam) contours are also plotted for context.
Figure 3.37: Composite change in 250-mb zonal wind component [contours] (m/s) and statistical significance [color fill] at hour 48 for (a) Miller-A and (b) Miller-B cyclones. A 20-point Gaussian weighted smoothing function is applied to the fields prior to computing the difference and statistical significance. Statistical significance is plotted for the 90, 95 and 99% confidence levels using a 2-sample, 2-tailed Student’s t-test for equal sample sizes and variances ................................................................................................................................110

Figure 3.38: Histograms of 250-mb wind speed (m/s) within the area bound from 20° to 49°N and 120° to 45°W for all Miller-A and Miller-B simulations. Data for the (a) present, (b) future and (c) difference histograms of Miller-A simulations are output from 00 to 63 hours. Data for the (d) present, (e) future and (f) difference histograms of Miller-B simulations are output from 00 to 69 hours. Wind speed bins are every 1 m/s.....................111

Figure 3.39: Change in composite 250-mb wind speed (m/s) [contour] and divergence (10^{-6} \text{ s}^{-1}) [fill] at hour 48 for (a) present Miller-A, (b) future Miller-A, (c) present Miller-B, and (d) future Miller-B cyclones ........................................................................................................112

Figure 3.40: Composite pressure (mb) [fill] and wind (kts) [barbs] on the 1.5 PVU (1 PVU = 10^{-6} \text{ m}^2 \text{ s}^{-1} \text{ kg}^{-1} \text{K}) surface at hour 48 for (a) present Miller-A, (b) future Miller-A, (c) present Miller-B, and (d) future Miller-B cyclones........................................................................................................113

Figure 3.41: Change in composite pressure (mb) on the 1.5 PVU (1 PVU = 10^{-6} \text{ m}^2 \text{ s}^{-1} \text{ kg}^{-1} \text{K}) surface at hour 48 for (a) Miller-A and (b) Miller-B cyclones..............................................114

Figure 3.42: Composite pressure (mb) [fill] and wind (kts) [barbs] on the 1.5 PVU (1 PVU = 10^{-6} \text{ m}^2 \text{ s}^{-1} \text{ kg}^{-1} \text{K}) surface at hour 57 for (a) present Miller-A, (b) future Miller-A, (c) present Miller-B, and (d) future Miller-B cyclones........................................................................................................115

Figure 3.43: Time-averaged mean Ertel PV anomaly profiles, expressed in units of PVUs (1 PVU = 10^{-6} \text{ m}^2 \text{ s}^{-1} \text{ kg}^{-1} \text{K}), for Miller-A cyclones. Time-averaged profiles are computed for (a) hours 18-27, (b) hours 30-39, (c) hours 42-51, and (d) hours 54-63. Present (future) profiles are represented by blue (red) lines ...............................................................................................................................116

Figure 3.44: Time-averaged mean Ertel PV anomaly profiles, expressed in units of PVUs (1 PVU = 10^{-6} \text{ m}^2 \text{ s}^{-1} \text{ kg}^{-1} \text{K}), for Miller-B cyclones. Time-averaged profiles are computed for (a) hours 18-24, (b) hours 27-33, (c) hours 36-42, (d) hours 45-51, (e) hours 54-60, and (f) hours 63-69. Present (future) profiles are represented by blue (red) lines ......................................117

Figure 3.45: Time-averaged mean latent heating rate profiles, expressed in units of K/day, for Miller-A cyclones. Time-averaged profiles are computed for (a) hours 18-27, (b) hours 30-39, (c) hours 42-51, and (d) hours 54-63. Present (future) profiles are represented by blue (red) lines .............................................................................................................118
Figure 3.46: Time-averaged mean latent heating rate profiles, expressed in units of K/day, for Miller-B cyclones. Time-averaged profiles are computed for (a) hours 18-24, (b) hours 27-33, (c) hours 36-42, (d) hours 45-51, (e) hours 54-60, and (f) hours 63-69. Present (future) profiles are represented by blue (red) lines.

Figure 3.47: Time-averaged mean non-advective PV tendency profiles, expressed in units of PVU/hr, for Miller-A cyclones. Time-averaged profiles are computed for (a) hours 18-27, (b) hours 30-39, (c) hours 42-51, and (d) hours 54-63. Present (future) profiles are represented by blue (red) lines.

Figure 3.48: Time-averaged mean non-advective PV tendency profiles, expressed in units of PVU/hr, for Miller-B cyclones. Time-averaged profiles are computed for (a) hours 18-24, (b) hours 27-33, (c) hours 36-42, (d) hours 45-51, (e) hours 54-60, and (f) hours 63-69. Present (future) profiles are represented by blue (red) lines.

Figure 3.49: Time-averaged mean 2-m potential temperature anomalies, expressed in units of K, for Miller-A cyclones. Anomalies are computed for (a) hours 18-27, (b) hours 30-39, (c) hours 42-51, and (d) hours 54-63. Present (future) profiles are represented by blue (red) lines.

Figure 3.50: Time-averaged mean 2-m potential temperature anomalies, expressed in units of K, for Miller-B cyclones. Anomalies are computed for (a) hours 18-24, (b) hours 27-33, (c) hours 36-42, (d) hours 45-51, (e) hours 54-60, and (f) hours 63-69. Present (future) profiles are represented by blue (red) lines.

Figure 3.51: Change in time-averaged mean 2-m potential temperature anomalies, expressed in units of K, for (a) Miller-A and (b) Miller-B cyclones. Here the change is calculated as the future minus the current anomaly. Changes in the Miller-A anomalies are computed for hours 18-27, 30-39, 42-51 and 54-63. Changes in the Miller-B anomalies are computed for hours 18-24, 27-33, 36-42, 45-51, 54-60, and 63-69.

Figure 4.1: Minimum sea-level pressure (mb) evolution of each (a) present and (b) future physics ensemble member for the January 3-6, 1994 event. Line colors designating each ensemble member are shown in the legends above.

Figure 4.2: Average minimum sea-level pressure (mb) evolution of the present [blue] and future [red] physics ensemble members for the January 3-6, 1994 event. Error bars denote the one standard deviation range.

Figure 4.3: Surface cyclone track of each (a) present and (b) future physics ensemble member for the January 3-6, 1994 event. Line colors designating each ensemble member are shown in the legends above.
Figure 4.4: Ensemble mean surface cyclone track of the (a) present and (b) future physics ensemble for the January 3-6, 1994 event. The one standard deviation (i.e. 1-sigma) cone for the present and future ensemble mean track is shown with blue dashed and red dashed lines, respectively ............................................................................................................................140

Figure 4.5: Minimum sea-level pressure (mb) evolution of the (a) GCM ensemble mean, and (b) individual GCM ensemble members for the January 3-6, 1994 event. Error bars in (a) denote the one standard deviation range. Line colors designating each ensemble member in (b) are shown in the legend above .........................................................................................141

Figure 4.6: Minimum sea-level pressure (mb) evolution of the GCM ensemble mean [black line], physics ensemble mean [green line] and control (EMB5) simulation [pink line] for the January 3-6, 1994 event. Error bars in denote the one standard deviation range for each ensemble ................................................................................................................................142

Figure 4.7: Surface cyclone track of each GCM ensemble member for the January 3-6, 1994 event. Line colors designating each ensemble member are shown in the legend above ......143

Figure 4.8: Ensemble mean surface cyclone track of the GCM ensemble for the January 3-6, 1994 event. The one standard deviation (i.e. 1-sigma) cone for ensemble mean track is shown with red dashed lines ................................................................................................................................144

Figure 4.9: Domain for simulations using Climate Forecast System Reanalysis (CFSR) data as initial and lateral boundary conditions ..........................................................................................................................145

Figure 4.10: Minimum sea-level pressure (mb) evolutions for present [blue lines] and future [red lines] simulations of the January 3-6, 1994 event. Evolutions are shown for simulations initialized using the CFSR [solid lines] and NARR [dashed lines] .........................................................146

Figure 4.11: Minimum sea-level pressure difference (mb) between the present and future (a) CFSR and (b) NARR simulations of the January 3-6, 1994 event. Difference is computed as the future minimum sea-level pressure minus the present minimum sea-level pressure.......147

Figure 4.12: Surface cyclone tracks for the (a) present and (b) future CFSR [black] and NARR [green] simulations of the January 3-6, 1994 event.................................................................148

Figure 4.13: Domains used for the horizontal resolution sensitivity study. The outermost domain has a horizontal grid spacing of 36 kilometers. The first inner nest (second largest domain) has a horizontal grid spacing of 12 kilometers while the innermost nest has a horizontal grid spacing of 4 kilometers ..........................................................................................................................149
Figure 4.14: Minimum sea-level pressure (mb) evolutions for present [blue line] and future [red line] simulations of the January 3-6, 1994 event. Evolutions are shown for simulations using (a) 36, (b) 12 and (c) 4 kilometer horizontal grid spacing ...........................................150

Figure 4.15: Minimum sea-level pressure difference (mb) between the present and future simulations of the January 3-6, 1994 event using (a) 36, (b) 12 and (c) 4 kilometer horizontal grid spacing. Difference is computed as the future minimum sea-level pressure minus the present minimum sea-level pressure ..........................................................................................................................151

Figure 4.16: Minimum sea-level pressure (mb) evolutions for present [blue line] and future [red line] (a) Miller-A and (b) Miller-B simulations. Evolutions are shown for simulations initialized using NARR composites ..................................................................................................................................................152

Figure 4.17: Surface cyclone tracks for present [blue line] and future [red line] (a) Miller-A and (b) Miller-B simulations. Evolutions are shown for simulations initialized using NARR composites ..................................................................................................................................................153
1. Introduction

1.1 Motivation

Extratropical cyclones are synoptic-scale weather systems typically consisting of an upper-level trough and a surface cyclone with warm and cold fronts. During the winter months, extratropical cyclones represent the dominant mode of extreme weather to impact the mid-latitudes. Extratropical cyclones are also important regulators of the general climate system. The atmosphere is responsible for a majority of the poleward energy transport in the Northern Hemisphere during the winter months, much of which comes through the growth of these systems (Fasullo and Trenberth 2008).

This study focuses on extratropical cyclone events that track close to or along the United States East Coast during the winter months, where winter is defined as December, January and February. Copious amounts of rain, damaging wind and coastal/inland flooding are among the threats these events pose to the tens of millions of inhabitants along the eastern seaboard. Some of these coastal extratropical cyclone events also produce heavy snowfall along the United States East Coast that can cripple local/state economies. In some instances, a one day shutdown of business and infrastructure could cost state economies between $300 and 700 million (American Highway Users Alliance and IHS Global Insight 2010).

Given their climatic and socioeconomic impacts, projecting potential climate warming impacts on extratropical cyclones is an area of very active research in the climatological and meteorological communities. Previous studies have investigated a wide range of topics including shifts in the mid-latitude storm tracks, and changes in the frequency and intensity of extratropical cyclones. Fewer studies have directly investigated the effect of
climate warming on the dynamics of individual cyclones themselves. The purpose of this study is to not only provide insight into how the synoptic evolution of extratropical cyclones may change along the United States East Coast in a warmer climate but to provide an explanation for why those changes may occur. Of particular interest is the impact that increased condensational heating in a warmer climate may have on the evolution of winter extratropical cyclone events in this region. Baroclinic instability theory, most notably the Eady Model, provides us with the means to understand how future thermodynamic changes may impact these events (Eady 1949). Likewise, the potential vorticity (PV) framework gives us a useful tool for diagnosing dynamical changes seen in present and future climate simulations.

General circulation models (GCMs) are typically limited in their horizontal and vertical grid spacing therefore the Weather Research and Forecasting (WRF) model is employed in this research to better resolve moist diabatic processes within extratropical cyclones. Likewise, WRF provides the ability to simulate individual events observed in nature. Composites of winter extratropical cyclone events along the United States East Coast in present and projected future thermodynamic environments are generated and analyzed. Two different compositing techniques are employed in this study which, along with the description of the WRF model, will be discussed further in the Methods section. The use of high resolution model simulations in tandem with composite-based analysis allows for diagnosis of changes in storm-scale dynamics common among composite members which would otherwise not be possible.
The remainder of this section is a literature review of research related to this study. A review of the various categorizations of extratropical cyclones, both globally and regionally, is first presented. A detailed overview of extratropical cyclogenesis and baroclinic instability from the potential vorticity framework is presented next. The role of moist diabatic processes in extratropical cyclone intensity is also presented within this framework. Various studies investigating the response of extratropical cyclones to climate warming are then presented. A review of cyclone compositing in past studies concludes the section.

1.2 Literature Review

1.2.1 Categorization of Extratropical Cyclones

Extratropical cyclones can be categorized based on a number of features including the manner in which they develop. Through their observational work, Petterssen and Smebye (1971) were able to categorize extratropical cyclones based on the role of upper and lower level features in the cyclogenesis process. Type-A cyclogenesis occurs when a surface cyclone develops in the absence of a pre-existing upper-level trough. Low-level thermal advection associated with the surface cyclone results in the development of an upper-level trough and intensification of both features continues until the upper trough and surface cyclone are superimposed on one another. Type-B cyclogenesis occurs when an upper-level trough moves over an area of low-level warm advection, increasing the low-level baroclinicity with time. Like Type-A cyclogenesis, mutual amplification of the upper and lower level features continues until they are superimposed on one another. It is noted by Petterssen and Smebye (1971) that Type-B cases are “rarely possible to find”. This has since
been refuted by Sanders (1986) who found that a pre-existing upper-level trough was present in each of the 48 explosively deepening cyclone events observed in the west-central North Atlantic between 1981 and 1984.

Plant et al. (2003) introduced a third type of extratropical cyclogenesis which he referred to as Type-C. Type-C cyclogenesis is similar to that of Type-B cyclogenesis in the sense that a pre-existing upper-level trough interacts with a weak surface baroclinic zone in the early stages. However, with Type-C cyclogenesis, the intensification of the surface cyclone is more dependent on diabatically generated low-level PV. Strong mid-level latent heating associated with the initial upper-level forcing results in the diabatic generation of positive PV at lower levels and negative PV at upper levels. This causes erosion of the upper positive PV anomaly (i.e. weakening of the trough) while enhancing the low-level cyclonic circulation (Figure 1.1). Type-C cyclogenesis can also lead to development of diabatic Rossby waves like those described by Parker and Thorpe (1995). Once a strong positive PV anomaly has developed at low levels, warm advection to its east can induce vertical motion and latent heating. This latent heating generates a positive low-level PV anomaly that acts to propagate the diabatic Rossby wave eastward. The role of diabatically generated PV on extratropical cyclone intensity will be discussed in further detail later in this section.

During the winter months, cyclogenesis events that specifically occur along the United States East Coast can also be classified according to the manner in which they develop (Miller 1946). In order to avoid confusion, these will be referred to as Miller-A and Miller-B events for the remainder of this study. Miller-A cyclogenesis events take place along low-level baroclinic zones located in the vicinity of the Gulf of Mexico or southeast
United States coastline. Once developed, these storms will progress northward and either hug the coastline or move out to sea depending on the synoptic pattern. Miller-B cyclogenesis events involve a much more complicated evolution. During these events, a primary cyclone interacts with the western edge of the Appalachian Mountains, weakens and subsequently redevelops to the south or southeast. As shown in Figure 1.2, the primary cyclone is initially deflected northward due to vortex stretching from the downslope flow along the cyclone's northeast flank (O’Handley and Bosart 1996). Once the upper-level cyclone has moved across the Appalachians, downslope flow to its south again leads to vortex stretching. This results in surface cyclone redevelopment to the south or southeast of the mountain chain.

In their study of forty cyclone events that interacted with the Appalachians, O’Handley and Bosart (1996) found that the exact location of the surface cyclone redevelopment during Miller-B cyclogenesis is dependent on the air mass in the lee of the mountains. When there is southerly flow east of the Appalachians, warm moist air is in place that acts to strengthen the baroclinic zone as the cold front crosses the mountains. In these instances, redevelopment was observed along or just ahead of the cold front and typically to the south of the primary low. If cold-air damming is in place east of the Appalachians, the redevelopment was observed to the southeast of the primary low. This is due to an inverted ridge east of the Appalachians and the presence of an enhanced baroclinic zone where the cold continental air mass meets the warm coastal waters.

A fundamental understanding of the different types of extratropical cyclone development is important to projecting how environmental changes may impact these systems in a warmer climate. Likewise, an understanding of how non-conservative processes
may alter extratropical cyclogenesis is also vital. The following section reviews research investigating the impacts of latent heating on extratropical cyclogenesis that also incorporates baroclinic instability theory and the PV framework. In doing so, this will provide the tools to not only diagnose what changes may occur as a result of climate warming but diagnose why those changes are occurring.

1.2.2 Extratropical Cyclogenesis in the Potential Vorticity Framework

One of the first attempts to describe the Eady Model of baroclinic instability in terms of PV was performed by Bretherton (1966) in which it was found that the necessary condition for baroclinic instability is that the meridional PV gradient must reverse signs within the interior of the domain. This must hold true in order for the volume integrated meridional PV flux to equal zero. Our modern day interpretation of extratropical cyclogenesis has expanded on this perspective by viewing this process as the mutual amplification and phase-locking of upper-level and lower-level PV (UPV and LVP) anomalies (Hoskins 1990, Lackmann 2011). Both of these processes are visually represented in Figure 1.3. When a LPV anomaly lies east of an UPV anomaly, their circulations can mutual amplify one another. The cyclonic circulation associated with the UPV extends down to the surface and acts to amplify the lower-level warm anomaly associated with positive PV. Likewise, the cyclonic circulation associated with the LPV anomaly extends up through the troposphere and can act to amplify the upper-level cold anomaly associated with positive PV. This mutual amplification is consistent with the necessary condition described in Bretherton (1966). Phase-locking, the process by which the phase speeds of the UPV and LPV
anomalies match each other, is also a result of this cross thermal advection (Lackmann 2011). If the UPV begins to catch up to the LPV, the cold advection projected onto the western edge of the UPV by the LPV will act to slow down the phase speed of the UPV. Simultaneously, warm advection projected onto the eastern edge of the LPV by the UPV will act to increase the phase speed of the LPV. The subsequent result is the “locking” of phase speeds in such a way that may ultimately lead to additional mutual amplification. Although the vertical structure of extratropical cyclones is observed in nature to change with time (i.e. not remain phase-locked), this provides a good theoretical overview of the intensification process.

Baroclinic instability itself is measured by the maximum Eady growth rate,

\[ \sigma_{\text{max}} = 0.31 \frac{f_0}{N} \frac{\partial U_0}{\partial z} \]

where \( \sigma_{\text{max}} \) is the maximum Eady growth rate, \( f_0 \) is the Coriolis parameter at some reference latitude, \( N \) is the Brunt-Väisälä frequency and \( \frac{\partial U_0}{\partial z} \) is the base-state wind shear (Eady 1949).

Through thermal wind balance, it is known that the base-state wind shear is proportional to the meridional temperature gradient. Assuming some constant reference latitude, it’s clear then that the maximum Eady growth rate is inversely proportional to the static stability and directly proportional to the meridional temperature gradient. While this solution only applies to adiabatic, frictionless flow and growth due to baroclinic instability, it does provide a good first order estimate of how favorable the environment is for the development and intensification of extratropical cyclones.

The Rossby penetration depth is another useful parameter in determining the favorability for mutual amplification and thus intensification to occur. The vertical reach a
disturbance’s circulation has within the troposphere is defined by the Rossby penetration depth,

\[ (1.2) \quad H_R = \frac{f_0}{N \sqrt{k^2 + l^2}} \]

where \( H_R \) is the Rossby penetration depth, \( f_0 \) is the Coriolis parameter at some reference latitude, \( N \) is the Brunt-Väisälä frequency, \( k \) is the zonal wavenumber and \( l \) is the meridional wavenumber. Like the maximum Eady growth rate, the Rossby penetration depth is inversely proportional to the static stability. This means that the cross thermal advection from UPV and LPV anomalies is stronger in more unstable environments. This is true for disturbances with large wavelengths as well since the Rossby penetration depth is inversely proportional to the total horizontal wavenumber. Knowledge of how some changes can impact the maximum Eady Growth rate and Rossby penetration depth is particularly important for determining how climate warming may impact extratropical cyclone intensity as will be discussed later.

The strengths of this approach to baroclinic instability are tied to the two main principles of PV: conservation and invertibility (Hoskins et al. 1985). In particular, they may be used to compute PV budgets and determine the relative contribution of conservative and non-conservative processes in extratropical cyclogenesis. Observational and modeling studies of extratropical cyclogenesis over the past several decades have shown that non-conservative processes, most notably the diabatic generation of PV by latent heating, can have varying degrees of importance in cyclone evolution. In the absence of a strong low-level warm anomaly, intense condensational heating can produce low-level diabatic PV anomalies that contribute significantly to surface cyclone intensity (Ahmadi-Givi et al.
2004). In some cases, this low-level diabatic PV anomaly is responsible for up to 40-70% of surface cyclone intensity (Davis and Emanuel 1991, Balasubramanian and Yau 1994, Stoelinga 1996). However, this is not typical of all extratropical cyclones. Several events have also been studied in which low-level diabatic PV only has a secondary contribution to cyclone intensity (Huo et al. 1999, Davis 1992, Davis et al. 1993). In one instance, latent heat release is even shown to accelerate the onset of cyclolysis through the decoupling of the UPV and LPV anomalies (Davis 1992).

In general, the impacts of diabatic PV on extratropical cyclogenesis appear to vary. In some cases the generation of low-level diabatic PV augments the interaction of the UPV and LPV anomalies while it can actually inhibit this interaction in others. With this in mind, it is of interest how changes in stability, the meridional temperature gradient and latent heating affect this interaction and thus extratropical cyclone structure/intensity in a warmer climate.

The following section reviews work investigating projected climate warming changes on precipitation and extratropical cyclone intensity. Projected changes associated with moisture and precipitation and their implications for extratropical cyclones are first addressed. Changes in extratropical cyclones from a variety of GCM and high resolution simulations are then presented.

1.2.3 Future Climate Projections

1.2.3.1 Projected Precipitation Changes and Impacts on Extratropical Cyclones

Future projections using GCMs indicate that the tropospheric moisture content will increase with warming. These projections show that relative humidity within the lower
troposphere remains roughly constant in the future, meaning that increases in atmospheric vapor pressure should scale with the Clausius-Clapeyron relation (Frei et al. 1988, Allen and Ingram 2002, Soden et al. 2005, Held and Soden 2006, Wentz et al. 2007). The Clausius-Clapeyron relation dictates that the saturation vapor pressure of water increases approximately 7% per degree Kelvin of warming,

\[
ln \frac{e_s}{6.11} = \frac{L}{R_v} \left( \frac{1}{273} - \frac{1}{T} \right)
\]

where \( e_s \) is the saturation mixing ratio of water, \( L \) is the latent heat of vaporization, \( R_v \) is the gas constant for water vapor and \( T \) is temperature. Given the projected increases in atmospheric water vapor content, it is not surprising that precipitation is also expected to increase in a warmer climate. This is particularly noteworthy for cases of extreme precipitation where studies have shown that precipitation increases may exhibit a super Clausius-Clapeyron relationship (Frei et al. 1988, Allen and Ingram 2002, Held and Soden 2006, Allan and Soden 2008). These extreme precipitation events may range in size from microscale to synoptic-scale phenomena meaning that many of these events could very well include extratropical cyclones (Trenberth 1999).

Watterson (2006) is a particularly insightful study in which two coupled atmosphere-ocean GCMs are used to investigate precipitation changes specifically associated with extratropical cyclones in a warmer climate. Considerable increases in the mean and extreme rainfall intensities were observed in future simulations which suggest that latent heating is also enhanced. However, it is acknowledged that high resolution simulations are necessary to quantify the relative contributions of such diabatic processes. Higher resolution studies
investigating extratropical cyclones in a warmer climate have since been conducted by Bengtsson et al. (2009) and Champion et al. (2011). Using the ECHAM5 GCM with 63 kilometer horizontal grid spacing, Bengtsson et al. (2009) investigated changes in extratropical cyclones due to climate warming over two 32-year periods (1959-1990 and 2069-2100) under the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4) A1B emissions scenario. Increases in mean precipitation associated with extratropical cyclones over North America and the North Atlantic during the winter months were a somewhat modest 4-14.4%; however, increases in extreme precipitation exceeded 20-30% thus clearly exhibiting a super Clausius-Clapeyron relation. Champion et al. (2011) expanded on the work Bengtsson et al. (2009) by conducting the same simulations with 40 km horizontal grid spacing. The frequency of extreme precipitation events (>2.5mm/hr) associated with extratropical cyclones were found to increase globally during the winter months with 1 in 5 year events becoming 1 in 0.5 year events. Focusing more regionally, Champion et al. (2011) found large increases in the track density of extreme precipitation events along the United States East Coast and western Atlantic basin.

Since the potential for increased precipitation, and thus increased diabatic PV generation, clearly exists it is important to determine how this may impact extratropical cyclone intensity. Recently, this question has been addressed by investigating the sensitivity of an idealized baroclinic wave to varying levels of moisture (Booth et al. 2012). Here it is found that increasing the moisture results in an increase in extratropical cyclone intensity based on several metrics including minimum sea-level pressure, surface wind speed and precipitation. However, moisture increases on the order of those projected by GCM
projections only result in small increases in intensity. Based on the discussion of baroclinic instability in the PV framework from the previous section, these increases in intensity may very well be tied to the development of stronger diabatic PV anomalies at low levels. However, this may not hold true when considering projected increases in both moisture and temperature. Using a pseudo-global warming approach, Lackmann (2013) performed high resolution simulations of the South-Central US flood of 2010 in present and future thermodynamic environments. The same approach is used in this study and will be discussed further in the Methods section. While Lackmann (2013) finds that both precipitation and low-level diabatic PV production increase with warming, the low-level circulation (more specifically the low-level jet) did not strengthen. Vertical velocities also increased with warming which suggests the possibility that parcels spend less time in the PV generation zone. This may partly explain why no change was seen in the low-level jet. However, it is noted that this may not be representative of all cases.

Although it appears that extratropical cyclones may have more diabatically generated PV in a warmer climate, it is yet unclear how exactly this may affect their intensity and/or structure. Likewise, there are other processes that may have significant impacts. The following section reviews a number of studies that investigate basin wide changes in North Atlantic extratropical cyclones including the factors that may impact future intensity and frequency.
1.2.3.2 Projected Changes in North Atlantic Extratropical Cyclones

Analysis of two 15-member coupled GCM ensembles using the IPCC AR4 20C and A1B emissions scenarios investigate some of the projected environmental changes within the Northern Hemisphere (Yin 2005). A poleward and upward shift of the North Atlantic storm track is generally observed among the ensemble members with an increase in upper tropospheric baroclinicity and decrease in lower level baroclinicity. These baroclinicity changes can be attributed to tropical upper tropospheric heating enhancing the meridional temperature gradient aloft and Arctic amplification weakening the meridional temperature gradient near the surface. Weak increases in static stability are also projected for much of the Northern Hemisphere. All of these projected changes, in addition to the possibility of increased diabatic PV generation, yield a number of competing processes that may affect future extratropical cyclone intensity. The question then becomes how do all of these competing processes impact extratropical cyclones in future climate projections?

Recent studies have projected extratropical cyclones to decrease in frequency in a warmer climate both globally and in the North Atlantic (Geng and Sugi 2003, Lambert and Fyfe 2006, Champion et al. 2011, Catto et al. 2011, Mizuta et al. 2011). While decreases in extratropical cyclone frequency are generally attributed to decreases in near-surface baroclinicity, the possible impacts of increased upper-level baroclinicity and increased latent heat release are also acknowledged. Catto et al. (2010) has shown that a GCM with horizontal grid spacing of 90 kilometers is capable of resolving prominent features of extratropical cyclones including the warm/cold conveyor belts and dry intrusion. However, Willison et al. (2013) have since shown that even this horizontal grid spacing is insufficient
to properly resolve moist diabatic processes within such systems. Changes in the frequency of intense extratropical cyclones within the Northern Hemisphere are also still unclear with some studies projecting decreases (Catto et al. 2011) while others project increases (Geng and Sugi 2003, Lambert and Fyfe 2006, Mizuta et al. 2011). The inconsistency among projections of intense extratropical cyclone frequency may be a result of differing metrics for intensity (Catto et al. 2011).

Several studies have focused more on the regional impacts of climate warming on extratropical cyclones over North America and the Western North Atlantic. Using the Coupled Climate System Model Version 3.0 (CCSM3), Teng et al. (2008) investigated future changes in winter extratropical cyclone activity over North America under the IPCC AR4 A1B scenario. Decreases in extratropical cyclone activity are projected along the United States East Coast which are attributed to a northward shift in the 300-mb jet and decreased lower tropospheric baroclinicity.

High resolution mesoscale models used in conjunction with GCM projections are yet another way to simulate extratropical cyclone events in present and future climate. Jiang and Perrie (2007) used output from the Coupled Global Climate Model (CGCM2) of the Canadian Centre for Climate Modeling and Analysis under the IPCC AR4 IS92a emissions scenario as lateral boundary conditions to drive the Canadian Mesoscale Compressible Community (MC2) model. Individual extratropical cyclone events in the northwest Atlantic during the autumn months from the present (1975-94) and future (2040-59) time periods were simulated with 20 kilometer horizontal grid spacing and 30 vertical levels. These events included tropical systems undergoing extratropical transition. No significant changes in
intensity in were found in the simulations of the future climate extratropical cyclones; however, composite analysis of the events showed some structural changes. Most notably, future extratropical cyclones are projected to become radially larger with a slightly larger but weaker wind field at low levels. Simulations of the future extratropical cyclone events also showed storms that track further north and propagate more rapidly.

A similar experiment was conducted by Perrie et al. (2010) for landfalling extratropical cyclones in the northwest Atlantic using the same GCM boundary conditions, mesoscale model and horizontal grid spacing. A 10-member initial conditions ensemble was performed using present and future cyclone composites to generate present and future ensemble means. Future landfalling extratropical cyclones are ultimately projected to be slightly weaker due to dynamic cooling by way of enhanced vertical motion near the storm center. This seems to be inconsistent with the prevailing thought that latent heating becomes more prominent with warming. An alternative approach for explaining the decreases Perrie et al. (2010) find in intensity could be to investigate the storm scale changes in PV.

Projections from some of the more recent Coupled Model Intercomparison Project 5 (CMIP5) GCM simulations have also been used to focus on changes associated with extratropical cyclones within this region. Focusing on what Colle et al. (2013) defined as the “Best 7” among a 15-member ensemble of CMIP5 GCMs using the RCP8.5 scenario, extratropical cyclone track density was found to increase 5-20% along eastern North America by the late 21st century (2069-2098). The number of intense cyclones (< 980 mb) over inland portions of the North American coastline is also projected to increase 10-40%. However, the number of intense cyclones immediately along those coastal waters is projected to remain
roughly the same. Analysis of two of the highest resolution and best performing GCMs (CCSM4 and MRI-CGCM3) found very little change in the Eady growth rate, upper-level jet strength or low-level temperature gradient along the eastern North American coast in the future. Given projected precipitation increases in this area of 5-30%, Colle et al. (2013) suggest that the projected increases in future extratropical cyclone intensity may instead be attributable to increased latent heat release.

It appears there is not yet a definitive answer to how extratropical cyclone intensity may change in a warmer climate. Similarly, there have been very few studies that actually investigate the storm scale dynamics responsible for projected changes in intensity. The use of high-resolution model simulations in conjunction with cyclone composites is one way to approach this problem. In the following section, the use of compositing as a means of extracting common denominators among storms is addressed.

1.2.4 Cyclone Compositing: Benefits and Limitations

Composites have been used to identify common features among extratropical cyclone events in observational studies for some time. Specifically, the use of compositing has shed a great deal of light on the climatology of and synoptic environments favorable for explosively deepening extratropical cyclones. Sanders and Gyakum (1980) generated composites of explosively deepening extratropical cyclones in the Northern Hemisphere and found they typically occur 400 nautical miles downstream of a mobile upper level trough. In their composite study, Lackmann et al. (1996) investigated explosively deepening extratropical cyclone events during the winter months in the western North Atlantic. By focusing on these
types of events within a specific region and time of year, it ensures less variability among the cases and thus a more accurate representation of specific features. As such, the authors were able to clearly identify distinct synoptic scale patterns in the days preceding such events.

More recently, the use of reanalysis data has allowed scientists to generate extratropical cyclone composites for specific regions over several decades rather than several years (Rudeva and Gulev 2011). Data available at each grid point and multiple levels is particularly useful as it allows for analysis of the horizontal and vertical structure of storm scale features throughout the evolution of extratropical cyclones.

Gridded data have also been used to generate extratropical cyclone composites in climate warming studies. As mentioned in the previous section, Jiang and Perrie (2007) and Perrie et al. (2010) both used cyclone composites to study the impacts of climate warming on autumn extratropical cyclones in the western North Atlantic. In particular, this allowed both studies to analyze structural changes in the horizontal wind and temperature fields. From a diagnostic perspective, Perrie et al. (2010) took it a step further by analyzing changes in vertical motion, divergence and computing a storm-relative heat budget. This type of analysis could be extended to look at various other fields of interest including PV. A composite-based PV budget would be particularly insightful as to what processes become more or less important to extratropical cyclogenesis in a warmer climate.

All of the previously mentioned studies averaged observations, gridded data or model output from individual extratropical cyclone events to generate cyclone composites; however, there is another approach to such analysis. A composite-initialized approach has also been used to produce pseudo-idealized simulations of atmospheric phenomena. In this
composite-initialized approach, variables of interest for each event are averaged together to generate a composite which is then used as initial and lateral boundary conditions to drive model simulations. The result is a pseudo-idealized simulation that retains the common aspects of the averaged events. Such composite-initialized simulations have been used recently to study changes in the frequency and intensity of hailstorms in Colorado due to climate warming (Mahoney et al. 2012). The benefits of such an approach are that only two simulations need to be done (present and future) as opposed to two simulations per event using the traditional composite approach, thereby saving time and computing memory. On the other hand, the ability to conduct statistical significance tests or other means of testing the robustness of results is lost using the composite-initialized approach. Both composite approaches are used and compared in this study and will be explained in greater detail in the Methods section.

While traditional composites allow scientists to identify common characteristics among storms and test for statistical significance, there are also several limitations to this approach. Composites of events are spatially and temporally averaged so there is inherent smearing of features as you move away from the time at the center of the composite (Lackmann et al. 1996). Differences in the track and propagation speed of each event ultimately result in a blurring of the signal. In the case of extratropical cyclones, the relative positioning of fronts within each storm may also differ and add to this spread. Nonetheless, the ability to identify common features among events of interest and test for statistical significance in one tool generally outweighs the limitations of composite smearing.
1.3 Science Questions

GCM and high-resolution modeling studies of climate warming project a number of changes in processes that may impact extratropical cyclones. These changes include a weakening of the near-surface meridional temperature gradient, small increases in static stability and increases in diabatic PV production. The maximum Eady growth rate is one way to determine the favorability of an environment for extratropical cyclone development/intensification but this does not account for non-conservative processes. With this in mind, the first question this study seeks to address is whether or not future changes in the environmental maximum Eady growth rate prior to cyclone development alone can explain projected changes in United States East Coast cyclone intensity.

If changes in extratropical cyclone intensity cannot be explained by the change in the maximum Eady growth rate alone then non-conservative processes such as enhanced diabatic PV generation may responsible. It is possible that the relative contributions of certain processes to extratropical cyclone intensification may change as well. In particular, it has been shown that low-level diabatic PV can augment the intensification process when the LPV anomaly is somewhat weak. However, enhanced vertical velocities may result in a shorter resonance time for diabatically generated PV at low-levels in some cases. If true, it could effectively negate the impact of increased diabatic PV production at low-levels. In light of this, the second question this study seeks to address is whether or not changes in United States East Coast cyclone intensity due to warming can be attributed to increased condensational heating and thus enhanced diabatic PV generation.
This study implements two different compositing techniques in order to identify and compare changes in United States East Coast extratropical cyclones due to climate warming. The traditional composite approach averages together fields of interest (i.e. sea-level pressure, temperature, wind, and exc.) from simulations of individual cyclone events. On the other hand, the composite-initialized approach first generates a cyclone composite from reanalysis data, which is then used as initial and lateral boundary conditions to drive simulations. The result is an ensemble mean of individual cyclone events and a pseudo-idealized cyclone event. For future climate simulations, the same thermodynamic changes are applied to the individual cases and the cyclone composite prior to initializing the model. The final question this study then seeks to address is whether or not the changes due to climate warming are qualitatively similar using the traditional and composite-initialized approaches.
Figure 1.1: Taken from Plant et al. (2003). (a) The surface cyclonic circulation induced by a positive upper-level potential vorticity anomaly during the early stages of an extratropical cyclone is shown. (b) The impacts of latent heating on the surface circulation and upper-level PV anomaly are shown. Negative diabatic PV tendencies aloft erode the upper-level positive PV anomaly while positive diabatic PV tendencies at low-levels enhance the surface circulation.
Figure 1.2: Taken from O’Handley and Bosart (1996). The process by which a surface cyclone track changes as it encounters the Appalachian Mountains is shown. The thick solid lines with arrows correspond to the circulation associated with the upper level cyclone which tracks to the east as shown by the dashed arrows. As the upper level cyclone moves from location 1 to location 2, vortex stretching to the north causes the surface cyclone to deflect northward (dot-dashed line). Similarly, as the upper level cyclone moves from location 2 to location 3 vortex stretching results in the redevelopment of the surface cyclone to the south of the Appalachians. The upper level cyclone and surface cyclone then come back into phase as they move away from the mountains.
Figure 1.3: Taken from Lackmann (2011) and adapted from Hoskins (1990). The mutual amplification of upper-level and low-level positive potential vorticity anomalies is shown. A cold upper-level trough and warm low-level trough is represented with isentrope contours (green and orange) at each level. The positive potential vorticity anomaly associated with the upper-level and low-level troughs are represented as a blue cross and red cross, respectively. The cross thermal advection from each anomaly onto the other are shown via dashed arrows.
2. Methods

This chapter describes the data and methods used in this study. The process by which cyclone events are categorized is described in detail first. Next, the manner in which WRF is configured for all simulations is addressed. A comprehensive description of the two compositing techniques implemented in this study follows. The application of thermodynamic changes, generation of future composites, and statistical significance testing is then discussed. A summary of the PV analysis utilized in this study concludes the chapter.

2.1 Categorization of Coastal Extratropical Cyclone Events

The North American Regional Reanalysis (NARR) is used substantially within this study to not only identify events but as initial/lateral boundary conditions for model simulations. The data used to initially identify and categorize coastal extratropical cyclone events along the United States East Coast is obtained from the Pennsylvania State University Electronic Map Wall (PSU e-Wall). The PSU e-Wall contains 3-hourly plots of the NARR dating back to January 1, 1979. The two different event types investigated along the United States East Coast for this study are those exhibiting the characteristics of Miller-A and Miller-B cyclogenesis during the winter months. Here the climatological definition of winter (December, January and February) is used.

An extratropical cyclone is categorized as a Miller-A event if it exhibits the following characteristics during its evolution. First, it must originally develop in or in the vicinity of the Gulf of Mexico. This definition comes from the observational study by Miller (1946) described in Chapter 1. For the purposes of this study, a cyclone is considered to have
developed once a closed surface low is observed using a contour interval of two millibars (mb). Second, the cyclone must track within 400 kilometers of the coastline until at least Cape May, New Jersey. Storms that track along or just off the coast are of particular interest because their proximity to the coast implies greater impacts to the major metropolitan areas within this region. Last, the cyclone must attain a minimum sea-level pressure below 995 mb within plus or minus 36 hours of its centering time ($T_0$). For Miller-A events, $T_0$ is defined as the time when the extratropical cyclone tracks closest to Cape Hatteras, North Carolina. A threshold of central surface pressure below 995 mb ensures the sampling of events that are of at least moderate dynamical strength.

The criteria for extratropical cyclones to be categorized as Miller-B events again come from Miller (1946). The cyclone must be a coastal redeveloping system in which the primary cyclone moves over the Western Appalachians resulting in the surface cyclone weakening on the windward side and reappearing on the leeward side. In order to focus on redevelopment events in the Mid-Atlantic and Southeastern United States, only cases where coastal redevelopments occur at or south of Lewes, Delaware are included. Likewise, the coastal cyclone must track within 400 kilometers of the United States East Coast until at least Cape May, New Jersey and attain a minimum sea-level pressure below 1000 mb by at least $T_0+12$ hours. These criteria are again implemented to focus on extratropical cyclones with high socioeconomic impacts and at least moderate intensity. For Miller-B events, $T_0$ is defined as the time when the coastal redeveloping surface low attains a minimum sea-level pressure less than or equal to the initial surface low.
United States East Coast extratropical cyclone events that do not meet all of the above criteria for either category are not included in this study. If a cyclone cannot unmistakably be defined as a Miller-A or Miller-B event it is also discarded to avoid introducing ambiguity in the dataset. This filtering is also applied to the simulations of individual cyclone events to remove cases for which the NARR initial conditions did not allow an adequate WRF simulation. Events meeting all criteria that fit Miller-A events in the NARR dataset but Miller-B events after being simulated, or vice versa, are also discarded. This is done to ensure the evolutions of events for each case type are consistent. Ultimately, this results in a dataset consisting of ten Miller-A and ten Miller-B extratropical cyclone events meeting all criteria since January 1, 1979. The dates and times of the events that make up each composite are shown in Table 2.1. The configuration of WRF and details of the simulations used in this study are addressed in the following section.

2.2 Configuration of the Weather Research and Forecasting Model

The Weather Research and Forecasting model (WRF) version 3.2.1 is used extensively in this research. WRF is a fully compressible, non-hydrostatic numerical weather prediction model developed by a number of the world’s leading meteorological research centers including the National Center for Atmospheric Research (NCAR), the National Oceanic and Atmospheric Administration’s (NOAA) National Center for Environmental Prediction (NCEP) and NOAA’s Earth System Research Laboratory (ESRL) among others. With the capability to simulate observed and idealized events, WRF is used quite often in both the operational forecasting and research communities. A more detailed description of
WRF including the numerical integration techniques implemented can be found in Skamarock (2008).

All simulations in this study are performed with 36-kilometer horizontal grid spacing and 41 vertical levels. Each run is initialized 36 hours prior to \( T_0 \) and simulated for 84 hours with a 60 second timestep. Output is generated every three hours. As previously mentioned the NARR is used as initial and lateral boundary conditions for all simulations contained within this study. With 32-kilometer horizontal grid spacing and 45 vertical levels, the NARR was chosen because it affords finer horizontal and vertical resolution than other reanalysis datasets (Mesinger and Coauthors 2006). Given the large number of cases that are simulated, WRF is configured to run with coarser resolution than the initial/lateral boundary conditions. This is done in order to reduce the amount of time/disk space required for each simulation.

One limitation in using the NARR is that the domain for WRF simulations is inherently limited by the NARR domain. Both the NARR domain and the domain used for WRF simulations are shown in Figure 2.1. A Lambert conformal conic map projection is used for the WRF simulations which limits how far east the model domain can extend. Because of this, the simulations are only analyzed up to a certain point in time in order to avoid possible contamination from the lateral boundaries. The specifics of this will be discussed in Section 3. The WRF domain covers the entire contiguous United States, much of Canada and Mexico, and portions of the Eastern Pacific and Western Atlantic Oceans. Such a broad domain is important because it allows for the analysis of the synoptic pattern upstream
of where the United States East Coast cyclogenesis events occur in addition to the events themselves.

Each simulation contained within this study also uses the same physics and dynamics options. Longwave radiation is represented using the Rapid Radiative Transfer Model (RRTM) while shortwave radiation is represented using the Dudhia scheme. In terms of microphysics, the WRF Single-Moment 5-Class (WSM5) scheme is used. The WSM5 allows for mixed-phase processes including snow and ice. It does not contain graupel but is slightly more computationally efficient than the WRF Single-Moment 6-Class (WSM6) scheme which does include graupel. The planetary boundary layer is represented using the Mellor-Yamada-Janjic (MYJ) scheme while the Betts-Miller-Janjic (BMJ) cumulus parameterization is used to represent sub-grid scale precipitation. The sensitivity of simulations to these parameterization schemes is addressed in a mini-ensemble study of one extratropical cyclone event.

With the model configuration and physics/dynamics options addressed, the focus can now move to the two compositing techniques are used in this study. The next two sections describe both of these techniques in detail. A description of the statistical significance testing used with the traditional compositing method is also included.

2.3 Cyclone Compositing

Compositing is a useful tool for determining synoptic patterns and storm scale features common between numbers of events of interest. This study implements two different compositing techniques. Neither of these techniques uses a storm relative approach. The
sensitivity of results to each of these techniques is investigated. Two separate extratropical cyclone composites are created based on the synoptic evolution of surface cyclones along the United States East Coast. These composites are made up of extratropical cyclone events that meet the previously mentioned criteria for Miller-A and Miller-B events. The Miller-A and Miller-B composites are made up of the ten cyclone events each.

2.3.1 WRF Composites

The first compositing technique implemented in this study averages individual WRF simulations together. Fields of interest at various vertical levels are averaged together from WRF simulations of an event type (i.e. Miller-A or Miller-B) which ultimately results in an average or composite cyclone event (Figure 2.2a). Temperature, specific humidity, geopotential height and the horizontal wind components are some of the fields averaged at each pressure level. Near surface parameters including the 10-m winds, 2-m temperature/relative humidity and sea-level pressure are also averaged. Several other parameters of interest including vertical velocities, liquid precipitation and snow water equivalent precipitation are averaged as well. Once all fields of interest have been averaged together, they can be analyzed to determine what storm scale characteristics of the storms have been retained.

This approach to cyclone compositing is quite useful because it makes identifying common features among United States East Coast extratropical cyclone events fairly simple. While features unique to each storm are lost in the composite, more in depth analysis of individual cases can still be conducted since each event has been simulated. This compositing
approach also allows for the testing of statistical significance which is important in any scientific study. However, it is also computationally and temporally expensive. With this in mind, another approach is also implemented in this study to determine whether the same qualitative results can be achieved using far less time and computing memory.

2.3.2 NARR Composites

The second compositing technique is different from the first in that it requires far fewer model simulations. Rather than simulating each event with WRF, a composite is first generated using the NARR dataset. Like the first technique, fields of interest at each vertical level are averaged together to generate Miller-A and Miller-B cyclone composites. Soil moisture data is also averaged together at four different levels. Once a NARR composite has been generated, it is then used as initial/lateral boundary conditions to drive a WRF simulation. This composite-initialized approach ultimately yields a pseudo-idealized simulation of a United States East Coast extratropical cyclone event. Figure 2.2b shows the general procedure of this approach. We acknowledge that this approach may yield initial/lateral boundary conditions that are no longer a true or balanced state of the atmosphere and could potentially lead to new or unresolved forcing terms. Here we are merely interested in the viability of this approach for cyclones as it has been used for other meteorological phenomena such as convective storms (Mahoney et al. 2012) in the past.

The initial and lateral boundaries for these pseudo-idealized simulations are effectively smoothed versions of the typical synoptic environment in which these systems develop. As such, the forcing for these events is likely somewhat weaker than what is
actually observed in nature. The expectation then is that the pseudo-idealized cyclone composites will exhibit characteristics similar to those seen in the traditional composites though slightly weaker in magnitude. In particular, the differences between the present and future composites are expected to qualitatively similar between the two approaches. This composite-initialized approach may prove to be an extremely useful and efficient tool for preliminarily diagnosing changes amid a large number of events in present and future thermodynamic environments. The manner in which these future thermodynamic environments are generated is addressed in the following section. The subsequent generation of future composites and statistical significance testing is also addressed.

2.4 Future Thermodynamic Changes, Composites and Statistical Significance

Extratropical cyclone events are simulated in future thermodynamic environments using the pseudo-global warming approach highlighted in a number of previous studies (Schär et al. 1996, Frei et al. 1998, Sato et al. 2007, Hara et al. 2008). Temperature changes, taken from GCM projections, are applied to the initial and lateral boundary conditions which ultimately drive regional climate models. Relative humidity is held constant meaning that the increases in atmospheric water vapor pressure scale with the Clausius-Clapeyron relation. As Schär et al. (1996) show in their study, the benefits of this approach are that it is simple to implement and both thermodynamically and dynamically consistent. More recently, this technique has been used in conjunction with high resolution numerical weather prediction models. Lackmann (2013) used this approach with WRF to investigate changes in the feedback between latent heating and the low-level jet for the South-Central United States
flood of May 2010. Similarly, Mallard et al. (2013) studied the impacts of warming on
Atlantic Hurricanes using this approach. An ensemble of GCMs was used to generate the
thermodynamic changes in each study.

In this study, we use projections of the IPCC AR4 A2 emissions scenario from a five-
member ensemble of GCMs to generate thermodynamic changes. The five GCMs that make
up this ensemble are the Bjerknes Centre from Climate Research Bergen Climate Model
version 2 (BCCR BCM2), Centre National de Recherches Météorologiques Coupled Global
Climate Model version 3 (CNRM-CM3), Institute of Numerical Mathematics Coupled Model
version 3 (INM-CM3.0), Max Planck Institute (MPI) ECHAM5 and third climate
configuration of the Met Office Unified Model (HadCM3). These GCMs were primarily
chosen because they have better underground interpolation of pressure levels than other
Coupled Model Intercomparison Project 3 (CMIP3) GCMs. The sensitivity of results to each
GCM’s derived thermodynamic changes is investigated in a mini-ensemble study of one
extratropical cyclone event.

Monthly averages of temperature at various vertical levels are generated from the
GCM ensemble and then averaged together over the periods 1990-1999 and 2090-2099 to
represent the present and future climate, respectively. The differences in average monthly
temperature over these two periods are then used to generate the monthly changes used in
this study. December temperature changes are then calculated as

\[
\Delta T_{Dec} = T_{Dec\ 2090's} - T_{Dec\ 1990's},
\]

with January and February temperature changes calculated in a comparable manner. Since
GCMs have fairly coarse vertical resolution, these changes must be vertically interpolated in

32
order to match the number of vertical levels in the NARR dataset. The appropriate
temperature changes are then tacked onto the initial/lateral boundary conditions used in the
present WRF simulations (i.e. NARR data for individual events and the NARR composite).
Cyclone events that occur in December have the December temperature changes ($\Delta T_{Dec}$)
applied to their initial/lateral boundary conditions with similar logic used for January and
February events. The NARR composite has the January temperature changes applied to it
because the average Julian day for all events falls in that month. Changes in sea surface
temperature (SST) are handled by applying the average skin temperature change derived
from the GCM ensemble. As previously mentioned, relative humidity is held constant.

After the thermodynamic changes have been applied, the new future initial/lateral
boundary conditions are used to initialize WRF simulations using the same configuration as
before. Given that we are applying changes projected over a period of 100 years, the future
simulations may then be thought of as how individual cyclone events or a pseudo-idealized
cyclone may evolve 100 years from now. Using the same method described in Section 2.3.1,
fields of interest are averaged together to generate a cyclone composite once each individual
cyclone event has been simulated in a future thermodynamic environment. Figure 2.3 shows
the process by which the future cyclone composite and future pseudo-idealized are generated.
The composite changes in 2-m temperature and 2-m temperature gradient along eastern
North America at model initialization are presented in Figure 2.4 for reference.

It is also important to test whether or not changes in particular fields between the
present and future WRF composites are statistically significant. However, there is a great
deal of noise within each composite which may unnecessarily inflate the standard deviation
of composite fields being investigated. In order to limit the noise and focus on the signal of interest, separate composites are used to test statistical significance. These composites are the same as the traditional composites except that a 20-point Gaussian weighted smoothing function is applied to each field prior to the averaging. A 2-sample, 2-tailed Student’s t-test for equal sample sizes and variances is conducted to test for statistical significance once the smoothed composites have been generated. Variance is determined to be equal using the F-test for the equality of two variances. The 2-sample, 2-tailed Student’s t-test for equal sample sizes and variances is expressed as

$$t = \frac{\bar{X}_1 - \bar{X}_2}{\sqrt{\frac{S_{x_1}^2 + S_{x_2}^2}{2} \cdot \frac{1}{n}}}$$

(2.2)

and

$$df = 2n - 2,$$

(2.3) where $t$ is the test statistic, $\bar{X}_i$ is the sample mean, $S_{x_i}$ is the sample standard deviation, $n$ is the sample size and $df$ is the degrees of freedom. This statistical significance test is applied to the averaged smooth fields of interest at each time and vertical level.

One important thing to note is that the use of a Student’s t-test is not particularly suited for this study. Given our limited sample size it is extremely difficult to extract a statistically significant signal. Likewise, fields such as total precipitation are not necessarily normally distributed at any one grid point over all cases. The fact that this study does not implement a storm relative composite also means the variability at any one grid point could potentially be quite large. Future work may necessitate the use of a storm relative composite in conjunction with the development of statistical significance metric than can account for the small sample size used here.
2.5 Potential Vorticity Analysis

The use of PV to study of climate change impacts on extratropical cyclones is unique to this study as most of the previous work that has focused on changes in intensity do not diagnose why those changes occur at the storm-scale level. Even the studies that have investigated storm-scale changes do not make use of PV despite its many advantages. As previously mentioned PV is conserved for adiabatic, frictionless flow; therefore, changes due to non-conservative processes such as latent heating can be easily diagnosed. This study computes Ertel PV (EPV) of the form:

\[ P = g(f \hat{k} + \bar{\nabla}_p \times \bar{V}_h) \cdot \bar{V}^3 \theta, \]

where \( P \) is potential vorticity, \( g \) is the gravitational constant, \( f \) is the Coriolis parameter, \( \bar{\nabla}_p \) is the quasi-horizontal vector gradient on a pressure surface, \( \bar{V}_h \) is the horizontal wind vector, \( \bar{V}^3 \) is the three-dimensional vector gradient in pressure coordinates, and \( \theta \) is potential temperature.

The effects of latent heating are investigated by calculating the third term on the right side of the PV tendency equation:

\[ \frac{\partial P}{\partial t} = -\bar{\nabla}_p \cdot (P \bar{V}_h) - \frac{\partial}{\partial p} (P \omega) - g \bar{V}^3 \cdot \bar{y}, \]

where \( \omega \) is the vertical velocity in pressure coordinates and \( \bar{y} \) is the nonadvective EPV flux (Cammas et al. 1994). The first two terms of equation 2.4 represent horizontal and vertical EPV flux convergence, respectively, while the third-term is the nonadvective EPV flux convergence. The nonadvective PV flux is expressed as

\[ \bar{y} = -\hat{\theta} \bar{\zeta}_a + \bar{V}^p \theta \times \bar{F}, \]
where $\dot{\theta}$ is the latent heating rate, $\zeta_a$ is the absolute vorticity vector and $\vec{F}$ is friction. The first two terms are the vertical diabatic and shear diabatic terms, respectively. While the effects of friction are known to be important, only the first two terms are computed in order to focus on the effects of latent heating. Latent heating is parameterized using the approximation from Emanuel et al. (1987)

$$\dot{\theta} = \omega \left( \frac{\partial \theta}{\partial t} - \frac{\gamma_m}{\gamma_d} \frac{\theta}{\theta_e} \frac{\partial \theta_e}{\partial p} \right),$$

where $\gamma_m$ is the moist-adiabatic lapse rate, $\gamma_d$ is the dry-adiabatic lapse rate and $\theta_e$ is the equivalent potential temperature.

The dynamic tropopause, a surface of constant EPV, is also computed and used to analyze the evolution of United States East Coast extratropical cyclones. The dynamic tropopause for this study is computed on the 1.0, 1.5 and 2.0 potential vorticity unit (PVU) surfaces (1 PVU = $10^{-6}$ m$^2$ s$^{-1}$ kg$^{-1}$ K). As Morgan and Nielsen-Gammon show in their 1998 study, computing and plotting the dynamic tropopause is an extremely efficient tool for showing the synoptic development of extratropical cyclones. The advantages of the dynamic tropopause lie in the ability to plot three-dimensional features such as tropopause folds in a two-dimensional space. Regions of non-conservation due to latent heat release, such as in the downstream ridge, can also be identified using this method since PV is conserved for adiabatic, frictionless flow. The use of dynamic tropopause maps in tandem with the PV budget and plots of EPV at various levels allow for a thorough analysis of extratropical cyclogenesis and baroclinic instability in the PV framework. In the next chapter, we make
full use of these tools to investigate changes in United States East Coast extratropical cyclone structure and intensity while also diagnosing why those changes occur.
Table 2.1: Starting and ending times for each case within the Miller-A and Miller-B composites.

<table>
<thead>
<tr>
<th>Case</th>
<th>Miller-A Start Time</th>
<th>Miller-A End Time</th>
<th>Miller-B Start Time</th>
<th>Miller-B End Time</th>
</tr>
</thead>
<tbody>
<tr>
<td>4</td>
<td>0000 UTC 3 January 1994</td>
<td>1200 UTC 6 January 1994</td>
<td>0300 UTC 28 December 1997</td>
<td>1500 UTC 31 December 1997</td>
</tr>
<tr>
<td>5</td>
<td>0900 UTC 6 January 1996</td>
<td>2100 UTC 9 January 1996</td>
<td>0300 UTC 18 December 2000</td>
<td>1500 UTC 21 December 2000</td>
</tr>
<tr>
<td>6</td>
<td>0300 UTC 15 February 1996</td>
<td>1500 UTC 18 February 1996</td>
<td>2100 UTC 3 February 2001</td>
<td>0900 UTC 7 February 2001</td>
</tr>
<tr>
<td>8</td>
<td>1500 UTC 10 February 2006</td>
<td>0300 UTC 14 February 2006</td>
<td>0500 UTC 21 January 2005</td>
<td>2100 UTC 24 January 2005</td>
</tr>
<tr>
<td>9</td>
<td>0000 UTC 18 December 2009</td>
<td>1200 UTC 21 December 2009</td>
<td>1500 UTC 12 February 2007</td>
<td>0300 UTC 16 February 2007</td>
</tr>
<tr>
<td>10</td>
<td>0300 UTC 25 December 2010</td>
<td>1500 UTC 28 December 2010</td>
<td>1500 UTC 8 February 2010</td>
<td>0300 UTC 12 February 2010</td>
</tr>
</tbody>
</table>
Figure 2.1: The domains for the North American Regional Reanalysis (NARR) dataset and WRF model simulations are shown in purple and pink, respectively.
Figure 2.2: Flow charts showing how the (a) WRF cyclone composites and (b) pseudo-idealized WRF simulations are generated. To generate the WRF cyclone composite, NARR data for each cyclone event is first fed into the WRF pre-processing software (WPS) which prepares the data to be read by WRF. The pre-processed data is then used by WRF to simulate individual cyclone events. Fields of interest from each of these simulated events are then averaged together to generate a WRF cyclone composite. To generate the pseudo-idealized WRF simulation, fields of interest for each individual cyclone event are averaged together using NARR data first. This generates a NARR composite. This NARR composite cyclone is then fed into WPS to pre-process the data. The pre-processed NARR composite data is then used by WRF to generate a pseudo-idealized simulation.
Figure 2.3: Flow charts showing how the (a) future WRF cyclone composites and (b) future pseudo-idealized WRF simulations are generated. The process by the future WRF cyclone composite is generated is exactly the same as that shown in Figure 2.2 except with one additional step. Prior to feeding the NARR data for each cyclone event into WPS, thermodynamic changes are applied. A similar step is also added for the generation of future pseudo-idealized WRF simulations as well. Prior to feeding the NARR cyclone composite into WPS, thermodynamic changes are also applied.
Figure 2.4: Future minus present change in 2-m (a) temperature (°C) and (b) temperature gradient (°C/10^5m) along eastern North America at model initialization (i.e. hour 00).
3. WRF Composite Analysis

This chapter describes the synoptic evolution of the Miller-A and Miller-B WRF composites in present and future thermodynamic environments. As noted in Chapter 2, these composites are made up of individual WRF simulations of Miller-A and Miller-B cyclone events. Each simulation is initialized 36 hours prior to the centering time for its respective composite and run for 84 hours. Specifics regarding the categorization of these events and this compositing technique are found in sections 2.1 and 2.3.1, respectively. A case by case assessment of the present WRF simulations versus the NARR is discussed first. Present to future changes in the evolution of Miller-A and Miller-B storms are then investigated including their track, intensity and extremes associated with them. Other parameters of interest including changes in the total precipitation and snowfall are also analyzed. The processes responsible for these changes are analyzed next by investigating changes in the lower and upper tropospheric dynamics. Finally, these changes are put into context using the potential vorticity framework.

3.1 Assessment of Present WRF Simulations

Before analyzing any present to future changes, it is important to determine how well the present cyclone simulations match the reanalysis data. Here we will assess how well the intensity and track for each case was simulated using WRF versus the NARR itself. The root mean square (RMS) error is computed for minimum sea-level pressure and track in units of millibars and kilometers, respectively. The results of this analysis for each Miller-A and Miller-B case are presented in Table 3.1. In general, the simulations appear to match the
NARR adequately well. Sixteen of the twenty simulations exhibit RMS intensity errors less than 5 mb and fifteen of the twenty simulations exhibit RMS track errors less than 300 km. It is noted that some of the individual cases exhibit large deviations from the NARR. While this is a bit troubling, the NARR has its own limitations. This is particularly true over the ocean where observations are scarce. Although some of the individual simulations have larger errors than we would like to see, the average simulation appears to do a sufficient job overall. The average Miller-A case has an RMS intensity error of ~3.3 mb and RMS track error of ~238 km. Similarly, the average Miller-B case has an RMS intensity error of ~3.2 mb and RMS track error of ~247 km. Given these results, we felt confident enough to move forward, simulate these events in future thermodynamic environments and analyze the composite changes.

3.2 Synoptic Overview of WRF Composites

The synoptic evolutions of both the present and future Miller-A composites are highlighted in Section 3.2.1. A similar synopsis is conducted for the present and future Miller-B composites in Section 3.2.2. It should be noted that although the middle and upper tropospheric geopotential heights are greater in future due to the thermodynamic changes applied, the overall synoptic pattern remains the same from present to future. As such, the synoptic patterns of the Miller-A and Miller-B composites are only highlighted here with analysis of present to future changes left for subsequent sections.
3.2.1 Synoptic Overview of Miller-A WRF Composites

Amplified upper-level ridges are seen over much of the Western United States at the initial Miller-A compositing time (Figures 3.1a, b). Less amplified, positively-titled upper-level troughs are also evident over the Central United States. At the surface (Figures 3.1c, d), broad areas of low-pressure exist over the western Gulf of Mexico (GOM), just east of the upper-level troughs, thus exhibiting a favorable vertical tilt for surface cyclone development. Consistent with Rossby wave downstream development, the upper-level ridges decrease in amplitude by hour 24 while the upper-level troughs over the Central United States clearly amplify (Figures 3.2a, b). The troughs exhibit fairly neutral axis tilts at this time. Surface cyclones have also developed east of the amplifying upper-level troughs by this point (Figures 3.2c, d).

Synoptic patterns consistent with downstream development continue at 48 hours as the amplitudes of the upstream ridges decrease and the downstream troughs amplify/dig into the Southeast United States (Figures 3.3a, b). The 500-mb level troughs also become negatively tilted with prominent regions of absolute vorticity within their base. Since the surface cyclones are also still located east of the upper-troughs, this allows them to deepen more rapidly by way of both baroclinic and barotropic energy conversion. This is evident in the sea-level pressure field as the surface cyclones in both the present and future Miller-A composites have deepened to central pressures of less than 996 mb (Figures 3.3c, d). By 72 hours, the 500-mb troughs have clearly lifted in both the present and future composites (Figures 3.4a, b). As the upper forcing lifts to the northeast, the surface cyclones do the same and track up into the Canadian Maritimes (Figures 3.4c, d).
3.2.2 Synoptic Overview of Miller-B WRF Composites

At the initial compositing time, the Miller-B composites exhibit upper-level ridges over the Western United States (Figures 3.5a, b). However unlike the Miller-A composites, the ridges in the Miller-B composites are embedded within a split flow pattern where the flow is fairly zonal to the south. Less amplified troughs are present through the Central United States, just east of the Rocky Mountains, as well. Broad regions of low-pressure are located over much of Texas and into Oklahoma associated with these upper-level troughs (Figures 3.5c, d). As was the case for the Miller-A composites, the surface low is positioned just east of the upper-trough meaning it is in a favorable location for growth through baroclinic energy conversion. Downstream development is evident at 24 hours as the 500-mb ridges weaken and the 500-mb troughs over the Central United States amplify and shift east (Figures 3.6a, b). The flow also becomes more confluent downstream of the upper-level troughs although there is a weak shortwave disturbance propagating through the Canadian Maritimes. At the surface, the amplified upper-level troughs have resulted in the development of surface cyclones to the south and east (Figures 3.6c, d). The primary surface cyclones are located west of the Appalachian Mountains with coastal redevelopment beginning along the South Carolina coast.

By 48 hours the upper-level troughs are further amplified with much more prominent absolute vorticity maxima in their base (Figures 3.7a, b). The troughs are negatively tilted at this point allowing the surface cyclones, which are still east of the upper-troughs, to intensify through both baroclinic and barotropic energy conversion. At this point in time, both the present and future surface cyclones have attained minimum central pressures of less than 996
mb (Figures 3.7c, d). Moving ahead to 72 hours, both the present and future upper-level troughs in the Miller-B WRF composites are lifting (Figures 3.8a, b). Similarly, the surface cyclones have made their way northeastward into the Canadian Maritimes (Figures 3.8c, d) like the Miller-A composites described in the previous section.

With the basic synoptic overview complete, the following chapter investigates the changes seen between the present and future composites of each cyclone type. In particular, changes in cyclone track, intensity, precipitation and wind are analyzed. The statistical significance of these changes are also highlighted and put into context.

3.3 Analysis of Present to Future WRF Composite Changes

One challenge in any composite analysis is the inherent variability in storm track and speed. Not only does this variability in track and speed result in a smearing of the composite signal as you move away from the centering time but it can also create issues along the lateral boundaries. As mentioned in Section 2, this is made particularly difficult given the limitation that the NARR domain puts on the northern and eastern lateral boundaries. If a cyclone tracks too closely to these lateral boundaries in a simulation, the model is not free to develop its own solution. Instead, the specified boundary conditions become the only possible solution. In order to limit possible contamination from the lateral boundaries, only the first 63 hours of the Miller-A simulations and the first 69 hours of the Miller-B simulations will be analyzed henceforth. Figure 3.9 shows the location of each surface low in each respective composite at its final analysis time. These times are chosen because they are the final times that all cyclones are well within the domain in both the present and future
simulations. Changes in the track, intensity, precipitation and extremes associated with these events are now analyzed for the above times. The storm-scale dynamics responsible for these analyzed changes are also highlighted using a variety of parameters, measurements and techniques.

3.3.1 Changes in Intensity and Track

In order to more accurately compute changes in intensity, the minimum sea-level pressure of the extratropical cyclone of interest is recorded in each simulation. These values are then used to compute the mean/standard deviation and to perform statistical analysis on the intensity of each subset of cyclones (i.e. present Miller-A, etc.). This is more useful than just using the composite sea-level pressure field itself because it allows us to analyze the mean of the minima instead of analyzing the minimum of the mean field. Similarly, it removes any smearing of the intensity signal due to track variability that is intrinsic to the composite. The latitudes and longitudes of the minimum sea-level pressure locations for each cyclone are also recorded and used to compute track statistics. A 20-point Gaussian weighted smoothing function is applied to the sea-level pressure field of each simulation prior to recording the minimum sea-level pressure and latitude/longitude. The filter is applied in order to focus the tracking on the synoptic scale low and remove any mesoscale lows (due to convection along front, etc.) that may generate unnatural jumping in the track. While the filter also reduces the magnitudes of the pressure minima used to compute the mean, the overall results are not sensitive to this change. Both the Miller-A and Miller-B analyses here
begin at hour 18, as it is the first time that a continuous track (i.e. tracking code doesn’t unnaturally jump for the subsequent 12 hours) exists for all present and future simulations.

The average minimum sea-level pressure evolutions of the Miller-A simulations in Figure 3.10 show that the average future simulation starts out weaker than its present counterpart at hour 18. However, 39 – 42 hours into the simulations the average future Miller-A case becomes slightly stronger than the average present Miller-A case. This trend continues until 63 hours when the future average minimum sea-level pressure is almost 3.5 millibars less than in the present average. The fact that the storms start weaker and end up stronger in terms of minimum sea-level pressure implies that they also deepen more rapidly in the future simulations. Looking at the minimum sea-level pressure difference (i.e. future minus present) of the average Miller-A case and each individual case in Figure 3.11 this trend becomes apparent. Each of the ten cases start out with a higher minimum sea-level pressure in the future at hour 18 and all but one of those has a lower minimum sea-level pressure by hour 63. Although nearly all of the future Miller-A simulations are more intense in terms of minimum sea-level pressure by hour 63, this increase in intensity is not statistically significant at the 90% confidence level using a 2-sample, 2-tailed student’s t-test. The one standard deviation ranges, represented by the error bars in Figure 3.10, indicate that the distributions overlap despite a shift towards deeper surface cyclones.

Projected changes seen in the average minimum sea-level pressure evolution of Miller-B cyclone simulations are similar to but smaller than those projected for Miller-A cyclones. The average future Miller-B simulation starts out weaker than the average present simulation in terms of minimum sea-level pressure but gradually catches up to and surpasses
it as seen in Figure 3.12. Between 54 and 57 hours, the average future simulation becomes more intense until ultimately attaining a minimum pressure just over 1 millibar less than the present average. This transition from a weaker to stronger future cyclone occurs twelve to fifteen hours later than what is seen in the evolution of the average Miller-A simulations. As was true for the future Miller-A simulations, each of the ten future Miller-B simulations start out weaker than the present one (Figure 3.13); however, only seven of those future simulations attain a minimum sea-level pressure that is less than in the present by hour 69. Given that the decrease in future average Miller-B minimum sea-level pressure is so small, it is not surprising to find that the change is not statistically significant at the 90% confidence level using a 2-sample, 2-tailed student’s t-test. This is especially apparent when looking at how much overlap there is between the average present and future one standard deviation distributions (Figure 3.12). The question of what distinguishes storms that weaken from those that strengthen is still unresolved and presents an avenue of future research.

Noticeable changes in the average Miller-A and Miller-B storm tracks are seen in addition to the above mentioned changes in intensity. The average present Miller-A cyclone, shown in Figure 3.14a, tracks out of the coastal Gulf of Mexico region and up along the eastern United States as expected. A similar track is seen for the average future Miller-A cyclone (Figure 3.14b) but the track is predominantly shifted eastward. The individual present and future tracks for each Miller-A simulation are shown in Figure 3.15 for context. The average present Miller-B cyclone also takes an expected track as it starts on the windward side of the Appalachians and subsequently jumps southeast to the coast when the coastal redeveloping surface low becomes dominant (Figure 3.16a). From there, it tracks up
along the eastern United States and into the Canadian Maritimes. A more gradual transition across the Appalachians is seen for the average future Miller-B cyclone in Figure 3.16b. The more gradual transition seen in the average future Miller-B track is due to some of the individual future cyclones tracking more like Miller-A cyclones, specifically cases 3 and 4 (Figure 3.17b). This increases the track variability and thus the smearing of the signal. After the average future Miller-B cyclone has moved out over the Atlantic Ocean, its track is also shifted eastward relative to the present. The eastward shifts in the average future Miller-A and Miller-B tracks lie well within the one standard deviation (1-sigma) cones of their present counterparts meaning the changes are likely not statistically significant (Figures 3.14, 3.15).

All of the above changes in the intensity and track of Miller-A and Miller-B cyclones have been shown to not be statistically significant meaning a statement on how these factors will change in a warmer climate cannot be made with any degree of confidence. This is due to the fact that the variability in track and intensity among the individual simulations themselves is greater than the changes that are projected to occur due to warming. It is also important to note that the statistical significance here is likely quite sensitive to this variability since the Miller-A and Miller-B samples consist of only ten events each. While increases in intensity and an eastward shift are observed in most cases, the outliers that do not follow the same trend limit the statistical significance of these changes.
3.3.2 Changes in Precipitation: Total Precipitation and Snowfall

Future projections of precipitation indicate that the amount of precipitation associated with extreme events such as extratropical cyclones is expected to increase with warming (see Section 1.2.3.1). Composite analysis of total precipitation associated with Miller-A (Figures 3.18a, b) and Miller-B (Figures 3.18 c, d) cyclone events in present and future climates show a similar trend. The Miller-A 63 hour composite total precipitation fields in Figure 3.18 show a distinct increase in areal extent of precipitation exceeding 35 millimeters (mm) in the future (b) relative to the present (a). The region of precipitation exceeding 35 mm in the Miller-B 69-hour composite total precipitation fields show comparable results (Figure 3.18 c, d). Likewise, increases in the maximum composite total precipitation can be seen for both Miller-A and Miller-B cyclones.

Computing the area-average of composite total precipitation for the area bound from 25° to 49°N and 95° to 45°W (region shown in Figure 3.18) yields an increase of ~14% for Miller-A cyclones and ~8.5% for Miller-B cyclones. The maximum composite total precipitation for Miller-A and Miller-B cyclones also increase ~26% and ~18%, respectively. Given an approximate 3.5K increase in area-average 850-mb temperature for this region, the expected increase in atmospheric water vapor content projected by the Clausius-Clapeyron relation is ~24.5%. Increases in water vapor content of ~23% and ~24% are found in this region for Miller-A and Miller-B cyclones, respectively, which is slightly sub Clausius-Clapeyron. The area-average precipitation increases for both Miller-A and Miller-B cyclones are well below the increases found in atmospheric water vapor content. The increase in maximum total precipitation associated with the Miller-B composite is also less than its
respective water vapor increase while the increase in maximum precipitation associated with the Miller-A composite is actually super Clausius-Clapeyron.

Difference fields of composite total precipitation highlight changes in the spatial distribution of precipitation associated with these systems more clearly (Figure 3.19). As expected based on the area-average statistics computed, most areas that experience a change in composite total precipitation do so in the form of an increase. For both the Miller-A (Figure 3.19a) and Miller-B (Figure 3.19b) composites, decreases are generally seen along the eastern periphery of the composite total precipitation distribution with warming. This is likely attributable to the eastward shift in future storm tracks seen in Section 3.3.1. Likewise, increases are generally observed along the coastal Mid-Atlantic, Southeastern United States and over much of the Western North Atlantic Ocean. The statistical significance of these spatially varying changes is shown in Figures 3.20a and 3.20c. As noted in Section 2, a 20-point Gaussian weighted smoothing function is applied to the precipitation field prior to taking the difference and computing the statistical significance in order to limit the variability due to varying cyclone tracks/speeds and focus on the signal of interest. While there is a region of statistically significant precipitation increase in the Miller-A composite at the 90-95% confidence level located just off the coast of the United States, it is clear that most of the spatially varying changes are not statistically significant (Figure 3.20a). This statement is particularly true for the Miller-B composite as there are virtually zero areas of statistically significant precipitation change above at or above the 90% confidence level.

Variability in the track of these storms can result in large variances of precipitation at any one grid point. In particular, small variations in the location of frontal features or other
regions of enhanced precipitation can highly inflate the variance thus making it difficult to extract a statistically significant signal. To determine whether there are changes in the overall occurrence and intensity of precipitation associated with these cyclones in a warmer climate, frequency histograms of total precipitation within the same domain over all simulations are computed (Figure 3.21). Looking at the histogram difference (future minus present) of total precipitation over all Miller-A (Figure 3.21c) and Miller-B (Figure 3.21f) simulations, two things are evident. First, there is an overall decrease in the number of grid cells with total precipitation less than or equal to ~30-40 mm in the future simulations. Second, the number of grid cells with total precipitation exceeding 40 mm clearly increases. This implies an overall decrease in light precipitation and increase in heavy precipitation associated with Miller-A and Miller-B cyclone events along the United States East Coast. Frequency histograms of composite reflectivity over all simulations are also presented in Figure 3.22. Difference histograms of composite reflectivity over all Miller-A (Figure 3.22c) and Miller-B (Figure 3.22f) simulations lend confidence to the above assertion as the number of reflectivities below ~30 dBZ generally decreases while the number of reflectivities above ~30 dBZ increases. Additional analysis finds that the number of grid cells with measureable precipitation (greater than 0.0 mm) decreases 3.35% and 1.47% over all Miller-A and Miller-B simulations, respectively. This further substantiates the idea that there is a shift toward more heavy and less light precipitation associated with these events.

Not only does warming impact the intensity and distribution of precipitation but also the type. Composite analysis of Miller-A and Miller-B total snowfall illustrates an overall decrease in snowfall associated with these systems (Figure 3.23). While there are fairly large
areas with composite total snowfall exceeding 14 kg/m² in the present for both Miller-A (Figure 3.23a) and Miller-B (Figure 3.23c) cyclones, these regions clearly decrease in the future (Figure 3.23b, d). Area-average composite snowfall is projected to decrease considerably for these types of cyclone events. Decreases of ~28% and ~29% are seen in the area-averaged composite snowfall fields for Miller-A and Miller-B cyclones, respectively. The maximum composite snowfall associated with these events is also projected to decrease. The maximum composite snowfall for Miller-A events decrease ~14% while the maximum for Miller-B events decreases ~15%.

The spatial distribution of changes in composite total snowfall clearly shows decreases in snowfall for most areas along the Atlantic seaboard (Figure 3.24). There are some exceptions to this with increases seen in northern Maine and into Canada in the Miller-A composite (Figure 3.24a) and increases in eastern Pennsylvania, northern New Jersey and northern Maine in the Miller-B composite (Figure 3.24b). These same areas experience an increase in total precipitation and align well with elevated terrain (i.e. cold enough to support snow). This implies that all of the precipitation increases that occur in these regions do so in the form of snow. Like with total precipitation, a 20-point Gaussian weighted smoothing function is applied to the fields prior to computing the statistical significance. The composite changes in the smoothed snowfall fields and the statistical significance of those changes at each grid point are plotted for Miller-A and Miller-B cyclones in Figures 3.20b and 3.20d, respectively. Statistical significance of the spatially varying snowfall changes is primarily limited to the southern Appalachians for Miller-A cyclones with a small region of decreases significant at the 90% confidence level. The statistically significant changes for Miller-B
cyclones encompass a slightly larger area extending from western New York down into central Tennessee.

Frequency histograms of total snowfall over all Miller-A and Miller-B simulations are generated in a similar fashion to those for total precipitation in order to investigate the overall occurrence of snowfall associated with these cyclones in a warmer climate (Figure 3.25). The domain for the previous plots/histograms contained within this section is again used. The histogram difference of snowfall over all Miller-A (Figure 3.25c) and Miller-B (Figure 3.25f) cyclone events clearly show a decrease in snowfall for most but not all bins. In general, decreases are seen for most bins which are consistent with the large decreases seen in area-average composite total snowfall. However, it is interesting to note that there are slight increases (< 10 grid cells) in the number of some of the larger snowfall bins (i.e. heavier snowfall). This implies that given the right conditions (i.e. cold enough, ample moisture, forcing for ascent) increases in total snowfall are possible and, in this case, the heaviest snowfall actually occurred in the future simulations. Likewise, it means that increases in total precipitation are not restricted to rainfall in areas that are cold enough to support snow. The overall trend for both Miller-A and Miller-B cyclones is for snowfall to decrease though which, given increases in total precipitation, means that there is likely more rain and less snow associated with these systems in a warmer climate.

3.3.3 Changes in Lower and Upper-Level Storm Dynamics

Given the changes in storm track and intensity shown in Section 3.3.1, one would expect to see consistent changes in the dynamics of Miller-A and Miller-B cyclones in a
warmer climate. At first glance, the composite analysis of 10-meter (10-m) wind speed at hour 54 appears to show extensive increases in the magnitude of wind speeds for both Miller-A (Figure 3.26a, b) and Miller-B (Figure 3.26c, d) cyclones. Hour 54 is presented here as it is far enough into the simulations that make up each composite such that the cyclones are well developed. The area of composite 10-m wind speed exceeding 12.5 meters per second (m/s) visibly increases for both cyclone types. In order to better assess extreme winds associated with these events, the maximum 10-m wind speed at each grid point over all times and simulations is also computed for Miller-A (Figure 3.27a, b) and Miller-B (Figure 3.27c, d) cyclones. There appear to be isolated regions where the maximum 10-m wind speed increases but overall the spatial distributions look very similar for both cyclone types.

Difference fields of the maximum 10-m wind speed over all Miller-A (Figure 3.28a) and Miller-B (Figure 3.28b) simulations reveal some of the spatial variability that does exist. Increases in maximum 10-m wind speed are primarily found along the Atlantic coastline and over the Atlantic Ocean for both Miller-A and Miller-B cyclones. The increases in maximum 10-m wind speed over all Miller-B simulations are mostly confined to an isolated strip while the increases for all Miller-A simulations are generally more widespread. This implies that the increases seen over all Miller-B simulations are likely dominated by a few storms (i.e. narrow swath of increases) whereas increases are seen over a larger number of cases for Miller-A storms. These findings are largely consistent with the minimum sea-level pressure differences shown for all Miller-A (Figure 3.11) and Miller-B (Figure 3.13) simulations in Section 3.3.1. While seven out of ten Miller-A cases had future simulations with minimum
central pressures at least 2 millibars lower than the present (i.e. pressure difference < -2 mb),
only four of ten Miller-B exhibited such a change.

One question regarding these increases in 10-m wind speed that needs to be resolved
is whether or not they are attributable solely to increases in intensity or to both changes in
intensity \textit{and} track. Histograms of 10-m wind speed exceeding zero m/s within the domain
shown in Figures 3.26-3.28 (area bound from 25° to 49°N and 90° to 45°W) are generated
over all Miller-A and Miller-B simulations to better diagnose these changes (Figure 3.29).
Looking first at the domain wide changes in 10-m wind speed, the same general trend is seen
in the difference histograms for Miller-A (Figure 3.29c) and Miller-B (Figure 3.29f)
simulations. Most notably, there are obvious increases in the bins exceeding 16-17 m/s.

Partitioning the difference histograms into the changes over land versus over water, it
becomes apparent that most of these changes are indeed due to the projected shift in track.
Both the Miller-A (Figure 3.30a) and Miller-B (Figure 3.30b) difference histograms over
land generally show increases in light winds (~1-5 m/s) and decreases in winds with
magnitudes of ~6-18 m/s. On the other hand, there are slight increases in wind speed bins
exceeding 18 m/s over all Miller-A simulations while there are decreases or no change in
these bins over all Miller-B simulations. Difference histograms of 10-m wind speed over
water for all Miller-A (Figure 3.31a) and Miller-B (Figure 3.31b) simulations look extremely
similar to those shown over all points in Figures 3.29. This implies that the changes seen in
10-m wind speed over all Miller-A and Miller-B simulations are largely dominated by
changes over water. The eastward shift in future tracks seen in Section 3.3.1 is likely
responsible for most of these changes given that more of the wind field associated with these
cyclones tracks over water in the future. The reduced surface drag over water allows for higher wind speeds in the future even if a storm does not exhibit a change in surface cyclone intensity (i.e. change in minimum sea-level pressure). The increases found in 10-m wind speeds exceeding 18 m/s over water and land for Miller-A cyclones suggests that it is primarily due to the shift in track but also slightly due to the increase in intensity. On the other hand, the fact that increases in 10-m wind speeds exceeding 18 m/s are only found over water for Miller-B cyclones suggests it is solely due to the shift in track.

Like near-surface wind speed, another dynamical feature of interest is the low-level jet (LLJ). The LLJ, as its name insinuates, is an isolated region of high wind speeds at low levels in the atmosphere. These features are commonly observed in the warm sector of extratropical cyclones ahead of the cold front and can be enhanced by strong regions of condensational heating. LLJs are quite important to extratropical cyclones as they help transport moisture into the system. As discussed in Section 1.2.3, increases in precipitation associated with these types of cyclones should result in increased diabatic PV generation at low levels. This may yield stronger southerly flow east of the diabatic PV anomaly and enhance the LLJ (Lackmann and Gyakum 1999, Lackmann 2002, Brennan et al. 2008).

Figure 3.32 presents composite analysis of the magnitude of the 850-mb wind and specific humidity at hour 57. Only contours exceeding 20 m/s are plotted in order to focus on the LLJ signal. For both the Miller-A (Figure 3.32a, b) and Miller-B (Figure 3.32c, d) composite analyses, there is an evident increase in both the future 850-mb wind speed (i.e. LLJ) and amount of moisture that has been transported northward. The maximum wind speed contour increases from 25 m/s in the present to 30 m/s in the future for each composite type.
Likewise, specific humidity values greater than 0.008 kilograms (kg) of water vapor per kg of moist air are present within the future LLJ cores, which clearly exceed the values seen in the present.

The statistical significance of changes in LLJ strength is determined by analyzing changes in the meridional wind components for the same time at 850 mb (Figure 3.33a, c). A couplet of decreases to the west and increases to the east is seen for both composites with large areas of statistically significant change at the 90% confidence level and higher. It is particularly worth noting that large portions of the future LLJs from Figure 3.32b and 3.32d show statistically significant change in Figure 3.33; this implies a statistically significant track change. These statistically significant changes encompass a larger area in the Miller-A composite than the Miller-B composite. It is a bit surprising that the decreases in the meridional wind component are larger and more significant than the increases given that the LLJ appears to strengthen and shift. This can be explained by looking at the wind vectors for the Miller-A (Figure 3.33b) and Miller-B (Figure 3.33d) composites. The areas with large v-wind decreases see large directional changes in their wind vectors with the vectors becoming more meridional (turning southward) and in some cases stronger (i.e. longer) as well. On the other hand, areas with v-wind increases generally become slightly less meridional (turning eastward) which would explain why the increases are smaller than the decreases in Figures 3.33a and 3.33c. However, looking at the wind vectors it is clear that wind speeds increase in the vicinity of the LLJ.

Based on this analysis alone it would appear that the LLJ associated with these types of cyclones increases in intensity in the future but some of this statistically significant change
may be due to the shift in track. Another issue is that the level at which the LLJ is maximized might differ in the future due to changes in the melting level and level of maximum heating. Histograms of wind speed greater than or equal to 25 m/s between 900 and 800 mb over all Miller-A and Miller-B simulations are generated in order to mitigate these issues and assess the overall change (Figure 3.34). These histograms include hours 00 to 63 for Miller-A simulations and 00 to 69 for Miller-B simulations within the same regions shown in Figures 3.31 and 3.32 (area bound from 25° to 49°N and 90° to 45°W). The frequency of most bins greater than ~30 m/s increases in the future over all Miller-A (Figure 3.34c) and Miller-B (Figure 3.34f) simulations. Decreases are found for most bins less than 30 m/s in both the Miller-A and Miller-B histograms as well which suggests a shift towards LLJ cores of greater strength. This is consistent with what was seen in the composite analysis (Figure 3.32) as the number of grid cells greater than 25 m/s but less than 30 m/s decreased and the number of grid cells exceeding 30 m/s increased. Section 3.3.4 will investigate whether or not these increases in LLJ strength are consistent with an increase in low-level PV.

Tropical upper-tropospheric heating and polar lower-stratospheric cooling is projected to strengthen the upper-level meridional temperature gradient. Cross-sections of zonally averaged temperature and meridional temperature gradient change from 25° to 49°N along the 90°W meridian at 48 hours are presented in Figure 3.35. The temperature gradient becomes more negative in much of the upper troposphere above 300 mb for both Miller-A (Figure 3.35b) and Miller-B (Figure 3.35d) cyclones which represents a strengthening of the meridional gradient. Thermal wind balance dictates that there must be an increase in westerly wind shear to compensate for this strengthening of the upper-level temperature gradient.
Composite analysis of the change in 250-mb zonal wind component for Miller-A (Figure 3.36a) and Miller-B (Figure 3.36b) cyclones at the same time shows that there is indeed an increase in the westerly winds aloft over most areas. Future geopotential height contours are plotted in Figure 3.36 for context. There are also regions with large decreases in westerly winds over the northeastern United States and Canadian Maritimes. These changes become more robust as the cyclones evolve and are likely due to enhanced downstream ridging as a result of increased latent heat release associated with future cyclones.

The statistical significance of these changes is assessed for the Miller-A (Figure 3.37a) and Miller-B (Figure 3.37b) simulations using the smoothed fields as previously described. Like many of the other fields investigated, there is limited statistical significance due to the large variability in the location of features of interest. In this case, variability in the exact location of the upper-level jet increases the variance such that it is difficult to extract a statistically significant signal. With that said, there are some isolated regions within the core of zonal wind increases that are statically significant at the 90% confidence level and higher. In order to diagnose common changes independent of location, histograms of the 250-mb zonal wind component over all cases and times within each composite are presented here (Figure 3.38). Difference histograms for Miller-A (Figure 3.38c) and Miller-B (Figure 3.38f) cyclones show that the number of grid cells with zonal wind components exceeding ~25 m/s generally increases over all times and cases for both composite types. This is consistent with the composite analysis in Figures 3.36 and 3.37 which showed increases over most areas at hour 48.
Increases in upper-level wind speed have important implications on the dynamics of surface cyclones. Composite analysis of the 250-mb jet and divergence is presented in Figure 3.39. Comparing the present and future analysis of the Miller-A (Figure 3.41a, b) and Miller-B (Figure 3.41c, d) composites at hour 48, two things are apparent. First, the future jet cores are slightly stronger in intensity with larger areas exceeding 60 m/s for both composite types. Second, a comparable increase in upper-level divergence is seen with the increase in jet strength. Since the future composites exhibit lower central surface pressures and increased deepening rates, it is not surprising to find enhanced divergence aloft. As dictated by mass continuity, the amount of mass being evacuated from the column above the cyclone must increase in the future if there are lower surface pressures. Although this is not necessarily true in a composite sense because of smearing, it is encouraging to see that upper divergence shows this change. Here the strongest divergence is found in the left jet exits and is reasonably positioned above the tracks seen in Figures 3.14 and 3.16.

Not only do increases in upper-level wind speed impact the intensity but also the track of surface cyclones. The speed at which an upper-level trough moves can dictate the development and general motion of a surface cyclone as described in Section 1.2.2. Although the systems studied here are baroclinic in nature, the barotropic Rossby wave equation provides an estimate of the zonal phase speed of an upper-level trough. The barotropic Rossby wave phase speed equation is expressed as,

$$(3.1) \quad C_x = \bar{U} - \frac{\beta}{(k^2 - l^2)}$$
where $C_x$ is the zonal phase speed, $\bar{U}$ is the mean zonal wind speed, $\beta$ is the Rossby parameter and $k$ and $l$ are the zonal and meridional wavenumbers, respectively. Based on this equation, one would then expect changes in the phase speed of an upper-level trough given a change in the base state zonal flow, latitude or the zonal/meridional wavelengths. The large increases seen in the zonal wind component are then clearly consistent with an increase in the phase speed of upper-level troughs and may thus explain, to an extent, the eastward shift seen in future simulations for both Miller-A and Miller-B cyclones.

Composite analysis of pressure and wind on the 1.5 PVU surface reveals that the upper-level troughs do indeed propagate eastward more quickly in the future. Hour 48 is again analyzed here because the future mean storm tracks have deviated from their present counterparts quite a bit by this point in their evolution (see Figures 3.14 & 3.16). For both the Miller-A (Figure 3.40a,b) and Miller-B (Figure 3.40c,d) composites, the lowest pressures on the dynamic tropopause in the future are shifted east relative to the present. Figure 3.41 shows this shift more clearly by computing the difference (future minus present) of pressure on the dynamic tropopause. Couplets of pressure increases and decreases representing the future shift are evident for both the Miller-A (Figure 3.41a) and Miller-B (Figure 3.41b) composites. This is consistent with the future increases seen in upper-level westerlies and our understanding of the barotropic Rossby wave phase speed equation. Though no attempt is made here to analyze changes in the average latitude or zonal/meridional wavelengths, it appears that any changes in these that may act to slow the upper-trough speed do not exceed the increase in base state westerly flow.
Now that we have a physical explanation for the changes in future Miller-A and Miller-B tracks, it is important to shift the focus toward the processes responsible for the changes in intensity. The following section investigates changes in EPV and the background Eady growth rate in an attempt to assess their contributions to changes in the intensity of Miller-A and Miller-B cyclones in a warmer climate.

3.3.4 Eady Growth Rate and Potential Vorticity Analysis

Having already diagnosed the eastward shifts seen in future Miller-A and Miller-B storm tracks, this section employs the maximum Eady growth rate and EPV to assess the changes in extratropical cyclone intensity seen in Section 3.3.1. The maximum Eady growth rate, expressed by equation 1.1, is a measure of a system’s baroclinic potential for growth. Here we compute the area average of 1000–500-mb maximum Eady growth rate for each composite type at hour 00. We chose to analyze hour 00 in order to assess the baroclinic potential for growth within the environment prior to the cyclogenesis process. The area average maximum Eady growth rate is found to increase for both Miller-A and Miller-B cyclones in a warmer climate. The Miller-A growth rate increases ~1.99% while the Miller-B growth rate increases ~2.66%. Neither of these increases is statistically significant at the 90% confidence level when a 2-sample, 2-tailed Student’s t-test for equal sample size and variance is applied. It is somewhat surprising to find that the maximum Eady growth rate increases more so for Miller-B than Miller-A cases given their respective increases in intensity. This suggests one of three things: other changes must occur that offset the increase in maximum Eady growth rate for Miller-B cases, something changes structurally that does not allow
Miller B cases to fully realize their potential growth, or there are large changes in the diabatic processes affecting Miller-B cases.

The Rossby penetration depth is another useful parameter for determining how favorable the environment is for cyclogenesis. As discussed in Section 1.2.2, it is a measure of how far a PV anomaly’s circulation projects vertically (i.e. UPV circulation onto surface, LPV circulation onto UPV anomaly). Like the maximum Eady growth rate, the Rossby penetration depth is inversely proportional to static stability (see equations 1.1 & 1.2). Here stability is computed between the dynamic tropopause (1.5 PVU surface) and 2-m level,

\[
(3.2) = -\frac{\partial \theta}{\partial p} = -\frac{(\theta_{\text{dyn trop}} - \theta_{2m})}{(p_{\text{dyn trop}} - p_{2m})}
\]

where \(-\frac{\partial \theta}{\partial p}\) is the stability, \(\theta\) is potential temperature and \(p\) is pressure. The subscripts “2m” and “dyn trop” represent values at the 2-m level and on the dynamic tropopause, respectively. One over the stability (i.e. \((\frac{\partial \theta}{\partial p})^{-1}\)) gives us what will be referred to as the coupling potential. Changes in this coupling potential represent changes in the ability of the UPV and LPV anomalies’ circulations to project onto one another. If the future increases in intensity are partly due to baroclinic processes, we would expect to see increases in the coupling potential (i.e. increases in Rossby penetration depth) which would allow the UPV and LPV anomalies to interact more.

Like the maximum Eady growth rate, we compute the area average of coupling potential at hour 00 in order to assess the favorability of the environment prior to cyclogenesis. The environmental coupling potential is found to decrease ~5.48% for Miller-A cyclones and ~5.83% for Miller-B cyclones with warming. This suggests that while there
may be a greater baroclinic potential for growth associated with these storms in the future, it may be more difficult for UPV and LPV anomalies to actually couple. However, the ability of the UPV and LPV anomalies to couple not only depends on the Rossby penetration depth but on the strength of the anomalies themselves.

In addition to the changes in phase speed of the UPV anomalies described in Section 3.3.3, there are also changes in their vertical extent and accordingly strength. Referring back to the analysis of pressure on the dynamic tropopause for the Miller-A composite at hour 48 (Figures 3.40a,b), we see that the 1.5 PVU surface extends down to 450-500 mb in the present but extends down past 500 mb in the future. This analysis at the same time for the Miller-B composite (Figures 3.40c,d) shows no change in the vertical extent of the 1.5 PVU surface. However, looking ahead 9 hours we find clear changes in the dynamic tropopause analysis (Figure 3.42). Although the 1.5 PVU surface in the future Miller-A composite no longer extends below 500 mb, there is clearly a larger area that extends down to 450-500 mb relative to the present (Figures 3.42a,b). In the case of the future Miller-B composite, the 1.5 PVU surface now extends down to 450-500 mb while the present only extends down to 400-450 mb (Figures 3.42c,d). With the future 1.5 PVU surface extending closer to the surface in both composites, it allows more stratospheric high PV air to move downwards and enhance cyclogenesis through the coupling process.

Given the large future increases in precipitation shown in Chapter 3, we anticipate that enhanced diabatic PV generation plays a large role in affecting the intensity of these storms. Here EPV anomalies are computed for each simulation by subtracting the base-state meridional PV gradient from the full EPV. Profiles of these EPV anomalies are then
generated by averaging the EPV anomalies within 400 a kilometer horizontal radius of the surface cyclone center at each isobaric level, every 25 mb, from 950 to 500 mb. This results in a grid point average within the 400 kilometer radius. A radius of 400 kilometers is used to make sure that important features, such as the low-level diabatic PV anomaly along the warm front, are captured. Once these the profiles have been generated for each simulation, they are averaged over all cases and several consecutive times to generate a mean time-averaged profile. Shorter time-averaging intervals are used for the Miller-B profiles in order to better capture some of the changes that occur due to the interaction with topography and coastal redevelopment process.

The evolutions of the time-averaged Miller-A and Miller-B EPV anomaly profiles are shown in Figures 3.43 and 3.44, respectively. Looking first at the evolution for Miller-A cyclones, we see that the future EPV anomalies are initially weaker below ~600 mb but those anomalies increase and eventually surpass the present anomalies with time. By the final time averaging period, the maximum future anomaly and anomaly increase is found at ~825 mb which is about where you would expect the diabatic PV maxima to be found (Figure 3.43d). A quite different evolution is found for the time-averaged EPV anomaly profiles of Miller-B cyclones. The present and future low-to-mid-level PV anomalies are initially much stronger than those seen for Miller-A cyclones and then suddenly weaken (Figures 3.44a-c). This is a result of the original cyclone interacting with the Appalachians by hours 36-42 and subsequently undergoing redevelopment. Like the Miller-A profiles, the EPV anomalies are initially weaker below ~600 mb. After the redevelopment occurs the EPV anomalies actually
remain weaker during the Miller-B evolution except for at 950 mb (Figure 3.44d-e). This is quite surprising since increases in future precipitation were found for both cyclone types.

Profiles of latent heating and non-advective PV tendency are generated in a similar manner to determine if there are any large discrepancies in the diabatic generation of PV between the two cyclone types. Latent heating is found to increase at all levels from 950 to 500 mb and over all time-averaged periods for both Miller-A and Miller-B cyclones (Figures 3.45 & 3.46). Similarly, increases in non-advective PV tendency are found at low-levels (below ~700 mb) over all time-averaged periods for both cyclone types (Figures 3.47 & 3.48). In general, the increases in low-level non-advective PV tendency are larger for Miller-A cyclones which is consistent with Miller-A cyclones having larger precipitation increases.

How can the low-level non-advective PV tendencies increase in the future for both cyclone types but only Miller-A cyclones experience an increase in their low-level diabatic PV anomalies? The answer likely lies in the terms we have not computed or analyzed from equations 2.5 and 2.6. Here we have not computed the horizontal and vertical EPV flux convergences (first two terms of equation 2.5) or the effects of friction (third term in equation 2.6). A calculation of the full EPV budget is necessary to determine the relative importance of each of these processes and is left for future work.

Finally, changes in the LPV anomaly are assessed. As described in Bretherton (1966), a boundary potential temperature anomaly can act as a PV anomaly where warm (cold) surface potential temperature anomalies are the equivalent of cyclonic (anticyclonic) PV anomalies. The approach used here is similar to the one used to generate the EPV anomaly profiles except now we compute the mean 2-m potential temperature anomaly. The base-state
The meridional temperature gradient is subtracted from the full 2-m potential temperature and the grid scale average within a 400 kilometer horizontal radius of the surface cyclone is calculated. The same time-averaging intervals used for the profiles are also used here. Analysis of the mean 2-m potential temperature anomalies for present (Figure 3.49a) and future (Figure 3.49b) Miller-A cyclones reveal that the LPV anomalies do indeed increase in intensity in the future. The present anomalies for Miller-A cyclones are negative over all time-averaging intervals, likely a result of the large averaging radius, while the future anomalies become positive during the final three intervals. Increases are also seen for Miller-B cyclones when comparing its present (Figure 3.50a) and future (Figure 3.50b) 2-m potential temperature anomalies. Unlike the Miller-A anomalies, only the first three time-averaging intervals for present Miller-B cyclones are exhibit negative anomalies. More importantly, the present and future anomalies are much larger during the middle and late stages of the cyclone evolution for Miller-B cyclones. In particular, the future 2-m potential temperature anomalies for Miller-A and Miller-B cyclones at their final time-averaging intervals are 1.76 K and 4.20 K, respectively. This suggests that the LPV anomaly may play a larger role in the development of Miller-B cyclones than Miller-A cyclones (i.e. much more dominant feature). Increases in the mean 2-m potential temperature anomaly for Miller-A (Figure 3.51a) and Miller-B (Figure 3.51b) cyclones are somewhat comparable although the largest increases are found in the Miller-B composite.

The changes analyzed here paint a complex picture of how the storm-scale dynamics of these systems may change in the future. A number of environmental and structural changes occur in the future which are used here to explain the increases we find in intensity.
To some extent, the future systems appear to have an increased potential for baroclinic growth within this region. At the same time, the environmental potential for coupling is actually found to decrease. Future UPV anomalies appear to extend closer to the surface during the latter stages of their evolution while the future LPV anomalies are more intense over all times. It is possible that the increases in UPV/LPV anomaly strength may offset the decreases found in environmental coupling potential. The largest differences between the two composites are found in the lower tropospheric diabatic PV maxima. During the latter stages of their evolution, Miller-A cyclones see a future increase in their diabatic PV maximum while Miller-B cyclones actually see a decrease. This is despite the fact that both produce more precipitation in the future and see increases in latent heating and diabatic PV generation. A summary of the factors that may lead to stronger and weaker storms for each cyclone type is presented in Table 3.2.

In the case of Miller-A cyclones, the enhanced LPV and diabatic PV anomalies in the future appear to augment each other thus potentially allowing them to couple with the UPV anomaly more easily. On the other hand, the enhanced LPV and weakened diabatic PV anomalies in future Miller-B cyclones appear to somewhat offset one another. This effectively limits the potential for coupling and may consequently delay the intensification process until the LPV anomaly increases can adequately compensate for the diabatic PV anomaly decreases. Although several questions regarding the evolution of these storms remain, the conceptual framework for the increases in future Miller-A and Miller-B cyclone intensity proposed above provides a conceptual basis with which to work off in the future.
Table 3.1: Root mean square (RMS) error is presented for each Miller-A and Miller-B simulation in terms of minimum sea-level pressure (MinSLP) and track. RMS error for minimum sea-level pressure and track are expressed in units of millibars (mb) and kilometers (km), respectively.

<table>
<thead>
<tr>
<th>Case</th>
<th>Miller-A MinSLP (mb)</th>
<th>Miller-A Track (km)</th>
<th>Miller-B MinSLP (mb)</th>
<th>Miller-B Track (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>6.2151</td>
<td>268.5085</td>
<td>5.846</td>
<td>288.6023</td>
</tr>
<tr>
<td>2</td>
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<td>1.7364</td>
<td>187.2185</td>
</tr>
<tr>
<td>3</td>
<td>5.3336</td>
<td>334.4745</td>
<td>3.4429</td>
<td>168.4254</td>
</tr>
<tr>
<td>4</td>
<td>4.2365</td>
<td>443.1385</td>
<td>1.8006</td>
<td>403.4018</td>
</tr>
<tr>
<td>5</td>
<td>2.9408</td>
<td>306.6858</td>
<td>1.9547</td>
<td>308.6329</td>
</tr>
<tr>
<td>6</td>
<td>1.2944</td>
<td>140.0055</td>
<td>4.2494</td>
<td>280.6756</td>
</tr>
<tr>
<td>7</td>
<td>2.2814</td>
<td>221.312</td>
<td>2.2108</td>
<td>186.562</td>
</tr>
<tr>
<td>8</td>
<td>4.4251</td>
<td>148.6219</td>
<td>5.2212</td>
<td>226.5259</td>
</tr>
<tr>
<td>9</td>
<td>1.0678</td>
<td>151.1394</td>
<td>1.0904</td>
<td>199.3618</td>
</tr>
<tr>
<td>10</td>
<td>2.6325</td>
<td>226.6576</td>
<td>4.1544</td>
<td>222.7832</td>
</tr>
</tbody>
</table>
Table 3.2: Factors that lead to stronger and weaker Miller-A and Miller-B cyclones in the future.

<table>
<thead>
<tr>
<th>Future Miller-A Cyclones</th>
<th>Future Miller-B Cyclones</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Factors for Strengthening</strong></td>
<td><strong>Factors for Weakening</strong></td>
</tr>
<tr>
<td>Increased area-average maximum Eady growth rate</td>
<td>Decreased area-averaged coupling potential</td>
</tr>
<tr>
<td>Stronger UPV anomaly during second half of evolution</td>
<td></td>
</tr>
<tr>
<td>Stronger diabatic PV maximum by latter stages of evolution</td>
<td></td>
</tr>
<tr>
<td>Stronger low-level warm (PV) anomaly throughout evolution</td>
<td></td>
</tr>
</tbody>
</table>
Figure 3.1: 500-mb (a,b) and surface (c,d) analysis of the present (a,c) and future (b,d) Miller-A WRF composite at hour 00. Geopotential height (dam) [contours] and absolute vorticity (10^{-5} s^{-1}) [color fill] are plotted at the 500-mb level. Mean sea-level pressure (mb) and simulated radar reflectivity (dBZ) are plotted at the surface.
Figure 3.2: As in Figure 3.1 except for hour 24.
Figure 3.3: As in Figure 3.1 except for hour 48.
Figure 3.4: As in Figure 3.1 except for hour 72.
Figure 3.5: 500-mb (a,b) and surface (c,d) analysis of the present (a,c) and future (b,d) Miller-B WRF composite at hour 00. Geopotential height (dam) [contours] and absolute vorticity ($10^{-5} \text{ s}^{-1}$) [color fill] are plotted at the 500-mb level. Mean sea-level pressure (mb) and simulated radar reflectivity (dBZ) are plotted at the surface.
Figure 3.6: As in Figure 3.5 except for hour 24.
Figure 3.7: As in Figure 3.5 except for hour 48.
Figure 3.8: As in Figure 3.5 except for hour 72.
Figure 3.9: Location of the surface cyclones, determined by minimum sea-level pressure, that make up each composite [red X]. The (a) present Miller-A and (b) future Miller-A simulations are analyzed up to hour 63. The (c) present Miller-B and (d) future Miller-B simulations are analyzed up to hour 69.
Figure 3.10: Average minimum sea-level pressure (mb) evolution of the present [blue] and future [red] Miller-A simulations. Error bars denote the one standard deviation range.
Figure 3.11: Minimum sea-level pressure difference (mb) between the present and future Miller-A simulations. Difference is computed as the future minimum sea-level pressure minus the present minimum sea-level pressure. The average difference is plotted as the thick black line with error bars denoting the one standard deviation range. The difference for each case [see above legend for colors] is also plotted for context.
Figure 3.12: Average minimum sea-level pressure (mb) evolution of the present [blue] and future [red] Miller-B simulations. Error bars denote the one standard deviation range.
Figure 3.13: Minimum sea-level pressure difference (mb) between the present and future Miller-B simulations. Difference is computed as the future minimum sea-level pressure minus the present minimum sea-level pressure. The average difference is plotted as the thick black line with error bars denoting the one standard deviation range. The difference for each case [see above legend for colors] is also plotted for context.
Figure 3.14: Tracks of the average (a) present [solid blue] and (b) future [solid red] Miller-A surface cyclones for hours 18-63. The one standard deviation (i.e. 1-sigma) cone for the average present and average future track is shown with blue dashed and red dashed lines, respectively.
Figure 3.15: Tracks of the individual (a) present and (b) future Miller-A surface cyclones for each case. Each track is shown for hours 18-63 of its respective simulation. The track for each case is labeled with a different color as shown in the legends above.
Figure 3.16: Track of the average (a) present [solid blue] and (b) future [solid red] Miller-B surface cyclones for hours 18-69. The one standard deviation (i.e. 1-sigma) cone for the average present and average future track is shown with blue dashed and red dashed lines, respectively.
Figure 3.17: Tracks of the individual (a) present and (b) future Miller-B surface cyclones for each case. Each track is shown for hours 18-69 of its respective simulation. The track for each case is labeled with a different color as shown in the legends above.
Figure 3.18: Composite total precipitation for Miller-A and Miller-B cyclones. The (a) present and (b) future Miller-A composite total precipitations are for hours 00 through 63. The (c) present and (d) future Miller-B composite total precipitations are for hours 00 through 69. Precipitation is plotted in units of millimeters (mm).
Figure 3.19: Composite total precipitation change for Miller-A and Miller-B cyclones. The (a) Miller-A composite total precipitation change is for hours 00 through 63. The (b) Miller-B composite total precipitation change is for hours 00 through 69. Precipitation is plotted in units of millimeters (mm).
Figure 3.20: Composite (a,c) total precipitation change [contours], (b,d) snowfall change [contours] and statistical significance [color fill] for Miller-A and Miller-B cyclones. The (a,b) Miller-A composite changes are for hours 00 through 63. The (c,d) Miller-B composite changes are for hours 00 through 69. Precipitation is plotted in units of millimeters (mm). A 20-point Gaussian weighted smoothing function is applied to the fields prior to computing the difference and statistical significance. Statistical significance is plotted for the 90, 95 and 99% confidence levels using a 2-sample, 2-tailed Student’s t-test for equal sample sizes and variances.
Figure 3.21: Histograms of total precipitation (mm) values greater than zero over all Miller-A and Miller-B simulations. The (a) present, (b) future and (c) difference histograms for Miller-A simulations are over 63 hours. The (d) present, (e) future and (f) difference histograms for Miller-B simulations are over 69 hours. Precipitation bins are every 1 mm.
Figure 3.22: Histograms of composite reflectivity (dBZ) values greater than zero over all Miller-A and Miller-B simulations. The (a) present, (b) future and (c) difference histograms for Miller-A simulations are over 63 hours. The (d) present, (e) future and (f) difference histograms for Miller-B simulations are over 69 hours. Composite reflectivity bins are every 1 dBZ.
Figure 3.23: Composite snowfall for Miller-A and Miller-B cyclones. The (a) present and (b) future Miller-A composite snowfall totals are for hours 00 through 63. The (c) present and (d) future Miller-B composite snowfall totals are for hours 00 through 69. Snowfall is plotted in units of kilograms per meter squared (km/m$^2$).
Figure 3.24: Composite snowfall change for Miller-A and Miller-B cyclones. The (a) Miller-A composite snowfall change is for hours 00 through 63. The (b) Miller-B composite snowfall change is for hours 00 through 69. Snowfall is plotted in units of kilograms per meter squared (kg/m²).
Figure 3.25: Histograms of total snowfall (kg/m²) values greater than zero over all Miller-A and Miller-B simulations. The (a) present, (b) future and (c) difference histograms for Miller-A simulations are over 63 hours. The (d) present, (e) future and (f) difference histograms for Miller-B simulations are over 69 hours. Snowfall bins are every 1 kg/m².
Figure 3.26: Composite 10-m wind speed (m/s) [fill] and wind barbs at hour 54 for (a) present Miller-A, (b) future Miller-A, (c) present Miller-B, and (d) future Miller-B cyclones.
Figure 3.27: Maximum 10-m wind speed (m/s) at each grid point over all times for all (a) present Miller-A, (b) future Miller-A, (c) present Miller-B, and (d) future Miller-B simulations.
Figure 3.28: Change in maximum 10-m wind speed (m/s) at each grid point over all times for all (a) Miller-A and (b) Miller-B simulations.
Figure 3.29: Histograms of 10-m wind speed (m/s) exceeding 0.0 m/s within the area bound from 25° to 49°N and 90° to 45°W over all Miller-A and Miller-B simulations. Data for the (a) present, (b) future and (c) difference histograms of Miller-A simulations are output from 00 to 63 hours. Data for the (d) present, (e) future and (f) difference histograms of Miller-B simulations are output from 00 to 69 hours. Wind speed bins are every 1 m/s.
Figure 3.30: Histogram difference (future minus present) of 10-m wind speed (m/s) exceeding 0.0 m/s over land within the area bound from 25° to 49°N and 90° to 45°W over all times for all (a) Miller-A and (b) Miller-B simulations. Miller-A values are output from hours 00 to 63. Miller-B values are output from hours 00 to 69.
Figure 3.31: Histogram difference (future minus present) of 10-m wind speed (m/s) exceeding 0.0 m/s over water within the area bound from 25° to 49°N and 90° to 45°W over all times for all (a) Miller-A and (b) Miller-B simulations. Miller-A values are output from hours 00 to 63. Miller-B values are output from hours 00 to 69.
Figure 3.32: Composite 850-mb wind speed exceeding 25 m/s [contour] (m/s) and specific humidity [fill] (kg water vapor/kg moist air) at hour 57 for (a) present Miller-A, (b) future Miller-A, (c) present Miller-B, and (d) future Miller-B cyclones.
Figure 3.33: Composite change in meridional wind component [contours] (m/s) and statistical significance [color fill] at hour 57 for (a) Miller-A and (c) Miller-B cyclones. Wind vectors for present [blue] and future [red] (b) Miller-A and (d) Miller-B composites are also shown. A 20-point Gaussian weighted smoothing function is applied to the fields prior to computing the difference and statistical significance. Statistical significance is plotted for the 90, 95 and 99% confidence levels using a 2-sample, 2-tailed Student’s t-test for equal sample sizes and variances.
Figure 3.34: Histograms of 850-mb wind speed (m/s) greater than or equal to 25 m/s within the area bound from 25° to 49°N and 90° to 45°W for all Miller-A and Miller-B simulations. Data for the (a) present, (b) future and (c) difference histograms of Miller-A simulations are output from 00 to 63 hours. Data for the (d) present, (e) future and (f) difference histograms of Miller-B simulations are output from 00 to 69 hours. Wind speed bins are every 1 m/s.
Figure 3.35: Vertical cross section of temperature change (K) and change in meridional temperature gradient (10^-6 K/m) at hour 48 for the (a, b) Miller-A and (c, d) Miller-B composites. Cross section is computed from along the 90°W meridian from 25° to 45°N.
Figure 3.36: Change in the 250-mb composite zonal wind component [fill] (m/s) at hour 48 for the (a) Miller-A, and (b) Miller-B composite. 250-mb geopotential height (dam) contours are also plotted for context.
Figure 3.37: Composite change in 250-mb zonal wind component [contours] (m/s) and statistical significance [color fill] at hour 48 for (a) Miller-A and (b) Miller-B cyclones. A 20-point Gaussian weighted smoothing function is applied to the fields prior to computing the difference and statistical significance. Statistical significance is plotted for the 90, 95 and 99% confidence levels using a 2-sample, 2-tailed Student’s t-test for equal sample sizes and variances.
Figure 3.38: Histograms of 250-mb wind speed (m/s) within the area bound from 20° to 49°N and 120° to 45°W for all Miller-A and Miller-B simulations. Data for the (a) present, (b) future and (c) difference histograms of Miller-A simulations are output from 00 to 63 hours. Data for the (d) present, (e) future and (f) difference histograms of Miller-B simulations are output from 00 to 69 hours. Wind speed bins are every 1 m/s.
Figure 3.39: Change in composite 250-mb wind speed (m/s) [contour] and divergence ($10^6 \text{s}^{-1}$) [fill] at hour 48 for (a) present Miller-A, (b) future Miller-A, (c) present Miller-B, and (d) future Miller-B cyclones.
Figure 3.40: Composite pressure (mb) [fill] and wind (kts) [barbs] on the 1.5 PVU (1 PVU = $10^{-6}$ m$^2$ s$^{-1}$ kg$^{-1}$K) surface at hour 48 for (a) present Miller-A, (b) future Miller-A, (c) present Miller-B, and (d) future Miller-B cyclones.
Figure 3.41: Change in composite pressure (mb) on the 1.5 PVU (1 PVU = $10^6 \text{ m}^2 \text{s}^{-1} \text{kg}^{-1} \text{K}$) surface at hour 48 for (a) Miller-A and (b) Miller-B cyclones.
Figure 3.42: Composite pressure (mb) [fill] and wind (kts) [barbs] on the 1.5 PVU (1 PVU = $10^{-6}$ m$^2$ s$^{-1}$ kg$^{-1}$ K) surface at hour 57 for (a) present Miller-A, (b) future Miller-A, (c) present Miller-B, and (d) future Miller-B cyclones.
Figure 3.43: Time-averaged mean Ertel PV anomaly profiles, expressed in units of PVUs (1 PVU = 10^{-6} m^2 s^{-1} kg^{-1} K), for Miller-A cyclones. Time-averaged profiles are computed for (a) hours 18-27, (b) hours 30-39, (c) hours 42-51, and (d) hours 54-63. Present (future) profiles are represented by blue (red) lines.
Figure 3.44: Time-averaged mean Ertel PV anomaly profiles, expressed in units of PVUs (1 PVU = 10^{-6} \text{ m}^2 \text{s}^{-1} \text{ kg}^{-1} \text{K}), for Miller-B cyclones. Time-averaged profiles are computed for (a) hours 18-24, (b) hours 27-33, (c) hours 36-42, (d) hours 45-51, (e) hours 54-60, and (f) hours 63-69. Present (future) profiles are represented by blue (red) lines.
Figure 3.45: Time-averaged mean latent heating rate profiles, expressed in units of K/day, for Miller-A cyclones. Time-averaged profiles are computed for (a) hours 18-27, (b) hours 30-39, (c) hours 42-51, and (d) hours 54-63. Present (future) profiles are represented by blue (red) lines.
Figure 3.46: Time-averaged mean latent heating rate profiles, expressed in units of K/day, for Miller-B cyclones. Time-averaged profiles are computed for (a) hours 18-24, (b) hours 27-33, (c) hours 36-42, (d) hours 45-51, (e) hours 54-60, and (f) hours 63-69. Present (future) profiles are represented by blue (red) lines.
Figure 3.47: Time-averaged mean non-advective PV tendency profiles, expressed in units of PVU/hr, for Miller-A cyclones. Time-averaged profiles are computed for (a) hours 18-27, (b) hours 30-39, (c) hours 42-51, and (d) hours 54-63. Present (future) profiles are represented by blue (red) lines.
Figure 3.48: Time-averaged mean non-advective PV tendency profiles, expressed in units of PVU/hr, for Miller-B cyclones. Time-averaged profiles are computed for (a) hours 18-24, (b) hours 27-33, (c) hours 36-42, (d) hours 45-51, (e) hours 54-60, and (f) hours 63-69. Present (future) profiles are represented by blue (red) lines.
Figure 3.49: Time-averaged mean 2-m potential temperature anomalies, expressed in units of K, for Miller-A cyclones. Anomalies are computed for (a) hours 18-27, (b) hours 30-39, (c) hours 42-51, and (d) hours 54-63. Present (future) profiles are represented by blue (red) lines.
Figure 3.50: Time-averaged mean 2-m potential temperature anomalies, expressed in units of K, for Miller-B cyclones. Anomalies are computed for (a) hours 18-24, (b) hours 27-33, (c) hours 36-42, (d) hours 45-51, (e) hours 54-60, and (f) hours 63-69. Present (future) profiles are represented by blue (red) lines.
Figure 3.51: Change in time-averaged mean 2-m potential temperature anomalies, expressed in units of K, for (a) Miller-A and (b) Miller-B cyclones. Here the change is calculated as the future minus the current anomaly. Changes in the Miller-A anomalies are computed for hours 18-27, 30-39, 42-51 and 54-63. Changes in the Miller-B anomalies are computed for hours 18-24, 27-33, 36-42, 45-51, 54-60, and 63-69.
4. Sensitivity Studies and Additional Experiments

Chapter 4 presents a series of experiments that investigate the sensitivity of results to model physics, applied thermodynamic changes, domain size, horizontal resolution and compositing technique. In particular, the sensitivity of cyclone track and intensity will be highlighted here. Miller-A Case 4 (January 3-6, 1994) is used for most of the following studies. This case was chosen because it showed the largest increase in intensity with warming where minimum sea-level pressure is used as the metric for intensity. Since this case was particularly sensitive to climate warming, it is of interest to determine whether changing different aspects of the original experiment yields significantly different results. As noted in Chapter 3, some cases exhibited strengthening with warming while other did not. By testing only one case here, and one that was atypical, we are assuming some level of similarity in terms of the sensitivity to each of the factors mentioned above.

First, the sensitivity of cyclone track and intensity to model physics will be assessed by performing an ensemble of simulations using different model configurations. Next, a similar experiment in which the thermodynamic changes from individual GCMs are used instead of the GCM average is conducted. The sensitivity to domain size is then investigated by initializing a run with a significantly larger domain and global reanalysis dataset. Following this, the sensitivity of cyclone intensity to horizontal resolution is addressed by running simulations with high resolution inner nests. Finally, we investigate simulations in which the NARR composite is used as the initial and lateral boundary conditions. With any study of this nature, a number of choices must be made in terms of the experimental design.
The goal of each of these experiments is to quantify the sensitivity of our results to some of the more significant choices we made.

4.1 Physics Ensemble Study

The sensitivity of results to model physics is addressed by performing a physics ensemble study. Six different configurations are used in which the surface layer scheme, planetary boundary layer (PBL) scheme, cumulus parameterization (CP scheme) and microphysics scheme vary. The variations in these configurations can be seen in Table 4.1. The EMB5 configuration represents the control setup used for the simulations analyzed in Chapter 3 and described in Section 2.2. The MYK5 and QQK5 configurations are the two reference configurations recommended by the Developmental Testbed Center (DTC) for WRF Version 3.2.1. Configuration names ending with a six are the same as those ending with a five except that the WRF Single-Moment 6-Class (WSM6) microphysics scheme is used instead. All other settings used for the control simulations in Chapter 3 remain unchanged.

The minimum sea-level pressure evolutions of each of the present and future physics ensemble members are presented in Figures 4.1a and 4.1b, respectively. For each of the three base configurations (EMB, MYK, QQK) there is very little difference in the final minimum sea-level pressure when the microphysics scheme is changed from WSM5 to WSM6. Therefore, the previously presented results do not appear to be particularly sensitive to our microphysics choice. Larger differences in the final intensity are found when comparing the base configurations themselves (i.e. EMB vs. MYK, etc.). Looking again at Figure 4.1 we
see that simulations using the EMB and MYK configurations produce very similar storm intensities (within ~1 mb) by hour 72. This is true for both the present and future ensemble. On the other hand, simulations using the QQK configuration produce surface cyclone intensities that are clearly stronger than those using the other two configurations. In fact, this difference is actually larger in the future ensemble than it is in the present ensemble. These results suggest that these storms may be more sensitive to PBL and cumulus parameterizations than the microphysics parameterizations. It is also worth noting that the configuration used for the analysis shown in Chapter 3 (EMB5) is found to be the weakest member in both the present and future ensembles. This suggests that the configuration used in the main study is a fairly conservative one. The minimum sea-level pressure evolutions of the present and future physics ensemble means are shown in Figure 4.2. The one standard deviation range for each physics ensemble, represented by error bars, is much less than those seen for the present and future composites in Figures 3.10 and 3.12. This suggests that variability due to model physics, at least for this case, is smaller than the case-to-case variability seen in the composites.

Figure 4.3 presents the tracks of each present and future physics ensemble member. For both the present (Figure 4.3a) and future (Figure 4.3b) ensembles, the surface cyclones take very similar tracks and are located in nearly the same position by hour 72. There is some disagreement in the present physics ensemble on how the surface cyclone tracks through the southeastern United States but the cyclones still end up at practically the same location. The ensemble mean tracks shown in Figure 4.4 show a shift of the future ensemble mean to the northeast relative to the present. This is consistent with what was seen in the composite
analysis. In terms of the variability of the tracks, the one standard deviation ranges are much smaller than the ranges seen in the composite analysis. Like the physics ensemble analysis of intensity, this suggests that the variability in track due to physics for this case is less than the case-to-case variability in the composite.

4.2 GCM Ensemble Study

An approach similar to the previously discussed physics ensemble study is used here to investigate the sensitivity of our results to the thermodynamic changes applied to the initial conditions. As mentioned in Section 2.4, thermodynamic changes derived from an ensemble of 5 GCMs were applied to the initial and lateral boundary conditions of each case and then simulated to produce future cyclones. While the ensemble mean thermodynamic changes were used in the main experiment, the changes from each of the individual GCMs are applied to the initial and lateral boundary conditions and then simulated here. Doing so will divulge the variability due to the GCM changes themselves and provide insight on the sensitivity of our results to those changes. Each simulation is referred to by the institute at which the GCM was run. Those institutes are the Bjerknes Centre from Climate Research (BCCR), Centre National de Recherches Météorologiques (CNRM), Institute of Numerical Mathematics (INM), Max Planck Institute (MPI) and UK Met Office (UKMO). The specific GCMs can be found in Section 2.4.

The minimum sea-level pressure evolution of the GCM ensemble mean and individual GCM ensemble members are presented in Figure 4.5. The GCM ensemble mean (Figure 4.5a) is plotted with the one standard deviation range as error bars. Overall, the
individual GCM ensemble members all fall within ~5 mb of one another with the CNRM simulation producing the most intense cyclone (Figure 4.5b). Figure 4.6 puts the variability of both the GCM and physics ensembles into context. The GCM and future physics ensemble mean minimum sea-level pressures are plotted together with their respective one standard deviation ranges. The minimum sea-level pressure evolution of the control set up (EMB5) used in the main experiment is also shown for context. Given that the average changes derived from the 5-member GCM ensemble are applied to EMB5’s initial and lateral boundary conditions, it is no surprise that it almost exactly follows the GCM ensemble mean here. The GCM and future physics ensembles clearly overlap thus producing a net one standard deviation range (i.e. net range covered by error bars) of ~7-8 mb. Given that the one standard deviation ranges of the future composites are ~17-18 mb (Figures 3.10, 3.12), the combined variability due to physics and the applied thermodynamic changes appears to be less than the future case-to-case variability.

There is a bit more variability in the tracks of the individual GCM ensemble members (Figure 4.7) compared to the individual future physics ensemble members (Figure 4.3b). This is represented more clearly in the one standard deviation ranges being slightly greater in the GCM ensemble (Figure 4.8). That being said, the overall paths the individual members take are fairly similar as the surface cyclones track along the eastern United States coastline and then up into Nova Scotia. Once again, the variability here is much less than that projected for the mean future Miller-A and Miller-B tracks meaning that the case-to-case variability within the composites is greater than the sensitivity to which GCM changes we apply.
4.3 Sensitivity to Domain Size

Present and future simulations of this case are also conducted in which an expanded domain and a global reanalysis dataset are used with the purpose of examining the sensitivity of our results to domain size. We are concerned with the sensitivity to domain size because the model solutions are locked together at the downstream outermost domain boundary. This could be artificially increasing the similarity between present and future simulations, especially as they approach these boundaries late in the storms’ life cycles. The Climate Forecast System Reanalysis (CFSR), a global reanalysis dataset with horizontal grid spacing of ~55 kilometers and 37 vertical levels, is used to initialize the simulations in this study (Saha and Coauthors 2010). Figure 4.9 shows the expanded WRF domain used for the CFSR simulations of this event. The domain has been expanded northward and eastward in order to allow the storm to fully evolve without contamination from the lateral boundaries. It also captures somewhat less of the Pacific Ocean compared to the domain used for the NARR simulations (Figure 2.1). The CFSR simulations are analyzed for all 84 hours since they fall well within the domain throughout their evolution.

The minimum sea-level pressure evolutions of the original present and future NARR simulations of this event are plotted along with those of the present and future CFSR simulations in Figure 4.10. At first glance it appears that the evolutions are quite a bit different with the present and future NARR simulations producing cyclones that attain lower minimum sea-level pressures than their CFSR counterparts by hour 72. The cyclones in the CFSR simulations do not attain minimum sea-level pressures as low as those seen in the NARR simulations until the end of their 84 hour simulations. This is a bit surprising seeing
as these simulations are initialized with two reanalyses of largely the same data. One potential explanation for this discrepancy is that they are attributable to the differences in horizontal and vertical resolution between the two reanalyses. Although the differences between the CFSR and NARR simulations are a bit concerning, it is more important to determine how the projected present to future changes compare with one another. The minimum sea-level pressure changes (future minus current) for the CFSR and NARR simulations, presented in Figure 4.11, are actually remarkably similar. In fact, the minimum sea-level pressure changes are nearly identical at 72 hours. This lends confidence to the belief that our intensity results are not particularly sensitive to the domain being limited along its northern and eastern boundaries.

Figure 4.12 presents the present and future tracks for both the CFSR and NARR simulations. Both of the present tracks bring the surface cyclone up into the Canadian Maritimes with the CFSR track extending further northeast as you would expect given that it is analyzed for an additional 12 hours. However, the tracks do diverge in the southeastern United States within the same region of disagreement seen in the present physics ensemble. The future CFSR and NARR tracks, on the other hand, both track northeastward along the Atlantic coastline and up into the Canadian Maritimes. At hour 72 the present to future track differences for the CFSR and NARR simulations are ~188 km and ~104 km, respectively. It is interesting to note that the present CFSR simulation ends up at roughly the same longitude and slightly north of the future simulation. However, the future CFSR cyclone tracks east of the present CFSR simulation for most of the first 72 hours which is consistent with what is seen in the simulations initialized with NARR in Chapter 3.
4.4 Sensitivity to Resolution

Previous work by Willison et al. (2013) has shown that simulated extratropical cyclone intensity is particularly sensitive to horizontal resolution. They found that under resolving mesoscale latent heating results in a weakened diabatic feedback and leads to weaker cyclones. To assess the sensitivity of our results to horizontal resolution, additional simulations are performed here with two-way inner nests. The outermost domain has a horizontal grid spacing of 36 kilometers while the two inner nests have horizontal grid spacings of 12 and 4 kilometers. Here the NARR is again used as the initial and lateral boundary conditions. Figure 4.13 shows the area encompassed by the outermost domain and two inner nests. For this study, surface cyclone intensity is only assessed at times when the cyclone remains well within all three domains (hours 12-72).

The present and future minimum sea-level pressure evolutions within each of the domains are presented in Figure 4.14. The general evolutions of the 36 (Figure 4.14a), 12 (Figure 4.14b) and 4 (Figure 4.14c) kilometer nests are very similar with the future cyclones ultimately attaining lower minimum sea-level pressures by 72 hours. Consistent with Willison et al. (2013), the final minimum sea-level pressures achieved in the 12 and 4 kilometer nests are lower than those seen in the 36 kilometer nest. This is true for both the present and future simulations despite the fact that the storms have not yet reached peak intensity (i.e. still deepening). The changes in cyclone intensity are also greater in the 12 and 4 kilometer nests (Figure 4.15). Both the 12 and 4 kilometer nests have minimum sea-level pressures that are ~1.5 mb lower than that projected in the outermost domain.
Overall, these results suggest that our choice of 36 kilometer grid spacing is underestimating the future increases in intensity by ~14.8%. It also is interesting to note that while there is an increase in intensity when going from 36 to 12 kilometer horizontal grid spacing there is very little change when going from 12 to 4 kilometers. This is likely due to the fact the convective parameterization (CP) scheme is not particularly active for this event. As such, going from 12 kilometers, where the CP scheme is on, to 4 km, where convection is explicitly resolved, does not result in as much of a change.

4.5 Sensitivity to Compositing Technique

Finally, the sensitivity of our results to compositing technique is assessed by initializing present and future simulations with cyclone composites instead of generating present and future cyclone composites made up of individual simulations. As noted in Section 2.3, NARR composites consisting of the same Miller-A and Miller-B events are used to initialize these simulations. By initializing WRF with composites, we are essentially generating pseudo-idealized simulations of Miller-A and Miller-B cyclone events. The evolutions of these pseudo-idealized Miller-A and Miller-B cyclones are presented here for hours 15-69. These represent the times that continuous tracks (i.e. tracking code doesn’t unnaturally jump for the subsequent 12 hours) exist and remain well within the domain for all present and future simulations.

Minimum sea-level pressure evolutions of Miller-A and Miller-B cyclone events using NARR composites as initial conditions are presented in Figure 4.16. The results found using this technique are somewhat different from those found in Chapter 3. Here both the
Miller-A (Figure 4.16a) and Miller-B (Figure 4.16b) cyclones weaken in the future whereas the results in Chapter 3 suggest that future cyclones may slightly increase in intensity. However, the cyclone track changes in this experiment are similar to those found in Chapter 3. Both the future Miller-A (Figure 4.17a) and Miller-B (Figure 4.17b) cyclones take a more eastward track relative to the present. This begs the question: why do these composite initialized simulations produce different intensity results but similar track results?

Referring back to the results from Chapter 3, not all of the future simulations exhibited lower minimum sea-level pressures in the future. While most future simulations had cyclones with minimum sea-level pressures that were lower, four of the twenty cases did not (Figures 3.11 & 3.13). On the other hand, eastward track shifts were evident in each of the twenty future simulations (Figure 3.15 & 3.17). Given these findings, it is not inconceivable to have a future simulation in which the storm takes a more eastward track but is also weaker. A likely explanation for this result is that the composite initial conditions are too smooth to allow for rapid cyclogenesis. Comparing the minimum sea-level pressure evolutions of Miller-A and Miller-B simulations using this technique (Figure 4.16) with the average evolutions of Miller-A and Miller-B simulations from chapter 3 (Figures 3.10 & 3.12) it is apparent that this technique spawns much weaker cyclones. The present simulations using NARR composites as initial conditions produce cyclones with minimum sea-level pressures just below 1000 mb while the average present cyclones in chapter 3 attain minimum sea-level pressures around 980 mb. Similar differences are seen between the future cyclones of these two techniques.
Although this technique has been shown to be useful for other types of weather phenomena (Mahoney et. al 2012), it was unable produce surface cyclones of at least moderate intensity here. This result may due to the fact that we built our composite around the exact location of the surface cyclone instead of the incipient upstream vorticity maximum aloft. Had we built the composite around the upper-level vorticity maximum, there would be less smearing of the forcing aloft and a stronger surface cyclone would have likely developed. Additional analysis is necessary to determine the exact reasons why intense storms were not spawned using the NARR composite as initial conditions and is left for future work.
Table 4.1: Configurations for each of the six physics ensemble members: EMB5, EMB6, MYK5, MYK6, QQK5 and QQK6. The EMB5 configuration uses the ETA similarity surface layer scheme, Mellor-Yamada-Janic PBL scheme, Betts-Miller-Janjic CP scheme and WRF Single-Moment 5-Class (WSM5) microphysics scheme. The MYK5 configuration uses the MM5 surface layer scheme, Yonsei University (YSU) PBL scheme, Kain-Fritsch (KF) CP scheme and WSM5 microphysics scheme. The EMB5 configuration uses the Quasi-Normal Scale Elimination (QNSE) surface layer scheme, QNSE PBL scheme, KF CP scheme and WSM5 microphysics scheme. The EMB6/MYK6/QQK6 configurations are the same as the EMB5/MYK5/QQK5 configurations except that the WRF Single-Moment 6-Class (WSM6) microphysics scheme is used instead of WSM5.

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Figure 4.1: Minimum sea-level pressure (mb) evolution of each (a) present and (b) future physics ensemble member for the January 3-6, 1994 event. Line colors designating each ensemble member are shown in the legends above.
Figure 4.2: Average minimum sea-level pressure (mb) evolution of the present [blue] and future [red] physics ensemble members for the January 3-6, 1994 event. Error bars denote the one standard deviation range.
Figure 4.3: Surface cyclone track of each (a) present and (b) future physics ensemble member for the January 3-6, 1994 event. Line colors designating each ensemble member are shown in the legends above.
Figure 4.4: Ensemble mean surface cyclone track of the (a) present and (b) future physics ensemble for the January 3-6, 1994 event. The one standard deviation (i.e. 1-sigma) cone for the present and future ensemble mean track is shown with blue dashed and red dashed lines, respectively.
Figure 4.5: Minimum sea-level pressure (mb) evolution of the (a) GCM ensemble mean, and (b) individual GCM ensemble members for the January 3-6, 1994 event. Error bars in (a) denote the one standard deviation range. Line colors designating each ensemble member in (b) are shown in the legend above.
Figure 4.6: Minimum sea-level pressure (mb) evolution of the GCM ensemble mean [black line], physics ensemble mean [green line] and control (EMB5) simulation [pink line] for the January 3-6, 1994 event. Error bars in denote the one standard deviation range for each ensemble.
Figure 4.7: Surface cyclone track of each GCM ensemble member for the January 3-6, 1994 event. Line colors designating each ensemble member are shown in the legend above.
Figure 4.8: Ensemble mean surface cyclone track of the GCM ensemble for the January 3-6, 1994 event. The one standard deviation (i.e. 1-sigma) cone for ensemble mean track is shown with red dashed lines.
Figure 4.9: Domain for simulations using Climate Forecast System Reanalysis (CFSR) data as initial and lateral boundary conditions.
Figure 4.10: Minimum sea-level pressure (mb) evolutions for present [blue lines] and future [red lines] simulations of the January 3-6, 1994 event. Evolutions are shown for simulations initialized using the CFSR [solid lines] and NARR [dashed lines].
Figure 4.11: Minimum sea-level pressure difference (mb) between the present and future (a) CFSR and (b) NARR simulations of the January 3-6, 1994 event. Difference is computed as the future minimum sea-level pressure minus the present minimum sea-level pressure.
Figure 4.12: Surface cyclone tracks for the (a) present and (b) future CFSR [black] and NARR [green] simulations of the January 3-6, 1994 event.
Figure 4.13: Domains used for the horizontal resolution sensitivity study. The outermost domain has a horizontal grid spacing of 36 kilometers. The first inner nest (second largest domain) has a horizontal grid spacing of 12 kilometers while the innermost nest has a horizontal grid spacing of 4 kilometers.
Figure 4.14: Minimum sea-level pressure (mb) evolutions for present [blue line] and future [red line] simulations of the January 3-6, 1994 event. Evolutions are shown for simulations using (a) 36, (b) 12 and (c) 4 kilometer horizontal grid spacing.
Figure 4.15: Minimum sea-level pressure difference (mb) between the present and future simulations of the January 3-6, 1994 event using (a) 36, (b) 12 and (c) 4 kilometer horizontal grid spacing. Difference is computed as the future minimum sea-level pressure minus the present minimum sea-level pressure.
Figure 4.16: Minimum sea-level pressure (mb) evolutions for present [blue line] and future [red line] (a) Miller-A and (b) Miller-B simulations. Evolutions are shown for simulations initialized using NARR composites.
Figure 4.17: Surface cyclone tracks for present [blue line] and future [red line] (a) Miller-A and (b) Miller-B simulations. Evolutions are shown for simulations initialized using NARR composites.
5. Summary and Conclusions

Previous studies investigating climate warming impacts on extratropical cyclones have focused on a broad spectrum of topics. One area of particular interest, and uncertainty, in recent decades is the impact of warming on extratropical cyclone intensity. Past studies have primarily made use of GCMs to diagnose hemispheric and regional changes in cyclone intensity. Most of these changes have been attributed to changes in environmental baroclinicity with latent heating also hypothesized to play some role (Geng and Sugi 2003, Lambert and Fyfe 2006, Watterson 2006, Teng et al. 2008, Bengtsson et al. 2009, Catto et al. 2011, Mizuta et al. 2011, Colle et al. 2013). Given their fairly coarse horizontal resolution, most ~100 km or more, GCMs typically do not lend themselves to diagnosing changes at a storm scale level. Champion et al. (2011) performed GCM simulations with horizontal grid spacings of 40 kilometers however their work was largely focused on the frequency of weather extremes. As such, very few studies outside of Perrie et al. (2010) have actually investigated the impacts of warming on the storm scale dynamics of extratropical cyclones to our knowledge.

The current study uses WRF to perform simulations of wintertime Miller-A and Miller-B cyclone events along the Eastern United States in present and future thermodynamic environments. The horizontal grid spacing used in these simulations allow for the diagnosis of changes in moist diabatic processes that may be unresolved by GCMs. Cyclone composites are generated from these simulations in order to assess the impacts of warming on the synoptic evolution of these storm systems. An alternate approach in which composites are used as initial and lateral boundary conditions is also used. The potential vorticity
framework is used extensively in this work to analyze changes in the storm scale dynamics due to warming.

The average Miller-A and Miller-B cyclone event is found to increase in intensity with warming where minimum sea-level pressure is used as the metric for intensity. Minimum sea-level pressure decreases of ~3.5 mb and ~1 mb are projected for the average Miller-A and Miller-B cyclone, respectively, with warming. Both types of cyclone event also exhibit enhanced deepening rates on average as the future systems start out with higher and end up with lower central pressures (see Figures 3.10 & 3.12). The increases in intensity are found to be statistically insignificant despite the fact that nine out of ten Miller-A and seven out of ten Miller-B events exhibited lower central pressures in the future. Eastward track shifts are evident with warming for both Miller-A and Miller-B cyclones as well. Although this is also found to be statistically insignificant, all twenty cases exhibited an eastward track shift which implies more confidence in this result. The statistical significances of these results are likely limited by the small sample size.

Consistent with previous studies, the amount of precipitation associated with these cyclone events is projected to increase with warming. Atmospheric water vapor content is projected to increase ~23% for Miller-A events with a ~14% increase in area-average precipitation and ~26% increase in maximum precipitation. A similar water vapor increase of ~24% is projected for Miller-B events with area-average and maximum precipitation increases of ~8.5% and ~18%, respectively. While these increases are not statistically significant at any one location, the overall trend is toward decreased light precipitation and increased heavy precipitation associated with Miller-A and Miller-B cyclone events. This is
consistent with the results of Allen and Ingram (2002). Snowfall associated with these events is generally projected to decrease as the future Miller-A and Miller-B composites exhibit area-average snowfall decreases of \(~28\%\) and \(29\%\), respectively. However snowfall is actually projected to increase in regions of elevated terrain, which means that increases in total precipitation may not be restricted to rainfall if conditions permit.

The thermodynamic changes imposed on the initial and lateral boundary conditions of our simulations result in a strengthening of the upper tropospheric meridional temperature gradient. Consistent with thermal wind balance, a compensating increase in upper tropospheric westerly winds occurs. The increases in upper tropospheric westerlies cause upper-level troughs to progress eastward more quickly and provide a physical explanation for the eastward track shift exhibited in each future simulation. This result in consistent with our theoretical understanding of Rossby wave dynamics and represents one of the principal findings of the current study. The LLJ is also found to increase in both composites.

As noted above, changes in the maximum Eady growth rate have generally been used to diagnose changes in future extratropical cyclone intensity. Here we find that the area averaged maximum Eady growth rate associated with Miller-A and Miller-B events in this region increases 2-3% prior to cyclogenesis onset. On the other hand, area averaged coupling potential decreases 5-6%. This represents a decrease in the mean environmental Rossby penetration depth prior to cyclogenesis. Given this information, the coupling of UPV and LPV anomalies may become more difficult (i.e. require stronger anomalies) in the future. The UPV anomalies associated with the average future Miller-A and Miller-B systems extend closer to the surface during the latter stages of their evolution. This allows more
stratospheric high PV air to extend downwards and aid the cyclogenesis process. It is unclear what mechanisms are responsible for this change and thus requires further investigation. Low-level warm (i.e. +LPV) anomalies in the average Miller-A and Miller-B cyclone also exhibit future strengthening throughout their life cycle. The future low-level warm anomaly strength for the average Miller-B event is more than twice that found for the average Miller-A event. By the end of its evolution, the average Miller-A cyclone exhibits a stronger diabatic PV maximum at low-levels with warming while the average Miller-B diabatic PV maximum actually weakens.

The series of sensitivity studies performed in Chapter 4 suggest that the model configuration used for this study, specifically the choice of model physics and horizontal grid spacing, may have resulted in overly conservative results. The default physics configuration (EMB5) used in the main experiment produced the weakest surface cyclone of the six configurations studied. Likewise, our choice of 36 kilometer horizontal grid spacing resulted in present to future intensity increases that were ~14.8% smaller than those in 12 and 4 kilometer simulations. Our choice of applied thermodynamic changes and domain size is not found to have a significant impact on the results. Finally, simulations in which NARR composites are used as initial and lateral boundary conditions produce qualitatively similar track changes but vastly different intensity changes. This is largely due to how we constructed the composite (i.e. surface based).

The first science question this study sought to answer was whether or not changes in the maximum Eady growth rate could explain the changes we find in cyclone intensity. Increases in the maximum Eady growth rate were found in this study but the average Miller-
B event exhibited larger increases with warming than the average Miller-A event. Given that the average Miller-A event exhibited a greater increase in intensity with warming, changes in intensity clearly cannot be explained by changes in the maximum Eady growth rate alone. It should be noted that we have only accessed the impacts of the environmental (area-averaged) maximum Eady growth rate here. Another factor of importance is the amount of time a system remains within the peak baroclinic zone (i.e. region where maximum Eady growth rate is largest). Given the increases in propagation speed, a storm may actually remain in the peak zone for a shorter time resulting in intensity changes that are less than expected. This aspect requires further investigation and is left for future work.

The second science question this study sought to answer was whether or not changes in cyclone intensity could be attributed to increases in condensational heating (e.g. diabatic PV generation). Both the Miller-A and Miller-B composites exhibit a strengthening of the UPV and LPV anomalies with warming. The discrepancy in intensity increase with warming between the two composites is ultimately tied to the difference in the low-level diabatic PV maxima. The enhanced diabatic PV anomaly seen in the average future Miller-A cyclone appears to augment the LPV anomaly thus allowing for stronger cross advection between the upper and lower levels. The opposite is true for Miller-B cyclones as the weakened diabatic PV anomaly partially offsets the LPV anomaly and weakens this cross advection. The timing of the intensity “flip” from weaker to stronger storms seen in each composite is also related to the PV structure. Since the future UPV anomalies move out over the Gulf Stream (region with low static stability/large Rossby depth) earlier in their evolution, future cyclones are able to couple earlier on in their evolutions. It is evident that changes in intensity can at least
be partly attributed to enhanced condensational heating/diabatic PV generation but several problems remain. These unresolved problems include determining what distinguishes storms that strengthen from those that weaken and why the diabatic PV maximum weakens for Miller-B events.

The third and final science question we sought to answer was whether or not the changes due to warming would be qualitatively similar between the WRF composites and composite initialized simulations. Similar track changes with warming are observed between the two techniques but the intensity results are quite different. This is primarily due to the fact that our composite was built around a surface feature instead of the upper forcing. As a result, the upper forcing was smeared heavily and not strong enough to produce an intense surface cyclone. Additional work is necessary to determine the true viability of this technique for cyclones.

The work presented here paves the way for several paths of future work. While we have identified some of the storm scale changes that may occur in these events with warming, more work must be done. In particular, we need to determine why some events increase in intensity while others weaken. This will necessitate doing analysis on individual cyclone events as opposed to the composite analysis presented here. Likewise, a larger subset of cases is necessary to truly determine whether there is any statistically significant change in Miller-A or Miller-B storm tracks or intensity. Analyzing present and future seasonal simulations within the Atlantic basin is one way this may be remedied. The question of why the diabatic PV maximum weakened for Miller-B cyclones also remains unresolved. Computing a full Ertel PV budget of the Miller-B composites would likely yield significant
insight into this problem. It would also be worthwhile to compare the results presented here to a similar study in which thermodynamic changes from the more recent CMIP5 models are used.

As previously noted, one of the primary weaknesses of this work is the statistical significance testing that is implemented. Given our small sample of cases, a 2-sample Student’s t-test is likely not best suited here. The fact that we did not use a storm relative composite also makes it especially difficult to extract a statistically significant signal. Variability in the tracks and propagation speeds of these events result in an inflation of the variance. Going forward, there are two ways in which we plan on addressing this weakness. The first way is to generate storm relative composites that alleviate some of the inherent smearing issues in the compositing approach used here. The second way is to develop a metric that can test for statistical significance while accounting for the limited sample size used here. This will likely necessitate collaborative work with the statistics department in the near future.

Finally, a brief caveat about the present study is necessary. The goal of this study was to determine how the synoptic evolutions of Miller-A and Miller-B cyclone events may change in a warmer world. As such, it should be kept in mind that the results presented here are not representative of what may happen to all extratropical cyclones with climate warming.


Geng, Q., and M. Sugi, 2003: Possible change of extratropical cyclone activity due to enhanced greenhouse gases and sulfate aerosols—study with a high-resolution AGCM. J. Climate, 16, 2262–2274.


