UNDERSTANDING EXTREME TORNADO EVENTS UNDER FUTURE CLIMATE CHANGE THROUGH THE PSEUDO-GLOBAL WARMING METHODOLOGY

BY

MATTHEW J. WOODS

THESIS

Submitted in partial fulfillment of the requirements for the degree of Master of Science in Atmospheric Sciences in the Graduate College of the University of Illinois Urbana-Champaign, 2021

Urbana, Illinois

Master’s Committee:

Professor Robert J. Trapp, Advisor
Professor Sonia Lasher-Trapp
Professor Ryan Sriver
ABSTRACT

This study is a first step in determining how extreme tornado events may change under future anthropogenic climate change using the “pseudo global warming” (PGW) methodology. To properly apply this methodology, current-day extreme tornado events must be adequately simulated (the control; CTRL), and then simulated again with climate-change differences, or “deltas”, applied to the original 3D meteorological forcing. These climate change deltas represent differences in future and historical general circulation model projections from the coupled model intercomparison project-phase 5 (CMIP5). Several delta construction methods were employed to provide a larger ensemble and assess the importance of delta configuration.

Given its classic supercellular nature, and extremely strong tornado (EF-5) with great societal impact, the 20 May 2013 tornado event in Moore, Oklahoma was one event chosen for application of the PGW methodology. Using high-resolution CTRL and PGW simulations using the Weather Research and Forecasting model (WRF), it was found that the convective storms under PGW are more numerous, but updraft velocities do not scale directly with future increases in convective available potential energy (CAPE). It was also found that instances of a baseline tornado proxy decreased in the PGW simulations, while instances of a high-end tornado proxy increased, indicating fewer but stronger tornadoes under PGW. The convective response also showed some sensitivity to the configuration of the climate change deltas, especially at upper thresholds of the parameters assessed.

Separate WRF simulations of the 10 February 2013 tornado outbreak, which included an EF-4 tornado in Hattiesburg, Mississippi, were also conducted. Here, uniform sea-surface temperature (SST) perturbations were applied to the Gulf of Mexico to investigate the influence
of SSTs on cool-season tornado events. Warmer SSTs were found to produce more favorable thermodynamic environments that manifested in increased convective coverage and intensity. Additionally, tornado proxy was more (less) numerous and intense in simulations with warmer (cooler) SSTs.

To complement the WRF experiments for the 20 May 2013 and 10 February 2013 events, idealized cloud-resolving model simulations were conducted using Cloud Model 1 (CM1). These warm and cool-season events, respectively, were used to determine if the influence of climate change on extreme tornado events exhibits any seasonality. In the idealized simulations, despite smaller and shorter-lived supercells, assessment of “tornado power” suggests more powerful tornado-like vortices under PGW in both cases. This is consistent with the stronger tornado proxy found in the WRF PGW simulations.
ACKNOWLEDGMENTS

There are a lot of people and organizations that made this all possible. First, I would like to thank my incredible parents, Mike and Dottie. They have worked and given so much just so I could pursue my dreams. I am also grateful for my friends, many of which I met during this journey. They provided many great laughs and memories, and always supported me. Thank you to the University of Oklahoma and the University of Illinois for providing me with a great education. I am especially thankful for my advisor Jeff Trapp, who reached out and gave me this opportunity. Jeff is not only an amazing scientist, but also an awesome person with a great taste in music (Foo Fighters). I have learned and accomplished much under his advisement, for which I will forever be grateful. Finally, thank you to the National Science Foundation (AGS 1923042) for funding this research.
# TABLE OF CONTENTS

CHAPTER 1: Introduction .............................................................................................................1

CHAPTER 2: Methodology ..........................................................................................................8

CHAPTER 3: WRF PGW Results ..............................................................................................21

CHAPTER 4: SST Experiments with a Cold-season Event .......................................................79

CHAPTER 5: CM1 Idealized PGW Simulations ......................................................................92

CHAPTER 6: Conclusions .........................................................................................................111

REFERENCES .........................................................................................................................115
CHAPTER 1: INTRODUCTION

Hazardous convective weather (HCW) in the form of damaging winds, hail, and tornadoes poses a serious threat to life and property in the United States. From 1980 to 2019, there have been 113 HCW events that produced over $1 billion in damage (NCEI, 2020). The frequency of these events has increased markedly since the start of the 21st century, owing in part to increased exposure and population density (Strader et al., 2017), but also potentially anthropogenic climate change (ACC).

While some studies have found increasing trends in certain convective hazards, such as large hail (Tang et al., 2019), a nationwide trend in tornado occurrence has not been detected (Gensini and Brooks, 2018). However, historical trends in HCW frequency and intensity can be difficult to diagnose due to non-meteorological factors influencing storm reporting (Doswell et al., 2005; Trapp et al., 2006; Allen and Tippett, 2015; Blair et al., 2017).

An alternative to the use of reports or observations of storms (and their associated hazards) is the analysis of storm-environment information. The hazards produced by deep convective storms are primarily dictated by convective intensity and convective mode (e.g., Smith et al. 2012). These traits depend on the buoyancy and vertical wind shear in the storm environment (Weismann and Klemp, 1982), both of which are predicted to change under ACC (e.g., Trapp et al., 2007a). Elevated greenhouse gas concentrations warm the atmosphere and augment moisture content in part through increased evaporation (Trenberth, 1999). Warming and moistening of planetary boundary layer air yields increases in convective available potential energy (CAPE). Conversely, rapid warming of the high latitudes relative to the low latitudes weakens the meridional temperature gradient. The thermal wind relationship:
\( \frac{\partial \vec{V}}{\partial z} \approx \frac{g}{fT} \hat{k} \times \nabla_h T \quad (1) \)

where vertical wind shear \((\partial \vec{V}/\partial z)\) is directly proportional to the horizontal temperature gradient \((\nabla_h T)\), indicates that such Arctic amplification will decrease vertical wind shear in the future on average. Research shows that the projected decreases in shear are disproportionately smaller than the projected increases in CAPE, suggesting that an increase in frequency and/or intensity of future HCW events is likely to occur under anthropogenic climate change (Del Genio et al., 2007; Trapp et al., 2007a, 2009; Van Klooster and Roebber, 2009; Gensini et al., 2014; Diffenbaugh et al., 2013; Gensini and Mote, 2014; Trapp and Hoogewind, 2016; Hoogewind et al., 2017; Rädler et al., 2019; Trapp et al., 2019).

These studies use general circulation model (GCM) or regional climate model (RCM) output to analyze basic environmental parameters such as CAPE and 0-6 km vertical wind shear, which are assumed to be properly resolved by the GCMs (Del Genio et al., 2007; Trapp et al., 2007a, 2009; Van Klooster and Roebber, 2009; Gensini et al., 2014; Diffenbaugh et al., 2013; Rädler et al., 2019). However, because the grid spacing of these models is too coarse to explicitly resolve hazardous convective storms, these studies assume that any increase in HCW environments will be manifest in the form of increased HCW events. Recent studies have shown that this is not necessarily a good assumption, especially during the summer where many GCMs indicate excessive convective inhibition (CIN) that can prevent storm initiation (Diffenbaugh et al., 2013). Indeed, CIN has been increasing at a rate of 3-5% per decade across the United States (Taszarek et al., 2020). Some studies have attempted to circumvent the uncertainty of convective initiation by utilizing the convective parameterization in the model, assuming that specified convective precipitation threshold exceedances indicate the initiation of a storm (Trapp et al.,
the convective hazards that may occur in a given storm.

With computational cost and/or lack of computational resources preventing the widespread use of convection-permitting GCMs for long-term climate simulations, other techniques have been pursued to address these convective-scale issues. Trapp et al. (2007b) showed that realistic convective-allowing simulations, i.e., those with horizontal grid spacings of ~4 km or less, and thus those run without parameterized convection, could be generated by “dynamically downscaling” coarse-resolution data over regional domains. In dynamical downscaling, a high-resolution regional model is used to physically reduce large-scale climate information down to regional scales. Subsequent studies began dynamically downscaling global reanalysis or GCM output at convective-allowing resolutions to address current and future trends in convective storms (Trapp et al., 2011; Robinson et al., 2013; Gensini and Mote, 2014; Gensini and Mote, 2015; Hoogewind et al., 2017; Prein et al., 2017; Trapp et al., 2019). This permits more detailed analysis and understanding of the influence of the larger-scale environment on convective initiation, intensity, and mode. For instance, despite favorable convective environments, Hoogewind et al. (2017) found reduced HCW events in future summers (2071-2100) due to increases in CIN. This indicates the possibility of less frequent but more extreme HWCs, consistent with the results of other downscaling studies (Trapp et al., 2019; Taszarek et al., 2020). Thus, high-resolution dynamical downscaling provides a good comparison to studies using environmental proxies.

Yet, even high-resolution dynamical downscaling has limitations: It only grants indirect, statistical conclusions on HCW because the frequency and intensity of HCW events are assessed through storm-scale proxies. For example, one may use updraft helicity to determine that rotating
storms in a future climate are stronger than those of the past climate. While still a valuable conclusion, it does not easily address the underlying physical mechanisms responsible for increases in intensity. Dynamical downscaling also does not allow an answer to the following question of relevance herein: Would a historical tornado event be more intense under the influence of anthropogenic climate change?

Thus, this research employs the pseudo global warming (PGW) methodology, which focuses on a specific event and analyzes the mechanisms through which ACC impacts the event. This is accomplished by comparing simulations of the event using the original 3D meteorological forcing with simulations using a climate-change-modified version of the original 4D meteorological forcing.

The PGW approach was developed by Kimura and Kitoh (2007) and Sato et al. (2007), who built upon the surrogate warming method of Schar et al. (1996) and Frei et al. (1998). The methodology is centered around the use of climate change differences (or “deltas”) that are derived from differences in the future and historical climates of GCMs. These climate change deltas are then added to the 4D meteorological forcing used initially to produce an accurate control simulation. This essentially allows the historical event to be simulated in a future climate. PGW methods have been employed in a variety of scenarios that may encompass a variety of timescales, including snowfall, flooding, hurricane, and tornado events (Rasmussen et al., 2011; Lackmann, 2013, 2015; Trapp and Hoogewind, 2016; Dougherty and Rasmussen, 2020). The focus of this study is event-based PGW, where 1-day tornado events are examined in detail. While similar to Trapp and Hoogewind (2016), this research features novel PGW work incorporating idealized simulations and diurnally varying (DV) climate change deltas. Previous PGW experiments implemented time-constant (TC) deltas computed from the difference in
monthly means of a future and historical decade, which assumed no change in the future diurnal cycle. This study incorporates time-dependent deltas to capture changes in the diurnal cycle of a future climate. Since the traditional delta construction includes all days of the given month, including those unfavorable for convection, we also formulate a new delta to isolate higher-end convective days. These “CAPE deltas” are constructed only from days that meet certain CAPE and CIN criteria in an effort to capture the most extreme responses supported by climate change. It is hypothesized here that tornado intensity will increase under ACC, especially during the months of October-February, hereinafter referred to as the “cool season”. Warmer and moister air of a future climate will be realized as larger CAPE values. The “thermodynamic speed limit” of Fiedler and Rotunno (1986) suggests that greater atmospheric instability would yield stronger tornadoes. This theory is based on the premise that a near-surface pressure deficit caused by an updraft will constrain, and must be matched by, the tornadic-vortex intensity via cyclostrophic balance considerations. While a bit simplistic, the thermodynamic speed limit illustrates that changes in CAPE would influence maximum tornado intensity. With CAPE projected to increase under ACC, it is reasonable to assume that the ceiling for future, extreme tornadoes would be higher; however, increased CIN may temper convective initiation. Additionally, tornadogenesis first requires the formation of storm-scale rotation that is then intensified near the surface. Mesocyclogenesis is at least partially modulated by vertical wind shear, thus the environmental winds also play a role in tornado strength. Because of projected decreases in vertical wind shear, it is hypothesized that the largest increases in tornado intensity will occur during the cool season, when tornadoes often occur in the presence of abundant shear but limited instability. The degree of this cool-season instability is also augmented by oceanic characteristics. Future boundary layer warming and moistening may be amplified by ACC-induced warming of the Gulf of
Mexico. During the cool-season, low-level moisture – and subsequently, instability – is often the limiting factor in severe convection (Childs et al., 2018). It has been shown that over 50% of the moisture in cool-season tornadic thunderstorms originates in the Gulf of Mexico, and warmer sea surface temperatures (SSTs) enhanced the moisture content of overlying air masses (Molina and Allen, 2019). Because of these connections, it is also hypothesized here that increases in the SSTs over the Gulf of Mexico will significantly increase storm intensity during cool-season tornado outbreaks.

Because of their profound human impacts and characterization as extreme (and thus rare) events, the 10 February 2013 and 20 May 2013 tornado outbreaks were selected for PGW experimentation. Statistically, EF-4 and EF-5 tornadoes like those that struck Hattiesburg, Mississippi and Moore, Oklahoma on 10 February and 20 May respectively, account for approximately 1% of all tornadoes (Brooks and Doswell, 2001). Together, these tornadoes were responsible for 24 fatalities, 300+ injuries, and approximately $2 billion in damage (NWS Norman, NWS Jackson). The representative environment on 20 May was characterized by a mixed-layer CAPE (MLCAPE) of 3120 J kg$^{-1}$, which exceeds the 75th percentile for significant, Southern Plains spring tornadoes (Grams et al., 2012). Vertical wind shear was moderate, with values falling within the inner-quartile ranges specified by Grams et al. (2012). The environment for the Hattiesburg tornado provides a stark contrast to the Moore event. These soundings featured CAPE that was only one quarter of that of the Moore case, but 0-1 km vertical wind shear over twice as strong at ~20 m s$^{-1}$. It should be noted that these environmental characteristics are common among significant tornadoes in the cool-season (Grams et al., 2012). The wide variety of extreme tornado environments illustrates the importance of investigating both warm and cool-season events.
This study aspires to assess how extreme tornado events may unfold under climate change. An attempt will be made to explain any changes to convective intensity and evolution through ACC-induced environmental modifications and current understanding of convective storms. Because it allows for process-based understanding and helps isolate the effects of ACC, the PGW methodology is applied to the 10 February and 20 May 2013 tornado outbreaks. This work also aims to provide guidance for future PGW studies by investigating the importance of climate-change delta construction methods.

Further details of the PGW methodology are provided in Chapter 2. Results of the PGW experiments are described in Chapter 3, followed by a brief discussion of Gulf of Mexico SST experimentation in chapter 4. Chapter 5 will feature results from idealized simulations. Finally, conclusions and thoughts on future research are given in Chapter 6.
CHAPTER 2: METHODOLOGY

2.1: PGW METHOD

In essence, the PGW method involves simulation of some event under its actual, present-day forcing (the control; CTRL), which is compared to a simulation of the event under a modified version of the original forcing. This modification comes from the addition of a climate change delta, which represents the average difference between a future and historical climate over a time period such as a month. Once quantified, the climate change deltas (hereinafter, deltas) were then added to the meteorological forcing of the control simulation. This process is shown by the following equation,

\[ T(x, y, z, t) = T(x, y, z, t) + \Delta T_{\text{month}} \]  \hspace{1cm} (2)

where temperature is used as an example. Represented by the \( \Delta T \) term, the deltas were added to the forcing at their respective times. Simulations were then run using the climate-change-modified initial and boundary conditions.

To construct these deltas, historical and Representative Concentration Pathway 8.5 (RCP8.5) simulations from five GCMs (GFDL-CM3, MIROC5, NCAR-CCSM4, IPSL-CM5A-LR, and NorESM-1M) were used to quantify five climate change scenarios at the end of the 21st century. RCP8.5 projections were used to represent a “worst case scenario” future climate. GCM data were obtained from the Coupled Model Intercomparison Project phase 5 (CMIP5) archive maintained by the Earth System Grid Federation. Following Trapp and Hoogewind (2016), these five GCMs were selected because of how well the historical frequency of convective-storm environments compared to reanalysis (Diffenbaugh et al., 2013; Seeley & Romps, 2015). The delta computation is illustrated by Eq. 3, where temperature is again used as an example:
\[ \Delta T_{\text{month}}(x, y, z) = \overline{T_{\text{month}}(x, y, z)}|_{\text{future}} - \overline{T_{\text{month}}(x, y, z)}|_{\text{past}} \]  

(3)

The overbar in (3) indicates a time average of the variable. The averaging periods are typically decadal, although some have used longer periods, for example 30 years (Liu et al., 2017).

Previous PGW studies have used this approach to produce time-constant (TC) deltas. Such TC deltas imply that the diurnal cycles in meteorological variables such as surface temperature and surface specific humidity do not change in the future climate. This is a questionable assumption, based on GCM projections of relatively warmer nights (e.g., Schoof and Robeson, 2016). To investigate the importance of this assumption, in the present study diurnally-varying (DV) deltas were created that capture modelled changes to the diurnal cycle that can influence the timing and development of deep convection. These DV deltas were created with 6-hourly GCM data (at 00, 06, 12, 18 UTC); each time period is an average across all days of a month. The DV deltas were then linearly interpolated to include times at 03, 09, 15, and 21 UTC to match the 3-hourly reanalysis data used to force the PGW simulations. To test the importance of diurnal variation and averaging periods, 30-year (“30y”; 1970-1999/2070-2099) DV, 10-year (“10y”; 1990-1999/2090-2099) DV, and 10y TC deltas were calculated from each set of GCM output.

Averaging across multiple GCMs to achieve a single set of deltas is a common practice (Liu et al., 2017; Poujol et al., 2020; Nayak and Takemi, 2020). This is presumably done because of limited computational resources, as it incorporates climate-change signals from the multiple GCMs but only requires one simulation. However, it has been shown that deltas of individual GCMs and the subsequent PGW responses can vary substantially (Lackmann, 2015; Trapp and Hoogewind, 2016). Here, a “composite delta” is also calculated by averaging individual deltas from the 5 GCMs to determine if the response to a mean forcing is analogous to the mean response of several independent forcings.
An additional set of DV deltas was created using only days of the month conducive to deep convection. These “CAPE deltas” were constructed in the same manner as the DV deltas, but solely utilize days meeting CAPE and CIN criteria based on 18 UTC GCM output. The instability calculations were done by averaging 18 UTC CAPE and CIN values over a domain encompassing Oklahoma. CAPE values in the 50th percentile were sorted by lowest CIN, where the top 10% were selected. These CAPE deltas were designed to elicit the most extreme responses fostered by ACC.

2.2: 20 MAY 2013 WRF SIMULATIONS

The CTRL and PGW simulations of the 20 May 2013 event were performed using the Weather Research and Forecasting model with the Advanced Research core (WRF; Skamarock et al. 2008) at a convective-permitting resolution. The parent domain had a horizontal grid-spacing of 3 km and 40 vertical levels. As indicated in Fig. 1, a subdomain of 1 km grid-spacing was nested within the parent domain over central Oklahoma; two-way feedback between the nested and parent domains was enabled. The simulations were initialized at 12 UTC on 20 May and terminated at 06 UTC on 21 May. This allowed for nearly 6 hours of “spin-up” time before convection began to initiate. Initial and boundary conditions were derived from 3-hourly North American Mesoscale Forecast System (NAM) analysis data with 12 km horizontal grid spacing and 42 vertical levels. Additional details regarding WRF model setup and parameterizations can be found in Table 1.

The control simulation (hereinafter, CTRL) was established by comparing simulations with various model configurations to observed radar characteristics and tornado reports. A
composite radar reflectivity analysis was produced over central Oklahoma (nested domain region) using the PyART (Helmus and Collis, 2016) package applied to data from the Oklahoma City, Lawton, and Tulsa, OK WSR-88D radars. The reflectivity data from these radars were interpolated to a uniform horizontal grid of 1 km grid-spacing, at a level 1 km above ground level, to match the output of the nested domain. This allowed for a direct comparison between observed and simulated reflectivity. Using simulated updraft helicity (UH) exceedances of 500 m$^2$ s$^{-2}$ as a proxy for tornado occurrence (see Chapter 3b for justification), the location and timing of tornadic storms in the CTRL were compared with observations. The accuracy of the CTRL simulation was determined quantitatively by searching for tornado proxy within 15 km of observed tornado paths.

At 2000 UTC, the time of the Moore tornado, the CTRL simulation lacked some of the observed convective storms south of Moore (Fig. 2a), as shown also in the comparative distributions of reflectivity at 20 UTC (Figure 2b). However, the CTRL simulation captured the Moore supercell well (Figure 2a; cell containing black dot), and later also captured the initiation of cells south of this supercell. Figure 3 shows the locations of CTRL tornado proxy and the observed EF-5 tornado path. Spatially and temporally, the CTRL simulation sufficiently represented the path of the Moore tornado, as well as several other observed tornadoes (not shown). After finding the CTRL simulation a sufficient representation of the historical case, the different delta types were then added to the input forcing to produce a total of 21 PGW simulations that are discussed in Chapter 3.2. These PGW simulations were analyzed in comparison to the CTRL, examining both environmental and storm characteristics, to evaluate possible effects of a warmer climate upon this event if it occurred at the end of the century.
2.3: 10 FEBRUARY 2013 WRF SIMULATIONS: SST SENSITIVITY

It was hypothesized in Chapter 1 that with anthropogenically-increased SSTs in the Gulf of Mexico, cool-season tornado outbreaks will become more intense. To test this hypothesis from a PGW framework, a series of WRF simulations were conducted with uniformly perturbed SSTs over the Gulf of Mexico. For these simulations, the 10 February 2013 tornado outbreak was simulated at 4 km horizontal grid spacing with uniform SST deltas of -5, -2.5, -1, +1, +2.5, and +5°C across the entire Gulf of Mexico basin; no other variables were modified. The magnitude of these perturbations was consistent with GFDL and NCAR SST deltas (4.2 and 3.0°C, respectively). As with the PGW simulations, initial and boundary conditions were set by 3-hourly NAM analysis data. A single model domain with 4 km grid spacing was positioned to encompass the entire Gulf of Mexico and the southeastern United States (Fig. 4). Model parameterizations and other configuration information are listed in Table 2. The WRF model was run continuously for 78 hours (8 February 00 UTC – 11 February 06 UTC) for all simulations, allowing for nearly three days of integration time prior to the onset of the outbreak. A longer integration time leading up to the event was necessary to give the perturbed SSTs adequate time to precondition the environment over the land. Given the extended integration time, spectral nudging was used to help confine the synoptic-scale evolution to that of the NAM analysis data and provide a more accurate simulation. The nudging was only applied to geopotential heights, horizontal winds, and temperatures above the planetary boundary layer. While constraining the large-scale atmosphere, the spectral nudging technique used here permitted free evolution of mesoscale features and atmospheric moisture. The storms and environments of the six perturbed SST simulations were then compared to the CTRL simulation for this event.
To garner a one-to-one storm comparison between the historical event and PGW simulations, Cloud Model 1 (CM1; Bryan and Fritsch, 2002) was used. Novel for PGW applications, idealized simulations feature steady, horizontally homogeneous environments. This allows storm characteristics such as intensity to be related to the initial environment more easily; upon adding a climate change delta to the initial environment, this further allows for a more straightforward assessment of the effect of climate change on intensity. Deep convection was initiated by a prescribed forcing, and a “translating” model domain was used to continually center the storm within the domain, allowing for a smaller domain and grid spacing. In both simulations, a 180 x 180 x 18.5 km model domain was used. Grid stretching was employed in the horizontal as well as vertical directions. In the horizontal, grid spacing was 64 m over the inner 80 x 80 km of the domain and then was increased to 2.5 km at the domain edges. Vertical grid spacing varied from 20 m in the lowest 300 m to 250 m in the upper 6000 m, with 125 m grid spacing in between. Additional details regarding CM1 model setup and parameterizations can be found in Table 3.

The initial and boundary conditions used for the CM1 CTRL simulations were derived from single soundings preceding the events. The Moore CTRL simulation environment was a modified version of the 20 May 2013, 18 UTC sounding launched from Norman, Oklahoma. The capping inversion was weakened by modifying lapse rates in the 1-5 km layer (Fig. 5a) to achieve a long-lived supercell representative of the May 20th event. The 11 February 2013 00 UTC Slidell, Louisiana sounding was used for the Hattiesburg supercell simulation, and required no modifications (Fig. 5b). In both simulations, convection was initiated via updraft nudging (Naylor and Gilmore, 2012) that persisted for 20 minutes, a technique that has been adopted in
other studies (e.g. Naylor and Gilmore, 2014; Coffer and Parker, 2017). After 20 minutes, the storms were free of any artificial forcing.

The CM1 initial and boundary conditions were represented by a base-state atmosphere at a specific location and time, so the added PGW deltas must also be representative of that location and time. Specifically, for this idealized modeling application of PGW, delta profiles were computed using GCM grid point data nearest to the time of the respective events, and within a 100 x100 km horizontal box around the geographic locations. Using temperature again to illustrate the method, the deltas for the idealized PGW simulations were computed as

$$
\Delta T_{\text{month}}(z) = \langle T_{\text{month}}(x, y, z, t) \rangle_{\text{future}} - \langle T_{\text{month}}(x, y, z, t) \rangle_{\text{past}} \quad (4)
$$

where \( t = 1800 \text{ UTC or 0000 UTC}, \) \( \text{month} = \) May or February, and \( \langle \rangle \) indicates the 100 x100 km horizontal average. The delta profile from each GCM was then added to the input sounding (Fig. 5), to conduct the PGW simulations, and differences from the CTRL simulations were used to investigate possible physical mechanisms resulting from ACC.
Figure 1: WRF parent domain (black box) and nested domain (red box) for 20 May 2013 PGW simulations.
Table 1: Parameterizations and settings for 20 May 2013 WRF simulations.

<table>
<thead>
<tr>
<th>Parameterization/Scheme</th>
<th>Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Microphysics</td>
<td>NSSL 2-moment</td>
</tr>
<tr>
<td>Land Surface</td>
<td>Unified Noah land-surface model</td>
</tr>
<tr>
<td>PBL Physics</td>
<td>Mellor-Yamada-Janji TKE</td>
</tr>
<tr>
<td>Radiation</td>
<td>Goddard shortwave and longwave</td>
</tr>
<tr>
<td>Domain Interaction</td>
<td>Feedback enabled</td>
</tr>
</tbody>
</table>
Figure 2: a). Observed 1 km reflectivity (left panel) and simulated 1 km reflectivity at 20 UTC. b). Quantitative analysis of (a) in grid point exceedances of various dBZ thresholds.
Figure 3: 30-minute updraft helicity swath (color fill) from the CTRL simulation with Moore, OK EF-5 tornado track (red line).
Figure 4: WRF domain for SST sensitivity experiment.
Table 2: Parameterizations and settings for WRF SST experiment.

<table>
<thead>
<tr>
<th>Parameterization/Setting</th>
<th>Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vertical levels</td>
<td>40</td>
</tr>
<tr>
<td>Input forcing</td>
<td>3-hourly NAM analysis</td>
</tr>
<tr>
<td>Microphysics</td>
<td>NSSL 2-moment</td>
</tr>
<tr>
<td>Land surface</td>
<td>Unified Noah land-surface model</td>
</tr>
<tr>
<td>PBL physics</td>
<td>YSU scheme</td>
</tr>
<tr>
<td>Spectral nudging coefficient</td>
<td>0.0004</td>
</tr>
</tbody>
</table>

Figure 5: a). Modified 18 UTC Norman, OK observed sounding with PGW-modified versions overlayed. b). Same as in (a), but for 00 UTC Slidell, LA observed sounding.

Table 3: CM1 model setup and parameterizations.

<table>
<thead>
<tr>
<th>Parameterization/Scheme</th>
<th>Type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Moisture</td>
<td>Morrison 2-moment</td>
</tr>
<tr>
<td>Coriolis</td>
<td>Off</td>
</tr>
<tr>
<td>Lower boundary condition</td>
<td>Free-slip</td>
</tr>
<tr>
<td>Forcing</td>
<td>Updraft nudging (30 min.)</td>
</tr>
</tbody>
</table>
  - Velocity: 10 m/s
  - Horizontal radius: 10 km
  - Vertical radius: 1.5 km
  - Center height: 1.5 km
CHAPTER 3: WRF PGW RESULTS

3.1: COMPARISON OF THE DELTAS

To begin, comparisons are made between the diurnally varying (DV) and time constant (TC) deltas used in the PGW simulations for the month of May. These are computed using 10y averages, from 1990 to 1999 and 2090 to 2099. Difference fields (DV delta minus TC delta) in 2-meter temperature (T2m) and specific humidity illustrate the magnitude of diurnal variation in each GCM (Figs. 6 and 7). Relative to the TC deltas, DV deltas are smaller (but still positive) in the overnight hours (06/12 UTC) and larger during the afternoon and evening (18/00 UTC). This is expected, as solar radiation and subsequent sensible heating is maximized during the day. Differences of ±3°C are found in the Desert Southwest and northern Mexico for the deltas associated with the GFDL GCM. Elsewhere, difference magnitudes are generally less than 1.5°C, which are also characteristic of differences across the U.S. for all the other GCMs examined. Specific humidity differences of up to 1 g kg\(^{-1}\) are found across much of the CONUS for all the GCMs. Decreases in 2-meter moisture during the afternoon and evening are to be expected as diurnal heating promotes deeper boundary layer mixing and may play a significant role in convective initiation in PGW simulations. Aloft, differences in variables reduce to less than 10\% of their surface differences (not shown), as impacts of the diurnal cycle are mostly confined to the planetary boundary layer.

When comparing DV deltas computed using 30y to those using 10y averaging periods, small changes in magnitude are noted. The 30y deltas were comprised of years from 1970 to 1999 and 2070 to 2099. As evident in Fig. 8, temperature deltas are larger with the 10y averaging period, both at 2 m and at the 700 hPa pressure level. Considering that the 10 years used are at
the end of the 30y periods, this is unsurprising. In other words, the 30y deltas consist of earlier (cooler) years that result in smaller climate change differences. Generally, larger changes in all meteorological variables occur in the 10y deltas. However, spatial similarities between the 30y and 10y deltas exist over much of the CONUS, with Pearson correlation coefficients greater than 0.8 and 0.9 for 700 hPa T and T2m, respectively. Root-mean-squared differences for these fields were less than 0.75°C for 700 hPa T, and less than 0.5°C for T2m (Table 4).

To address the argument that 10y deltas may be capturing more natural climate variability than anthropogenic climate change, and thus that 30y deltas may be more appropriate for PGW considerations (Liu et al., 2017), CAPE distributions from the averaging periods are examined. CAPE was chosen because of its relevance to deep convection as well as its dependence on atmospheric variables such as temperature and boundary layer moisture. Using a domain centered on Oklahoma (Fig. 9), domain-averaged 18 UTC CAPE distributions were created using daily data for the month of May for each 10y and 30y period. Statistical similarity is seen in both the historical and future periods for each GCM (Table 5). This is supported by a t-test, with only the GFDL future having statistically different distributions between the 10y and 30y periods at the 95% confidence interval (Table 5). Thus, because of this similarity in the distributions, there does not appear to be a compelling reason to use a longer averaging period to create the deltas.

Additional considerations were taken to determine whether the 10y averaging period was long enough to achieve an approximately balanced flow. It is important that the climate-change deltas represent a sufficiently balanced base state (e.g., in terms of geostrophy) such that it can be added to the perturbation-like weather event. Evaluation of the degree of balance was conducted by computing differences in actual and geostrophic 500 hPa winds for historical and
future averaging periods. Monthly-averaged GCM data were interpolated to the NAM grid, and the geostrophic winds were then calculated as:

\[ U_{geos} = -\frac{g}{f} \frac{\delta Z}{\delta y} \]  
\[ V_{geos} = \frac{g}{f} \frac{\delta Z}{\delta x} \]  

where \( Z \) is geopotential height, \( f \) is the Coriolis parameter, \( g \) is the acceleration due to gravity, and \( \delta x \) and \( \delta y \) represent the horizontal distances between every 5th grid point (which was used to reduce the number of calculations) (Fig. 9). The domain-averaged departure from geostrophy (i.e., \( |U - U_{geos}| \) and \( |V - V_{geos}| \)) was generally less than 10% of the 500 hPa wind speeds (Table 6). Additionally, geostrophy departures of the 10y and 30y averaging periods were nearly identical (Table 6). Thus, it is concluded that 10y deltas also represent a sufficiently balanced state and are suitable for PGW experimentation.

Unlike typical deltas that use all days of the averaging period, CAPE deltas were constructed from days with the 10% lowest CIN and 10% highest CAPE. Most of the CAPE deltas are associated with greater low-level moisture across central Oklahoma (Fig. 10). Plots of 700 hPa T and T2m reveal that these variables are of greater amplitude in the CAPE deltas as compared to the DV deltas (e.g., Fig. 8), although not necessarily over the domain of CAPE/CIN computations (Fig. 11). There is significant variation in the spatial patterns of the deltas between GCMs. Given that the purpose of the CAPE deltas was to locate the most extreme convective environments over Oklahoma occurring during the month of May, it is intriguing to see such synoptic-scale variance between GCMs. As will be shown later, not all of the CAPE deltas manifested extreme environments, nor extreme storms, in their simulations.
Figure 6: Difference fields (diurnally varying [DV] deltas minus time constant [TC] deltas) in 2-m temperature (T2m) (°C), at 00 UTC, 06 UTC, 12 UTC, and 18 UTC, for deltas based on CMIP5 simulation output from GFDL-CM3, IPSL-CM5A-LR, MIROC5, NCAR CCSM4, and NorESM-1M.

Figure 7: Same as in Fig. 6, but with 2-m specific humidity (g/kg).
Figure 8: 700 hPa temperature (top panel) and 2-m temperature (bottom panel) plots of 10y DV and 30y DV deltas based on CMIP5 simulation output from GFDL-CM3, IPSL-CM5A-LR, MIROC5, NCAR CCSM4, and NorESM-1M.

Figure 9: Computational domain for WRF simulations. Red dots indicate gridpoints used in calculations of geostrophic winds and departure from geostrophy (see text). Bold outline indicates approximate domain used for the CAPE/CIN calculations for CAPE deltas.
Figure 10: 850 hPa specific humidity (top row) and 2-m specific humidity (bottom row) “CAPE” deltas based on CMIP5 simulation output from GFDL-CM3, IPSL-CM5A-LR, MIROC5, NCAR CCSM4, and NorESM-1M. CAPE/CIN computation domain (red box) is shown in the upper-left panel.

Figure 11: 700 hPa temperature (top row) and 2-m temperature (bottom row) “CAPE” deltas based on CMIP5 simulation output from GFDL-CM3, IPSL-CM5A-LR, MIROC5, NCAR CCSM4, and NorESM-1M. CAPE/CIN computation domain (red box) is shown in the upper-left panel.

Table 4: Comparison of deltas in 2m temperature (T2m; °C) and 700 hPa temperature (°C) constructed using 10y versus 30y averages, and based on CMIP5 simulation output from GFDL-CM3, MIROC5 and NCAR: Root-mean-square (RMS) differences and Pearson correlation coefficients (r).

<table>
<thead>
<tr>
<th>GCM</th>
<th>RMS, T2m</th>
<th>r, T2m</th>
<th>RMS, 700 hPa T</th>
<th>r, 700 hPa T</th>
</tr>
</thead>
<tbody>
<tr>
<td>GFDL</td>
<td>0.51</td>
<td>0.97</td>
<td>0.74</td>
<td>0.80</td>
</tr>
<tr>
<td>MIROC</td>
<td>0.42</td>
<td>0.98</td>
<td>0.66</td>
<td>0.85</td>
</tr>
<tr>
<td>NCAR</td>
<td>0.38</td>
<td>0.94</td>
<td>0.42</td>
<td>0.83</td>
</tr>
</tbody>
</table>
Table 5: Mean May (mixed-layer) CAPE (J kg\(^{-1}\)) computed across an Oklahoma-centered domain (see Figure 1) using 10y versus 30y averaging periods. Summary of t-test in which the CAPE distributions computed from 10y versus 30y averaging periods are compared.

<table>
<thead>
<tr>
<th>GCM</th>
<th>30-yr historical</th>
<th>10-yr historical</th>
<th>30-yr future</th>
<th>10-yr future</th>
<th>t-statistic: historical/future</th>
<th>p-value: historical/future</th>
</tr>
</thead>
<tbody>
<tr>
<td>GFDL</td>
<td>447.4</td>
<td>485.8</td>
<td>943.0</td>
<td>1110.9</td>
<td>-1.65/-3.87</td>
<td>0.100/0.000</td>
</tr>
<tr>
<td>MIROC</td>
<td>1217.2</td>
<td>1235.9</td>
<td>1754.5</td>
<td>1645.0</td>
<td>-0.31/1.51</td>
<td>0.756/0.130</td>
</tr>
<tr>
<td>NCAR</td>
<td>546.1</td>
<td>564.7</td>
<td>914.4</td>
<td>969.2</td>
<td>-0.55/-1.11</td>
<td>0.585/0.266</td>
</tr>
</tbody>
</table>

Table 6: Domain-averaged departure from geostrophy (m s\(^{-1}\)) in the zonal (U) and meridional (V) wind components, i.e., |U - U\(_{\text{geos}}\)| and |V - V\(_{\text{geos}}\)|, at 500 hPA, during the month of May, over the subdomain indicated in Figure 1. The geostrophy departures are computed using 10y versus 30y averaging periods.

| GCM   | |U - U\(_{\text{geos}}\)| historical | |U - U\(_{\text{geos}}\)| future | |V - V\(_{\text{geos}}\)| historical | |V - V\(_{\text{geos}}\)| future |
|-------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|
| GFDL  | 0.235           | 0.260           | 0.685           | 0.710           |
| MIROC | 0.220           | 0.235           | 0.635           | 0.605           |
| NCAR  | 1.575           | 1.560           | 2.805           | 2.815           |
3.2: MAY 20TH PGW RESPONSE

Diurnally-varying deltas, time-constant deltas, 10y average deltas, 30y average deltas, and the CAPE deltas from the five GCMs, as well as a composite 10y DV delta, were added to the CTRL forcing to provide 21 PGW simulations in total for the 20 May case. Herein, the analyses presented will focus on the sub-domain (“domain 02”) positioned over central Oklahoma (Fig. 1), as this is where the Moore EF-5 tornado occurred. Storms in the southern portion of this region initiated along a dryline, whereas a stationary front was responsible for storm initiation north of Moore.

Time series of grid point occurrences of 40 dBZ simulated reflectivity, used here as a general indicator of deep convection, reveal that convective storm initiation was delayed 1-3 hours in the PGW simulations relative to the CTRL simulation (Fig. 12). However, in many PGW cases the amount of convection over the domain quickly equaled and exceeded that of the CTRL, indicating rapid development of deep and widespread convection. More intense deep convection, indicated here by the exceedance of a simulated reflectivity of 55 dBZ, often occurred earlier in the PGW simulations and was more extensive than in the CTRL simulation (Fig. 13). In addition, as convection (including intense convection) began to decline after 00 UTC in the CTRL simulation, in many PGW simulations it often continued to persist, up to the end of the simulation period.

When the simulated convection is examined cumulatively, overall trends become clearer. Two thirds of the PGW experiments featured a greater amount (spatially and/or temporally) of reflectivity exceedances (40 dBZ threshold) than the CTRL simulation (Fig. 14). Those PGW simulations that had fewer 40 dBZ occurrences than the CTRL simulation utilized 30y averaged deltas, or the CAPE deltas. On average, the PGW simulations had over 9% more convection than
the CTRL simulation, but with significant variability (-57.1% for IPSL CAPE to 46.1% for GFDL TC). All but two of the PGW simulations had more intense convection than the CTRL simulation (Fig. 15), yielding an average increase of 109.7%, but again with significant variability (-24.9% for NORESM DV to 283.4% for GFDL TC). Box plots showcasing the spread of PGW reflectivity responses can be seen in Fig. 16. These changes appear to be modulated by the increased amount of “upscale growth” in the PGW simulations, i.e., development of mesoscale convective systems that thus cover more area (Figs. 17 and 18). Overall, in terms of simulated reflectivity, the future environments yielded more extensive and intense convective storms.

Storm motion was 5.6% slower on average in the PGW simulations, owing to decreases in wind speeds aloft that will be discussed later. This gave PGW storms greater residence time within the Oklahoma subdomain. However, reflectivity exceedances and thus convection is poorly related to the $u$-component (the dominant component) of storm motion (computed based on Bunkers et al. 2000), explaining less than 3% of the variability (Fig. 19). Thus, most of the difference in reflectivity exceedances was due to the presence of more widespread convection in the PGW simulations rather than the slower storm motion.
Figure 12: Time series of gridpoint exceedances of 40 dBZ simulated radar reflectivity factor across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 13: Time series of gridpoint exceedances of 55 dBZ simulated radar reflectivity factor across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 14: Cumulative gridpoint exceedances of 40 dBZ simulated radar reflectivity factor across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 15: Cumulative gridpoint exceedances of 55 dBZ simulated radar reflectivity factor across the WRF computational domain, for the control and all PGW-delta experiments.

Figure 16: Box plots of PGW simulated reflectivity exceedance response at 40 and 55 dBZ. Responses are characterized as percent differences relative to the CTRL.
Figure 17: Simulated radar reflectivity factor across the WRF computational domain, for the control and all PGW-delta experiments at 21 UTC.
Figure 18: Simulated radar reflectivity factor across the WRF computational domain, for the control and all PGW-delta experiments at 00 UTC.

<table>
<thead>
<tr>
<th>Control</th>
<th>PGW, 1km Delup</th>
<th>PGW, 2km Delup</th>
<th>PGW, 3km Delup</th>
<th>PGW, 10km Delup</th>
<th>PGW, 20km Delup</th>
<th>PGW, 30km Delup</th>
</tr>
</thead>
<tbody>
<tr>
<td>Control</td>
<td>PGW, 1km Delup</td>
<td>PGW, 2km Delup</td>
<td>PGW, 3km Delup</td>
<td>PGW, 10km Delup</td>
<td>PGW, 20km Delup</td>
<td>PGW, 30km Delup</td>
</tr>
<tr>
<td>Control</td>
<td>PGW, 1km Delup</td>
<td>PGW, 2km Delup</td>
<td>PGW, 3km Delup</td>
<td>PGW, 10km Delup</td>
<td>PGW, 20km Delup</td>
<td>PGW, 30km Delup</td>
</tr>
</tbody>
</table>

1 km Simulated Reflectivity at 00:00z

![Simulated radar reflectivity factor across the WRF computational domain.](image)
Figure 19: Linear regression analysis of u-component storm motion and normalized cumulative 40 dBZ reflectivity exceedances. Exceedances were normalized by the cumulative exceedances of the CTRL.
Although the hydrometeor size and concentration that define simulated reflectivity are functions of convective strength, other parameters, such as updraft velocity and updraft helicity, provide a more direct assessment of storm intensity. Time series of the number of grid points having a (column maximum) updraft velocity exceeding 25 m s\(^{-1}\) imply similar convective evolution to that analyzed using reflectivity thresholds, with the PGW simulations having delayed convective initiation but stronger storms that persist longer, into the overnight hours (Fig. 20). Cumulatively, all but five PGW simulations demonstrated higher frequencies of 25 m s\(^{-1}\) updrafts compared to the CTRL simulation (Fig. 21). The five PGW simulations that did not were predominantly the CAPE delta simulations, which had the most variable results, with percentage differences ranging from -58.8% (MIROC) to 92.5% (GFDL). These findings persisted at higher updraft velocity thresholds. At updraft velocity thresholds of 40 and 50 m s\(^{-1}\), 15 and 17 of PGW runs, respectively, had greater updraft frequencies than the CTRL simulation (Figs. 22 and 23). The mean PGW response at 25, 40, and 50 m s\(^{-1}\) thresholds was +33.4%, +68.8%, and 250.3%, respectively, in terms of percentage differences. An increased presence of strong updrafts under ACC is to be expected, given projected increases in CAPE; however, run-maximum updraft velocities exceeded the control in only 11 of the 21 PGW simulations (Fig. 24). Boxplots illustrating the range of PGW updraft responses are presented in Fig. 25. These results indicate that convective updrafts would be more numerous under ACC, consistent with the increase in storm coverage, but would not necessarily be stronger. Potential explanations for this result will be discussed in Chapter 3.3.
Updraft helicity (UH) is a local, integrated measure of updraft rotation, and generally indicates the simulated presence of a mesocyclone, a prerequisite for tornadogenesis. It is computed as:

$$UH = \int w\zeta \, dz$$  \hspace{1cm} (6)$$

where $w$ is the vertical component of the velocity vector, $\zeta$ is the vertical component of the vorticity vector, and the integration is typically performed over the 2-5 km layer of the atmosphere. UH above 500 m$^2$s$^{-2}$ is utilized here as a proxy for tornado occurrence; a proxy is necessary given the inadequate grid spacing to resolve a tornado-like vortex. This threshold was derived from the 95th percentile of grid points with at least some degree of updraft rotation (i.e. $UH > 100$ m$^2$s$^{-2}$). In all but five of the PGW simulations, the cumulative UH was less than in the CTRL simulation (Fig. 26). The mean PGW response in terms of the percentage difference in tornado proxy occurrence was -14.7%, which was partially skewed by the three GFDL runs that exceeded the CTRL by 48%, 53.6%, and 96.5%, respectively. However, using a greater threshold of 1000 m$^2$s$^{-2}$, representing the 99.9th percentile of grid points exhibiting some updraft rotation, UH exceedances increased in 15 of the 21 PGW runs, with a mean response of +126.9% (Fig. 27). Additionally, peak PGW UH values were greater than the control in 15 simulations (Fig. 28). Boxplots summarizing the overall PGW UH response are shown in Fig. 29. The strongest PGW mesocyclones often featured stronger updraft speeds and less vertical vorticity, suggesting that the increase in UH was driven more by increases in CAPE.

The decrease in UH occurrences at lower thresholds, but increase at higher thresholds, in the PGW simulations may indicate two possible scenarios. First, these results suggest fewer but stronger tornadoes under ACC for this event. This scenario seems to be realized in some of the MIROC and NORESM simulations, where UH swaths are fewer but more intense (Fig. 30). The
other possibility is that the tornadoes are stronger but shorter-lived. This is most apparent in some of the GFDL and NCAR runs, where UH swaths are more intense but less contiguous (Fig. 30). Focusing on the UH track near Moore, Oklahoma, the simulated Moore supercell was discernable in 18 of the PGW runs, albeit displaced north of Moore, and offers a surprisingly close one-to-one comparison with the CTRL simulation (Fig. 31). Most PGW simulations feature fewer exceedances at lower UH thresholds, but more at the higher UH threshold, compared to the CTRL simulation. These results are indicative of a stronger, but more transient supercell in the PGW simulations. In terms of the maximum UH value, 17 PGW runs exceeded the CTRL simulation near Moore, OK. This suggests the possibility, and increased probability, of stronger tornadoes in a future climate.

The last storm-related metric examined for this event was the accumulated precipitation. Spatial distributions between the CTRL and most PGW simulations appear similar (Fig. 32). Sixteen PGW simulations surpassed the CTRL in terms of 5 mm precipitation exceedances (Fig. 33), which is used here as another measure of convective area. Given the relatively low threshold used, it is suspected that the increase was primarily the result of more widespread convection rather than slower propagation. Also, 5 mm and 40 dBZ exceedances correlate well ($r^2 = 0.86$) between simulations, supporting the idea of more storm coverage. The relatively larger convective area was not always manifested in the form of prolonged heavy rainfall, however. When computing the grid point exceedances of 50 mm (~2 inches) precipitation accumulation, only about half of the PGW runs surpassed the CTRL simulation (Fig. 34), with an average increase of 8.1%. This result may be attributable to the more transient nature of the PGW convection. However, when 100 mm (~4 inches) exceedances were considered, 16 PGW simulations surpassed the control (Fig. 35), with an average increase of 56.2%. Changes in
precipitation response are summarized in Fig. 36. Linear regression analysis between 100 mm exceedances and storm motion indicates that the heaviest precipitation amounts are very poorly correlated with changes in storm motion (Fig. 37), suggesting that greater precipitable water and/or instances of locally stronger convection in the PGW simulations are candidate reasons for this result. Nonetheless, it appears ACC would enhance extreme precipitation potential in the 20 May 2013 event.
Figure 20: Time series of gridpoint exceedances of 25 m/s vertical velocity across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 21: Cumulative gridpoint exceedances of 25 m/s vertical velocity across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 22: Cumulative gridpoint exceedances of 40 m/s vertical velocity across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 23: Cumulative gridpoint exceedances of 50 m/s vertical velocity across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 24: Maximum vertical velocity across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 25: PGW updraft responses shown as percentage differences relative to the CTRL. The top three rows show change in updraft velocity exceedances while the bottom row shows change in maximum updraft velocity.
Figure 26: Cumulative gridpoint updraft helicity exceedances of $500 \text{ m}^2\text{s}^{-2}$ across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 27: Cumulative gridpoint updraft helicity exceedances of 1000 m^2/s^2 across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 28: Maximum updraft helicity value across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 29: PGW updraft helicity responses shown as percentage differences relative to the CTRL. The top three rows show change in updraft helicity exceedances while the bottom row shows change in maximum updraft helicity.
Figure 30: Updraft helicity swaths across the WRF computational domain, for the control and all PGW-delta experiments.

Figure 31: Updraft helicity swaths near Moore, Oklahoma, for the control and all PGW-delta experiments.
Figure 32: Simulation-total accumulated precipitation across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 33: Gridpoint total accumulated precipitation exceedances of 5 mm across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 34: Gridpoint total accumulated precipitation exceedances of 50 mm across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 35: Gridpoint total accumulated precipitation exceedances of 100 mm across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 36: Box plots of PGW accumulated precipitation exceedance response of 5, 50, and 100 mm. Responses are characterized as percent differences relative to the CTRL.
Figure 37: Linear regression analysis of u-component storm motion and gridpoint exceedances of 100 mm total accumulated precipitation.
3.3: PGW ENVIRONMENTS

Based on the preceding storm-scale analysis, PGW-induced changes enhanced convective activity in many, but not all, simulations. On average, occurrences of intense simulated reflectivity and stronger updrafts were more numerous than in the CTRL simulation. Many of the updrafts exhibited less helicity, but the presence of some particularly strong mesocyclones, and potentially tornadoes, was more likely in PGW simulations. Here, analysis of the local, pre-convective environment will be used to explain these findings. Convective initiation in the CTRL simulation was approximately 18:30 UTC, with initiation in most PGW runs occurring at least one hour later. To provide atmospheric comparisons that are direct and free of convective contamination, all analyses are conducted at 18:00 UTC within 30 km of Moore, Oklahoma.

Based on the 18 UTC observed sounding at Norman, Oklahoma, the mesoscale environment of 20 May 2013 featured bulk wind shear of 27 m s\(^{-1}\) and 0-3 km storm-relative helicity (SRH) of 156 m\(^2\) s\(^{-2}\), which are typical for significant, Southern-Plains spring tornadoes, and MLCAPE of 3120 J kg\(^{-1}\), which exceeds the mean value for this region and season (Grams et al., 2012). The addition of climate change deltas not only influenced the characteristics of this environment, but also the location, by shifting the position of synoptic and mesoscale boundaries. Most of the PGW simulations exhibited a northwestward shift in the dryline and stationary front present over central Oklahoma. This can be seen in domain plots of 10-meter wind direction and 2-meter specific humidity (Figs. 38 and 39). As a result, a similar shift was seen in the location of convective initiation (not shown). Within a 60x60 km box centered on Moore, Oklahoma, the mean PGW response of 2-meter temperature and specific humidity was +16.7\% and +25.5\%, respectively, in terms of percentage differences relative to the CTRL. The increase in boundary layer temperature and moisture bolstered CAPE values by 25\% (Table 7).
The only PGW simulation in which CAPE decreased was the IPSL CAPE delta, where the dryline and frontal boundary had surged east, thus significantly reducing low-level moisture. This also led to a substantial increase in CIN, at nearly +67000% (~247 J kg\(^{-1}\); Table 7). Excluding the IPSL CAPE delta, the mean PGW CIN response was +463%, which corresponds to an increase of 1.7 J kg\(^{-1}\). The large percentage difference is somewhat misleading due to the minimal CIN in the control simulation (0.37 J kg\(^{-1}\)). In scenarios where forcing is weaker or initial CIN is stronger, these increases in CIN may pose a more serious threat to convective initiation.

Despite nearly ubiquitous increases in CAPE, simulation-maximum updraft speeds exceeded the control in only 11 PGW runs. The failed manifestation of more potent thermodynamic environments into stronger updrafts was also noted in Trapp and Hoogewind (2016). This may be explained by increased hydrometeor loading in the future convective updrafts (e.g., as discussed by Trapp and Hoogewind 2016). When looking at updrafts greater than 25 m s\(^{-1}\), hydrometeor mixing ratios were 45.1% larger on average (Fig. 40). Precipitation loading was exacerbated by an increase in precipitable water, which was 26.4% higher than the CTRL on average (Table 7). Thus, it seems evident that the PGW updraft velocities were hindered by increased precipitation content.

While peak updraft speeds increased in only 11 simulations, maximum UH values were higher in 15 PGW runs. Given that UH is dependent on the colocation of upward vertical motion and low-level horizontal vorticity, discrepancies in UH should be partially explained by changes in updraft velocity and/or storm-relative helicity (SRH). Five PGW simulations (IPSL 30y and 10y DV, MIROC 10y TC and DV, and NORESM 10y TC) that failed to produce a stronger updraft than the control featured higher UH values. This was surprising, as 0-1 and 0-3 km SRH
values at 18 UTC were (Table 8) lower than the CTRL simulation in all five of these simulations. Because UH depends on the colocation of an updraft and SRH, it is possible that the displacement between the strongest updrafts and higher SRH was less in PGW runs. This would also account for the increase in 1000 m$^2$s$^{-2}$ UH exceedances. Overall decreases in SRH, especially from 0-3 km, likely led to the reduction in UH exceedances at the lower thresholds. The decline in SRH can be tied to weakened vertical wind shear. The mean PGW response for both 0-1 and 0-6 km vertical wind shear was approximately -8% (Table 8). Vertical shear in the lowest 1 km has been used to discriminate between significantly tornadic and nontornadic supercells (Thompson et al., 2003), thus this slight decrease in 0-1 km shear may preclude tornadogenesis in some future storms.

Convective mode also plays a role in tornado likelihood, as discrete supercells account for the majority of observed significant tornadoes (Thompson et al., 2012). Upscale growth, from discrete cells into a mesoscale convective system, appears to be more prevalent in PGW simulations due to enhanced convective cold pools. This is evident by the increase in simulated reflectivity exceedances seen in Figs. 12-15. The lifted condensation level (LCL) of the PGW environments increased by 12.3% on average, which corresponded to +125 m (Table 7). This, along with relatively drier low levels, led to greater evaporative cooling and stronger low-level downdrafts. Subsequently, cold pools were cooler and more negatively buoyant, as cold pool equivalent potential temperature differences in PGW simulations were 29% larger than the control on average (Table 7). Here, the cold pool equivalent potential temperature was defined as the hourly-averaged, near-surface value beneath regions of 40 dBZ convection. Differences were then computed by subtracting the 18 UTC equivalent potential temperature near Moore from the minimum cold pool value. More robust cold pools combined with slower storm motion resulted
in outflow-dominant convection. As outflow expanded radially, more convection was initiated along the dryline and cold front, leading to the faster upscale growth into an MCS as determined by contiguous 40 dBZ reflectivity (Figs. 17 and 18). Colder outflow has been shown to be detrimental to tornadogenesis, as rear-flank downdraft parcels become more negatively buoyant (Markowski et al., 2002). This may preclude tornadogenesis or limit tornado duration, and may be responsible for the transient nature of the PGW UH swaths.

Table 7: 2-m temperature (°C), 2-m mixing ratio (g kg⁻¹), most-unstable CAPE and CIN (J kg⁻¹), precipitable water (mm), lifted condensation level (LCL) heights (m), and equivalent potential temperature (θ-e; °C) deficits for the CTRL and all PGW experiments. Values are computed at 18 UTC and represent the average environment within a 30 km radius of Moore, Oklahoma.
Table 8: Same as in table 7, but for 0-1 km and 0-6 km vertical wind shear (m s\(^{-1}\)) and 0-1 km and 0-3 km storm relative helicity (SRH; m\(^2\)s\(^{-2}\)).

<table>
<thead>
<tr>
<th>Variable</th>
<th>0-1 km shear</th>
<th>0-6 km Shear</th>
<th>0-1 km SRH</th>
<th>0-3 km SRH</th>
</tr>
</thead>
<tbody>
<tr>
<td>GFDL CAPE</td>
<td>6.42</td>
<td>29.92</td>
<td>74.9</td>
<td>95.9</td>
</tr>
<tr>
<td>GFDL hourly</td>
<td>3.91</td>
<td>26.06</td>
<td>36.4</td>
<td>71.5</td>
</tr>
<tr>
<td>GFDL monthly</td>
<td>3.47</td>
<td>25.83</td>
<td>33.3</td>
<td>66.7</td>
</tr>
<tr>
<td>GFSL 30y</td>
<td>3.12</td>
<td>25.42</td>
<td>27</td>
<td>61.2</td>
</tr>
<tr>
<td>IFSL CAPE</td>
<td>4.91</td>
<td>38.04</td>
<td>11.1</td>
<td>244</td>
</tr>
<tr>
<td>IFSL hourly</td>
<td>2.85</td>
<td>26.11</td>
<td>26.9</td>
<td>63.7</td>
</tr>
<tr>
<td>IFSL monthly</td>
<td>2.8</td>
<td>26.06</td>
<td>26.3</td>
<td>63.8</td>
</tr>
<tr>
<td>IFSL 30y</td>
<td>2.62</td>
<td>27.44</td>
<td>20.7</td>
<td>58.1</td>
</tr>
<tr>
<td>MIROC CAPE</td>
<td>3.1</td>
<td>18.57</td>
<td>25.2</td>
<td>43.7</td>
</tr>
<tr>
<td>MIROC hourly</td>
<td>3.09</td>
<td>21.31</td>
<td>26.8</td>
<td>50.9</td>
</tr>
<tr>
<td>MIROC monthly</td>
<td>3.24</td>
<td>21.26</td>
<td>30.4</td>
<td>61.6</td>
</tr>
<tr>
<td>MIROC 30y</td>
<td>3.15</td>
<td>22.22</td>
<td>25.6</td>
<td>57</td>
</tr>
<tr>
<td>NCAR CAPE</td>
<td>4.35</td>
<td>26.51</td>
<td>42.8</td>
<td>173.8</td>
</tr>
<tr>
<td>NCAR hourly</td>
<td>3.26</td>
<td>23.63</td>
<td>33.1</td>
<td>69</td>
</tr>
<tr>
<td>NCAR monthly</td>
<td>3.18</td>
<td>23.72</td>
<td>33.4</td>
<td>71.6</td>
</tr>
<tr>
<td>NCAR 30y</td>
<td>3.69</td>
<td>23.33</td>
<td>32.7</td>
<td>68.8</td>
</tr>
<tr>
<td>NORESM CAPE</td>
<td>4.6</td>
<td>18.63</td>
<td>36.5</td>
<td>74</td>
</tr>
<tr>
<td>NORESM hourly</td>
<td>2.88</td>
<td>21.25</td>
<td>27.7</td>
<td>64.8</td>
</tr>
<tr>
<td>NORESM monthly</td>
<td>2.94</td>
<td>21.06</td>
<td>29.2</td>
<td>66.9</td>
</tr>
<tr>
<td>NORESM 30y</td>
<td>2.91</td>
<td>22.2</td>
<td>29.1</td>
<td>66</td>
</tr>
<tr>
<td>Composite</td>
<td>2.89</td>
<td>23.71</td>
<td>27.9</td>
<td>66.3</td>
</tr>
<tr>
<td>Control</td>
<td>3.86</td>
<td>27.21</td>
<td>30.5</td>
<td>121.5</td>
</tr>
</tbody>
</table>
Figure 38: 18 UTC 10-m wind direction across the WRF computational domain, for the control and all PGW-delta experiments.

Figure 39: 18 UTC 2-m specific humidity across the WRF computational domain, for the control and all PGW-delta experiments.
Figure 40: Average hydrometeor mixing ratio within updrafts greater than 25 m s\(^{-1}\).
3.4: GCM RESPONSE COMPARISONS

Of the five GCMs utilized in this study, four (GFDL, MIROC, NCAR, and NORESM) were examined by Seeley and Romps (2015). Two (GFDL and NORESM) were designated as “high-performing” models for continued similarity to reanalysis data during March-August, though all four exhibited similar correlation scores during March-May (Seeley and Romps, 2015). Based on projected changes to CAPE and vertical wind shear, the GFDL produced the largest increase in severe thunderstorm environments (Seeley and Romps, 2015), a result which was mirrored here with the GFDL producing the strongest convection and proxy tornadoes.

When ranked by UH exceedances and maximum values (i.e. tornado proxy occurrence and intensity), the GCMs of this study are positioned as follows: GFDL, NCAR, NORESM, MIROC, and IPSL. Each UH category and delta type were given equal weight during the ranking process. To explain the differences seen between GCMs, atmospheric variables measured prior to convective initiation were averaged across delta type for each GCM. The GCMs were then ranked for each variable based on these averages. Variables that exhibited identical rankings to the UH criteria were low-level relative humidity and 0-1 km SRH. For reference, the GFDL had the highest low-level relative humidity and 0-1 km SRH, while the IPSL had the driest boundary layer and weakest 0-1 km SRH. CIN and LCL height, variables which are detrimental to tornadogenesis when large, exhibited the inverse ranking. This makes physical sense, as greater boundary layer relative humidity would yield lower LCLs and less CIN. Higher LCLs with drier air beneath also yields greater evaporative cooling and thus colder convective outflow. This has been shown to be detrimental to tornadogenesis (Rasmussen and Blanchard, 1998). The importance of boundary layer humidification has been noted in previous studies as well (Trapp et al., 2007a; Trapp et al., 2009; Diffenbaugh et al., 2013; Seeley and Romps, 2015). Lastly, the
relationship between UH response and 0-1 km SRH of each GCM is not surprising given the importance of low-level kinematic fields to tornadogenesis. It has been shown that 0-1 km vertical wind shear is a useful discriminator between significantly tornadic and nontornadic supercells (Thompson et al., 2003). Thus, it appears that the low-level humidification and kinematic projections were the primary drivers of inter-GCM response differences.

3.5: DELTA RESPONSE COMPARISONS

3.5.1: 30-YEAR VERSUS 10-YEAR AVERAGED DLTAS

Different climate change deltas constructed from averaging periods of 10 and 30 years were utilized in this study. Liu et al. (2017) argue that 30y deltas capture more climate change signal and less interannual variability, such that they are better suited for PGW experimentation. As discussed in Chapter 3.1, statistical similarity in distributions of CAPE from these two averaging periods suggest their variability is similar. Nonetheless, the averaging period lengths do influence the magnitude of the deltas, and are thus worthy of further comparison.

Beginning with simulated reflectivity, 10y deltas resulted in more 40 dBZ exceedances than the 30y deltas in three of the five GCMs (Fig. 14). On average, 30y deltas led to 2.9% less convection. However, when the reflectivity benchmark was raised to 55 dBZ to represent more intense convection, 30y deltas exceeded the 10y in four of the GCMs (Fig. 15). With a mean 30y minus 10y difference in intense convection of 12.5%, it appears that the delta averaging period has some significance in 1-km simulated reflectivity response, especially at higher thresholds.

Updraft velocity exceedances also illuminate dissimilarities in the responses of 30y and 10y deltas. At the 25, 40, and 50 m s\(^{-1}\) thresholds, the 10y delta simulations updrafts were more
prevalent than those of the 30 y deltas in three of the five GCMs (GFDL, IPSL, and NCAR) (Figs. 21-23). The mean 30 y minus 10 y difference at each threshold was -4.1%, -0.7%, and -23.3%, respectively. However, the range of percentage differences was broad at each threshold. At 25 m s\(^{-1}\), 30 y minus 10 y differences ranged from -35.8% to 51.7%. At 40 m s\(^{-1}\), they varied from -47.1% to 71%. Not until the 50 m s\(^{-1}\) threshold did GCM consensus and mean difference clearly suggest a negative 30 y minus 10 y difference, with values ranging from -70.5% to 15.4%. These results suggest that the 10 y deltas elicit a stronger updraft response, especially in high-end updrafts. This finding seems to contradict that from the convective activity based upon simulated reflectivity, where 30 y deltas produced a stronger convective response. However, this may be partially explained by considering temperature changes aloft. The 10 y deltas have relatively warmer air at all levels throughout the mid-troposphere, which results in a relatively higher melting level. Hail or any other hydrometeor must then fall farther through above-freezing temperatures, which increases melting/evaporation and decreases hydrometeor size. Because of smaller hydrometeor size, simulated radar reflectivity decreases.

Regarding UH response, 30 y deltas from three of the five GCMs produced less frequent UH at 500 m\(^2\)s\(^{-2}\), and also produced lower maximum UH values (Figs. 26 and 28). At the 1000 m\(^2\)s\(^{-2}\) threshold, the 30 y deltas of three GCMs produced more frequent high UH values compared to their 10 y counterparts (Fig. 27). Despite GCM consensus generally indicating a weaker UH response with 30 y deltas, these differences were smaller, such that the mean 30 y minus 10 y difference was positive at each threshold, with values ranging from +2.7% to +47.1%. The differing signals of mean response and ensemble consensus across thresholds suggest that there is no clear difference in tornado potential based on UH thresholds between 30 y and 10 y deltas.
Spatial patterns in accumulated precipitation appear to be relatively consistent between 30y and 10y deltas (Fig. 32). The consensus and mean response diverge at the 5 mm threshold, with three GCMs indicating more widespread precipitation in 10y deltas, but a mean 30y minus 10y difference of +3% (Fig. 33). At larger thresholds, consensus and average response began to align, and the range of PGW responses shifted towards smaller values. Again, 30y exceedances of 50 mm were fewer in three GCMs, but the mean response also showed a decrease of 4.7% (Fig. 34). Differences at 50 mm varied from -28.8% to 15.5%. Instances of extreme precipitation (i.e. 100 mm; Fig. 35) were 20.5% fewer in 30y deltas. Percentage differences for 30y minus 10y ranged -72.4% to 17.6%, as again three GCMs demonstrated a stronger response in 10y deltas. Thus, the 10y deltas generally induced a stronger precipitation response, especially at greater thresholds.

Comparisons between simulations with 30y and 10y deltas suggest that the overall convective responses are not especially different. Differences became more significant at higher thresholds, with 10y deltas procuring a stronger thermodynamic and precipitation response. This makes sense intuitively, as temperature and moisture differences were larger in the 10y deltas. Boxplots summarizing the response differences can be found in Fig. 41.
3.5.2: DIURNALLY-VARYING VERSUS TIME-CONSTANT DELTAS

Prior PGW studies have employed a time-independent delta based on monthly-averaged data. While this adequately captures the basic ACC signal derived from the GCM output, it omits potential climate-change-induced alterations to the diurnal cycle. For example, observed trends and GCM projections suggest warmer nights in a future climate (Schoof & Robeson, 2016; Li et al., 2018). Changes to the diurnal cycle may influence the timing of convective initiation and decay. Diurnally varying (DV) 10y deltas were created to investigate any projected diurnal cycle changes. Here, the response of these deltas is compared to time-constant (TC) deltas constructed from the same 10y periods.

Simulated reflectivity differences were minor at the 40 dBZ threshold. While four GCMs indicated an increase in convection occurrence with TC deltas, the mean TC-DV difference was only +3.5% (Fig. 14). At the 55 dBZ threshold, the mean difference in occurrence of stronger convection increased to +12.5%, with again four GCMs supporting this trend (Fig. 15). These
trends can at least be partially explained by differences in precipitable water, which were 0.2% higher in TC simulations. Considering that radar reflectivity factor is dependent on the sixth power of hydrometeor diameter, small changes in hydrometeor size can have large effects on simulated reflectivity. Otherwise, given the parity between TC and DV environments, these differences are most likely just noise, especially at 55 dBZ where exceedances are fewer.

Responses in updraft velocity were also similar between TC and DV deltas. Mean TC-DV differences were +2.1% and +3.2% at the 25 and 40 m s\(^{-1}\) thresholds, respectively (Figs. 21 and 22). The range of differences was relatively small, with values less than +/- 11%. At 50 m s\(^{-1}\), TC deltas produced on average nearly 25% fewer occurrences (Fig. 23). GCM consensus supported this as well, with DV deltas having more of the stronger updrafts in all five GCMs. While the percent difference magnitude may be influenced by the relatively small number of 50 m s\(^{-1}\) occurrences, the unanimous GCM agreement on the sign of change may be explained by subtle instability differences. The mean CAPE response was 0.8% smaller in TC deltas (Table 7), which corresponds to a difference of less than 100 J kg\(^{-1}\). Perhaps more importantly, CIN was 44.5% larger in the TC deltas on average (Table 7). This represents a change of less than 4 J kg\(^{-1}\) at 18 UTC (when calculations were performed), but differences may have been larger in the overnight hours when CIN magnitude increases.

Unlike updraft velocity, UH exceedances at every threshold were larger in TC delta simulations than in DV delta simulations. As thresholds increased, so did the mean TC-DV difference. Occurrences of 500 m\(^2\) s\(^{-2}\) were almost 9% more frequent with TC deltas (Fig. 26). The greatest difference occurred at the 1000 m\(^2\) s\(^{-2}\) threshold, where TC deltas had 104% more UH exceedances than DV deltas (Fig. 27), although the large percent change at this highest threshold had higher uncertainty due to the limited number of values. However, the consistency
of mean response and GCM consensus across thresholds suggest that there is physical reasoning
behind the differences. Surprisingly, TC deltas had more potentially-tornadic mesocyclones
based on UH exceedances, but the strongest updrafts were more common in DV deltas. Thus, it
appears that the disparities in UH, particularly at the upper thresholds, are primarily driven by
kinematic factors. Vertical vorticity was increased in TC updrafts, as 0-1 and 0-3 km SRH values
in TC simulations were 1.8% and 4.2% higher, respectively (Table 8). Indeed, when examining
instances of UH exceedance, average 2-5 km vertical vorticity in TC simulations was greater in
at least three GCMs. The GCMs in which the average vorticity was greater varied across
thresholds, such that TC simulations exhibited greater 2-5 km vorticity in every GCM at least
once. It is not clear whether the differences in SRH were the sole cause of UH discrepancies, but
they appear to at least be a contributing factor, especially at lower UH thresholds.

Accumulated precipitation responses were similar between the two delta types. Amongst
all thresholds, percentage differences ranged from -20.5% to 39.3%. Despite the breadth of
responses, mean TC-DV differences were +4.8%, +3.9%, and -1.7% at the 5, 50, and 100 mm
thresholds, respectively (Figs. 33-35). Slight increases in TC precipitable water may account for
the differences at lower thresholds, but do not explain the decrease in 100 mm occurrence. This
might be explained by the fewer number of 50 m s⁻¹ TC updrafts and ~1% faster average storm
motion. Nonetheless, precipitation differences appear to be relatively small.

Overall, most differences in TC and DV delta responses are quite small. Given the
environmental similarities, this is no surprise. Boxplots summarizing the response differences
can be found in Fig. 42. The more significant responses in UH and greater reflectivity may at
least partially attributed to subtle differences in SRH and precipitable water. It is worth
reiterating that the magnitudes at higher thresholds are greatly influenced by the limited number
of occurrences. However, GCM consensus supports the sign of these changes, and thus suggests that adding a time-dependence to PGW deltas will influence the response.

![Figure 42: Response differences in exceedances and maximum values of simulated reflectivity, updraft velocity, updraft helicity, and accumulated precipitation. Differences are noted as percent changes relative to the 10y DV deltas.](image)

### 3.5.3: CAPE DELTAS

Experimentation with a new delta type was conducted in this study. Constructed from the 90th percentile days of a 10y CAPE and CIN distribution, these CAPE deltas were intended to promote the most extreme climate change convective response. Other delta types include days that are unfavorable for the development of deep convection, whereas CAPE deltas only utilize 15 future and 15 historical days that were most conducive for convective storms (having the most CAPE, and least CIN). This led to highly variable and amplified delta distributions, as mentioned in Chapter 3a. Because they have a time-dependence and use the same 10y future and historical periods, the CAPE delta response is compared to that of the 10y DV deltas.

Surprisingly, in most GCMs, the CAPE deltas produced a weaker simulated reflectivity response. Exceedances of 40 dBZ were fewer in four GCMs, with a mean CAPE-10y difference of -30.7% (Fig. 14). When raised to 55 dBZ, exceedances due to CAPE deltas were less in three
GCMs, but with an average difference of +11.1% (Fig. 15). The positive mean was heavily skewed by a +92.5% response in the NORESM because of earlier initiation and more widespread convection. Despite having less convection, the GFDL CAPE delta produced the strongest convection of all PGW simulations. It was the only CAPE delta to produce more occurrences of updraft velocity and UH at all thresholds (Figs. 21-23, 26, 27). For reference, mean CAPE-10y differences at all updraft velocity and UH exceedance thresholds varied from -25.2% to -37.8%. The GFDL CAPE delta environment was characterized by +105.8% and +34.1% increases in 0-1 and 0-3 km SRH, respectively, relative to the 10y DV (Table 8). While CIN did increase 160.5% (~1.25 J kg\(^{-1}\)), CAPE surged by 13.1% (~600 J kg\(^{-1}\); Table 7). These changes greatly improved the mesoscale environment, which resulted in the strongest PGW storms. However, in the other GCMs, instability decreased in CAPE delta simulations. The mean increase in CIN and decrease in CAPE seen in the CAPE delta runs seem to be the most plausible causes of fewer and weaker storms, as kinematic parameters increased. Because of fewer and weaker storms, CAPE deltas also produced less precipitation. Average CAPE-10y differences were -26.6%, -24.9%, and -35.8% at the 5, 50, and 100 mm thresholds, respectively (Figs. 33-35). Despite having the strongest convection, the GFDL produced the fewest 100 mm exceedances, likely due to a 21.8% faster u-component of storm motion.

With the exception of the GFDL, the goal of extracting the most extreme climate change response failed with the CAPE deltas. Boxplots summarizing the response differences can be found in Fig. 43. However, perhaps the true response is that future climates will be increasingly detrimental to organized, deep convection. To achieve a more definite result, it is recommended that more GCMs be utilized and a new CAPE delta construction technique that incorporates
more days be employed. This should promote a more-balanced delta with a cleaner climate-change signal.

![CAPE-10y DV Difference in Storm Parameters](image)

Figure 43: Response differences in exceedances and maximum values of simulated reflectivity, updraft velocity, updraft helicity, and accumulated precipitation. Differences are noted as percent changes relative to the 10y DV deltas.

### 3.5.4: COMPOSITE DELTA

Creating a composite climate-change delta has become a common practice in recent PGW studies (Liu et al., 2017; Poujol et al., 2020; Nayak & Takemi, 2020). This practice saves computational resources by combining output from multiple GCMs into a single set of deltas, thus only needing one PGW simulation. While efficient, it potentially masks the range of possible climate change responses. The viability of a 10y DV composite delta was assessed by comparing its response to the mean response of the 10y DV deltas.

Beginning with simulated reflectivity, a 21 UTC comparison plot shows that the composite delta spatially captures the mean convective response near initiation (Fig. 44). Even at 01 UTC, the composite delta reflectivity was spatially close to that of the average response (Fig. 45), but averaging the responses of the five 10y DV simulations results in a loss in convective intensity. The response of the composite delta was greater than the mean response by 20.7% and
15.6\%, for the convection and intense convection thresholds, respectively (Figs. 14 and 15). Similarly, spatial patterns of accumulated precipitation in the composite delta and mean response were comparable (Fig. 46). When evaluated quantitatively, however, the response of the composite delta was comparatively larger at the upper and lower thresholds. Composite-mean differences at 5, 50, and 100 mm thresholds are +25.1\%, +6.9\%, and +25.3\%, respectively.

The largest discrepancies arose when comparing the strength of the updrafts of the simulated convective storms. While relatively close at the 25 m s\(^{-1}\) threshold (+8.1\%; Fig. 21), the composite-mean difference became more significant at 40 m s\(^{-1}\) (-19.9\%) and 50 m s\(^{-1}\) (-68.6\%; Figs. 22 and 23). A similar trend in differences was seen in UH comparisons. The composite delta response was within five percent of the mean (-4\%) at 500 m\(^2\) s\(^{-2}\), but much less at the 1000 m\(^2\) s\(^{-2}\) (-59.9\%) threshold (Figs. 26 and 27). Even the maximum UH value was 25.2\% lower than the mean response (Fig. 28). These differences were seen despite slightly higher CAPE (+0.2\%) and 0-3 SRH (+3.6\%) values in the composite delta (Tables 7 and 8).

Comparisons between the composite delta and mean response illustrate the importance of having a PGW ensemble. It is clear that the response of a mean input is not equivalent to the mean of several responses, as the atmosphere is a non-linear system. Boxplots summarizing the response differences can be found in Fig. 47. Taking an ensemble approach to PGW methodology allows researchers to display a range of possible outcomes along with the associated uncertainty. Using a single, composite delta provides only one solution, with no reference point as to where this solution falls on the spectrum of possible solutions, and thus is not recommended for PGW considerations.
Figure 44: 1 km simulated reflectivity of the composite 10y DV delta (left panel) and the mean 10y DV response (right panel) at 21 UTC.
Figure 45: 1 km simulated reflectivity of the composite 10y DV delta (left panel) and the mean 10y DV response (right panel) at 01 UTC.

Figure 46: Total accumulated precipitation of the composite 10y DV delta (left panel) and the mean 10y DV response (right panel) at 21 UTC.
Figure 47: Response differences in exceedances and maximum values of simulated reflectivity, updraft velocity, updraft helicity, and accumulated precipitation. Differences are noted as percent changes relative to the mean response of the 10y DV deltas.
CHAPTER 4: SST EXPERIMENTS WITH A COLD-SEASON EVENT

Recent studies have examined the importance of Gulf of Mexico SSTs to cool-season tornado outbreaks (Molina et al., 2018; Molina and Allen, 2019; Molina et al., 2020). To assess this relationship from a PGW-like framework, a series of simulations were run with uniformly perturbed SSTs. Perturbations ranged from -5°C to +5°C, comparable to the magnitude of end-of-the-century Gulf of Mexico warming projected by GCMs: Average increases in Gulf of Mexico SSTs in the GFDL and NCAR were +4.2°C and +3.0°C, respectively. Here, storm and environmental parameters are compared among seven simulations of the 10 February 2013 (cold-season) tornado outbreak. This outbreak was characterized by tornadic supercells that had initiated ahead of a quasi-linear convective system (QLCS) in the Southeastern U.S.

Simulated reflectivity at 19 UTC shows substantial differences in event evolution and intensity among the seven simulations (Fig. 48). Simulations with cool SST perturbations featured a slower and weaker QLCS than those with warm SSTs. Ahead of the QLCS, the warm-SST simulations generally had more cellular convection. To quantify these and other differences in storm characteristics, grid point exceedances of variables with various thresholds were computed. The quantitative analysis of reflectivity and other storm variables was conducted with location in mind, as convection over the open ocean may have been due to the rapid, artificial increase in low-level lapse rates in the warm-SST simulations. To keep this convection from skewing the inland results, only locations north of 30°N latitude were considered (see Fig. 48). It should be noted, however, that when the entire domain was included in the quantitative analyses, results and trends improved.
When summed over the time of the outbreak (10 February 12 UTC – 11 February 06 UTC), statistically significant trends were noted in convective activity as represented by reflectivity exceedances at various thresholds (Fig. 49b). The warm-SST simulations had nearly 100% more inland convection than those performed with cool SSTs (Fig. 49a). Changes in storm coverage from warm SSTs were not proportional to those from cool SSTs (Fig. 49). This can be partially explained by a simulation time that was too short for maximum environmental modification by the warm SSTs (discussed later); however, there were also convective interactions that influenced this result. In the +5°C simulation, a complex of storms concentrated off the southern coast of Louisiana (south of 30°N) disrupted the environment to the north (Fig. 48). This acted to suppress the QLCS and any warm-sector convection in that area, thus reducing convective activity.

Updraft velocity and UH were the primary measures of storm strength used in this study. Peak updraft velocity changed little between simulations (Fig. 50c). The largest change in updraft velocity resulted from the 2nd coolest and 2nd warmest runs, with values of 20 m s⁻¹ and 31 m s⁻¹, respectively. Acknowledging the relatively small sample size represented by these experiments, these differences were not proportional to changes in CAPE and were not statistically significant at the 95% confidence level. However, there was a significant trend in updraft velocity exceedances of 15 m s⁻¹ (Figs. 50a, 50b), generally increasing with warmer SST perturbations. Likewise, UH swath area increased with warmer SSTs (Fig. 51). When UH was examined through the tornado proxy threshold of 50 m² s⁻² (which is different here owing to the coarser grid point spacing), it was much more prevalent in the warm-SST simulations (Fig. 52a). Unlike updraft velocity, the trend in UH exceedances north of 30°N was not statistically significant due to a large decrease in the +5°C simulation (Fig. 52b). Because of the large
concentration of convection and of UH swaths south of 30°N, the +5°C simulation had fewer tornado proxy occurrences than the other warm runs. Despite this, increases in tornado proxy were better correlated with SSTs than peak UH values (Fig. 52c), in agreement with the modest correlation in peak updraft velocity (Fig. 50c). Together, these results seem to indicate that convective coverage over this region, rather than intensity, is more sensitive to SST perturbations in the Gulf of Mexico.

Environmental variables exhibited positive correlations with SST perturbations, and can be used to explain the differences in the convective-scale response just shown. Inland 2-meter temperatures ahead of the QLCS were anomalously warm (relative to the control simulation) when positive SSTs modifications were applied (Fig. 53a). Conversely, cool SSTs led to cooler temperatures in this region (Fig. 53a). Careful interpretation of temperature, dewpoint, and CAPE spatial plots is important, as the control simulation had a different pre-outbreak evolution than the perturbed runs. The difference in evolution is suspected to be the reason for anomalously high temperatures, dewpoints, and CAPE in southern Mississippi and southwestern Alabama in the cool simulations (Figs. 53a, 54a, 55a). Thus, the magnitude of these anomalies is examined with respect to the other perturbed simulations. Differences in peak temperatures at Hattiesburg, MS, were nearly 5°C between the warmest and coldest simulations (Fig. 53b). The nonproportional changes in 2-meter temperatures relative to SSTs were likely the result of inland heating, as well as the relatively short simulation time that allowed the SSTs to modify the overlying air mass. The latter is likely to have played a role in other environmental trends, as will be discussed later. Linear correlation between SSTs and inland temperatures proved to be statistically significant (Fig. 53b), though the sample size is limited here. The 2-m dewpoint temperatures were better correlated with SSTs than air temperatures. Positive dewpoint
anomalies existed ahead of the QLCS in the warm simulations (Fig. 54a). When SSTs were cooled, these anomalies became negative (Fig. 54a). Peak dewpoint temperatures at Hattiesburg generally increased as SSTs were warmed (Fig. 54b). However, the maximum dewpoint temperatures increased very little beyond +1°C SST perturbation. This likely indicates that the simulation time prior to the outbreak was not long enough. Despite this, dewpoint temperatures exhibited a statistically significant linear trend with the SSTs, having a Pearson correlation coefficient of 0.846 (Fig. 54b).

With the thermodynamic profile of the free atmosphere largely constrained via spectral nudging, changes to CAPE values in the experiments can be attributed to changes in 2-meter temperature and dewpoint. Thus, it is not surprising to see increases in CAPE with increases in the SST perturbations (Fig. 55). East of the QLCS, the most unstable CAPE (MUCAPE) anomalies become smaller and eventually negative as SSTs were cooled (Fig. 55a). Again, the positive anomalies in some of the cooler-SST simulations are believed to be the result of a different pre-outbreak evolution rather than an actual increase in instability. The decrease in MUCAPE anomaly magnitude with cooling SSTs supports this claim (Fig. 55a). When looking at Hattiesburg, peak MUCAPE increases by nearly 1500 J kg⁻¹ from the coolest to warmest simulation (Fig. 55b). However, MUCAPE is not necessarily a good indicator of instability for tornadic environments, as the most unstable parcel may be rooted above the boundary layer. Indeed, when surface-based CAPE (SBCAPE) is inspected, the increase in instability is even larger at almost 2000 J kg⁻¹ (Fig. 55c). These increases in CAPE likely played a significant role in the more frequent updraft and UH exceedances. As with temperature and dewpoint, the peak CAPE values leveled off after +1°C of warming (Figs 55b, 55c). Again, this suggests that the simulations were likely too short. Figure 55c also illustrates the large difference between the
control and the perturbed simulations. SBCAPE is smaller than the coldest SST simulation and is nearly 0 J kg\(^{-1}\) in the control run. Again, this is believed to be tied to the differences in pre-outbreak evolution. Despite the evolutionary differences, changes in both MUCAPE and SBCAPE were statistically significant.

It is believed that had there been more model integration time prior to the event, the perturbed SSTs would have had more time to precondition the overlying atmosphere to a greater extent. This would have likely bolstered the linear relationships between SST perturbations and the thermodynamic variables analyzed here. Nonetheless, it is clear that warmer SSTs produce a more favorable thermodynamic environment than the control. These enhanced environments yield convection and tornadoes that are not only more widespread, but also more intense. Conversely, cooler SSTs generated diminished thermodynamic environments, weaker convection, and fewer tornadoes.
Figure 48: 1 km simulated reflectivity for the CTRL and SST perturbation simulations at 19z. Red line in control plot represents 30°N latitude.
Figure 49: a) Cumulative reflectivity exceedances of 40 dBZ (North of 30°N from 12z-06z). b) Scatter plot of (a).
Figure 50: Cumulative updraft velocity exceedances of 15 m/s (North of 30°N from 12z-06z). b) Scatter plot of (a). c) Scatter plot of peak updraft velocity (North of 30°N from 12z-06z).
Figure 51: Updraft helicity swathes from 12z-06z.
Figure 52: Cumulative tornado proxy (UH > 50) density plot. b) Scatter plot of cumulative tornado proxy occurrences (North of 30N from 12z-06z). c) Scatter plot of peak updraft helicity (North of 30N from 12z-06z).
Figure 53: a) 2-meter temperature anomalies relative to the CTRL. b) Peak temperature at Hattiesburg, MS from 12z-06z.
Figure 54: a) 2-meter dewpoint temperature anomalies relative to the CTRL. b) Peak dewpoint temperature at Hattiesburg, MS from 12z-06z.
Figure 55: a) MUCAPE anomalies relative to the CTRL. b) Peak MUCAPE at Hattiesburg, MS from 12z-06z. c) Peak SBCAPE at Hattiesburg, MS from 12z-06z.
CHAPTER 5: CM1 IDEALIZED PGW SIMULATIONS

Cloud Model 1 (CM1; Bryan and Fritsch, 2002) was utilized to produce a series of idealized PGW simulations of the 20 May 2013 and 10 February 2013 events. Idealized modeling provides an alternative, and perhaps better assessment of the PGW response than the WRF simulations because a the horizontally-homogenous atmosphere in CM1 means that changes in convective characteristics can be attributed only to local environmental changes. A “translating” domain was used to permit a smaller domain size, and thus finer grid spacing. With a 64 m horizontal grid spacing within the interior portion of the domain, tornado-like vortices can be nominally resolved. This meant that the tornadic potential of these events in a future climate could be better assessed. With the prescribed updraft nudging present for the first 20 minutes of the simulation, analyses begin at minute 30 to limit potential contamination by residual effects of the nudging. The results of these idealized PGW simulations are presented below.

5.1: 20 MAY 2013 CM1 SIMULATIONS

Beginning with simulated reflectivity, variations in supercell evolution and intensity existed between simulations (Figs. 56 and 57). Storm size was quantified by the number of 40 dBZ exceedances at grid points south of 20 km in the domain, to prevent any of the left-moving supercell values from contaminating the results. Time series of 40 dBZ exceedances reveal that all PGW supercells were smaller than the CTRL supercell (Fig. 58). Compared to the CTRL, the maximum area of the simulated future storms was 28.6% smaller on average. Also evident in the reflectivity time series was the earlier dissipation of the PGW supercells. Exceedances of 40 dBZ
begin to decrease and diverge from the CTRL between minutes 60-80, contrasting the longer-lived PGW convection seen in the WRF simulations. The weakening of convection during this time is corroborated with examination of simulated reflectivity plots.

Updraft cores in the PGW simulations were notably smaller as well. Updraft core width was derived from the number of grid points exceeding the 99th percentile of mean 3-7 km updraft velocities in the CTRL. For the 20 May case, this was 9.39 m s\(^{-1}\). These grid points were assumed to represent a circular updraft core area from which diameter could be computed. Calculations were performed with mean 3-7 km updraft velocities. Time series of updraft width highlight the continuously smaller diameter of PGW updrafts, apart from the NCAR simulation which briefly exceeded the control (Fig. 59). PGW updrafts also experience an earlier demise, which explains the trends seen in simulated reflectivity. When evaluating maximum updraft velocities at all levels, the control exhibits stronger updrafts at most analysis times (Fig. 60). Beyond minute 60, even the most resilient PGW updrafts begin to fade while the control is relatively consistent. The weaker updrafts seen in the future environments appear to be at least partially due to increased precipitation loading. Hydrometeor mixing ratios within updraft cores were 49.4% larger in PGW environments (Fig. 61). This increase was driven by greater precipitable water values (Fig. 62). The additional weight and drag of these hydrometeors reduced the vertical velocities of PGW updrafts.

Tornado potential of CTRL and PGW simulations was evaluated through the metric of tornado power. This metric describes surface energy dissipation, and is based on Fricker et al. (2017), who applied it to observed tornadoes. Tornado power accounts for both tornado wind speed and width. It can be thought of as a parallel to accumulated cyclone energy (ACE) with tropical cyclones. Tornado power is represented mathematically with the following equation,
\[ Tornado\ Power = \sum \pi r^2 \rho V^3 \quad (7) \]

where \( r \) represents the average radius of maximum winds, \( \rho \) is the air density (assumed to be 1 g kg\(^{-1}\)), \( V \) is the maximum surface wind speed, and the summation is taken over the lifetime of the tornado. Tornado power was calculated using only the strongest tornado-like vortex (TLV) present at each analysis time. TLVs were identified by examining surface fields of vertical vorticity and Okubo-Weiss (OW) parameter (Markowski et al. 2011). To be considered a TLV, vertical vorticity and OW had to exceed 0.1 s\(^{-1}\) and 0.03 s\(^{-2}\), respectively, and be collocated with vertical velocities greater than 5 m s\(^{-1}\) over more than 12 grid points in the lowest 2 km (similar to Sherburn and Parker, 2019). The TLVs were not present at all times during simulations. Upon locating the strongest TLV, maximum and minimum of \( u \) (i.e., \( x \)-direction) and \( v \) (i.e., \( y \)-direction) wind components were found within 500 m of the vortex center. The locations of these maxima and minima were used to determine an average radius of maximum winds. The maximum ground-relative wind speed at the surface was also found within this search radius, thus providing a value for \( V \) in Eq. 7. Tornado power was then summed over all analysis times to yield total tornado power.

Calculations of tornado power reveal that four of the five PGW simulations exhibited greater tornado power than the CTRL (Fig. 63). On average, tornado power increased by +86.6% in the future environments. In these four experiments, average instantaneous tornado power was greater at each analysis time with a TLV present. The mean PGW response in average instantaneous tornado power was +108.8%. This indicates that increases in tornado power were driven by stronger or wider tornadoes, and not longer-lived or more numerous tornado segments. In fact, only one PGW simulation (IPSL) featured more analysis times with a TLV present. Increases in tornadic wind speeds occurred in all PGW experiments with greater tornado power.
The average change in tornadic winds was +11.9% relative to the control. Wider TLVs were only found in the GFDL and MIROC simulations, although the mean PGW response was a +6.2% increase in vortex width. Thus, enhancement of TLV wind speeds was the primary contributor to increased tornado power in future climates.

All PGW environments experienced an increase in CAPE and/or SRH, both of which are believed to be important contributors to tornado intensity. Mean and median PGW responses in CAPE were nearly +5% (Fig. 62). Three of the five PGW simulations had greater 0-1 km and 0-3 km SRH, with average increases were +2.8% and +2.6%, respectively. However, it is unclear why the NCAR simulation produced less tornado power given the slight increases in CAPE and SRH. This could be a result of the TLV detection algorithm, but time series of maximum near-surface wind speeds and vertical vorticity show the NCAR simulation frequently at the lower end of the PGW spectrum, especially at earlier analysis times (Fig. 64). Nonetheless, increases in CAPE and/or SRH in the other PGW simulations were sufficient to overcome other environmental changes that were detrimental. Decreased boundary layer relative humidity resulted in higher lifted condensation levels (LCL) and levels of free convection (LFC) (Fig. 62). LCL and LFC heights were raised 50% and 72.7%, respectively. A higher LCL height has been shown to be less favorable for tornadogenesis, as it supports greater evaporational cooling below the cloud base and thus stronger outflow (Rasmussen and Blanchard, 1998). As a result of higher LFCs, convective inhibition (CIN) increased in all future environments. Changes ranged from +7% to +264%, with a mean response of +97% (Fig. 62). This contributed to the earlier demise of the future supercells, as shown in Fig. 58.
Figure 56: CM1 1 km simulated reflectivity for the CTRL and all Moore PGW simulations at minute 40.
Figure 57: CM1 1 km simulated reflectivity for the CTRL and all PGW simulations at minute 60.

Figure 58: Time series of gridpoint exceedances of 40 dBZ simulated reflectivity for the CTRL and all Moore PGW simulations.
Figure 59: Time series of updraft core width for the CTRL and all Moore PGW simulations using the 99th percentile of positive vertical velocities.

Figure 60: Time series of maximum vertical velocity for the CTRL and all Moore PGW simulations. Dashed vertical line represents the start of analysis period.
Figure 61: Mean hydrometeor mixing ratios of within updraft cores as defined Chapter 5.1. Hydrometeors included in the calculation were rain, graupel, and hail.
Figure 62: CTRL-relative percentage differences in most-unstable CAPE and CIN, lifted condensation level (LCL), level of free convection (LFC), 0-1 km and 0-3 km storm relative helicity (SRH), and precipitable water (PW) for all Moore PGW simulations.
Figure 63: CTRL-relative percentage differences in tornado power, average instantaneous power, average TLV velocity, and average TLV width for all Moore PGW simulations.
Figure 64: Time series of maximum near-surface vertical vorticity (top panel) and wind speeds (bottom panel) for the CTRL and all PGW simulations. Dashed vertical line represents the start of analysis period.
5.2: 10 FEBRUARY 2013 CM1 SIMULATIONS

Like the Moore case, the Hattiesburg PGW simulations featured supercells that were smaller and shorter-lived (Figs. 65 and 66). Time series of 40 dBZ exceedances show two distinctly different scenarios in the PGW storms (Fig. 67). The GFDL and IPSL simulations were clustered closely together throughout the analysis time, with significantly smaller storm sizes than the others. Size changed very little after nudging ceased, and substantial weakening began to occur as early as minute 45. In the MIROC, NCAR, and NORESM runs, storm size grew for 10-15 minutes after nudging had ended. While still smaller than the CTRL, these supercell evolutions mirrored that of the control until minute 60. At that point, the CTRL supercell increased in size while the PGW supercells continued weakening to their ultimate demise. At their peak sizes, the PGW supercells were 52.9% smaller than the CTRL supercell.

Updrafts in the PGW supercells were notably weaker than those of the CTRL supercell, especially beyond minute 33 of the simulations (Fig. 68). Additionally, time series of updraft core width, as derived from exceedances of the control simulation 99th percentile updraft velocity, show a stark difference in the CTRL and PGW simulations (Fig. 69). Again, GFDL and IPSL simulations had the smallest and briefest updraft cores. Even at their peak size, right after nudging had ceased, the MIROC, NCAR, and NORESM runs featured updraft widths that were approximately half that of the control. By minute 50, the updraft cores of these three simulations were practically nonexistent.

Despite the brevity of the PGW supercells, three of the five (GFDL, IPSL, and NCAR) PGW simulations produced more tornado power than did the CTRL (Fig. 70), including the two shortest-lived storms. These simulations also exhibited greater average instantaneous tornado power. However, upon closer examination, it was found that these increases were driven by
stronger but shorter-lived TLVs. In the GFDL, IPSL, and NCAR simulations, 49%, 57%, and 91% of the total power came from one analysis time (Table 9). The skewed nature of the NCAR tornado power distribution resulted in average TLV widths and wind speeds that were smaller than the control. The increase in total power seen in the GFDL simulation appears to be driven by a 30% increase in TLV width, as the average TLV wind speed decreased by 9.4%. With PGW consensus suggesting increases in tornado power but decreases in tornado wind speeds and width, these results imply the presence of stronger future tornadoes whose lives and/or increased intensity are short-lived.

It was quite surprising to see increases in tornado power in the PGW simulations given their thermodynamic environments. Similar to the May 20th simulations, boundary layer relative humidity decreased in future environments. This led to significantly higher LCLs and LFCs (Fig. 71). When combined with warmer temperatures aloft, CAPE was reduced by 31.3% on average. CIN increased by over 2000%, corresponding to an increase of 33.1 J kg⁻¹. These changes to CAPE and CIN values were likely the main reasons for the weaker and narrower updrafts seen in the February 10th PGW runs. Subtle increases in 0-1 km and 0-3 km SRH (Fig. 71) likely aided updraft rotation, and in turn tornadogenesis. However, decreases in both SRH parameters occurred in the IPSL environment, where tornado power was greater than the CTRL.

PGW experimentation within CM1 revealed that supercells in these future environments would likely have shorter lifespans due to increased convective inhibition. Similarly, TLVs produced by these storms were also briefer. However, these simulated tornadoes produced greater tornado power in most of the PGW experiments, primarily due to stronger vortex wind speeds. Increases in TLV intensity were likely influenced by increases in SRH and, in some PGW simulations, CAPE. The common denominator in many of the convective and
environmental changes seen in the PGW simulations was a drier boundary layer. Greater temperature-dewpoint spreads near the surface increased LCL heights, CIN, and sometimes actually reduced CAPE. This not only reduced storm longevity, but the higher LCL heights and resultant stronger outflow likely reduced tornado longevity. The findings here support those of the WRF simulations, with future environments producing fewer and/or shorter-lived tornadoes that are more powerful.

Figure 65: CM1 1 km simulated reflectivity for the CTRL and all Hattiesburg PGW simulations at minute 40.
Figure 66: CM1 1 km simulated reflectivity for the CTRL and all Hattiesburg PGW simulations at minute 60.

Figure 67: Time series of gridpoint exceedances of 40 dBZ simulated reflectivity for the CTRL and all Hattiesburg PGW simulations.
Figure 68: Time series of maximum vertical velocity for the CTRL and all Hattiesburg PGW simulations. Dashed vertical line represents the start of analysis period.

Figure 69: Time series of updraft core width for the CTRL and all Hattiesburg PGW simulations using the 99th percentile of positive vertical velocities.
Figure 70: CTRL-relative percentage differences in tornado power, average instantaneous power, average TLV velocity, and average TLV width for all Hattiesburg PGW simulations.
Table 9: Analysis of tornado power from the 20 May and 10 February 2013 CM1 experiments.

<table>
<thead>
<tr>
<th></th>
<th>CTRL</th>
<th>GFDL</th>
<th>IPSL</th>
<th>MIROC</th>
<th>NCAR</th>
<th>NORESM</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Tornado power</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(20 May 2013)</td>
<td>9.4E+09</td>
<td>2.91E+10</td>
<td>2.94E+10</td>
<td>1.45E+10</td>
<td>1.38E+09</td>
<td>1.33E+10</td>
</tr>
<tr>
<td>% of power</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>from peak</td>
<td>12.98%</td>
<td>32.13%</td>
<td>18.09%</td>
<td>18.68%</td>
<td>25.19%</td>
<td>21.03%</td>
</tr>
<tr>
<td>analysis time (20</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>May 2013)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Tornado power</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(10 Feb. 2013)</td>
<td>1.43E+10</td>
<td>1.82E+10</td>
<td>8.27E+10</td>
<td>1.42E+10</td>
<td>4.5E+10</td>
<td>4.84E+09</td>
</tr>
<tr>
<td>% of power</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>from peak</td>
<td>36.69%</td>
<td>49.28%</td>
<td>56.51%</td>
<td>36.90%</td>
<td>90.87%</td>
<td>47.95%</td>
</tr>
<tr>
<td>analysis time (10</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Feb. 2013)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure 71: CTRL-relative percentage differences in most-unstable CAPE and CIN, lifted condensation level (LCL), level of free convection (LFC), 0-1 km and 0-3 km storm relative helicity (SRH), and precipitable water (PW) for all Hattiesburg PGW simulations.
CHAPTER 6: CONCLUSIONS

Through the use of the pseudo-global warming (PGW) methodology, two historical tornado events were effectively simulated in environments influenced by anthropogenic climate change (ACC). Future projections from five GCMs were employed to create five climate-change deltas that varied in averaging period length, time dependence, and the convective potency of days used. All deltas were then utilized in high-resolution Weather Research and Forecasting model (WRF) simulations of the 20 May 2013 tornado outbreak, providing a 21-member PGW ensemble from which the event’s possible future intensity could be assessed. Additionally, WRF simulations of the 10 February 2013 tornado event were run with uniformly-perturbed sea surface temperatures (SSTs) across the Gulf of Mexico. With perturbation magnitude approximating end-of-the-century projections of GCMs, the influence of SSTs on this cool-season tornado outbreak was examined. Finally, the novel application of the PGW methodology to idealized cloud-resolving model simulations was used to better isolate the climate-change response in these two tornado events.

WRF simulations of the 20 May 2013 event suggest that more convective storms would occur under ACC. PGW simulations exhibited greater exceedances of 40 and 55 dBZ reflectivity occurrences, indicative of a greater area of deep convection and intense storms. Accumulated precipitation generally increased as well. More areas received measurable rainfall, and instances of extreme rainfall – more than 100 mm – increased by more than 50% on average. While updrafts exceeding speed thresholds of 25 m s\(^{-1}\), 40 m s\(^{-1}\), and 50 m s\(^{-1}\) were more numerous, maximum updraft velocities were not necessarily greater due to enhanced precipitation loading. Defined as updraft helicity (UH) exceeding 500 m\(^2\)s\(^{-2}\), instances of baseline tornado proxy
decreased in the PGW environments. However, increases in higher-end tornado proxy and maximum UH values indicate the possibility of fewer but stronger tornadoes under ACC. These tornado proxy observations hold true for the UH swath representative of the Moore supercell seen in ~85% of the PGW simulations. Thus, it appears that the 20 May 2013 tornado event would be more intense under ACC.

Delving deeper into the spectrum of PGW responses in the May 20th simulations revealed some discrepancies between delta types. The purpose of using several delta configurations was to address questions and concerns raised by previous PGW studies regarding optimal delta construction. A summary of the results includes:

- When comparing deltas of different length averaging periods, 10y deltas produced a stronger thermodynamic and precipitation response than the 30y deltas. However, changes were neither major nor unambiguous. Given similarities in their spatial patterns, as well as the parity in CAPE variability between averaging periods, there seems to be no obvious advantage in using 30y deltas.

- Differences in the responses of 10y diurnally varying (DV) and time constant (TC) deltas were relatively insignificant in most convective parameters. The largest differences arose in the upper tiers of UH thresholds, where the limited number of exceedances tended to exaggerate the percentage changes. Nonetheless, agreement between the mean UH response and GCM consensus suggests that adding a diurnal dependence to deltas has an influence on rotation of PGW storms.

- Despite capturing some of the spatial characteristics of the mean response, the composite delta failed to represent the intensity of the mean response. The use of a composite delta results in a loss of variability between GCM responses, which has been shown here to be
quite significant. Thus, it is advised that future PGW work utilizes an ensemble approach to retain some measure of uncertainty.

- Finally, deltas created solely from the most favorable days for deep convection resulted in the most polarized responses. While the GFDL CAPE delta produced the most extreme PGW response, CAPE deltas of the other four GCMs elicited the weakest response in most convective metrics.

By perturbing SSTs in the simulated 10 February 2013 tornado outbreak, it was concluded that warmer (cooler) SSTs over the Gulf of Mexico produce a more (less) favorable thermodynamic environment for convection across the southeastern United States. This resulted in more widespread and stronger storms, although increases in convective coverage, rather than intensity, were better correlated with SST perturbations. However, the relationships found here are not without their nuances and caveats. As found in the simulations of Molina et al. (2020), ongoing/widespread convection can contaminate the influence of SST perturbations, which may contribute to the non-linear relationship between SSTs and convective intensity. Additionally, the simulation time prior to the outbreak was likely insufficient for complete preconditioning of the mesoscale environment, especially in the warmer simulations. A longer integration time would have likely resulted in even more favorable environments in the +2.5°C and +5°C runs, and may have bolstered the trends found here. Lastly, the seven different simulations conducted here represents a rather small sample size. Thus, future work will feature more SST perturbations and longer simulations. Higher resolution grid spacing will also be considered. It is believed that this work can contribute understanding to seasonal-to-subseasonal forecasting as well as the influence of climate change on severe storms.
Finally, idealized, single-storm CM1 simulations of the 10 February 2013 and 20 May 2013 events reveal that despite being detrimental to supercell size and longevity, future environments produced increases in total tornado power. These increases were likely driven by subtle increases in SRH, as well as occasional increases in CAPE. In all instances of increased total power, average tornado power was greater as well, indicating the presence of stronger tornado-like vortices (TLVs). The amount of time with a TLV present was often equivalent or less than that of the control simulations, supporting the WRF simulation conclusion of fewer but stronger tornadoes.

While the conclusions presented here are believed to be robust, additional work is already being conducted to bolster these findings. Applying the traditional PGW approach, rather than SST perturbations, to WRF simulations of the 10 February 2013 event will provide a seasonal contrast to the conclusions of the May 20th WRF experiment. Also, including more GCM projections to expand the ensemble approach would provide an opportunity for a clearer consensus and uncertainty measurement in this PGW work. In the CM1 simulations, investigating if the PGW response was neither sensitive to perturbations in the initial conditions nor the forcing mechanism would supply additional confidence in the results shown here. Future experimentation with wind-only and thermodynamic-only deltas within CM1 will also be conducted. This will allow simulated storm changes to be better attributed to the competing kinematic and thermodynamic effects of climate change.
REFERENCES


NCEI billion-dollar events----- NOAA National Centers for Environmental Information (NCEI)

https://www.ncdc.noaa.gov/billions/, DOI: 10.25921/stkw-7w73


Tang, Brian & Gensini, Vittorio & Homeyer, Cameron. (2019). Trends in United States large hail environments and observations. npj Climate and Atmospheric Science. 2. 45. 10.1038/s41612-019-0103-7.


