SEVERE WEATHER DURING THE NORTH AMERICAN MONSOON AND ITS RESPONSE TO RAPID URBANIZATION AND A CHANGING GLOBAL CLIMATE WITHIN THE CONTEXT OF HIGH RESOLUTION REGIONAL ATMOSPHERIC MODELING

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ABSTRACT

The North American monsoon (NAM) is the principal driver of summer severe weather in the Southwest U.S. With sufficient atmospheric instability and moisture, monsoon convection initiates during daytime in the mountains and later may organize, principally into mesoscale convective systems (MCSs). Most monsoon-related severe weather occurs in association with organized convection, including microbursts, dust storms, flash flooding and lightning. The overarching theme of this dissertation research is to investigate simulation of monsoon severe weather due to organized convection within the use of regional atmospheric modeling.

A commonly used cumulus parameterization scheme has been modified to better account for dynamic pressure effects, resulting in an improved representation of a simulated MCS during the North American monsoon experiment and the climatology of warm season precipitation in a long-term regional climate model simulation. The effect of urbanization on organized convection occurring in Phoenix is evaluated in model sensitivity experiments using an urban canopy model (UCM) and urban land cover compared to pre-settlement natural desert land cover. The presence of vegetation and irrigation makes Phoenix a “heat sink” in comparison to its surrounding desert, and as a result the modeled precipitation in response to urbanization decreases within the Phoenix urban area and increase on its periphery.

Finally, analysis of how monsoon severe weather is changing in association with observed global climate change is considered within the context of a series of retrospectively simulated severe weather events during the period 1948-2010 in a numerical weather prediction paradigm. The individual severe weather events are identified by favorable thermodynamic conditions of instability and atmospheric moisture (precipitable water). Changes in precipitation extremes are evaluated with extreme value statistics. During the last several decades, there has been intensification of organized convective precipitation, but these events occur with less frequency. A more favorable thermodynamic environment for monsoon thunderstorms is the driver of these changes, which is consistent with the broader notion that anthropogenic climate change is presently intensifying weather extremes worldwide.
CHAPTER 1: INTRODUCTION

Severe weather is an extreme meteorological event or phenomenon that represents a real hazard (to human life and property). The definition of severe weather is often impact-based, and usually defined by thresholds. Severe weather occurs all over the world, but developing countries are more socially economically vulnerable to damages because of less well developed infrastructure. Specific severe weather phenomena vary, depending on the latitude, altitude, topography, and atmospheric conditions.

The focus region of this study is the tropical to subtropical latitudes, where localized convection is the most important factor in severe weather during the warm season. In contrast to the mid latitude, vertical motion is not due to quasi-geostrophic forcing. Rather, latent heat release within cumulus cloud is the main source energy for upward vertical motion (Krishnamurti et al., 2013). There is also strong interaction among various scales of cumulus convection from small air-mass thunderstorms to mesoscale convective systems that cover hundreds to thousands of square miles. How convection rose and organized is extremely important for severe weather forecasting for tropical to subtropical regions but presently not well understood. This dissertation work focused on organized, propagating convective systems in the region of the North American Monsoon (NAM). We demonstrate the importance of representing convective precipitation reasonably to answer important questions such as how urbanization effect monsoon thunderstorms and how monsoon thunderstorm maybe changing in associating with global scale climate change.

A background on the NAM climatology is briefly presented in section 1.1. Section 1.2 mentions literature for observed atmospheric condition and extreme precipitation trends due to climate change. Modeling severe weather during the NAM season is discussed in section 1.3 with regard to treatment of MCSs by convection scheme and treatment of the land surface changes due to urbanization in a desert city. And the last section 1.4 lists research objectives.

1.1 The North American monsoon
The NAM is the principal driver of summertime severe weather in the Southwest U.S. Increases in moisture, instability, and precipitation during the summer months in Arizona is associated with the NAM System (NAMS) in mid to late summer (July-September), an important regional climatological feature of the Southwest United States that accounts for approximately half the annual precipitation (e.g. Adams and Comrie, 1997).

Through explicit moisture tracing in a modeling framework, it is known that monsoon moisture may come from both the land surface, through local evapotranspiration, and atmospheric moisture transport from the Gulf of California, Eastern Pacific, and Gulf of Mexico (e.g. Hu and Dominguez, 2015). With respect to specifically in south central Arizona, low level moisture from the Gulf of California, in the form of Gulf surges (Rogers and Johnson, 2007) and mid-level moisture from the Gulf of Mexico are probably the most important. Gulf surges, or pulses of low-level moisture from the Gulf of California, are typical during monsoon bursts period and enhance the low-level atmospheric moisture necessary to sustain organized convection west of the Mogollon Rim in Arizona (Zehnder, 2004).

Synoptic-scale features, for example transient inverted troughs (Pytlak et al., 2005; Bieda et al., 2009; Seastrand et al. 2014; Lahmers et al., 2015), facilitate quasi-geostrophic upward vertical motion and provide favorable wind shear profiles. Given favorable dynamic conditions, monsoon thunderstorms may organize into MCSs that propagate off the terrain in the direction of the upper-level steering flow (e.g. Castro et al., 2007; Finch and Johnson, 2010; Newman and Johnson, 2012).

The strong influence of complex terrain on convective development is a key difference of Southwest cities (e.g. Phoenix, Tucson) as compared to cities in the central and eastern U.S., most of which are not adjacent to a large mountain range. Monsoon convection with air-mass type thunderstorms forms over the highest terrain in early to mid-afternoon (Janowiak et al., 2007). Precipitation generated by thunderstorms on the mesoscale is linked to the diurnal cycle of heating of terrain and generation of mountain-valley circulations (e.g. Diem and Brown, 2003). On the days where organized convection occurs, convective propagation is made possible through successive outflow boundaries (or cold pools) from leading convective lines. The outflow boundaries
mechanically lift moist and unstable air at the surface (Corfidi, 2003) thus may initiate new lines of convection through their convergence zones (Smith and Gall, 1989). Resultant MCSs may be sustained well into the late evening hours, long after reaching maximum in diurnal heating (Lang et al., 2007; Rotunno et al., 1988).

In these types of conditions, organized convection, in the form of MCSs and squall lines, is able to propagate westward into the Phoenix area with easterly mid to upper level steering flow in the atmosphere. These storms can originate on either Mogollon Rim, to the northeast of the city, or in the mountain ranges near the city of Tucson (e.g. Catalinas) to the southwest. Monsoon-related severe weather hazards include microbursts, dust storms, flash flooding and lightning are most favored during these monsoon burst periods (e.g. Douglas et al., 1993; Adams and Comrie, 1997; Ray et al. 2007).

1.2 NAM extreme precipitation trend relating to changes in atmospheric conditions.

There is likely to be an increase in heat and aridity in subtropical arid to semi-arid regions due to the retreat of the mid-latitude jet and expansion of sub-tropical highs (e.g. Lu et al., 2009; Archer and Caldeira, 2008; Seidel et al., 2008), for example as concluded in the recent Climate Change Assessment for the Southwest (Garfin et al., 2013) and Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report (IPCC, 2013). Another conclusion within these two climate assessment reports is that there will be an increase in precipitation intensity and more extreme weather, due to the exponential increase in atmospheric water vapor holding capacity of the atmosphere with a higher mean temperature (e.g. Meehl et al., 2000). This argument is supported by data from recent observational record. A number of analyses have documented significant positive trends in atmospheric water vapor globally and within the United States, in terms of both surface specific humidity and column-integrated precipitable water (e.g. Karl and Knight 1998; Groisman et al. 2005; Willett et al. 2007; Santer et al. 2007).

Karl and Trenberth (2003) have empirically demonstrated that for the same annual precipitation totals, warmer climates tend to generate more extreme precipitation events than cooler climates, implying that increases in extreme weather worldwide reflect more intense convective precipitation. Min et al. (2009, 2011) explicitly linked the
observed increases in extreme precipitation globally during the past several decades to anthropogenic greenhouse gas increases.

In the Southwest U.S., observation records show that 20-yr return period thresholds of daily maximum precipitation have exhibited an upward trend (Kunkel et al., 2013), matching what has been documented on a global scale. Extreme dry periods within the NAM region have increased 21% from 1910 to 2010, while extreme wet periods have increased 18% over the same period (Petrie et al., 2014). In an analysis of direct precipitation gauge measurements within the Southwest, Anderson et al. (2010) showed increases in monsoon precipitation and precipitation event frequency in Utah and Colorado, on the northern periphery of the monsoon. Increases in monsoon observed precipitation intensity in the Southwest within the context of the Climate Prediction Center daily gridded precipitation product during the period 1950-2010 has been revealed especially over the most mountainous areas in Chang et al. (2015).

With the distinct NAM thermodynamic pattern mentioned in section 1.1, the NAM-related severe weather events could be closely related to the changes in the large scale forcing but the relationships have not been examined in detail. One of my research interests is to objectively analyze the trend of NAM thermodynamic patterns and identify the linkage to the trend of weather extremes to establish a uniform tool of severe weather forecasting and projection to be used for climate change impact assessment.

1.3 Modeling severe weather during monsoon season

Global and regional atmospheric models have been used to generate future projections and retrospective simulations of the NAM for impacts assessment. Current future projections of the NAM, based on global and regional climate models have largely focused on the mean change in NAM precipitation within multi-model ensembles of climate change projection models. An analysis of all the Coupled Model Intercomparison Project, Version 5 (CMIP 5) global climate models suggest there will be a delay in monsoon onset in early summer due to increased atmospheric stability and an increase in monsoon precipitation in late summer once the stability barrier is overcome (Cook and Seager, 2013). The land-sea thermal contrast will increase in the context of the CMIP 5
models. This as a physical cause for weakening of the retreat of monsoon and increasing monsoon precipitation in early fall (Torres-Alvarez et al., 2015).

The use of models to represent monsoon convection, however, comes with some important caveats. Global climate models (GCMs) and global atmospheric reanalyses are generally challenged to represent the NAM as a salient climatological feature, in terms of its seasonal maxima in precipitation that occurs during July and August. This problem has been previously noted in reference to both atmospheric reanalyses and global seasonal forecast models (Castro et al. 2007, 2012) and the CMIP 3 climate change projection models (Bukovsky et al., 2015). Albeit there has been some notable improvement in the mean climatological representation of the NAM in more recent CMIP 5 models (Sheffield et al. 2013), a substantial number of the GCMs still erroneously delay the onset of the monsoon until August or September and are too wet during the early fall months (Geil et al., 2015). Dynamical downscaling of atmospheric reanalyses and global climate models to a grid spacing at the meso-β scale (10s of kilometers), for example as done in North American Regional Climate Change Assessment Program (NARCCAP), can improve a modeled climatology of NAM precipitation on a regional scale (Bukovsky et al. 2013; 2015). But regional models at the meso-β scale, even when “perfect” boundary forcing of an atmospheric reanalysis is applied, tend to overestimate monsoon precipitation in mountainous regions and underestimate precipitation associated with organized, propagating convection (Castro et al. 2012; Bukovsky et al. 2013).

Coarse resolution numerical atmospheric models (whether they are global or regional) generally have a poor representation of the terrain-forced diurnal cycle of convection (e.g., Collier and Zhang, 2007; Lee et al., 2007). This also adversely affects their ability to represent organized, propagating convection, for example occurring during the North American monsoon as described in analyses of radar data (Lang et al., 2007; Nesbitt et al., 2008). Increasing the horizontal resolution from 50 to 10km improves the ability of the model to capture individual NAM extreme (above 90th percentile) summer storm events (Tripathi and Dominguez, 2013). However, there is still consistent overestimation of precipitation and very small hit rates in the low desert at both resolutions.

The likely reasons why coarse-resolution models fail to represent the diurnal cycle well is because of their poor representation of terrain forcing, mesoscale
meteorological features (e.g. gulf surges), land–atmosphere coupling, and, most importantly, parameterized convective precipitation. Therefore, in order to simulate severe weather event well, MCSs have to be reasonably represented during the NAM season. This could be done by better treatments of both convective processes and land surface physics (discussed in subsection 1.3.1 and 1.3.2, respectively).

1.3.1 Convective parameterization

A substantial body of work over the past thirty years has demonstrated the importance of the meso-γ-scale for simulation of convection, or the convective-permitting scale on the order of a kilometer (e.g., Klemp and Wilhelmson, 1978a,b; Finley et al., 2001; Cai and Wakimoto, 2001; Xu and Randall, 2001). At this scale, the dynamic pressure (or non-hydrostatic) effects within thunderstorms may be explicitly simulated, and are critical to the evolution and movement of organized convection. This would specifically include the structure of MCSs, with leading convective lines and trailing stratiform precipitation regions (e.g. Cassell et al., 2015). Meso-γ orographic effects on the dynamic structure of airflow over mountains have also been well documented. Doyle and Durran (2002) depicted the formation of low-level rotors, horizontal vorticity, and waves propagating to upper levels caused by a 600-m-high mountain. Chu and Lin (2000) and Chen and Lin (2005) have demonstrated the presence of a dynamic pressure perturbation which may be associated with flow and precipitation over complex terrain.

A commonly used numerical atmospheric model, the Weather Research and Forecasting (WRF) model, is basically incapable of simulating a MCS during an Intensive Observing Period of the North American Monsoon Experiment (NAME) with a ten kilometer grid spacing (Cassell et al., 2015). This conclusion is consistent with other work that considered simulation of organized convection during NAME with either different regional atmospheric models and/or model domain and parameterization configurations (Li et al., 2008; Newman and Johnson, 2012). However it may be possible to modify existing commonly used cumulus parameterization schemes in atmospheric models to better represent organized convection, as opposed to convective-resolving simulations or through use of superparameterization, with the advantage of less of a computational expense. Truong et al. (2009) proposed a modification to the Kain Fritsch
(KF) convective parameterization scheme (CPS) to better account for dynamic pressure effects in complex terrain for coarse-resolution, meso-β scale simulations. These modifications account for dynamic pressure effects by taking the vertical gradient of the pressure perturbation into account. With the modified CPS, the model simulated precipitation yielded much better results in terms of the precipitation verification with available observed gauge data and satellite-derived precipitation. Moreover, the modified scheme produced larger and deeper stratiform clouds, increasing the resolved (non-convective) precipitation and a better represented a large, well-developed MCS that covered the central portion of the country. This modified scheme is very appropriate to apply in the NAM with similar complex orographic setup (study 1 - appendix A).

1.3.2 Rapid urbanization effect and its impact on monsoon precipitation

The second proposed treatment to the model is urbanization physics. Large urban regions in the middle of deserts occupying on the order of a thousand square miles, modify their surrounding environment in a variety of ways. The increase in the surface roughness, due to buildings, increases mechanically generated turbulence downwind and generates upward vertical motion (Bornstein and LeRoy, 1990; Diem and Brown, 2003). Urbanized surfaces reduce albedo and absorb more of the incoming solar radiation that is converted to sensible heat flux in large cities surrounded by highly vegetated natural landscapes like forests and croplands. Higher surface temperatures and deeper planetary boundary layers (PBL) are present over the city, as compared to the surrounding rural areas (e.g. Rozoff et al., 2003). Deeper PBL in association with the urban “heat island” causes low-level convergence that is favorable for the enhancement of convective precipitation downwind of the city (Bornstein and Lin, 2000, Rozoff et al., 2003, Cotton and Pielke, 2007).

However, a city like Phoenix surrounded by a desert has quite different urbanization respond. Irrigated croplands, like those southeast of Phoenix, as well as recreational irrigated lands have substantially greater evapotranspiration as compared to that of the surrounding natural desert vegetation. The added moisture from the irrigated lands introduces a relatively large amount of latent heating, or evapotranspiration. The increase in moisture within the urban area can actually reduce surface temperature. A desert urban area like Phoenix is essentially a “heat sink” rather than “heat island” (Diem
and Brown, 2003). Balling and Brazel (1987) showed there is a local blocking pattern in the isobaric surfaces in the city by analysis of surface pressure data. They suggested that the blocking pattern inhibits the cold pools from propagating into the city and thus suppresses convective activity. Shepherd (2006) evaluated long-term changes in observed mean monsoon precipitation from rain gauge measurements within and surrounding the Phoenix area, from a pre-urban period (1895-1949) and a post-urban period (1950-2003). They found a statistically significant increase (more than 10%) in monsoon rainfall in stations located to the north and east of the city, and speculated that landscape and agriculture in the Phoenix urban core enhance the moisture available for convection.

The impact of urbanization in Phoenix has been investigated explicitly through model sensitivity experiment (Georgescu, 2008, 2009a,b; Yang, 2015). However, precipitation change in the domain as a result of land surface change is not statistical significant, though there was signal of increases in precipitation to the north and east of Phoenix. The Phoenix “heat sink” only suppresses the convection that originates from elsewhere and passes through the city and thus can only be revealed on severe weather days due to organized, propagating convection. Urbanization can substantially change the severe weather precipitation pattern around a desert city with the use of an urban canopy model (study 2 - appendix B).

1.5 Objectives

The lack of research in NAM extreme precipitation motivates us to conduct this investigation. The scientific objective of this investigation is to physically characterize NAM-related severe weather precipitation and its response to urbanization and changing climate using a high resolution regional atmospheric model. It is hypothesized that urban “heat sink” in a desert city can only be revealed in the context of severe weather events. And the representation of the NAM MCSs can either be achieved by better account for dynamic pressure effects in complex terrain for coarse-resolution (meso-β scale) simulations or by convective-permitting grid spacing (meso-γ scale) simulations.

In that context, I am proposing the following three studies that are the main structure of the dissertation. In the first study, we identify the improvement in the representation of the North American Monsoon convective precipitation using the
modified KF CPS (Appendix A). In the second study, we reveal the urban “heat sink”
effect on the North American monsoon precipitation in Arizona within the context of
modeled severe weather events (Appendix B). In the third study, we explore the more
extreme nature of monsoon precipitation in the Southwest U.S. as revealed by a long-
term climatology of simulated severe weather events (Appendix C).
CHAPTER 2: PRESENT STUDY

The methods, results, and conclusions of this investigation are presented in three manuscripts formatted for submission and appended in this dissertation. The following is a summary of the most relevant findings.

2.1 Appendix A: Improvement in the representation of the North American Monsoon convective precipitation using a Modified Kain-Fritsch convective parameterization scheme

In the first study, the research question is to evaluate if representation of NAM MCSs can be improved with better account for dynamic pressure effects in complex terrain for coarse-resolution. We have demonstrated the value added of the modified KF scheme originally described in Truong et al. (2009) in dynamical downscaling paradigms of numerical weather prediction and regional climate modeling with WRF. We first verified that the modified scheme is able to reproduce the results of the extreme precipitation event in Vietnam considered in their original RAMS study. Next, we repeated the NAME IOP 2 WRF modeling experiments described in Cassell et al. (2015) with the modified KF. Clear and dramatic improvements in the model simulations were found in the following aspects: representation of the vertical structure and timing of the gulf surge, organization and propagation of a MCS, and alleviation of model-simulated precipitation biases, by decreasing the precipitation over complex terrain and increasing more off the terrain in lowland desert regions toward the Gulf of California.

We also repeated the dynamical downscaling of an atmospheric reanalysis originally described in Chang et al. (2015) with the modified KF scheme. Similar to the NWP-type simulations of NAME IOP2, use of the modified KF scheme yields results that compare better with observed precipitation, alleviating well-known positive biases in the regional model simulated rainfall over the SMO, for example as described in Castro et al. (2012). Monsoon precipitation also increases in more lowland areas at further distance from terrain, where MCS-type convection would account for a greater proportion of the total monsoon precipitation. When the RCM simulations are further dynamically downscaled to convective-permitting scales (2.5 km) in WRF to simulate
extreme precipitation events, we find that the use of the RCM simulation with modified KF also has a positive impact on the representation of organized convection, even though the CPS is not activated on this grid.

2.2 Appendix B: The impact of urbanization on the North American monsoon precipitation in Arizona within the context of modeled severe weather events

In the second study, we evaluate how urbanization affects a monsoon precipitation. A regional model case study in Phoenix thunderstorms during a specific severe weather event has been simulated. A new urban canopy model (UCM) is used within WRF that more explicitly accounts for the urban surface in terms of roughness, agricultural and landscape irrigation, increased heat storage and anthropogenic heating. Our basic conclusion from this model sensitivity experiment is that the Phoenix urban area seems to act as “heat sink” as suggested by Diem and Brown (2003), behaving in an opposite way to a cities of similar size in the central and eastern U.S. with a documented “heat island.” Greater portioning of surface heat fluxes to latent heating due to the presence of more vegetation and irrigated areas in the urban core, as compared to the surrounding desert, lead to a shallower PBL, higher surface pressure, and a more stable atmosphere. These factors suppress convective precipitation, causing the monsoon thunderstorms to bypass and divert around the city, increasing the precipitation in the areas surrounding the urban core in a horseshoe pattern to the north, east, and south. Immediately downwind of the city to the southwest in the direction of the upper-level steering flow, precipitation decreases.

We repeated these experiments on a series of 25 of the most severe weather event cases for the Phoenix area, within the context of most thermodynamically favorable days identified in a long-term dynamically downscaled atmospheric reanalysis for the period 1991-2010. The composite results of these experiments essentially confirm the same behavior observed for the 13-14 August 2003 case also applies, in climatological sense, to all the days which would be most conducive for severe monsoon weather in Arizona. The average difference pattern of precipitation for the “urban” minus “desert” simulations comports with the findings of Shepherd (2006) that showed statistically
significant changes in observed monsoon precipitation in association with the growth of Phoenix during the twentieth century.

2.3 Appendix C: The more extreme nature of monsoon precipitation in the Southwest U.S. as revealed by a long-term climatology of simulated severe weather events

In the third study, we have evaluated long-term changes in precipitation intensity during the North American monsoon in the Southwest U.S., through the use of convective-permitting model simulations of objectively identified severe weather events during “historical past” (1950-1970) and “present day” (1991-2010) periods. These severe weather events are the days when the highest atmospheric instability and moisture occur in the Southwest within a long-term regional climate simulation (35 km grid spacing) that dynamically downscales a global atmospheric reanalysis over the contiguous U.S. and Mexico.

The severe weather event simulations appear to reasonably represent the diurnal cycle of convective precipitation during the period of the Stage IV product, in terms of the development of precipitation over the highest terrain during the day and convective organization and propagation into the evening hours. The comparisons with hourly precipitation data from NCDC coop stations also shows that the model simulations effectively captures the differences in the timing of convective precipitation during the monsoon in relation to elevation.

While mean daily monsoon precipitation in the Southwest has decreased from the analysis of the CPC dataset, the most extreme monsoon precipitation has become more in these same data during the days with most favorable thermodynamic and dynamic conditions to support organized, propagating convection. We observe a similar increasing intensity of extreme monsoon precipitation in the convective-permitting severe weather event simulations. In the model simulations, precipitation is becoming more intense within the context of the diurnal cycle of convection. Largest modeled increases in extreme event precipitation occur is located in central and southwest Arizona, where MCSs account for a majority of monsoon precipitation during the evening hours. The conclusion of this study is a more favorable thermodynamic environment in the
Southwest U.S. is facilitating stronger organized monsoon convection during the last twenty years.
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APPENDIX A: IMPROVEMENT IN THE REPRESENTATION OF THE NORTH AMERICAN MONSOON CONVECTIVE PRECIPITATION USING A MODIFIED KAIN-FRITSCH CONVECTIVE PARAMETERIZATION SCHEME

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Abstract

A commonly noted problem in the simulation of warm season convection in the North American monsoon region has been the inability of atmospheric models at the meso-β scales (10s to 100s of kilometers) to simulate organized convection, principally mesoscale convective systems (MCSs). With the use of convective parameterization, high precipitation biases in model simulations are typically observed over the peaks of mountain ranges, for example the Sierra Madre Occidental in western Mexico. To address this issue, the Kain-Fritsch (KF) cumulus parameterization scheme has been modified with a new diagnostic equation to compute the updraft velocity, the convective available potential energy (CAPE) closure assumption, and the convective trigger function. This modified scheme was originally tested for the simulation of an extreme precipitation event in Vietnam with the Regional Atmospheric modeling system. The scheme has been adapted for use in WRF and tested for Vietnam and North American monsoon region. A numerical weather prediction-type simulation is conducted for the North American Monsoon Experiment (NAME) Intensive Observing Period 2 and a regional climate simulation is performed, by dynamically downscaling the NCEP-NCAR Reanalysis (1950-2010). In both of these applications, there are notable improvements in the WRF model-simulated precipitation due to the better representation of organized, propagating convection. The use of the modified KF scheme for atmospheric model simulations at the meso-β scale may provide a more computationally economical alternative to improve the representation of organized convection, as compared to convective-permitting, meso-γ simulations at the kilometer scale or a superparameterization approach.
1. Background and Motivation

The North American monsoon (NAM) is the principal driver of summertime severe weather in the Southwest U.S. Given sufficient atmospheric instability and moisture, monsoon thunderstorms typically initiate during the daytime over the highest terrain in association with the mountain-valley circulations. Given favorable dynamic conditions of vertical wind shear and synoptic-scale upward vertical motion, these thunderstorms may organize into mesoscale convective systems that propagate off the terrain in the direction of the upper-level steering flow (e.g. Castro et al., 2007; Finch and Johnson, 2010; Newman and Johnson, 2012). During these types of monsoon burst periods convective precipitation associated with MCSs tends to propagate from the high terrain into the low deserts (e.g. Watson et al. 1994). For example the convective environment is much more favorable for MCSs to reach the urban areas of Tucson and Phoenix and westward into the low southwest deserts and the Colorado River valley. Monsoon-related severe weather hazards are most favored during these periods, and include microbursts, dust storms, flash flooding and lightning (e.g. Douglas et al., 1993; Adams and Comrie, 1997; Ray et al. 2007).

Though there are synoptic-scale dynamical features that facilitate convective organization during these monsoon burst periods, in particular transient inverted troughs (Pytlak et al., 2005; Bieda et al., 2009; Seastrand et al., 2014; Lahmers et al., 2015), the convection itself occurs on the meso-γ scale (e.g. Pielke and Pearce, 1994) with a spatial extent on the order of one to tens of kilometers. Development of monsoon convection is strongly tied to a diurnal cycle, with air-mass type thunderstorms forming over the highest terrain in early to mid-afternoon (Janowiak et al., 2007). On the days when
organized convection occurs, convective outflow boundaries and their convergence zones may initiate new lines of convection (e.g. Smith and Gall, 1989; Watson et al. 1994; McCollum et al. 1995). Gulf surges, or pulses of low-level moisture from the Gulf of California, are also typical during monsoon bursts period and enhance the low-level atmospheric moisture necessary to sustain organized convection west of the Mogollon Rim in Arizona (e.g. Douglas 1995; Zehnder, 2004; Schiffer and Nesbitt 2012).

Propagating mesoscale convective systems may be relatively long-lived, persisting for more than several hours into the evening and early morning (e.g. Rotunno et al., 1988; McCollum et al. 1995; Lang et al., 2007).

Numerical atmospheric models can be used to represent monsoon convection. However, as we discuss in Castro et al. (2012) with reference to the warm season in the western and central United States, regional models used at spatial resolution on the order of tens of kilometers, or the meso-β scale, have a particular spatial pattern of bias in their simulated precipitation. They tend to overestimate precipitation in areas of complex terrain and underestimate precipitation in areas with more homogeneous terrain, where organized convection accounts for a greater proportional of the precipitation. Coarse resolution, meso-α to meso-β numerical atmospheric models (whether they are global or regional) generally have a poor representation of the terrain-forced diurnal cycle of convection (e.g., Collier and Zhang, 2007; Lee et al., 2007). This also adversely affects their ability to represent organized, propagating convection, for example MCSs that occur during the North American monsoon as described in analyses of radar data (Lang et al., 2007; Nesbitt et al., 2008). The likely reasons why coarse-resolution models fail to represent the diurnal cycle well is because of their poor representation of terrain forcing.
mesoscale meteorological features (e.g. gulf surges), land–atmosphere coupling, and, most importantly, parameterized convective precipitation. We have demonstrated in prior work that a commonly used numerical atmospheric model, the Weather Research and Forecasting (WRF) model, is basically incapable of simulating a mesoscale convective system during an Intensive Observing Period of the North American Monsoon Experiment (NAME) with a ten kilometer grid spacing (Cassell et al., 2015). These simulations used a parameterization configuration nominally similar to that of the quasi-operational WRF forecasting system at the Department of Atmospheric Sciences at the University of Arizona. This conclusion in Cassell et al. (2015) is consistent with other work that considered simulation of organized convection during NAME with either different regional atmospheric models and/or model domain and parameterization configurations (Li et al., 2008; Newman and Johnson, 2012).

A substantial body of work over the past thirty years has demonstrated the importance of the meso-γ-scale for simulation of convection, or the convective-permitting scale on the order of a kilometer (e.g., Klemp and Wilhelmson, 1978a,b; Finley et al., 2001; Cai and Wakimoto, 2001; Xu and Randall, 2001). At this scale, the dynamic pressure (or non-hydrostatic) effects within thunderstorms may be explicitly simulated, and these are critical to the evolution and movement and of organized convection. This would specifically include the structure of mesoscale convective systems, with leading convective lines and trailing stratiform precipitation regions (e.g. Cassell et al., 2015). Meso-γ orographic effects on the dynamic structure of airflow over mountains have also been well documented. Doyle and Durran (2002) depicted the formation of low-level rotors, horizontal vorticity, and waves propagating to upper levels...
caused by a 600m high mountain. Chu and Lin (2000) and Chen and Lin (2005), along with several others, have directly or indirectly demonstrated the presence of a dynamic pressure perturbation which may be associated with flow and precipitation over complex terrain. One possible alternative to convective-permitting simulations at the meso-γ is the technique of superparameterization (e.g. Grabowski and Smolarkiwicz, 1999; Grabowski, 2001; Randall et al., 2003), in which a two-dimensional cloud-resolving model is essentially statistically imbedded within the grid cell of a coarser resolution model. The use of superparameterization has been demonstrated to add value in organized convective processes in global climate models, for example producing a Madden-Julian Oscillation when the employment of a more typical quasi-equilibrium-type cumulus parameterization (e.g. Arakawa-Schubert scheme [Arakawa and Schubert, 1974]) cannot.

It may be possible to modify existing commonly used cumulus parameterization schemes in atmospheric models to better represent organized convection, as opposed to convective-resolving simulations or superparameterization, with the advantage of reduced computational expense. Our previous work in Truong et al. (2009) proposed a modification to the Kain Fritsch (KF) convective parameterization scheme (CPS) to better account for dynamic pressure effects in complex terrain for coarse-resolution, meso-β scale simulations. Specifically, that study modified the scheme with a new diagnostic equation to compute the updraft velocity, the convective available potential energy (CAPE) closure assumption, and the convective trigger function. These modifications better account for dynamic pressure effects by taking the vertical gradient of the pressure perturbation into account. The modified KF CPS was initially incorporated into the Regional Atmospheric Modeling System (RAMS) and used to
simulate an extreme heavy rainfall event from 24 to 26 November 2004 in the mountainous provinces of central Vietnam that resulted in severe flooding along local rivers. With the modified CPS, the model simulated precipitation was more consistent with available observed gauge data and satellite-derived precipitation estimates. Moreover, the modified scheme produced larger and deeper stratiform clouds, increasing the resolved (non-convective) precipitation, and it better represented a large, well-developed mesoscale convective system that covered the central portion of the country.

Though the study of Truong et al. (2009) considered only one specific case, it highly motivates further testing of the modified KF scheme in environments of complex terrain where organized mesoscale convective systems are climatologically very important but relatively challenging to represent in a numerical atmospheric model. We investigate in this work whether the modified KF CPS scheme applied in Vietnam may help alleviate some of the aforementioned problems in modeling organized convection in the North American monsoon region. The nature of North American monsoon precipitation, particularly in Mexico, is actually quite similar to that of Vietnam during the warm season. In both locations, development and propagation of convection is strongly linked to the topography, there is a proximity to ample sources of deep tropical moisture (from the Eastern Pacific/Gulf of California and the South China Sea, respectively), and severe weather, aside of tropical cyclones, is commonly associated with organized mesoscale convective systems.

The main goal of this study is to demonstrate the effectiveness of the modified KF CPS within the WRF model for a well-documented case of organized convection in the North American monsoon region. Specifically, the case we consider here is Intensive
Observing Period 2 from NAME, which has been previously studied with the use of a convective-permitting NWP-type simulation in Cassell et al. (2015), and Newman and Johnson (2012). These two works both show that the explicit representation of convection at the meso-γ scale, without the use of a convective parameterization, is necessary for a reasonable representation of a large, westward propagating MCS in the Mexican state of Sonora, which occurred during the IOP, and its associated outflow boundary that enhanced a gulf surge. In using the same simulation design as in Cassell et al. (2015), we aim to show that application of the modified KF CPS on the meso-β scale helps to better represent the physical structure of the MCS and its associated precipitation, in a similar way as the previously discussed Vietnam case.

2. Data Analysis and Methods

a. Overview of NAME Intensive Observing Period 2

We conduct our atmospheric model sensitivity experiments on Intensive Observing Period 2 of the North American Monsoon Experiment (NAME) during the summer of 2004. As we have summarized previously in Cassell et al. (2015) and basically restate here from that work in this sub-section, the Intensive Observing Periods (IOPs) are periods of higher frequency data collection that focus on specific meteorological features of interest during the monsoon, embedded within an overall enhanced observing period (EOP). Some specific scientific goals of the IOPs included better characterization and physical understanding of: 1) mean moisture flux over the NAME Tier I region (basically northwest Mexico encompassing the Mexican states of Sonora, Sinaloa and part of Durango as well as southern Arizona in the U.S.), 2) the structure of the low-level jet in the Gulf of California, 3) the genesis and propagation of
gulf surges, 4) middle and upper-level easterly inverted troughs, 5) inland penetration of the sea breeze and planetary boundary layer evolution and convective development over the Sierra Madres, and 6) the evolution of mesoscale convective systems. Potentially improved representation of organized convection in numerical atmospheric models is most closely aligned with the latter two of these NAME IOP science objectives.

Over the entire course of NAME, a total of ten IOPs were called. The detailed weather forecast and post-event discussions are maintained on the NAME website (http://catalog.eol.ucar.edu/name/catalog/missions.html). IOP 2, which took place 11 July 2004 to 14 July 2004, is the most important of the IOPs considered, and has been the most comprehensively investigated (Rogers and Johnson, 2007; Finch and Johnson, 2010; Newman and Johnson, 2012; Cassell et al., 2015). Therefore, our results presented in Section 3 will be for this event. IOP 2 was the first major outbreak of organized convection during the 2004 season, and heralded the (climatologically late) onset of the monsoon in Arizona. It had a major gulf surge triggered by the passage of a tropical cyclone (Blas) near the southern end of the Gulf of California, an inverted trough located just east of the core monsoon region, and well organized MCSs in both Sonora and Arizona that propagated all the way to the Colorado River Valley and Gulf of California over multiple days. Analysis of the NAME-generated observational datasets and convective-resolving simulations considering IOP 2, from aforementioned references, have demonstrated: 1) the importance of northeasterly flow and anomalous northeasterly shear, due the presence of the inverted trough, in creating an environment favorable for convective organization, and 2) the influence of MCSs in strengthening Gulf surges, through their convective outflow boundaries.
**b. Regional Atmospheric Model Simulation Design for NAME IOP 2**

The IOP 2 is retrospectively simulated using the Advanced Research Weather Research and Forecasting (ARW-WRF) Model, version 3.1.1 (Skamarock et al., 2005). A multiple grid nesting procedure is used that approximately mimics the NAME Tier region structure (Fig. 1). The coarsest grid (132 x 134 grid points at 30 km grid spacing) covers most of the contiguous U.S.; and the finer grid (265 x 262 grid points at 10 km grid spacing) covers the Southwest U.S. and northern Mexico. There are 37 levels in the vertical on eta levels. The model simulations of NAME IOP 2 is approximately 48 hours in length and are initialized 6 hours prior to the start of the IOP 2, as this is a sufficient period to allow for model spin-up and consistent with practices used to generate quasi-operational WRF numerical weather prediction (NWP) forecasts at the University of Arizona, Department of Atmospheric Sciences (UA-ATMO) at a grid spacing of 1.8 km (http://www.atmo.arizona.edu/index.php?section=weather).

The model physical parameterizations are also mostly consistent with those of the existing UA-ATMO WRF NWP system. Common parameterizations include: the Morrison double moment scheme for microphysics (Morrison et al., 2005); Eta surface layer (Janjic, 1996, 2002); Mellor-Yamada-Janic (MYJ) planetary boundary layer (Janjic, 1990, 1996, 2002); and the NOAH land surface model (Ek et al., 2003). The IOP simulations additionally use Dudhia Shortwave radiation (Dudhia, 1989) and the Rapid Radiative Transfer Model (RRTM) Longwave radiation (Mlawer et al., 1997). Convection is parameterized using the Kain-Fritsch scheme (Kain and Fritsch, 1993, 2004), in its standard form available within the WRF package and with the modifications of Truong et al. (2009), to be specifically summarized later here in sub-section e.
c. Observational Data

Atmospheric lateral boundary forcing data for WRF simulations is specified by the Global Forecast System (GFS) final (FNL) analyses prepared by the National Centers for Environmental Prediction (NCEP). The FNL Operational Global Analysis data are on one degree grids prepared operationally every six hours. These data are derived from the Global Data Assimilation System (GDAS), which continuously collects observational data from the Global Telecommunications System (GTS), and other sources (see http://rda.ucar.edu/datasets/ds083.2/). North American Regional Reanalysis (NARR, Mesinger et al. 2006) data is used to initialize the soil moisture in lieu of FNL data as FNL data for 2004 only has two soil moisture levels. NARR soil moisture is generated by the NOAH land surface model, which is equivalent to the land surface model that is used within the WRF simulation. For verification, both the quality controlled NCEP Stage IV radar+gauges data (http://data.eol.ucar.edu/codiac/dss/id=21.093) and the Tropical Rainfall Measuring Mission (TRMM) satellite data were used to compare measured rainfall to model rainfall. Geostationary Operational Environmental Satellite (GOES) 10 satellite data is used in comparison against the WRF simulations. We note that in Cassell et al. (2015), the results from the WRF simulations with application of the native KF CPS were compared to the Colorado State University (CSU) NAME surface dataset, discussed in detail in Johnson et al. (2007) and Ciesielski and Johnson (2008). As our main objective here is to show differences in model simulations, we refer the reader to that previous paper for the observational comparisons of the IOP 2 simulations, but do reference these observational data in Section 4 as necessary.

d. Regional climate model simulation
In addition to the simulation of NAME IOP-2, we also repeat the dynamical downscaling of the NCEP-NCAR Reanalysis with WRF as recently described in Chang et al. (2015) but with the modified KF CPS for continuous sixty year period 1950-2010. This regional climate model (RCM) simulation encompasses a domain of the contiguous U.S. and Mexico at a grid spacing of 35 km. The model parameterizations are nominally similar to the NAME IOP2 simulation and spectral nudging is employed to maintain the variability of large-scale circulation features. Our objective in conducting this RCM simulation is to evaluate whether the modified KF scheme can help alleviate some of the aforementioned known biases in warm season precipitation that seem to be common to RCMs, overestimation of precipitation in mountainous areas and underestimation of precipitation in areas with more homogeneous terrain where the organized convection accounts for a greater proportion of the warm season precipitation (e.g. Castro et al., 2012).

e. Modifying the Kain-Fritsch CPS to account for vertical dynamic pressure effects

We modified the KF CPS in WRF following Truong et al (2009), and summarize the key points of the scheme modifications here for brevity. We refer the reader for a more thorough derivation and theoretical consideration of the scheme modifications to that previous paper. We start from the equation of motion in a locally steady state, so just the advective components are retained in an Eulerian framework of expansion of the total derivative of wind. Considering the expression for the grid-scale vertical advective component (left hand side of equation below)

$$
\bar{w} \frac{\partial \bar{w}}{\partial z} = g \left( \frac{T'}{T_0} - \frac{R}{C_p} \frac{p'}{p_0} \right) - \theta_0 \frac{\partial \pi'}{\partial z} + ADV
$$

(1)
Where the right side of the equation the first term is the buoyancy in its full form, accounting for both perturbation temperature and dynamic pressure \( (p') \), the second term is the vertical gradient of the Exner function perturbation (basically equivalent to the vertical gradient of the dynamic pressure), and the last term \( (\text{ADV}) \) is the advection of vertical velocity due to the grid-scale horizontal velocity. Note here that the perturbation quantities (prime terms) are computed as the difference between the grid-resolved quantity minus a synoptic-base state computed from hydrostatic balance.

We calculate the ratio of vertical gradient of the pressure perturbation and buoyancy force, expressed with these difference terms, as

\[
P_B = \frac{\theta_0 \frac{\partial (\tilde{\pi} - \pi_0)}{\partial z}}{g \left( \frac{T_u - T_0}{T_0} - \frac{R \bar{p} - p_0}{\bar{C}_p} \right)}
\]

We introduce modified buoyancy at every model vertical level that takes the dynamic pressure perturbations with the above ratio

\[
F(z) = g \frac{T_u(z) - \bar{T}(z)}{\bar{T}(z)} \left[ 1 + PB(z) \right]
\]

This function basically adds a weighting factor to buoyancy that explicitly accounts for the vertical gradient of dynamic pressure. This can be used to create a modified Convective Available Potential Energy (CAPE) closure assumption

\[
\text{CAPE} = \int_{LCL}^{CT} F(z) \, dz
\]

We then add a new convective trigger that considers this dynamic pressure-modified CAPE. A potential updraft source layer (USL) that initiates convection must satisfy 2 conditions:
1) The modified CAPE within an updraft source layer must be positive in order to support convection

\[
\begin{align*}
\left[w_{\text{mix}}^2 + 2 \int_{\text{USL}_{\text{base}}}^{\text{LCL}} F(z)dz \right] > 0, & \quad w_{\text{mix}} > 0 \\
\left[-w_{\text{mix}}^2 + 2 \int_{\text{USL}_{\text{base}}}^{\text{LCL}} F(z)dz \right] > 0, & \quad w_{\text{mix}} < 0
\end{align*}
\]

(5)

2) There must be positive modified CAPE within an updraft source layer (USL) that is sufficient to overcome any convective inhibition through the depth of the USL to the lifting condensation level.

\[F_{\text{USL}} = \int_{\text{USL}_{\text{base}}}^{\text{USL}_{\text{top}}} F(z)dz > 0 \quad (6)\]

3. Evaluation of the modified scheme for the Vietnam case in WRF

We first test the modified scheme with the same case study for Vietnam in Truong et al. (2009) simulated with the RAMS model, to verify the value gained by a new trigger function similarly exists in WRF before applying it to simulation of North American monsoon organized convection for NAME IOP2. We use a nominally similar experimental design in terms of domain configuration, NCEP-NCAR reanalysis boundary forcing data, and WRF model parameterizations as previously described in Section 2. There are two nested grids with a grid spacing of 40 and 10 km, respectively. Additionally, we add a third convective-permitting grid of 2.5 km, with the convective parameterization scheme turned off. For brevity, we just show the final comparison results for model simulated precipitation for the 48 h model period starting 0 UTC 24 November 2004. Reproduction of all the accompanying model-simulated atmospheric variables is omitted for the Vietnam case, as that will be done in more detail for the NAME IOP2 simulation later to demonstrate the value added of the modified KF CPS on
the atmospheric circulation fields. The main objective in that previous paper (Truong et al., 2009) was to better represent the precipitation maxima in central Vietnam associated with a mesoscale convective system, estimated to be 721 mm from an in-situ rain gauge measurement at the station of Thuong Nhat (16.128N, 107.688E) and 713 mm in remote-sensed satellite data from the Tropical Rainfall Measuring Mission (TRMM) product.

In the original RAMS simulation in Truong et al. (2009), the modified KF CPS increases the precipitation maxima at Thuong Nhat station from 345 mm to 673 mm on a 10 km intermediate grid (see their Fig. 7). Application of the native KF scheme in WRF simulations produces comparatively less rain compared to RAMS (249 mm as in Fig. 2). Even though the modified KF with WRF does not produce double the amount of precipitation as in RAMS at the Thuong Nhat station location, it still does yield 335 mm, about a 100 mm more than the simulation with the native KF scheme. A third nested interior domain with 2.5km grid spacing shows the impact of applying the modified KF scheme as boundary forcing to the convective-permitting grid where the CPS scheme has been turned off. Results of the same 48-hour rainfall are shown in Fig. 3. Considering boundary forcing from the intermediate grid with the native KF scheme, the convective-permitting grid produces local maxima of 653 mm and 665 mm near the Thuong Nhat station. However, use of modified KF scheme on the intermediate grid downscaled to the convective-permitting 2.5 km grid produces a local maximum that is nearly twice as much at 1285 mm.

Though the absolute precipitation amounts are different between the RAMS and WRF simulations, basically the same general conclusions hold in comparing simulations with the native KF scheme and modified KF. In both cases, the modified KF scheme not
only produces more precipitation on the intermediate 10 km resolution domain that is closer to the observed values from rain gauge and satellite-derived estimates. But perhaps more importantly, application of the modified KF scheme is substantially modifying the model-simulated precipitation on the convective-permitting 2.5 km resolution domain. Therefore, at least having verified a similar behavior in WRF on the intermediate domain as Truong et al. (2009) with their RAMS simulation for the Vietnam case, we proceed to apply the modified KF scheme to the WRF simulations of interest in the NAM region.

4. Application of the modified KF scheme in WRF simulation of NAME IOP2

The objective of the presentation of WRF model simulation results of NAME IOP2 in this Section is to show that the modified KF scheme produces improved results, as compared to that with the native KF scheme, in two respects. First, we want to show that the modified KF scheme produces precipitation that compares better to available observed products of precipitation, as archived in NAME data inventories, especially in reducing the positive precipitation bias over the highest terrain. Second, we want to evaluate if the modified scheme more realistically in physically represents the gulf surge and MCS that occurred during the IOP. The presentation of results is meant to be as consistent as possible with our prior results in Cassell et al. (2015) that simulated NAME IOP2.

a) Review of meteorological conditions during NAME IOP2

The meteorological conditions during NAME IOP2 are well illustrated by GOES enhanced infrared satellite imagery as shown in Fig. 4 near the time of 3Z 14 July 2004 (7pm local time in Arizona). From the satellite-derived outgoing longwave radiation
it is clear that there is organized convective precipitation occurring, with multiple mesoscale convective systems present in northern Sonora and southern Arizona. At this time, these MCSs are propagating westward toward the Gulf of California and the Colorado River Valley from their origination points in the Sierra Madre Occidental and Mogollon Rim. The leading edge of the strongest MCS has reached the northern coast of the Gulf of California at this time. As we have discussed in Cassell et al. (2015), the development of these MCSs is facilitated by low-level moisture (below about 800-mb) supplied by a strong gulf surge in the Gulf of California, triggered by the passage of tropical storm Blas located several hundred miles southwest of the southern tip of Baja California.

The strong gulf surge that traveled the entire length of the Gulf of California is one of the most distinguishing features of NAME IOP2, and has been documented extensively from the NAME observational network in Rogers and Johnson (2007). Fig. 5 shows, taken from this paper, the wind profiler data during the surge for Puerto Peñasco, located at the northern end of the Gulf of California. The surge was most pronounced at this location, as compared to the other NAME wind profiler stations along the coast of the Gulf of California. The surge is strongest during the approximate period 9 UTC to 18 UTC (2am to 11am local Arizona time) on 13 July, with a maximum wind speed of more than 20 m s\(^{-1}\) measured by the wind profiler instrument half a kilometer above the ground surface at 15UTC (8 am local Arizona time).

b) Model Representation of the Gulf Surge

As mentioned in the Introduction, the value added of representing the vertical structure of the gulf surge in convective-permitting simulations in WRF has already been
demonstrated in Cassell et al. (2015) and Newman and Johnson (2012), with relatively poorer representations of the gulf surge on their courser grids on the meso-β scale. Here we use the metrics of Cassell et al. (2015) for surge evaluation applied to the WRF model output at Puerto Peñasco, but just considering the intermediate 10 km grid where convective parameterization is applied in Fig.6. The simulation with the native KF has two clear deficiencies, as compared to the observed wind profiler data. The strength of the low level moisture surge is not strong enough during the observed time of the surge. Model simulated northerly winds are only on the order of 10 to 15 m s\(^{-1}\) within the lowest 1000m above the land surface. The timing of the maximum of the surge is also incorrect with maximum wind values of 17.5 m s\(^{-1}\) in the lowest kilometer above the surface and occurring nearly a day after the observed event, and this was a problem we also noted in Cassell et al. (2015). The modified scheme improves upon both the magnitude and timing of the gulf surge, producing a wind speed on the order of 15 to 20 m s\(^{-1}\) in the lowest kilometer within 6 hours of the time when the observed 20 m s\(^{-1}\) wind maximum occurred. Though the wind maximum of the gulf surge is still underestimated, and the surge is not as deep, the modified scheme still performs much better in simulating the NAME IOP2 gulf surge, particularly in terms of the timing.

c) Cross sectional analysis of condensate mixing ratio at high temporal resolution

Cross sections of the total condensate mixing ratio and wind field simulated by the model are shown for latitude 26.8 N in Fig. 7a in approximate relation to the time of MCS development during the afternoon of 13 July, from 7 UTC to 12 UTC (12 noon to 5pm local time). The simulations are saved at ten minute intervals, as that is the necessary temporal scale to visualize the convective development and propagation.
However hourly panels plotted in figure 7a are sufficient to show the MCS propagation. Note that we are only showing in these figures the results with the modified scheme. Consistent with what has been reported previously in Cassell et al. (2015), with use of just the native KF scheme on the 10 km grid, convection develops directly over the peak of the Sierra Madre Occidental mountains and does not propagate, hence the corresponding figure for native KF is not shown. By contrast, the use of the convective-permitting scale produces convection with a MCS-type structure, with a leading convective line, organized inflows and outflows, and a trailing stratiform precipitation region (Fig. 7b). Thus the use of the modified KF scheme produces a result that much more closely matches the Cassell et al. (2015) simulation on their convective-permitting grid, even though not as detail due to simulated on a much coarser grid (4 times coarser). Both figures 7a and 7b reproduce MCSs in their mature stage per criteria of Houze (1993).

The time evolution of the total condensate mixing ratio and wind velocity clearly show convective organization and propagation via the mechanism of outflow boundaries triggering new convective lines. As the simulation progresses, the leading convective line becomes more intense, with higher total condensate mixing ratios and stronger updrafts and a well-defined gust front as the MCSs moves westward off the Sierra Madre Occidental towards the Gulf of California. Near the end of the period in Fig. 7a during the hour of 11Z to 12Z there is even a signature of a trailing stratiform precipitation region and rear inflow jet near 700-mb. The very important conclusion from the cross sectional analysis is that the use of the modified KF scheme permits the development of MCS-type convection during the NAM in a manner that is very similar to what can be
achieved using convective-permitting simulation, at finer grid spacing and more computational expense. It is the MCS outflow boundary that is also responsible for intensifying the modeled gulf surge. Our observations identified that the modified scheme better represents MCS-type cloud structure and the process of convective development and propagation basically matches what was found for the RAMS simulations in the Vietnam case.

*d) Differences in model simulated precipitation and comparison with observations*

The 24h observed and modeled precipitation for NAME IOP2 is shown in Fig. 8 for the 10 km WRF grid, along with the differences in model-simulated precipitation from the modified KF minus native KF. The specific period is 12 UTC 13 July to 12 UTC 14 July. Observed precipitation is from the NAME observational rain gauge network. On the top left of the figure is the model-accumulated precipitation using the native KF scheme. The native KF scheme, as suspected, overestimates precipitation over the Sierra Madre Occidental (SMO) where convection is initiated but underestimates it further west, more a distance from the mountains and toward the coast of the Gulf of California. Use of the modified KF scheme, on the lower left of the figure, reduces the precipitation over the SMO but increases the precipitation further west toward the coast, thus matching the NAME gauge-derived precipitation observations better. Specifically considering the differences between modified KF minus native KF in the lower right panel, the modified KF scheme clearly reduces the precipitation that occurs directly over the SMO. The reduction in the 24h model simulated precipitation is greater than 25 mm (one inch) in some places. Increases in precipitation nearer to the coast with modified KF, for example to the north and west of the city of Hermosillo, are on the order of 8-16
mm. This is about half the total model simulated precipitation. Interestingly, there seems to be specific elevation transition point of approximately 1000 m that effectively separates those geographic areas on the domain where use of the modified KF scheme is decreasing precipitation versus where it is increasing it. For reference the 1000 m elevation contour is included on the figure. So below this elevation, it is fair to conclude that more of the precipitation is accounted for by westward propagating mesoscale convective system that occurred during the IOP, as discussed earlier in the cross sectional analysis.

To further illustrate the improved representation of the propagating MCS in the simulation with the modified KF scheme, we look more closely at the diurnal cycle of precipitation. The observed diurnal pattern in precipitation is shown in Fig. 9 for two six hour periods corresponding to 18 UTC 13 July to 0 UTC 14 July (11 am – 5pm local Arizona time) and 0 UTC 14 July to 6 UTC 14 July (5pm – 11pm local Arizona time). Two sources of observational data are shown in Fig. 9, both have sufficient temporal resolution to capture this diurnal evolution of precipitation, the Stage IV combined radar-gauge product and the TRMM satellite-derived precipitation. Stage IV is the better product in terms of spatial and temporal resolution, but its coverage does not extend south of the US-Mexico border while the TRMM product does. These two periods approximately reflect the two distinct periods of development of monsoon thunderstorms in the context of the diurnal cycle. Air-mass type thunderstorms develop in early to late afternoon over complex terrain, as indicated by isolated areas of precipitation occurring more over the peaks of the terrain. Larger, more organized convective lines form with more intense precipitation during the late afternoon to evening hours, reflecting the
organization of westward propagating MCSs. The TRMM data indicate that the strongest MCS, with highest rainfall during this period exceeding 20mm, is located in northwestern Sonora, northwest of the city of Hermosillo, corresponding well the areas of lowest OLR and coldest cloud tops on the enhanced infrared imagery of NAME IOP2 shown in Fig. 4.

Since we already know that the modified scheme is mostly improving the ability of WRF to simulate organized MCS-type convection, we show the model simulated precipitation differences between the modified KF minus native KF in Fig. 10. This figure is nearly identical to Fig. 8, but just focuses to the late afternoon to early evening period when the MCSs are most apparent in the TRMM data and enhanced infrared satellite imagery. The difference in model-simulated precipitation on the far right of the figure is entirely consistent with Fig. 8, but the pattern differences previously noted become even more apparent during the peak time of MCS development. Use of the modified KF scheme decreases the precipitation above an elevation of 1000 m, directly over the SMO, and increases below 1000 m, further west in lowland desert areas toward the Gulf of California and away from the mountain range.

5. Value added in a regional climate modeling experiment

Given the improvements observed in the simulation of organized, propagating convection for the NAME IOP2 case, we want to evaluate if the modified KF scheme can help alleviate the noted common biases in the simulation of warm season precipitation within the North American monsoon region in long-term, regional climate model (RCM) simulations. As described earlier in the methods section, the dynamical downscaling of the NCEP-NCAR reanalysis for the period 1950-2010 described in Chang et al. (2015) is
repeated with use of the modified KF scheme. This RCM simulation is categorized as type 2 dynamical downscaling, from the classification of Castro et al. (2005), in which the initial conditions within the regional model are forgotten and lateral boundary forcing may be considered from a retrospective atmospheric reanalysis, as a “perfect” representation of the observed atmosphere.

a) Changes in model simulated monsoon precipitation in the 60 year RCM experiment

The sixty-year average precipitation of July and August is shown in Fig. 11 for the simulation with the native KF scheme, modified KF scheme, and a 0.5° by 0.5° gridded National Oceanic and Atmospheric Administration (NOAA) long-term US-Mexico precipitation data set product (P-NOAA) (Vose et al., 2014) over the region of the Southwest US and northwest Mexico. The maximum in precipitation in the P-NOAA data is located just the west of the crest of the SMO, with precipitation amounts during the monsoon on the order of 200 – 250 mm per month. In the RCM simulation with the native KF scheme, not surprisingly, monsoon precipitation is overestimated directly over the crest of the SMO, with amounts exceeding 400 mm at the highest elevations, nearly double that of observed. The modified KF scheme is able to substantially reduce the model simulated precipitation over the SMO, but about half, and shift the maxima in precipitation further west toward the coast, generally more in accordance with the P-NOAA observations.

The bias pattern in downscaled reanalysis precipitation during the warm season in North America with WRF has been described in Castro et al. (2012), as previously mentioned. Fig. 12, taken from their paper, shows this bias pattern. Generally speaking for the North American monsoon region, WRF when used as a RCM overestimates
precipitation over the highest complex terrain and underestimates it at lower elevations.

We show the warm season monthly differences between the use of the modified KF scheme minus the native KF scheme in Fig. 13. Consistent with the earlier analysis of Fig. 11, the modified scheme decreases precipitation over the highest mountains of the SMO and Mogollon Rim and increases it at lower elevations where organized convection accounts for a greater proportion of monsoon precipitation (Fig. 13). The monthly average precipitation differences can be as high as 60 mm (more than 2 inches) in some places. There is also a notable overall regional reduction in monsoon precipitation during the initiation of the monsoon in June in Mexico. During this time any monsoon precipitation would be from air-mass type thunderstorms that would typically be relatively weaker with higher cloud bases because the atmosphere has not moistened enough. These early season thunderstorms would be more likely to be confined to mountain areas, with less of a tendency to organize. The positive differences in precipitation off the peaks of the mountain ranges become apparent from July onward through September, when the monsoon is in its mature phase and organized convection would be more likely to occur. The basic conclusion is that the same types of improvements in model simulated precipitation demonstrated for a numerical weather prediction-type experiment of NAME IOP2 are also present in the 60-year RCM experiment.

b) Simulating extreme weather events at convective-permitting scales

One of our research interests is to use long-term sources of dynamically downscaled data as a database for simulation of extreme weather events, for short-term numerical weather prediction-type simulations. How would use of downscaled reanalysis
data with the modified KF scheme as the WRF boundary condition affect the simulation of organized convection for specific extreme weather events simulated at a convective-permitting scale, with the CPS shut off? We show a sample simulation result of model-simulated organized convection at 2.5 km grid spacing for an event that took place on 16 July 2006 in Fig. 14. This simulation uses the long term WRF RCM experiments just described with the native KF and modified KF as boundary forcing. WRF model results and the corresponding Stage IV data are shown in the form of a Hovmoller diagram of hourly precipitation averaged latitudinally across the state of Arizona during the peak of the convective event. The Hovmoller figures are constructed such that each point displays the maximum rainfall of the current longitude taken from 30.6 N to 33.6N:

\[ r_m(x,t) = \max_{30.6 \leq y \leq 33.6} r(x,y,t) \]  

(6)

Where \( r(x,y,t) \) is hourly rainfall at longitude \( x \), latitude \( y \), and time \( t \). \( r_m(x,t) \) is latitudinal maximum rainfall at longitude \( x \), and time \( t \). This is to show the propagation of the storm. The comparison of WRF model-simulated hourly precipitation is much better overall on the convective-permitting grid with use of boundary forcing from the RCM simulation with the modified KF. It is able to simulate an organized mesoscale convective system that propagates westward toward the city of Yuma, very similar to the Stage IV observations. By contrast, the same simulation using the boundary forcing from the RCM simulation with the native KF presents a comparatively weaker MCS that does not propagate westward across the entire state of Arizona.

6. Summary and discussion

We have demonstrated the value added of the modified KF scheme originally described in Truong et al. (2009) in dynamical downscaling paradigms of numerical
weather prediction and regional climate modeling with WRF. We first verified that the modified scheme is able to reproduce the results of the extreme precipitation event in Vietnam considered in their original RAMS study. Though the WRF simulated precipitation amounts are different than RAMS for the Vietnam case, we find that use of the modified KF scheme similarly increases the modeled precipitation amounts on a ten kilometer grid such that they are closer to observations. Next, we repeated the NAME IOP 2 WRF modeling experiments described in Cassell et al. (2015) with the modified KF. Clear and dramatic improvements in the model simulations were found in the following aspects: representation of the vertical structure and timing of the gulf surge, organization and propagation of a mesoscale convective system, and alleviation of model-simulated precipitation biases, by decreasing the precipitation over complex terrain and increasing more off the terrain in lowland desert regions toward the Gulf of California.

We also repeated the dynamical downscaling of an atmospheric reanalysis originally described in Chang et al. (2015) with the modified KF scheme. Similar to the NWP-type simulations of NAME IOP2, use of the modified KF scheme yields results that compare better with observed precipitation, alleviating well-known positive biases in the regional model simulated rainfall over the SMO, for example as described in Castro et al. (2012). Monsoon precipitation also increases in more lowland areas at further distance from terrain, where MCS-type convection would account for a greater proportion of the total monsoon precipitation. When the RCM simulations are further dynamically downscaled to convective-permitting scales (2.5 km) in WRF to simulate extreme precipitation events, we find that the use of the RCM simulation with modified
KF also has a positive impact on the representation of organized convection, even though the CPS is not activated on this grid.

The modified KF scheme is better because it produces a more physically realistic representation of the formation and development of mesoscale convective systems during the North American monsoon on the meso-β scale. It thus helps to address a fundamental problem in regional modeling of North American monsoon precipitation that has been noted in multiple studies in the NAME literature, including some of our own prior work. We suggest future work should evaluate the use of modified KF in comparison to convective-permitting simulations and atmospheric model simulations that utilize superparameterization approaches, in terms of the aspects of physical representation of organized convection and computational efficiency.

**Acknowledgements**

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**References**


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**Figures**
**Figure 1:** Illustrative maps corresponding with the study. The top left panel is from the NAME project and shows the NAME domains as defined. The top right panel shows the WRF domains from this study. The bottom left panel shows WRF domain 3 zoomed in, with the locations of cities and landmarks referenced by this study. The MCS cross sectional area is also displayed. The bottom right panel shows the elevation map of the study area.
Figure 2: Accumulated precipitation (mm) for 48-hour of domain 2 (10km resolution) for WRF with native KF scheme (top-left), WRF with modified KF scheme (top-right), Objective analyses of the observed precipitation (bottom-left), and TRMM data (bottom-right). Numbers are the absolute maxima measured at Thuong Nhat station (16.128N, 107.688E).
Figure 3: Accumulated precipitation (mm) for 48-hour of domain 3 (2.5km resolution) for WRF with native KF scheme (top-left), WRF with modified KF scheme (top-right), Objective analyses of the observed precipitation (bottom-left), and TRMM data (bottom-right). Numbers are the absolute maxima measured at Thuong Nhat station (16.128N, 107.688E).
Figure 4: GOES 10 satellite enhanced infrared imagery (4km_ch4_thermal-IR) at 0254 UTC 14 July 2004 during NAME IOP 2 (NCAR/EOL).
Figure 5: Puerto Peñasco (a) wind profiler and (b) surface data from 0000 UTC 12 July to 0000 UTC 15 July. In (a), wind speed (m s$^{-1}$) is plotted every half-hour and wind barbs every 2 h. One full barb equals 5 m s$^{-1}$. Data not plotted for times and heights of low data confidence. In (b), surface temperature (°C, red), dewpoint temperature (°C, blue), and pressure (hPa, green) are averaged and plotted every half-hour. This figure is from Rogers and Johnson (2007), their Fig. 4.
Figure 6: WRF model simulation of the gulf surface at Puerto Peñasco during NAME IOP 2 from native KF scheme (top), and modified KF scheme (bottom), presented equivalently the wind profiler observations as shown in Fig. 5a.
Figure 7a: Crosssection of modified WRF model simulated total water condensation (kg kg$^{-1}$) from 7Z to 12Z 14 July. Wind vectors are scaled such that the horizontal wind is ten times larger than the vertical wind in order to show updrafts and downdrafts. The vertical planes to construct the cross section are defined at latitude 26.8 N extending in the x-axis from 109.5 W to 106.5 W. The y-axis is log pressure scaling from 1000mb at surface to 100mb near the tropopause.
Figure 7b: Cross sections of WRF model simulated radar reflectivity (dBZ) (left panels) on domains 2 and 3 at 0300 UTC 14 July. Wind vectors on radar reflectivity panel are scaled such that the horizontal wind is ten times larger than the vertical wind. The vertical planes to construct the cross section are defined intersecting a point at 29.69° N and 111.4° W and extending along constant latitude from 112.5° W to 107.5° W, and the frames have a height of about 20 km. They are averages of parallel planes north and south of the center and extend through the depth of the model. From Cassell et al., 2015 figure 11.
Figure 8: Accumulated precipitation (mm) for 24-hour (12 UTC 13 July to 12 UTC 14 July) of NAME IOP2 for domain 2 (10km) of WRF with native KF scheme (top-left), WRF with modified KF scheme (bottom-left), Tier-1 observation (top-right), and different between 2 WRF simulations: Modified minus Native (bottom-right). Thin black contour is 1000m elevation.
Figure 9: Accumulated precipitation (mm) for the Stage IV radar data (left), and TRMM satellite data (right), respectively, for the 6-hour periods 1800 UTC 13 July to 0000 UTC 14 July (top) and 0000 UTC 14 July to 0600 UTC 14 July (bottom) during NAME IOP 2.
Figure 10: Accumulated precipitation (mm) for the WRF with native KF scheme (left), WRF with modified KF scheme (middle), and difference of them (Modified minus Native: right), respectively, for the 6-hour periods 0000 UTC 14 July to 0600 UTC 14 July. Thin black contour is 1000m elevation.
Figure 11: Monthly mean precipitation (mm month$^{-1}$) averaged over 60 year (1951-2010) of July (top) and August (bottom) for WRF with native KF scheme (left), P-NOAA observation (middle), and WRF with modified KF scheme (right).
Figure 12: Average warm season (JJAS) precipitation bias (mm day$^{-1}$) for WRF-NCEP 
(P-BIAS = RCM minus P-NOAA observation). Red (blue) colors indicate precipitation 
overestimation (underestimation) by WRF. From Castro et al, 2012.
Figure 13: Differences of monthly mean precipitation (mm month$^{-1}$) averaged over 60 years (1951-2010) from June to September. Difference = Modified KF minus Native KF.
Figure 14: Hovmoller diagrams of hourly precipitation of the main storm propagating West-Northwesterly toward Yuma on July 16th 2006 for Modified KF scheme run (top), Native KF scheme run (bottom right), and Stage IV observation (bottom left).
APPENDIX B: THE IMPACT OF URBANIZATION ON THE NORTH AMERICAN MONSOON PRECIPITATION IN ARIZONA WITHIN THE CONTEXT OF MODELED SEVERE WEATHER EVENTS

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Abstract

This study investigates the effect of incorporating an urban canopy model (UCM) and urban land cover, in a series of sensitivity experiments for severe weather events during the North American monsoon in Arizona. One single severe weather event is first simulated with the Weather Research and Forecasting (WRF) model at a convective-permitting scale (2.5 km grid spacing). The “urban” simulation uses the UCM and 2006 land cover classifications and the “desert” simulation replaces the Phoenix and Tucson areas with their pre-settlement, natural desert land cover. The particular event took place during the period 13-14 August 2003, a date with recorded severe weather impacts in the Phoenix area. The “urban” simulation produces less precipitation within the urban core of the city and more precipitation around the periphery of the city. Instead of being a “heat island” like cities of comparable size in the eastern and central United States, the presence of vegetation and irrigation makes Phoenix a “heat sink” in comparison to its surrounding desert. Greater partitioning of surface energy to latent heat fluxes causes the boundary layer to be suppressed and local increase in stability and surface pressure. Repeating the same experiment on a series of objectively identified severe weather events in the Phoenix area for the period 1991-2010 confirms that the result of the 13-14 August 2003 test case applies to an entire climatology of severe weather events. The model sensitivity study results support ideas brought forth in several earlier studies that suggested urbanization has had a statistically significant impact on observed long-term changes in the surface meteorology and precipitation in the Phoenix area during the monsoon.
1. Background and Motivation

During the recent historical period of approximately the last thirty to forty years, there has been rapid urbanization and population growth within the southwestern United States. In particular, the Phoenix, Arizona, area has greatly expanded in geographic size, resulting in an overall change from a natural desert-type vegetative land surface cover to one with a combination of urban development and heavily irrigated areas (Georgescu et al., 2008). Large urban regions in the middle of deserts occupying on the order of a thousand square miles in geographic area, like Phoenix or Las Vegas, modify their surrounding environment in a variety of ways. Changes in land use/land cover alter the surface energy and moisture exchanges with the atmosphere and thus the thermodynamic characteristics of the planetary boundary layer. The increase in the surface roughness, due to buildings, increases mechanically generated turbulence downwind and generates upward vertical motion (Bornstein and LeRoy, 1990; Diem and Brown, 2003). Irrigated croplands (e.g. cotton fields), like those southeast of Phoenix, as well as recreational irrigated lands (e.g. golf courses) have substantially greater evapotranspiration as compared to that of the surrounding natural desert vegetation. This increase in atmospheric moisture effectively allows for a decrease in atmospheric moist static stability and the generation of vertical motion in the synoptic-scale which act to enhance convection in the region and thus precipitation (Diem and Brown, 2003). Increased pollution in urbanized regions can lead to greater amounts of cloud condensation nuclei. Cloud growth may thus be enhanced but the clouds exhibit smaller cloud droplets as compared to unpolluted rural areas (Diem and Brown, 2003). Cloud microphysical processes in polluted urban environments may either enhance or suppress precipitation,
dependent on the specific climate conditions and physiography of a given location (Shepherd, 2005).

The most well recognized effect of cities on modifying their surrounding environment is that urbanized surfaces reduce albedo and absorb more of the incoming solar radiation that is converted to sensible heat flux. In large cities surrounded by highly vegetated natural landscapes like forests and croplands the city will thus tend to have a “heat island.” Higher surface temperatures and deeper planetary boundary layers (PBL) are present over the city, as compared to the surrounding rural areas (e.g. Rozoff et al., 2003). However, a city like Phoenix surrounded by a desert is quite different in this respect. The added moisture from the irrigated lands introduces a relatively large amount of latent heating, or evapotranspiration, as compared to the surrounding desert. The increase in moisture within the urban area can actually reduce surface temperature as compared to the surrounding rural areas, not increase it. A desert urban area like Phoenix is essentially a “heat sink” rather than “heat island” (Diem and Brown, 2003).

It has been well documented in cities in the central and eastern United States, that the deeper PBL in association with the urban “heat island” causes low-level convergence that is favorable for the enhancement of convective precipitation downwind of the city (Bornstein and Lin, 2000, Rozoff et al., 2003, Cotton and Pielke, 2007). The numerous case examples evaluated in the context of atmospheric model sensitivity experiments and observational analyses include New York City, New York (Bornstein and Leroy, 1990); Atlanta, Georgia (Bornstein and Lin, 2000; Shem and Shepard, 2009); St. Louis, Missouri (Changnon, 1981; Rozoff et al., 2003); Chicago, Illinois (Changnon, 2001); Houston, Texas (Orville et al., 2001; Burian and Shepherd, 2005; Shepherd et al., 2010)
and Oklahoma City (Hand and Shepherd, 2009). However, there has been comparatively less research on anthropogenic land-use change has affected convective precipitation in a desert city like Phoenix, especially in terms of numerical atmospheric modeling in the vein of the Rozoff et al. (2003) study for St. Louis, where the land use/land cover classifications are changed in a model sensitivity experiment.

Increases in moisture, instability, and precipitation during the summer months in Arizona is associated with the North American Monsoon System (NAMS) in mid to late summer (July-September), an important regional climatological feature of the Southwest United States that accounts for approximately half the annual precipitation (e.g. Adams and Comrie, 1997). During the monsoon, organized convection produces severe weather, for example flash flooding, strong winds, dust storms, and lightning (e.g. Ray et al., 2007). If there is any impact of urbanization on thunderstorm development and evolution in the Phoenix area it should be most evident during this time of the year. Through explicit moisture tracing in a modeling framework, it is known that monsoon moisture may come from both the land surface, through local evapotranspiration, and atmospheric moisture transport from the Gulf of California, Eastern Pacific, and Gulf of Mexico (e.g. Hu and Dominguez, 2015). With respect to specifically in south central Arizona, low level moisture from the Gulf of California, in the form of Gulf surges (Rogers and Johnson, 2007) and mid-level moisture from the Gulf of Mexico are probably the most important.

Monsoon precipitation is generated by thunderstorms on the mesoscale, and intimately linked to the diurnal cycle of heating of terrain and generation of mountain-valley circulations (e.g. Diem and Brown, 2003). The strong influence of complex
terrain on convective development is another key difference of Phoenix as compared to cities in the central and eastern U.S., most of which are not adjacent to a large mountain range. Monsoon thunderstorms begin to develop generally over the areas of highest terrain in the early afternoon. Synoptic-scale features, for example transient inverted troughs (Pytlak et al., 2005; Bieda et al., 2009; Seastrand et al. 2014; Lahmers et al., 2015), facilitate quasi-geostrophic upward vertical motion and provide favorable wind shear profiles for convective organization and propagation off the high terrain. In these types of conditions, organized convection, in the form of mesoscale convective systems and squall lines, is able to propagate westward into the Phoenix area with easterly mid to upper level steering flow in the atmosphere (around 600-mb). These storms can originate on either Mogollon Rim, to the northeast of the city, or in the mountain ranges near the city of Tucson (Catalinas, Santa Ritas, and Rincons) to the southeast. Convective propagation is made possible through successive outflow boundaries, or cold pools, from leading convective lines. The outflow boundaries mechanically lift moist and unstable air at the surface thus supporting more convection (Corfidi, 2003). Resultant mesoscale convective systems may be sustained well into the late evening hours, long after reaching maximum in diurnal heating.

The question of interest in this work is does the aforementioned “heat sink” due to the presence enhanced vegetative cover in urbanized area of Phoenix locally affect the monsoon thunderstorms? There have been some observational analyses that at least suggest, from a statistical standpoint, that there has been an impact of urbanization on Phoenix on monsoon precipitation. Balling and Brazel (1987) showed there is a local blocking pattern in the isobaric surfaces in the city by analysis of surface pressure data.
They suggested that the blocking pattern inhibits the cold pools from propagating into the city and thus suppresses convective activity. Shepherd (2006) evaluated long-term changes in observed mean monsoon precipitation from rain gauge measurements within and surrounding the Phoenix area, from a pre-urban period (1895-1949) and a post-urban period (1950-2003). They found a statistically significant increase (more than 10%) in monsoon rainfall in stations located to the north and east of the city, and speculated that landscape and agriculture in the Phoenix urban core enhance the moisture available for convection.

One of the important few series of studies that have investigated the impact of urbanization in Phoenix explicitly through an atmospheric model sensitivity experiments is that of Georgescu (2008) and Georgescu (2009a,b) with the Regional Atmospheric Modeling System (RAMS). They physically changed the land use/land cover tiling to reflect urbanization over time and then conducted seasonal model simulations with a convective-permitting (2 km) grid spacing. In their 2008 paper, high resolution (2 km) simulations of the North American Monsoon region were conducted for 3 “dry” summers and 3 “wet” summers, to evaluate how precipitation is affected by changes in land surface cover. Two types of land cover surfaces were used in that experiment: a natural surface like that of pre-1900 achieved by pixel replacement and an urbanized scenario based on satellite observations circa 1992 (Georgescu et al., 2008). During the “dry” years there was a noticeable increase in precipitation generally east and north of the urban area for the urban landscape. However, the “wet” years showed no discernible precipitation change in the domain as a result of land surface change. Their later work (Georgescu et al., 2009a, b) similarly replaced the land use tiling according to 1973,
1992, and 2001 land use datasets and simulated a series of wet and dry years. Though there was not a consistent and clear precipitation response to the land use change for both wet and dry years, they did find some very interesting results for simulations of dry years where they hypothesized the influence of the land surface on the atmosphere is more important. For example, in their simulations for July 1979 and July 1989 they found increases in precipitation to the north and east of Phoenix at least in some of their simulations with the more recent land use specifications, conceptually matching the results of Shepherd et al. (2006). Another study to evaluate the changes in regional climate that could arise due to projected urbanization in the Phoenix-Tucson corridor has been published recently using land cover data for 2005 and projections to 2050 (Yang et al., 2015). Ten-year high resolution simulation (2 km grid spacing) shows reductions in precipitation over the mountainous regions northeast of Phoenix and decreased evening precipitation over the newly-urbanized area. This precipitation results however is not robust and inconclusive. They hypothesize that because the ambient air is very dry, impacts of urbanization on energy partitioning at the surface will not result in significant changes in precipitation because there is simply not enough available specific humidity.

Our objective in this study is to evaluate how changes in land use/land cover in the Phoenix urban area impact the most severe thunderstorm events during the North American monsoon. The methodological approach is quite similar to previous works of Georgescu et al. (2008, 2009a, b) and Yang et al. (2015), in that we consider changes in land use/land cover associated with urbanization in convective-permitting simulations. But there is important difference: we evaluate changes in most severe event-based
numerical weather type simulations, rather than long term continuous regional climate model-type simulations and.

Our underlying hypothesis that the impact the urban heat sink effect of Phoenix on monsoon precipitation should be most apparent during days when the strongest thunderstorms associated with organized convection occurs, and there should be a physically consistent and clear response with the model simulations of the most severe weather events, in terms of changes in simulated thunderstorm evolution and precipitation patterns.

2. Analysis and Methods

a. Severe weather event selection and overview of model simulation strategy

We simulate only objectively identified individual severe weather events in a numerical weather prediction (NWP) mode with a convective-permitting regional atmospheric model. The specific severe weather event days in Arizona during the monsoon are selected from a long-term dynamically downscaled NCEP-NCAR Reanalysis (1950-2010) with the Weather Research and Forecasting (WRF) that was recently described in Chang et al. (2015). The event days are identified in this long-term regional climate model simulation based on thresholds in thermodynamic conditions of heat and moisture. The thermodynamic variables we use are the daily maximum values of column-integrated precipitable water (PW) and most unstable convective available potential energy (MUCAPE). The days with the top 20% of MUCAPE and PW within the Southwest U.S. are selected for numerical weather type-simulations at convective-permitting scale, to be described later, and are defined as severe weather event days (Jares et al., 2014). The final list of severe weather event days we found
corresponds with approximately 70% of the recorded severe weather reports in southern Arizona during the monsoon over the last twenty years of simulation (Jares et al., 2014). That same work also found that the upper-level 500mb monsoon ridge is in a favorable orientation for severe weather in Arizona during the identified severe weather event days, matching the first two severe weather configurations described originally in Maddox et al. (1995), with the center of the ridge located to the north and west and north and east of Arizona. In these configurations, easterly flow is favored at upper-levels, facilitating thunderstorm organization and westward propagation off the mountain ranges toward the Phoenix area. All the severe weather event days we identify exceed the known operational thresholds for forecasting severe monsoon thunderstorms per operational practice of the National Weather Service (NWS) in Tucson, with CAPE values exceeding 500 J kg⁻¹ and PW values exceeding 25mm (~1 in).

The strategy for simulating identified severe weather event days and evaluating results is as follows. First, we consider one very obvious and well documented severe weather event that occurred in the Phoenix area and retrospectively model it in a short-term, numerical weather forecast-type mode, in a series of two simulations accounting for desert and urban land use/land cover. The specific event is during 13-14 August 2003, to be described in more detail in the next subsection. We evaluate these particular simulations in detail as an idealized case study, in the context of the changes in model-simulated precipitation and the physical causes. Next, we simulate all the identified severe monsoon weather events with high CAPE and PW criteria in the downscaled reanalysis for the period 1991-2010, a total of 255 events. However, because these events are based on thermodynamic conditions regionally within the entire Southwest
U.S., precipitation in Phoenix does not necessarily occur during all of these days. So we further sub-selected a series of 25 of these severe weather event days, to specifically investigate the issue of sensitivity to land use/land cover. This subset of what we consider the most extreme of the severe weather event days falls in the top 10% of the 255 days, in terms of the simulated precipitation in Phoenix at the convective-permitting grid spacing. They also all satisfy the condition that upper-level steering winds at 500-mb must be from southeasterly to northeasterly direction. We evaluate whether the changes in precipitation and causes for it determined in the August 2003 test case apply in a statistical sense to the lot of 25 most severe modeled weather events in Phoenix area since 1991. Thus, in contrast to the previous approach in the studies by Georgescu et al. (2008, 2009a, b), we have specifically designed our model experiments in this study to only focus on the effects of land use/land cover change on monsoon thunderstorms on the days when the most precipitation actually occurred in Phoenix and is most likely to be associated with well-organized, propagating, and extreme convection.

b) Summary of severe weather event 13-14 August 2003

The specific severe weather event we choose to model first as an idealized test case occurred during the 24h period of 13-14 August 2003. The observed background atmospheric environment during this day is illustrated on Fig. 1, by the 500-mb geopotential height analysis and Phoenix morning sounding, taken at the Salt River Project site. The 500-mb map shows an upper-level ridge centered in east central Texas. Though this is not quite an ideal position of the 500-mb ridge for severe weather in Arizona by the Maddox et al. (1995) classification, nonetheless winds at 500-mb are generally from a favorable easterly direction (recorded as northeasterly in Tucson), such
that any thunderstorms that develop along the Mogollon Rim during the day may propagate toward the Phoenix area. An “inverted V” morning (12 UTC) sounding on 13 August is indicative of microbursts, in which precipitation from thunderstorms evaporates in a deep, dry boundary layer, producing negatively buoyant air that sinks rapidly to the surface. The boundary layer extends up to approximately the height of 600-mb, as the temperature decreases dry adiabatically to this point. Relative humidity in the boundary layer is on the order of 30-40%, for example 30% at 900-mb, with a relatively more moist layer above. Winds from the sounding above 500-mb are generally from an easterly direction. During this particular day the MUCAPE in Phoenix recorded in the 12 UTC sounding was 892 J kg\(^{-1}\) and PW was 39.1 mm (Fig. 1), satisfying NWS thresholds for severe monsoon thunderstorms.

This day was quite typical for a monsoon severe weather day, with development of convection over the mountain ranges in mid to late afternoon and convective organization and propagation into the Phoenix urban area during the late afternoon to evening hours. Impacts of this event can be found on the NOAA Storm Events Database maintained by the National Climate Data Center (NCDC) (www.ncdc.noaa.gov/stormevents), and these include during the evening of 13 August: 50-60 kt wind gusts from microbursts recorded in Maricopa and Pinal counties from 8-10 pm local time, which damaged a trailer after being blown 100 feet; hail up to 2 inches in diameter; weak tornadoes (F0); and widespread flash flooding that caused two fatalities. It is also interesting to note that during the defined “wet” years in the Georgescu et al. studies, they suggest the impact of urbanization on convective precipitation becomes less
apparent. At least for this event, we will show that urbanization does seem to play an important role in the context of a “wet” environment.

c. Convective-permitting regional atmospheric model simulations

For the August 2003 case study and the rest of the identified severe weather event days as described in the previous sub-section a, the Weather Research and Forecast (WRF) model version 3.4 (Skamarock et al., 2005), is used to generate the convective-permitting simulations over the Southwest. The thermodynamically favorable severe weather event days for the period 1991-2010 during the monsoon seasons (from June to September) are retrospectively simulated. The dynamically downscaled NCEP-NCAR long-term RCM simulation (1950-2010) with grid spacing of 35 km is used to identify these days, as mentioned, and provide lateral boundary forcing to produce short-term numerical weather forecast-type simulations of 24 hours duration of a convection day from 12 UTC of the current day to 12 UTC of the next day, with a six hour model spin-up period. The model grid structure is shown in Fig. 2. An intermediate domain with 10 km grid spacing covers the southwestern United States and northwestern Mexico. The convective-permitting domain with 2.5 km grid-spacing extends over the entirety of Arizona and New Mexico and also includes portions of California, Colorado, Nevada, Texas, and Utah. The first six hours of the simulation before the 24 hour analysis period is the model spin-up period, and so is not considered for calculating total precipitation or the modeled physical behavior of the atmosphere.

Parameterizations used in these WRF model simulations are mostly consistent with what we previously used to simulate organized convection during the North American monsoon experiment (Cassell et al., 2015). This includes use of the Morrison
double moment scheme for microphysics (Morrison et al., 2005), the NOAH land surface model (Ek et al., 2003), the Eta surface layer (Janjic, 1996, 2002), and the Mellor-Yamada-Janic planetary boundary layer scheme (Janjic, 1990, 1996, 2002). It additionally uses Dudhia Shortwave radiation (Dudhia, 1989) and the Rapid Radiative Transfer Model (RRTM) Longwave radiation (Mlawer et al., 1997). An important addition for this study is the use of a one layer Urban Canopy Model (UCM) within NOAH applied over urban areas with optimized anthropogenic parameters for the southwestern United States, as adapted from Grossman-Clarke (2010) and described in more detail in the next subsection. The specific areas in which the UCM is applied are those in which the land use tiling is defined as “urban” in Fig. 3. The urban land use tiles then are replaced with natural desert land cover in the “desert” simulation, as also shown on the figure. Though our main interest here is the effect of the Phoenix area on convection, it is important to note that the city of Tucson is also clearly indicated as urban area, albeit about a third as large, in the land use tiling.

d. Urban canopy model and land surface scenario sensitivity experiments

In WRF, the Noah UCM and LSM are applied to the fraction of a model grid cell with built and natural surfaces, respectively. The Noah UCM considers urban geometry in the surface energy balance and momentum flux calculations (Chen et al., 2011). A multi-layer heat conductivity equation is solved for roof, wall, and road temperature profiles. Sensible heat fluxes from the respective surfaces are aggregated into the total flux \( H_{\text{urban}} \). An anthropogenic heat flux, \( Q_d \), is activated and added to the \( H_{\text{urban}} \). Appropriate \( Q_d \) for Phoenix were derived by the Sailor and Lu (2004) method. Maximum \( Q_d \) occur during the evening rush hour (LST 1700) and amount to \(~30\text{W m}^{-2}\) with slight
variation between urban land use/land cover (LULC) classes (Grossman-Clarke et al., 2005). Three urban LULC classes are included in WRF for Phoenix (state of LULC in 2005): commercial/industrial, mesic residential and xeric residential, which are distinguished by the fractional cover of built, vegetation and soil surfaces, building heights, roof and road widths (Grossman-Clarke et al., 2010, their Fig. 2). Here these LULC have been adjusted: commercial–industrial, urban mesic residential and urban xeric residential, which are distinguished by the type of vegetation and irrigation (no vegetation, well-watered flood or overhead sprinkler irrigated, and drought-adapted vegetation with drip irrigation, respectively). The fractional cover of vegetation surfaces comprises 0.10 in the xeric residential and 0.23 in the mesic residential categories.

We apply the UCM over all areas defined as urban in the Southwest in 2.5 km domain. In order to sustain landscape vegetation and agricultural productivity in the Phoenix region, irrigation is necessary all year. The Noah LSM does not account for irrigation but assumes soil moisture contents as obtained from WRF’s initial conditions. The preferred practice for urban vegetation is drip irrigation over an extended period of time. This technique ensures sufficient water in the root zone for plant transpiration, but soil surfaces in between plants are usually not irrigated resulting in low soil evaporation. In order to account for the irrigation in WRF, the initial soil moisture content in the urban areas was increased for the sub-surface layers but not for the top soil layer. For urban vegetation, as well as irrigated agricultural land, the soil moisture content was adjusted in WRF to the reference soil moisture that corresponds to field capacity.

The predominant soil categories in the region are “sandy loam” and “loam,” with field capacities of 0.383 and 0.329 m$^3$ m$^{-3}$ and wilting points of 0.047 and 0.066 m$^3$ m$^{-3}$. 

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In the simulations the soil moisture content for the sandy loam dropped to 0.35, 0.36, and 0.32 m$^3$ m$^{-3}$ for soil layer 1 to 3 and stayed at field capacity for layer 4. Plants did not experience significant water stress. In comparison the climatological soil moisture content as obtained from the NCEP Eta Model initial data are 0.10, 0.11, 0.12, and 0.14 m$^3$ m$^{-3}$.

Similar to that of Georgescu et al. (2008, 2009a, b), two land surface scenarios are used in two model sensitivity experiments for any given severe weather event: 1) an urbanization scenario which uses current land coverage of the Phoenix region (referred to the “urban” simulation) and 2) a natural vegetation scenario (referred to as “desert” simulation). The two simulations are compared to evaluate any salient differences. Precipitation plots computed are the accumulated precipitation and the diurnal cycle. The diurnal cycle considers hourly increments of precipitation to show the propagation of convection across the region. Additionally, to further illustrate the propagation of convection through the Phoenix area, vertical cross sections of the vertical winds are extracted along a storm track. These plots show the cold-pool propagation of thunderstorms, via an updraft-downdraft coupling. Atmospheric sounding profiles before and after the convection (12Z, 5AM local time) in Phoenix are shown to see how the initial thermodynamic conditions change. For verification, the quality controlled Stage IV radar data (based on NEXRAD and rain gauge measurements: http://data.eol.ucar.edu/codiac/dss/id=21.093) is used to compare observed measured rainfall to modeled rainfall.

3. Effect of urbanization for severe weather event of August 13, 2003
The simulation of August 13, 2003 (the selected severe weather event) is used as the test case to evaluate the land coverage effects on precipitation. As mentioned, the land surface covers for the “urban” simulation have the city signature of both Phoenix and Tucson present in the convective-permitting domain through use of the Urban Canopy scheme (Fig. 3). Information regarding how precipitation changes and why this might be the case, from the standpoint of changes in meteorological conditions at the surface and in the local atmosphere in the Phoenix and Tucson areas, are described in the subsequent three sub-sections.

a. Changes in model-simulated precipitation and relationship to urbanization

Observed and modeled precipitation for Phoenix and Tucson is shown in Fig. 4. Consistently more rain occurs over the cities for the “desert” simulation compared to the “urban” simulation. Using urban land-use land cover and the UCM in “urban” simulation reduces the daily average precipitation in both Phoenix and Tucson, by over 7 mm day\(^{-1}\) and nearly 3 mm day\(^{-1}\), respectively. These results are closer to the observed Stage IV precipitation at the equivalent model grid points, particularly in the urban areas of Phoenix where little rainfall actually occurred (less than 1 mm day\(^{-1}\)). These total precipitation values in Fig. 4 just show the change in simulated precipitation within the defined (urban tiled) metro areas of Phoenix and Tucson per Fig. 3. A spatial representation of the accumulated precipitation is needed to indicate the specific geographic locations in Arizona where precipitation is locally maximized or minimized. Fig. 5 shows the model simulated and the equivalent Stage IV observed precipitation over southern Arizona for the local time period 5am 13 August 2003 to 5am 14th August 2003, along with the difference in simulated precipitation from the “urban” minus
“desert” simulation. The model simulated precipitation amounts in areas that are relatively far (generally greater than a 100 km) from Tucson and Phoenix is relatively similar for the two simulations, for example in the northern part of Arizona near Flagstaff. Comparing the “urban” to the “desert” simulation in the Phoenix area it is apparent that less precipitation occurs in the “urban” simulation precipitation, with a simulated average daily precipitation difference that is nearly 10 mm. There are also decreases in model simulated precipitation in Tucson, though the area of precipitation decrease is not geographically as large. More precipitation occurs in areas immediately surrounding the Phoenix urban core to the northwest and northeast of the city in the “urban” simulation. In particular, the daily average precipitation increases in the northwest of Phoenix is on the order of 10 mm. However, to the south and west of the city, outside the urban core, there are decreases in precipitation in the “urban” simulation, notable because this area is immediately downwind of the Phoenix urban core given the mean upper-level steering flow at 500-mb (Fig. 1). A clear spatial pattern of precipitation differences between both the “urban” and “desert” simulations is apparent (Fig. 5, bottom left).

In the equivalent observed Stage IV precipitation for the date, there is a clear local minimum in precipitation immediately within the Phoenix urban core, with local maxima in precipitation values immediately to the east and northwest of the city. The “urban” simulation much better captures the pattern of a local precipitation minimum in the Phoenix urban core, with local precipitation maxima occurring in areas that ring the urban core generally to the north and east. By contrast, in the “desert” simulation there is no discernible difference in precipitation amounts between the urban core and its
surrounding areas. It should be noted that the WRF model simulations tend to overestimate precipitation rates in the whole domain, at least in comparison to the Stage IV product. Fig. 6 shows the same information as Fig. 5 but zooming in to Phoenix metropolitan area. At this scale, the precipitation difference between the “urban” minus “desert” simulation even more clearly shows suppressed convection in the urban core and immediately downwind of it.

b. Evolution of monsoon thunderstorms in relation to Phoenix urban “heat sink”

Evaluation of this severe weather event at an hourly time scale gives more physical insight to storm evolution and how sensitive that process is to the presence of Phoenix as a large urban area. There are two distinct periods of modeled precipitation that impacted the city during this particular severe weather event, though they appear differently in the “urban” and “desert” simulations, as shown in the hourly precipitation totals in Fig. 7. These two events will be referred as the first and second round storms. The “desert” simulation has the first round storm occurring in the evening after sunset, with precipitation on the order of 0.3 mm hr⁻¹ near 9 pm local time, and a second round storm occurring early in the morning hours around 3 am local time, with precipitation on the order of 2.5 mm per hr⁻¹. The “urban” simulation also shows two peaks in hourly precipitation on the order of 0.5 mm hr⁻¹ over the same time period. However, the first round storm does not occur at the same time as in the “desert” simulation, but rather about two hours later at 11 pm local time. The second round storm also occurs around 3 am local time, but with a much reduced precipitation amount.

Decreases in temperature also occur corresponding to these thunderstorms. The diurnal cycles of air temperature at 2m for both simulations are plotted in Fig. 7. The
temperature of the “urban” simulation is higher in late morning to mid-afternoon (on the order of 3°F). However, before sunset to midnight the “desert” case has an overall higher simulated air temperature. At midnight, the “desert” case begins to have a lower temperature. During the most active rainfall period (8pm to 5am), both the “desert” and “urban” simulations have a dramatic temperature decrease, with the steepest gradient from “desert” simulation (-9.6°F compared to -7.3°F of “urban” simulation during the time period) (Fig. 7). The greater decrease in temperature in the “desert” simulation is a result of greater precipitation from the more intense second round storm.

The propagation of storms during both the rounds is evaluated in the “desert” and “urban” model simulations at an hourly time scale in Fig. 8-9. The first round storm in both the “urban” and “desert” runs occurs during the hours of 8pm to 11pm local time on 13 August 2003 (Fig. 8). It approaches the Phoenix area from the northeast, traveling southwestward toward the city from the Mogollon Rim. As the storm reaches the northeast extent of the Phoenix urban area in the “desert” simulation, it weakens but continues propagating through it with reduced rainfall amounts. However, as the same first round storm approaches Phoenix in the “urban” simulation it almost completely dissipates as it reaches the city and then reorganizes north and west of the city. Thus, the convection avoids the metro area and continues to the northwest. Though it tends to bypass Phoenix, the convection of the first round storm in the “urban” simulation outlasts that of the “desert” run by a few hours once the convection reorganizes.

The second round storm affects the Phoenix area from 1am to 4am local time on 14 August 2003, as seen in the hourly sequence on precipitation in Fig. 9. The “desert” simulation of the second round storm shows that a convective cell moves directly over
the Phoenix urban area. From 1 am to 2 am local time the eastern portion of Phoenix (east of 112°W) and from 3 am to 4 am the western portion of Phoenix (west of 112°W) experiences rainfall. The western portion of the city experiences higher and more prolonged rainfall rates than the eastern portion. While the entire city of Phoenix experiences rainfall in the “desert” simulation for the second round storm, this does not occur in the “urban” simulation. During the peak time of precipitation in the “desert” simulation, the second round storm cell in “urban” run instead moves to the southeast only precipitates over the very southern extent of Phoenix from 1 am to 3 am local time. Overall, we observe that both storms in the “urban” simulation tend to bypass the metro area as compared to the “desert” simulation, either moving to the right (first round storm) or left (second round storm) of the upper-level steering wind.

Does the presence of an urban “heat sink” help explain the deviation of the modeled, propagating thunderstorm cells around the Phoenix area in the “urban” simulation? To more specifically consider this question, we show the diurnal cycle of the modeled surface heat fluxes (sensible and latent) in Fig. 10. During the daytime hours, the sensible heat flux of the “desert” simulation is higher than the “urban” run (on the order of 50 W m⁻²). Reasons for this are 1) plant cover in agricultural areas reduces the amount of absorbed solar radiation and subsequently heat storage in the agricultural soils and 2) irrigation. Evaporation reduces heat storage relative to desert soils even though the albedo of moist soils is lower and heat capacity higher. From 7 pm to 9 pm local time, both the “desert” and the “urban” simulations show a sharp decrease in sensible heat flux, as this is the time at which the first round storm occurs and cools down the surface. The sensible heat flux of the “urban” simulation becomes just slightly higher than that of the
“desert” simulation at night, due to a lesser amount of precipitation occurring over Phoenix during the second round storm.

The latent heat flux of the “urban” simulation is larger during the day due to irrigation within the city of Phoenix, with differences on the order of 100 W m$^2$. There is basically comparatively little difference in latent heat flux between the simulations after sunset. The most important point to be noted in evaluation of the surface heat fluxes, consistent with the “heat sink” idea, is that the urban area tends to lower the sensible heat fluxes and increase the latent heat fluxes during the daytime hours. These simulated diurnal heat fluxes are consistent with Grossman-Clarke (2010) and Brazel et al. (2000). The latter study found no significant increase in average maximum June daytime air temperatures after 1940 in the center of the urban area, and on average slightly higher maximum daytime temperatures at a rural site located southeast of the Phoenix metropolitan area.

The evolution of PBL height between the simulations is shown in Figure 11. The “heat sink” behavior of the “urban” simulation, with greater portioning of the surface heat flux to latent heating, leads to a lower PBL height in Phoenix during the entire day prior to the occurrence of any convective activity. The lower PBL height therefore locally increases the atmospheric stability over the urban area, in comparison to the surrounding desert. In the last section of the results, we will explicitly show that the use of the urban scheme locally lowers the MUCAPE over Phoenix for this reason.

c. Vertical cross section of vertical motion along direction of storm propagation

Although convective propagation can be assessed somewhat in the plots of the hourly rainfall rates in Figs. 8 and 9, examination the vertical motion of the simulated
storms provides additional physical insight. Our goal is to evaluate if the cold-pool convection propagation mechanism exists and how it may be different in the “desert” and “urban” simulations. A cross section of the second round storm is taken, that extends from the low-lying area southwest of Phoenix, through Phoenix, to the higher elevation areas of the Mogollon Rim (Fig. 12). This cross sectional axis reflects the direction of storm propagation from the point of convective initiation in the mountains. In a propagating storm we would expect to see coupling of updrafts and downdrafts. Fig. 13 shows the vertical cross sections of the vertical wind component at 12 AM (1st row), 1 AM (2nd row), 2 AM (3rd row) and 3 AM (4th row) local time for the “urban” run (left) and “desert” run (right). At 12 AM, the “urban” simulation shows a well-developed storm cell with updraft-downdraft coupling over the higher terrain. At the same time, the “desert” simulation shows two downdrafts over the higher terrain with two updrafts northwest of Phoenix and on the edge of the city. Progressing to 1 AM, the “urban” simulation shows comparatively weaker vertical motion, implying that there is little to no convection along the cross section axis. However, in the “desert” simulation the convection from the previous time has become more intense and organized. Several coupled updrafts and downdrafts reach the edge of the city (Fig. 13). Continuing to 2 AM, the “urban” simulation continues to show little to no evidence of convection. The “desert” simulations shows updraft-downdraft couplets reflecting storm passage directly through the city. Although this storm appears to have weakened from the previous time step, it is still propagating with the correct structure to the southwest (downwind). In the last cross section at 3 AM local time, there is again no signature of strong updrafts or downdrafts in the “urban” simulation cross section. In the “desert” simulation, the storm
indicated by the downdraft and updraft couplets has continued to propagate southwest, traversing the entire city by this point. From these cross sections in Fig. 13, we can basically observe that the second round storm in the “desert” simulation is propagating through Phoenix while in the “urban” simulation it is not.

4. Results for the series of modeled severe weather event cases (1991-2010)

Twenty-five of the most severe weather events around Phoenix out of 255 are selected objectively to form a statistical set, as described in the Methods section. The specific events are shown in Table 1. During these events there are always modeled monsoon thunderstorms in the vicinity of Phoenix. The objective threshold criteria for defining a severe monsoon precipitation event in Phoenix are 1) there is rainfall of more than 1 mm spatially averaged within 100 km$^2$ (4x4 grid box) of the city, and 2) MCSs propagation generally following easterly upper-level steering flow. Locally statistically significant differences in precipitation between “urban” and “desert” simulation are evaluated by a two-tailed t-test for difference of means, with statistical significance at the 90% level and above.

The ensemble mean of the “desert” and “urban” experiment precipitation (mm day$^{-1}$) is shown in Fig. 14. Note that during the defined severe weather events in Phoenix, there is precipitation coverage over most of Arizona, including the more climatologically drier southwestern portion of the state where organized, propagating convection accounts for a greater proportion of monsoon precipitation. This spatial pattern is somewhat different than the pattern of seasonally averaged monsoon precipitation, which is maximized over the highest terrain with comparatively much less precipitation in the low desert areas southwest Arizona. Therefore we may infer that the
severe weather events defined for Phoenix are reflective of monsoon “burst” periods where organized, propagating convection occurs in Arizona, principally in the form of mesoscale convective systems that traverse from east to west across the state in a generally easterly upper-level steering flow.

The differences in precipitation between “urban” and “desert” simulations within the Phoenix severe weather event set are quite noticeable, and basically match what has been found in the previously described test case of 13-14 August 2003 (Fig. 14). Note that Fig. 14 shows the results both at the native model grid spacing and with a nine-point smoother. Differences in precipitation are significantly positive in a horseshoe pattern around the Phoenix urban core, with these positive amounts generally to the north, east, and south of the city (on the order of 1-3 mm day$^{-1}$). The precipitation difference within the urban core of Phoenix is significantly negative, especially when considering the smoothed precipitation. Area averaged over the city precipitation produced by “urban” simulations is 1.81 mm less than, and about a quarter of average precipitation in the “desert” simulations. Precipitation simulated by “urban” is decreased about a quarter to half of its standard deviation over Phoenix and is increased at the same relative magnitude around the urban core (Fig. 14). Interestingly, there are also precipitation decreases of similar magnitude, at least in the model smoothed results, over the Tucson area as well.

Are these simulated precipitation differences for the severe weather set due to a consistently modified thermodynamic environment over the city? Figs. 15 and 16 show the average of sensible and latent heat fluxes at 18 UTC or 11 AM local time, respectively. The sensible heat flux of the “desert” simulation is approximately 60-90 W
m$^2$ higher than the “urban” simulations within the Phoenix and Tucson urban areas. The two cities basically appear as clear “bullseyes” in the difference plot, with very distinct behavior from the surrounding desert landscape. The latent heat flux is correspondingly higher in the “urban” simulation and these differences are locally maximized where the largest sources of moisture locally are, in the agricultural irrigated areas on the western and southern edges of the city. Within these locations, the difference in latent heat flux is on the order of 30-45 W m$^{-2}$, or an equivalent evapotranspiration rate of 1-1.5 mm day$^{-1}$.

The average PBL height of the “urban” simulations is lower than the “desert” simulations as a result of these changes in heat fluxes. The “urban” simulation has a shallower PBL height, on average, over the urban areas and the difference is as much as 400 m in northwest Phoenix or Tucson (Fig. 17). The atmospheric instability, measured through MUCAPE, is thus also decreased in Phoenix for “desert” simulations (Fig. 18). The decrease can be as large as 50-100 J kg$^{-1}$, or on the order of about 10% for severe weather event days that exceed the NWS CAPE threshold.

Finally, Fig. 19 shows the difference of mean sea level pressure. The “urban” simulation produces a relatively higher surface pressure in Phoenix (+0.1 to 0.2 mb). Consistent with the idea Balling and Brazel (1987), propagating thunderstorms are deflected around Phoenix because of a positive surface pressure anomaly, leading to local surface divergence. The propagation of monsoon thunderstorms via a cold pool mechanism is therefore much more effective in the desert experiment, without the existence of the isobaric blocking pattern. In summary, the consideration of a set of modeled severe weather events strongly supports the “heat sink” idea of Diem and Brown (2003), with the simulated monsoon thunderstorms having more of a tendency to bypass
the Phoenix urban area, during the objectively identified days when we know that convection would be most favored to be the strongest and most well organized.

5. Discussion and Conclusions

The 13-14 August 2003 regional model case study in Phoenix is designed to evaluate how urbanization affects a monsoon thunderstorms during a specific severe weather event. A new urban canopy model (UCM) is used within WRF that more explicitly accounts for the urban surface in terms of roughness, agricultural and landscape irrigation, increased heat storage and anthropogenic heating. Two 30-hr simulations convective-permitting are performed: the “urban” simulation uses the UCM with optimized anthropogenic parameters; the “desert” run removes the city and replaces it with natural desert land surface cover. We examined changes in the simulations relative to accumulated precipitation, hourly precipitation evolution, surface sensible and latent heat fluxes, PBL height, and the vertical cross section of vertical motion in the direction of storm propagation. Our basic conclusion from this model sensitivity experiment is that the Phoenix urban area seems to act as “heat sink” as suggested by Diem and Brown (2003), behaving in an opposite way to a cities of similar size in the central and eastern U.S. with a documented “heat island.” Greater portioning of surface heat fluxes to latent heating due to the presence of more vegetation and irrigated areas in the urban core, as compared to the surrounding desert, lead to a shallower PBL, higher surface pressure, and a more stable atmosphere. These factors suppress convective precipitation, causing the monsoon thunderstorms to bypass and divert around the city, increasing the precipitation in the areas surrounding the urban core in a horseshoe pattern to the north, east, and
south. Immediately downwind of the city to the southwest in the direction of the upper-level steering flow, precipitation decreases.

We repeated these experiments on a series of 25 of the most severe weather event cases for the Phoenix area, within the context of most thermodynamically favorable days identified in a long-term dynamically downscaled atmospheric reanalysis for the period 1991-2010. The composite results of these experiments essentially confirm the same behavior observed for the 13-14 August 2003 case also applies, in climatological sense, to all the days which would be most conducive for severe monsoon weather in Arizona.

The average difference pattern of precipitation for the “urban” minus “desert” simulations comports with the findings of Shepherd (2006) that showed statistically significant changes in observed monsoon precipitation in association with the growth of Phoenix during the twentieth century. Also interesting, and surprising, is that these same type of effects seem to exist for the Tucson area as well, even though it is a city of smaller size and closer to mountain ranges which are the initiation points for monsoon convection.

What are some reasons our present findings might be different from the Georgescu et al. (2008; 2009a,b) study? They also use convective-permitting modeling approach and adjusted the land surface tiling in their model sensitivity experiments in a somewhat similar way, but found the pattern of precipitation increases around the city only in select cases for identified “dry” monsoon years but not the “wet” years. One possibility is that we use the UCM in the urban tiled model grid cells, which has its own unique model treatment of an urban area that is very different from NOAH. We are not simply swapping the land use tiling within the same land surface model. But we actually
think the framing of the methodological approach is the key difference. While they considered model simulated changes in precipitation in the context of seasonal regional climate model-type simulations of “wet” and “dry” monsoon years, we only are considering the changes associated with actual severe weather events on the days where it is clear the Phoenix area is experiencing precipitation due to organized, propagating convection. If Phoenix is acting as a “heat sink” as our results suggest, then this effect would be most apparent only during the days where the propagating convection, originating from the surrounding mountains, would be favored to pass directly through the city. In other words, the Phoenix “heat sink” only would suppress the convection that originates from elsewhere and passes through the city during only a very few number of days during the monsoon. It does not generate air-mass type thunderstorms as a city with a “heat island” does. We note that we did perform these model sensitivity experiments on all of the 255 thermodynamically favorable severe weather event days, which would include the days that Phoenix did not experience any precipitation (not shown here). Similar to the previous Georgescu et al. results, we did not find any statistically significant differences in the “urban” minus “desert” simulations in that case.

One unanswered question in the present study is which particular aspect(s) of the UCM physics is most critical to explaining the model simulated differences in precipitation. To answer this question would require additional model experiments that investigate the sensitivity to the individual UCM physical components and their possible interactions, using a factor separation approach similar to the study of Rozoff et al. (2003) for the St. Louis area. This will be the subject of a future companion paper that complements our present study.
Acknowledgements

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References


Tables and figures

**Table 1**: List of most severe events selected and their simulated rainfall, CAPE, and PW in Phoenix, AZ

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Figure 1: Upper level observation of geopotential height, temperature, and wind at 500mb (top) and thermodynamic diagram observed at station 74626 (Phoenix Salt river project) (bottom) at 12Z on 13 August 2003.
Figure 2: Position of domain 2 and 3 with 10 km and 2.5 km resolution, respectively. Domain 2 uses forcing data from a downscaled Reanalysis as described in Chang et al. (2015) and domain 3 is a convective-permitting nested domain within domain 2.
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Figure 6: Accumulated daily precipitation over Arizona (mm day$^{-1}$) from 12Z 13 to 12Z 14 August 2003. Upper-left: “urban” run, upper-right: “desert” run, lower-left: difference between the “urban” and “desert” runs, and lower-right: Stage-IV observation. Phoenix and Tucson are marked in bold black contours.
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Figure 9: Hourly precipitation (mm day$^{-1}$) over Arizona of the second storm from 1AM 14 August 2003 to 4AM 14 August 2003, local time. Left column: “urban” run, right column: “desert” run. Phoenix is marked in bold black contour.
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Figure 12: Terrain map of Arizona displaying the vertical cross section axis for vertical motion cold-pool propagation analysis. The red arrow indicates the axis for the cross section while the star shows the location of Phoenix, AZ.
Figure 13: Vertical cross section contouring w-wind (m s\(^{-1}\)) at 12 AM, 1 AM, 2 AM, and 3 AM local time. Phoenix area is indicated by black box and overlying star symbol. Left column is “urban” and right column is “desert” simulation.
Comparison within the most severe weather events. Top-Left and Top-Right panels are the “desert” and “urban ensemble precipitation, respectively. Middle-left panel is difference of two experiment (Urban – Desert) and middle-right panel is a smoothed pattern of the difference. Bottom panels are differences divided by it standard deviation.
Figure 15: Ensemble mean of difference of sensible heat flux (W m$^{-2}$) at 18Z included for severe weather event cases
Figure 16: Ensemble mean of difference of latent heat flux (W m$^{-2}$) at 18Z included for severe weather event cases.
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APPENDIX C: THE MORE EXTREME NATURE OF MONSOON PRECIPITATION IN THE SOUTHWEST U.S. AS REVEALED BY A LONG-TERM CLIMATOLOGY OF SIMULATED SEVERE WEATHER EVENTS

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Abstract

The North American monsoon is the period of convective precipitation in the Southwest and when increases in precipitation intensity are most likely to occur in the Southwest U.S. Long-term changes in precipitation intensity in the Southwest U.S. are evaluated through the use of convective-permitting model simulations of objectively identified severe weather events during “historical past” (1950-1970) and “present day” (1991-2010) periods. Severe weather events are days when the highest atmospheric instability and moisture occur within a long-term regional climate simulation. The event day simulations are performed with convective-permitting (2 km) grid spacing. Results from the severe weather event simulations are compared to available observed precipitation data to evaluate the model physical performance and verify any statistically significant model simulated trends in precipitation. Statistical evaluation of precipitation extremes is performed using peaks-over-threshold (POT) approach with a generalized Pareto distribution. An observed, statistically significant long-term increase in atmospheric moisture and instability causes an increase in extreme monsoon precipitation in the severe weather event simulations. Precipitation is becoming more intense within the context of the diurnal cycle of convection. Largest modeled increases in extreme event precipitation occur is located in central and southwest Arizona, where mesoscale convective systems account for a majority of monsoon precipitation during the evening hours. Therefore, we conclude that a more favorable thermodynamic environment in the Southwest U.S. is facilitating stronger organized monsoon convection during the last twenty years.
1. Background and Motivation

Arid to semi-arid regions located in subtropical zones are projected to experience some of the most adverse impacts of climate change. There is likely to be an increase in heat and aridity due to the retreat of the mid-latitude jet and expansion of sub-tropical highs (e.g. Lu et al., 2009; Archer and Caldeira, 2008; Seidel et al., 2008), for example as concluded in the recent Climate Change Assessment for the Southwest (Garfin et al., 2013) and Intergovernmental Panel on Climate Change (IPCC) Fifth Assessment Report (IPCC, 2013). Another conclusion within these two climate assessment reports is that there will be an increase in precipitation intensity and more extreme weather, due to the exponential increase in atmospheric water vapor holding capacity of the atmosphere with a higher mean temperature (e.g. Meehl et al., 2000). Data from recent observational record supports this is already happening. A number of analyses have documented significant positive trends in atmospheric water vapor globally and within the United States, in terms of both surface specific humidity and column-integrated precipitable water (e.g. Karl and Knight 1998; Groisman et al. 2005; Willett et al. 2007; Santer et al. 2007). Karl and Trenberth (2003) have empirically demonstrated that for the same annual or seasonal precipitation totals, warmer climates tend to generate more extreme precipitation events than cooler climates, implying that increases in extreme weather worldwide reflect more intense convective precipitation. Min et al. (2009, 2011) explicitly linked the observed increases in extreme precipitation globally during the past several decades to anthropogenic greenhouse gas increases.

Our geographic area of interest in this study with respect to changes in precipitation extremes is the Southwest United States. Observed 20-yr return period
thresholds of daily maximum precipitation in the Southwest have exhibited an upward trend (Kunkel et al., 2013), matching what has been documented on a global scale. Any long-term increases in precipitation intensity should be most apparent during the North American monsoon (NAM) in late summer (Adams and Comrie, 1997), as this is the period of warm season severe weather caused by convective thunderstorms. There is some observational evidence to suggest that the warm season climate during the NAM is becoming more extreme. Extreme dry periods within the NAM region have increased 21% from 1910 to 2010, while extreme wet periods have increased 18% over the same period (Petrie et al., 2014). In an analysis of direct precipitation gauge measurements within the Southwest, Anderson et al. (2010) showed increases in monsoon precipitation and precipitation event frequency in Utah and Colorado, on the northern periphery of the monsoon. More recently, our own study (Chang et al., 2015) has revealed increases in monsoon observed precipitation intensity in the Southwest within the context of the Climate Prediction Center daily gridded precipitation product during the period 1950-2010, but especially over the most mountainous areas.

We must necessarily depend on global and regional atmospheric models to generate future projections and retrospective simulations of the NAM for impacts assessment, but the use of these modeling tools for this purpose comes with some important caveats. Global climate models (GCMs) and global atmospheric reanalyses are generally challenged to represent the NAM as a salient climatological feature, in terms of its seasonal maxima in precipitation that occurs during July and August. This problem has been previously noted in reference to both atmospheric reanalyses and global seasonal forecast models (Castro et al. 2007, 2012) and the Coupled Model
Intercomparison Project, Version 3 (CMIP 3) climate change projection models (Bukovsky et al., 2014). Albeit there has been some notable improvement in the mean climatological representation of the NAM in more recent CMIP 5 models (Sheffield et al. 2013), a substantial number of them still erroneously delay the onset of the monsoon until August or September and are too wet during the early fall months (Geil et al., 2015). Dynamical downscaling of atmospheric reanalyses and global climate models to a grid spacing at the meso-β scale (10s of kilometers), for example as done in North American Regional Climate Change Assessment Program (NARCCAP), can improve a modeled climatology of NAM precipitation on a regional scale (Bukovsky et al. 2013; 2015). But regional models at the meso-β scale, even when “perfect” boundary forcing of an atmospheric reanalysis is applied, tend to overestimate monsoon precipitation in mountainous regions and underestimate precipitation associated with organized, propagating convection (Castro et al. 2012; Bukovsky et al. 2013).

Current future projections of the NAM, based on global and regional climate models as just described, have largely focused on question of the mean change in NAM precipitation within multi-model ensembles of climate change projection models. A recent analysis of all the CMIP 5 global climate models suggest there will be a delay in monsoon onset in early summer due to increased atmospheric stability and an increase in monsoon precipitation in late summer once the stability barrier is overcome (Cook and Seager, 2013). Torres-Alavez et al. (2014) showed in the context of the CMIP 5 models that the land-sea thermal contrast will increase and suggests this as a physical cause for weakening of the retreat of monsoon and increasing monsoon precipitation in early fall. The NARCCAP regional climate models (RCMs), by contrast, exhibit a low level of
statistical agreement in their future projections of NAM precipitation, with a minority of models (2) projecting a wetter monsoon and a majority (8) drier one (Bukovsky et al., 2014). There is a wide range of variability in the NARCCAP RCMs precipitation solutions even with the same boundary forcing is applied, illustrating the high sensitivity to the physics (or parameterized processes) of the particular regional models and use of an interior (spectral) nudging option.

When considering the question of changes in extreme precipitation during the NAM, global and regional climate models with a grid spacing on meso-β scale or coarser are inadequate to explicitly represent monsoon thunderstorms. The statistical representation of precipitation extremes may change with increasing spatial resolution in a regional model. Tripathi and Dominguez (2013) found that increasing the horizontal resolution from 50 to 10 km improves the ability of the model to capture individual extreme (above 90\textsuperscript{th} percentile) summer storm events of 24 h and 3 h duration, simulating the NAM within Arizona and New Mexico. However, there is still consistent overestimation of precipitation at both resolutions. Regional atmospheric model simulations at the meso-γ scale (on the order of a kilometer grid spacing) are referred to convective-permitting, because thunderstorms may be explicitly represented without the use of convective parameterization. In the context of numerical weather prediction-type simulations, we and others have previously established that a convective-permitting grid spacing is required to reasonably represent mesoscale convective systems during the NAM, one of the most common meteorological triggers for severe weather events (Cassell et al., 2015).
The present study evaluates historical changes in the intensity of NAM precipitation within the Southwest U.S., in the context of convective-permitting simulations of severe weather events during the period 1950-2010. We select the specific severe weather event days from a long-term dynamically downscaled reanalysis with a similar spatial resolution as the NARCCAP models, based on thermodynamic conditions of atmospheric instability and precipitable water. If long-term increases in atmospheric moisture are present in the Southwest during the NAM and appropriately represented in the dynamically downscaled reanalysis, then we hypothesize that there should be increases in model-simulated precipitation intensity in the severe weather event simulations over the sixty year period, consistent with the observational analysis of CPC precipitation data in Chang et al. (2015). Our objective is to create a historical database of severe weather event simulations during the NAM that can be used for 1) climate impacts assessment purposes, specifically at U.S. military installations within the Southwest, and 2) a future comparison with similarly simulated severe weather events using dynamically downscaled CMIP3 and CMIP5 global climate models. More broadly, we hope to illustrate the high value added of convective-permitting modeling for representing changes in warm season precipitation extremes. Where possible, we try to compare our model simulated extreme precipitation trends to the long-term, gauge-derived CPC observed daily precipitation product.

2. Severe weather event selection and NWP-type modeling

a. Severe weather event selection

The severe weather event days for numerical weather prediction-type simulations at convective-permitting grid spacing are selected from an existing long-term regional
climate simulation described in Chang et al. (2015). This RCM simulation dynamically
downscales NCEP 1 global analysis data during the complete historical period 1948-2010
with the Weather Research and Forecasting (WRF) model at 35 km grid spacing for a
domain that spans the contiguous U.S. and Mexico. Specific severe weather event days
are chosen based on favorable thermodynamic characteristics for monsoon convection in
the Southwest. The thermodynamic variables used are daily maximum values of
column-integrated precipitable water (PW) and most unstable convective available
potential energy (henceforth MUCAPE). When both modeled PW and MUCAPE are in
the top 20% of the distribution during the period June-September (i.e. 122 days per year
during monsoon period) within the Southwest, the event is flagged as a severe weather
event day. Identified severe weather event days in the downscaled reanalysis correspond
well with National Weather Service observed storm reports, with a hit rate of
approximately 70% (Jares et al., 2014). A convective day is defined as starting at 12
UTC to ensure uniformity with radiosonde observational data. The locations of seven
operational radiosonde stations throughout the Southwest have been used to objectively
define the PW and MUCAPE characteristics for the entire region with this downscaled
reanalysis, described in Jares et al. (2014). As shown in Fig. 1, these station locations are:
Tucson, Arizona (TUS), Phoenix, Arizona (PHX), Flagstaff, Arizona (FLG), San Diego,
California (SAN), Las Vegas, Nevada (LAS), Albuquerque, New Mexico (ABQ), and El
Paso, Texas (ELP).

PW is computed from six-hourly water vapor mixing ratio \( (q_v) \) measurements
present at the nineteen height intervals within the WRF model as:
\[ PW = \int_{z_0}^{z} q_v \rho dz = -\frac{1}{g} \int_{z_0}^{z} q_v dp \]  
(7)

where \( z_0 \) is the lowest modeling layer, \( z \) highest modeling layer and \( g \) is the gravitational constant. Atmospheric instability, in the form of MUCAPE, is calculated as:

\[ MUCAPE = \int_{z_{LFC}}^{z_{EL}} \frac{T_{v,p} - T_{v,e}}{g} dz \]  
(8)

in which \( z_{EL} \) is the height of the equilibrium level, \( z_{LFC} \) is the height of the level of free convection, \( g \) is the gravitational constant, \( T_{v,p} \) is the virtual temperature profile of an ascending air parcel, \( T_{v,e} \) is the virtual temperature profile of the environment. Craven et al. (2002) notes for elevated deep convection, it is appropriate to consider MUCAPE for a rising air parcel originating from the lowest 300 hPa above ground level. Therefore, MUCAPE is used as the instability metric, rather than surface or mixed layer CAPE.

A time series of PW and MUCAPE is extracted from downscaled reanalysis by means of area-averaging over the Southwest, defined as the convective-permitting (d3 domain) in Fig. 2. In previous work, Mazon et al. (2014) and Jares et al. (2014) respectively evaluated severe weather event days in the Southwest identified within radiosonde data and the dynamically downscaled reanalysis. In both cases, the identified severe weather event days by PW and MUCAPE criteria basically reflect the known atmospheric circulation patterns associated with severe weather in the Southwest found by Maddox et al. (1995) related to the specific geographic positioning of the upper-level 500-mb high (or monsoon ridge) in relation to Arizona. The downscaled reanalysis also shows long-term increases in CAPE and PW (Jares et al., 2014) and these are consistent with evaluation of observed radiosonde data, as described in Lahmers et al. (2015) to be mentioned later.
b. Regional atmospheric model simulations of severe weather events

Thermodynamically favorable severe weather event days identified in the WRF dynamically downscaled reanalysis during a retrospective “present day” period (1991-2010) are first simulated with a convective-permitting grid spacing of 2.5 kilometers. These results from these simulations are compared with hourly observed precipitation (derived from gauge and radar data) and daily radiosonde data, to verify that the atmospheric model can reasonably represent the diurnal cycle of precipitation and its relationship to convective organization and propagation. Severe weather event days within a retrospective “historical past” period (1951-1970) are then simulated. Differences in the behavior of precipitation between the “present day” and “historical past” periods can help reveal the impact of long-term observed changes in atmospheric moisture and instability on severe monsoon weather. The total number of simulated days in the “present day” and “historical past” periods is 255 and 268, respectively.

The WRF model experimental design for simulation of severe weather events uses a two-domain nesting strategy, with an intermediate domain of 10 km grid spacing, and a convective-permitting, meso-γ domain of 2.5 km grid spacing. The intermediate domain utilizes data from the aforementioned long-term dynamically downscaled NCEP 1 reanalysis as boundary forcing and covers the southwest United States and northwest Mexico. The 2.5 km convective-permitting domain is centered over central and southern Arizona and extends over entirety of Arizona and New Mexico, including portions of California, Colorado, Nevada, Texas, and Utah (Fig.2). WRF model parameterization options on the convective-permitting domain are nominally similar what is used for generating real-time WRF quasi-operational monsoon forecasts at the
University of Arizona, Department of Atmospheric Sciences (UA-ATMO), but with two notable differences with respect to land surface modeling and convective parameterization. Common parameterization options include: a bulk microphysics scheme (Thompson et al., 2004); Mellor-Yamada-Janic planetary boundary layer scheme (Janjic 1990, 1996, 2002) with Eta surface layer (Janjic 1996, 2002); the NOAH-MP land surface model (Niu et al., 2011); Dudhia Shortwave radiation (Dudhia 1989); and the Rapid Radiative Transfer Model (RRTM) Longwave radiation (Mlawer et al. 1997). On the intermediate domain, the Kain-Fritsch cumulus parameterization scheme is applied (Kain and Fritsch 1993, 2004), with the modified convective trigger and CAPE closure assumption of Truong et al. (2009) that better accounts for dynamic pressure effects in complex terrain. An Urban Canopy Model (UCM), with optimized anthropogenic parameters for the southwestern United States as adapted from Grossman-Clarke (2010), is applied at grid points with a defined urban land use classification, including the cities of Tucson, Phoenix, Albuquerque and Las Vegas. For a given identified severe weather event, that satisfies the thermodynamic threshold criteria as previously described, the convective-permitting WRF simulation is performed as follows in a numerical weather prediction-type mode. The event simulation is initialized at 6 UTC (11 pm MST) the day prior to the event and the simulation is executed for thirty-hours, ending at 12 UTC (5 am MST) the day following the event. The initialization in the evening prior allows for 6 hours of model spin-up time, consistent with UA-ATMO operational forecast practices. Model output from the convective-permitting grid is saved hourly, as this temporal resolution resolves well the diurnal cycle of convection.

c. Observational verification data for precipitation
For verification, the quality controlled Stage IV combined NEXRAD radar-gauge precipitation product (http://data.eol.ucar.edu/codiac/dss/id=21.093) is used to compare observed precipitation to WRF-modeled precipitation on the convective-permitting domain. The specific dataset we use is the Hourly Precipitation Data (HPD), digital data set DSI-3240 archived at the National Climatic Data Center (NCDC). Though we consider the Stage IV data as observational truth in the context of this study, it is important to note that these data still have problems with respect to estimating precipitation in complex terrain, due to issues of lack of rain gauge observations and radar beam blockage (e.g. Adams et al., 2014). Hourly Coop station data is used to verify modeled precipitation diurnal cycle in some specific location (http://www.ncdc.noaa.gov/cdo-web/search?datasetid=PRECIP_HLY). For the historical climate trend maps, we use CPC quarter degree daily US unified gauge-based analysis of precipitation data (Higgins et al., 1996).

3. Statistical analysis methods of precipitation distributions and trends

The representation of the likelihood of receiving a specific rainfall amount is best accomplished by fitting a statistical distribution (e.g. gamma distribution). This parameterized distribution is a continuous function when fitted for every local grid points on the map will show no discontinuity. These distributions make it possible to estimate the likelihood of rainfall being within a specified range. The gamma distribution for example is a good tool to fit a precipitation curve overall. However, extreme climate values in the tail of the distribution may not necessarily fit well to a theoretical PDF that applies to the entire data set. The solution is to fit generalized Pareto distribution, a peak-over-threshold method, to better describe the behavior in the tail (e.g. Katz, 2010;
Dominguez et al., 2012). This methodology is useful when dealing with a limited number of time slices (on the order of 250) are available. It allows for a larger sample size than the generalized extreme value (GEV) approach.

A Poisson distribution is used to characterize the extreme precipitation rate at which the threshold is exceeded and a Generalized Pareto (GP) distribution is used to characterize the amount at which the threshold is exceeded (termed “Poisson–GP model”). For the distribution of excesses above a high threshold, the cumulative distribution and quantile functions for the GP are given by:

\[ F(x; \sigma^*, \gamma) = 1 - \left[ 1 + \gamma \left( \frac{x}{\sigma^*} \right)^1 \right]^{-\frac{1}{\gamma}}, \quad \sigma^* > 0; \quad 1 + \gamma \left( \frac{x}{\sigma^*} \right) > 0 \quad (9) \]

\[ F^{-1}(1 - p; \sigma^*, \gamma) = \frac{\sigma^*}{\gamma} (p^{-\gamma} - 1), \quad 0 < p < 1 \quad (10) \]

Here \( \sigma^* \) and \( \gamma \) are the scale and shape parameters, respectively. We set the threshold for the POT distribution to the 98th percentile. The tails of the distribution with an intensification of events above this threshold is what is of interest.

Once the parameters are estimated, their accuracy in approximating the true rainfall distribution is evaluated. This study tests the goodness-of-fit using a \( \chi^2 \) test that compares the histogram and the discrete density function. In this statistical test, the null hypothesis is that the data are consistent with the specified distribution. The test statistic random variable \( \chi^2 \) is defined by the following equation.

\[ \chi^2 = \sum_{\text{classes}} \frac{(O_i - E_i)^2}{E_i} \quad (11) \]

where \( O_i \) is the observed frequency count for the \( i^{\text{th}} \) level of the categorical variable, and \( E_i \) is the expected frequency count for the \( i^{\text{th}} \) level of the categorical variable.
The local statistical significance (for each grid box) of the estimated daily precipitation above the 98th percentile is determined by a bootstrap resampling procedure using a total of 1000 samples. The future changes in the return values are considered statistically significant at the 0.01 (99th percentile) level if their 90% confidence intervals do not overlap (Kharin and Zwiers, 2005).

4. Long-term changes in atmospheric thermodynamic conditions in the Southwest

Favorable thermodynamic conditions are primary requirement for development of any monsoon thunderstorms in the NAM region. Low-level moisture is available during the monsoon season. Moist, rising air occurs during the day over mountain ranges, due to the differential heating of the mountains relative to the surrounding air. In a conditionally unstable atmosphere, where CAPE exceeds convective inhibition (CIN), water vapor in rising air may condense to form cumuliform clouds. Cumulonimbus clouds will form in a conditionally unstable atmosphere and will extend through the entire depth of the troposphere when monsoon thunderstorms fully mature. Thunderstorms begin to develop over mountain ranges in late morning to early afternoon and produce precipitation by late afternoon. The formation of monsoon thunderstorms thus requires sufficient instability and moisture.

Atmospheric instability and moisture during the monsoon have substantially changed over the past thirty years in the context of the downscaled reanalysis, as have reported earlier in Jares et al. (2014). The mean difference in MUCAPE (Fig. 3, left) shows an overall increase in atmospheric instability over the entire Southwest U.S., excluding the Gulf of California. This increase of MUCAPE is maximized over northern Arizona and the southern parts of Nevada and Utah. The corresponding results for PW
(Fig. 3, right) show that PW has increased throughout the entire Southwest as well. The increase is maximized at the northern end of the Gulf of California and extends northward to the area where increases in MUCAPE are observed. The largest increase in PW occurs over central Arizona where Phoenix is located. The increase in atmospheric moisture and instability over the monsoon region during the past 60 years in the downscaled reanalysis is consistent with observations from radiosonde sounding data in the region (Table 1), as reported by Lahmers et al. (2015). Therefore, we would assert that these modeled MUCAPE and PW changes are likely not artifacts of the global reanalysis that are appearing due to changes in instrument observing systems, principally the introduction of satellite data into the data assimilation after the late 1970s. In the next sections we will test our hypothesis that these observed increases in atmospheric moisture and instability are more conducive to heavier convective precipitation during the modeled severe weather event days.

5. **Severe weather simulations during period of Stage IV product**

To verify the performance of the WRF severe weather event simulations on the convective-permitting grid prior to any evaluation of long-term trends in precipitation intensity, we compare model-simulated precipitation to the Stage IV combined radar-gauge observed precipitation product during the nine-year period 2002-2010. We first consider precipitation climatology of the severe weather event days, and then the modeled treatment of the diurnal cycle of precipitation from the hourly data.

a. **WRF simulated daily precipitation vs. Stage IV during the period 2002-2010**

The model simulated daily average precipitation (over 24 hours of the convective day) and corresponding Stage IV precipitation for all thermodynamically favorable
severe weather event days is shown in Figure 4. The Stage IV data shows that the thermodynamically identified severe weather event days have widespread precipitation over all of the Southwest, particularly Arizona and New Mexico. The average observed daily precipitation is a maximum (greater than 5 mm) over the highest elevations because the mountains are the focal point for convective initiation. For example, the highest average observed precipitation in Arizona occurs over the Mogollon Rim, which roughly bisects the state from the southeast to the northwest corners. Generally, the convective-permitting severe weather event simulations exhibit similar behavior in terms of capturing the terrain-dependence of monsoon precipitation and show precipitation occurring throughout the Southwest. There is an underestimation of precipitation in the model simulations on the order of 1-2 mm day$^{-1}$.

b. Diurnal cycle of convection

The convective-permitting simulations are able to reasonably simulate the diurnal cycle of convection, with a maximum in precipitation centered over areas of high terrain during the afternoon (11am to 5pm local time), westward propagation of the precipitation off the terrain of the Mogollon Rim during early evening (5pm to 11pm local time), and weakening and dissipation of precipitation during the late evening to early morning (11pm to 5am local time). The propagation of monsoon precipitation during the late afternoon to early evening is demonstrated in Figure 5 in terms of the precipitation rate averaged over the six-hour time period. The precipitation at later times in the day reflects the presence of organized convection, principally mesoscale convective systems, that originate as air-mass type thunderstorms on the Mogollon Rim and tend to organize and propagate westward across the western part of Arizona toward the Colorado River Valley.
during the late afternoon and evening hours, as we have previously discussed in Cassell et al. (2015) in reference to simulations during Intensive Observing Period 2 of the North American Monsoon Experiment.

6. Long-term changes in monsoon precipitation in the Southwest

Long-term changes in monsoon precipitation are evaluated in this section from the perspective of both observations and results of the severe weather event simulations. Observations in this case are station NCDC coop precipitation data and the daily gridded CPC precipitation product. In the presentation of these results, we attempt to distinguish between the mean changes in precipitation versus the changes in precipitation extremes in the tails of the distribution. For evaluation of the latter, we apply the extreme value statistical analysis peak over threshold technique described earlier.

a. Broadening and flattening of the distribution of daily monsoon precipitation

The histogram of NCDC coop station precipitation data shows that Phoenix has experienced an increase in precipitation extremes during the “present” period 1991-2010 as compared to the “historical” period 1951-1970 (Fig. 6, left). The red and blue lines on the figure are the peak-over-threshold generalized Pareto distributions fitted to the right tail of the PDF for precipitation events above the 98th percentile. The fitted distributions satisfy the Chi square goodness-of-fit test at significance level $\alpha = 0.1$. The increase in extremes in monsoon precipitation in Phoenix can be interpreted as a broadening and flattening of the daily precipitation distribution. The differences are statistically significant tested with bootstrapping with 1000 samples at significance level $\alpha = 0.01$. The right side of Fig. 6 shows the same analysis per
formed for the model simulated severe weather events. Though modeled precipitation exhibits a dry bias, as discussed earlier with reference to comparison with the Stage IV product, it shows the same type of general behavior as the observed station data.

Does the idea of a broadening and flattening of the daily precipitation distribution also hold over the entire Southwest, in the context of comparing the observed CPC gridded precipitation product with the simulated severe weather events within the convective-permitting domain? The “present” minus “historical” period change in mean daily CPC observed precipitation for all days during the monsoon months of July and August is shown in Fig. 7, with the absolute changes on the right and the percentage changes on the left. From the observational standpoint, mean daily monsoon precipitation has decreased as a whole in Arizona during the “present days” period, as compared to the period of the “historical past”. The largest absolute precipitation decreases occur over the Mogollon Rim (1 mm day$^{-1}$ or greater than 30%). The largest precipitation percentage decreases occur in the Colorado River Valley and over southwest Arizona (40-50%), an area of the state where more infrequent, organized convection accounts for the majority of monsoon precipitation (e.g. Castro et al. 2007).

However, the corresponding pattern of changes in extreme monsoon precipitation is quite different than that of the mean just shown. Characterizing the changes in extreme event precipitation within the CPC daily precipitation data is done with the top 20% heavy precipitation days. Using just this limited subset of days, we characterize the changes in extreme precipitation above the 98th percentile using peaks over threshold analysis technique. With these filters applied, Figure 8 shows that CPC observed
extreme event monsoon precipitation has generally increased over Arizona, by as much as 30 mm day\(^{-1}\) in some locations. It increases more in the eastern and northern portions of AZ at higher elevations. This seems the opposite for the high resolution (2.5 km grid spacing) downscaled data. Extreme event monsoon precipitation above the 98\(^{th}\) percentile from the set of severe weather event simulations at convective-permitting scale similarly exhibits an increase, but especially south and west of the Mogollon Rim at lower elevations where organized convection accounts for more monsoon precipitation. Coarse resolution (35km grid spacing) downscaled Reanalysis, on another hand, shows an increased pattern over the Mogollon Rim and high elevation regions. The wide spread positive pattern, even though more similar to observation, could be erroneous due to the nature of the coarse grid spacing and the use of a convective parameterization. Precipitation from RCMs simulated at this grid spacing has a known positive bias in the mountainous regions (Chang et al., 2015). These analyses of gridded observed and modeled data substantiates the conceptual idea of broadening and flattening of the distribution of daily monsoon precipitation over the past twenty years in Arizona, with decreasing mean daily precipitation but increasing extreme event precipitation intensity.

b. The diurnal cycle of convection in model vs. coop station observations

Atmospheric models are generally challenged to represent the diurnal cycle of convective precipitation during the warm season (e.g. Trenberth et al., 2003). In the Southwest the diurnal cycle of monsoon thunderstorms is intimately linked to mountain valley circulations and propagation of thunderstorms off the terrain when upper-level winds are from a favorable direction (easterly in Arizona). To evaluate how well the convective-permitting simulations perform in representing the diurnal cycle of
precipitation during severe weather event days, we compare them with select NCDC coop observing stations. Only a very limited number of coop stations in the Southwest have sufficient hourly precipitation records since the mid twentieth century to make such a comparison. Not surprisingly they are located in the largest cities in the Southwest U.S. region: Phoenix (PHX), Tucson (TUS), Flagstaff (FLG), and Albuquerque (ABQ). PHX and TUS in these comparisons are classified as the low elevation stations (337 m and 777 m, respectively) located downwind of mountain ranges; FLG and ABQ are classified as high elevation stations (2,139 m and 1,619 m, respectively). As the smallest wavelength an atmospheric model can physically resolve is approximately four times the grid spacing (Pielke and Pearce, 1994), the modeled precipitation is computed as the average of 25 model grid points surrounding all stations with the exception of PHX. Because Phoenix is so large, in this case we simply average all the grid points with urban land use tiling.

The diurnal cycle for the two low elevation stations is shown in Figure 9 and for the two high elevation stations in Figure 10. The horizontal axis in these figures is local time (MST), where a convective day starts at 5am and ends at 5am the next day. Hourly precipitation is accumulated from the previous hour to the time plotted. We add 24 hours to the local time of the next day to make a continuously increasing x-axis (so t=29 is equivalent to 5am of the next day). The vertical axis is plotted in (mm hr$^{-1}$), albeit with different scaling for model and observations for two reasons: 1) we are more interested in whether the model simulation can reproduce the pattern of the diurnal cycle of precipitation, not necessarily the exact magnitude of precipitation at each hour, and 2) hourly observation for the “historical” period are not as reliable as those of the “present
period.” These periods are denoted as blue and red lines, respectively, on the figures with model simulation results as M51-70, M91-10, and observations as O51-70, O-91-10. We are not explicitly assessing statistical significance of the hourly differences between the two time periods here, though we do point out some physically interesting results in the following discussion.

The observed severe weather event day precipitation evolution over Phoenix (Fig. 9) shows that most of the precipitation at this station occurs during the late evening hours (10 pm to 2 am local time). The precipitation maximum at this time reflects the passage of more organized, westward propagating convection through the city at night that develops during the previous day over the mountains, most predominantly the Mogollon Rim to the north and east of the city. During the “present day” period this late evening precipitation maxima diminishes (on the order of 0.10 mm hr⁻¹) and begins to occur about 1 to 2 hours earlier. The equivalent model results for PHX in the severe weather event simulations are not capturing well but similarly show a nocturnal maximum in precipitation (at 10 pm local time) in the “historical past” and shifting this maximum to occur approximately five hours earlier (at 5 pm local time) in the “present day”. Therefore, a more favorable thermodynamic environment in the “present day” seems to facilitate thunderstorms to occur in Phoenix earlier during the evening. The observed maximum in convective precipitation in Tucson occurs earlier in the day than in Phoenix, or in late afternoon to early evening (3 pm to 5 pm local time), due to its closer proximity to mountain ranges surrounding the city. Though coop station observations would indicate a decrease in precipitation intensity at the peak convection time, the modeled precipitation results show little appreciable change between the two periods.
The model representation of the diurnal cycle of precipitation is much better at the higher elevation stations, where precipitation is generally due to locally forced air mass-type monsoon thunderstorms rather than organized convection. The peak intensity of observed precipitation at Flagstaff is during the early to midafternoon (2 pm to 3 pm local time), and at Albuquerque during the late afternoon to early evening (5 pm to 7 pm local time). The timing of the maxima of modeled and observed precipitation is similar. Another note for all the modeled diurnal cycle plots is that there is a precipitation peak during the first 6 hours of the model simulation, likely due to a poor model treatment of any remnant convective precipitation from the previous day.

c. Change in model simulated propagation of convection in conjunction with the diurnal cycle.

Maps of hourly precipitation are plotted over the course of the diurnal cycle of convection (11am until 5am of the next day) to show development and propagation of monsoon precipitation during the severe weather event days (Figs.11). This 18-hour window has been chosen based on the fact that NAM monsoon thunderstorms start to develop around noon. Simulations during both the “historical” and “present day” periods show an exceptional level of detail in the spatial patterns of precipitation and appear to capture the diurnal cycle of monsoon precipitation across the terrain well, when compared to, for example, the analysis of radar data during the North American Monsoon Experiment in Nesbitt et al. (2008). Precipitation maxima centered over the highest terrain in early to mid afternoon are indicative of air-mass type monsoon thunderstorms that develop in a thermodynamically favorable environment. During the evening hours, the precipitation maxima over the mountains diminishes as precipitation moves south and
westward toward the urban centers of Phoenix and Tucson and then toward the Colorado River Valley. It is at this time that organized, propagating convection, in the form of mesoscale convective systems, is most likely to occur in Arizona. Most organized convection dissipates by the early morning hours of the following day.

The differences in simulated precipitation for the most extreme precipitation events (top 10%) between the “present day” and “historical period” are revealed every four hours in Figs. 12. These difference plots suggest that simulated monsoon precipitation is becoming more intense in relation to the time of day it is most favored to occur at a given location. The largest increases in simulated precipitation occur over the mountains in the middle of the day (12 pm to 3 pm), when air mass-type thunderstorms are developing. In contrast, the Phoenix area and the southwest desert areas of Arizona experience the largest increases in precipitation in the late afternoon and early evening hours (4 pm to 6 pm), in association with more intense mesoscale convective systems. The increases in daily extreme event precipitation are maximized in southwest Arizona because the largest increases in instability and water vapor have occurred there (Jares et al., 2014). It appears that the region with the greatest increases in precipitation extremes is also the one with low mean values and heavy precipitation depends on propagation of MCSs in the late afternoon to the evening.

7. Discussion and Conclusions

The overall objective of this study is to evaluate long-term changes in precipitation intensity during the North American monsoon in the Southwest U.S., through the use of convective-permitting model simulations of objectively identified severe weather events during “historical past” (1950-1970) and “present day” (1991-
2010) periods. These severe weather events are the days when the highest atmospheric instability and moisture occur in the Southwest within a long-term regional climate simulation (35 km grid spacing) that dynamically downscales a global atmospheric reanalysis over the contiguous U.S. and Mexico. The event day simulations are performed in a numerical weather prediction-type mode with the WRF model over the region of the Southwest with convective-permitting (2 km) grid spacing, using the long-term regional climate simulation as the boundary forcing. Results from the severe weather event simulations are compared to observed precipitation data from the Stage IV combined radar-gauge product, the CPC daily precipitation product, and NCDC coop station data in urban centers to evaluate the model physical performance and verify any statistically significant model simulated trends in precipitation. Statistical evaluation of precipitation extremes is performed using peaks-over-threshold (POT) approach with a generalized Pareto distribution.

The severe weather event simulations appear to reasonably represent the diurnal cycle of convective precipitation during the period of the Stage IV product, in terms of the development of precipitation over the highest terrain during the day and convective organization and propagation into the evening hours. The comparisons with hourly precipitation data from NCDC coop stations also shows that the model simulations effectively captures the differences in the timing of convective precipitation during the monsoon in relation to elevation. Model simulated severe weather event precipitation tends to be slightly underestimated in comparison to the Stage IV.

There has been an observed, statistically significant long-term increase in atmospheric moisture and instability in the Southwest over the past sixty years per trends
in observed radiosonde data, which is realized in the long-term regional climate simulation. While mean daily monsoon precipitation in the Southwest has decreased from the analysis of the CPC dataset, the most extreme monsoon precipitation has become more intense in these same data during the days with most favorable thermodynamic and dynamic conditions to support organized, propagating convection. We observe a similar increasing intensity of extreme monsoon precipitation in the convective-permitting severe weather event simulations. In the model simulations, precipitation is becoming more intense within the context of the diurnal cycle of convection. Where the largest modeled increases in extreme event precipitation occur is located in central and southwest Arizona, where mesoscale convective systems account for a majority of monsoon precipitation during the evening hours. Therefore, we conclude that a more favorable thermodynamic environment in the Southwest U.S. is facilitating stronger organized monsoon convection during the last twenty years.

We are currently applying the same methodological approach to dynamically downscaled global climate change projection models, as an ensemble analysis of future projection with level of model agreement, similar to what has been done for NARCCAP simulations (Bukovsky et al., 2014). These types of model experiments will hopefully more explicitly attribute the increases in monsoon precipitation intensity, documented in a historical sense here, to anthropogenic global climate change.

Acknowledgements

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References


Petrie, M., Collins, S., Gutzler, D., Moore, D., 2014: Regional trends and local variability in monsoon precipitation in the northern Chihuahuan Desert, USA. Journal of Arid Environments, 63–70


**Tables and figures**

**Table 1**: Correlation Coefficients and t-statistics for annual average CAPE and Precipitable Water (15 June – 15 September) long-term trends. Bold font denotes statistical significance. Adapted from Lahmers et al., 2015.

<table>
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Figure 1: Geographic map of the analysis region with highlighted operational radiosonde stations. Names of stations have been abbreviated: San Diego, California (SAN), Las Vegas, Nevada (LAS), Flagstaff, Arizona (FLG), Phoenix, Arizona (PHX), Tucson, Arizona (TUS), Albuquerque, New Mexico (ABQ) and El Paso, Texas (ELP).
Figure 2: Position of domain 2 and 3 with 10 km and 2.5 km resolution, respectively. Domain 2 uses forcing data from a downscaled Reanalysis as described in Chang et al. (2015) and domain 3 is a convective-permitting nested domain within domain 2.
Figure 3: Results of the difference of means test for most unstable convective available potential energy (MUCAPE) in J kg\(^{-1}\) (left) and column-integrated precipitable water (PW) in mm (right). The means of daily maximum MUCAPE and PW gridded data of the late period (1980-2010) were subtracted from the mean of the early period (1950-1980). Adapted from Jares et al. (2014)
Figure 4: Composite daily mean of precipitation [mm day$^{-1}$] of all selected severe weather event days during 2002-2010 for model (left) and Stage IV observation (right)
Figure 5: Peak-hour composite 6-hourly mean of precipitation [mm hr$^{-1}$] of all selected severe weather event days during 2002-2010 for model (left), and Stage IV observation (right) plotting from 11am to 5pm (top), and from 5pm to 11pm (bottom).
Figure 6: Station comparison of probability distribution of daily precipitation extremes for Phoenix (1991-2010 red versus 1951-1970 blue). Red and blue bars are station histogram. Red and blue lines are peak-over-threshold General Pareto distributions fitted into the right tail of the PDF with events on the right of 98 percentile.
Figure 7: Mean precipitation climatology trends (1991-2010 minus 1951-1970) for July and August from CPC plotted in an absolute scale [mm day$^{-1}$] (left) and relative scale [%] (right)
Figure 8: Extreme precipitation climatology trends (1991-2010 minus 1951-1970) of intensity of events above the 98th percentile [mm day$^{-1}$]. Left panel is observation from CPC; Middle and right panels are from high resolution and coarse resolution downscaled Reanalysis, respectively. Maps are plotted with grid points that satisfy a bootstrapping statistical significance at the 0.01 level.
Figure 9: Mean diurnal cycle at two low desert stations, Phoenix (top) and Tucson (bottom) for modeled (left) and observed (right) precipitation. Local time is in hour and on MST time zone.
Figure 10: Mean diurnal cycle at two high desert stations, Flagstaff (top) and Albuquerque (bottom) for modeled (left) and observed (right) precipitation. Local time is in hour and on MST time zone.
Figure 11: Hourly mean precipitation [mm hr$^{-1}$] of the top 10% most severe events from 11am to 12pm (top), from 4pm to 5pm (middle), and from 9pm to 10pm (bottom) for historical period (left), and present period (right).
Figure 12: Different in hourly mean precipitation [mm hr⁻¹] of the top 10% most severe events from 12pm to 1pm (top left), from 4pm to 5pm (top right), from 8pm to 9pm (bottom left), and from 12am to 1am (bottom right) for the two periods (present minus historical).