The International H2O Project (IHOP_2002) was a field study conducted over the Southern Great Plains (SGP) during the summer of 2002. The main goal of IHOP_2002 was to study water vapor distribution in the atmosphere in order to improve predictions of warm season rainfall over the SGP. This research is aimed toward studying the effects of interactions between the land and atmosphere on mesoscale model simulations of convection and precipitation. The convection case examined during this study occurred from 17-20 June 2002 during IHOP_2002. A combination of synoptic forcing and surface features contributed to the development and continuance of convection during this time period. An observational analysis of this case showed evidence of surface dryline boundaries, accompanied by temperature and moisture gradients. Gradients of surface sensible and latent heat fluxes across the boundaries were also seen throughout the time period. Convection formed along the boundaries, suggesting that the land-surface features are important in the development of deep moist convection, even for cases with strong synoptic forcing.

Two sets of sensitivity studies were conducted for this case using the MM5 model. The first involved testing the sensitivity of different surface data assimilation schemes on simulations of convection. A Control Simulation was performed using standard MM5 data assimilation, followed by an Experimental Simulation using a Flux-Adjusting Data Assimilation System (FASDAS). Significant differences were seen
between these two simulations, especially in regards to accumulated precipitation. The Control Simulation generated very little precipitation, while the Experimental Simulation produced significantly more precipitation, and overall appeared to match more closely with observations than the Control. The Experimental Simulation also produced more varied heat flux fields, setting up strong gradients which may have led to the enhanced precipitation. The Experimental Simulation also overall matched observed heat fluxes than the Control Simulation. This study suggests that improvements in surface data assimilation may drastically improve predictions of convection and precipitation.

The second study involved testing the effects of using two different coupled land-surface models (LSM) and planetary boundary layer (PBL) schemes on simulations of convection. The same convection case (17-20 June 2002) was used for this study. The first simulation used the Noah LSM coupled to the MRF PBL, and the second used the Pleim-Xiu (PX) LSM coupled to the ACM PBL. The precipitation fields generated by both simulations were very similar to each other; however they differed from the observations in terms of accumulations and timing. Significant differences were seen in the evolution of soil moisture and temperature between the two simulations, which was to be expected due to the different soil layers in each LSM used. There was a more immediate soil-precipitation feedback in the PX run, where the top soil layer was closer to the surface. A similar feedback could be seen in the heat flux fields. Results of this study provide evidence of the importance of the land-surface in mesoscale modeling, while also suggesting that improvements are needed in modeling land-surface processes.
The Effects of Land-Atmosphere Interactions on Convection Initiation and Quantitative Precipitation Forecasts During the International H$_2$O Project (IHOP_2002)

By

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BIOGRAPHY

Aneela Laurel Qureshi was born on September 10th, 1980 to parents Judy and Nazir Qureshi in the quaint little town of Sterling, Illinois. Her interest in weather began early in life, when she was terrified of thunderstorms and used to beg her parents to let her stay up to watch the weather report at night. However, she didn’t consider meteorology as a career option until towards the end of her time at Sterling High School, where she graduated with honors in 1998. After high school, Aneela went on to college at Valparaiso University in Valparaiso, Indiana. During college, she had a wonderful opportunity to participate in an REU program with the Space Physics Research Laboratory at the University of Michigan, working with the Microwave Geophysics Group to study remote sensing of soil moisture. It was this experience that ultimately led to her interest in land-atmosphere interactions. In May of 2002, Aneela became the first person in her family to graduate from college, earning a Bachelor of Science in Meteorology, with minors in Mathematics and Classical Civilizations. After graduating, she opted to take a year off of school before pursuing graduate studies. During this time, she worked with the AmeriCorps Benefits Children Literacy Program in northwestern Illinois, tutoring elementary school students. In the fall of 2003, Aneela started a Master’s program at North Carolina State University, where she studied land-atmosphere interactions. Inspired by her year with the AmeriCorps program, Aneela is currently working toward a Master’s in Education, specializing in Environmental Education. She plans to use this degree in conjunction with her extensive meteorology background to specialize in educating the public about weather and weather safety.
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All of the observational data used for this research was obtained from the IHOP_2002 field catalog and data archive (http://www.joss.ucar.edu/ihop). Access to data was provided by Dr. Steve Williams of UCAR. I would also like to acknowledge all of the participants of the IHOP_2002 field study.

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CHAPTER 1: INTRODUCTION

1.1 Importance of Land-Surface Processes

Despite a growing body of evidence regarding the importance of land-atmosphere interactions, the operational forecasting community has lagged in considering the land-surface feedback in numerical weather models. Interactions between the surface of the Earth and the atmosphere are vital to the development of deep moist convection. Land-surface heterogeneities, including horizontal differences in soil moisture, soil texture and vegetation, have been shown significantly impact the formation of cumulus convection (Anthes 1984; Chen and Avissar 1994; Avissar and Liu 1996; Wetzel et al. 1996; Pielke et al. 2001).

1.2 Overview of the Southern Great Plains

The Southern Great Plains (SGP) region of the United States is often the location of convection studies. There are several reasons for this, including (1) this region is a “hot-spot” for active convection, (2) large gradients of both water vapor and rainfall exist over the SGP, (3) the region is a prime location for the convergence of several air masses, i.e., cool, dry air from Canada, hot dry air from the Mexican Plateau and warm, moist air from the Gulf of Mexico, and (4) horizontally varying fields of soil moisture, vegetation and the slope of the terrain help lead to the formation of surface boundaries. Due to the active convection over the SGP, a vast assortment of operational and experimental instruments can be found in this region, which also makes it a prime location for studying convection (Weckworth et al. 2004).
1.3 Study Objectives

Outstanding issues in the study of how land-atmosphere interactions affect deep moist convection still remain. The International H₂O Project (IHOP_2002) field study was designed to address some of these issues. In particular, the participants of IHOP_2002 sought to improve measurements of water vapor distribution in the boundary layer, along with studying the relationship between this distribution and surface and boundary layer processes. The ultimate goal of the project was to create a better understanding of how these processes impact convective initiation (CI) in order to make improvements in warm-season quantitative precipitation forecasts over the SGP (Weckworth et al. 2004).

Land-surface processes and land-atmosphere interactions have previously examined primarily for cases with synoptically weak conditions, where the main forcing for convection has been caused by surface features. However, studies appear to be lacking for cases with strong synoptic forcing. Under strong synoptic conditions, boundary layer features still may play an important role, via the surface-atmosphere feedback cycle associated with moisture availability, convective initiation, thunderstorm development, precipitation, changes in cloud cover, etc. Outstanding questions remain on how these feedbacks affect the evolution of mesoscale and synoptic scale events.

The objective of the following studies follows along the same goals as the IHOP_2002 field study. The impacts of land-atmosphere interactions on CI and QPF, as well as the general four-dimensional structure of the SGP are examined through in-depth studies of a convection case which occurred during IHOP_2002. The MM5 model is used to conduct sensitivity studies involving (a) the use of two different surface data
assimilation schemes and (b) the use of two different coupled LSM/PBL schemes; both of which aim to examine the aforementioned impacts of land-atmosphere interactions.

The layout of these studies is as follows: A description of both the IHOP_2002 field study and the MM5 model is provided in Chapter 2; an observational analysis of the convection case is presented in Chapter 3; Chapter 4 focuses on the sensitivity of surface data assimilation schemes; the use of different coupled LSM/PBL schemes is in Chapter 5; and finally conclusions are in Chapter 6.
CHAPTER 2: SUMMARY OF THE IHOP_2002 FIELD STUDY AND THE MM5 MODELING SYSTEM

2.1 Introduction

The case study used throughout this research occurred during the International H₂O Project (IHOP_2002) field study. This study was conducted during the summer of 2002 over the Southern Great Plains region of the United States. The data ingested into the model, as well as observations used for comparison purposes were collected and archived during IHOP_2002.

All simulations were completed using version 3.6.2 of the MM5 modeling system. The fifth-generation NCAR/Penn State Mesoscale Model (MM5) is the latest version of a mesoscale model first used and developed at The Pennsylvania State University in the early 1970’s. MM5 is a hydrostatic, primitive equation model that uses a non-dimensional terrain-following σ-vertical coordinate system.

2.2 International H₂O Project Overview

An overview and preliminary highlights of IHOP_2002 are presented by Weckworth et al. (2004). The IHOP_2002 field study was conducted over the southern great plains of the United States, mainly covering portions of Kansas, Oklahoma and Texas (Figure 2.1), between 10 May and 25 June 2002. One of the main goals of IHOP_2002 was to study water vapor distribution in the atmosphere in order to improve predictions of summertime convection and associated rainfall over the Southern Great Plains.
Figure 2.1 Map of the IHOP_2002 Domain.
IHOP_2002 came about primarily because of the lack of skill in quantitative precipitation forecasts (QPF) over the SGP region during the summer. Accurate rainfall forecasts are imperative during the warm season, due to the hazards associated with excessive rainfall, such as flash flooding. Part of the inaccuracies in QPF lies in the parameterization of convection in operational models. Numerical weather models simply are unable to capture the complex structure and physical processes of convective storms. One major improvement that is needed are more accurate estimates of boundary layer water vapor, since this is such an important factor in predicting buoyancy within clouds and precipitation rates.

IHOP_2002 was a collaboration between many organizations from several countries. The scientists involved sought to address the issues relating to measurements of water vapor and accurate prediction of rainfall. The research was divided up into four categories, representing improvements in QPF relating to a greater accuracy in water vapor measurements, convection initiation, atmospheric boundary layer processes and their feedbacks with convection, and improvements in instrumentation to measure water vapor. Some of the accomplishments of IHOP_2002 included the measurements made of water vapor distribution in three-dimensional time varying scales, the coordination of instruments in the field and the use of new sensors in the field. The data collected during this study will ideally be used to improve the representation of water vapor distribution in the lower atmosphere and subsequent forecasts of convection initiation and rainfall. For more information, refer to Weckworth et al. (2004) or go to http://www.atd.ucar.edu/dir_off/projects/2002/IHOP.html.
2.3 MM5 Modeling System

A flow chart of the MM5 modeling system is shown in Table 2.1. The MM5 modeling system is broken down into three components: 1. Main Programs, 2. Data Sets, and 3. Additional Capabilities. TERRAIN, REGRID, RAWINS, INTERPF, and MM5 are the main programs included in the MM5 model. Programs TERRAIN and REGRID interpolate terrestrial and isobaric atmospheric data in a latitude-longitude mesh to a variable high-resolution model domain. Mesoscale detail is added to the REGRID data with surface and upper air observations obtained from the IHOP_2002 network of surface and rawinsonde stations incorporated into the LITTLE_R program. The LITTLE_R program was modified to assimilate all available IHOP_2002 meteorological data (Figure 2.xx) into the REGRID analysis at 6-hr intervals. Atmospheric data are then interpolated from pressure levels to the vertical sigma coordinate system using the INTERPF program. MM5 is the final main program and is the numerical weather prediction component of the model. The MM5 program includes the various physics options and the governing equations.

Eta model analyses, produced by the National Center for Environmental Prediction (NCEP) and archived by the National Center for Atmospheric Research (NCAR) were used to prescribe initial and lateral boundary conditions. The resolution of the archived data is approximately 40 km. The above data are interpolated onto the model grid to serve as initial values and to provide lateral boundary conditions for the simulation. The analysis corresponding to 0000 UTC 17 June 2002 was utilized as the initial condition. The model was integrated up to a period of 72h ending 0000 UTC 20 June 2002.
2.4 MM5 Physics Options

Table 2.2 shows the MM5 physics configuration used in each study. For the surface data assimilation studies discussed in Chapter 4, surface layer similarity is used for the constant flux layer and the Eta Mellor-Yamada (Eta M-Y) planetary boundary layer (PBL) parameterization scheme for the mixed layer (Betts and Chen 1997). The Eta M-Y is a 2.5 level 1.5-order TKE closure model used in NCEP’s operational Eta model. The scheme requires a prediction equation for the turbulent kinetic energy (TKE) and parameterization of TKE sources and sinks for each model layer. Prediction of the TKE gives better representation of mixing by sub-grid scale eddies that develop as a result of vertical wind shear. Since the diffusion rates at each model layer in the PBL are determined by the wind, moisture, and temperature conditions at the layer's top and bottom interfaces, the PBL closure is considered to be local. The mixing that is emulated in each time step only takes place through the interface between adjacent model layers. The scheme also requires a soil model that calculates ground temperature at multiple levels.

The LSM/PBL study presented in Chapter 5 uses the MRF and ACM PBL schemes. The asymmetric convective model (ACM) is summarized by Alapaty et al. (1997). The ACM is a non-local closure scheme based on Blackadar (1979). The underlying assumption of the Blackadar scheme is symmetric turbulent mixing in the boundary layer. However, both observations and large-eddy simulation modeling have shown that in the convective boundary layer, there is asymmetric mixing (Schumann 1989). The ACM was therefore modified by Pleim and Chang (1992) to add
asymmetrical vertical mixing. As such, this model is only valid for convective conditions, and should not be used for stable situations. The ACM is able to simulate rapid upward transport and slower downward subsidence. The sensible heat flux is used to estimate the mixing rates. Refer to Pleim and Chang (1992) for more details. The medium range forecast (MRF) PBL scheme, also referred to as the Hong-Pan PBL scheme (Hong and Pan 1996) is also a non-local closure scheme based on the representations of the countergradient term and K profile in the well-mixed PBL developed by Troen and Mahrt (1986). Vertical diffusion in the MRF makes use of a fully implicit scheme in order to allow longer time steps and calculate updates of prognostic variables. The PBL height is determined through the use of the Bulk-Richardson number. During unstable conditions, the surface sensible and latent heat fluxes are used to determine an enhancement to the virtual potential temperature, which is in turn used to re-determine the PBL height. Refer to Hong and Pan (1996) for more details.

MM5 uses explicit equations for cloud water, rainwater, ice and water vapor. The Simple Ice scheme was used to account for the ice phase processes. There is no supercooled water and immediate melting of snow below the freezing level.

The Kain-Fritsch cumulus parameterization scheme was used to account for the sub-grid water cycle (Kain, 2004) in the 12 km domain for the first study, while the inner 4 km domain used only explicit moisture physics to account for precipitation processes. The Kain-Fritsch parameterization is a complex cloud-mixing scheme that is capable of solving for entrainment and detrainment processes. The scheme also removes the available buoyant energy in the relaxation time. Updraft and downdraft properties are
also predicted. The influence of shear effects on the precipitation efficiency is also considered by the Kain-Fritsch scheme. The updated version of this scheme, the Kain-Fritsch 2 scheme, was used in the second study.

The Dudhia cloud-radiation scheme was used to account for the interaction of shortwave and longwave radiation with clouds and the clear air. The scheme provides an important contribution in simulating the atmospheric temperature tendencies. Surface radiation fluxes are also considered in this scheme.

The Noah Land Surface Model (LSM) was used to represent all land surface processes in the first study. The Noah LSM is the latest high-resolution land surface model developed and implemented by NCAR and NCEP scientists (Ek et al. 2003). In this LSM, soil temperature and soil moisture are predicted at 4 levels (10, 30, 60 and 100 cm). Soil water/ice, canopy water and snow cover are also predicted. The soil heat flux explicitly includes contributions from both the snow and non-snow covered portions of a model gridbox. This scheme is capable of resolving diurnal temperature variations that result in a more rapid response from the surface temperature. The Noah LSM is an updated version of the Oregon State University (OSU) LSM and was fully implemented into the operational Eta model in spring 2004. Equations for bare soil evaporation and soil thermal conductivity have been revised for use in MM5. The Pleim-Xiu land-surface model is used for one of the simulations in the second study. More information on this LSM will be presented in Chapter 5.
Table 2.1 Flowchart of the MM5 Modeling System
Table 2.2 MM5 Physics Configuration

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<td>Cloud Radiation Physics</td>
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CHAPTER 3: OBSERVATIONAL ANALYSIS OF 17-20 JUNE 2002
CONVECTION EVENT DURING IHOP_2002

3.1 Introduction

Surface features are an integral part of the processes leading to convective initiation. Important surface features in this study include surface convergence, temperature and moisture gradients, and gradients of sensible and latent heat fluxes. The dates of this case study are from 0000UTC on 17 June 2002 to 0000UTC on 19 June 2002, during IHOP_2002.

All of the observations used in this study were taken during the IHOP_2002 field study. The flux data was obtained from the NCAR Integrated Flux Facility (ISFF), which is comprised of nine stations located in Texas, Oklahoma and Kansas (Figure 3.1). ISFF towers are equipped with sensors to measure fluxes of momentum, sensible and latent heat, trace gases, and radiation, in addition to standard surface and atmospheric variables. More specifically, the standard instrumentation at these sites includes:

- Prop vane anemometer, RM Young 9101, for mean wind speed and direction at 10 m agl
- Vaisala 50Y Humitter to measure air temperature and RH; in NCAR aspirated radiation shield at 2 m agl
- Barometer, Vaisala PTB220B, with a Ser single-disk static pressure port
- MRI model 302/303/304 tipping bucket rain gauge, with Alter wind screen at sites 1-3
- Sonic anemometer, initially either Campbell CSAT3 or ATI-NUW
- Fast-response hygrometer, Campbell KH2O
- Net radiometer, REBS Q*7
• Incoming longwave (Eppley PIR) and shortwave (Eppley PSP or Kipp & Zonen CM 21) radiometers

• PAR (Photosynthetically Active Radiation) sensor, Li-Cor 190SA Quantum Sensor

• Infrared surface temperature sensor, Everest 4000.4GL

• Near-surface soil heat flux plate, REBS HFT-3

• Near-surface soil temperature (REBS) and water content (Campbell Scientific CS-615) sensors

• Soil profile consisting of a Campbell Scientific 107 temperature sensor, a Campbell Scientific 229 heat dissipation matric water potential sensor and a Decagon ECH2O Dielectric Aquameter at 6 depths: 7.5, 15, 22.5, 37.5, 60, and 68-90 cm.

Figure 3.1 Map showing the locations of the NCAR Integrated Surface Flux Facility stations.
For more information on ISFF, visit http://www.atd.ucar.edu/rtf/projects/ihop_2002/isff/report.shtml. The surface sensible and latent heat fluxes were determined using the eddy correlation method. Sensible heat fluxes were calculated by the covariance between the vertical velocity and temperature fluctuations ($\overline{w\theta'}$). Likewise, the latent heat fluxes were calculated by the covariance between the vertical velocity and mixing ratio fluctuations ($\overline{w'q'}$). Vertical velocity fluctuations were measured via a 3-dimentional sonic anemometer. The temperature and mixing ratio fluctuations were obtained using a fast-response temperature sensor and a fast-response ultra-violet absorption hygrometer, respectively.

3.2 Synoptic Overview

Synoptic conditions for this time period began under northwest flow aloft over the IHOP_2002 region. The ridge axis was oriented SW-NE over the central Rockies. Conditions were clear on the morning of 17 June, with strong southwest winds at the surface. By the afternoon, a deepening surface low was over central Colorado, with an ENE-WSW oriented dryline moving SSE out of Colorado into NW Kansas. Convection began to fire along this dryline in Kansas and the Oklahoma panhandle between 1900 and 2000 UTC (Figure 3.5). Low-level jet development in western Oklahoma and Kansas supported continuing convection propagating ESE across Kansas overnight.

On 18 June, the region was still under weak northwest flow, becoming more westerly overnight. A boundary was in the vicinity of Garden City, Kansas by late morning (Figure 3.8), appearing to be a combination of a weak cold front and dryline. Convection was occurring during the morning ahead of this boundary in the central to
eastern Kansas region. The boundary began to drift back to the NNW during the afternoon. Weak convection occurred around 2000 UTC over the Front Range along the dryline in SE Colorado (Figure 3.11) and moved into north central Kansas; however convection did not intensify overnight.

The upper air pattern shifted to more WSW flow on 19 June. There was a cold front in the morning located west of Goodland, Kansas. Between 1740 and 1905 UTC, a NNE-SSW oriented boundary was observed near Goodland, moving towards Colby, Kansas (Figure 3.12). Convection began in SE Colorado around 2000 UTC, and rapidly fired along the boundary in Kansas, propagating eastward overnight.

3.3 Observations

Multisensor Precipitation Estimate (MPE) data was used to plot accumulated precipitation for this case study. MPE corrects radar estimates of precipitation with measurements from surface rain gauges. This method gives a better spatial coverage of precipitation than radar estimates alone (Johnson et al. 1999, Krajewski 1987, Seo 1998, Young et al. 2000). The 24 hour accumulated precipitation from MPE data, ending at 0000 UTC 18 June, is shown in Figure 3.2. The precipitation is not particularly widespread over the first 24 hours. The largest accumulations are in extreme southwestern Kansas into the central Oklahoma panhandle, spreading southward into the Texas panhandle, where small pockets of precipitation reached upwards of 7.5 cm. Lighter areas of precipitation fell across northeastern New Mexico, central Kansas and southeastern Nebraska into northwestern Missouri, with the majority of accumulations ranging from 0.75 cm to 2.75 cm, along with couple of very small regions in southeastern
Nebraska of 4.5cm. Accumulated precipitation from MPE data over the second 24-hour period, ending 0000 UTC 19 June, is shown in Figure 3.3. Hardly any precipitation fell from 0000UTC 18 June through 0000UTC 19 June 2002. There was a small region just east of central Kansas, with accumulations ranging from 0.5cm to a tiny area of 4.75cm. Small amounts of precipitation also fell in southwestern Iowa, northwestern and west-central Missouri. Accumulations here were fairly small, mainly 0.5cm to 1.0cm, with a very small area in southwestern Iowa of 2.5cm to 4.25cm. There were also tiny regions of around 0.5cm accumulations in north-central and northwestern Oklahoma, as well as northeastern New Mexico. The majority of the precipitation between 0000UTC 17 June and 0000UTC 20 June 2002 fell during the last 24 hours, as is shown in the MPE data in Figure 3.4. Accumulations were the greatest in western Kansas, south-central Nebraska, and eastern Colorado, along with a very small region in southeastern New Mexico. Amounts in these areas ranged from 0.5cm to 8cm, along with a small region of greater than 8.5cm in western Kansas. Widespread precipitation also occurred across northeastern Colorado and central Nebraska, with patchy areas in western Texas, and eastern Texas into Louisiana. These areas were mainly 0.5cm to 2.5cm.

A surface map for the central United States valid 2000 UTC on 17 June is shown in Figure 3.5. The dryline boundary can clearly be seen in western Kansas in this map, accompanied by a temperature gradient and a strong moisture gradient. Across the dryline between Hill City (HLC) and Goodland, Kansas (GLD), a distance of approximately 137 miles (219km), the dew point drops from 60ºF to 46ºF (a 14ºF drop), while the temperature rises from 92ºF to 98ºF. Similarly, the dew point gradient between Dodge City (DDC) and Elkhart, Kansas (EHA) is 14ºF across a distance of 121 miles.
(194km). Also, there is some evidence of convergence in the wind pattern across the dryline region. In southeastern Colorado, the winds are primarily from the WNW, while in southwestern Kansas, they are from the southwest. There is also evidence of weak speed convergence in southwestern to central Kansas, with 25kt winds at Liberal (LBL), 20kt winds at Dodge City and Great Bend (GBD), and 15kt winds at Hays (HYS). In the satellite image valid for the same time (Figure 3.6), convection can be seen firing along this dryline boundary in western Kansas and the Oklahoma panhandle.

Surface sensible and latent heat fluxes for 17 June are shown in Figure 3.7. These flux measurements were taken at Elmwood, Oklahoma, (Figure 3.7A) and Zenda, Kansas (Figure 3.7B), which are approximately 180 miles (288km) apart. Refer to Figure 3.1 for the locations of these stations. Elmwood is located in the hot, dry air to the west of the dryline and Zenda is located in the warm, moist air to the east of the dryline. Evidence of this can be seen in the surface fluxes. Higher sensible heat fluxes are observed over Elmwood, where the maximum is approximately 300 W m\(^{-2}\), while at Zenda the maximum is around 100 W m\(^{-2}\). The opposite is seen for the latent heat fluxes, with lower values maxing out at around 200 W m\(^{-2}\) over Elmwood and higher values of about 325 W m\(^{-2}\) over Zenda. This is evidence of a significant flux gradient across the dryline, which may have helped fuel the convection.

In Figure 3.8, a surface map valid 1500 UTC 18 June is shown. Once again, a boundary can be easily located by looking at the temperature and dew point gradients, as well as a shift in the surface winds. The largest gradient appears to be between HLC and GLD, where the temperature rises from 78\(^\circ\)F to 88\(^\circ\)F (10\(^\circ\)F change), and the dew point drops from 63\(^\circ\)F to 39\(^\circ\)F (24\(^\circ\)F change).
Figure 3.2 Accumulated precipitation valid 0000 UTC 17 June through 0000 UTC 18 June
Figure 3.3 Accumulated precipitation valid 0000 UTC 18 June through 0000 UTC 19 June
Figure 3.4 Accumulated precipitation valid 0000 UTC 19 June through 0000 UTC 20 June
Figure 3.5 Surface map valid 2000 UTC 17 June. A surface dryline is outlined in black.
Figure 3.6 Visible satellite image valid 2003 UTC 17 June.
Figure 3.7 Surface fluxes on 17 June at Elmwood, OK (Figure 3.7A – top) and Zenda, KS (Figure 3.7B – bottom).
Figure 3.8 Surface analysis valid 1500 UTC 18 June. A surface dryline is outlined in black.
Between DDC and EHA, the change is less drastic, with the temperature rising from 77°F to 82°F (5°F) and the dew point dropping from 64°F to 57°F (7°F). In general, there is a large moisture gradient on either side of the boundary, with dew points throughout most of Oklahoma, Kansas and eastern Nebraska in the low to mid 60’s, while the dew points in Colorado and western Nebraska range from the lower 30’s to the lower 50’s. Surface convergence can also be seen in Figure 3.9 by examining the wind field. Winds to the west (in Colorado, western Nebraska and extreme northwestern Kansas) are from the WNW, while to the east they are primarily SSW to SSE. Convection is occurring ahead of the boundary at this time, as shown in the satellite image in Figure 3.9. There is also convection forming along the boundary in western Nebraska.

Surface sensible and latent heat fluxes for 18 June are shown in Figure 3.10. Once again, these are observations taken at Elmwood, Oklahoma (Figure 3.10A) and Zenda, Kansas (Figure 3.10B). As on the previous day, Elmwood is in hot dry air and Zenda is in warm, moist air. The sensible heat fluxes on this day are higher over Elmwood, where the maximum is about 300 W m⁻². The sensible heat fluxes over Zenda have a maximum around 125 W m⁻². On the other hand, the latent heat fluxes are lower over Elmwood, with the maximum being around 150 W m⁻². Over Zenda, the latent heat fluxes are higher, with a maximum of approximately 300 W m⁻². Just as on the previous day, this is evidence of a strong flux gradient, which may have aided in supporting the convection on 18 June.

A surface analysis valid 2000 UTC 18 June is shown in Figure 3.11. Weak convection was occurring at this time along a dryline in western Kansas into southeastern Colorado. A surface analysis valid 1900 UTC 19 June is shown in Figure 3.12.
Convection initiation occurred along the boundary located across western Kansas around 2000 UTC. Surface boundaries and fluxes appeared to play a significant role in the formation of deep, moist convection over this 72-hour time period.
Figure 3.9 Visible satellite image valid 1515 UTC 18 June.
Figure 3.10 Surface fluxes on 18 June at Elmwood, OK (Figure 3.7A – top) and Zenda, KS (Figure 3.7B – bottom).
Figure 3.11 Surface analysis valid 2000 UTC 18 June. A surface boundary is outlined in black.
Figure 3.12 Surface analysis valid 1900 UTC 19 June. A surface boundary is outlined in black.
4.1 Introduction

As reliance on numerical weather prediction models continues to increase, more accurate and detailed data assimilation systems are essential. Data assimilation is based on the concept of combining current and past meteorological data in an explicit dynamical model. Four-dimensional data assimilation (FDDA) is the time dependent dynamical coupling of various numerical fields with the model’s prognostic equations. FDDA provides a logical extension between objective analysis methods and dynamic relationships of atmospheric variables.

At least two types of sequential FDDA are currently used in operational and research models. The first is a process of initializing an explicit prediction model, using subsequent forecast cycles (typically 3-12 hrs) as a first guess in the static three-dimensional objective analysis, and then repeating this step for future forecasts. This process is used in many current operational forecast models. A second common method of FDDA uses a continuous (i.e., every time step) dynamical assimilation where forcing functions are added to the governing equations to “nudge” the model state toward the observations. This type of FDDA is often used in the research community to study various mesoscale features. Users of the MM5 modeling system frequently use continuous nudging FDDA. Nudging was initially developed and tested by Kistler (1974) and by Anthes (1974). Refer to Stauffer and Seaman (1990) for a more detailed review of these techniques.
4.2 FASDAS

The Flux-Adjusting Surface Data Assimilation System (FASDAS) has been continuously developed over the last few years (Alapaty et al. 2001a,b,c). FASDAS builds on the works of Mahfouf (1991) and Bouttier et al. (1993). They used the surface layer temperature and humidity to estimate the soil moisture and temperature evolutions in numerical model predictions. These schemes work well, but assume that the largest errors in the simulated surface energy budget are due to errors only in the soil moisture parameter. However, these errors are not always significant, especially during cloudy conditions. Both soil moisture and soil temperature are important variables in the development of deep convection. FASDAS was developed not only to address these issues but also to include direct and indirect assimilation components utilizing the FDDA methodology employed by Stauffer and Seaman (1990).

The FASDAS algorithm is outlined by Alapaty et al. (2001a,b,c). Observations of surface temperature and dew point temperature are assimilated directly into the lowest model atmospheric layer (the surface layer) and interpolated onto the model grid. Forecasted temperature and dew point data are then compared with these interpolated observations. FASDAS calculates the difference between the model analyses and the observations, and then adjusts this difference based on a number of weighting terms, including a nudging factor. This creates an FDDA term, which is then used to calculate adjustments to the simulated surface sensible and latent heat fluxes. These adjustments are added to the surface heat fluxes simulated by the model. The heat flux adjustments are also used in the prognostic ground temperature and soil moisture equations, which in turn have an effect on the simulated surface heat fluxes in the subsequent time step. This
process alters the surface energy budget and then the soil temperature and moisture. This method helps maintain consistency between the ground temperature and soil moisture with the surface-layer mass variables.

FASDAS uses continuous data assimilation, which occurs at each time step throughout the model integration period. Continuous data assimilation has an advantage over intermittent data assimilation, since it reduces the chance of mass and dynamical imbalances being discontinuously brought into the model. This method, however, was not necessarily designed to cause model-simulated fluxes to match observations. Instead, simulated air temperature and humidity are meant to converge towards observations. This allows for a realistic evolution of the simulated structure of the atmosphere, regardless of the source of errors for simulated temperature and humidity (Alapaty et al. 2001a). For further details on the FASDAS process, refer to Alapaty et al. (2001a,b,c).

Preprocessing and analysis packages within MM5 provide an accessible interface for processing. The quality assurance and quality control (QA/QC) for the observations are performed as a part of the pre-processing of the data. A flow chart describing the FASDAS procedure is shown in Table 4.1.

In the MM5 model, a preprocessing system provides initial estimates of soil moisture and soil temperature at various depths, water-equivalent snow depth, sea ice, and canopy moisture. These fields are obtained from the Eta Advanced Weather Interactive Processing (AWIP) analyses. Chen and Dudhia (2001) show that much uncertainty still exists in the initial soil moisture estimates, which in turn contributes to model prediction errors. Alapaty et al. (2001b,c) demonstrate how FASDAS corrections for soil moisture and soil temperature can significantly improve model forecasts and
reduce errors. For example, using FASDAS, the RMS errors for PBL height over the FIFE study region in the Great Plains were reduced from 405 m to 155 m (Alapaty et al. 2001a). Similarly, Alapaty (2001b,c) reported consistent improvements in model estimated soil temperature and soil moisture, which lead to improvements in model simulated dew point temperature, lifting cloud level, cumulus convection and

Table 4.1 Flowchart of the FASDAS algorithm.

Interpolated observations \( T_s, q_s \) — Model Analysis or Forecasts \( \hat{T}_s, \hat{q}_s \)

Data Assimilation Term
\(((\text{Weighting Terms})(\hat{T}_s - T_s)) \)
\(((\text{Weighting Terms})(\hat{q}_s - q_s)) \)

Adjustments to surface sensible and latent heat fluxes

Adjustment to ground temperature and moisture via surface energy budget
precipitation. Therefore, initial studies indicate that the MM5-FASDAS coupling improves model performance by enhancing land-atmosphere feedbacks. For the Experimental Simulation described here, MM5 was altered to allow for the full implementation of the FASDAS scheme. The code was modified to allow for observational nudging of the mass fields, and to allow for direct interaction between atmospheric variables and related surface fluxes. The observations used for the data assimilation were obtained from the IHOP_2002 hourly surface mesonet composite. A map of all the stations included in this composite is shown in Figure 4.1. Observational nudging is prescribed at an interval of 180 minutes throughout the 72-hr model integration.

4.3 Results and Discussion

Plots of accumulated precipitation over the domain are shown in Figures 4.2 through 5.4. For each of these four-panel plots, A is from the MPE data (also shown in Chapter 3), B is from the Control (Non-FASDAS) Simulation, C is from the Experimental (FASDAS) Simulation and D is the difference plot between the two simulations. Refer to Figures 3.3, 3.4 and 3.5 for a description of the MPE plots. The 24 hour accumulated precipitation ending 0000 UTC 18 June for the Control Simulation is shown in Figure 4.2B. This simulation produced small amounts of precipitation in two main areas: central Nebraska and eastern New Mexico. These areas are also seen in the observations (Figure 4.2A); however the model over-predicted the spatial coverage and under-predicted the intensity. The model also completely missed the precipitation in
Kansas and the Texas panhandle regions, along with producing precipitation in Arkansas not seen in the observations.

Figure 4.1 Map of surface stations included in the IHOP_2002 hourly surface mesonet composite.
The Experimental Simulation generated significantly more precipitation during the first 24 hours of the simulation (Figure 4.2C) than the Control Simulation. A large area of precipitation ranging from 0.5 to greater than 8.5 cm is located in central to southeastern Nebraska into northeastern Kansas. When compared to the observations, the Experimental Simulation over predicted both the amount and spatial coverage of the precipitation in this region. The other large area of precipitation in this simulation is located in Colorado into New Mexico and the Texas panhandle. Accumulations in these regions range from 0.5 to 3 cm to as high as 5 to 7 cm. Once again, the model significantly over predicted the precipitation in this area. No precipitation fell in Colorado during the first 24 hours, and very little fell in New Mexico. The precipitation in Texas is not nearly as widespread in the observations as in the Experimental Simulation. Just as in the Control Simulation, the Experimental Simulation generated precipitation in Arkansas, and also in southwestern Missouri and eastern Oklahoma, none of which was in the observations. The model also missed the precipitation in central Kansas.

In Figure 4.2D, the difference between the Control and Experimental Simulations is shown. It is clear from this plot that the Experimental Simulation produced far more precipitation during the first 24 hours than the Control Simulation. The largest differences are seen in eastern Nebraska and Colorado into New Mexico and the Texas panhandle. The Experimental Simulation generated greater than 3.5 cm more precipitation than the Control in both eastern Nebraska and the Texas panhandle, as well as a small region in southeastern New Mexico. In Colorado, the Experimental
Figure 4.2 Accumulated precipitation (cm) valid 0000 UTC 17 June through 0000 UTC 18 June. MPE data is shown in A, the Control Simulation in B, the Experimental Simulation in C and the difference between the Control and Experimental in D.
Simulation produced 0.25 to about 2.5 cm more precipitation. There is a very small region in eastern New Mexico where the Control Simulation produced approximately 0.75 cm more precipitation than the Experimental.

The 24 to 48 hour accumulated precipitation ending 0000 UTC 19 June is shown in Figure 4.3. During this time period, the Control Simulation (Figure 4.3B) did not generate any of the precipitation seen in the observations (Figure 4.3A) in Kansas or Missouri, with the exception being southwestern Iowa. However, the model over predicted the spatial coverage of precipitation in Iowa, while under predicting the intensity. The Control Simulation also produced precipitation in eastern and central Nebraska, which did not occur in the observations.

The Experimental Simulation (Figure 4.3C) was not as intense for the second 24 hours of the simulation as for the previous time period; however it once again generated more precipitation than the Control. This simulation captured the precipitation in central Kansas, but over predicted the spatial coverage. The precipitation located in Missouri is missing in this simulation, while the southwestern Iowa precipitation has been shifted westward into Nebraska. The model also generated small, widespread swaths of precipitation in Colorado, New Mexico and the Texas panhandle that are not in the observations.

The difference plot between the Control and Experimental Simulations for this time (Figure 4.3D) once again clearly shows that the Experimental Simulation produced more precipitation overall. In central to southeastern Nebraska into northern Kansas, the Experimental Simulation produced 0.25 up to greater than 3.5 cm more precipitation. Also, the swaths of precipitation found in Colorado, New Mexico and Texas in the
Figure 4.3 Same as Figure 4.2, except valid from 0000 UTC 18 June through 0000 UTC 19
June. Experimental Simulation are not present in the Control Simulation. One region where the Control Simulation generated more precipitation is eastern Nebraska into western Iowa, where the difference ranges from 0.25 to 1.25 cm.

During the final 24 hours of the simulation, the Control Simulation barely produced any precipitation over the entire domain (Figure 4.4B). The only areas of precipitation seen are in western Nebraska and eastern Texas into western Louisiana, where accumulations are less than 0.75 cm. While there is precipitation in these areas in the observations (Figure 4.4A), the Control Simulation underpredicted the amount and spatial coverage. This simulation did not generate any of the precipitation seen in Kansas and Colorado.

The Experimental Simulation (Figure 4.4C) once again produced more precipitation than the Control Simulation during the final 24 hour time period. However, this simulation did not capture all of the precipitation that fell in the observations. The model generated precipitation in western Nebraska and eastern Colorado, and a small amount in western Kansas. The intensity and spatial coverage of the simulated precipitation are less than the observed, and also the convection in central Nebraska appears to have been shifted westward by the model. The Experimental Simulation also produced a small bull’s-eye of precipitation at the corner of Missouri, Kansas and Nebraska, which is not in the observations. An interesting thing to note is that the majority of the observed precipitation fell during the final 24 hours; however in the Experimental Simulation, the majority of the precipitation fell during the first 24 hours. There are two main areas where the Experimental Simulation produced more precipitation than the Control, as seen in Figure 4.4D. The main area is in southwestern
Figure 4.4 Same as Figures 4.2 and 4.3, except valid 0000 UTC 19 June through 0000 UTC 20 June.
Nebraska into Colorado, where the Experimental Simulation generated 0.25 to greater than 3.5 cm more precipitation than the Control. In southeastern Colorado, the difference is about 0.5 cm, and in the small bulls eye in the Missouri/Kansas/Nebraska region, the Experimental Simulation produced 0.25 to 1.5 cm more precipitation. Overall, over the entire 72-hour integration, it appears that the Experimental Simulation produced significantly more precipitation than the Control Simulation, while also coming a little closer to the observations.

Spatial plots of surface sensible and latent heat fluxes are shown in figures 4.5 through 4.12. In each of these plots, A is from the Control Simulation and B is from the Experimental Simulation. In Figure 4.5, sensible heat fluxes from 1800 UTC 18 June 2002 are shown. Over the domain, the fluxes are overall higher and more uniform in the Control Simulation at this time (Figure 4.5A), mainly ranging from 150 to 350 W m$^{-2}$, with a few small patches of 25 to 75 W m$^{-2}$. On the other hand, the Experimental Simulation produced more variation at this time (Figure 4.5B). There is a large flux gradient across Kansas, with fluxes of approximately 350 W m$^{-2}$ in western Kansas and approximately 25 to 50 W m$^{-2}$ in eastern Kansas. There are also some small patches where the fluxes are 400 to 450 W m$^{-2}$ in northern Oklahoma and eastern New Mexico.

Sensible heat fluxes for 0000 UTC 19 June 2002 are shown in Figure 4.6. The Control Simulation (Figure 4.6A) is once again more uniform, with fluxes ranging from 25 to 75 W m$^{-2}$. The Experimental Simulation (Figure 4.6B) is similarly uniform, except for the southwestern portion of the domain, where fluxes range from 25 to 125 W m$^{-2}$. Over portions of central Kansas, western Nebraska, southern Oklahoma and north-central Texas, the Control Simulation produced higher fluxes at this time. Conversely, over
Figure 4.5 Surface sensible heat fluxes (W m$^{-2}$) valid 1800 UTC 18 June. A is from the Control Simulation and B is from the Experimental Simulation.
Figure 4.6 Surface sensible heat fluxes (W m$^{-2}$) valid 0000 UTC 19 June.
portions of eastern Nebraska, Colorado and New Mexico, the Experimental Simulation produced higher sensible heat fluxes.

In Figure 4.7, sensible heat fluxes for 1800 UTC 19 June 2002 are shown. Just as in Figure 4.5, the Control Simulation (Figure 4.7A) produced much more uniform and generally higher fluxes over the entire domain than the Experimental Simulation (Figure 4.7B). Several areas of larger flux gradients are seen in the Experimental Simulation at this time, especially in the northwest Kansas/northeast Colorado/western Nebraska region, as well as eastern Kansas into Missouri and spreading southward into eastern Oklahoma. There are also some higher flux gradients in eastern New Mexico in the Experimental Simulation. The Control Simulation produced less strong gradients, with the exception of the Texas panhandle region.

The final sensible heat flux spatial plot is shown in Figure 4.8, valid 0000 UTC 20 June 2002. Over the eastern portion of the domain, there are higher fluxes (and a larger flux gradient) in the Control Simulation (Figure 4.8A). Over the western half of the domain, the Control Simulation appears to have overall produced larger areas of higher fluxes, however the Experimental Simulation (Figure 4.8B) this time has larger flux gradients, especially in areas of Colorado and New Mexico. The differences seen in the flux fields are possibly due to differences in simulated solar radiation due to cloud cover in the model. This issue will be discussed further with the flux time-series plots.

Figures 4.9 through 4.12 all contain spatial plots of surface latent heat fluxes. Latent heat fluxes valid 1800 UTC 18 June 2002 are shown in Figure 4.9. The fluxes at this time are higher in the Control Simulation (Figure 4.9A), with values mainly ranging from 200 to 650 W m$^{-2}$. The Experimental Simulation (Figure 4.9B) has generally lower fluxes and a
greater variation, with fluxes ranging from 50 to 600 W m$^{-2}$. There are also higher gradients in the Experimental Simulation, especially across eastern Kansas, central Nebraska and the Texas panhandle.
Figure 4.7 Surface sensible heat fluxes (W m$^{-2}$) valid 1800 UTC 19 June.
Figure 4.8 Surface sensible heat fluxes (W m$^{-2}$) valid 0000 UTC 20 June.
Figure 4.9 Surface latent heat fluxes (W m$^{-2}$) valid 1800 UTC 18 June.
The latent heat fluxes on 0000 UTC 19 June 2002 (Figure 4.10) are much more uniform and higher in the Control Simulation (Figure 4.10A) than in the Experimental Simulation (Figure 4.10B). Large gradients are present across the eastern half of the domain in the Experimental Simulation, with fluxes ranging from less than 50 W m\(^{-2}\) in the extreme eastern portion to approximately 350 W m\(^{-2}\) in the central portion of the domain. This large gradient is not present in the Control Simulation at this time.

On 1800 UTC 19 June 2002 (Figure 4.11), there appear to be a few gradients of latent heat flux present in both the Control (Figure 4.11A) and Experimental (Figure 4.11B) Simulations. In the Control Simulation, there is a large gradient in the Texas panhandle region, with values ranging from 50 to 400 to a small area of 600 W m\(^{-2}\). Fluxes in this simulation are higher in the southeastern portion of the domain. Across the eastern half of the domain, the Experimental Simulation produced large gradients, ranging from 50 to 700 W m\(^{-2}\). There is also a region on northeast Colorado into western Nebraska where the fluxes are smaller than the surrounding area.

The final latent heat flux spatial plot is valid 0000 UTC 20 June 2002 (Figure 4.12). Overall, the Control Simulation (Figure 4.12A) produced higher and more uniform fluxes. The Experimental Simulation (Figure 4.12B), on the other hand, produced generally lower fluxes and much greater flux gradients over the entire domain, especially in Kansas and Oklahoma. These higher gradients seen in the Experimental Simulation throughout most of the 72 hour time period most likely are what led to the increased amount of precipitation as compared to the Control Simulation.
Figure 4.10 Surface latent heat fluxes (W m\(^{-2}\)) valid 0000 UTC 19 June.
Figure 4.11 Surface latent heat fluxes (W m$^{-2}$) valid 1800 UTC 19 June.
Figure 4.12 Surface latent heat fluxes (W m$^{-2}$) valid 0000 UTC 20 June.
A map of the nine NCAR flux stations, as well as the various sounding sites around the IHOP region, is shown in Figure 4.13. The NCAR stations used to examine time series of fluxes and soil data are Elmwood, Oklahoma; Spivey, Kansas and Atlanta, Kansas; while the sounding stations used for theta profiles are North Platte, Nebraska; Dodge City, Kansas and Amarillo, Texas.

Surface heat fluxes for Elmwood, Oklahoma; Spivey, Kansas and Atlanta, Kansas, valid 0000 UTC 17 June through 0000 UTC 20 June 2002, are shown in Figures 4.14, 4.15 and 4.16, respectively. For each of these figures, surface sensible heat fluxes (W m$^{-2}$) are shown in Figure A and surface latent heat fluxes (W m$^{-2}$) are shown in Figure B.

Figures 4.14A, 4.15A and 4.16A indicate that the Experimental Simulation appears to have performed better for the surface sensible heat fluxes. For both Atlanta and Elmwood, the Experimental simulated fluxes matched the observations fairly closely over most of the time period. Sensible heat fluxes were under estimated over Spivey in the Experimental Simulation; however by 19 June they were closer to the observations. The Experimental Simulation did seem to have problems during the first 24 hours of the integration, especially for Elmwood and Spivey, but seemed to perform better in the latter part of the simulation. The Control Simulation did not perform as well for Atlanta, where fluxes were over estimated throughout the entire simulation, or for Elmwood, where fluxes were under estimated. However, the Control Simulation performed well for Spivey, where simulated fluxes matched the observations fairly well.

Neither simulation seemed to overall perform better than the other for the surface latent heat fluxes (Figures 4.14B, 4.15B and 4.16B). The Experimental Simulation came
Figure 4.13 Map showing the locations of the NCAR Integrated Surface Flux Facility stations. Cross section locations are indicated by the solid red lines.
Figure 4.14 Time series of surface sensible (A) and latent (B) heat fluxes (W m$^{-2}$) for Elmwood, OK.
Figure 4.15 Time series of surface sensible (A) and latent (B) heat fluxes (W m$^{-2}$) for Spivey, KS.
Figure 4.16 Time series of surface sensible (A) and latent (B) heat fluxes (W m\(^{-2}\)) for Atlanta, KS.
close to observations for Atlanta on 17 and 18 June, but then over estimated fluxes significantly on 19 June. For both Elmwood and Spivey, the Experimental simulated latent heat fluxes were under estimated on 17 June, then over estimated for the remainder of the integration. The Control Simulation consistently over estimated fluxes for all three sites, however for both Atlanta and Spivey the Control simulated latent heat fluxes were closer to the observations than the Experimental for the final 24 hours of the simulation.

Time series of root mean square (RMS) errors for the surface sensible and latent heat fluxes shown in Figure 4.17. These errors were calculated using the model output and observational data from all nine NCAR flux stations. The equation used to calculate the RMS error for any given variable $x$ is

$$\text{rmse}(x) = \left[ \frac{1}{N} \sum_{n=1}^{N} (x_n^f - x_n^o)^2 \right]^{1/2}, \quad (4.1)$$

where $N$ is the total number of values (in this case nine), $f$ is the model output value and $o$ is the observed value.

The sensible heat flux RMS errors are shown in Figure 4.17A. During the first 24 hours of the simulation, the errors for both the Control and Experimental simulations are very similar, with the main difference being a time lag between the error peaks. However, over the majority of the remaining 48 hours, the Experimental Simulation has lower RMS errors than the Control, with the highest difference between the two on the final day being around 75 W m$^{-2}$.

Latent heat flux RMS errors are shown in Figure 4.17B. For the first 60 hours, the Experimental Simulation has considerably lower RMS errors than the Control, with a maximum difference of approximately 125 W m$^{-2}$ on both the first and second days. During the final 12 hours, however, the RMS errors for the Experimental Simulation are
Figure 4.17 Time series of RMS errors of sensible (A – top) and latent (B – bottom) heat fluxes (W m\(^{-2}\)) for all nine NCAR ISFF stations.
a bit higher than the Control, except for the end of the integration, when the error for the Experimental Simulation begins to approach zero.

Time series of incoming solar (shortwave) radiation (W m$^{-2}$) over Elmwood, Spivey and Atlanta are shown in Figures 4.18A, 4.18B and 4.18C, respectively. Differences in the amount of shortwave radiation generated by each simulation are likely due to differences in the simulated cloud cover. For each site, the observed solar radiation and both simulations converge during the final 24 hours of the simulation. Over Elmwood, both simulations match the observations during the final 48 hours. For all three sites, the Experimental Simulation underestimated the solar radiation during the first 24 hours, then performed better during the final 48 hours. The Control Simulation matched well with observations over the entire 72-hour period for Elmwood and Spivey, while overestimating solar radiation during the second 24 hours over Atlanta. These differences in solar radiation values may have caused the differences in the fluxes generated by each simulation.

Spatial plots of 10 cm soil temperature (°C) and soil moisture are shown in Figures 4.19 through 4.26. For the soil temperature plots, A is from the Control Simulation and B is from the Experimental Simulation. Soil moisture output was not available for the Control Simulation, so those plots are from the Experimental Simulation only. On 1800 UTC 18 June 2002, the Control Simulation (Figure 4.19A) yielded higher soil temperatures across the entire domain than the Experimental Simulation (Figure 4.19B). The 10 cm soil temperatures in the Control Simulation range from about 24 to 30°C over the northeastern portion of the domain, 27 to 33°C over much of Kansas, Oklahoma, Texas and New Mexico (reaching 35°C in parts of eastern Oklahoma and
Texas as well as western Kansas), and as high as 36 to 39ºC in western Nebraska and northeastern and southern Colorado. Temperatures in the Experimental Simulation at this time range from 18 to 30ºC, with the lowest temperatures (18 to 20ºC) in southern Iowa into northern Missouri, as well as the Texas panhandle. The highest temperatures (27 to 31ºC) are in Colorado and western Kansas. Over the rest of the domain, the temperatures range from about 20 to 25ºC, with the eastern half of the domain being cooler than the western half.

While the soil temperatures in the Control Simulation on 0000 UTC 19 June 2002 (Figure 4.20A) are overall higher over the domain than the Experimental Simulation (Figure 4.20B) (with the exception of parts of the western half of the domain), there is a much higher gradient present in the latter simulation. Across the eastern half of the domain in the Control Simulation, the temperatures mainly range from 27 to 30ºC, with a small region of lower temperatures (~21 to 23ºC) in Iowa and higher temperatures (~32ºC) in Oklahoma and Texas.
Figure 4.18 Time series of incoming solar (shortwave) radiation (W m$^{-2}$) over Elmwood, OK (A), Spivey, KS (B) and Atlanta, KS (C).
Figure 4.19 Spatial plots of 10 cm soil temperature (°C) valid 1800 UTC 18 June.
Figure 4.20 Spatial plots of 10 cm soil temperature (°C) valid 0000 UTC 19 June.
Temperatures are higher in the northwestern portion of the domain, where they range from 30 to 35ºC. The highest temperatures in this simulation are in southern Colorado. As was previously mentioned, there is a strong soil temperature gradient in the Experimental Simulation at this time. Temperatures in this simulation range from 20 to 28ºC across the eastern half of the domain, as well as in the Texas panhandle, up to 29 to 38ºC in the western half. The warmest temperatures are in Colorado and western Kansas, and also in New Mexico. In these regions, the temperatures are warmer than in the Control Simulation, however the rest of the domain has cooler temperatures than in the Control.

On 18 UTC 19 June 2002 (Figure 4.21), the Control Simulation (Figure 4.21A) once again produced higher 10 cm soil temperatures than the Experimental Simulation (Figure 4.21B). In the Control Simulation, temperatures across the eastern parts of Kansas, Oklahoma and Texas are very warm, ranging from 30 to 36ºC. The same temperatures are seen in western Nebraska into northeastern Colorado, as well as southern Colorado. The coolest temperatures are seen in the Texas panhandle, where they range from 20 to 28ºC. There is a large temperature gradient between the Texas panhandle region and eastern Texas/Oklahoma. Most of western Kansas, eastern Nebraska and Iowa contain soil temperatures of 28 to 30ºC. The temperatures in the Experimental Simulation are much cooler than the Control, and the gradient is not as large as in the previous time (Figure 4.20B). The coolest temperatures this time are seen in northern Colorado and western Nebraska, as well as eastern third of the domain. Temperatures in these regions range from 18 to 24ºC. Across the central portion of the
Figure 4.21 Spatial plots of 10 cm soil temperature (°C) valid 1800 UTC 19 June.
domain, temperatures are 25 to 28°C, and there is a small area in southern Colorado of 29 to 32°C temperatures.

On 0000 UTC 20 June 2002 (Figure 4.22), temperatures in the Control Simulation (Figure 4.22A) are generally warmer in the east and cooler in the west than the Experimental Simulation (Figure 4.22B). Across eastern Kansas and Oklahoma, as well as northeastern and south central Colorado, the Control simulated temperatures range from 30 to 33°C. Over the rest of the domain, temperatures are cooler, ranging from 23 to 27°C. In the Experimental Simulation, temperatures range from 20 to 27°C across much of the eastern half and northwestern portion of the domain. Temperatures are warmer in the southwestern portion and southern Colorado, where they reach upwards of 30 to 35°C. The southwestern portion is warmer in this simulation is warmer than in the Control Simulation, whereas the eastern half is cooler than the Control. Overall, it appears that the Control Simulation generated warmer temperatures over the duration of the simulation.

The 10 cm soil moisture for 1800 UTC 18 June 2002 from the Experimental Simulation is shown in Figure 4.23. The highest soil moisture values at this time can be found in southeastern Nebraska, where the range is from 0.32 to 0.42. A secondary maximum is located in the Texas panhandle, where the moisture ranges from 0.30 to 0.36. Moisture values in the remainder of the eastern half of the domain are moderate, around 0.26 to 0.32. To the west, the soil is much drier, with values of mainly 0.16 to 0.24 and a small patch in northern Colorado of 0.10 to 0.14. There is not much change between this time and 0000UTC 19 June 2002 (Figure 4.24). On 1800 UTC 19 June 2002 (Figure 4.25), the moisture values are once again very similar to the previous two
Figure 4.22 Spatial plots of 10 cm soil temperature (°C) valid 0000 UTC 20 June.
Figure 4.23 Spatial plots of 10 cm volumetric soil moisture from the Experimental Simulation valid 1800 UTC 18 June.
Figure 4.24 Spatial plots of 10 cm soil moisture from the Experimental Simulation valid 0000 UTC 19 June.
Figure 4.25 Spatial plots of 10 cm volumetric soil moisture from the Experimental Simulation valid 1800 UTC 19 June.
Figure 4.26 Spatial plots of 10 cm volumetric soil moisture from the Experimental Simulation valid 0000 UTC 20 June.
times. The main exceptions are less moisture in southeastern Nebraska and the Texas panhandle regions, and a small swath of higher moisture in eastern Colorado. By 0000 UTC 20 June 2002 (Figure 4.26), the main difference is the swath of higher moisture in Colorado has spread to central Nebraska, where there is a patch of high moisture ranging from 0.32 to 0.42.

Time series of 10cm soil temperature and soil moisture are shown in Figures 4.27, 4.28 and 4.29. Once again, the NCAR stations chosen for comparisons are Elmwood, Oklahoma; Spivey, Kansas and Atlanta, Kansas. The soil temperature time series feature comparisons between the Control Simulation, Experimental Simulation, and Observations. However, as has been previously mentioned, soil moisture output was not available for the Control Simulation, therefore those time series are comparisons between the Experimental Simulation and Observations.

Overall, the Experimental Simulation performed better for the 10cm soil temperatures. For Elmwood (Figure 4.27A), the Control Simulation overestimated the soil temperatures during the day and underestimated them at night, while the Experimental Simulation overestimated them during the first 18 hours, and then underestimated them throughout the remainder of the simulation. The observations were in between the Control and Experimental simulations over the first 12 hours. The Control Simulation was closer to observations during the second 12 hours, then during the third 12 hours both simulations were very similar, with neither being closer to the observations. For the remaining 36 hours, the Experimental Simulation was closer to the observations. Similar results can be seen for Spivey (Figure 4.28A). Once again, the Control Simulation overestimated the soil temperatures during the day and
underestimated them at night, while the Experimental Simulation overestimated temperatures for the first 18 hours and underestimated them throughout most of the remaining 54 hours, with the exception of a slight overestimation during part of 19 June. During the first day of the simulation, the Control Simulation produced closer soil temperature values to the observations. However, throughout the rest of the simulation, the Experimental Simulation was significantly closer to the observed temperatures. For Atlanta (Figure 4.29A), the Control Simulation once again overestimated temperatures during the day and overestimated them at night. The Experimental Simulation mainly underestimated temperatures, except for a short period towards the beginning of the simulation, and again near the end where this simulation slightly overestimated soil temperatures. Overall, the Experimental Simulation produced closer values of soil temperature to the observations than the Control, with the exception of a short time period during the first day and at the very end of the simulation.

As previously stated, 10cm soil moisture output was only available for the Experimental Simulation. For Elmwood (Figure 4.27B), the Experimental Simulation captured the soil moisture trend fairly well; however it significantly underestimated the values by as much as 13%. The model performed much better for Spivey (Figure 4.28B), where it captured the trend very well (although missed some of the peaks seen in the observations), and only slightly underestimated the moisture values throughout the simulation. On the other hand, the model did not do well at all for Atlanta (Figure 4.29B). The modeled and observed soil moisture almost matched exactly at the very beginning. However, for the remainder of the simulation, the Experimental Simulation
Figure 4.27 Time series of 10 cm soil temperature (A) and soil moisture (B) for Elmwood, OK.
Figure 4.28 Time series of 10 cm soil temperature (A) and soil moisture (B) for Spivey, KS.
Figure 4.29 Time series of 10 cm soil temperature (A) and soil moisture (B) for Atlanta, KS.
not only underestimated the moisture values, but it also produced an opposite trend than the observations, with decreasing model values while the observations were increasing.

A time series of RMS errors for the 10 cm soil temperature is shown in Figure 4.30. Just as for the surface heat fluxes, these errors were calculated using model output and observations for all nine NCAR surface flux stations. Overall, the Experimental Simulation has lower RMS errors than the Control, with values fluctuating between about 1.5°C and 6.5°C for the Control and between 2°C to 4.5°C for the Experimental Simulation. Towards the beginning of the simulations, there are two short periods where the Experimental RMS errors are higher than the Control, and at the end the errors are roughly the same, but for the rest of the simulation, the Experimental Simulation produced lower errors for the rest of the simulation.

Spatial plots of planetary boundary layer (PBL) height are shown in Figures 4.31 through 4.34. As before, A is from the Control Simulation and B is from the Experimental Simulation. At 1800 UTC 18 June 2002 (Figure 4.31), the PBL heights in the Experimental Simulation (Figure 4.31B) were overall higher than in the Control Simulation (Figure 4.31A). This is consistent throughout the simulation. Over much of the domain at this time, the Control simulated PBL heights range from 500 to 1000 m. This is mainly over Nebraska, western Iowa, most of Kansas, eastern Colorado, western Oklahoma and most of the Texas panhandle. Over southeastern Kansas, most of Oklahoma and the northwestern part of the Texas panhandle, the PBL heights range from 1000 to 1500 m. The highest PBL heights are seen over parts of central and southern Colorado, as well as southeastern Wyoming, where they range from 1500 to 3500 m,
Figure 4.30 Time series of RMS errors of 10 cm soil temperature for all nine NCAR ISFF stations.
Figure 4.31 Spatial plots of planetary boundary layer heights valid 1800 UTC 18 June.
while the lowest heights of less than 50 m are located in eastern Nebraska and
northwestern Missouri. The Experimental Simulation produced a larger variation in PBL
heights across the domain at this time. The lowest heights, ranging from 500 to 1000 m,
are located in eastern Nebraska into eastern Kansas, northeastern Oklahoma, western
Missouri and Arkansas, as well as most of the Texas panhandle. Across most of
Oklahoma and western Kansas, and also Iowa and central Missouri, PBL heights are
1000 to 1750 m. The highest PBL heights are found over northern New Mexico, much of
Colorado northwestern Kansas, western Nebraska and southeastern Wyoming, where the
range is from 1750 to greater than 3500 m.

By 0000 UTC 19 June 2002, there appears to be an even larger difference in PBL
heights between the two simulations (Figure 4.32). Over most of the domain in the
Control Simulation (Figure 4.32A), the model simulated PBL heights of less than 50 m.
If observations were available for comparison, this simulation would likely not match
well with observed PBL heights, considering around 0000 UTC (6:00pm local time), the
PBL should be at its maximum height. Over most of Texas and New Mexico, and small
regions of Oklahoma, the PBL height ranges from 100 to 1000 m. There is a small area
in New Mexico where the PBL height is 1250 to 1750 m. Most of the northwestern
portion of the domain (Colorado, western Nebraska and southeastern Wyoming), the
model simulated PBL heights of 50 to 1250 m, with small regions in Nebraska of 1750 to
2000 m, and a few pockets in Colorado and Wyoming of 2000 to 3000 m. In
comparison, the Experimental simulated PBL heights (Figure 4.32B) are significantly
higher. Across the northern half of the domain, the heights are lowest in the east (500 to
1000 m across Missouri and eastern Kansas), becoming increasingly higher to the west.
Figure 4.32 Spatial plots of planetary boundary layer heights valid 0000 UTC 19 June.
Across most of Kansas, eastern Nebraska and Iowa, PBL heights are 1000 to 1500 m. There is a very high gradient of PBL heights in central Nebraska, where they increase from 800 to 2000 m over a very short distance. The highest heights are once again over the northwestern portion of the domain (Colorado, northwestern Kansas, western Nebraska and southeastern Wyoming), where PBL heights reach upwards of greater than 3500 m. Across the southern half of the domain, there are lower PBL heights, with the lowest heights of 100 to 1000 m in Arkansas and eastern Oklahoma, as well as the Texas panhandle. Over most of Oklahoma, the heights reach 1000 to 1500 m, and are the highest in Wyoming, where the range is 1000 to 3500 m.

At 1800 UTC 19 June, the Control simulated PBL heights (Figure 4.33A) are fairly low across the central and most of the northern part of the domain, ranging from 100 to 900 m, with the lowest heights in northwestern Kansas and the Texas panhandle. The southeastern portion of the domain has PBL heights of 1000 to 1500 m. The maximum heights reach upwards of 1250 to 2500 m in very small patches in Colorado. In the Experimental Simulation (Figure 4.33B), most of simulated PBL heights are 1000 to 1500 m. The lowest heights are located in the northwestern part of the domain, as well as a few small regions in Oklahoma and Missouri, ranging from 500 to 900 m. In southern Colorado into New Mexico, the maximum PBL heights are seen, ranging from 1750 to greater than 3500 m.

Similar to the previous day, the Control Simulation produced very low PBL heights of less than 50 m over most of the domain at 0000 UTC 20 June (Figure 4.34A). Across the western part of the domain, the heights range from 50 to 1750 m, with the lowest heights being in Nebraska and Texas, and the highest in western Nebraska into
Figure 4.33 Spatial plots of planetary boundary layer heights valid 1800 UTC 19 June.
Wyoming. The Experimental Simulation once again produced higher PBL heights at this time (Figure 4.34B). Across the eastern half of the domain, the heights are mainly 1000 to 1750 m, with small areas in Arkansas and Oklahoma of 500 to 900 m. The maximum heights occur in a swath from central Nebraska spreading southwestward across Kansas, the Oklahoma panhandle and over most of New Mexico and southern Colorado, reaching from 1750 to 3500 m. Over northwestern Kansas and eastern Colorado, the lowest heights are seen, ranging from 400 to 900 m. There is also a large gradient in PBL heights in western Kansas, where the height changes very quickly from 400 to 2500 m.

10 m wind vectors and boundary layer mixing ratio is shown in Figures 4.35 through 4.38. At 1800 UTC 18 June, the winds in the Control Simulation (Figure 4.35A) are mainly southerly. Weak easterly winds are present over eastern Texas the southeastern Oklahoma, turning southeasterly toward the west and southerly toward the north. Over northern New Mexico and southern Colorado into southwestern Kansas, winds are southwesterly. The northern half of Colorado has very weak 10 m winds, while the winds in southwestern Wyoming and western Nebraska are stronger and from the west. The strongest winds are over Kansas and Nebraska, as well as the northeastern Texas panhandle. Weak convergence in the wind field can be seen in western Kansas and also a little in western and southern Nebraska. The model did not simulate much moisture at this time, with small pockets of 7 to 8 g kg⁻¹ in north central Kansas, eastern Nebraska and southern Colorado. In contrast, the winds in the Experimental Simulation at this time (Figure 4.35B) exhibit more directional variation. Over eastern Texas and southeastern Oklahoma, there are weak easterly winds, turning southeasterly toward the
Figure 4.34 Spatial plots of planetary boundary layer heights valid 0000 UTC 20 June.
Figure 4.35 Spatial plots of 10 m wind vectors (m s$^{-1}$) and mixing ratio (g kg$^{-1}$) valid 1800 UTC 18 June.
west and north, finally becoming southerly over north central Oklahoma, eastern Kansas into eastern Nebraska, and the central Texas panhandle. Winds over northwestern New Mexico and the northern Texas panhandle into northwestern Oklahoma and most of western Kansas are southwesterly and also appear fairly strong. Very weak winds are present over most of Colorado and northwestern Kansas. Over Wyoming into the Nebraska panhandle, winds are westerly, turning northeasterly over western Nebraska and northeastern Colorado, and then turning to westerly and southwesterly over central Nebraska. There is more evidence of convergence in the Experimental Simulation, especially in western and central Kansas and Nebraska. The mixing ratios in this simulation are higher than in the Control Simulation, with a large region of 6 to 10 g kg$^{-1}$ in eastern Kansas and southeastern Nebraska, as well as a smaller region in the Texas panhandle.

By 0000 UTC 19 June, the winds in the Control Simulation (Figure 4.36A) are primarily southeasterly to southerly. Winds across the southern part of the domain, including northern Texas, most of Oklahoma, the southern Texas panhandle and southern New Mexico are mainly southeasterly, with weaker winds to the east and stronger winds to the west. To the north of these regions, winds turn to become southerly, especially in central Kansas into Nebraska, the northern Texas panhandle, and northern New Mexico. Over Colorado, winds are southwesterly in the south, turning southerly and then southeasterly and back to southerly northward. Main areas of convergence in this simulation are located in central to northern Colorado, western Kansas and western Nebraska. The moisture at this time is primarily located in Colorado, New Mexico and western Nebraska, ranging from 6 to 9 g kg$^{-1}$. In the Experimental Simulation (Figure
Figure 4.36 Spatial plots of 10 m wind vectors (m s$^{-1}$) and mixing ratio (g kg$^{-1}$) valid 0000 UTC 19 June.
the winds are also primarily south/southeasterly, and are also overall weaker in several regions than the Control Simulation. When compared to the Control Simulation, weaker winds are seen in southern Oklahoma, eastern Texas, Louisiana, northeastern Colorado, northwestern Kansas and western Nebraska. Across northern New Mexico, the northern Texas panhandle, and most of Kansas and eastern Nebraska, strong south to southeasterly winds are present. Areas of convergence in this simulation appear to be located in northwestern Kansas, southern Colorado and central Nebraska. As far as boundary layer moisture is concerned, the Experimental Simulation once again produced larger areas of higher moisture than the Control. There is a large region of 6 to 10 g kg$^{-1}$ mixing ratios across eastern Kansas into northwestern Missouri and southwestern Iowa, as well as another large swath from south central Nebraska extending southwestward across western Kansas and southeastern Colorado into northwestern New Mexico.

At 1800 UTC 19 June, the winds in the Control Simulation are once again primarily from the south/southeast (Figure 4.37A). The main areas of interest at this time are in Colorado and Nebraska, where there is the greatest evidence of convergence. Winds in the Nebraska panhandle into northern Colorado are from the north/northeast, turning to southerly winds in central to southern Colorado and southwesterly in western to central Nebraska. The changes in wind direction in these regions clearly denote a boundary of some sort accompanied by convergence in the wind field. There is also very little significant moisture at this time, with only a small patch of 6 to 8 g kg$^{-1}$ mixing ratios in central Colorado. In the Experimental Simulation at this time (Figure 4.37B), there is a circulation present in the wind field in northwestern Kansas/southwestern Nebraska. The winds in Colorado and the panhandle of Nebraska are northerly, while
Figure 4.37 Spatial plots of 10 m wind vectors (m s\(^{-1}\)) and mixing ratio (g kg\(^{-1}\)) valid 1800 UTC 19 June.
winds in Kansas are southwesterly, turning easterly into southwestern Nebraska. Convergence can be seen in the vicinity of this circulation, extending into southeastern Colorado as well as western Nebraska. Also associated with this circulation is an area of higher mixing ratios, ranging from 6 to 12 g kg\(^{-1}\).

By 0000 UTC 20 June, the Control simulated northerly winds in Wyoming, Nebraska and Colorado have shifted to northeasterly, with more southerly winds spreading further north in Colorado. There is once again clear evidence of convergence in western Nebraska, where the winds shift from southerly to northerly, extending into northern Colorado, where there is a shift from southerly to northeasterly winds. Along this boundary there are areas of high moisture, with mixing ratios ranging from 6 to 12.5 g kg\(^{-1}\). In New Mexico, there are also two areas of higher moisture, with mixing ratios of 6 to 11 g kg\(^{-1}\). In the Experimental Simulation at this time (Figure 4.38B), the circulation seen in the previous figure has moved into central Nebraska. The boundary associated with this circulation has become stronger by this time, with convergence extending from southwestern Nebraska southwestward across western Kansas and finally westward across southern Colorado. The winds along this boundary shift from north/northwesterly to south/southeasterly in Kansas and from east/northeasterly to south/southeasterly in Colorado. In the area of the circulation, as well as along the boundary and across eastern New Mexico, there is a large swath of high mixing ratios ranging from 6 to 11 g kg\(^{-1}\).

Boundary layer vertical velocities (m s\(^{-1}\)) and relative humidity (% contoured) are shown in Figures 4.39 through 4.42. At 1800 UTC 18 June (Figure 4.38), the Control Simulation produced the highest upward vertical velocities over the northern half of the domain (Figure 4.39A). Vertical velocities in this simulation range from 0.01 to 0.7 m s\(^{-1}\).
Figure 4.38 Spatial plots of 10 m wind vectors (m s$^{-1}$) and mixing ratio (g kg$^{-1}$) valid 0000 UTC 20 June.
over central and western Kansas, eastern and southern Colorado and most of Nebraska. Relative humidity (RH) values are fairly low over most of the domain, mainly ranging from 10 to 40%. The highest RH values of 50 to 60% are found in north central Kansas through eastern Nebraska into western Iowa. In contrast, the Experimental simulated RH values are significantly higher, especially over eastern Nebraska and Kansas into northern Oklahoma (Figure 4.39B). The RH values over this area range from 40 to as high as 90%, most likely providing sufficient moisture to support convection. A secondary area of higher RH values is located in the Texas panhandle, where the range is 50 to 70%.

Enhanced vertical velocities are associated with and behind the highest area of RH, in southeastern Nebraska and central Kansas. Velocities range from 0.01 to 0.8 m s$^{-1}$ in this region. The model also simulated regions of upward motion over central Nebraska into northwestern Kansas and eastern Colorado, portions of western and central Oklahoma, and parts of Texas and Missouri.

By 0000 UTC 19 June, Control simulated vertical velocities are higher over the northeastern portion of the domain than in the previous time (Figure 4.40A). This area of elevated vertical motion of 0.01 to 0.5 m s$^{-1}$ is associated with the highest RH values of 50 to 60% over western Nebraska and eastern Iowa. Other regions of enhanced vertical motion are seen over Colorado and western Nebraska. The relative humidity over most of the domain in this simulation is low, ranging from 10 to 40%. In the Experimental Simulation (Figure 4.40B), the region of highest RH values has shifted slightly eastward to over eastern Kansas, extending northward into Iowa and southward into Oklahoma. The elevated vertical velocities associated with this region have also spread eastward along with the higher moisture. There is also a swath of both higher RH and vertical velocities.
Figure 4.39 Spatial plots of vertical velocity (m s$^{-1}$) and relative humidity (% contoured) valid 1800 UTC 18 June.
Figure 4.40 Spatial plots of vertical velocity (m s$^{-1}$) and relative humidity (% contoured) valid 0000 UTC 19 June.
velocities extending from central Nebraska across western Kansas into southeastern Colorado. The RH in this region is about 50 to 60% and the vertical velocity tops out around 0.5 m s\(^{-1}\). Both simulations produced higher vertical motion in central to eastern Kansas, western and central Nebraska, as well as Colorado.

The Control simulated RH at 1800 UTC 19 June are low over the entire domain, ranging mostly from 10 to 30%, with a small area of 40 to 50% over central Colorado (Figure 4.41A). The highest vertical motions at this time are found over Colorado, western Nebraska, eastern Nebraska into western Iowa, and northern Kansas, where the range is 0.01 to about 0.5 m s\(^{-1}\). By sharp contrast, the RH is once again significantly higher in the Experimental Simulation (Figure 4.41B). The region of high RH values over eastern Kansas has moved eastward to over Missouri and has also become lower, ranging now from 40 to 70%. There is very little vertical motion associated with this region at this time. The maximum RH values of up to 90% are found in southwestern Nebraska, associated with the circulation in this area. Accompanied by this high moisture are the highest vertical velocities, reaching as high as 2 m s\(^{-1}\). Regions of vertical motion are also over Colorado, central Kansas, northwestern Oklahoma and much of Nebraska.

By 0000 UTC 20 June, higher RH values have advected into the western portion of the domain in the Control Simulation (Figure 4.42A). High RH values of 70 to 90% are seen over northern New Mexico, northern Colorado and western Nebraska. The highest vertical motions are also over the western portion of the domain, as well as across most of Nebraska, showing mainly the same range as has been seen in previous times in the Control Simulation. As the circulation produced by the Experimental Simulation has
Figure 4.41 Spatial plots of vertical velocity (m s$^{-1}$) and relative humidity (% contoured) valid 1800 UTC 19 June.
Figure 4.42 Spatial plots of vertical velocity (m s\(^{-1}\)) and relative humidity (% contoured) valid 0000 UTC 20 June.
propagated northeastward, the region of maximum RH and vertical velocity has moved with it (Figure 4.42B). The RH around this circulation is still 70 to 90%, with the vertical velocity being greater than 2 m s$^{-1}$. Enhanced RH values can also be seen across northwestern Kansas into the Oklahoma and Texas panhandles and eastern New Mexico. There are also other regions of elevated vertical motion over central Kansas, eastern Kansas into Missouri, northwestern Kansas into Colorado, Oklahoma and New Mexico. The Experimental Simulation overall produced a larger variation of relative humidity over the entire domain, as well as overall greater vertical motion.

Model simulated zonal cross sections of potential temperature contours (isentropes), vertical velocity (shaded areas) and horizontal wind vectors shown in Figures 4.43 through 4.46. These cross sections are located at three latitudes: 39.1°N (A to A'), 40.0°N (B to B'), and 41.3°N (C to C'), with a west to east extent of 102°W to 93.5°W. Refer back to Figure 4.13 for the locations of each of these three cross sections. At 1800 UTC 18 June, the upper-level northwest flow can clearly be seen in cross section A to A', located at 39.1°N (Figure 4.43). In the Control Simulation at this time (Figure 4.43A), the highest vertical motions are seen around 99°W longitude (north central Kansas). The vertical motion starts at the surface at around 0.01 m s$^{-1}$ and reaches a maximum between 750 and 700 mb of 0.1 m s$^{-1}$. Another region of higher surface vertical motion, ranging from about 0.01 to 0.06 m s$^{-1}$, is between 102°W and 101°W (northwestern Kansas), which extends throughout the entire atmospheric column in this region. This is concurrent with the boundary seen in the surface winds in Figure 4.35A. Vertical motion can be seen aloft (~450 to 400 mb) around 99°W and 94.5°W. Between 97°W and 94.5°W are areas of smaller surface vertical motions. Surface winds are
southerly, veering around to the southwest to west in the mid levels (750 to 600 mb), finally becoming northwest aloft (550 to 300 mb). This indicates warm-air advection in this region. The strongest winds are located above the boundary layer and also in the upper levels (500 to 300 mb). The isentropes over northwestern Kansas are spread far apart in the mid levels, indicating instability in this region. Closer to the surface there is more stability. Also over central to eastern Kansas, the atmosphere is moderately unstable, especially in the mid levels.

The same cross section (39.1°N) from the Experimental Simulation is shown in Figure 4.43B. Surface vertical velocities are once again fairly weak in this simulation, with the strongest area of near-surface vertical motion being 0.01 to 0.06 m s⁻¹ around 97.8°W (central to eastern Kansas). There appears to be weak surface convergence in this region in Figure 4.35B. This area of vertical motion extends vertically from the surface through the entire atmospheric column, tilting back to the west to around 100°W (northwestern Kansas), with velocities reaching a maximum of 0.08 m s⁻¹ between 800 and 550 mb and also between 350 and 300 mb. Associated with the boundary in northwestern Kansas (around 100°W) is another area of weak vertical motion of around 0.01 m s⁻¹, reading 0.06 m s⁻¹ around 800 mb. A region of similar vertical velocities is seen between 96°W and 95°W, with another weaker region located to the east. In the upper levels (500 to 300 mb), there are stronger vertical motions (0.2 to 1.0 m s⁻¹) located at about 96.5°W. Weak vertical velocities aloft surround this particular region. In the mid-levels (700 to 550 mb), an area of moderate vertical velocities of 0.1 to 0.4 m s⁻¹ can be seen around 97°W. Just as in the Control Simulation at this time, the horizontal winds veer from south-southwest at the surface to westerly in the mid-levels to northwesterly
Figure 4.43 Zonal cross section at location A to A’ of vertical velocity (m s\(^{-1}\)), horizontal winds (m s\(^{-1}\)) and potential temperature (K) valid 1800 UTC 18 June.
aloft, showing warm air advection over the region. The largest area of instability
(indicated by the spread of the isentropes) is located over the western half of the region,
around 800 to 600 mb, with moderate to low instability further east. The downward
slope of the isentropes between 102ºW and 98ºW seem to indicate the boundary located
in northwestern Kansas.

In Figure 4.44, cross sections across the same latitude (39.1ºN) at 0000 UTC 19
June are shown. In the Control Simulation (Figure 4.44A), there is an area of vertical
motion of 0.03 to 0.06 m s⁻¹ around 102ºW, corresponding with an area of weak
convergence in eastern Colorado seen in Figure 4.36A. Also at the same longitude,
stronger vertical velocities reaching 0.1 m s⁻¹ are located between 400 and 300 mb. A
large area of vertical motion is located between 99.5ºW to 95ºW, which does not appear
to be associated with any visible convergence zones in Figure 4.36A. These vertical
motions are strongest in the mid-levels (850 to 550 mb), reaching from 0.03 to 0.08 m s⁻¹.
The wind pattern from the surface to the upper levels is essentially the same as in the
previous time, veering from the south-southwest at the surface to northwest aloft.
Instability at this time is weakest at the surface (where the isentropes are closer together),
and stronger in the mid-levels (especially 750 to 600 mb). The western area of the region
is more unstable in the mid-levels than it is to the east.

There is higher vertical motion in the Experimental Simulation at this time
(Figure 4.44B) associated with the boundary in northwestern Kansas in Figure 4.36B.
Here, the vertical velocities range from 0.1 m s⁻¹ at the surface to 0.3 m s⁻¹ between 825
and 650 mb. Also along the boundary slightly to the east are weaker surface vertical
motions of 0.01 to 0.06 m s⁻¹. Between 98ºW and 97ºW is an area of weak vertical
Figure 4.44 Zonal cross section at location A to A’ of vertical velocity (m s⁻¹), horizontal winds (m s⁻¹) and potential temperature (K) valid 0000 UTC 19 June.
motion possibly associated with some weak convergence in northern Kansas in Figure 4.36B. The same vertical motions can be seen in the mid-levels around the same area and to the east. Around 95°W, there are stronger vertical velocities of 0.1 to 0.3 m s⁻¹ between 500 and 300 mb, with some weaker vertical motion to the east. The wind pattern at this time is similar to the previous time, however now the stronger northwest winds extend further down to the middle levels. Also, the mid-level winds to the east are stronger at this time than before. The isentropic pattern is also similar to the previous time period, with the maximum instability once again located to the west.

The 1800 UTC cross section on 19 June (Figure 4.45) is located at 40°N (B to B' on Figure 4.13). In the Control Simulation (Figure 4.45A), vertical motion can be seen around 105W associated with the boundary in northeastern Colorado seen in Figure 4.37A. The vertical velocities in this area are weak, ranging from 0.01 m s⁻¹ at the surface to 0.04 m s⁻¹ around 750 mb. Vertical velocities of 0.01 m s⁻¹ extend up to 550 mb. A few patches of weak vertical motion at the surface can be seen to the east. Across the eastern two-thirds of the cross section, there are weak vertical velocities extending mainly from 650 to 350 mb. The wind pattern at this location shows almost no variation in direction from the surface to the upper levels, with winds from the southwest to west. The stronger winds are located just above the boundary layer and are weaker aloft. As before, the strongest instability is located mainly to the west and in the mid levels, with the surface being much more stable.

In Figure 4.37B, a circulation is located along the western Kansas/Nebraska border. The Experimental Simulation produced surface vertical motion associated with this circulation in Figure 4.45B. The velocities at this location only reach 0.06 m s⁻¹,
Figure 4.45 Zonal cross section at location B to B' of vertical velocity (m s$^{-1}$), horizontal winds (m s$^{-1}$) and potential temperature (K) valid 1800 UTC 19 June.
however there is a region of much higher vertical velocities directly to the west. This region extends all the way from the surface to 300 mb, ranging from 0.5 m s⁻¹ at the surface to greater than 2.5 m s⁻¹ between 500 and 300 mb. This is possibly due to effects from the Rocky Mountains. Just as in the Control Simulation, there are a few patches of weak vertical motion at the surface to the east. Around 99.5W, the weak vertical motion extends from the surface to 300 mb. A secondary area of higher vertical velocities (0.01 to 0.5 m s⁻¹) is located at 96W, between 550 and 300 mb. The wind pattern at this time is similar to the Control, except that the winds are more southerly in the mid to upper levels. There is also once again higher instability in the mid levels to the west.

Cross sections located at 41.3N (C to C' on Figure 4.13) at 0000 UTC 20 June are shown in Figure 4.46. The Control Simulation produced weak vertical motions across nearly the entire cross section at this time. The strongest surface vertical velocities are around 0.03 to 0.08 m s⁻¹ and are associated with the boundary in western Nebraska seen in Figure 4.38A. The rest of the surface velocities are weak, around 0.01 m s⁻¹. Towards the east, there are some stronger vertical motions in the middle levels. The wind pattern is much the same as the previous time, with winds from the southwest to west from the surface to 300 mb.

The Experimental Simulation produced higher vertical motions ranging from 0.01 to 0.1 m s⁻¹ (Figure 4.46B), associated once again with the circulation that has moved into Nebraska and the boundary in western Nebraska seen in Figure 4.38B. These vertical motions extend all the way up to 300 mb between 100W and 98W. The stronger vertical motions located to the west are still present at this time, however they are weaker than the previous time period. Weaker vertical motions are located to the east, with
Figure 4.46 Zonal cross section at location C to C' of vertical velocity (m s\(^{-1}\)), horizontal winds (m s\(^{-1}\)) and potential temperature (K) valid 0000 UTC 20 June.
velocities of mainly 0.01 m s$^{-1}$. The winds in the Experimental Simulation are mainly southwesterly to southerly throughout the entire cross section.

Theta profiles at three sounding sites are shown in Figures 4.47 through 4.52. These sites are North Platte, Nebraska; Dodge City, Kansas and Amarillo, Texas. Figures 4.47, 4.48 and 4.49 are at 0000 UTC 19 June (6:00pm LT). For North Platte (Figure 4.47), both the Control and Experimental Simulations were fairly close to each other. They came fairly close to the observed theta profile in terms of temperature, however the structure did not entirely match the observations. The observed PBL height at this location was around 850 mb, however the modeled PBL heights are difficult to discern from these profiles. At Dodge City (Figure 4.48), the modeled profiles are once again close to each other. They start out cooler than the observations, then around 800 mb the temperatures match fairly well. The main difference between the modeled and observed profiles is once again the PBL height. The observed height is around 800 mb, while both of the simulated PBL heights are around 950 mb. Both the Control and Experimental Simulations also under predicted the PBL height at this time at Amarillo (Figure 4.49). The observed height is around 850 mb, while the simulated heights are both around 800 mb. The model once again starts out cooler than observations, then converges towards observations above the boundary layer.

Figures 4.50, 4.51 and 4.52 are at the same locations at 0000 UTC 20 June. At North Platte (Figure 4.50), the observed PBL height is around 800 mb. The Control simulated PBL height is around 950 mb, while the Experimental height is a little closer to 800 mb. The two simulations start out spread apart, and then converge towards each other and towards observations above the boundary layer, while once again diverging.
Figure 4.47 Theta profile at North Platte, NE valid 0000 UTC 19 June 2002.
Figure 4.48 Theta profile at Dodge City, KS valid 0000 UTC 19 June 2002.
Figure 4.49 Theta profile at Amarillo, TX valid 0000 UTC 19 June 2002.
Figure 4.50 Theta profile at North Platte, NE valid 0000 UTC 20 June 2002.
Figure 4.51 Theta profile at Dodge City, KS valid 0000 UTC 20 June 2002.
Figure 4.52 Theta profile at Amarillo, TX valid 0000 UTC 20 June 2002.
aloft. For Dodge City (Figure 4.51), both simulations are once again close to each other, but neither converges towards observations until the upper levels. The observed PBL height is around 925 mb, while the modeled PBL heights are around 950 mb. For the final profile at Amarillo (Figure 4.52), the model once again starts off cooler than observations, then converges towards the observed profile above the boundary layer. The observed PBL height is high, reaching 700 mb. The simulated PBL heights, however, are much lower, reaching 800 mb. Overall, it appears that neither simulation performed better for these theta profiles.

4.4 Summary

There is overall a significant difference between the Control (without FASDAS) and Experimental (with FASDAS) simulations. The Experimental Simulation generated significantly more precipitation over the 72-hour integration, while coming a little closer to the observations than the Control Simulation, which hardly produced any precipitation. Surface heat fluxes seemed more uniform for the Control Simulation, while exhibiting larger gradients in the Experimental Simulation. These gradients possibly helped fuel and enhance the convection that occurred in the Experimental Simulation. The RMS errors over the nine NCAR stations for both the surface sensible and latent heat fluxes as well as the soil temperatures were lower for the Experimental Simulation than for the Control. The PBL heights were overall much higher in the Experimental Simulation than for the Control (in the spatial plots – they were mainly the same for the three sounding locations). Surface boundaries and associated convergence was much more defined in the Experimental Simulation, as were vertical motions associated with these boundaries.
This enhanced convergence, along with higher moisture values also seen in this simulation, supported the greater amounts of precipitation that were produced by the Experimental Simulation. While the Control Simulation did produce some convergence and vertical motions, very little precipitation was generated during this simulation.
5.1 Introduction

Interactions between the surface of the Earth and the atmosphere are vital to the development of deep moist convection.

According to Pielke (2001), the effects of the land-surface on the development of convection stem from the surface energy and moisture budgets. These can be written as

\[ R_N = Q_G + H + L(E + T) \] \hfill (5.1)

\[ P = E + T + RO + I \] \hfill (5.2)

where the net radiative fluxes \( R_N = Q_s (1 - A) + Q_{LW}^\uparrow - Q_{LW}^\downarrow \), \( P \) is precipitation, \( E \) is evaporation, \( T \) is transpiration, \( Q_G \) is the soil heat flux, \( H \) is the turbulent sensible heat flux, \( L(E + T) \) is the turbulent latent heat flux, \( L \) is the latent heat flux of vaporization, \( RO \) is runoff, \( I \) is infiltration, \( Q_s \) is insolation, \( A \) is albedo, \( Q_{LW}^\downarrow \) is downwelling longwave radiation, upwelling longwave radiation \( Q_{LW}^\uparrow = (1 - \varepsilon)Q_{LW}^\downarrow + \varepsilon \sigma T_s^4 \), \( \varepsilon \) is the surface emissivity, and \( T_s \) is the surface temperature. Land use change can lead to changes in one or more of these variables, thus directly affecting the fluxes of heat, moisture and momentum in the boundary layer. This can result in changes in surface convergence, the vertical structure of the atmosphere, local wind circulations and vertical motions, all of which have an impact on the development of deep, moist convection.

Since the surface plays such an important role in thunderstorm development, it is important to be able to simulate land-surface processes, and subsequently land-atmosphere interactions using mesoscale models. For example, past simulations...
conducted without soil moisture gradients did not result in the formation of a dryline, unlike simulations where these gradients were present (Ziegler et al. 1995; Shaw et al. 1997; Grasso 2000). Land surface models (LSMs) have evolved over time from a simple implicit bucket parameterization (Manabe 1969) to more complex, multi-soil layer models which better represent the processes by which the land-surface impacts the atmosphere (Pitman 2003). There are, however, still outstanding issues regarding the accuracy of modeling land surface processes. For example, efforts are currently underway to modify the LSM within MM5/WRF. The objective of this particular study is to test the sensitivity of land-atmosphere interactions in MM5 simulations of a convection event, focusing on whether or not there is a significant impact between using a simple or more complex LSM.

5.2 Case Overview

This study covers a 72-hour period beginning 0000 UTC 17 June and ending 0000 UTC 20 June 2002, during which convection occurred at various times along several surface boundaries. Surface as well as synoptic processes contributed to the formation of these boundaries throughout the study period. A more in-depth discussion of these processes and boundaries is provided in Chapter 3. 17 June began with weak northwest upper-level flow over the IHOP_2002 region, with a ridge axis oriented SW-NE across the central Rockies. The morning of 17 June was marked by clear conditions and strong southwest winds at the surface. An ENE-WSW oriented dryline propagated from Colorado into NW Kansas during the afternoon, pushed by a deepening surface low over central Colorado. Between 2000 and 2100 UTC, convection began to fire in northwest
Kansas and the Oklahoma panhandle along this dryline boundary. Convection continued to propagate east-southeast across Kansas overnight, supported by low-level jet dynamics in western Oklahoma and Kansas.

The region was still under weak northwest flow on 18 June, becoming more westerly overnight. During the afternoon, a dryline boundary, located near Garden City, Kansas began to drift towards the north-northwest. Around 2000 UTC, weak convection developed over the Front Range in southeast Colorado, moving into north central Kansas during the next few hours. Convection dissipated quickly at nightfall.

On 19 June, the upper level pattern had shifted to a more west-southwest flow. By late morning, a NNE-SSW surface boundary was located near Goodland, Kansas, which was observed propagating towards Colby, Kansas between 1740 and 1905 UTC. Convection quickly developed in SE Colorado and along the boundary in Kansas, propagating eastward overnight.

What is the effect of the land surface – atmospheric boundary layer feedback on the simulation of this convection event is the focus of our current study.

5.3 Methodology

This study was conducted using version 3.6.2 of the MM5 modeling system. Two simulations were performed, one using the Noah LSM coupled to the MRF PBL scheme, and the other using the Pleim-Xiu (PX) LSM coupled to the ACM PBL scheme. There are several important differences between the Noah and the PX schemes (Chen and Dudhia 2001; Ek et al. 2003; Xiu and Pleim 2001). First, Noah has four soil layers, located at depths of 10cm, 30cm, 60cm and 100cm, while PX is a simple two-layer
model, with depths of 1cm and 100cm. The additional soil layers in Noah are expected to provide a more vertically heterogeneous soil moisture and soil temperature structure across the domain. The top soil layer generally responds to sub-hourly variations and is expected to exhibit obvious differences between the two simulations. The second important difference is the PX scheme is implicitly coupled to an ACM PBL scheme, while the Noah has been coupled with the MRF PBL scheme for this study. The asymmetric convective model (ACM) is summarized by Alapaty et al. (1997). The ACM is a non-local closure scheme based on Blackadar (1979). The underlying assumption of the Blackadar scheme is symmetric turbulent mixing in the boundary layer. However, both observations and large-eddy simulation modeling have shown that in the convective boundary layer, there is asymmetric mixing (Schumann 1989). The ACM was therefore modified by Pleim and Chang (1992) to add asymmetrical vertical mixing. As such, this model is only valid for convective conditions, and should not be used for stable situations. The ACM is able to simulate rapid upward transport and slower downward subsidence. The sensible heat flux is used to estimate the mixing rates. Refer to Pleim and Chang (1992) for more details. The medium range forecast (MRF) PBL scheme, also referred to as the Hong-Pan PBL scheme (Hong and Pan 1996) is also a non-local closure scheme based on the representations of the countergradient term and K profile in the well-mixed PBL developed by Troen and Mahrt (1986). Vertical diffusion in the MRF makes use of a fully implicit scheme in order to allow longer time steps and calculate updates of prognostic variables. The PBL height is determined through the use of the Bulk-Richardson number. During unstable conditions, the surface sensible and latent heat fluxes are used to determine an enhancement to the virtual potential temperature,
which is in turn used to re-determine the PBL height. Refer to Hong and Pan (1996) for more details. These two schemes are expected to lead to differences in the manner in which surface fluxes are coupled to the boundary layer and cause impacts on the evolution of the boundary layer convection and the cloud/radiation in the model case study.

Beyond these two significant differences, there are five other contrasting model features may have subtle but sometimes significant impacts via feedbacks on the model results. These differences are, for instance, in the manner in which surface run-off and infiltration is parameterized. In Noah, the surface runoff and infiltration is more explicitly modeled as based on Schaake et al. (1996). This, for a precipitation case, may have implications in the manner of which soil moisture alterations occur as a response to the precipitation amount and distribution in the model. Additionally, in the Noah LSM, the soil thermal conductivity is modeled as a function of changes in soil moisture content, following Peters-Lidard et al. (1998). This allows for relatively higher thermal conductivity for drier soils, leading to higher soil heat fluxes and relatively damped soil 2-meter air temperature differences in the evolution and redistribution of net radiation between the ground, latent and sensible heat fluxes. The conductivity and ground heat flux is also modified as a function of the vegetation fraction, providing a shielding to the ground heat flux as a function of radiation. The PX scheme does not include these features, therefore it is expected that this will lead to additional differences in the surface temperature and soil moisture fields, which will in turn have a feedback on the model results. EDAS soil moisture and soil temperature fields are used to initialize the model. Due to the similarity between EDAS and Noah, it is expected that the use of EDAS will
be more adaptive to the Noah scheme and may introduce larger errors for the PX. Another important difference is the manner in which stomatal resistance, and subsequently its scaling to canopy resistance, is achieved. The PX scheme adapts a modification of the Ball-Berry stomatal model, which represents stomatal resistance as a function of relative humidity (Ball et al. 1987; Collatz et al. 1991). Noah uses a Jarvis (1976) type approach, which utilizes a minimum canopy resistance prescription. The Jarvis model has feedback for soil moisture and radiation changes, while the Ball-Berry has a humidity feedback. LAI is constant in Noah, while it varies in PX as a function of the vegetation type. Ultimately, the discrepancies between the Noah and PX schemes are expected to lead to different impacts on the way the model calculates transpiration and sensible and latent heat fluxes. Some cold-season processes, such as snow cover and snow albedo, also differ between the two schemes; however, these are irrelevant for this particular study.

The remaining model physics used in the model configuration were identical for each simulation, and included the Simple Ice scheme for the explicit moisture physics, the Kain-Fritsch 2 cumulus parameterization (for the 12km domain only), and Dudhia Cloud Radiation physics. Initial and lateral boundary conditions for both simulations were prescribed using NCEP Eta model analyses. The analysis valid 0000 UTC 17 June 2002 was used to provide the initial conditions for the model. Both simulations were integrated over a period of 72 hours, beginning 0000 UTC 17 June 2002 and ending 0000 UTC 20 June 2002. Standard MM5 data assimilation was used to incorporate the IHOP_2002 Hourly Surface Meteorological Composite dataset into the model. Refer to Chapter 2 for a more in-depth description of the physics used.
Figure 5.1 Illustration of land-surface processes.

Figure 5.2 Illustration of planetary boundary layer processes.
5.4 Results and Discussion

Accumulated precipitation from the first 24 hours (0000 UTC 17 June – 0000 UTC 18 June) of the period is shown in Figure 5.3. 24 hour accumulated precipitation from the Noah Simulation is shown in Figure 5.3B. The Noah-simulation over the first 24 hours captured one major area of precipitation, while the remainder lagged westward of the observations (Figure 5.3A). This simulation did not capture the majority of the precipitation in central Kansas, aside from one small region towards western Kansas. The precipitation in southwestern Kansas into the Oklahoma and Texas panhandles is present in this simulation, however it is further west and less intense than the observations. In eastern/northeastern New Mexico the precipitation is largely overestimated, covering almost that entire region. Other regions where the spatial coverage of precipitation was over-simulated include central Nebraska (where the precipitation is also less intense than the observations), south-central Colorado, and northern Arkansas (the latter two regions not receiving any accumulations in the observations).

24 hour accumulated precipitation from the Pleim-Xiu (PX) simulation is shown in Figure 5.3C. As far as spatial distribution of precipitation, the PX simulation is very similar to Noah at this time. The main difference between the two simulations lies in the amount of accumulated precipitation, which will be better examined by looking at the difference plot. The PX was generally similar to the Noah simulation. Once again, the PX simulation did not generate the precipitation in central Kansas, with the exception of one small region located in the same area as in the Noah simulation. The convection in southwestern Kansas into the Oklahoma and Texas panhandle regions is also in the PX
simulation, but it is once again shifted to the west and less intense than observations. The PX simulation, however, did produce greater accumulations of precipitation in this area, making it a little closer to the observations. In east-central Nebraska, this simulation produced a greater spatial coverage of precipitation than observations, which is also shifted further west and north. Once again, PX produced greater accumulations, which are closer to the observed accumulations over this time period.

In Figure 5.3D, the difference between the Noah and PX simulated 24 hour precipitation is shown. The PX simulation overall produced more precipitation than the Noah simulation over the first 24 hours. The differences are mostly fairly small, with PX producing accumulations of 0.5 cm to 1.0 cm greater than Noah. However they are larger in a few regions, including southwest Kansas, eastern Nebraska, east-central New Mexico, western Texas panhandle, and central Arkansas, where the PX precipitation is around 2.0 cm to 2.5 cm greater than Noah. Noah produced greater precipitation in some very small regions, such as in east-central and northeastern New Mexico, the northwest corner of the Texas panhandle, extreme southwestern Kansas and central Nebraska, where the difference is about 0.5 cm.
Figure 5.3 Accumulated precipitation (cm) valid 0000 UTC 17 June through 0000 UTC 18 June. MPE data is shown in A, the Noah Simulation in B, the PX Simulation in C and the difference between the Noah and PX simulations in D.
In Figure 5.4, 24-48 hour accumulated precipitation is shown. While the Noah Simulation (Figure 5.4B) did produce precipitation in central Kansas, southwestern Iowa, northwestern Missouri, and northeastern New Mexico during the second 24 hours of the simulation, it over-predicted the precipitation over the entire domain when compared to the observations (shown in Figure 5.4A). Accumulations in the Noah simulation are spread over much of western, central and northeastern Kansas, central into southeastern Nebraska, southwestern Iowa into northwestern Missouri, and eastern New Mexico, with small regions in the Texas panhandle and central Arkansas. The precipitation in west-central Missouri is not present in this simulation; rather it once again appears to have been shifted to the west. The greatest accumulations are in north central and south-central Kansas, with total amounts reaching from 2.5cm to 4.75cm. Accumulations elsewhere mostly range from 0.5cm to 1.75cm, with greater amounts of around 2.25 cm to 3.0 cm in central Nebraska.

The PX simulated accumulated precipitation over the second 24 hours of the simulation is shown in Figure 5.4C. The PX simulation once again generated precipitation in mainly the same areas as the Noah simulation. The largest differences are again in the amounts of precipitation produced, which will be examined later in the difference plot. The precipitation in the PX simulation is more widespread across central and southeastern Kansas, eastern Nebraska, northwestern Missouri and the Oklahoma panhandle. The greatest accumulations are in south-central Kansas, with a small area reaching upwards of 6.5 cm. However, just as in the Noah simulation, the PX does not match well with the observations, over-predicting precipitation over the entire domain.
Figure 5.4 Same as Figure 5.2, except valid from 0000 UTC 18 June through 0000 UTC 19 June.
The difference between the Noah and PX simulated precipitation over the second 24-hour period is shown in Figure 5.4D. Overall, it appears that the PX simulation produced greater amounts of precipitation over the second 24 hours of the simulation. This is especially evident in central to south-central and southeastern Kansas, northwestern Missouri, northwestern Kansas, southwestern Kansas into the Oklahoma panhandle, central Arkansas, and southeastern New Mexico into western Texas. The differences in these areas range from 0.5 cm to as high as 2.5 cm. In small regions such as north central and just west of central Kansas, southern Nebraska, the central Texas panhandle and eastern New Mexico, the Noah simulation generated more precipitation, with accumulations ranging from larger areas of 0.5 cm to tiny areas of 1.75 cm greater than the PX simulation.

Accumulated precipitation during the final 24 hours of the simulation is shown in Figure 5.5. During this final 24-hour period, The Noah simulation, shown in Figure 5.5B, once again produced a greater spatial coverage of precipitation than the observations (Figure 5.5A). This simulation did capture the convection in Nebraska (which was shifted to the north and west by the model), western Kansas (albeit the extent was not as far east as the observations), southeastern New Mexico into western Texas, Louisiana, and a little bit of eastern Colorado. The model, however, was not as intense as the observations, and also generated precipitation over Iowa and northern Missouri, as well as eastern New Mexico into the western Texas panhandle, spreading northward into the Oklahoma panhandle. This simulation did not produce most of the precipitation in central and northwestern Colorado. The greatest accumulations in the Noah simulation are located in central Nebraska and parts of the northwestern Texas panhandle, where
amounts range from 2.5 cm to upwards of 5.0 cm. The rest of the precipitation generated mostly ranges from 0.5 to 2.0 cm.

During the final 24 hours, the PX simulation again produced basically the same spatial coverage as the Noah simulation, as shown in Figure 5.5C, although PX also generated precipitation in eastern Texas, along with the northeastern Texas panhandle into the eastern part of the Oklahoma panhandle. The latter is not present in the Noah simulation. The intense precipitation in south-central Nebraska in the observations is present in the PX simulation; however the model has it shifted to the northwest. Also, this area of accumulated precipitation is much higher than in the Noah simulation, with a small area greater than 8.5 cm in central Nebraska. A little bit of the convection in central Colorado is also in this simulation, although the model did not capture the majority of this precipitation. A small region of precipitation in southeastern Wyoming is also present in the PX simulation. PX also simulated some of the convection in eastern Texas and Louisiana.

The PX simulation generated more precipitation over this time period across central and southwestern Nebraska into parts of northeastern Colorado and northwestern Kansas, as shown in the difference plot in Figure 5.5D. The same is true over a small part of southeastern New Mexico, and portions of Texas, including eastern Texas into Louisiana, the southern half of the Texas panhandle and the northeastern part of the Texas panhandle extending into Oklahoma. Differences in these areas range from 0.5 cm to greater than 3.5 cm in Nebraska and Texas. Across southwestern Kansas, southwestern and part of western Colorado and the western Oklahoma and Texas panhandle regions into eastern New Mexico, the Noah simulation generated greater
Figure 5.5 Same as Figures 5.3 and 5.4, except valid 0000 UTC 19 June through 0000 UTC 20 June.
amounts of precipitation during the final 24 hours. The differences mainly range from 0.5 cm to 2.0 cm, with a couple of small regions in eastern New Mexico and northwestern Texas reaching greater than 3.5 cm more than the PX simulation.

Spatial plots of sensible heat flux are shown in Figures 5.6 through 5.9. For each of these figures, A is from the Noah Simulation and B is from the PX Simulation. At 1800 UTC on 18 June, the Noah simulation generated overall higher sensible heat fluxes than PX (Figure 5.6). Over much of the domain, the Noah-simulated sensible heat fluxes range from 150 to 300 W m\(^{-2}\) at this time (Figure 5.6A). Higher fluxes of 400 to as high as 600 W m\(^{-2}\) are seen over portions of Texas, Oklahoma, New Mexico and Colorado. Across eastern Kansas, as well as a small patch of southeastern New Mexico, the fluxes are much lower, ranging from less than 25 to 100 W m\(^{-2}\). These regions of lower fluxes largely correspond to areas in Figure 5.4B where precipitation fell during this time. This precipitation would have led to lower temperatures due to evaporative cooling, and therefore lower sensible heat fluxes. Higher flux gradients are seen in areas such as the western Texas panhandle and southeastern Colorado, which likely contributed to the precipitation that fell in these areas during the final 24 hours of the simulation (Figure 5.5B).

Overall lower sensible heat fluxes are seen in the PX Simulation at this time (Figure 5.6B). Just as in the Noah Simulation, the fluxes over much of the domain range from 150 to 300 W m\(^{-2}\). The main differences between the two simulations lie in the areas of higher and lower fluxes. The highest sensible heat fluxes in the PX Simulation are seen in western Colorado, where the fluxes reach as high as 500 to 550 W m\(^{-2}\). Small patches of similar fluxes are seen in New Mexico and Oklahoma. The lowest fluxes are
Figure 5.6 Surface sensible heat fluxes (W m\(^{-2}\)) valid 1800 UTC 18 June.
across eastern Nebraska and Kansas into northern Oklahoma, as well as small areas in New Mexico, Texas and Louisiana, where values range from 25 to 150 W m\(^{-2}\). The region of lower fluxes in eastern Kansas into Nebraska and Oklahoma is much larger than in the Noah Simulation, and also corresponds to greater amounts of precipitation in the PX Simulation seen over these areas in Figure 5.5C. Once again, this enhanced area of precipitation likely led to evaporative cooling, thus lowering the sensible heat fluxes.

Sensible heat fluxes at 0000 UTC on 19 June are shown in Figure 5.7. The Noah Simulation (Figure 5.7A) once again generated overall higher fluxes than the PX Simulation (Figure 5.7B). Fluxes in the Noah Simulation range from 25 to 200 W m\(^{-2}\). The lowest fluxes are over the eastern portion of the domain, as well as a swath extending across eastern New Mexico. The latter region, along with patches over central and eastern Kansas partly correspond to areas of precipitation that prior to this time. The PX Simulation generated fluxes of mostly 25 W m\(^{-2}\) over the eastern 3/4 of the domain, with slightly higher fluxes of 50 to 200 W m\(^{-2}\) over the western 1/4 of the domain. The areas of lowest fluxes do not appear to be due to precipitation that fell in the PX Simulation prior to this time.

In Figure 5.8, sensible heat fluxes at 1800 UTC on 19 June are shown. Just as in the previous two figures, the Noah Simulation (Figure 5.8A) overall produced higher fluxes across the domain in comparison to the PX Simulation (Figure 5.8B). The majority of the flux values in Figure 5.8A are comparable to those seen in Figure 5.6A, with fluxes of 150 to 300 W m\(^{-2}\) over most of the domain and patches of higher fluxes reaching upwards of 500 to 600 W m\(^{-2}\) in the western and southern portions of the domain. A swath of lower sensible heat fluxes extends in large patches from central and
Figure 5.7 Surface sensible heat fluxes (W m\(^{-2}\)) valid 0000 UTC 19 June.
Figure 5.8 Surface sensible heat fluxes (W m\(^{-2}\)) valid 1800 UTC 19 June.
western Nebraska down through western Kansas into the Texas panhandle. Just as before, these lower fluxes correspond to areas where precipitation was simulated, as seen in Figure 5.5B. In the PX Simulation (Figure 5.8B), there are larger regions of low sensible heat fluxes. These regions are found in Nebraska, Iowa, western and north central Kansas, northern Oklahoma, the Texas panhandle, and eastern Texas into Louisiana. The fluxes in these areas are lower than in the Noah Simulation, and also correspond to areas of simulated precipitation found in Figure 5.5C. In general, the lowest fluxes are seen where the highest amounts of precipitation fell.

The final sensible heat flux plot is valid at 0000 UTC 20 June 2002 (Figure 5.9). Both the Noah (Figure 5.9A) and PX (Figure 5.9B) simulations are very similar to Figure 5.7. The only notable exception lies in the Noah simulated fluxes in Figure 5.9A, where the area of lowest fluxes in the central portion of the domain is larger than in Figure 5.7A. This larger area corresponds with greater amounts of simulated precipitation seen in Figure 5.5B.

Spatial plots of latent heat fluxes are shown in Figures 5.10 through 5.13. At 1800 UTC on 18 June, the Noah simulated latent heat fluxes do not exhibit a significant amount of variation across the domain, especially across the eastern half (Figure 5.10A). Fluxes here range from approximately 300 to 500 W m⁻², with a region across eastern Kansas of 50 to 100 W m⁻². The lowest fluxes are seen in the southwest corner of the domain. The latent heat fluxes in central to western Kansas, central Nebraska and eastern New Mexico are slightly higher, where the model simulated accumulations of precipitation. The PX Simulation simulated overall higher latent heat fluxes across the domain at this time (Figure 5.10B). The highest fluxes are seen over central and southern
Figure 5.9 Surface sensible heat fluxes (W m$^{-2}$) valid 0000 UTC 20 June.
Figure 5.10 Surface latent heat fluxes (W m$^{-2}$) valid 1800 UTC 18 June.
Nebraska into western Kansas, as well as portions of Iowa, Missouri, eastern Texas, Louisiana and a small region in eastern New Mexico, where the range is 500 to 700 W m$^{-2}$. The majority of these regions of higher fluxes correspond to areas where the model simulated precipitation, since the moisture added to the ground would have led to higher latent heat fluxes. Across the rest of the domain, the fluxes range mainly from 200 to 400 W m$^{-2}$, with a few regions in the west of 50 to 100 W m$^{-2}$.

Latent heat fluxes at 0000 UTC on 19 June are shown in Figure 5.11. At this time, the Noah Simulation (Figure 5.11A) generated lower fluxes across the entire domain than the PX Simulation (Figure 5.11B). The highest fluxes in both simulations are found over Kansas into Nebraska, in areas where precipitation was seen in Figure 5.4. In these areas of higher latent heat fluxes, the Noah Simulation produced lower fluxes, ranging from 200 to 300 W m$^{-2}$, while the PX highest simulated fluxes were spread over a larger area and ranged from 200 to 400 W m$^{-2}$.

The Noah simulated latent heat fluxes at 1800 UTC on 19 June (Figure 5.12A) are similar to those seen 24 hours earlier in Figure 5.10A. The main differences between the two figures are that the fluxes in eastern Kansas as well as eastern New Mexico into western Texas are higher in Figure 5.12A. The higher fluxes in New Mexico into Texas are over areas where precipitation was simulated. The PX Simulation once again produced higher fluxes at this time (Figure 5.12B). The highest fluxes are over central and southern Nebraska, northeastern Kansas into northern Missouri and southern Iowa, northern Oklahoma, and small patches in Texas and Louisiana. The majority of these higher fluxes are in regions of precipitation, with the exception of Oklahoma and Kansas into Missouri.
Figure 5.11 Surface latent heat fluxes (W m$^{-2}$) valid 0000 UTC 19 June.
Figure 5.12 Surface latent heat fluxes (W m\(^{-2}\)) valid 1800 UTC 19 June.
At 0000 UTC on 20 June, the latent heat fluxes are again lower in the Noah Simulation as compared to those simulated in the PX Simulation (Figure 5.13). The range over most of the domain in the Noah Simulation (Figure 5.13A) is 50 to 150 W m\(^{-2}\), with some patches reaching 200 to 250 W m\(^{-2}\). In contrast, the fluxes over most of the domain in the PX Simulation (Figure 5.13B) range from 200 to 350 W m\(^{-2}\), with a few large patches of much lower fluxes mostly across the western and southeastern portions of the domain.

Time series of sensible and latent heat fluxes over three of the NCAR flux station are shown in Figures 5.14 through 5.16, with the RMS errors shown in Figure 5.17. The stations chosen for comparison here are Booker, Texas, Zenda, Kansas and Grenola, Kansas. Refer to Figure 5.13 for the locations of these sites. The sensible heat flux time series for each of these three sites are shown in Figures 5.14A, 5.15A and 5.16A, respectively. Neither simulation appeared to overall perform better than the other for these fluxes. Over Booker (Figure 5.14A), both simulations overestimated the fluxes, except for the PX Simulation underestimated them slightly towards the end. The PX Simulation produced values that were closer to observed sensible heat fluxes for this site. Over Zenda (Figure 5.15A), both simulations once again overestimated the fluxes over most of the 72-hour time period, except for 18 June, when PX slightly underestimated the fluxes. The Noah Simulation produced closer values of sensible heat fluxes on the 17 June, while the PX Simulation came closer on 18 June, and neither simulation was closer to the observations on 19 June. Over Grenola (Figure 5.16A), the Noah Simulation came the closest to the observations during the entire simulation, although both simulations overestimated fluxes on the first and third days and underestimated them on the second
Figure 5.13 Surface latent heat fluxes (W m$^{-2}$) valid 0000 UTC 20 June.
Figure 5.14 Time series of sensible and latent heat fluxes (W m\(^{-2}\)) for Booker, TX.
Figure 5.15 Time series of surface sensible and latent heat fluxes (W m$^{-2}$) for Zenda, KS.
Figure 5.16 Time series of surface sensible and latent heat fluxes (W m$^{-2}$) for Grenola, KS.
Figure 5.17 Time series of RMS errors of sensible and latent heat fluxes (W m$^{-2}$) for all nine NCAR ISFF stations.
day. The sensible heat flux RMS error time series (Figure 5.17A) shows that neither simulation overall performed significantly better than the other. Noah produced slightly higher errors on the first day, and then had slightly lower errors on the second and third days.

The time series of latent heat fluxes also do not indicate that one simulation performed better than the other. Over Booker (Figure 5.14B), the Noah Simulation came significantly closer to observations, although both simulations over predicted the fluxes. On 17 June over Zenda (Figure 5.15B), the observations appear to be in between the Noah and PX simulated fluxes, while the Noah Simulation was closer to observations on the 18 June and neither simulation was closer on 19 June. The Noah Simulation overestimated fluxes throughout the entire simulation, while PX underestimated fluxes on the first day and overestimated them during the remaining two days. The PX Simulation was closer to observations throughout the simulation over Grenola (Figure 5.16B). Both simulations again overestimated the fluxes, except for 17 June, when the PX simulated fluxes were lower than observed. The latent heat flux RMS error time series indicates that the Noah Simulation did indeed perform better than PX over all nine NCAR stations, especially on 18 and 19 June. The errors on 17 June are about the same for both simulations; with the Noah Simulation producing errors of up to 75 W m\(^{-2}\) lower than the PX on 18 and 19 June.

Little difference can be seen between the two simulations and the observations in the time series plots of incoming solar radiation over Booker (Figure 5.18A), Zenda (Figure 5.18B) and Grenola (Figure 5.18C). This indicates that differences in cloud
cover generated by both LSMs did not play a significant role in altering the surface fluxes and temperatures.

Spatial plots of soil temperature and moisture are shown in Figures 5.19 through 5.26. For each of these figures, A is from the Noah Simulation and is at a depth of 10 cm, while B is from the PX simulation and is at a depth of 1 cm. Since the soil output from the simulations are at two different levels, the exact values cannot be compared to each other. Instead, comparisons of soil temperature and moisture trends will be discussed, as well as how each simulation compares to the model predicted precipitation and surface fluxes.

Simulated 10 cm soil temperature from the Noah Simulation valid at 1800 UTC on 18 June is shown in Figure 5.19A. There is an area of lower temperatures in eastern Kansas, corresponding to the same area of lower sensible heat fluxes in Figure 5.6A. As was previously mentioned, precipitation was simulated in this area, leading to evaporative cooling and therefore lower temperatures. The lowest temperatures are found in the mountainous regions of western Colorado. In areas where higher sensible heat fluxes were seen in Figure 5.6A, including western Texas, southern New Mexico and southeastern Colorado, there are also higher 10 cm soil temperatures.

The same area of lower temperatures is found in the simulated 1 cm soil temperatures from the PX Simulation at the same time (Figure 5.19B). This also corresponds to an area of lower sensible heat fluxes seen in Figure 5.6B. The same correlation between simulated precipitation, lower sensible heat fluxes and lower temperatures is found in small patches in eastern New Mexico and also in portions of both western and eastern Texas into Louisiana. Just as in the Noah Simulation, western
Colorado is cooler than surrounding areas. The warmest temperatures are seen across eastern Colorado into western Nebraska and western Kansas, as well as most of Oklahoma, Texas and New Mexico.

Figure 5.18 Incoming solar (shortwave) radiation (W m-2) over Booker, TX (A), Zenda, KS (B) and Grenola, KS (C).
Figure 5.19 Spatial plots of 10 cm soil temperature (°C) from the Noah Simulation and 1 cm soil temperature (°C) from the PX Simulation valid 1800 UTC 18 June.
At 0000 UTC on 19 June, the lowest 10 cm soil temperatures in the Noah Simulation are once again over western Colorado (Figure 5.20A). Cooler temperatures across the eastern portions of Nebraska, Kansas and Oklahoma, extending across into Iowa, Missouri and Arkansas, correspond to regions of lower sensible heat fluxes seen in Figure 5.7A. Similarly, the highest temperatures once again are located mainly in areas of higher sensible heat fluxes found in eastern Colorado, western Nebraska, Kansas and Oklahoma, as well as New Mexico and the panhandle of Texas.

The PX simulated 1 cm soil temperatures are shown in Figure 5.20B. The coolest temperatures at this time are located in the same regions as in the Noah Simulation, with the addition of northern and eastern New Mexico, where model-simulated precipitation lowered the temperatures. Also, the warmest temperatures are found mainly in the same areas as in the Noah Simulation, only they are not quite as warm in the PX simulation. The lower temperatures seen in the PX Simulation are likely due to faster cooling of temperatures near the surface, while at the 10 cm depth in the Noah Simulation, the heat is trapped in the soil for a longer period of time.

By 1800 UTC on 19 June, the Noah simulated 10 cm soil temperatures have cooled across most of the domain (Figure 5.21A). Just as in the previous two times, the lowest temperatures are in western Colorado. Cooler temperatures in portions of Nebraska, Kansas and the Texas panhandle correspond to areas of low sensible heat fluxes, mainly due to precipitation in these areas. The warmest temperatures are seen across the southern part of the domain, especially in southern New Mexico into western Texas.
In the PX Simulation at this time (Figure 5.21B), lower 1 cm temperatures are seen in central to western Nebraska into northeastern Colorado, as well as western Kansas, the panhandle of Texas and western Louisiana, where the sensible heat fluxes were also the lowest. Once again, western Colorado has cooler temperatures than the surrounding areas. The warmest temperatures are in southeastern Colorado, central Kansas and Nebraska, Oklahoma into Arkansas, and much of Texas and New Mexico. The near-surface temperatures in the PX simulation have clearly warmed much faster than those at the lower depth in the Noah Simulation.
Figure 5.20 Spatial plots of 10 cm soil temperature (°C) from the Noah Simulation and 1 cm soil temperature (°C) from the PX Simulation valid 0000 UTC 19 June.
Figure 5.21 Spatial plots of 10 cm soil temperature (°C) from the Noah Simulation and 1 cm soil temperature (°C) from the PX Simulation valid 1800 UTC 19 June.
By 0000 UTC on 20 June, the 10 cm soil temperatures in the Noah Simulation have warmed throughout the domain (Figure 5.22A). A few patches of cooler temperatures correspond to regions where precipitation was simulated, including central to western Nebraska, southwestern Kansas, and portions of eastern New Mexico into the panhandle of Texas. Warmer temperatures are found over central Kansas, eastern Colorado, most of New Mexico and Texas, western Oklahoma and Louisiana. The areas of highest temperatures are located in southeastern Colorado and southern New Mexico into western Texas.

Similar to the previous day, the 1 cm soil temperatures generated by the PX Simulation have become much cooler by this time (Figure 5.22B). Larger areas of cooler temperatures corresponding to where precipitation fell are seen in this simulation than in the Noah Simulation, likely due to more direct effects closer to the surface. These regions include central to western Nebraska into northeastern Colorado, western Kansas, western Oklahoma, eastern New Mexico, and parts of the Texas panhandle and western Texas. Just as in the Noah Simulation, the warmest temperatures at this time are found in southeastern Colorado, as well as southern New Mexico into western Texas.

The highest values of 10 cm soil moisture at 1800 UTC on 18 June are found across central Kansas northward into eastern Nebraska (Figure 5.23A). This is a region where precipitation was simulated by the model, thus leading to greater amounts of
Figure 5.22 Spatial plots of 10 cm soil temperature (°C) from the Noah Simulation and 1 cm soil temperature (°C) from the PX Simulation valid 0000 UTC 20 June.
moisture in the soil. Latent heat fluxes in this region (Figure 5.10A) are also slightly higher than surrounding areas. Similarly, higher amounts of moisture are located in parts of western Kansas, eastern New Mexico, and portions of western Texas. The aforementioned regions of higher moisture also correspond to areas where precipitation was generated by the Noah Simulation. Large areas of higher soil moisture are found over parts of Arkansas, southern Missouri and southeastern Oklahoma; however, no precipitation was generated in these areas by the model. A few areas of lower moisture corresponding with the lower values of latent heat fluxes can be seen in Colorado, western New Mexico, eastern Nebraska and parts of the western Texas panhandle.

In the PX Simulation at this time (Figure 5.23B), the highest values of 1 cm soil moisture are seen over eastern Nebraska and Kansas. Similar to the Noah Simulation, this is also the region where the most model-simulated precipitation is found. However, the spatial coverage and amounts of moisture are higher in the PX Simulation, since the rainfall affects the soil near the surface faster than at 10 cm. Enhanced values of soil moisture are also located in western Kansas, extending southward through the Oklahoma panhandle into Texas and eastern New Mexico. Once again, these are areas where precipitation fell in the model. The aforementioned regions also correspond with higher latent heat fluxes seen in Figure 5.10B. Low amounts of soil moisture across the western portion of the domain largely correspond with the lower latent heat fluxes in the same area.

At 0000 UTC on 19 June, the spatial distribution of soil moisture in the Noah Simulation is very similar to the previous time (Figure 5.24A). The areas of highest moisture values are in the same locations, with the main exception between this time and
Figure 5.23 Spatial plots of 10 cm soil moisture from the Noah Simulation and 1 cm soil moisture from the PX Simulation valid 1800 UTC 18 June.
the previous time being that there is slightly less moisture present in the soil at this time. This could be due to the precipitation having occurred earlier in the simulation, thus giving the soil time to dry out a little. Similarly, the PX Simulation also exhibits almost the same distribution of soil moisture as the previous time (Figure 5.24B). Across the previous areas of higher moisture in Kansas and Nebraska, the moisture has decreased slightly. The moisture in this region has also now spread to the east, with the maximum amounts now located in northwestern Missouri and southwestern Iowa. In eastern New Mexico, the moisture is higher than it was before. These regions of higher moisture are once again due to precipitation that was generated by the model.

The spatial pattern of soil moisture produced by the Noah Simulation at 1800 UTC on 19 June is once again similar to the previous time (Figure 5.25A). Across eastern Kansas and Nebraska, as well as Missouri, Arkansas and southeastern Oklahoma, the areas of higher moisture are in the same locations, but the amounts are lower than before. Significantly higher moisture is now present in a swath extending from southwestern Kansas southward into the western Texas panhandle and eastern New Mexico. This corresponds to where the model generated precipitation prior to this time. In contrast, the areas of higher moisture in the PX Simulation have become much more widespread at this time (Figure 5.25B). The soil has become drier in eastern New Mexico, southeastern Nebraska, eastern Kansas, Missouri and western Iowa, while higher moisture has now spread to eastern Iowa. Where precipitation was simulated in Nebraska, northeastern Colorado, western Kansas, the panhandles of Oklahoma and Texas, and eastern Texas into Louisiana in Figure 5.5C, the soil is now significantly moister than prior to this time.
Figure 5.24 Spatial plots of 10 cm soil moisture from the Noah Simulation and 1 cm soil moisture from the PX Simulation valid 0000 UTC 19 June.
Figure 5.25 Spatial plots of 10 cm soil moisture from the Noah Simulation and 1 cm soil moisture from the PX Simulation valid 1800 UTC 19 June.
By 0000 UTC on 20 June, the moisture pattern in the Noah Simulation still has not changed significantly (Figure 5.26A). The important differences at this time lie in central to western Nebraska and eastern New Mexico, where model-simulated precipitation has led to patches of higher soil moisture. In the PX Simulation, the moisture across western Nebraska into northwestern Kansas has increased from the previous time (Figure 5.26B). There is also higher moisture in central Colorado, southeastern Wyoming and eastern New Mexico, as well as a swath of higher moisture extending from the northeastern Texas panhandle into central Kansas. Eastern Kansas, Missouri and Iowa are now drier than before. Overall, the Noah simulated 10 cm soil moisture exhibited less variation over the course of the simulation than the 1 cm soil moisture in the PX Simulation. The reason for this is likely due to more direct effects of the rainfall on moisture towards the surface in the PX Simulation than at the lower depth in the Noah Simulation.

In Figures 5.27 through 5.30, plots of surface winds and mixing ratio are shown. For each of these plots, the anomalous wind pattern seen in the extreme western portion of the domain is due to mountain effects. At 1800 UTC on 18 June, the surface winds across most of the domain in the Noah Simulation are from the south-southeast (Figure 5.27A). However, in northeastern Colorado, winds are from the west-northwest, creating a convergence zone in western Kansas through central Nebraska. Convection formed along and ahead of this boundary. Higher mixing ratios are seen associated with the precipitation in northern Kansas and central Nebraska, along with southeastern New Mexico. The highest mixing ratios are located in the Texas panhandle, which may have helped to fuel the convection that occurred on the following day. A similar surface wind
Figure 5.26 Spatial plots of 10 cm soil moisture from the Noah Simulation and 1 cm soil moisture from the PX Simulation valid 0000 UTC 20 June.
Figure 5.27 Spatial plots of boundary layer wind vectors (m s\(^{-1}\)) and mixing ratio (g kg\(^{-1}\))
valid 1800 UTC 18 June.
pattern is found in the PX Simulation at this time (Figure 5.27B). Surface convergence can also be seen in western Kansas and central Nebraska in this simulation, once again corresponding with precipitation that formed in these regions. There is also convergence associated with the convection in eastern New Mexico. Mixing ratios at this time are the highest in Kansas and Nebraska, and also western Texas.

At 0000 UTC on 19 June, the surface winds in the Noah Simulation are still primarily from the south-southeast (Figure 5.28A). The surface boundary in Kansas and Nebraska now extends further to the south-southwest, through Colorado and New Mexico. Precipitation formed along and ahead of this boundary, corresponding mainly with the areas of higher mixing ratios found in Kansas, Nebraska and Texas. The surface boundary in the PX Simulation appears to be shifted further to the east (Figure 5.28B). In this simulation, the boundary layer mixing ratios are much higher than in the Noah Simulation. Once again, the highest mixing ratios are located mainly where the most convection occurred in this simulation.

The surface boundary in the Noah Simulation at 1800 UTC on 19 June is still located in the same general area (Figure 5.29A). Along and ahead of this boundary is where the highest mixing ratios are found, along with most of the convection that formed. Across most of Kansas southward through Oklahoma, lower mixing ratios are found where no convection occurred. The PX Simulation at this time once again produced higher mixing ratios, mostly along and ahead of the surface boundary (Figure 5.29B). Precipitation again formed along and ahead of this boundary.

In the final surface plot, valid at 0000 UTC on 20 June, there is still convergence from central Nebraska southwestward through Colorado and New Mexico (Figure
Figure 5.28 Spatial plots of boundary layer wind vectors (m s$^{-1}$) and mixing ratio (g kg$^{-1}$) valid 0000 UTC 19 June.
Figure 5.29 Spatial plots of boundary layer wind vectors (m s\(^{-1}\)) and mixing ratio (g kg\(^{-1}\)) valid 1800 UTC 19 June.
5.30A). Enhanced moisture is along this boundary, helping to fuel the convection that was present at this time. Higher mixing ratios have spread into eastern Kansas and Oklahoma; however less moisture is still located just to the west of there, where no convection formed in this simulation. The PX Simulation at this time has the boundary further to the east, and once again produced much higher mixing ratios along and to the east of the boundary (Figure 5.30B). This would explain why the PX Simulation generated greater accumulations of precipitation during the final 24 hours of the simulation that the Noah Simulation.

5.5 Summary

The purpose of this study was to examine the impacts of land-surface and boundary layer coupling on simulations of convective initiation and precipitation. Two sets of simulations were performed, one coupling the Noah LSM with the MRF PBL scheme and the other using the Pleim-Xiu LSM and the ACM PBL scheme. In terms of precipitation, the PX and Noah simulations exhibited similar features to each other. However, dramatic differences on timing of convection are apparent when compared to the observations. In both simulations, the model lagged behind the observations and was considerably slower in evolving the system. Significant differences were seen in the evolution of soil moisture and temperature, which was to be expected due to the different soil layers. There was a more immediate soil-precipitation feedback in the PX run, where the top soil layer was closer to the surface. A similar feedback could be seen in the heat flux fields. Differences in surface circulations and convergence were also seen in the two simulations, probably also due to the previously mentioned feedbacks.
Figure 5.30 Spatial plots of boundary layer wind vectors (m s$^{-1}$) and mixing ratio (g kg$^{-1}$) valid 0000 UTC 20 June.
The use of different LSM/PBL coupled schemes in the same modeling system was expected to yield subtle differences between the two simulations. However, the results of this study showed that significant differences occurred, thus providing evidence that the land-surface is an important component to a mesoscale model. Results indicate that improvements in land-surface models may have a dramatic improvement in forecasts of convective initiation and precipitation.
CHAPTER 6: CONCLUSIONS

The purpose of this research was to study the effects of land-atmosphere interactions on simulations of convective initiation and associated rainfall. The case study used for this research occurred during the International H2O Project (IHOP_2002) field study. IHOP_2002 was conducted over the Southern Great Plains region of the United States, between 10 May and 25 June 2002, with its primary goal being to explore improvements of water vapor measurements in the lower atmosphere in order to improve summertime QPF. The case used was from 17-20 June 2002, where convection occurred along various dryline boundaries under strong synoptic forcing.

Two studies were conducted using the MM5 modeling system. The first study examined the impacts of the surface data assimilation within the model. Two simulations were conducted for this study, the Control Simulation using standard MM5 data assimilation and the Experimental Simulation using the Flux-Adjusting Surface Data Assimilation System (FASDAS). Results of this study showed a significant difference between the two simulations, with an overall improved performance in the Experimental Simulation. The Experimental Simulation produced more well-defined surface features, including surface boundaries and heat flux gradients, which likely led to the enhanced precipitation generated by the model. The Control Simulation, on the other hand, generated very little precipitation and more uniform surface features. The results of this study suggest that improvements in the methods of assimilating surface data into mesoscale operational models may contribute to a greater skill in QPF.

The second study involved the use of two different land-surface model (LSM) and planetary boundary layer (PBL) coupled schemes. The first simulation used the Noah
LSM and the MRF PBL, while the second simulation used the Pleim-Xiu LSM coupled with the ACM PBL scheme. Both simulations produced similar precipitation patterns; however dramatic differences in the amount and timing of precipitation were seen when compared to the observations. Significant differences were seen in the evolution of surface and soil features in each simulation. Results of this study provide evidence of the importance of the representations of both land-surface and PBL processes, and their subsequent effects on the generation of convection within the model.

Overall, the results of both of these studies provide comprehensive evidence that changes in land-surface processes and land-atmosphere feedbacks play an important role in the development of deep moist convection even during strong synoptic forcing. Therefore, improvements are necessary in the way these feedbacks are represented in operational forecasting models.

Future research is planned to examine this matter further, including a study using different land-surface models coupled to the same PBL scheme, assimilation of other forms of data, such as satellite data, and simulations over other regions other than the SGP. Studies such as these will ideally lead to improvements in the overall skill of forecasts of convection initiation and rainfall.
REFERENCES


