Potential effects of regional soil moisture anomalies on the Great Plains low-level jet

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POTENTIAL EFFECTS OF REGIONAL SOIL MOISTURE ANOMALIES ON THE GREAT PLAINS LOW-LEVEL JET

by

Matthew A. Campbell

A Thesis
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ABSTRACT

The Great Plains low-level jet (GPLLJ) contributes to Great Plains (GP) warm season water resources (precipitation), wind resources, and severe weather outbreaks. Past research has shown that synoptic and local mesoscale physical mechanisms (Holton and Blackadar mechanisms) are jointly required to explain GPLLJ variability. Although local mechanistic theories hinge upon soil moisture-planetary boundary layer (PBL) interactions, the effect of regional soil moisture anomalies on GPLLJ speed, northward penetration, and propensity for severe weather is not well known.

In this thesis, four 31-member WRF-ARW Stochastic Kinetic Energy Backscatter Scheme ensembles simulate two synoptically-contrasting typical warm season GPLLJs under CONUS-wide wet and dry soil moisture scenarios. In the GP (48°–24°N and ~103°–90°W), wet and dry ensemble mean differences in sensible heating and PBL height of ~10–100 W m⁻² and ~100–700 m, respectively, at 2100 UTC (afternoon) culminate in GPLLJ 850-hPa windspeed differences of ~1.0–4.0 m s⁻¹ nine hours later (0600 UTC; early morning) in both case studies. Greater heat accumulation in the daytime PBL over dry soil impacts the zonal temperature gradient in the GP (Holton mechanism) thereby impacting the mass field in the northern GP, causing increases in the geostrophic windspeeds. Larger diurnal contrasts in turbulent mixing over dry soil impacts the PBL structure (Blackadar mechanism) and leads to increased ageostrophic winds speeds overnight. Together, the geostrophic and ageostrophic winds generally constructively interact to compose the faster nocturnal GPLLJs over dry soil. Ensemble differences in CIN (~50–150 J kg⁻¹ K⁻¹) and CAPE (~500–1000 J kg⁻¹ K⁻¹) have implications for severe weather predictability.
I would like to thank my advisors Dr. Craig Ferguson and Dr. Lance Bosart for their advisement during completion of this thesis. Their ideas and suggestions greatly added to the science presented here. Significant contributors to this thesis also include members from the Dr. Ferguson’s research group including Dr. Alex Burrows, Dr. Geng Xia and Mark Beauharnois. Without the support of the research group, this work would have been extremely difficult to complete and would likely be missing many key insights. Special thanks is also expressed toward Ariel Cohen, Robert Fovell, Ryan Torn, Matthew Vaughan, Tomer Burg, Massey Bartolini, and Kevin Tyle for discussions and technical aid that contributed to this work.

I would also like to thank my family and friends for their support during the past six years of schooling. Encouragement and backing throughout the completion of my undergraduate and graduate degrees has made achieving my career goals possible. Additionally, I am grateful for the experiences I have received at The Ohio State University in Columbus, Ohio and the National Weather Center Research Experiences for Undergraduates (REU) Program at the University of Oklahoma in Norman, Oklahoma. I would not be where I am personally or professionally without the friends, professional/educational experiences, and personal growth gained during my undergraduate and REU programs.

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5.1 Discussion and conclusions

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References
1. Introduction

1.1 Motivation

The United States Great Plains (GP) is currently well-studied in the hydrological and meteorological communities for two main reasons: it serves as the corridor of a prominent low-level jet (LLJ) and it is a “hot-spot” for modeled land-atmospheric coupling strength. Both research themes are socially motivated by the need to better understand and predict precipitation in a region that is highly susceptible to seasonal droughts, such as the 2012 “flash drought” (e.g., Hoerling et al. 2014), and prolonged multi-year droughts, such as the Dust Bowl of the 1930s (NCEI 2019). Furthermore, the GP’s $92 billion per year crop and livestock industry critically depends on GP low-level jet (GPLLJ)-related precipitation (Basara et al. 2013; Melillo et al. 2014) and GPLLJ-related wind events in Texas, Oklahoma, Kansas, and Nebraska collectively support up to 45% of the United States’ installed wind energy generation capacity (AWEA 2018).

First objectively defined by Bonner (1968), the GPLLJ is characterized by a diurnally oscillating, low-level wind maximum below three kilometers above ground level with decreasing winds above (e.g., Helfand and Schubert 1995; Mitchell 1995; Whiteman et al. 1997) and below (e.g., Walters and Winkler 2001) the nose of the LLJ. The GPLLJ is most frequent over the central and southern GP during the warm season (e.g., Bonner 1968; Fig 1.1). Importantly, the GPLLJ acts as the primary conveyor of atmospheric moisture from the Gulf of Mexico northward, anywhere from Texas to southern Manitoba and Saskatchewan, along an axis which varies between approximately 90°W and 103°W. GP warm season precipitation (e.g., Means 1954; Dirmeyer and Brubaker 1999), severe weather, tornado outbreaks (e.g., Muñoz and
Enfield 2011), and mesoscale convective systems (MCSs) are largely driven, sustained, and modulated by the GPLLJ (e.g., Maddox 1983; Cotton et al. 1989; Coniglio et al. 2010) through moisture convergence (e.g., Helfand and Schubert 1995; Higgins et al. 1997; Wang and Chen 2009) and vertical wind shear (e.g., French and Parker 2010) at the jet’s northern exit region.

The GP land-atmosphere coupling “hotspot” terminology was born out of the GEWEX Global Land Atmospheric Coupling Experiment (Koster et al. 2004; Guo et al. 2006; Koster et al. 2006), which demonstrated, using an ensemble of coarse resolution global circulation models, that realistic soil moisture can improve predictability of GP 2-meter temperature and precipitation on sub-seasonal timescales. The dependence of weather predictability on soil moisture led to the GP being highlighted as a region of strong land-atmospheric coupling (Fig. 1.2). Later, using atmospheric reanalysis datasets, it was shown that the strength of the region’s land-atmospheric coupling is also largely dependent upon moisture convergence associated with the GPLLLJ and soil moisture availability (Song et al. 2016). In the context of GPLLLJ-forced sub-seasonal to seasonal precipitation variability, which cannot be fully explained by internal atmospheric variability (e.g., Yu et al. 2017) or sea-surface temperature forced variability (e.g., Schubert et al. 2004; Song et al. 2005; Yang et al. 2007, Kam et al. 2014; Krishnamurthy et al. 2015; Yu et al. 2017), many outstanding questions remain. Two particularly important questions are: how sensitive is the GPLLJ to regional soil moisture anomalies and how can this sensitivity be interpreted/understood dynamically?

1.2 Literature review

1.2.1 Coupled versus uncoupled GPLLLJs
Uccellini (1980) laid the framework for differentiating GPLLJs according to the relative strength of their large-scale synoptic forcing (e.g., ridge or trough-ridge systems) as compared to local land-atmospheric interactions. He showed that the GPLLJ’s have two distinct synoptic states: 1) uncoupled from the upper-level jet stream and dominated by local land and planetary boundary layer (PBL) processes, and 2) coupled to the upper-level jet stream via a thermally indirect circulation associated with the upper-level jet exit region (Uccellini and Johnson 1979).

According to Uccellini (1980), uncoupled, terrain-process dominated GPLLJs tend to exhibit a clear diurnal cycle and occur when there is a large ridge over the central U.S. with weak flow aloft over the southern GP (Fig. 1.3). More specifically, uncoupled GPLLJs develop around sunset, veer overnight, maximize in intensity between 0600 UTC and 0900 UTC (early morning) (e.g., Song et al. 2005; Berg et al. 2015) and occur above the nocturnal PBL within the residual mixed layer (Blackadar 1957). The frequency of uncoupled GPLLJ occurrence tends to peak in the warm season due to favorable synoptic conditions caused by the climatological placement of the subtropical high (Wexler 1961) and long-term heating of the GP’s gently westward-upward-sloping terrain leading to the thermal wind forcing conducive to GPLLJ development (e.g., Parish 2017; Burrows et al. 2019).

According to Uccellini (1980), coupled GPLLJs and their supportive upper-level jets occur when there is an upper-level trough over the western U.S. that is propagating toward the GP and jet streaks are ejecting into the central U.S. from the southwest (Fig. 1.4). Coupled GPLLJs may not always exhibit as clear a diurnal signal as uncoupled jets because the physical mechanisms driving coupled GPLLJs are only partially explained by diurnally driven local terrain and PBL processes. Synoptically-forced GPLLJs peak frequency occurs during the
shoulders of the warm season, specifically in May and September (Burrows et al. 2019), when the synoptic environment is conducive to increased baroclinicity and cyclonic wave activity, which is often lacking in the middle of the summer.

1.2.2 GPLLJ formation mechanisms

The GP local land conditions favor three primary terrain mechanisms for GPLLJs: 1) the Blackadar mechanism (Blackadar 1957), 2) the Holton mechanism (Holton 1967), and 3) the Wexler mechanism (Wexler 1961). The Blackadar mechanism describes how the GPLLJ results from inertial wind oscillations in the PBL due to nocturnal reductions in friction of the low-level flow. According to Blackadar (1957), during the warm season in the GP, daytime insolation heats the surface of the earth and the convective boundary layer grows as turbulent vertical mass exchange (vertical mixing) occurs. Vertical mixing helps to reduce the vertical wind shear within the boundary layer and reduce the difference between the maximum and minimum wind speed within the boundary layer during the day (Fig. 1.5). After sunset, convectively-driven vertical mixing rapidly declines as radiative cooling leads to temperature decreases at and near Earth’s surface. As the ground cools, the lower atmosphere stabilizes and a nocturnal temperature inversion forms within a few hundred meters of the ground. Just above the temperature inversion, in the residual layer, the aforementioned reduction in vertical mixing is associated with a rapid reduction in friction that can lead to an inertial wind oscillation. The frequency of the inertial wind oscillation is 24 hours at 30°N (within the GP) and the inertial wind has a southerly component during the overnight hours in the GP. During the day the inertial wind has a northerly component, but the inertial wind is generally not considered important during the
daytime due to the diurnal cycle of daytime heating and friction-associated PBL structure in the midlatitudes. Overall, the inertial wind oscillation is quantified as an ageostrophic wind component, rotates (veers) clockwise, and adds to background geostrophic flow just atop the nocturnal temperature inversion when southerly background flow is present (e.g., Blackadar 1957; Parish 2017) (Fig. 1.6). The stable temperature profile at night prevents near GPLLJ winds from mixing down to the ground and being slowed by surface friction. As long as a stable surface layer is present, a sharp vertical gradient in wind speed can be maintained all night with the wind maximum (i.e., the GPLLLJ) just atop the temperature inversion (Fig. 1.5).

The Holton mechanism describes how diurnal fluctuations in the lower-tropospheric thermal profile across sloping terrain contributes in GPLLJ formation. Holton (1967) demonstrates that radiatively cooled air can drain downslope from the elevated terrain in the western GP and contribute to the diurnal fluctuations of GPLLJ strength during the summertime via mass field and geostrophic wind adjustments. Holton’s results suggest that the Blackadar mechanism may not be the only forcing mechanism important for GPLLJ formation. According to the Holton mechanism, the elevated and dry western GP experiences greater diurnal ranges in lower-tropospheric temperature than the lower, more humid, eastern GP. Consequentially, the zonal temperature gradient across the entire GP reverses from day to night, with warmer temperatures in the western GP during the daytime and cooler temperatures in the western GP overnight. The diurnal reversal of zonal temperature gradients impacts the low-level mass field and results in enhanced geostrophic wind contributions to the GPLLJ during the nighttime compared to the daytime. In essence, the Blackadar mechanism focused on the ageostrophic component of the wind contributing to the often-observed diurnal patterns of GPLLJ wind speed. The Holton mechanism builds on the Blackadar mechanism and demonstrates that diurnal
changes in the geostrophic wind, associated with a varying zonal temperature gradient, are also important for dynamical understanding the GPLLJ.

The Wexler (1961) mechanism describes how mechanical blocking and northward steering of tropical easterlies by the Sierra Oriental/Rocky Mountain cordillera contributes to GPLLJ formation. Wexler (1961) applied inertial boundary layer theory, originally developed to explain boundary currents in the ocean, to the atmosphere and terrain of North America in order to explain the GPLLJ. Wexler’s results demonstrate that the northward veering of the trade-winds, which are associated with GPLLJ, are somewhat analogous to the Gulf Stream current off the coast of eastern North America. The Wexler mechanism is active when easterly flow south of the Bermuda High, which is a common climatological feature in the warm season, is blocked by the Rocky/Sierra Madre Mountains (Fig. 1.7) similarly to how North Atlantic Ocean currents are blocked by the terrain of North America. The easterly flow is forced northward and results in synoptic conditions favoring southerly flow conducive to GPLLJ formation. These synoptic flow conditions are associated decreasing lower atmospheric geostrophic heights from east to west across central and eastern North America with southerly geostrophic wind. Wexler (1961) acknowledges that local land-atmosphere and PBL processes are needed in order fully explain the super-geostrophic and diurnal aspects of the GPLLJ but emphasis the importance of conducive background synoptic conditions needed for southerly GPLLJ formation.

More recent idealized numerical modelling efforts have expanded our understanding of the Blackadar, Holton, and Wexler mechanisms (e.g., Ting and Wang 2006; Shapiro and Fedorovich 2010; Du and Rotunno 2014; Shapiro et al. 2016; Fedorovich et al. 2017), in the way that these three mechanisms drive the GPLLJ, especially during periods of synoptic quiescence. The importance of many additional parameters associated with the three main GPLLJ forcing
mechanisms, including but not limited to: slope of the terrain, lower atmospheric stability, the maximum and minimum surface buoyancy, and sunset and sunrise times (length of day in relation to inertial wind) have all been considered. Regardless of the mechanism or combination of mechanisms that are active for a given GPLLJ event, it is hypothesized that perturbations to the soil moisture state may lead to modulation, suppression, or enhancement of the GPLLJ. The GPLLJ Blackadar and Holton mechanisms are locally-driven processes that are inherently tied to the terrain and local land conditions. Soil moisture has the ability to impact the PBL structure (Blackadar mechanism) through PBL height (PBLH) adjustments via modified turbulent eddy fluxes (e.g., Ek and Mahrt 1994; Ek and Holtslag 2004; Santanello et al. 2018) (Fig. 1.8). Any modifications to PBL structure could have the potential to impact the inertial oscillation associated with the Blackadar mechanism thereby impacting the GPLLJ. Modifications to soil moisture can also impact lower tropospheric temperature (e.g., Ek and Mahrt 1994; Ek and Holtslag 2004) which can lead to the altered zonal temperature gradients and thermal wind profiles that can impact the GPLLJ (McCormle 1988; Fast and McCormle 1990; Zhong et al. 1996).

1.3 Research goals and thesis structure

An ensemble modeling approach using the Advanced Weather Research and Forecasting (WRF-ARW, hereafter WRF) version 3.9.1.1 (Skamarock et al. 2008) model is employed to assess the impact of regional soil moisture availability on the GPLLJ and associated weather. Specifically, the impact of extreme wet and dry soil moisture states on a “classic” uncoupled GPLLJ and a coupled GPLLJ is investigated. Modeling the impact of soil moisture on the
GPLLJ may contribute to the dynamical understanding of GPLLJ and help to improve the accuracy of seasonal and sub-seasonal forecasts in the GP. The goals of this thesis are to: 1) quantify the impact of wet and dry soil moisture anomalies on the GPLLJ, sensible weather, and severe weather parameters, 2) dynamically link any apparent changes to the GPLLJ, sensible weather, and severe weather parameters to the prescribe soil moisture anomalies, and 3) assess the potential societal importance of any observed modifications to the GPLLJ, sensible weather, and severe weather parameters. Four additional coupled cases and four additional uncoupled cases, which display similar results to the two cases studies analyzed in theses, were also conducted but are not discussed.

The organization of this thesis is as follows. Chapter 2 describes the criteria employed to select the case studies in this thesis, the WRF model configuration, experimental design, and significance testing. Chapter 3 presents a synoptic overview of an uncoupled GPLLJ case, the impacts of soil moisture prescription on the uncoupled GPLLJ, sensible weather, and severe weather parameters, and dynamically links atmospheric changes to soil moisture prescription as well as severe weather and societal impacts. Chapter 4 presents a synoptic overview of a coupled GPLLJ case, the impacts of soil moisture prescription on the coupled GPLLJ, sensible weather, and severe weather parameters, and dynamically links atmospheric changes to soil moisture prescription as well as severe weather and societal impacts. Chapter 5 succinctly compares, contrasts, and summarizes key details of the uncoupled and coupled case studies and concludes with suggestions for future research.
FIG 1.1. Frequency distributions of GPLLJs (Bonner class 1) observations within 30º class intervals of wind direction at the level of maximum wind. Distribution is for the summer months. [Figure 6 and caption adapted from Bonner (1968).]
FIG 1.2. The land-atmospheric coupling strength diagnostic for boreal summer (the omega difference, dimensionless, describing the impact of soil moisture on precipitation), averaged across 12 models participating in Global Land Atmosphere Coupling Experiment. (Insets) Aerially averaged coupling strengths for the 12 individual models over the outlined, representative hotspot regions. No signal appears in the obscured regions of southern South America and South Africa. [Figure 1 and caption adapted from Koster et al. 2004.]
FIG 1.3. Schematic of upper tropospheric 300-hPa flow patterns for 3 uncoupled GPLLJ cases. The 3 GPLLJ cases are taken throughout the entire year. [Figure 2 and caption adapted from Uccellini (1980).]
FIG 1.4. Schematic of upper tropospheric 300-hPa flow patterns for 12 coupled GPLLJ cases. The 12 GPLLJ cases are taken throughout the entire year. [Figure 2 and caption adapted from Uccellini (1980).]
FIG 1.5. Schematic illustration explaining the evolution of a boundary layer jet profile. The wind profile on the left can be thought of as a typical daytime wind profile during the warm season over the GP. The wind profile on the right can be thought of as a typical nighttime wind profile during the warm season over the GP. [Figure 10 and caption adapted from Blackadar (1957).]
FIG. 1.6. Composite analysis of warm season GPLLJ cases in three-hour increments. Black vectors represent the total wind at the level of the GPLLJ. Red vectors represent the geostrophic wind at the level of the GPLLJ. Blue vectors represent the ageostrophic wind at the level of the GPLLJ. [Figure 12a and caption adopted from Parish (2017).]
FIG. 1.7. Schematic sketch of trade-winds entering the Caribbean Sea, Gulf of Mexico, and southern United States. [Figure 3 and caption adapted from Wexler (1961).]
FIG. 1.8. Important interactions between the surface and atmospheric boundary layer for conditions of daytime surface heating. Solid arrows indicate the direction of feedbacks that are normally positive. Dashed arrows indicate negative feedbacks. Two consecutive negative feedbacks make a positive one. Note the many positive and negative feedback loops that may lead to increased or decreased relative humidity and cloud cover. [Figure 1 and caption adapted from Ek and Mahrt (1994).]
2. Data and methodology

2.1 Case selection

In order to investigate the possible impacts of soil moisture on a typical summertime GPLLLJ, the criteria set by Uccellini (1980) to describe a “classic”, uncoupled GPLLLJ is employed to select the synoptic environment for the uncoupled case used in this thesis. A LLJ criteria, similar to the one first introduced by Bonner (1968), and later modified by Walters and Winkler (2001), to describe the vertical wind shear profile of a LLJ, is also used. To be considered a LLJ, the windspeed must increase with height above the surface and ultimately peak in the lower troposphere before decreasing with height above the level of maximum winds. More specifically, the 850-hPa windspeed, which is used to analyze the GPLLLJ, must be at least 4.0 m s\(^{-1}\) greater than the surface and 700-hPa windspeed. The use of 4.0 m s\(^{-1}\) of vertical wind shear above and below the GPLLLJ is consistent with the vertical wind shear criteria used by Walters and Winkler (2001). Using ERA-Interim reanalysis (Dee et al. 2011) and taking into account these two primary considerations, the “classic” uncoupled GPLLLJ on the morning of 21 July 2017 (Fig. 2.1) is selected for further analysis. ERA-Interim data was obtained via the Research Data Archive at the National Center for Atmospheric Research (https://rda.ucar.edu/datasets/ds627.0/).

Similarly, the criteria set by Uccellini (1980) to describe a coupled GPLLLJ is employed to select the synoptic environment for the coupled GPLLLJ case used in this experiment. The LLJ criteria described above is also used to ensure the vertical wind shear profile matches that of a GPLLLJ. Taking these two primary considerations into account, the coupled GPLLLJ on the morning of 10 June 2012 (Fig. 2.2) is selected for further analysis.
2.2 The model

For the two case studies examined in this thesis, the WRF is employed with a horizontal resolution of 27 kilometers, which is suitable for GPLLJ studies (e.g., Helfand and Schubert 1995; Zhong et al. 1996; Doubler et al. 2015), and 50 sigma coordinate vertical levels between the surface and 50 hPa. The domain (Fig. 2.3) size that is selected ensures the GPLLJs do not approach the domain boundaries of the simulations. The domain also incorporates dynamically important regions for GPLLJ variability, such as the Gulf of Mexico, the western North Atlantic, and the Rocky Mountains. Instead of running various parametrization schemes in order to generate forecast ensembles, the current experiments utilize the selected parameterizations listed in Table 2.1 along with the Stochastic Kinetic Energy Backscatter Scheme (SKEBS; Shutts 2005) described in Chapter 2.5 in order to limit degrees of freedom in the experiment and test for statistical significance.

2.3 Experimental design

The initial conditions for all model states, except for soil moisture, and all boundary conditions for WRF are provided by the ERA-Interim reanalysis dataset at approximately 0.7-degree horizontal resolution across 37 pressure levels extending from 1000 hPa to 1 hPa and a surface level. The uncoupled GPLLJ simulations are initialized at 0000 UTC 20 July 2017, which is 24 hours prior to the uncoupled GPLLJ of interest, and end at 1800 UTC 21 July 2017, which is after the cessation of particular uncoupled GPLLJ analyzed in this study. The coupled
GPLLJ simulations are initialized at 0000 UTC 09 June 2012, which is 24 hours prior to the coupled GPLLJ of interest, and end at 1800 UTC 10 June 2012, which is after the cessation of particular coupled GPLLJ analyzed in this study. Atmospheric boundary conditions are updated at the temporal resolution of ERA-Interim, which is every six hours.

In order to test GPLLJ sensitivity to soil moisture, two distinct soil moisture scenarios are prescribed in all four soil levels at model initialization for both simulations. The soil moisture prescriptions are generated using an identically configured Noah Version 3.6 Land Surface Model (Noah LSM) as in WRF. The Noah LSM is run from 1981–2016, with one year of spin-up, using the NASA Land Information System (LIS; Kumar et al. 2006; Peters-Lidard et al. 2007) software framework in order to create a spun-up, daily, model-based warm season (May to September) climatology of soil moisture values at each gridpoint over the domain. The Noah LSM soil moisture simulation uses the same model physics, soil, and surface vegetation parameterizations, and grid mesh as the coupled WRF model. The hourly meteorological forcing for the Noah LSM was taken from the Modern-Era Retrospective Analysis for Research and Applications version 2 (MERRA-2; Gelaro et al. 2017; Reichle et al. 2017) and included bias-corrected precipitation. For the WRF simulations, the 5th percentile and 95th percentiles of 1982–2016 warm-season daily volumetric soil moisture content at each gridpoint and for each of Noah LSM’s four soil layers are prescribed at initialization. The 5th percentile represents dry soil conditions (Fig. 2.4b) at each gridpoint and the 95th percentile represents wet soil conditions (Fig. 2.4a) at each gridpoint. The differences between the wet and dry soil moisture prescriptions in each of the four soil layers is plotted in Fig. 2.4c-f. The volumetric top layer (0–10 cm layer) soil content differences in the GP are approximately of 0.05–0.15 m$^3$ m$^{-3}$. The 95th and 5th percentiles of climatological warm season soil moisture are used in order ensure a reasonably
large difference between soil moisture values occur at each gridpoint over the entire domain. If large differences in soil moisture cannot dynamically generate a GPLLJ response, it is hypothesized that smaller differences cannot either.

2.4 Significance testing

Quantifying model uncertainty with a full WRF multi-physics ensemble (e.g., Mooney et al. 2017) was deemed beyond the scope of this study. Instead, a SKEBS (Berner et al. 2011, 2015) ensemble-based approach is used to test for statistical significance. Prior studies have demonstrated SKEBS can account for physical uncertainty associated with subgrid-scale turbulence (e.g., Berner et al. 2011, 2015; Judt et al. 2016). Following the most common SKEBS approach, suggested default white noise perturbations are applied to the potential temperature and rotational wind component fields continuously, at every model time step and only within the model domain interior. As described in the WRF user manual (NCAR 2017), these default values are tuned for the synoptic-scale perturbations in middle latitudes, consistent with our use case. The lateral boundary conditions are not perturbed as synoptic uncertainty associated with the ERA-Interim reanalysis dataset is not the focus of this study.

For each GPLLJ case study, one ensemble group consists of members initialized with wet soil moisture conditions and the other ensemble group consists of members initialized with dry soil moisture conditions at each gridpoint in each soil layer. Each ensemble group contains 31 total members: 1 control run without SKEBS perturbations and 30 runs generated by setting the random number seed in SKEBS from 1 to 30. A difference of means $t$-test is iterated over each gridpoint in order to test if the two ensemble groups within each respective case study are
statistically different at the 95 percent confidence level (i.e., $p \leq 0.05$) for selected meteorological fields. Using SKEBS to generate ensembles and testing for statistical significance should help to increase the probability that apparent differences between the wet and dry ensembles are the result of physically-relevant processes and are not due to random chance or model uncertainty. The wet and dry mean ensemble fields are compared and contrasted within the GP from approximately 48°–24°N and 103°–90°W. 0900 UTC 21 July 2017 and 0900 UTC 09 June 2012 are selected as the primary analysis times since GPLLJ wind differences between ensembles are maximized at this time.
FIG. 2.1. ERA-Interim reanalysis derived synoptic environment on (a) 0000 UTC, (b) 0600 UTC, (c) 1200 UTC 21 July 2017. 850-hPa windspeed (shaded; m s$^{-1}$), wind vectors (barbs; half barb is 2.5 m s$^{-1}$, full barb is 5.0 m s$^{-1}$, and flag is 25.0 m s$^{-1}$), and 500-hPa geopotential height (black contours; every 5.0 dam) are shown. Land is shaded in grey and oceans are white.
FIG. 2.2. ERA-Interim reanalysis derived synoptic environment on (a) 0000 UTC, (b) 0600 UTC, (c) 1200 UTC 10 June 2012. 850-hPa windspeed (shaded; m s$^{-1}$), wind vectors (barbs; half barb is 2.5 m s$^{-1}$, full barb is 5.0 m s$^{-1}$, and flag is 25.0 m s$^{-1}$), and 500-hPa geopotential height (black contours; every 5.0 dam) are shown. Land is shaded in grey and oceans are white.
FIG. 2.3. Topography over the WRF model domain, based on GTOPO30 (Gesch et al. 1999). Horizontal grid size is 27 km and 50 vertical levels are used between the surface and 50 hPa.
<table>
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<tr>
<th>WRF Parameterizations</th>
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<tr>
<td>Microphysics</td>
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<td>Surface Layer</td>
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<tr>
<td>Shortwave Radiation</td>
<td>Iacono et al. 2008</td>
<td>RRTMG</td>
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</tbody>
</table>

TABLE 2.1. Parametrizations selected for the WRF simulations in this study.
FIG. 2.4 a) 95th and b) 5th percentile Noah v3.6 daily warm season (May-September) 0-10 cm (top layer) volumetric soil moisture (m$^3$ m$^{-3}$) and c) their difference (95th-minus-5th percentile). Fields are calculated from a 34-year (1982-2016) NASA-LIS offline simulation with identical surface model physics, (soil and vegetative) parameters, and grid mesh as the coupled WRF configuration. d) Same as 3c but for 10–40 cm layer. e) Same as 3c but for 40–100 cm layer. f) Same as 3c but for 100–200 cm layer.
3. Uncoupled case study

3.1 Synoptic setup

The GPLLJ of 21 July 2017 exhibits the defining characteristics of a “classic” uncoupled GPLLJ (Uccellini 1980). According to the ERA-Interim reanalysis, a large upper-level ridge at 500 hPa was present over the central U.S. at 0600 UTC 21 July 2017 (Fig. 2.1b). The GPLLJ at 850 hPa was located directly under the upper-level ridge axis. There was weak flow aloft (not shown), especially over the central and southern GP, indicative of a lack of upper-level jet stream support for GPLLJ formation. Clear sky conditions were present and warm conditions persisted over much of the region. Similar analysis at 0000 UTC (Fig. 2.1a) and 1200 UTC (Fig. 2.1c) 21 July 2017 reveal a nocturnally oscillating GPLLJ at 850 hPa with veering winds overnight and the GPLLJ first appearing after 0000 UTC. Furthermore, as shown later in this chapter, the uncoupled GPLLJ WRF simulations conducted for this study similarly reproduce the GPLLJ as it appears in the ERA-Interim reanalysis and confirm the presence of defining characteristics used to classify a “classic”, uncoupled GPLLJ.

3.2 GPLLJ response to soil moisture

At 2100 UTC 20 July 2017, WRF simulations depict a broad area of 5.0–15.0 m s⁻¹ southerly winds over the central GP at 850 hPa (Fig. 3.1a, b). Concurrently, an 850-hPa geopotential height field analysis reveals a ridge centered over the southern Mississippi River Valley and a trough located over the northern GP resulting in south-southwesterly geostrophic flow over the central GP. Although there are southerly winds over the GP, there is no strong
evidence of a GPLLJ at 850 hPa. The vertical wind shear profile over the GP does not meet the shear requirement for a GPLLJ (not shown). Furthermore, the differences between the wet and dry ensembles’ mean 850-hPa wind forecasts are small and generally less than 1.0 m s\(^{-1}\) (Fig. 3.1c).

At 0000 UTC 21 July 2017, there is still no GPLLJ and 850-hPa winds are between 5.0–15.0 m s\(^{-1}\) (Fig. 3.1d, e). However, after sunset, between 0000 UTC and 0300 UTC, the GPLLJ forms as the 850-hPa wind veers, increases by approximately 5.0 m s\(^{-1}\), and the required GPLLJ vertical wind shear profile is achieved. By 0300 UTC a somewhat discontinuous southerly GPLLJ is analyzed over the GP (Fig. 3.1g, h) with 850-hPa windspeed exceeding 15.0 m s\(^{-1}\) in two areas. A southernmost wind maximum is located in southwestern Texas/northeastern Mexico and a northernmost wind maximum located around the Kansas/Nebraska border. During the early evening hours of 0000 UTC and 0300 UTC, differences between wet and dry ensemble 850-hPa GPLLJ winds (Fig. 3.1f, i) remain negligible (less than 1.0 m s\(^{-1}\)).

At 0600 UTC and 0900 UTC, a well-organized, continuous, southwesterly GPLLJ at 850 hPa extends from Texas to Nebraska (Fig. 3.2a, b, d, e) with the core of the jet located over Kansas. GPLLJ winds reach peak intensity at 0600 UTC. The maximum windspeed develops over dry soils where windspeed exceeds 20.0 m s\(^{-1}\) over Kansas and Nebraska. The GPLLJ is best organized at 0900 UTC. At this time, closed 850-hPa windspeed contours of 15.0 m s\(^{-1}\) span the entire GP. In these early morning hours of 0600 UTC and 0900 UTC, winds veered from 0300 UTC and both wet and dry soil moisture simulations show the GPLLJ in the same location evolving similarly over time. The 850-hPa wind direction between ensembles is also comparable, but windspeed differences greater than 1.0 m s\(^{-1}\) are apparent (Fig. 3.2c, f). The GPLLJ is slower over wet soil compared to dry soil, especially at 0900 UTC. At 0900 UTC, in the GPLLJ
entrance and exit regions, the wet ensemble mean 850-hPa windspeed is approximately 1.0–4.0 m s\(^{-1}\) slower when compared to dry ensemble mean (Fig. 3.2f). GPLLJ 850-hPa windspeed differences are also statistically significant at 0600 UTC and 1200 UTC (Fig. 3.2i) as differences between ensemble mean 850-hPa windspeed are between 1.0–3.0 m s\(^{-1}\). At 1500 UTC (post sunrise), after the PBL has already grown considerably, ensemble differences diminish. Windspeed differences at 850 hPa are < 1.0 m s\(^{-1}\) (not shown). The soil moisture-GPLLJ wind interaction is most detectable in the early morning from 0600–1200 UTC according to this case study.

3.3 Sensible weather response

Sensible weather is considerably influenced by soil moisture prescription. Surface temperature and dewpoint temperature fields exhibit robust differences between the two ensembles over broad regions of the domain at 0900 UTC 21 July 2017 (Fig. 3.3a-f). The wet ensemble mean two-meter temperature ranges from 0.5–2.0 °C cooler over the GP when compared to the dry ensemble. Simultaneously, two-meter dewpoint temperature is 1.0–5.0 °C higher over wet soil in the same region. The tendency for the wet ensemble to be more moist and cooler at the surface than the dry ensemble originates during the mid-morning hours around 1500 UTC (not shown) and carries into the night. After sunrise and starting with identical two-meter temperature and dewpoint temperature fields from model initialization, the temperature over dry soil tends to heat faster than over wet soil (e.g., Ek and Mahrt 1994). Concurrently, the surface dewpoint temperature over dry soil decreases faster or rises slower than the surface dewpoint temperature over wet soil during the daytime in the current simulation. The trend in increasing
differences in surface temperature and dewpoint temperature between the two ensembles ceases around sunset. This suggests that physical processes associated with the daytime insolation and vertical mixing are essential for developing ensemble group temperature and dewpoint temperature heterogeneities.

Although the two-meter temperature and dewpoint temperature fields exhibit noticeable differences between ensembles, 10-meter wind (surface wind) lacks sensitivity to soil moisture conditions. The surface wind does not exhibit any clear difference between ensemble means at any hour of the simulation, including 0900 UTC 21 July 2017 (Fig. 3.3g-i). Windspeed differences are generally less than 1.0 m s\(^{-1}\). There is possibly a weak tendency for surface winds to be stronger over the GP during the daytime and slightly weaker at night over wet soil. However, the differences between ensemble means is small and the \(t\)-test results (not shown) do not exhibit any clear spatial patterns of statistically significant results.

3.4 Severe weather parameters

Differences between the ensembles’ sensible weather and GPLLJ forecasts leads to potentially important changes in parameters associated with severe weather occurrence. In this subsection, the thermodynamic profile is examined using the maximum convective available potential energy (CAPE) and maximum convective inhibition (CIN) at each gridpoint. CAPE and CIN are calculated on the parcel with the highest theta-\(e\) in the column within the first 3000 meters of the ground. Parcel moisture and temperature attributes are then computed by averaging over 500-m around this layer. The kinematic environment is examined using surface to 1-km (0–1-km) shear. For this calculation, the vector difference between the 10-meter wind and the 1-km
above ground level wind is calculated at each gridpoint and then the magnitude of the shear vector is computed. Additional shear calculations, including surface to 3-km (0–3-km) shear are calculated, in a similar manner. The only differences between shear calculations is the uppermost layer of wind being used.

At 0900 UTC 21 July 2017, more CAPE is modelled over wet soils as compared to dry soils (Fig. 3.4a-f). CAPE over wet soil is greater by 100–500 J kg⁻¹ K⁻¹ over most of the GP. Contrasts between wet and dry ensemble CAPE are largest near the exit region of the GPLLLJ in South Dakota, ranging from 500–1500 J kg⁻¹ K⁻¹. Differences in the spatial coverage of CIN between the wet and dry ensembles is not as broad and is fairly noisy. Statistically significant values of less CIN in South Dakota over wet soils is shown and there is more CIN over parts of Nebraska and Texas at the same time. However, over most of the GP, no clear spatial pattern of change in CIN values emerges. The differences between ensemble mean CAPE and CIN first develop during the daytime as heating due to insolation warms the earth’s surface and creates a surface-based mixed layer. Daytime modifications are then carried into the nighttime hours.

Important dissimilarities in 0–1-km shear (Fig. 3.4g-i) are analyzed at 0900 UTC 21 July 2017. There is less 0–1-km shear over wet soil in the GP. Shear values over wet soil are generally lower by 1.0–3.0 m s⁻¹ compared to dry soils. The difference in 0–1-km shear is associated mostly with windspeed differences; the directional shear differences between ensembles are negligible. It is important to note that shear differences do not appear until GPLLLJ windspeed differences form. Over wet soils, the GPLLLJ at 850 hPa is 1.0–3.0 m s⁻¹ slower in the central GP compared to dry soil at 0900 UTC. However, surface winds are generally the same over wet and dry soils. Thus, the differences in GPLLLJ aloft are likely the cause the observed 0–1-km shear differences. Additional measures of vertical shear, such as 0–3-km shear, are also
computed in order to explore the depth of shear and differences in shear between the wet and dry ensembles (not shown). It is found that the vertical shear profile only changes within the first two kilometers above ground level during the early morning hours of 21 July 2017. The upper level winds above 850 hPa, along with the surface winds, are comparable under both soil moisture conditions. The layer of wind that is impacted by soil moisture is between tens of meters above the ground and a couple kilometers.

3.5 Surface fluxes

As previously discussed, during the daytime hours of 20 July 2017, despite having the same synoptic conditions, the surface temperature over wet soil warm more slowly than over dry soil while the dewpoint temperature becomes higher. In this subsection it is proposed that soil moisture content drives contrasts in surface latent and sensible heat flux which causes differences in PBL structures over the two distinct soil moisture prescriptions thereby leading to the aforementioned sensible weather responses.

At 2100 UTC 20 July 2017, which is a representative time of other daylight hours in the GP, there is more surface latent heat flux and less surface sensible heat flux over wet soil (Fig. 3.5a-f) compared to dry soil over almost all of the domain. Differences in sensible and latent heat flux generally range from 10–100 W m². Isolated areas in the southern GP experience differences exceeding 100 W m². The importance of these differences is that greater sensible heating throughout the day over dry soil is leading to higher surface temperatures over dry soil compared to wet soil. Over wet soil, a greater fraction of total energy goes toward latent heating than in the dry soil cases. Concurrently, at 2100 UTC 20 July 2017, PBLH (i.e., PBL height) is
100–700 m higher over dry soil compared to wet soil (Fig. 3.5g-i). Deeper PBL growth over dry soil, which is being driven by increased sensible heating, is leading to enhanced entrainment of relatively dry, free tropospheric air into the daytime mixed layer and causing reductions in surface dewpoint temperature during the daytime. In contrast, over wet soil, less entrainment of dry, free tropospheric air occurs and the ground acts as a moisture source to the PBL causing more latent heating. An increasing surface dewpoint temperature during the daytime is the result.

It is important to note that the differences between PBLH and surface heat fluxes are largest during the daytime hours when comparing between the wet and dry ensembles. Overnight, the differences between the two ensembles are negligible (not shown) as the PBL collapses and surface heat fluxes approach zero in both simulations in the absence of radiational heating. However, the impacts of the daytime surface heat fluxes ultimately manifest themselves as dynamical drivers of the nocturnal GPLLJ wind differences.

3.6 Geostrophic and ageostrophic wind response

Meaningful mass field contrasts start to develop between wet and dry soil from 1500 UTC to 2100 UTC (daytime) 20 July 2017 and continue to strengthen in intensity after surface heat flux and PBLH differences diminish around 0000 UTC 21 July 2017 as evidenced by analysis of 850-hPa geopotential difference maps (not shown). By 0900 UTC 21 July 2017 (Fig. 3.6a-f), the temperature and geopotential height fields at 850 hPa depict cooler temperatures and higher geopotential heights in the GP over wet soil as compared to dry soil. Outside of the GP, the 850-hPa geopotential height and temperature fields are nearly identical. Prior to 0900 UTC the magnitude of the 850-hPa geopotential height differences are limited and generally less than
0.5 dam (not shown). Importantly, by 0900 UTC, the 850-hPa geopotential height fields depict a relative minimum in geopotential height (~146.0–148.0 dam) corresponding to a relative maximum in 850-hPa temperature (~26.0–28.0 °C) over the northern GP indicating that a thermal low is present. The thermal low is deeper over dry soil by approximately 0.5–1.0 dam. More heat accumulates throughout the daytime in the PBL over dry soil and the excess heat persists in the residual layer overnight, which includes the 850-hPa level in the GP. Ultimately, the warmer temperature in the lower troposphere result in a deeper thermal low as represented in the 850-hPa geopotential height field over dry soil.

At 0900 UTC 21 July 2017, dry soil simulations have a broad region of steeper geopotential height gradients compared to wet soil simulations over the GP at 850 hPa due to the geographically isolated enhancement of the thermal low and relatively constant mass fields east of the GP. In order to quantify the impact of the enhanced thermal low over dry soil, the geostrophic wind is calculated on the 850-hPa pressure level. These calculations reveal that the 850-hPa geostrophic wind explains most of the 850-hPa GPLLJ total wind as the wind vectors and location of the GPLLJ closely resembles the geostrophic wind field (Fig. 3.7a, b). When comparing geostrophic windspeed between ensembles, the differing height gradients across the GP result in a slight, mostly non-statistically significant, enhancement of the 850-hPa geostrophic wind over dry soils in the GP of approximately 0.0–2.0 m s⁻¹ (Fig. 3.7c) at 0900 UTC 21 July 2017. While there is some evidence for a slight enhancement of the geostrophic wind across the region over dry soil, the differences in GPLLJ are not fully explained by the differences in geostrophic wind between ensembles. Accordingly, mass field adjustments cannot fully explain the total wind differences in the GPLLJ forecasts and ageostrophic wind differences must be considered.
Past studies have demonstrated that the ageostrophic component of GPLLJ is dynamically explained by the nocturnal decoupling of the PBL and the associated inertial oscillation (e.g., Blackadar 1957). The ageostrophic wind component, which we use as a rough proxy to describe the inertial oscillation, is calculated by subtracting the geostrophic wind from the total wind field at 850 hPa at 0900 UTC 21 July 2017. The contribution of the ageostrophic wind to the total wind is between 5.0–10.0 m s\(^{-1}\) in Texas. Elsewhere, the contribution of the ageostrophic wind to the GPLLJ total wind is < 5.0 m s\(^{-1}\). A statistically significant increase between 1.0–2.0 m s\(^{-1}\) of the 850-hPa ageostrophic wind over dry soil compared to wet soil (Fig. 3.7d-f) is evident over Texas and other isolated regions of the GP. While not statistically significant everywhere, the ageostrophic wind over dry soil is generally greater in magnitude than over wet soil in the vicinity of the GPLLJ. The increase in ageostrophic wind over dry soil compared to wet soil is the same magnitude as the geostrophic wind increase but, as with the geostrophic wind, the ageostrophic wind cannot fully explain the total wind differences. Ageostrophic and geostrophic responses to soil moisture must be taken into account together in order to physically explain the total GPLLJ changes.

While the ageostrophic response does not fully explain the GPLLJ wind response, the increase in the 850-hPa ageostrophic wind over dry soil compared to wet soil is dynamically important. According to Shapiro and Fedorovich (2010), increased diurnal differences in turbulent exchange is dynamically linked to stronger GPLLJs. Here it is hypothesized that the dynamical linkage between the 850-hPa ageostrophic wind differences over wet and dry soil is driven by larger diurnal reductions in turbulent exchange over dry soil compared to wet soil. To support this claim, we attempt to physically relate turbulent exchange to surface heat fluxes and PBL structure. Daytime PBL structure, as represented by PBLH (i.e., PBL height), is related to
surface heat fluxes. Sensible heating in particular, which is influenced by soil moisture, can be important in determining the PBLH (e.g., Ek and Mahrt 1994). Greater sensible heating occurs over dry soil in GP (~10–100 W m⁻²). Greater sensible heating leads to deeper and more rapid increases in PBLH during the day over dry soil compared to wet soil (not shown). The increased rate of PBL growth over dry soil is likely due to deeper and stronger turbulent eddies (i.e., turbulent exchange) resulting from increased convective turbulence over the warmer dry soil surface during the daytime. At night, surface fluxes in both ensembles approach zero in the absence of insolation and turbulent mixing weakens. The PBL collapses after sunset, as evidenced by PBLH values decreasing (not shown). Since surface fluxes and PBLH at night are nearly identical in both ensemble groups, it is hypothesized that nighttime turbulent exchange in each ensemble at night is comparable as well. Thus, maximum daytime values, rather than minimum nighttime values, of turbulent exchange likely determine the diurnal range of turbulent exchange. Over dry soil, the turbulent exchange during the daytime is likely larger than over wet soil. Consequentially, the diurnal difference in turbulent exchange is larger over dry soil. The result is a stronger ageostrophic component to the flow over dry soil in the GP at 850 hPa.

3.7 GPLLJ formation mechanisms

In an effort to build upon past literature discussing the three main mechanisms for GPLLJ formation, the GPLLJ windspeed differences are discussed in relation to the large-scale environment (i.e., Wexler mechanism), diurnal oscillations (i.e., Blackadar mechanism), and zonal temperature gradient across the sloping terrain of the GP (i.e., Holton mechanism). The Wexler (1961) mechanism appears to be synoptically important because the large-scale pressure
and height gradients lead to a synoptic setup that is favorable for GPLLJ development with southerly geostrophic flow in place over the entire forecast period. However, the diurnal cycle of pressure fluctuations of the Bermuda High and time invariant terrain-induced steering cannot explain the diurnal cycle or soil moisture driven GPLLJ wind differences highlighted in this case study. The local geostrophic and ageostrophic response to soil moisture-dependent heating rates and PBL structure are critical contributors.

The Blackadar (1957) mechanism for GPLLJ formation is impacted by altered PBL structure over wet and dry soil. Dry soil experiences more sensible heating in the GP compared to wet soil during the daytime. As a result, the surface layer above dry soil experiences greater temperature gains during the daytime when compared wet soil. The thermodynamic response is for stronger and taller turbulent eddies to develop over dry soil compared to wet soil, as reflected by taller daytime PBL. Upon nightfall, the PBL, along with turbulent exchange, collapse in both simulations. Thus, larger differences in diurnal turbulent exchange occur over dry soil and support the stronger 850-hPa ageostrophic response over dry soil, or a stronger inertial oscillation, to first order.

The Holton (1967) mechanism is impacted by differential heating rates across the terrain in both simulations. Overnight in the northern GP, a deeper thermal low at 850 hPa develops over dry soils in response to warmer conditions. The 850-hPa geostrophic response is a slight increase in geostrophic wind over dry soil due to the thermal low deepening. Ultimately, the geostrophic response and ageostrophic responses constructively interact to result in a stronger GPLLJ at 850 hPa over dry soil. Accordingly, both the Blackadar and Holton mechanisms are crucial to understanding the soil moisture-driven GPLLJ differences illustrated in this case study.
3.8 Severe weather implications

On the morning of 21 July 2017, the most favorable environment for storm development is located in the northeast GP along a northwest-southeast oriented frontal boundary. In the vicinity of this front, a strip of modest to high CAPE values (i.e., > 1000 J kg\(^{-1}\) K\(^{-1}\)), along with low CIN values (i.e., < 250 J kg\(^{-1}\) K\(^{-1}\)) overlap and extend from Minnesota to Illinois (Fig. 3.4a, b, d, e). Low-level forcing for ascent and wind shear profiles are enhanced near the terminus of the GPLLJ. Deep layer shear is also present as the northernmost extent of the upper-level ridge is approached (not shown). Elsewhere in the GP, the absence of deep shear, a strongly capped (i.e., CIN > 250 J kg\(^{-1}\) K\(^{-1}\)), low CAPE (i.e. <1000 J kg\(^{-1}\) K\(^{-1}\)) environment, and lack of any potential forcing depicts an unfavorable environment for substantial severe storm development.

Focusing on the northeast GP, atmospheric profiles generally contain 1000–3000 J kg\(^{-1}\) K\(^{-1}\) of CAPE and CIN values near 100 J kg\(^{-1}\) K\(^{-1}\). Dry soil simulations average 500–1000 J kg\(^{-1}\) K\(^{-1}\) less CAPE than wet soil simulations and have approximately 50–150 J kg\(^{-1}\) K\(^{-1}\) of additional CIN (Fig. 3.4c, f). Thermodynamically, this means that wet soil simulations have a more favorable environment for convection. However, both wet and dry soil simulations have sufficient instability to support severe storm development with CAPE values > 1000 J kg\(^{-1}\) K\(^{-1}\).

In terms of the kinematic environment, the only changes to the atmospheric wind profile between wet and dry simulations occurs around the level of the GPLLJ. Deep layer shear profiles in wet and dry simulations are unchanged, but low-level vertical shear values are greatly impacted by soil moisture due to the GPLLJ forecast differences. In the vicinity of the front in the northeast GP, at the northmost terminus of the GPLLJ, 0–1-km shear values are 10.0–15.0 m s\(^{-1}\) (Fig. 3.4g, h). Dry soil simulations exhibit 1.0–3.0 m s\(^{-1}\) more shear than wet soil just south
of the frontal boundary (Fig. 3.4i). The enhanced low-level shear profile over dry soil, especially as the 15.0 m s$^{-1}$ threshold is approached, is more favorable for severe convection than wet soil.

When taking into account the differences in thermodynamic profiles, and lower-atmospheric kinematic profiles between wet and dry soil simulations in the northeast GP, it is difficult to determine which environment is more favorable for severe convection. While wet soil simulations exhibit a more favorable thermodynamic environment for storms, dry soil simulations have a more favorable shear profile for storms. As one favorable ingredient for severe convection is gained, the other is reduced. The conclusion can at least be made that regional soil moisture anomalies lead to some potentially impactful shifts in the CAPE and shear phase space that governs convective mode and organization, which is extremely important for the forecasting of organized, long-lived severe convective storms (e.g., Rotunno et al. 1988).

3.9 Sensible weather implications

Keeping the scope of the experiment in mind, simulations reveal that dry soil conditions can theoretically result in different societal impacts than wet soil, especially when considering the changes to sensible weather and the GPLLJ that are modelled between ensemble groups. In the dry soil scenario, stress on vegetation, irrigation systems, and surface water reservoirs is increased when compared to the wet soil scenario. The additional stress is caused by hotter and drier boundary layer conditions leading to larger evaporative demand by the atmosphere due to warmer air being able to hold more water (i.e., Clausius-Clapeyron relation). The cooling demands of homes and businesses is larger due to hotter surface temperatures which can lead to increased stress on the electric grid due to heightened energy consumption (e.g., Colombo et al.
1999; Miller et al. 2008). A deeper daytime PBL, along with increased vertical wind shear at night associated with a stronger GPLLJ, may influence aircraft in the lower atmosphere. Severe weather potential is also impacted. These are just a few of the practical real-life impacts that soil moisture could have on society. Many additional societal consequences could likely be discussed which only further emphasizes that the results from this case study may be considered far beyond their statistical and meteorological significance. When the forecast differences over wet and dry soil are placed in point of view of the potential societal impacts, the importance of understanding the impact of soil moisture on the GP weather only grows.
FIG. 3.1. a) WRF derived mean 850-hPa windspeed (shaded; m s$^{-1}$), wind vectors (barbs; m s$^{-1}$) and geopotential height (dashed contours; every 2.0 dam) for the wet ensemble at 2100 UTC 20 July 2017. b) Same as 3.1a except for dry ensemble. c) Difference between wet and dry ensemble mean 850-hPa windspeed (shaded; m s$^{-1}$). Solid green outline denotes regions where difference of means t-test for ensemble 850-hPa windspeed is statistically significant ($p \leq 0.05$). d) Same as 3.1a except at 0000 UTC 21 July 2017. e) Same as 3.1b except at 0000 UTC 21 July 2017. f) Same as 3.1c except at 0000 UTC 21 July 2017. g) Same as 3.1a except at 0300 UTC 21 July 2017. h) Same as 3.1b except at 0300 UTC 21 July 2017. i) Same as 3.1c except at 0300 UTC 21 July 2017. Land is shaded in grey and oceans are white. Values from -1.0 to 1.0 m s$^{-1}$ (small differences) in the difference plots are not shaded.
FIG. 3.2. a) WRF derived mean 850-hPa windspeed (shaded; m s$^{-1}$), wind vectors (barbs; m s$^{-1}$) and geopotential height (dashed contours; every 2.0 dam) for the wet ensemble at 0600 UTC 21 July 2017. b) Same as 3.2a except for dry ensemble. c) Difference between wet and dry ensemble mean 850-hPa windspeed (shaded; m s$^{-1}$). Solid green outline denotes regions where difference of means $t$-test for ensemble 850-hPa windspeed is statistically significant ($p \leq 0.05$). d) Same as 3.2a except at 0900 UTC 21 July 2017. e) Same as 3.2b except at 0900 UTC 21 July 2017. f) Same as 3.2c except at 0900 UTC 21 July 2017. g) Same as 3.2a except at 1200 UTC 21 July 2017. h) Same as 3.2b except at 1200 UTC 21 July 2017. i) Same as 3.2c except at 1200 UTC 21 July 2017. Land is shaded in grey and oceans are white. Values from -1.0 to 1.0 m s$^{-1}$ (small differences) in the difference plots are not shaded.
FIG. 3.3.  a) WRF derived mean 2-meter temperature (shaded; °C) for the wet ensemble at 0900 UTC 21 July 2017. b) Same as 3.3a except for dry ensemble. c) Difference between wet and dry ensemble mean 2-meter temperature (shaded; °C). Solid green outline denotes regions where difference of means $t$-test for 2-meter temperature is statistically significant ($p \leq 0.05$). d) WRF derived mean 2-meter dewpoint temperature (shaded; °C) for the wet ensemble at 0900 UTC 21 July 2017. e) Same as 3.3d but for dry ensemble. f) Same as 3.3c but for 2-meter dewpoint temperature (°C). g) Mean 10-meter windspeed (shaded, m s$^{-1}$) and wind vectors (barbs; m s$^{-1}$) for the wet ensemble at 0900 UTC 21 July 2017. h) Same as 3.3g except for dry ensemble. i) Same as 3.3c but for 10-meter windspeed (m s$^{-1}$). Land is shaded in grey and oceans are white. Values from -0.5 to 0.5 °C (2-meter temperature), -0.5 to 0.5 °C (2-meter dewpoint temperature), and -0.5 to 0.5 m s$^{-1}$ (10-meter windspeed) (small differences) in the difference plots are not shaded.
FIG. 3.4. a) WRF derived mean CAPE (shaded; J kg$^{-1}$ K$^{-1}$) for the wet ensemble at 0900 UTC 21 July 2017. b) Same as 3.4a except for dry ensemble. c) Difference between wet and dry ensemble mean CAPE (shaded; J kg$^{-1}$ K$^{-1}$). Solid green outline denotes regions where difference of means t-test for CAPE is statistically significant ($p \leq 0.05$). d) WRF derived mean CIN (shaded; J kg$^{-1}$ K$^{-1}$) for the wet ensemble at 0900 UTC 21 July 2017. e) Same as 3.4d but for dry ensemble. f) Same as 3.4c but for CIN (J kg$^{-1}$ K$^{-1}$). g) Mean 0–1-km wind shear magnitude (shaded; m s$^{-1}$) and shear vectors (barbs; m s$^{-1}$) for the wet ensemble at 0900 UTC 21 July 2017. h) Same as 3.4g but for dry ensemble. i) Same as 3.4c but for 0–1-km wind shear magnitude (m s$^{-1}$). Land is shaded in grey and oceans are white. Values from -100 to 100 J kg$^{-1}$ K$^{-1}$ (CAPE), -25 to 25 J kg$^{-1}$ K$^{-1}$ (CIN), and -1.0 to 1.0 m s$^{-1}$ (0–1-km shear) (small differences) in the difference plots are not shaded.
FIG. 3.5. a) WRF derived mean surface sensible heat flux (shaded; W m$^{-2}$) for the wet ensemble at 2100 UTC 20 July 2017. b) Same as 3.5a except for dry ensemble. c) Difference between wet and dry ensemble mean sensible heat flux (shaded; W m$^{-2}$). Solid green outline denotes regions where difference of means t-test for ensemble sensible heat flux is statistically significant ($p \leq 0.05$). d) WRF derived mean surface latent heat flux (shaded; W m$^{-2}$) for the wet ensemble at 2100 UTC 20 July 2017. e) Same as 3.5d but for dry ensemble. f) Same as 3.5c but for latent heat flux (W m$^{-2}$). g) Mean planetary boundary layer height (PBLH) (shaded; m) for the wet ensemble at 2100 UTC 20 July 2017. h) Same as 3.5g except for dry ensemble. i) Same as 3.5c but for PBLH (m). Land is shaded in grey and oceans are white. Values from -10 to 10 W m$^{-2}$ (sensible heat flux), -10 to 10 W m$^{-2}$ (latent heat flux), and -100 to 100 m (PBLH) (small differences) in the difference plots are not shaded.
FIG. 3.6. a) WRF derived mean 850-hPa temperature (shaded; °C) for the wet ensemble at 0900 UTC 21 July 2017. b) Same as 3.6a except for dry ensemble. c) Difference between wet and dry ensemble mean 850-hPa temperature (shaded; °C). Solid green outline denotes regions where difference of means $t$-test for ensemble 850-hPa temperature is statistically significant ($p \leq 0.05$). d) Mean 850-hPa geopotential height (shaded; dam) for the wet ensemble at 0900 UTC 21 July 2017. e) same as 3.6d except for dry ensemble) f) Same as 3.6c except for geopotential height (dam). Land is shaded in grey and oceans are white. Values from -0.5 to 0.5 °C (850-hPa temperature) and -0.2 to 0.2 dam (850-hPa geopotential height) (small differences) in the difference plots are not shaded.
FIG. 3.7. a) WRF derived mean 850-hPa geostrophic windspeed (shaded; m s\(^{-1}\)), geostrophic wind vectors (barbs; m s\(^{-1}\)), and geopotential height (dashed contours; every 2.0 dam) for the wet ensemble at 0900 UTC 21 July 2017. b) Same as 3.7a except for dry ensemble. c) Difference between wet and dry ensemble mean 850-hPa geostrophic windspeed (shaded; m s\(^{-1}\)) and geostrophic wind vectors (barbs; m s\(^{-1}\)). Solid green outline denotes regions where difference of means \(t\)-test for ensemble 850-hPa geostrophic windspeed is statistically significant \((p \leq 0.05)\). d) Same as 3.7a but for ageostrophic wind. e) Same as 3.7b but for ageostrophic wind. f) Same as 3.7c but for ageostrophic wind (m s\(^{-1}\)). Land is shaded in grey and oceans are white. Values from -0.5 to 0.5 m s\(^{-1}\) (small differences) in the difference plots are not shaded.
4. Coupled case study

The case study examined in this chapter occurs under different synoptic background conditions than the case study examined in chapter three. The GPLJ in chapter three occurs under an upper-level ridge and exhibits defining features used to describe an uncoupled GPLJ (Uccellini 1980). There is no cold front or trough immediately upstream of the uncoupled GPLJ and upper-level jet stream support for GPLJ formation is absent. In contrast, the case study in presented in this chapter exhibits characteristics of a coupled GPLJ (Uccellini 1980). There is an approaching trough, cold front, and upper-level jet streak providing synoptic support for the coupled GPLJ. Direct comparison between the case study in chapter three and chapter four is conducted in chapter five.

4.1 Synoptic setup

The GPLJ on the morning of 10 June 2012 exhibits defining characteristics of a coupled GPLJ (Uccellini 1980). According to the ERA-Interim reanalysis, a deep trough at 500 hPa was present over the western U.S. and a ridge was present over the central and eastern U.S. (Fig. 2.2). 500-hPa geopotential height and wind analysis reveal a jet streak rounding the base of the western U.S. trough before ejecting into the GP between 0000 UTC and 1200 UTC 10 June 2012 (Fig. 4.1). During this same time period, the GPLJ forms at 850 hPa to the east of 500-hPa trough between 0000 UTC and 0600 UTC 10 June 2012 (Fig. 2.2a, b). At 0600 UTC the northern end of the GPLJ at 850 hPa is located near the exit region of the upper-level jet streak in the vicinity of the U.S./Canada border. The 850-hPa GPLJ is strongest at 0600 UTC 10 June
2012 and weakens by daybreak (1200 UTC) (Fig. 2.2c). Noteworthy is that surface analysis reveals a cold front approaching the GP from the west. A sharp gradient in surface temperature, surface dewpoint temperature, and surface wind shift (shown later) along a trough of low pressure (not shown) is associated with the cold front. Overall, the coupled GPLLJ WRF simulations examined later in this chapter similarly reproduce the GPLLJ and weather conditions depicted in the ERA-Interim reanalysis thereby confirming the presence of defining characteristics used to classify a coupled GPLLJ.

4.2 GPLLJ response to soil moisture

At 2100 UTC 09 June 2012, WRF simulations depict 15.0–20.0 m s⁻¹ southerly winds over portions of the central and northern GP at 850 hPa (Fig. 4.2a, b). At 0000 UTC 10 June 2012, 850-hPa wind analysis reveals a larger area of 15.0–20.0 m s⁻¹ southerly winds over the same region (Fig. 4.2d, e), but shifted slightly further east compared to 2100 UTC. Enlargement of the area enclosed by the 15.0–20.0 m s⁻¹ contours at 0000 UTC signals the 850-hPa winds are strengthening between 2100 UTC and 0000 UTC. The 850-hPa geopotential height field analysis at 2100 UTC and 0000 UTC reveals an eastward progressing trough, associated with a cold front, over the western GP and a ridge centered over the central east coast of the U.S. The resulting southwest to northeast geopotential height gradient over the GP results in southwesterly geostrophic flow. However, during the daytime at 2100 UTC 09 June 2012 and 0000 UTC 10 June 2012, WRF simulations depict southerly flow over the GP at 850 hPa (Fig. 4.2a, b, d, e) indicating winds crossing the contours toward lower heights. This cross-contour flow implies that ageostrophic wind is an important component to the total wind during these daylight hours.
and is likely related to frictional influences of the PBL. At 2100 UTC and 0000 UTC, the differences between the wet and dry ensemble mean 850-hPa windspeed are negligible (<1.0 m s\(^{-1}\); Fig 4.2c, f). Furthermore, the vertical wind shear profile required for a GPLLJ is achieved by 0000 UTC (not shown) as the 850-hPa windspeed increased from 2100 UTC.

At 0300 UTC (after sunset) 10 June 2012, an organized GPLLJ is analyzed over of the central and northern GP at 850 hPa. Windspeed over the GP increased substantially from 0000 UTC. Windspeeds greater than 20.0 m s\(^{-1}\) at 850 hPa span much of the GP (Fig. 4.2g, h). The 850-hPa winds veered from 0000 UTC, becoming southwesterly, indicating that there is no longer substantial cross-geopotential height contour flow as the wind aligns with the geopotential height contours ahead of the approaching trough. At 0300 UTC, contrasts between the wet and the dry ensemble mean 850-hPa GPLLJ windspeed are not spatially organized, but isolated pockets of windspeed differences between 1.0 and 2.0 m s\(^{-1}\) are apparent (Fig. 4.2i). In these isolated areas, the wet ensemble mean windspeed is slower than the dry ensemble mean windspeed.

At 0600 UTC 10 June 2012, the southwesterly GPLLJ, which extends from Texas to Minnesota, east of the trough at 850 hPa reaches peak intensity. The 850-hPa wind analysis reveals the area enclosed by windspeeds greater than 20.0 m s\(^{-1}\) reaches maximum spatial coverage at this time (Fig. 4.3a, b). The GPLLJ starts to weaken by 0900 UTC as the windspeed away from the core of the GPLLJ, located over the Kansas/Nebraska border, starts drop below 20.0 m s\(^{-1}\) (Fig. 4.3d, f). The weakening trend continues as GPLLJ windspeed is generally between 15.0–20.0 m s\(^{-1}\) by 1200 UTC (Fig. 4.3g, h). Overall, the GPLLJ moves eastward between 0600 UTC and 1200 UTC as the 850-hPa trough progresses towards the east. Over this same time period no substantially veering of the 850-hPa winds take place. Importantly, at 0600
UTC, 850-hPa windspeed differences become statistically significant and spatially organized in the GP (Fig. 4.3c). Within the GPLLJ at 0600 UTC, extending almost the entire meridional length of the GP, the wet ensemble mean 850-hPa windspeed is 2.0–4.0 m s$^{-1}$ less than the dry ensemble mean windspeed. The magnitude and spatial coverage of ensemble mean windspeed differences peaks at 0900 UTC (Fig. 4.3f) and 1200 UTC (Fig. 4.3i), characterized by the wet ensemble mean 850-hPa windspeed being 2.0–5.0 m s$^{-1}$ slower than the dry ensemble. Ensemble mean 850-hPa windspeed differences weaken and drop below 1.0 m s$^{-1}$ after 1200 UTC (after sunrise; not shown). Wind direction in the wet and dry ensembles is comparable at all forecast times.

4.3 Sensible weather response

Surface analysis reveals a cold front stretching the meridional extent of the GP at 0900 UTC 10 June 2012. A sharp gradient in two-meter dewpoint temperature is associated with a wind shift and weak two-meter temperature gradient across the front. West of the cold front, in the western GP, dewpoint temperatures are $<10.0 \, ^\circ\text{C}$ (Fig. 4.4d, e). East of the cold front dewpoint temperatures are $>10.0 \, ^\circ\text{C}$ (Fig. 4.4d, e). Along the dewpoint gradient a 10-meter wind shift is apparent. Winds are westerly west of the front and winds are southerly east of the front (Fig. 4.4g, h). Two-meter temperatures west of the front are $<16.0 \, ^\circ\text{C}$ and temperatures east of the front are $<16.0 \, ^\circ\text{C}$ (Fig. 4.4a, b). Additionally, the cold front is associated with a pressure trough as revealed by mean sea level pressure analysis (not shown).

Surface temperature (two-meter temperature) and surface dewpoint temperature (two-meter dewpoint temperature) fields exhibit statistically significant differences between the wet
and dry ensembles at 0900 UTC 10 June 2012. The wet ensemble mean two-meter temperature ranges from 0.5–1.5 °C cooler over the GP when compared to the dry ensemble (Fig. 4.4a-c). Simultaneously, two-meter dewpoint temperature is 0.5–5.0 °C higher over wet soil in the same region (Fig. 4.4d-f). The wet ensemble becomes more moist and cooler at the surface than the dry ensemble during daylight hours (not shown) after starting with identical model initializations. Temperature and dewpoint temperature differences between the wet and dry ensemble peak during the daytime and diminish during the night (not shown).

The surface wind at 10m exhibits some minor differences between wet and dry ensemble means at 0900 UTC 10 June 2012 (Fig. 4.4g-i). The wet ensemble mean surface wind is approximately 0.5–1.0 m s⁻¹ slower over portions of the GP compared to the dry ensemble, but the spatial pattern of the windspeed differences is sporadic. Approximately half of the GP experiences windspeed differences less than 0.5 m s⁻¹ at 0900 UTC. In general, from 0600 UTC to 1200 UTC, the surface windspeed differences between the wet and dry ensembles are characterized by isolated pockets of windspeed differences > 1.0 m s⁻¹ with a large portion the GP exhibiting little to no difference in windspeed (< 1.0 m s⁻¹). However, there appears to be evidence of the wet ensemble tending to have slower surface winds than the dry ensemble between 0600 UTC and 1200 UTC 10 June 2012.

4.4 Severe weather parameters

Differences between wet and dry ensemble sensible weather and GPLLJ forecasts lead to important changes in parameters associated with severe weather (e.g., CAPE, CIN, and vertical wind shear). At 0900 UTC 10 June 2012, more CAPE is modelled over wet soils as compared to
dry soils (Fig. 4.5a-c). CAPE over wet soil is greater by 100–1000 J kg⁻¹ K⁻¹ in the GP east of the approaching cold front. CIN, in the same area (east of the cold front), is lower by approximately 25–150 J kg⁻¹ K⁻¹ over wet soil compared to dry soil (Fig. 4.5d-f). The tendency for wet regions to have more CAPE and less CIN relative to dry regions develops during the daytime and is persists into the nighttime hours.

Important, statistically significant, dissimilarities between wet and dry ensemble 0–1-km wind shear are analyzed at 0900 UTC 10 June 2012 (Fig. 4.5g-i). There is 1.0–4.0 m s⁻¹ less 0–1-km shear over wet soil compared to dry soil in the GP east of the cold front. Generally, decreased 0–1-km shear (~1.0–4.0 m s⁻¹) over dry soil east of the cold front is present from 0600 UTC to 1200 UTC (not shown). In contrast to the nighttime hours, wet and dry ensemble 0–1-km shear differences are generally small (< 1.0 m s⁻¹) and are not statistically significant during the daytime hours (not shown). The 0–1-km shear differences between ensembles are mostly associated with GPLLJ windspeed dissimilarities between ensembles. Surface winds only account for approximately 1.0 m s⁻¹ of the 0–1-km shear differences between ensembles.

However, the 1-km AGL wind, which is approximately at the 850-hPa pressure level over the GP, is faster by 1.0–5.0 m s⁻¹ between 0600 UTC and 1200 UTC (not shown) in the dry soil ensemble to wet soil ensemble. This suggests the GPLLJ wind differences between ensembles are responsible for a bulk of the lower troposphere wind shear differences. Additional deep-layer shear calculations, such as 0–3-km shear, generally do not exhibit clear patterns of statistical significance between wet and dry ensembles (not shown). The vast majority of vertical wind shear discrepancies between the wet and dry ensembles in the GP occurs between tens of meters and two kilometers above ground.
4.5 Surface fluxes

During the daytime hours, differences between wet and dry ensemble surface sensible and latent heat flux lead to surface temperature, surface dewpoint temperature, and PBLH (i.e., PBL height) contrasts between ensembles. At 2100 UTC 09 June 2012 (afternoon), there is more surface latent heat flux over wet soil compared to dry soil (~10–100 W m$^{-2}$; Fig. 4.6d-f). At the same time there is less sensible heat flux (~10–100 W m$^{-2}$) over wet soil compared to dry soil in the same region (Fig. 4.6a-c). In general, during the entire daytime period (not shown), more sensible heat flux in the GP contributes to higher PBL temperatures over dry soil compared to wet soil east of the cold front as more energy goes directly toward warming the air temperature rather than evaporating soil moisture. At night surface heat fluxes are comparable between ensembles (not shown) as surface heat fluxes approach zero in the absence of insolation.

Daytime PBLH differences can be related to surface sensible heat flux differences. At 2100 UTC 09 June 2012, PBLH is 100–400 m lower over wet soil compared to dry soil (Fig. 4.6g-i) east of the cold front. The PBL is deeper over dry soil due to greater sensible surface heating, which leads to increased vertical mixing during the daytime compared to wet soil. Furthermore, deeper PBL growth over dry soil leads to more entrainment of relatively dry free-tropospheric air into the convective boundary layer that further dries out the PBL over dry soil. Over wet soil, less vertical mixing and entrainment of dry air occurs. PBLH differences follow the same diurnal trend of surface heat fluxes. PBLH differences between ensembles are maximized during the daytime and are negligible at night as the PBL collapses without solar insolation (not shown). The impacts of daytime surface heat fluxes on PBL daytime evolution ultimately lead to GPLLLJ wind differences at night because daytime differences in PBL
temperature and height impact the nocturnal mechanisms for GPLLJ formation associated with low-level mass fields and frictional profiles.

4.6 Geostrophic and ageostrophic wind responses

Lower tropospheric mass field contrasts between the wet and dry ensemble start developing during the daytime (~1500 UTC–2100 UTC 09 June 2012) and continue amplifying through the evening (not shown). By 0900 UTC 10 June 2012 (nighttime) statistically significant 850-hPa geopotential height differences between the wet ensemble and dry ensemble develop near a shortwave trough located over the central GP (Fig. 4.7d, e). The 850-hPa geopotential heights over the trough axis and east of trough axis are 0.2–1.0 dam higher over wet soil compared to dry soil (Fig. 4.7f). In other words, the trough is deeper over dry soil in the GP. The 850-hPa geopotential heights outside of the GP exhibit negligible differences between ensembles (< 0.2 dam). The 850-hPa temperature analysis at 0900 UTC depicts a narrow corridor of relatively warm air extending from southwest Texas to the northern GP east of the trough axis (Fig. 4.7a, b). Cooler air is located in the northern GP in the vicinity of the 850-hPa trough indicating that the trough is associated with cold air, not a thermal low. Within the narrow axis of warm air, ensemble mean 850-hPa temperature differences between the wet and dry ensemble is less than 0.5 °C. Just outside the warm air tongue, the wet ensemble mean 850-hPa temperature is 0.5–2.0 °C cooler in the GP compared to the dry ensemble (Fig. 4.7c). Cooler 850-hPa temperatures over the wet ensemble are limited to the GP region and are likely influenced by less daytime surface sensible heat flux over the wet soil compared to dry soil.
The trough depth in the GP has important GPLLJ implications relating to the geopotential height gradient across the GP. At 0900 UTC there is a steeper 850-hPa geopotential height gradient across the GP over dry soil compared to wet soil due to the deeper trough (i.e., lower geopotential heights) in the GP over dry soil. In order to quantify the impact of the deeper 850-hPa trough over dry soil the geostrophic wind is computed at 850 hPa at 0900 UTC 10 June 2012. The 850-hPa geostrophic wind calculations reveal 0.5–4.0 m s\(^{-1}\) faster geostrophic winds over dry soil compared to wet soil, especially over the northern GP (Fig. 4.8a-c). Furthermore, the geostrophic wind explains most of the GPLLJ because the total 850-hPa wind closely matches the 850-hPa geostrophic wind. However, the ageostrophic component of the wind must be taken into account since the geostrophic wind does not explain all GPLLJ differences between ensembles and the geostrophic wind does not perfectly depict the 850-hPa total winds.

The ageostrophic wind is calculated at 0900 UTC 10 June 2012 on the 850-hPa pressure level. Ageostrophic wind calculations reveal 0.5–4.0 m s\(^{-1}\) slower ageostrophic winds over wet soil compared to dry soil in the southern and central GP (Fig. 4.8e-f). In the northern GP the ageostrophic wind differences between ensembles are not organized and the sign of the ageostrophic windspeed differences between the wet and dry ensemble alters between positive and negative. In general, the ageostrophic wind is the most important contributor to the total GPLLJ 850-hPa windspeed differences between wet and dry ensembles over the southern and central GP and results in a faster GPLLJ at night over dry soil. The geostrophic component of the wind is the most important contributor to the faster nocturnal GPLLJ in the northern GP for the dry soil ensemble. Together, the ageostrophic and geostrophic wind comprise the total wind and depict a faster GPLLJ over dry soil compared to wet soil in the majority of the GP.
As discussed in chapter 3, section 6, the ageostrophic wind, which is a rough proxy for the inertial wind, has been related to PBL structure and GPLLJ strength (Shapiro and Fedorovich 2010). Larger diurnal differences in turbulent exchange lead to a stronger GPLLJ via increased ageostrophic windspeeds. The coupled GPLLJ case study results are consistent with this finding. During the daytime there is more sensible heat flux over dry soil than over wet soil. As a result, the surface temperature increases faster and the PBL grows deeper over dry soil compared to wet soil. Increased PBL growth is related to stronger turbulent eddies over dry soil resulting from more sensible heating. At night, after sensible heat flux goes toward zero and the PBL collapses, the vertical mixing within the residual layer in the GP (~850 hPa) rapidly reduces. The nocturnal reduction in vertical mixing east of the cold front is largest over dry soil since vertical mixing is stronger during the day over dry soil compared to wet soil. The enhanced reduction in vertical mixing over dry soil results in a stronger ageostrophic acceleration of the wind overnight and a faster GPLLJ over dry soil.

4.7 GPLLJ formation mechanisms

The synoptic environment is extremely important in driving the coupled GPLLJ of 10 June 2012 but the classic Wexler (1961) mechanism for GPLLJ formation is not active. On the morning of 10 June 2012, a synoptic-scale trough and cold front is approaching the GP. Ahead of the cold front southerly flow is present due to the geopotential height gradient across the GP. The synoptic disturbance is likely the primary forcing mechanism for the northward turning of the tropical trade winds and GPLLJ formation. The Rocky/Sierra Madre Mountains appear to be secondary mechanisms to the trough. Evidence supporting the GPLLJ not being directly tied to
topography is associated with the apparent eastward progression of the GPLLJ during the morning of 10 June 2012. The topography does not move but the GPLLJ moves east along with the trough and cold front.

The Blackadar (1957) mechanism is dynamically important for explaining the diurnal cycle of GPLLJ formation on 10 June 2012 and nocturnal 850-hPa GPLLJ differences between the wet and dry ensembles. The coupled GPLLJ of 10 June 2012 clearly exhibits nocturnal windspeed enhancements. During the daytime the 850-hPa total winds are ~5.0–15.0 m s\(^{-1}\) slower compared to nighttime over most of the GP in both ensembles. Geostrophic and ageostrophic wind analysis reveals that the nocturnal increase of 850-hPa wind cannot be fully explained by the geostrophic wind (not shown). The ageostrophic wind, which is associated with the inertial wind and friction (i.e., Blackadar mechanism), exhibits a vector magnitude change of ~2.5–12.5 m s\(^{-1}\) between the daytime and nighttime in the GP. The ageostrophic wind is easterly ~2.5–7.5 m s\(^{-1}\) during the daytime (opposing the southwesterly geostrophic flow; not shown) and becomes southwesterly ~2.5–5.0 m s\(^{-1}\) (adding to southwesterly geostrophic flow) during the night (Fig. 4.8 d-e). Without the reduction of friction at night and the contributions of southerly/southwesterly ageostrophic wind (related to the inertial wind oscillation due to reduction of friction), the total nocturnal acceleration of the 850-hPa GPLLJ winds cannot be achieved in this case study. The Blackadar mechanism also explains a portion of the coupled GPLLJ windspeed differences between the wet and dry ensembles since ageostrophic windspeed is impacted by soil moisture prescription. Greater diurnal reductions in turbulent mixing over dry soil leads to stronger 850-hPa ageostrophic wind over dry soil compared to wet soil at night. Stronger nocturnal ageostrophic wind over dry soil greatly contributes to the faster GPLLJ over the southern and central GP.
The Holton (1967) mechanism is impacted by soil moisture prescription as evidenced by the geostrophic wind explaining most of the nocturnal GPLLJ windspeed differences between wet and dry ensembles in the northern GP. During the daytime hours of 09 June 2012 differential heating rates across the terrain between the wet and dry soil moisture scenarios start to impact the temperature and geopotential height fields at 850 hPa. By the early morning of 10 June 2012, a deeper trough forms over dry soil leading to stronger geostrophic wind over dry soil compared to wet soil at 850 hPa in the GP. Stronger geostrophic wind leads to a faster GPLLJ over dry soil, especially over the northern GP. Additionally, as previously mention, the Holton mechanism also helps explain the diurnal cycle of GPLLJ formation since the southwesterly geostrophic wind is faster at night compared to during the daytime (~2.5-5.0 m s\(^{-1}\), not shown).

Ultimately the ageostrophic wind (related to the Blackadar mechanism) and the geostrophic wind (related to the Holton mechanism) each contribute 0.5–4.0 m s\(^{-1}\) to the total GPLLJ windspeed differences between ensembles in the GP, although the importance of each component of the wind varies across the GP (Fig. 4.8a-f). The Blackadar mechanism is most important in causing the GPLLJ windspeed differences in the southern GP where ageostrophic windspeed differences between the wet and dry ensembles range from 0.5–4.0 m s\(^{-1}\). The Holton mechanism is most important in the northern GP where the geostrophic windspeed differences between the wet and dry ensembles range from 0.5–4.0 m s\(^{-1}\). Additionally, the geostrophic and ageostrophic winds combine to together lead to ~5.0–15.0 m s\(^{-1}\) total wind difference between the daytime and nighttime at 850 hPa. Additionally, the ageostrophic wind is the larger contributor (~2.5–12.5 m s\(^{-1}\)) to total diurnal wind differences in both ensembles compared to the geostrophic wind (~2.5-5.0 m s\(^{-1}\)).
4.8 Severe weather implications

At 0900 UTC 10 June 2012, the thermodynamic environment throughout most of GP is not favorable for severe weather. Despite the presence of forcing for ascent associated with a cold front approaching from the west, shear associated with the cold front, and CAPE east of the cold front, warm air aloft creates a temperature inversion (i.e., cap) east of the cold front that strongly inhibits convection. The warm air aloft, which is associated with the cap, is well represented by the tongue of higher temperature air in the GP revealed by the 850-hPa temperature analysis at 0900 UTC 10 June 2012 (Fig. 4.7a, b). The warm air aloft leads to high CIN values. Given the warm air and CIN distribution in the GP, the most suitable thermodynamic environment for severe storms exists in the northern GP near the exit region of the GPLLJ where CAPE values exceed 1000 J kg\(^{-1}\) K\(^{-1}\) and mesoscale areas of CIN values below 100 J kg\(^{-1}\) K\(^{-1}\) overlap. The rest of the GP is capped and has CIN values ranging from 100–300 J kg\(^{-1}\) K\(^{-1}\) that are unfavorable for severe storm formation. Additionally, the region of strongest quasi-geostrophic and upper-level jet support for ascent is located in the northern GP. Rising motion due to low-level convergence associated with the northernmost GPLLJ exit region also provides forcing for ascent that can help to overcome the local minimum of CIN in the northern GP.

More specifically, near the GPLLJ exit region in the Dakotas east of the cold front at 0900 UTC 10 June 2012, CAPE values range from 1000–3000 J kg\(^{-1}\) K\(^{-1}\) (Fig. 4.5a, b). In this area, there is 100–1000 J kg\(^{-1}\) K\(^{-1}\) more CAPE over wet soil compared to dry soil (Fig. 4.5c). In this same region, CIN values range from 50–250 J kg\(^{-1}\) K\(^{-1}\) (Fig. 4.5d-f) with pockets of CIN < 100 J kg\(^{-1}\) K\(^{-1}\). CIN values are 25–150 J kg\(^{-1}\) K\(^{-1}\) less over wet soil than dry soil. Overall,
increased CAPE and decreased CIN in the GPLLJ exit region over wet soil indicate that the thermodynamic environment is much more favorable for convection over wet soil; there is greater instability with less convective inhibition.

The lower-tropospheric kinematic environment in the GPLLJ exit region is also impacted by soil moisture prescription. The 0–1-km shear in the Dakotas ranges from 5.0–15.0 m s\(^{-1}\) over wet soil and 5.0–20.0 m s\(^{-1}\) over dry soil (Fig. 4.5g, h). More specifically there is approximately 2.0–5.0 m s\(^{-1}\) more 0–1-km shear over dry soil compared to wet soil (Fig. 4.5i). Hence, there is a more favorable shear profile for severe storms over dry soil compared to dry soil. The 0–1-km shear differences are primarily caused by GPLLJ windspeed differences at 1km above ground level since surface winds are comparable between the wet and dry ensembles. Deep layer shear is not impacted by soil moisture prescription as wind differences between the wet and dry ensembles are generally limited to below two kilometers above ground level.

When considering the thermodynamic and kinematic profile differences between the wet and dry ensembles in the Dakotas it is difficult to determine with a high degree of confidence which environment is most favorable for severe weather. The dry ensemble has a less favorable thermodynamic environment for severe convection than the wet ensemble but has a more conducive lower-tropospheric shear profile for organized convection. On the other hand, the wet ensemble has a more favorable thermodynamic environment for convection with much less convective inhibition than the dry ensemble. It can only be concluded that potentially important shifts in the CAPE and shear phase space occur between wet and dry soil ensembles that can lead to impactful consequences for long-lived organized thunderstorm development and sustenance through changes to convective mode and organization (e.g., Rotunno et al. 1988; Weisman and Rotunno 2004).
4.9 Sensible weather implications

Simulations conducted for the coupled GPLLJ case of 10 June 2012 reveal notably different sensible weather, PBL structure, and nocturnal GPLLJ conditions east of the cold front in the dry and wet soil scenarios. Over dry soil warmer surface temperatures and lower surface dewpoint temperatures east of the cold front may have potential societal impacts. Higher two-meter temperatures could lead to higher electricity consumption due, for example, to the cooling demands of homes and businesses being larger (e.g., Colombo et al. 1999; Miller et al. 2008). Atmospheric evaporative demand is higher in the dry soil simulations since warmer air can hold more water vapor (i.e., Clausius-Clapeyron relation) thereby leading to more evapotranspiration from plants and evaporation from surface water reservoirs. Dry soil simulations are also associated with a deeper daytime PBL and increased nocturnal vertical wind shear due to a stronger GPLLLJ. The PBL and vertical wind shear profiles could potentially impact turbulence felt by aircraft due to increased daytime eddy mixing and increased vertical wind shear at night over dry soil. Overall, results from this coupled GPLLJ case study demonstrate that soil moisture may not only impact meteorologically-important variables under coupled GPLLJ synoptic scenarios, but societal impacts should be considered as well.
FIG. 4.1. ERA-Interim reanalysis derived synoptic environment on (a) 0000 UTC, (b) 0600 UTC, (c) 1200 UTC 10 June 2012. 500-hPa windspeed (shaded; m s$^{-1}$), wind vectors (barbs; half barb is 2.5 m s$^{-1}$, full barb is 5.0 m s$^{-1}$, and flag is 25.0 m s$^{-1}$), and 500-hPa geopotential height (black contours; every 5.0 dam) are shown. Land is shaded in grey and oceans are white.
FIG. 4.2. a) WRF derived mean 850-hPa windspeed (shaded; m s\(^{-1}\)), wind vectors (barbs; m s\(^{-1}\)) and geopotential height (dashed contours; every 2.0 dam) for the wet ensemble at 2100 UTC 09 June 2012. b) Same as 4.2a except for dry ensemble. c) Difference between wet and dry ensemble mean 850-hPa windspeed (shaded; m s\(^{-1}\)). Solid green outline denotes regions where difference of means \(t\)-test for ensemble 850-hPa windspeed is statistically significant (\(p \leq 0.05\)). d) Same as 4.2a except at 0000 UTC 10 June 2012. e) Same as 4.2b except at 0000 UTC 10 June 2012. f) Same as 4.2c except at 0000 UTC 10 June 2012. g) Same as 4.2a except at 0300 UTC 10 June 2012. h) Same as 4.2b except at 0300 UTC 10 June 2012. i) Same as 4.2c except at 0300 UTC 10 June 2012. Land is shaded in grey and oceans are white. Values from -1.0 to 1.0 m s\(^{-1}\) (small differences) in the difference plots are not shaded.
FIG. 4.3. a) WRF derived mean 850-hPa windspeed (shaded; m s$^{-1}$), wind vectors (barbs; m s$^{-1}$) and geopotential height (dashed contours; every 2.0 dam) for the wet ensemble at 0600 UTC 10 June 2012. b) Same as 4.3a except for dry ensemble. c) Difference between wet and dry ensemble mean 850-hPa windspeed (shaded; m s$^{-1}$). Solid green outline denotes regions where difference of means t-test for ensemble 850-hPa windspeed is statistically significant ($p \leq 0.05$). d) Same as 4.3a except at 0900 UTC 10 June 2012. e) Same as 4.3b except at 0900 UTC 10 June 2012. f) Same as 4.3c except at 0900 UTC 10 June 2012. g) Same as 4.3a except at 1200 UTC 10 June 2012. h) Same as 4.3b except at 1200 UTC 10 June 2012. i) Same as 4.3c except at 1200 UTC 10 June 2012. Land is shaded in grey and oceans are white. Values from -1.0 to 1.0 m s$^{-1}$ (small differences) in the difference plots are not shaded.
FIG. 4.4. a) WRF derived mean 2-meter temperature (shaded; °C) for the wet ensemble at 0900 UTC 10 June 2012. b) Same as 4.4a except for dry ensemble. c) Difference between wet and dry ensemble mean 2-meter temperature (shaded; °C). Solid green outline denotes regions where difference of means $t$-test for 2-meter temperature is statistically significant ($p \leq 0.05$). d) WRF derived mean 2-meter dewpoint temperature (shaded; °C) for the wet ensemble at 0900 UTC 10 June 2012. e) Same as 4.4d but for dry ensemble. f) Same as 4.4c but for 2-meter dewpoint temperature (°C). g) Mean 10-meter windspeed (shaded, m s$^{-1}$) and wind vectors (barbs; m s$^{-1}$) for the wet ensemble at 0900 UTC 10 June 2012. h) Same as 4.4g except for dry ensemble. i) Same as 4.4c but for 10-meter windspeed (m s$^{-1}$). Land is shaded in grey and oceans are white. Values from -0.5 to 0.5 °C (2-meter temperature), -0.5 to 0.5 °C (2-meter dewpoint temperature), and -0.5 to 0.5 m s$^{-1}$ (10-meter windspeed) (small differences) in the difference plots are not shaded.
FIG. 4.5. a) WRF derived mean CAPE (shaded; J kg$^{-1}$ K$^{-1}$) for the wet ensemble at 0900 UTC 10 June 2012. b) Same as 4.5a except for dry ensemble. c) Difference between wet and dry ensemble mean CAPE (shaded; J kg$^{-1}$ K$^{-1}$). Solid green outline denotes regions where difference of means $t$-test for CAPE is statistically significant ($p \leq 0.05$). d) WRF derived mean CIN (shaded; J kg$^{-1}$ K$^{-1}$) for the wet ensemble at 0900 UTC 10 June 2012. e) Same as 4.5d but for dry ensemble. f) Same as 4.5c but for CIN (J kg$^{-1}$ K$^{-1}$). g) Mean 0–1-km wind shear magnitude (shaded; m s$^{-1}$) and shear vectors (barbs; m s$^{-1}$) for the wet ensemble at 0900 UTC 10 June 2012. h) Same as 4.5g but for dry ensemble. i) Same as 4.5c but for 0–1-km wind shear magnitude (m s$^{-1}$). Land is shaded in grey and oceans are white. Values from -100 to 100 J kg$^{-1}$ K$^{-1}$ (CAPE), -25 to 25 J kg$^{-1}$ K$^{-1}$ (CIN), and -1.0 to 1.0 m s$^{-1}$ (0–1-km shear) (small differences) in the difference plots are not shaded.
FIG. 4.6. a) WRF derived mean surface sensible heat flux (shaded; W m$^{-2}$) for the wet ensemble at 2100 UTC 09 June 2012. b) Same as 4.6a except for dry ensemble. c) Difference between wet and dry ensemble mean sensible heat flux (shaded; W m$^{-2}$). Solid green outline denotes regions where difference of means t-test for ensemble sensible heat flux is statistically significant ($p \leq 0.05$). d) WRF derived mean surface latent heat flux (shaded; W m$^{-2}$) for the wet ensemble at 2100 UTC 09 June 2012. e) Same as 4.6d but for dry ensemble. f) Same as 4.6c but for latent heat flux (W m$^{-2}$). g) Mean planetary boundary layer height (PBLH) (shaded; m) for the wet ensemble at 2100 UTC 09 June 2012. h) Same as 4.6g except for dry ensemble. i) Same as 4.6c but for PBLH (m). Land is shaded in grey and oceans are white. Values from -10 to 10 W m$^{-2}$ (sensible heat flux), -10 to 10 W m$^{-2}$ (latent heat flux), and -100 to 100 m (PBLH) (small differences) in the difference plots are not shaded.
FIG. 4.7. a) WRF derived mean 850-hPa temperature (shaded; °C) for the wet ensemble at 0900 UTC 10 June 2012. b) Same as 4.7a except for dry ensemble. c) Difference between wet and dry ensemble mean 850-hPa temperature (shaded; °C). Solid green outline denotes regions where difference of means $t$-test for ensemble 850-hPa temperature is statistically significant ($p \leq 0.05$). d) Mean 850-hPa geopotential height (shaded; dam) for the wet ensemble at 0900 UTC 10 June 2012. e) same as 4.7d except for dry ensemble) f) Same as 4.7c except for geopotential height (dam). Land is shaded in grey and oceans are white. Values from -0.5 to 0.5 °C (850-hPa temperature) and -0.2 to 0.2 dam (850-hPa geopotential height) (small differences) in the difference plots are not shaded.
FIG. 4.8. a) WRF derived mean 850-hPa geostrophic windspeed (shaded; m s\(^{-1}\)), geostrophic wind vectors (barbs; m s\(^{-1}\)), and geopotential height (dashed contours; every 2.0 dam) for the wet ensemble at 0900 UTC 10 June 2012. b) Same as 4.8a except for dry ensemble. c) Difference between wet and dry ensemble mean 850-hPa geostrophic windspeed (shaded; m s\(^{-1}\)) and geostrophic wind vectors (barbs; m s\(^{-1}\)). Solid green outline denotes regions where difference of means \(t\)-test for ensemble 850-hPa geostrophic windspeed is statistically significant (\(p \leq 0.05\)).

d) Same as 4.8a but for ageostrophic wind. e) Same as 4.8b but for ageostrophic wind. f) Same as 4.8c but for ageostrophic wind (m s\(^{-1}\)). Land is shaded in grey and oceans are white. Values from -0.5 to 0.5 m s\(^{-1}\) (small differences) in the difference plots are not shaded.
5. Discussion, conclusions, and suggestions for future work

The goals of this thesis were to: 1) quantify the impact of wet and dry soil moisture anomalies on the GPLLJ, sensible weather, and severe weather parameters, 2) dynamically link any apparent changes to the GPLLJ, sensible weather, and severe weather parameters to soil moisture conditions, and 3) assess the potential societal importance of any modifications to the GPLLJ, sensible weather, and severe weather parameters associated with wet and dry soil moisture conditions. In order to investigate the role of soil moisture on the GPLLJ and associated weather this thesis uses WRF 3.9.1.1 to simulate a pair of GPLLJ case studies that occur under two contrasting synoptic background conditions that are common in the GP during the summertime. The first GPLLJ case study presented in this thesis is considered a “classic”, uncoupled GPLLJ in which the GPLLJ forms due to local land and PBL processes under an upper level ridge (Uccellini 1980). The second case presented in this thesis is considered a coupled GPLLJ in which the GPLLJ is strongly influenced by synoptic forcing and forms east of an approaching trough and cold front (Uccellini 1980). The GPLLJ sensitivity to soil moisture is tested by initializing WRF with two distinct soil moisture anomaly profiles for each case study. The 95\textsuperscript{th} percentile of soil moisture values from the modelled 1982–2016 daily warm season climatology represents wet soil conditions. The 5\textsuperscript{th} percentile represents dry soil conditions. Four additional uncoupled and four additional coupled case studies were conducted in a similar manner to the case studies presented in this thesis. Overall, the results from these additional cases are comparable with the cases presented in this thesis.

5.1 Discussion and conclusions
The two case studies examined in this thesis indicate that the GPLLJ can be impacted by regional soil moisture conditions consistent with prior literature (McCorcle 1988; Fast and McCorcle 1990; Zhong et al. 1996). In the uncoupled GPLLJ case study, between 0600 UTC–1200 UTC 21 July 2017, GPLLJ 850-hPa winds are modelled to be 1.0–3.0 m s\(^{-1}\) slower in the entrance and exit regions of the GPLLJ over wet soil when compared to dry soil. In the coupled GPLLJ case study, between 0600 UTC–1200 UTC 10 June 2012, coupled GPLLJ winds are modelled to be 2.0–5.0 m s\(^{-1}\) slower over the entire GPLLJ over wet soil when compared to dry soil. 850-hPa windspeed differences prior to 0600 UTC, and after 1200 UTC, are generally not statistically different and are less than 1.0 m s\(^{-1}\) for the uncoupled and coupled GPLLJ scenarios. Both case studies have GPLLJs that exhibit a distinct diurnal signal and peak in intensity overnight. Winds in the uncoupled GPLLJ veer overnight but GPLLJ winds do not veer overnight in the coupled case since the GPLLJ is strongly influenced by an eastward-progressing trough. Overall, despite having different synoptic conditions, the uncoupled and coupled GPLLJs are impacted similarly by soil moisture prescription. Both GPLLJs are most impacted by soil moisture conditions during the overnight hours between 0600 UTC and 1200 UTC with a slower GPLLJ occurring over wet soil. It should be noted that the entire coupled GPLLJ is slower over wet soil compared to only the entrance and exit regions of the uncoupled GPLLJ. Further examination of eight additional case studies reveals that the core of the GPLLJ does not exhibit clear patterns of sensitivity to soil moisture prescription regardless of if the GPLLJ is considered coupled or uncoupled (not shown). It has not been determined what causes the core of the GPLLJ to be impacted in some cases studies and not others.
Sensible weather and severe weather parameters are impacted by soil moisture prescription in the uncoupled and coupled GPLLJ case studies. WRF simulations depict warmer two-meter temperatures (> 0.5 °C) and lower two-meter dewpoint temperatures (< 0.5 °C) developing during the day and lasting into the night over dry soil compared to wet soil throughout the entire GP in the uncoupled GPLLJ case study and east of the cold front in the coupled GPLLJ case study. Unlike temperature and dewpoint temperature, ten-meter wind is generally not impacted by soil moisture in either case study (wind differences between wet and dry ensembles < 1.0 m s⁻¹). Changes to lower tropospheric temperature lead to CAPE and CIN profile modifications over wet and dry soil. In the uncoupled GPLLJ case study, more CAPE (~100–1000 J kg⁻¹ K⁻¹) is modelled over wet soil compared to dry soil in the GP and the spatial coverage and sign of CIN differences are fairly noisy. However, in the coupled GPLLJ case study, clear spatial patterns of decreased CIN (~25–150 J kg⁻¹ K⁻¹) coincide with increases in CAPE (~100–1000 J kg⁻¹ K⁻¹) over wet soil east of the cold front in the GP. Both simulations experience increases in CAPE over wet soil but the spatial patterns of decreased/increased CIN is dependent on the background synoptic conditions. Furthermore, 0–1-km shear is decreased over wet soil compared to dry soil (~1.0–4.0 m s⁻¹) in both case studies in the vicinity of the GPLLJ due to reductions in GPLLJ windspeed over wet soil.

The dynamical drivers of the GPLLJ, sensible weather, and severe weather parameter differences in the uncoupled and coupled case studies over wet and dry soil prescriptions are daytime surface sensible and latent heat fluxes. Despite comparable amounts of downward radiation (shortwave and longwave) reaching the surface in the uncoupled and coupled GPLLJ case studies (not shown), greater sensible heat flux ranging 10–100 W m⁻², along with less latent heat flux ranging 10–100 W m⁻², is modelled over dry soil compared to wet soil throughout the
It is hypothesized that deeper and stronger turbulent eddies form over dry soil compared to wet soil in response to greater sensible heat flux. Stronger turbulent eddies cause increased daytime PBL growth over dry soil and help to dry the PBL via entrainment of free tropospheric air. Accordingly, in the uncoupled G PLLJ case study, a hotter, drier, and deeper PBL results over the entire GP. In the coupled G PLLJ case study a hotter, drier, and deeper PBL is evident especially in areas east of the cold front. Daytime PBL height and temperature modifications then eventually lead to G PLLJ windspeed differences in both cases studies.

In the uncoupled G PLLJ case study, warmer lower-tropospheric conditions over dry soil leads to enhanced thermal low development overnight in the northern GP as evidenced by 850-hPa mass and temperature field analysis. The mass field response to the deeper thermal low over dry soil increases the 850-hPa geostrophic wind in the GP by 0.0–2.0 m s\(^{-1}\) between 0600 UTC–1200 UTC 21 July 2017. The 850-hPa ageostrophic flow is also enhanced by 0.0–2.0 m s\(^{-1}\) at the same time due to a larger diurnal difference in turbulent exchange in the GP over dry soil when compared to wet soil. Increased geostrophic and ageostrophic wind ultimately leads to a stronger G PLLJ over dry soil on the morning of 21 July 2017. Although a thermal low is not present in the coupled G PLLJ case study (cold air is associated with the 850-hPa trough axis) and the dynamics driving the 850-hPa trough are slightly different, lower-tropospheric conditions over dry soil drive a mass field response that results in a deeper trough over the GP during the night on 10 June 2012. The 850-hPa mass field response to the deeper trough drives 850-hPa geostrophic wind increases over dry soil by 0.5–4.0 m s\(^{-1}\) between 0600 UTC–1200 UTC 10 June 2012 in the northern and central GP. The 850-hPa ageostrophic flow is also enhanced by 0.5–4.0 m s\(^{-1}\) over dry soil mainly in the southern GP (related to diurnal mixing differences). Similar to the uncoupled G PLLJ case study, investigating both the geostrophic and ageostrophic
wind components is critical to interpreting the total GPLLJ windspeed differences between the wet and dry ensembles in the coupled GPLLJ scenario.

Overall, it has been shown that soil moisture can indeed impact the Holton (related to geostrophic wind) and Blackadar mechanisms (related to ageostrophic wind) for GPLLJ formation as well as sensible weather and parameters associated with severe weather occurrence via surface sensible heat fluxes that are a function of soil moisture. Additionally, results from this thesis confirm the importance of accurate soil moisture (e.g., Koster et al. 2004; Guo et al. 2006; Koster et al. 2006) to GP weather forecasts. Soil moisture appears to be an integral part of the weather in the GP. The case studies presented in this thesis demonstrate potential soil moisture modifications to short-term weather and weather-related impacts forecasts. The proper modeling and observation of soil moisture may be a crucial piece of accurate weather and climate prediction in the future.

5.2 Suggestions for future work

In order to further examine the importance of soil moisture on the GPLLJ, future work can leverage the experimental design used in this thesis to investigate sensitivity of the GPLLJ to less extreme soil moisture scenarios and isolate areas of soil moisture that drive GPLLJ forecast differences between ensembles. Investigation of a variety of soil moisture scenarios (e.g., 75th vs 25th percentiles, 60th vs 40th percentiles, 90th and 80th percentiles, etc.) is needed to quantify a threshold of how large soil moisture values must differ in order to result in altered GPLLJ forecasts. Determination of a soil moisture difference threshold could help determine the degree of accuracy needed for soil moisture initializations in future weather and climate simulations.
Furthermore, the current study does not geographically isolate the most important regions of soil moisture for the GPLLJ forecasts. Future work could focus on prescribing soil moisture scenarios in different areas inside the current domain (such as within the GP only) in order to determine dynamically important regions of soil moisture for the GPLLJ. Determination of important soil moisture source regions for GPLLJ forecasts would help build dynamical understanding of how regional soil moisture processes impact the local Blackadar and Holton mechanisms compared to the background synoptic state.

The experimental design used in this thesis could also be expanded to include investigation into the roles of different model physics parameterizations (incl., PBL, land surface, microphysics, cumulus scheme), horizontal model resolution, data assimilation, and vegetative growth state on the GPLLJ. Creating WRF ensembles by varying physics parameterizations would further help quantify the amount of GPLLJ variability, in addition to SKEBS perturbations used in this thesis, attributed to current physical/numerical modelling uncertainty. Increasing the horizontal model resolution would be particularly helpful in resolving the impact of convection on the GPLLJ and removing the need for a convective parameterization. Lastly, the impact of soil moisture data assimilation and vegetative growth state on the GPLLJ has yet to fully be determined.

Significant modifications to the current experimental blueprint can also be explored. More specifically, new experimental designs allowing for longer simulations using global models such as The Model for Prediction Across Scales (i.e., MPAS) can be considered. The overall downstream impacts or long-term consequences of soil moisture is challenging to attain using the current experimental design. The current framework is ideal for short-term forecasts (~1–3 days), but the overall ability to drift from the boundary conditions is limited for medium
and long-range forecasts beyond three days. Implementation of a global model would remove the horizontal boundary condition constraint and allow the investigation of possible global-scale circulation impacts resulting from GP soil moisture conditions.

Regardless of the modeling approaches that may be used in the future, building further understanding of how local-land processes interact with the atmospheric syntonic environment should be pursued. Outstanding issues that could not be addressed in this thesis, such determining why the GPLLJ core is sometimes impacted by soil moisture prescription, remain left to be determined. Answering outstanding questions would be greatly beneficial to society since the GPLLJ is an integral part of the weather/climate system in the GP and helps to sustain the food supply and economic well-being of millions of people. Improved ability to make accurate sub-seasonal and seasonal forecasts of GPLLJ speed, frequency, northward penetration, and structure would greatly help society prepare for future weather and climate states.
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