Nocturnal Destabilization Associated with the Summertime Great Plains Low-Level Jet

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ABSTRACT: The Great Plains low-level jet (LLJ) has long been associated with summertime nocturnal convection over the central Great Plains of the United States. Destabilization effects of the LLJ are examined using composite fields assembled from the North American Mesoscale Forecast System for June and July 2008–12. Of critical importance are the large isobaric temperature gradients that become established throughout the lowest 3 km of the atmosphere in response to the seasonal heating of the sloping Great Plains. Such temperature gradients provide thermal wind forcing throughout the lower atmosphere, resulting in the establishment of a background horizontal pressure gradient force at the level of the LLJ. The attendant background geostrophic wind is an essential ingredient for the development of a pronounced summertime LLJ. Inertial turning of the ageostrophic wind associated with LLJ provides a westerly wind component directed normal to the terrain-induced orientation of the isotherms. Hence, significant nocturnal low-level warm-air advection occurs, which promotes differential temperature advection within a vertical column of atmosphere between the level just above the LLJ and 500 hPa. Such differential temperature advection destabilizes the nighttime troposphere above the radiatively cooled near-surface layer on a recurring basis during warm weather months over much of the Great Plains and adjacent states to the east. This destabilization process reduces the convective inhibition of air parcels near the level of the LLJ and may be of significance in the development of elevated nocturnal convection. The 5 July 2015 case from the Plains Elevated Convection at Night field program is used to demonstrate this destabilization process.

KEYWORDS: Ageostrophic circulations; Atmospheric circulation; Dynamics; Jets

1. Introduction

The Great Plains low-level jet (LLJ) was a major topic of research as part of the Plains Elevated Convection at Night (PECAN) field study, which was conducted from 1 June to 15 July 2015 (e.g., Geerts et al. 2017). The LLJ is a nocturnal wind speed maximum that develops in the lowest kilometer of the atmosphere during warm-season months. Winds associated with the LLJ are predominantly from the south, with peak wind speeds often in excess of 20 m s⁻¹. LLJ frequency is highest over the southern Great Plains states of Texas, Oklahoma, and Kansas, with a maximum over northwestern Oklahoma and southern Kansas (Bonner 1968).

Nocturnal summertime convection, the focus of PECAN, is prevalent over an extended region of the Great Plains from Texas northward into Canada. More than 60% of summertime rainfall over the central Great Plains occurs at night (Kincer 1916) and is principally in the form of convective storms (e.g., Bleeker and Andre 1951; Means 1954; Wallace 1975; Balling 1985; Riley et al. 1987; Dai et al. 1999; Carbone and Tuttle 2008). The highest frequency of occurrence of convection during summertime months is situated in the north-central Great Plains states with a maximum in northeastern Kansas and southeastern Nebraska (see Fig. 1 in Geerts et al. 2017 or Fig. 2 in Weckwerth et al. 2019). The geographical preference of such convective features suggests a recurring mechanism at work in the destabilization of the nocturnal atmosphere.

The role of the LLJ in nocturnal convection has been a topic of discussion for many years. Means (1952) provides one of the first studies that link summertime nocturnal convection over the central Great Plains with the LLJ, emphasizing the cross isobar-isotherm patterns at 2000 m representing temperature advection by the geostrophic wind. Of relevance is his discussion regarding differential advection in the vertical by which warm-air advection decreases with height such as to destabilize the atmosphere. Since then, there have been ample studies noting the association between the warm-air advection that accompanies the LLJ and nocturnal convection (e.g., Bonner 1966; Maddox 1983; Stellen and Gallus 2017; Gebauer et al. 2018; Smith et al. 2019; Weckwerth et al. 2019; Weckwerth and Romatschke 2019).

Recently, Rattray et al. (2018) and Parsons et al. (2019) have also linked the LLJ to nocturnal changes in stability. Both note the importance of warm-air advection above the level of the LLJ where they also find a local maximum in the westerly flow. Figure 3 from the Parsons et al. (2019) illustrates that convective inhibition (CIN) displays a vertical decrease at night as

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determined from a composite of 3-h soundings from the Department of Energy’s Atmospheric Radiation Measurement (ARM) Program. Both Rattray et al. (2018) and Parsons et al. (2019) point to differential temperature advection as the key element affecting the destabilization of the nocturnal atmosphere above the radiatively cooled surface layer.

Destabilization preconditions the atmosphere for a number of convection initiation possibilities, some of which can themselves be associated with the LLJ. For example, several studies suggest mechanical forcing, through convergence, is an important link between the LLJ and nocturnal convection. Pitchford and London (1962) propose that regions of convergence associated with spatial variations in the speed of the LLJ are critical in the formation of convective storms. Similarly, Sangster (1979) notes the importance of the surface geostrophic wind in forecasting nocturnal convection. Patterns of isolachs and streamlines are assembled to provide a picture of the environment in which convection develops. A region of streamline diffuence at the northern edge of the LLJ is tied to areas of convection. This linkage of the northern terminus of the LLJ with convection has been a common theme in work that follows (e.g., Maddox 1983; Trier and Parsons 1993). Convergence at the northern terminus is thought to drive upward motion in a manner similar to that proposed by Pitchford and London (1962; see also Balling 1985; Easterling and Robinson 1985; Augustine and Caracena 1994; Tuttle and Davis 2006; Trier et al. 2010; Coniglio et al. 2010; Weckwerth et al. 2019; Weckwerth and Romatschke 2019).

While considerable work links the northern terminus of the LLJ with nocturnal convection, some researchers associate nighttime storms with the anticyclonic shear (eastern) side of the jet (e.g., Walters and Winkler 2001). Pu and Dickinson (2014) suggest that vorticity tendencies associated with the LLJ lead to vertical motions that can explain the observed pattern of the convection over the Great Plains. Based on results from their 20-yr climatology, Reif and Bluestein (2017) note that 24% of nocturnal convection occurs in the presence of a LLJ without nearby surface boundaries. Convection during these episodes tends to occur later on during the night [~0900 UTC or 0400 central daylight time (CDT)] and initiate on the anticyclonic shear side of the LLJ. Those authors note that the highest percentage of strong jets are found with these non-boundary cases. Gebauer et al. (2018) identified three cases from PECAN that match the no-boundary classification in Reif and Bluestein (2017). Results suggest that convergence of the east–west component of motion along the eastern edge of the LLJ helps provide necessary lift to initiate convection (see also Weckwerth et al. 2019).

The purpose of this paper is to extend the previous work by Means (1952) and the recent work by Rattray et al. (2018) and Parsons et al. (2019) by describing the process by which the summertime LLJ is linked to atmospheric destabilization. Herein, we show, fundamentally, why cross isobar–isotherm patterns as discussed by Means (1952) are tied to the LLJ and how the LLJ environment promotes differential temperature advection that serves to destabilize the nocturnal atmosphere on a recurring basis throughout the warm season over the central Great Plains. Similar to that shown in the Parsons et al. (2019), we demonstrate that such a destabilization process leads to a nighttime diminution of the CIN for air parcels above the LLJ, thus preconditioning the atmosphere for nocturnal convection above the radiatively cooled near-surface layer east of about 100°W. Further, we show that the recurrent elevated nighttime warm-air advection also acts to suppress surface-based convection over the central Great Plains the following day throughout the warm-season period.

2. Forcing of the LLJ

Over the past 60 years, numerous studies have been conducted on the Great Plains LLJ using observations (e.g., Bonner 1966; Parish et al. 1988; Stensrud 1996; Whiteman et al. 1997; Banta et al. 2002), modeling (e.g., Holton 1967; McNider and Pielke 1981; Fast and McCorele 1990; Ting and Wang 2006; Shapiro and Fedorovich 2009; Du and Rotunno 2014; Shapiro et al. 2016; Parish and Clark 2017), and analyses of gridded model output (e.g., Walters and Winkler 2001; Parish 2017).

Development of the nocturnal wind maximum is thought to result largely from processes described in Blackadar (1957) and Holton (1967). According to Blackadar (1957), the LLJ maximum results from an inertial oscillation of the ageostrophic component of the wind in response to the rapid decoupling from the surface layer around sunset. Wind speeds increase during the nighttime hours and reach a maximum typically a few hours after local midnight when the ageostrophic and geostrophic wind components become aligned along a common axis. This so-called Blackadar mechanism has been widely acknowledged to be a significant factor in the development of the nocturnal wind maximum that we know as the LLJ (e.g., Markowski and Richardson 2010; Shapiro et al. 2016 and references contained therein).

Some criticism has been leveled on Blackadar’s physical construct since it does not explain why the maximum LLJ frequency is found over the southern Great Plains states of Oklahoma and Kansas (Holton 1967; Bonner 1968; McNider and Pielke 1981; Jiang et al. 2007; Shapiro and Fedorovich 2009; Du and Rotunno 2014; Fedorovich et al. 2017; Gebauer et al. 2018). Holton (1967) develops a one-dimensional model incorporating the role of diurnal heating over sloping terrain representative of the Great Plains. Results from that work suggest that the diurnal heating when combined with the Blackadar (1957) mechanism can produce a LLJ structure and timing consistent with observations (see also Sangster 1967; Bonner and Paegle 1970; Jiang et al. 2007; Du and Rotunno 2014; Shapiro et al. 2016).

Parish (2017) takes a different approach in terms of understanding reasons for the LLJ geographical preference of the Great Plains. The central theme in that work is the critical importance of the background horizontal pressure gradient (PGF) and corresponding geostrophic wind at the level of the jet. Maximum summertime LLJ frequency over northern Oklahoma, such as shown in Bonner (1968), is simply a reflection of the maxima in the southerly geostrophic wind and thus the strongest background flow. Analyses of operational model output by Parish (2017) show that the fundamental difference between strong and weak LLJ episodes is the...
magnitude of the geostrophic wind at the level of the jet. Cases for strong LLJ episodes (maximum 0900 UTC or 0400 CDT winds exceeding 20 m s$^{-1}$) over the Oklahoma/Kansas region showed an 875-hPa geostrophic wind from the south at approximately 15 m s$^{-1}$; summertime episodes with no LLJ experienced geostrophic winds less than 5 m s$^{-1}$ (see Fig. 1 in Parish 2017). Existence of a strong southerly geostrophic wind provides the necessary base current in which the nocturnal wind maximum forms. The importance of the background geostrophic wind has been noted in the past by a number of investigators (e.g., Wexler 1961; Augustine and Caracena 1994; Shapiro et al. 2016). It is this background PGF and the surrounding environment that is the focus of the following discussion.

Fundamentally, the generation of the background geostrophic wind occurs via the same process as that considered by Sangster (1967) and Holton (1967), but over a considerably longer time scale. Parish (2017) notes that seasonal heating of the sloping terrain over the central United States is responsible for establishment of background isobaric temperature gradients in the lowest few kilometers of the atmosphere over the Great Plains during the warm weather months. The corresponding background PGF, and thermal wind forcing, then requires the establishment of a strong background geostrophic wind (e.g., Fig. 4 in Parish 2017).

To illustrate development of this background geostrophic wind, composite fields were assembled using gridded output from the North American Mesoscale Forecast System (NAM) for the months of June and July for the 5-yr period 2008–12. Grids used here are taken from the 0000 UTC run and are available every 3 h at 25-hPa increments. NAM grids have a horizontal resolution of 12 km. Output from the NAM for a grid point at 36.4°N, 98.6°W, near the center of the highest LLJ frequency (see Fig. 2 in Bonner 1968), was used to isolate “strong LLJ” (SJ) cases. Following Parish (2017), we define an SJ case as one in which a maximum wind speed of 20 m s$^{-1}$ or more was present in the lowest 1000 m above ground level (AGL) at 0900 UTC and with a 600-hPa wind less than 10 m s$^{-1}$. As noted in Parish (2017), about 21% of all June and July days hosted an SJ case. All SJ cases in the 5-yr record have wind maxima with a dominant southerly component.

Figure 1 shows composite isotherms for SJ days at a few standard levels of the atmosphere. Here we use averages from eight 3-hourly gridded fields [i.e., averaged over the period commencing at 1500 UTC (1000 CDT) the day prior to the LLJ occurrence to 1200 UTC (0700 CDT) the following day]. The purpose of Fig. 1 is to demonstrate the dominant influence played by summertime insolation of the sloping Great Plains terrain on the background isobaric temperature field during SJ days.

The fundamental feature of the lower atmosphere (i.e., 925, 850, and 700 hPa) over the southern Great Plains states of Kansas, Oklahoma, and Texas is the strong east–west isobaric temperature gradient that develops in response to the summertime heating of the sloping Great Plains terrain. Strongest heating occurs at the surface and, given the topographic slope of the Great Plains, warmest air is found to the west along a particular isobaric surface. Thus, isobaric temperature gradients are generated since air to the west is closer to the surface than air to the east where the isobaric surfaces are at higher levels above the ground. Isobaric temperature gradients are therefore a climatologically prominent feature of the lower atmosphere over the Great Plains region during SJ days of June and July from the 2008–12 study period (also see Smith et al. 2019). Recently, Gebauer and Shapiro (2019) have analyzed 19 years of surface data from the Oklahoma Mesonet and conclude that along-slope buoyancy gradients also contribute to the baroclinic structure in the lower atmosphere. As noted in that work, gradients shown in Fig. 1 reflect such buoyancy gradients as well as the general heating of the sloping terrain.

Figure 1a illustrates the composite isotherms at 925 hPa and the underlying terrain for SJ days. Isotherms at 925-hPa run essentially parallel to topographic contours over the Great Plains from Texas to Kansas, again simply reflecting the general heating of the sloping Great Plains terrain. Isotherm orientation takes on a slight north–south gradient in southeastern Texas in response to the relatively cool Gulf of Mexico.

Similar east–west isobaric temperature gradients over Kansas, Oklahoma, and Texas extend in the vertical to 850 (Fig. 1b) and 700 (Fig. 1c) hPa. Magnitudes of isobaric temperature gradients over the southern Great Plains decrease progressively with height, as evidenced in changes seen between Figs. 1b and 1c. An isotherm orientation that parallels the underlying terrain contours from Texas northward to Kansas is thus characteristic of the warm-season atmosphere throughout the lowest 2500 m or so and is especially prominent during SJ days.

By 500 hPa (Fig. 1d), however, isotherm orientation shows little relationship to the underlying terrain. Temperature gradients are directed primarily north–south and are considerably weaker than those found at lower levels of the atmosphere. It is thus apparent that the terrain-induced isobaric temperature gradients over the sloping Great Plains during June and July SJ days of the 2008–12 study period are limited to levels principally below about 500 hPa. Consequently, the mean flow at 500 hPa and above has a negligible southerly geostrophic flow for SJ days and hence provides little support for the development of the LLJ. This implies that for typical SJ conditions, development of a supporting southerly background geostrophic wind must commence below approximately 500 hPa.

As noted in Parish (2017), existence of isobaric temperature gradients over a significant depth of the lower atmosphere provides the prerequisite thermal wind forcing that is responsible for development of the strong southerly geostrophic wind, without which strong LLJs would not be a recurrent warm-season feature of the Great Plains. Presence of a strong LLJ is thus confirmation of the coincident existence of strong east–west isobaric temperature gradients throughout the lower atmosphere. Such mean thermal wind forcing for SJ days at 0000 and 0900 UTC is shown in Fig. 2 as cross sections between 37.5°N, 105°W and 33°N, 90°W (see gray line A in Fig. 1a). Figure 2a (0000 UTC) depicts mean isentropes and attendant geostrophic wind components normal to the cross section for SJ days from the surface to 500 hPa. Only weak potential temperature gradients are present near 500 hPa as seen by the nearly flat isentropic field. Isentropic slopes increase progressively moving downward from 500 hPa toward the surface in
Fig. 2a, again reflecting increasing isobaric temperature gradients such as seen in Fig. 1. Effects of late-afternoon heating can be seen by the well-mixed layers in the lowest 1–1.5 km. The isentropic slopes require that the normal component of the geostrophic wind intensify toward the surface, commencing near zero at 500 hPa and reaching a maximum of approximately 20 m s$^{-1}$ near the surface.

By 0900 UTC (Fig. 2b), large changes are evident near the ground in response to nocturnal radiative cooling of the surface. Note, however, that the isentropic slopes remain virtually unchanged above the cooled near-surface layer from that seen at 0000 UTC (i.e., see the 316-K isentrope). This demonstrates that isentropic slopes about 1.5 km or so above the surface are not a diurnally forced feature but represent the effects of longer-term (i.e., seasonal) heating. Note that the normal components of the geostrophic wind show a decrease at the surface but remain in excess of 12 m s$^{-1}$ about 1 km above the surface.

A second means to show the effects of thermal wind forcing is to illustrate composite $D$ values in cross-section form. As noted by Bellamy (1945), $D$ values are simply the difference between the height of a pressure surface and the height of that same isobaric surface in the U.S. Standard Atmosphere. For analyses presented here, we use the mean PECAN daytime atmosphere as described in Parish (2017) as the reference atmosphere rather than the U.S. Standard Atmosphere, 1976. Similar to conventional geopotential height, the gradient of $D$ values on an isobaric surface is proportional to the geostrophic wind. The advantage of using cross sections of $D$ values is that one can better infer the vertical variation of the horizontal PGF since the vertical component of the pressure gradient force is effectively removed; the link between the PGF and
The greater the isobaric slope of $D$ values, the greater the PGF. Figure 2c illustrates composite $D$ values and the normal component of the wind at 0000 UTC for SJ days. Isobaric $D$-value slopes increase moving downward toward the surface, representing a progressive intensification of the PGF. The lowest isobaric $D$ values are found at the surface over most of the domain and decrease along an isobaric surface to the west, indicating that the PGF in the lower levels of the atmosphere along the sloping Great Plains terrain and Rocky Mountains is directed from east to west. This is consistent with the existence of the daytime branch of a mountain–plains solenoidal circulation (e.g., Tripoli and Cotton 1989; Carbone and Tuttle 2008; Reif and Bluestein 2017). Maximum normal wind speed in excess of 10 m s$^{-1}$ occurs in response to the localized maximum in the PGF.

It has been proposed that such a mountain–plains circulation reverses at night (e.g., Carbone and Tuttle 2008). The presence of such a large-scale circulation was first proposed by Bleeker and Andre (1951) who offered that nighttime surface cooling establishes a large-scale drainage pattern over the eastern slopes of the Rocky Mountains and Great Plains. Convergence then occurs over the eastern plains with concurrent ascending motion in support of nocturnal convection. This idea has received support by several authors continuing up to Reif and Bluestein (2017). The composite fields for SJ days employed herein, however, are not in agreement with such a conceptual picture. By 0900 UTC (Fig. 2d), $D$ values retain much of the character of that at 0000 UTC with $D$-value slopes increasing downward. Of relevance is the isobaric $D$-value analysis at the surface which clearly shows the lowest $D$ values adjacent to the terrain. For a large-scale organized drainage flow to become established, the PGF needs to reverse direction from that of the afternoon. Composite $D$ values from SJ days show unequivocally that the PGF remains directed from east to west throughout the entire 24-h period. Based on this analysis, we posit that no large-scale drainage such as discussed by Bleeker and Andre (1951) or Reif and Bluestein (2017) can exist for SJ nighttime periods.

The normal component of the LLJ in Fig. 2d shows a 0900 UTC composite maximum wind speed exceeding 20 m s$^{-1}$. Keep in mind that the LLJ has an associated westerly component and processes such as discussed in Gebauer et al. (2018) may result in convergence to the east of the LLJ that no doubt assists rising motion during the evening.

Consequences of the thermal wind forcing shown in Figs. 1 and 2 are seen in both the isobaric height and wind fields. Figure 3 illustrates composite fields of isobaric height, temperature, wind vectors and wind speed at 925, 850, 700, and 500 hPa at 0600 UTC for SJ days. Owing to the thermal wind forcing, isobaric heights at 925 hPa over the southern Great Plains have a pronounced and persistent depression compared to other regions in the domain.
Plains (Fig. 3a) become directed such that a strong southerly geostrophic wind has become established. Wind vectors generally follow isobaric heights. A LLJ with maximum wind speeds of 18 m s\(^{-1}\) over north-central Oklahoma is evident. Not surprisingly, the maximum LLJ winds are collocated with the strongest height gradient.

At 850 hPa (Fig. 3b), the LLJ magnitude exceeds 20 m s\(^{-1}\), centered about 150 km or so to the west of that seen at 925 hPa. Height gradients are only slightly less intense than those at 925 hPa and are oriented clockwise from those at 925 hPa. Wind vectors are again mostly parallel to height contours, implying that the ageostrophic wind is aligned with the geostrophic wind. Warm-air advection is especially evident north of the jet core. The corresponding veering nature of the low-level geostrophic flow with height is a consequence in part due to thermal wind forcing (warm-air advection) and is seen nearly every night that a LLJ develops. Clockwise turning also occurs throughout the night associated with the inertial oscillation that ensures development of a more westerly flow.

Note that at 0600 UTC, relatively strong east–west temperature gradients are seen at 925, 850, and 700 hPa in the general vicinity of those found in Fig. 1. Despite the onset of surface cooling, those levels retain at least part of their seasonal isobaric temperature gradients. Such behavior has been recognized for some time. McNider and Pielke (1981) note that the daytime mixed layer is typically deeper than the nocturnally cooled layer leaving temperature gradients in the residual layer relatively undisturbed.

Thermal wind forcing effects are less pronounced at 700 hPa (Fig. 3c), an outcome of the diminishing temperature gradients found in the atmosphere above the LLJ core. Isotherms continue to reflect summertime heating of the sloping Great Plains.

FIG. 3. Composite 0600 UTC isobaric heights (black lines; m), temperatures (red lines; °C), wind vectors (black arrows), and wind speeds (blue lines; m, with shading of speeds greater than 10 m s\(^{-1}\)) for June and July SJ days from the 2008–12 study period at (a) 925, (b) 850, (c) 700, and (d) 500 hPa.
terrain, although the magnitude of temperature gradients at 700 hPa is reduced from that seen at 850 hPa. Isobaric height gradients suggest weaker geostrophic flow as compared to that seen at 850 hPa. Consistent with the above changes, the intensity of the LLJ has decreased to about 11 m s\(^{-1}\) and is from a more westerly direction than that seen at lower levels of the atmosphere. Wind vectors again show that the actual wind is parallel to height contours; warm-air advection is apparent across Kansas.

By 500 hPa (Fig. 3d), no evidence of the LLJ is seen. Rather, isobaric heights and associated winds take on a more typical climatological pattern with a broad swath of zonal winds across the northern states with wind vectors that closely follow height contours. Isotherms run roughly parallel to the isobaric heights. The “inflection zone” from northeast Kansas to eastern Nebraska, where isotherm orientation abruptly changes, is the target zone for a high probability of precipitation as noted by Sangster (1979). The spatial changes in wind and height fields associated with the inflection zone can be seen in Figs. 3a–c, where the northern extent of the 10 m s\(^{-1}\) isotach at 925 hPa is aligned with the inflection zone of the isotherms, consistent with thermal wind forcing. Height and wind patterns associated with the northern terminus of the LLJ at the inflection zone has been referred to as the “Sangster signal” in the work of Augustine and Caracena (1994), reflecting the situation shown in Fig. 1 of Sangster (1979; see also Fig. 1 in Augustine and Caracena 1994). We note that pronounced warm-air advection is found, by necessity, at the northern edge of the LLJ core where isobaric temperature gradients change orientation. That inflection zone requires that wind components shift from southerly to westerly, providing a wind direction nearly normal to the isotherm orientation and hence promoting significant warm-air advection within a region of strong convergence (see also Trier and Parsons 1993). Isotherms and wind vectors in Figs. 3a–c show that strong warm-air advection exists over a wide swath over the central Great Plains, and is especially enhanced over northern Kansas, southeast Nebraska and southwest Iowa, near the inflection zone discussed above. Such a situation is what Means (1952) referred to as “cross patterns.”

Nocturnal advective warming is maximized at a level of about 800 hPa in the composite fields (see below), typically 500 m or so above the level of the LLJ core. The location of the warm advection maximum above the maximum LLJ winds is a recurrent feature for individual cases as well, an example of which will be highlighted below. Such localized warming of the atmosphere occurs repeatedly for our SJ cases although the pattern shows significant night-to-night variability owing to transient synoptic disturbances that can modulate the position of the inflection zone shown in Fig. 3b.

Recall that negligible warm-air advection is present at 500 hPa in the composite fields (e.g., Fig. 3d). Differential temperature advection is thus occurring in the lower troposphere at 0600 UTC for June and July SJ days from the 2008–12 study period. Such vertically differential temperature advection, in which the lower levels are warmed relative to upper levels, is a classic means by which the atmosphere is destabilized, as noted by many including Means (1944, 1952), Pitchford and London (1962), and Banacos and Ekster (2010). We advocate that such differential temperature advection also represents one key link between the LLJ environment and destabilization of the nighttime lower troposphere over the central Great Plains throughout the warm season, a finding consistent with the recent work of Rattray et al. (2018) and Parsons et al. (2019). Such destabilization can work in concert with other lapse rate tendency processes (Banacos and Ekster 2010) and convection initiation mechanisms, such as isentropic convergence and forced ascent associated with frontal boundaries to the north (Means 1952; Pitchford and London 1962; Trier and Parsons 1993; Tuttle and Davis 2006), to foster nocturnal convection.

Given the general north–south orientation of isotherms in the lower atmosphere such as shown in Figs. 3a–c, winds associated with the LLJ must have a westerly component for warm advection to occur. Parsons et al. (2019) point to evidence of a nighttime maximum in the east–west component of the wind above the level of the wind maximum associated with the LLJ. Such a feature is a classic characteristic of the LLJ and is seen on nearly every night the jet appears and also in the composite fields (see Fig. 10 in Parish 2017). Development of such a westerly wind maximum is the result of inertial oscillations of the ageostrophic wind that commence near sunset. As shown by Parish (2017), the ageostrophic winds for SJ cases decrease in magnitude and veer with height. From Fig. 10 of that paper, ageostrophic winds at about 800 hPa early in the evening are typically from 140° as compared to 75° near the level of the eventual LLJ at about 875 hPa. With the onset of inertial turning of ageostrophic wind, enhancement of the east–west component of the wind first appears at levels well above the LLJ. As night advances, larger magnitude ageostrophic winds attain directions that permit acceleration of the east–west components at lower levels. This process leads to the appearance of an increase and a lowering of the maximum east–west component with time, though levels of such an east–west maximum remain above the LLJ core. It is hardly surprising then that the maximum warm-air advection during the night is found at levels above the LLJ core that contain a larger east–west component of the wind.

Patterns revealed by the composite fields offer indisputable evidence of the climatological persistence of warm-air advection associated with the LLJ and prevalence of differential temperature advection as a means of preconditioning the atmosphere over the central Great Plains for outbreaks of nighttime convection during summer. To illustrate, Fig. 4 shows resulting temperature changes from 0000 to 0900 UTC for SJ days. A vertical cross section from eastern Wyoming through central Missouri (Fig. 4a; see solid gray line B in Fig. 1a) shows nocturnal warming from 0000 to 0900 UTC in a layer of about 2000 m in depth centered near 800 hPa. Weak cooling, perhaps in response to radiative effects, is present above 600 hPa. Spatial patterns of warming at 800 hPa (Fig. 4b) match the general climatological position of the maximum frequency of mesoscale convective complexes such as shown in Geerts et al. (2017) or Weckwerth et al. (2019). Warming can be seen to encompass a broad region over the Great Plains.
from Oklahoma to North Dakota and extending eastward into Missouri and Iowa.

Three issues are apparent from Fig. 4. First, the warming around 800 hPa that is capped by weak cooling above 600 hPa implies that recurring vertically differential temperature advection is occurring during strong LLJ episodes. The LLJ environment is thus responsible for destabilizing the lower atmosphere above the cooled near-surface layer. A second issue concerns the broad spatial extent of the nighttime warm advection in the composite fields in Fig. 4b. Carbone and Tuttle (2008) have offered that the nocturnal 650-hPa rising motion maximum seen in their summertime climatology is best explained by a reversal in the mountain–plains solenoidal circulation. We suggest that such mean rising motion might also be explained as a quasigeostrophic response to the prevalence of geostrophic warm-air advection associated with the LLJ during summer (see Fig. 3). Previously we have established that no large-scale mountain–plains circulation reversal is possible for our SJ cases given that the near-surface horizontal pressure gradient documented in Fig. 2d shows a PGF adverse to any low-level drainage flows. Last, the recurring warm advection and nocturnal warming of an atmospheric layer above the radiatively cooled surface layer serves to increase the stability for surface-based convection the following day. For episodes of strong warming around 800 hPa or so, a significant cap becomes established that requires significant heating the following day to remove (see also Carlson et al. 1983). Days following an enhanced LLJ required additional forcing for convection to develop during the subsequent afternoon. This process at least supports the well-observed minimum in afternoon convection over the Great Plains and states to the east that traditionally has been explained as a consequence of subsidence associated with the descending branch of a mountain–plains circulation (e.g., Bleeker and Andre 1951; Carbone and Tuttle 2008).

Destabilization is typically accompanied by a reduction in CIN. Following Parsons et al. (2019), time–height profiles of CIN and CAPE have been constructed (Fig. 5) from vertical profiles of virtual temperature from the NAM composite grids at 40.5°N, 97.5°W (see the black dot in Fig. 4b).
nighttime warming is seen in a diminution of CIN. For example, CIN magnitude at 750 hPa decreases by 60 J kg\(^{-1}\) from 0000 to 0900 UTC. The low-CIN layer corresponds to the level of maximum nocturnal warming shown in Fig. 4a. Such a reduction in CIN provides at least a preconditioning mechanism for elevated nighttime convection above the radiatively cooled near-surface layer. Given the relatively unchanged CAPE field (Fig. 5b) above the radiatively cooled near-surface layer, destabilization provides a key preconditioning mechanism for subsequent convection. We point out that the relatively small magnitude of warming in the composite fields (e.g., Fig. 4b) masks the often-pronounced destabilization that can occur during individual cases, an example of which will be shown later.

To summarize our findings from composite fields for SJ days, the intensity of the nocturnal LLJ maximum is intimately linked to the strength of the PGF and hence the magnitude of the southerly geostrophic wind at the level of maximum winds. It also follows that the strength of the PGF is dependent on the integrated thermal wind forcing from about 500 hPa down to the level of the LLJ. In this way, the magnitude of the LLJ maximum is directly tied to the intensity of the seasonally forced isobaric temperature gradients above jet level. Given the presence of a nocturnal LLJ, it is also the case that significant isobaric temperature gradients must exist at and just above the level of the jet. Thus, the strength of LLJ is proportional to the intensity of the isobaric temperature gradients at jet level. Significant warm-air advection is maximized at a level 500 m or so above the LLJ core and is the source of destabilization. Warm-air advection decreases with height in response to diminishing wind speeds and weaker isobaric temperature gradients aloft. This connection between the atmosphere at

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**Fig. 6.** Isobaric heights (black lines; m), temperatures (red lines; °C), wind vectors (black arrows), and wind speeds (blue lines and shading; m s\(^{-1}\)) for speeds greater than 10 m s\(^{-1}\) from NAM grids valid at 0600 UTC 5 Jul 2015 for (a) 925, (b) 850, (c) 700, and (d) 500 hPa. The solid dark lines in (a) refer to cross section locations in Figs. 7 and 8a, below.
and above the level of the jet with the LLJ is at the heart of the destabilization process.

3. An example of LLJ-differential temperature advection destabilization process

To illustrate an example of the LLJ-differential temperature advection on the destabilization process, the 5 July 2015 case from the PECAN program has been selected. As noted by Gebauer et al. (2018), this is among the most studied of all PECAN cases (see also Reif and Bluestein 2017; Trier et al. 2017; Weckwerth et al. 2019; Weckwerth and Romatschke 2019). For consistency, we again use NAM grids to illustrate key processes at work for the 5 July 2015 case. It is not our intent here to examine details of individual convective elements, but rather to simply demonstrate the process by which LLJ-differential temperature advection destabilizes the nocturnal atmosphere.

Figure 6 depicts NAM grids of isobaric heights, temperatures, wind vectors and wind speed at 0600 UTC for four standard levels on 5 July 2015. At 925 hPa (Fig. 6a) pronounced temperature gradients are present over the Great Plains states from Texas northward to Nebraska despite the late nighttime hour. Isotherm orientation that follows the general shape of underlying terrain contours is evidence of the primal importance of heating in establishing the temperature field. Strong height gradients also persist over central Oklahoma and Kansas. A LLJ with speeds in excess of 15 m s\(^{-1}\) has developed; wind vectors indicate that the LLJ is predominantly from the south at 925 hPa. Larger temperature gradients are seen at 850 hPa (Fig. 6b), again reflecting general summertime terrain heating. The 850-hPa height gradient reaches farther north than that seen in the composite fields. Conditions are favorable for a 20 m s\(^{-1}\) LLJ that extends northward to the Nebraska–South Dakota border. Note that warm-air advection is present as evidenced by wind vectors occurs from northern Oklahoma northward to east-central Nebraska.

By 700 hPa (Fig. 6c), relatively strong temperature gradients continue, although wind speeds associated with the LLJ have decreased significantly to less than 10 m s\(^{-1}\). Note that the geostrophic wind takes on a more westerly direction; wind vectors indicate that the actual wind directions are nearly orthogonal to the isotherms. While actual winds are relatively weak, warm-air advection is apparent across the southern Great Plains states. Considerably weaker temperature gradients are found at 500 hPa (Fig. 6d) and isobaric heights suggest northwesterly flow across Kansas.

As with the composite fields, vertical cross sections (Fig. 7) have been prepared from the NAM grids to show the development of the background geostrophic wind in which the LLJ is embedded. Here the cross section cuts through northern Kansas (see solid dark line in Fig. 6a), a site of convection during the nighttime hours of 5 July 2015. Figure 7a shows the isentropes and resulting geostrophic wind components normal to the cross section at 0000 UTC on 5 July. As with the composite fields, isentrope slopes increase from 600 hPa to the

![Fig. 7. As in Fig. 2, but from NAM grids for 5 Jul 2015 along a line between 40°N, 105°W and 38.5°N, 95°W (see the solid dark line in Fig. 6a).](image-url)
surface. A well-mixed near-surface layer is also present; isentropes show the daytime boundary layer to be considerably deeper to the west, consistent with more arid conditions there (e.g., Gebauer et al. 2018). Unlike the composite cross sections, the PGF at 500 hPa on 5 July 2015 supports a northerly geostrophic wind component of roughly 8 m s⁻¹. Large changes in the normal component of the geostrophic wind occur from 600 hPa to the surface in response to the thermal wind forcing. Southerly geostrophic winds become established by 700 hPa and increase in magnitude downward; the largest magnitude of the normal component of the geostrophic wind is greater than 20 m s⁻¹ near the surface.

Destabilization is apparent in the isentropic pattern by 0900 UTC (Fig. 7b) in a layer just above the LLJ core. Associated with vertically differential temperature advection, isentropes show an increasing vertical separation. This is especially apparent between the 318- and 320-K isentropes along 98°W, the location where convection appeared around 0900 UTC. Characteristic of warm-air advection, the 318-K isentrope displays a pronounced downward displacement from 0000 to 0900 UTC. Reif and Bluestein (2017) have also argued that warm-air advection is an important factor for this case and Trier et al. (2017) discuss development of a 100-hPa-deep layer that is nearly dry adiabatic at 0600 UTC for this case. Close inspection also shows that isentropic slopes have increased from those at 0000 UTC and the maximum normal component of the geostrophic wind remains in excess of 20 m s⁻¹. Note also that the geostrophic wind maximum has moved downslope from its position earlier in the evening in response to the adjustment in the isentropes and hence thermal wind forcing.

Analyses of D values at 0000 UTC (Fig. 7c) show the transition from northerly geostrophic winds at 500 hPa to southerly by 700 hPa. As with the composite fields, maximum normal geostrophic winds as evidenced by D-value slopes are seen near the surface. A broad maximum in the zonal wind components in excess of 10 m s⁻¹ mark the location of the nighttime LLJ. D values and normal components of the wind are consistent with the sequence of events described above. At 0000 UTC, D values at 500 hPa show a PGF with higher heights to the west, implying northerly geostrophic winds. D-value slopes progressively decrease moving toward the surface in response to thermal wind forcing and switch signs about 700 hPa. The normal wind component shows a maximum collocated with the position of the maximum geostrophic wind in Fig. 7a.

By 0900 UTC (Fig. 7d) the D-value slopes at 500 hPa show a less pronounced northerly geostrophic wind compared with that at 0000 UTC. The PGF supports southerly geostrophic winds commencing about 600 hPa, and D-value slopes progressively increase toward the surface. Lowest pressure remains to the west over the entire domain at 0900 UTC. As in the composite case shown in Fig. 2d, the PGF at the surface retains support for southerly geostrophic winds and hence provides no evidence of a nocturnal reversal of the mountain-plain circulation. Hence, large-scale drainage flow such as discussed in Bleeker and Andre (1951) is not possible. Normal components of the wind show a strong LLJ that is collocated with the maximum normal component of the geostrophic wind.

The analyses show downslope translation of the LLJ during the evening. Maximum LLJ winds follow the maximum PGF that also progresses eastward in response to the adjustment of isentropes and hence thermal wind forcing. An easterly displacement of the LLJ wind maximum during the night has been observed by others (e.g., Smith et al. 2019).

That destabilization as shown in Fig. 7b is the result of warm-air advection can be inferred in the vertical cross section of temperature change between 0000 and 0900 UTC across northern Kansas (Fig. 8a; see solid line in Fig. 6a for cross section location). Maximum warming in excess of 4°C is evident at 750 hPa at 101°W, matching the location of the isentrope stretching shown previously. Significant destabilization has occurred over an approximately 2-km layer from about 800 to 600 hPa, again commencing about 500–1000 m above the LLJ core. As will be shown below, the position of the destabilization also matches the area where convection developed by 0900 UTC.

Figure 8b illustrates the spatial pattern of warming at the 750-hPa level from 0000 to 0900 UTC. Maximum warming is
found in northwestern Kansas, extending northward into Nebraska. It should also be noted that differential temperature advection associated with the LLJ extends eastward at lower levels. At 825 hPa (not shown), a maximum of 5°C of warming occurs between 0000 and 0900 UTC in northeast Nebraska and south-east South Dakota; convection also developed in this area during the early morning hours of 5 July 2015. Similar warming is also seen on other PECAN LLJ nights such as on 20 June 2015 as discussed by Smith et al. (2019).

Differential warm-air advection is responsible for nighttime diminution in CIN for this case. Figure 9 illustrates time–height sections of CIN and CAPE for 5 July 2015 at grid point located at 39.5°N, 100.5°W (see black dot in Fig. 8b). As noted above, the strongest warm-air advection throughout the night is near 750 hPa, just above the LLJ core. As with the composite fields and results from Parsons et al. (2019), CIN magnitude progressively decreases with height throughout the night (Fig. 9a). Inspection of the Fig. 9a shows a nighttime decrease in CIN magnitude in an atmospheric layer commencing about 1000 m above the LLJ core. The most significant decrease is in excess of 100 J kg⁻¹ from 0000 to 0900 UTC centered near 725 hPa, where there is the most significant warm advection (e.g., Fig. 8a). Nighttime diminution of CIN magnitude thus is consistent with effects of differential temperature advection. The evolution of CAPE (Fig. 9b) depicts an increase at 750 hPa between 0000 and 0900 UTC, providing additional support for development of convection during nighttime hours.

Destabilization can also be seen in vertical profiles taken from the NAM. Figure 10a shows the effect of the warm-air advection on the potential temperature profile during the nighttime hours of 5 July at the same grid point for Fig. 9, which corresponds to a site of convective activity at 0900 UTC. Again, note the maximum warming that occurs near 750 hPa. A feature from this case that is also seen in other PECAN cases is that the destabilization is often accompanied by little change or even a decrease in dewpoint temperature (Fig. 10b). For this case, dewpoints remained roughly constant (see also Trier et al. 2017). This finding is contrary to a longstanding position regarding the LLJ that has increasing dewpoints accompanying the formation of the LLJ, but it is consistent with recent work by Parsons et al. (2019) and supports the idea that low-level moisture is present in the atmosphere over the Great Plains during LLJ episodes as a result of seasonal transport from the Gulf of Mexico moisture (e.g., Helfand and Schubert 1995).

Winds (Fig. 10c) reveal classic development of the LLJ such as described in Shapiro et al. (2016) or Parish (2017). Maximum winds in excess of 20 m s⁻¹ develop at about 850 hPa. LLJ directions indicate veering with height. Maximum east–west components of the wind at 0300 UTC are 8.7 m s⁻¹ at 725 hPa, by 0600 UTC are 8.3 m s⁻¹ at 775 hPa, and at 0900 UTC are 10.5 m s⁻¹ at 800 hPa. Throughout the night, such east–west components are responsible for the bulk of the warm advection owing to the roughly north–south orientation of isobaric isotherms. Not surprisingly, the largest temperature increase during the night of 5 July 2015 near 750 hPa corresponds to the position of the maximum east–west components (see also Parsons et al. 2019), roughly 1000 m above the position of the LLJ core.

Convection elements on 5 July 2015 are collocated with the position of maximum destabilization. Figure 10d shows convection that has developed by 0900 UTC over western Kansas in the area that matches the maximum 0000 UTC to 0900 UTC warming shown in Fig. 8b.

4. Summary and conclusions

The Great Plains and states directly to the east experience enhanced nocturnal convection during the summer months. Carbone and Tuttle (2008) report that about 65% of nocturnal Great Plains rainfall is the result of thunderstorms that initiate over the elevated Rocky Mountain terrain and then translate eastward at night (see also Weckwerth and Romatschke 2019). Some of these precipitation episodes are locally augmented over the Great Plains; local initiation of Great Plains precipitation accounted for the remainder of nocturnal precipitation. Here we report on the unique atmospheric conditions that are present over this region of the United States that provide reinforcement for preexisting convection or support for in situ development of nighttime rainfall.

It is a necessary condition that terrain-induced temperature gradients become established above the sloping Great Plains topography owing to summertime insolation on both diurnal and longer-term time scales. Existence of strong LLJ episodes require large (>10 m s⁻¹) background geostrophic winds at the level of jet. The corresponding PGF develops in response to thermal wind forcing, arising from the isobaric temperature gradients in the lower atmosphere during the warm-season months. A prerequisite condition for strong LLJ episodes is the
presence of large isobaric temperature gradients at and just above the level of the nocturnal jet. The intensity of the LLJ is therefore a proxy measure of the magnitude of isobaric temperature gradients in the lower atmosphere. Warm-air advection is maximized at a level of about 500 to 1000 m above the level of the jet core in response to temperature gradients and the evolving ageostrophic inertial oscillations associated with the LLJ. Such a destabilization process is a recurring nocturnal feature during the warm-season months.

To explore the relationship between the LLJ and warm-air advection, composite fields have been assembled for strong LLJ cases from NAM 3-hourly grids for summer months June
and July over the 5-yr period 2008–12. Differential temperature advection often occurs within a vertical column along the axis of the LLJ just above the level of the jet and about 500 hPa during the nighttime hours throughout the summer months. Its effect is to steepen lapse rates during the night (e.g., Fig. 8a) and thereby decrease atmospheric stability above the radiatively cooled surface layer. As a result, CIN diminution during the night is a recurring and characteristic feature of air parcels just above the level of the LLJ over the central Great Plains during strong jet events. Working in concert with convection initiation mechanisms, such destabilization is a preconditioning mechanism that provides a favorable environment for intensification of existing transient convection or initiation of new convective elements.

The 5 July 2015 case from PECAN is used as an example of the destabilization process. Pronounced differential temperature advection occurs across Kansas and Nebraska that is associated with terrain-induced isobaric temperature gradients. Convection occurred during the nighttime hours of 5 July 2015, collocated with the zone of maximum atmospheric destabilization that developed in response to differential temperature advection in the lower atmosphere just above the LLJ. Such destabilization of the layer above the radiatively cooled surface layer and accompanying reduction in CIN magnitude can occur during warm weather months when a strong jet case is present and represents an active link between the LLJ environment and elevated nocturnal convection over the central Great Plains.

Nocturnal warming at about 800 hPa also acts to suppress surface-based convection the following afternoon. Nighttime heating at LLJ level combined with surface cooling leads to the establishment of a strong inversion by dawn that inhibits subsequent rising motion in response to insolation at the surface during the following day. Mixing associated with surface heating sharpens the inversion in the lower atmosphere and hence increases the CIN for surface-based air parcels during the following daytime hours (see also Carlson et al. 1983). In this way the LLJ environment over the Great Plains offers a special set of conditions that enhance the potential for elevated nocturnal convection while limiting the potential for surface-based convection during the subsequent daylight hours.

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REFERENCES


