A Comparative Study of the 3 June 2015 Great Plains Low-Level Jet

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ABSTRACT

Detailed ground-based and airborne measurements were conducted of the summertime Great Plains low-level jet (LLJ) in central Kansas during the Plains Elevated Convection at Night (PECAN) campaign. Airborne measurements using the University of Wyoming King Air were made to document the vertical wind profile and the forcing of the jet during the nighttime hours on 3 June 2015. Two flights were conducted that document the evolution of the LLJ from sunset to dawn. Each flight included a series of vertical sawtooth and isobaric legs along a fixed track at 38.7°N between longitudes 98.9° and 100°W.

Comparison of the 3 June 2015 LLJ was made with a composite LLJ case obtained from gridded output from the North American Mesoscale Forecast System for June and July of 2008 and 2009. Forcing of the LLJ was detected using cross sections of $D$ values that allow measurement of the vertical profile of the horizontal pressure gradient force and the thermal wind. Combined with observations of the actual wind, ageostrophic components normal to the flight track can be detected. Observations show that the 3 June 2015 LLJ displayed classic features of the LLJ, including an inertial oscillation of the ageostrophic wind. Oscillations in the geostrophic wind as a result of diurnal heating and cooling of the sloping terrain are not responsible for the nocturnal wind maximum. Net daytime heating of the sloping Great Plains, however, is responsible for the development of a strong background geostrophic wind that is critical to formation of the LLJ.

1. Introduction

Nocturnal low-level wind maxima have received considerable attention during the past few decades. In particular, the Great Plains low-level jet (LLJ) has been the topic of extensive study (e.g., Bonner 1968; Mitchell et al. 1995; Whiteman et al. 1997). Wind profiles in the lowest kilometer at Great Plains sites often show profound day-to-night differences. Weak southerly winds in the lowest several hundred meters often persist throughout the daylight hours only to be replaced by strong winds with a well-defined jet core during the nighttime. For LLJ cases, wind speeds increase significantly in the hours after sunset with a wind maximum often in excess of 20 m s$^{-1}$ developing at levels 300–800 m above the ground. LLJ wind speeds typically reach a maximum between midnight and dawn; wind directions commonly display a shift from southerly near midnight to southwesterly by dawn. Observations from soundings and profilers (e.g., Bonner 1968; Whiteman et al. 1997) and results from model simulations (e.g., Zhong et al. 1996) show that the summertime LLJ is centered geographically over the southern Great Plains from Texas northward to Nebraska with a maximum over northern Oklahoma and southern Kansas.

Two paradigms continue to be espoused in describing the forcing of the LLJ (e.g., Jiang et al. 2007; Du and Rotunno 2014). The first theory, proposed by Blackadar (1957), considered the LLJ to be supergeostrophic, the result of an inertial oscillation of the ageostrophic wind arising due to the sudden decay of turbulence in the boundary layer after sunset. Airborne measurements (Parish et al. 1988) and numerical experiments (e.g., Zhong et al. 1996) leave little doubt that LLJ winds are supergeostrophic and veer throughout the night in a manner prescribed by the Blackadar mechanism.

A second theory is that proposed by Holton (1967) and suggests that the oscillation in the wind is tied to the diurnal heating and cooling of the sloping Great Plains terrain. The influence of diurnal heating of the sloping terrain has also been discussed in Bonner and Paegle (1970). In that work, 3-hourly analyses of near-surface geostrophic winds were used to investigate periodic
variations in the horizontal pressure gradient force (PGF) resulting from the heating and cooling of the sloping terrain. Maximum geostrophic winds were observed at about 2100 UTC; an oscillation of about 3 m s\(^{-1}\) was seen (Bonner and Paege 1970). Such studies show that the maximum geostrophic winds occur nearly 180° out of phase with the observed maximum LLJ wind speeds, an issue that continues to confront those who champion the original Holton theory.

As a consequence of the two dominant themes used to describe the LLJ, authors (e.g., Jiang et al. 2007; Du and Rotunno 2014) have attempted to evaluate the LLJ in terms of these two theories or some combination of each. Here the forcing of the LLJ is addressed through the dynamics of the 3 June 2015 LLJ case from the Plains Elevated Convection at Night (PECAN) field program. Detailed airborne observations of wind and temperature were obtained from two flights using the University of Wyoming King Air research aircraft (UWKA). Emphasis is placed on the role of the evolving thermal structure over the sloping Great Plains terrain on the dynamics of the LLJ. Comparison is made of the LLJ observations taken during PECAN with a classic LLJ based on a 2-yr LLJ climatology. Here it will be put forward that the mechanism suggested by Blackadar (1957) is responsible of the formation and evolution of the LLJ. Diurnal oscillations in the geostrophic wind such as shown by Holton (1967) are not the reason for the development of the nocturnal wind maximum, although it has an impact on the actual wind profile.

2. Climatological forcing of the Great Plains LLJ

To place in situ aircraft observations of the evolution of the LLJ in the proper context, an examination has been made into the mean LLJ conditions as inferred from the operational 12-km horizontal resolution North American Mesoscale Forecast System (NAM). Here the focus is on summertime months of June and July for a 2-yr period 2008–09 to provide a composite gridded output set with which to compare the PECAN observations. To focus on the LLJ environment, model output was selected to include only those days for which a southerly LLJ was present. Bonner (1968) lists three criteria by which the intensity of the jet is categorized that have guided the selection process. Here conditions include a maximum jet speed of 15 m s\(^{-1}\) or greater at the 0900 UTC profile with a concurrent decrease such that the 700-hPa wind vector has a magnitude less than 10 m s\(^{-1}\). These LLJ criteria limited the composite output to 47% of the June/July cases for the 2-yr period. This percentage is in accord with statistics from the Whiteman et al. (1997) study.

The large-scale circulation in the lower atmosphere over the southeastern part of the United States during summer is associated with the Bermuda high (e.g., Pan et al. 2004). As a result, southerly flows are prevalent from the East Coast westward to the Rockies throughout warm season months. Examination of typical summertime conditions in the lower atmosphere over the Great Plains, however, shows a PGF over the Great Plains that is far more intense than that expected from the Bermuda high alone. As an example, Fig. 1a illustrates the mean 0000 UTC sea level pressure field from the composite 2-yr average June/July period for which the LLJ was active. Two features are prominent on the 0000 UTC map that represent the essence of boundary layer wind maxima over the sloping Great Plains topography. First, the PGF is strengthened east of the Rocky Mountains and is especially large across Texas, Oklahoma, and Kansas. This enhancement is responsible for a significant southerly geostrophic component at low levels of the atmosphere, a prerequisite for the LLJ.

Second, mean 850–950-hPa isotherms in the composite 0000 UTC map (Fig. 1a) underscores the fundamental role of solar insolation of the sloping terrain in the development of the summertime jet. A large east–west temperature gradient is present across the southern Great Plains in Fig. 1a. The mean isotherms track the general height contours of the topography. The most intense summertime heating occurs at the surface. Along a particular isobaric level, the warmest temperatures are found closest to the terrain and over the western Great Plains. Therefore, a strong isobaric temperature gradient develops as a response to the summertime heating of the sloping Great Plains terrain.

The gradient of temperature along an isobaric surface is proportional to the thermal wind. For the situation shown in Fig. 1a, a mean temperature gradient of about 8°C (500 km)\(^{-1}\) is present across Oklahoma and southern Kansas. Such a gradient corresponds to an increase in the southerly geostrophic wind moving downward from 850 to 950 hPa by 5.7 m s\(^{-1}\). From the composite gridded NAM output, diurnal oscillation of temperatures owing to daytime heating is strongest at the surface and is reduced to near zero by about 600 hPa. This suggests that it is the thermal wind between 600 hPa and the surface that is responsible for the development of the enhanced PGF shown in Fig. 1a.

Holton (1967) and others note that radiational cooling during nighttime hours may be sufficient for large-scale drainage flows to develop. Results from the LLJ cases in the NAM composite grids suggest otherwise. As an example, Fig. 1b shows the 1200 UTC sea level pressures and mean 850–950-hPa temperatures from the 2-yr summertime LLJ composite grids. This time roughly
corresponds to the coolest time of day, representing the integrated effects of nighttime cooling on the atmosphere. Outstanding features include a decrease in the intensity of the sea level PGF over the Great Plains and the corresponding mean 850–950-hPa temperature gradient as compared to that shown in Fig. 1a. Such changes are in response to the nighttime cooling that acts to diminish and at the lowest levels reverses the isobaric temperature gradients. Thus, the thermal wind is reduced considerably during the nighttime hours.

Note, however, that the 1200 UTC low-level PGF shown in Fig. 1b still supports southerly geostrophic winds over the Great Plains. Such a pressure field does not support a large-scale drainage flow at the surface since lower pressures remain over the highest terrain. Composite grids from the NAM also show that the PGF remains considerably stronger than the background PGF associated with the broad anticyclonic flow associated with the Bermuda high. This indicates that the net effect of the daytime heating of the sloping terrain is able to persist throughout the night. From the thermal wind perspective, the layer of air cooled through nighttime radiative processes is too shallow and the isobaric temperature gradients too weak to overcome the pressure field that has developed during the previous afternoon.

Close inspection of Fig. 1b shows that the mean 850–950-hPa temperature field at 1200 UTC is rather disorganized compared to 12 h earlier, but still with warmer temperatures to the west. Given that the horizontal pressure field in the composite retains a southerly geostrophic wind, it must be the case that the mean isobaric temperatures between middle tropospheric levels and the surface still show warmer temperatures to the west. Mean temperature gradients from the composite grids indicate that temperature gradients with warmer temperatures to the west persist throughout the night to 600 hPa; typical gradients of 3°C–4°C across Oklahoma are present between 700 and 800 hPa. This implies that the sea level pressure gradient must continue to reflect

FIG. 1. Sea level pressure (solid, dark lines; hPa) and mean temperatures between 950 and 850 hPa (dashed lines; °C) based on composites of LLJ cases from the 12-km NAM output during June and July of 2008 and 2009 at (a) 0000 and (b) 1200 UTC, and (c) the entire 24-h period.
thermal wind processes from aloft. From the 1200 UTC sea level pressure field shown in Fig. 1b, it can be concluded that summertime heating of the sloping terrain manifests itself in the enhancement of the PGF over the Great Plains that endures throughout the 24-h period. Effects of summertime heating of the sloping terrain are obvious. The western enhancement of the PGF is pronounced. From this figure it is again concluded that the enhanced sea level pressure gradient over the Great Plains is a consequence of the thermal wind constraints originating from the heating of the sloping terrain.

The aforementioned paradigm of the LLJ environment emphasizes the importance of both the terrain-induced temperature gradients and the synoptic forcing pattern at 600 hPa or so. Ideal conditions for the development of a strong LLJ consist of a favorable thermal wind pressure gradient at 600 hPa and abundant summertime solar radiation to warm the underlying sloping terrain. For situations in which 600-hPa geostrophic wind is directed with a southerly component and fair summertime weather conditions prevail to ensure a strong terrain-induced horizontal temperature gradient that extends from the surface to 600 hPa, significant enhancement of the PGF will result throughout lower levels of the atmosphere. A strong LLJ is possible under such conditions.

In contrast, if an extended period of cloud cover or excessive soil moisture exists, the terrain-induced temperature gradients associated with the sloping Great Plains terrain will be reduced. This will decrease the magnitude of the thermal wind within the lowest levels of the atmosphere and the low-level PGF. Further, if the 600-hPa geostrophic winds are from the north the terrain-induced temperature gradients and thus the thermal wind in layers below 600 hPa must first overcome the adverse PGF. If the initial 600-hPa northerly geostrophic winds are in excess of 10 m s$^{-1}$, even strong terrain-induced temperature gradients are insufficient to overcome the upper-level PGF and only a weak southerly geostrophic wind may develop near the surface. Under such conditions no significant LLJ will form.

The enhancement of the PGF over the Great Plains as shown in Figs. 1a–c is similar to the “westerly” intensification first discussed by Wexler (1961). In that paper, Wexler argued that the LLJ occurrence was forced in a manner similar to an inertial boundary layer current, being deflected northward by the North American Cordillera. Although such “mechanical” forcing has received some recent support (Ting and Wang 2006), Parish and Oolman (2010) have shown that previously proposed processes such as topographic blocking are not consistent with the observed pressure field that shows lower pressures to the west throughout the day. For blocking situations (e.g., Schwerdtfeger 1975), higher pressure must be situated adjacent to the obstructing orography.

In terms of the original Holton (1967) mechanism, it is proposed here that it is not the diurnal oscillation of the PGF that is relevant to the LLJ but rather the net effect of the solar insolation on the sloping Great Plains topography. For a LLJ to form, a large PGF is necessary. The high frequency of the LLJ over the Great Plains during summertime conditions is the result of the enhancement of the PGF such as shown in Fig. 1. Such an intensification of the PGF is the net effect of strong summertime heating of the sloping terrain.

A vertical profile showing the thermal wind process that occurs over the Great Plains during summertime LLJ days is illustrated in Fig. 2. Conditions represented in this figure are taken from the LLJ composite grids averaged along 37°N between 101° and 98°W for a time of 0000 UTC. During summer the elevated terrain to the west becomes heated with respect to the atmosphere to the east and, hence, a temperature gradient develops that is maximized near 0000 UTC. The strongest horizontal temperature gradients are generally found near the surface and decrease with height. Isobaric temperature gradients of approximately 1.5°C (100 km)$^{-1}$ are representative of the gradients shown in Fig. 1a. From Fig. 2, mean temperature gradients between 600 and 950 hPa are about 1°C (100 km)$^{-1}$. This corresponds to an increase in the southerly component of the geostrophic wind of about 14 m s$^{-1}$ between 600 and 950 hPa.

Figure 2 also shows the $y$ component of the geostrophic wind from the LLJ composite grids, increasing from less than 2 m s$^{-1}$ at 600 hPa to about 13.5 m s$^{-1}$ at 950 hPa. To show that such a geostrophic wind profile is consistent with that expected from the thermal wind, a geostrophic wind profile is also computed, starting with the geostrophic wind at 600 hPa and subtracting the $y$ component of the thermal wind computed between subsequent 25-hPa levels until 950 hPa. Note that the inferred geostrophic wind profile (dashed line in Fig. 2) follows the actual geostrophic wind with fidelity. This merely demonstrates that the PGF in the $x$ direction can be explained by simple thermal wind constraints. There is no doubt that the enhanced PGF that develops over the Great Plains as shown in Fig. 1a represents the effects of isobaric temperature gradients and the thermal wind.

The above argument specifically is focused on summertime conditions. During winter, profound differences no doubt emerge. Clear-sky conditions during winter can result in a horizontal temperature gradient opposite that shown in Fig. 1a and supportive of a...
thermal wind from the south. Under such conditions, support for low-level northerly winds is in place. Although the Blackadar (1957) frictional decoupling mechanism should be less active during winter, low-level wind maxima from the north may develop in response to the dual effects of friction in the near-surface layer and a decreasing PGF with height from thermal wind constraints.

Dynamics of the LLJ from the 2-yr summertime LLJ composite can be illustrated using cross sections of $D$ values. Bellamy (1945) conducted the seminal work on $D$ values, which are simply the difference between the height of a pressure surface and the height of that same pressure surface in the U.S. Standard Atmosphere. Bellamy (1945) points out that isobaric variation of $D$ values plays the same role as isobaric variation of geopotential heights. Both are proportional to the magnitude of the PGF and the geostrophic wind. One significant advantage of using cross-section analyses of $D$ values is that the visualization of the vertical variation of the PGF and, hence, the thermal wind is facilitated. In cross sections that simply display geopotential height, the PGF is difficult to assess since heights are dominated by the hydrostatic component and, hence, the relatively small isobaric gradients are swamped by the far larger vertical changes. The $D$-value analyses remove nearly all of the hydrostatic influence and, thus, allow the horizontal variations to be revealed on cross sections (Parish et al. 2016).

Choice of a U.S. Standard Atmosphere as a reference for $D$ values is appropriate for many studies but the concept of a $D$ value holds for any reference atmosphere. Here Bellamy’s $D$-value analysis is modified by invoking a reference atmosphere that reflects a well-mixed summer atmosphere with a lapse rate of $9.0^\circ \text{C} \text{ km}^{-1}$ and a sea level pressure and temperature of $1006 \text{ hPa}$ and $308 \text{ K}$, respectively. Cross sections using such criteria enhance visualization of the thermal wind process for the composite grids; the same reference atmosphere will be used in $D$-value cross sections throughout this paper.

Figure 3a illustrates potential temperatures, $D$ values, and wind speeds in excess of $10 \text{ m s}^{-1}$ at the 0000 UTC time for a cross section along $36^\circ \text{N}$ and between longitudes $103^\circ \text{W}$ and $98^\circ \text{W}$ for the composite LLJ grids. This period is approximately the warmest time of day and, hence, the temperature gradients should be most apparent and the low-level PGF should be most pronounced. Isentropes are oriented in a near-vertical direction in the lowest 1–2 km, indicative of a well-mixed boundary layer. The topographically induced temperature gradient is apparent with warmest temperatures to the west. The thermal wind in the lowest levels is from the north and out of the plane of the cross section.

The $D$ values provide a visual depiction of the vertical adjustment of the PGF through the thermal wind process. Note that the slope of the $D$-value contours increases toward the surface, and thus also the magnitude of the PGF. Since isobaric gradients of potential temperature are identical to isobaric variations of temperature, the slope of the isentropes as shown in Fig. 3 is also a measure of the thermal wind. From Fig. 3a, it can be readily seen that the PGF decreases with height in response to the thermal wind and corresponding isobaric temperature gradients. Vertical PGF changes are easily identified on the eastern part of the cross section above the well mixed boundary layer as isentrope slopes transition. Note that the reduction of the $D$ value slopes with height and, hence, decrease of the PGF with height is tied to the changing orientation of the isentropes.
As a consequence of nighttime cooling, significant changes occur in both potential temperature and $D$-value fields. The 1200 UTC 2-yr composite potential temperatures, $D$ values, and wind speeds are illustrated in Fig. 3b. The LLJ is well organized with maximum wind speeds in excess of 16 m s$^{-1}$ in the composite grids. Stabilization of the boundary layer is present in potential temperature contours; isentropes move from a nearly vertical alignment at 0000 UTC to a primarily horizontal orientation near dawn. Note that the isobaric temperature gradient remains directed such that at levels above a few hundred meters AGL, the warmest temperatures remain to the west. Such a picture is consistent with the thermal wind forcing described previously.

The $D$ values develop a classic jetlike profile in the lowest kilometer as a consequence of the nocturnal cooling. The inflection point of the $D$ values denotes the top of the radiatively cooled layer. Not surprisingly, the wind maximum is found just above the inflection point in
the $D$-value curves. As noted by Bonner and Paegle (1970), geostrophic winds in the radiatively cooled layer decrease throughout the night. The largest oscillations in the geostrophic wind magnitude occur near the surface and only small changes are present at the level of the LLJ. Analyses clearly show that the LLJ is supergeostrophic. Winds decrease sharply with height above the LLJ wind maximum in Fig. 3b, in response to the thermal wind forcing and attendant decreasing PGF.

Key profiles associated with the LLJ evolution from the composite grids are shown in Fig. 4. For these profiles an average is obtained between latitudes 35.2°–38.2°N and longitudes 99.5°–99.0°W, which is representative of the entire Great Plains environment. Wind speeds (Fig. 4a) increase dramatically after sunset as the surface layer stabilizes and the force balance of the daytime boundary layer is disrupted. As noted by Blackadar (1957), an inertial oscillation of the unbalanced motion components becomes initiated and the classic LLJ profile becomes established by 0600 UTC in the 2-yr composite grids. Maximum LLJ winds are reached between 0600 and 0900 UTC. Wind directions in Fig. 4a display the classic veering from southerly to southwesterly that is reflective of the inertial oscillation of the unbalanced components.

Profiles of potential temperature (Fig. 4b) show that the cooling is maximized at the lowest levels and extends upward from the surface to about 1 km or so. Most cooling takes place in the first few hours after sunset with minor cooling continuing between 0600 and 0900 UTC. Of particular interest in the initiation of the inertial oscillation is the frictional decoupling. Here profiles of turbulent kinetic energy (TKE) during the nighttime hours are shown. TKE values are maximized during the late afternoon hours near 2100 UTC. By 0000 UTC, TKE profiles already indicate a significant decay and are less than half those three hours earlier. By 0600 UTC, TKE profiles show a marked decrease to model background levels.
Effects of heating of the sloping Great Plains topography are represented by the low-level oscillation in the \( y \) component of the geostrophic wind \( u_y \) in Fig. 4c. As noted previously, maximum \( u_y \) values of about 13 m s\(^{-1} \) are seen in the composite grids at the lowest profile levels. With nighttime cooling, \( u_y \) values near the surface decrease throughout the night with minimum values about half that seen in the afternoon. Note that the diurnal oscillation in the \( u_y \) values is restricted to the lowest roughly 1 km or so. The largest geostrophic wind changes occur at the surface. Although heating and cooling of the sloping terrain must influence the actual wind profile in the lower atmosphere, note that the geostrophic wind oscillation is out of phase with the LLJ maximum. At the time of the LLJ maximum, geostrophic winds are at a minimum. Further, close inspection illustrates that at the level of LLJ core minimal diurnal oscillation in the PGF occurs. It must follow that the diurnal wind maximum itself is not the result of oscillations of the geostrophic wind.

No significant variation is present in the \( x \) component of the geostrophic wind \( u_x \) (Fig. 4c), which is not surprising since the terrain slope is primarily along the east–west axis and thus supporting \( y \)-component geostrophic wind changes.

An advantage of working with gridded composite output is that the dataset is dynamically constrained. This enables direct assessment of the ageostrophic components of motion. Figure 4d illustrates both \( x \) and \( y \) components of the ageostrophic wind, \( u_{ag} \) and \( v_{ag} \), respectively. The time variation of each of these components is critical to the development of the jet profile according to Blackadar (1957); ageostrophic wind components are important dynamic parameters with which to compare PECAN airborne observations. For LLJ flights, direct measurement of the \( y \) component of the ageostrophic winds is made. From the composite grids, the \( y \) component of the ageostrophic wind \( v_{ag} \) at sunset in the composite grids is about \(-7 \) m s\(^{-1} \) at the lowest levels, but increases with height such that at the level of the eventual top of the cooled layer it is about 5 m s\(^{-1} \). In the hours after sunset, \( v_{ag} \) increases in the lower levels of the atmosphere to produce the jetlike structure by 0600 UTC. The magnitude of the \( v_{ag} \) oscillations is maximized near the level of the LLJ. Above the LLJ nose, the \( v_{ag} \) values decrease with time. The profile of \( v_{ag} \) confirms the importance of the inertial oscillations and shows that such ageostrophic components vary by level. Similar to the theory proposed by Blackadar (1957), evolution of the ageostrophic motion is dependent on the initial unbalanced motion components concurrent with the frictional decoupling.

Profiles of the \( x \) components of the ageostrophic wind \( u_{ag} \) show only minor vertical changes in magnitude to start the inertial turning near sunset. Oscillations seen also are consistent with that expected from an inertial oscillation and the Blackadar (1957) theory, being approximately 90° out of phase with the \( y \) components.

### 3. Forcing of the 3 June 2015 Great Plains LLJ

PECAN was a large field campaign conducted during June and July 2015 based in central Kansas. The primary goals of the study were to document nocturnal convection over the Great Plains during summer and to advance our understanding of such storms. As part of PECAN, LLJ studies were also incorporated. A central objective was to examine the mesoscale kinematics and dynamics that contribute to the nocturnal wind maximum. In particular, the ability of an airborne platform to directly measure the dynamics of the LLJ is a significant advance from previous studies (e.g., Parish et al. 1988). LLJ cases during PECAN were selected on a noninterference basis with convection flights; as a result, clear skies and fair weather generally prevailed during LLJ case study flights.

Figure 5 presents a synoptic overview of conditions at 0000 UTC 3 June 2015 from the operational NAM. Figure 5a illustrates the sea level pressure field; a strong surface PGF exists across the Great Plains states of Oklahoma and Kansas at 0000 UTC with lower pressures to the west. The sea level pressure field over the Great Plains is similar to that shown in the composite case (Fig. 1a) and the strong PGF again reflects effects of the summertime heating of the sloping terrain. The 950-hPa heights (Fig. 5b) also reveal a strong PGF across the southern Great Plains states. The isotherms at 950 hPa again show the influence of the daytime heating of the terrain with a strong temperature gradient that is directed upslope, similar to that in the composite map shown in Fig. 1a.

The 850-hPa height field (Fig. 5c) also shows a PGF across the southern Great Plains states although not as strong as that at 950 hPa. Isotherms again reflect the diabatic heating of the Great Plains, running primarily in a north–south direction and parallel to terrain height contours. At the 600-hPa level (Fig. 5d), isotherms show only a weak temperature gradient across the southern Great Plains. This suggests that the diurnal oscillations induced by diabatic heating processes on this day are most pronounced below the 600-hPa level. The progressive changes in the height field from 600 hPa down to the surface are due to the thermal wind influence across the Great Plains and extending northward to the northern plains states.
The height field at 600 hPa over central Kansas in Fig. 5d shows geostrophic winds to be weak. The Dodge City, Kansas, sounding from 0000 UTC 3 June 2015 indicated a 6 m s\(^{-1}\) wind from 293°. The 600-hPa PGF and the large-scale synoptic pattern, thus, offer little opposition to the development of a modest LLJ in the lower levels during the nighttime hours.

Given the conditions illustrated in Fig. 5, a decision was made to launch the UWKA just after 0000 UTC 3 June 2015. Figure 6 shows the location of the PECAN study region. The center of operations for PECAN was in Hays, Kansas (HYS in Fig. 6); the UWKA was based out of Great Bend, Kansas (GBD in Fig. 6), about 75 km southeast of Hays. Two flights were executed with the first flight from 0132 to 0548 UTC and the second from 0636 to 1048 UTC. Each flight followed the same pattern with a climbout sounding being conducted from GBD and then repeating vertical sawtooth and isobaric legs along a fixed track at 38.7°N between longitudes 98.9° and 100°W (see dark solid line in Fig. 6a).

Details of the sawtooth profiling along the fixed track are shown in Fig. 6b. Starting on the east end of the fixed track, vertical sawtooth maneuvers were conducted between roughly 150 and 1000 m AGL. Ascent/descent rates were limited to about 300 m (min)\(^{-1}\) to ensure accurate wind measurement. This required a horizontal traverse of about 20 km per individual ascent/descent; a total of five ascent/descent soundings were conducted over the roughly 100-km leg.

For this case, flights beginning on the east end commenced at the upper level and, thus, ending at the lower level on the west end of the track; the pattern was then reversed by starting the higher level on the west end and finishing at the low level at the east end. After both sawtooth legs were completed, reciprocal isobaric legs were flown about 500 m AGL. The pattern was then repeated. The same strategy was used for each flight.

Two key measurements were made during the vertical sawtooth profiling. The series of soundings along the
track enable determination of the LLJ kinematics and the thermodynamic structure of the lower atmosphere. In addition, the differential GPS measurements (i.e., Parish and Leon 2013) enable precise vertical positioning of the UWKA and thereby allow instantaneous $D$ values to be determined. Sawtooth profiling provides data to allow construction of $D$-value cross sections (Parish et al. 2016). To check measurements of the PGF from the sawtooth maneuvers, isobaric legs were conducted along the track at a height roughly at the middle of the sawtooth legs. As noted in Parish et al. (2007), height measurements provided by these isobaric legs enable precise determination of the isobaric slopes of the UWKA, and the PGF that can be compared with measurements obtained by sawtooth profiling.

Figure 7 summarizes soundings obtained from the UWKA sawtooth profiles throughout the night. Sunset at Hays near the center of the flight track on 3 June is at 0155 UTC, about 25 min after takeoff of the first flight. Soundings shown in Fig. 7 are for the westernmost ascent/descent leg.

Figure 7a illustrates the potential temperature profiles as measured from the UWKA. The earliest potential temperature profile in Fig. 7a is from 0215 UTC, about 20 min after local sunset. Most of the sampled boundary layer is characterized by uniform potential temperature profile, indicating a well-mixed lower atmosphere. Close inspection shows potential temperatures at the lowest levels of the sawtooth profile (~150 m AGL) to have decreased by about 2°C from the upper boundary layer. Such cooling commenced prior to sunset and decoupling of the so called “residual” layer (e.g., Stull 1988) from the surface layer is seen to occur early and in a sudden manner. The impressive feature about the early part of the first flight was how rapidly the lower boundary layer stabilized and decoupling between the daytime boundary layer and the near-surface layer took place.

Stabilization of the lower boundary layer continues throughout the nighttime hours as shown in the subsequent potential temperature profiles in Fig. 7a. Cooling is greatest at the lower levels but can be traced to about 1200 m MSL in Fig. 7a, corresponding to a height above ground level of about 500 m. By the time of the second profile at 0430 UTC, an inversion of about 1°C has formed near 1000 m MSL (about 300 m above the terrain). This feature persists and by the third profile at 0720 UTC has lifted to a height of 1050 m MSL and the strength has increased slightly.

By 0910 UTC, potential temperatures in the boundary layer above about 1200 m MSL (500 m AGL) show that warming is occurring. Inspection of dewpoint temperature profiles (not shown) reveals a sharp decrease coincident with the potential temperature increase, offering an unmistakable signal of subsidence.

With the rapid stabilization shown in Fig. 7a, winds undergo an acceleration associated with an inertial oscillation that continues throughout the night. Wind speeds increase 4–5 m s$^{-1}$ at the 1000–1100-m AGL layer in the 2-h period 0215–0430 UTC. The temperature inversions discussed above at 1000 m AGL at 0430 UTC and at 1050 m MSL at 0720 UTC each have wind maxima just above the inversion level, suggesting a strong link between the thermal stratification and the wind field. Maximum jet core winds of nearly 28 m s$^{-1}$ were observed at 1100 m AGL at 0720 UTC during the third profile. The final profile was conducted about two hours after the jet core maximum winds were observed; wind speeds at the level of the jet core show a decrease of approximately 2 m s$^{-1}$. Wind directions (Fig. 7c) display a clockwise rotation throughout the night that roughly matches that expected from an inertial oscillation of the ageostrophic wind.
To evaluate the dynamics of the developing LLJ, measurements of the horizontal pressure gradient force must be made using differential GPS techniques described in Parish and Leon (2013). The vector difference between the actual wind and the geostrophic wind yields the ageostrophic wind. In absence of isallobaric effects, the ageostrophic wind rotates with an inertial period once the residual layer becomes decoupled from the surface. For vertical sawtooth profiling, $D$ values have been calculated and analyzed.

Figure 8 illustrates a cross section from the first vertical sawtooth leg conducted between 0152 and 0210 UTC. Potential temperatures show that a transition from daytime to nighttime is occurring. Isentropes show a near-vertical orientation at levels near 1000 m MSL, corresponding to levels between 350 and 550 m AGL. The warmest temperatures are seen over the higher western terrain. Stabilization of the atmosphere in the lowest levels of the sawtooth is apparent with the modification of isentrope orientation to a more horizontal pattern.

The $D$ values for the initial profile show a strong PGF in the lowest levels directed with lower heights to the west. The orientation of the $D$-value contours is similar to that shown in the composite case (Fig. 3a). Consistent with the thermal wind associated with the isobaric temperature gradient, $D$-value slopes are greatest near the surface and decrease slightly by the top of the sawtooth. Wind speeds already show a wind maximum over the western part of the fixed track with wind speeds in excess of 18 m s$^{-1}$.

Mean conditions from the entire vertical sawtooth leg are shown in Fig. 9. Stabilization of the potential temperature field (Fig. 9a) is again apparent. Of note is the geostrophic wind analysis based on a least squares linear fit to the $D$-value field at various levels along the sawtooth. Since the profiling occurs along an east–west track, only the $y$ component of the geostrophic wind can be detected (Fig. 9b). From the analysis, geostrophic winds have a maximum value of 20 m s$^{-1}$ at the lowest levels. Effects of the initial cooling are also seen in the slight decrease at the lowest levels of the vertical sawtooth. The $y$ component of the geostrophic wind decreases with height in a manner consistent with the thermal wind forcing discussed previously such that the geostrophic wind magnitude has decreased to about 11 m s$^{-1}$ at the top of the sawtooth profile.
Subtraction of the $y$ component of the geostrophic from the $y$ component of the UWKA-measured wind provides an estimate of the $y$ component of the ageostrophic wind (Fig. 9c). Values of $v_{ag}$ range from about $-5 \text{ m s}^{-1}$ at the lowest levels of the sawtooth to near $5 \text{ m s}^{-1}$ at the top level of the sawtooth near 1100 m MSL or 840 hPa, a profile consistent with that from the composite grids shown in Fig. 4d.

To check the validity of the PGF from the $D$-value analyses such as shown in Fig. 9, isobaric legs were conducted at a level of 875 hPa following completion of the first two vertical profiling legs from 0235 to 0315 UTC. As noted in Parish and Leon (2013), such legs are conducted along an isobaric surface using the autopilot; small deviations of the airplane from the mean pressure surface have been corrected using the hydrostatic equation. Figure 10 illustrates isobaric heights from such legs. Isobaric heights display a monotonic increase from west to east with little evidence of the wavelike structures that have been identified in other missions (e.g., Parish et al. 2014). In addition, the out-and-back portions of the isobaric legs show heights that mirror one another with extreme fidelity. This indicates that isallobaric effects are negligible during this portion of the flight.

Geostrophic wind magnitudes from the isobaric legs were $15.4 \text{ m s}^{-1}$, in agreement with the inferred geostrophic wind magnitude from the $D$-value analyses shown in the profile in Fig. 9, and lending support for the validity of the $D$-value analyses.

![Fig. 8. Cross section of data collected by UWKA along fixed track from 0150 to 0232 UTC 3 Jun 2012 showing $D$ values (black lines; m), wind speed (blue lines; m s$^{-1}$), and potential temperature (red lines; K). Dashed brown lines indicate path of UWKA. Heights shown are above mean sea level; actual terrain heights are roughly 100–150 m below lowest points of sawtooth.](image)

![Fig. 9. Mean UWKA profiles of (a) potential temperature (K), (b) the $y$ components of the geostrophic wind (solid line; m s$^{-1}$) and measured wind (dashed line), and (c) the $y$ component of the ageostrophic wind for the sawtooth profile from 0150 to 0232 UTC 3 Jun 2012 shown in Fig. 8.](image)
Profiling continued throughout the nighttime period as the LLJ became better established. The maximum jet wind speeds were observed between 0600 and 0900 UTC. Figure 11 illustrates results from the sawtooth profiling from 0907 to 0928 UTC. The LLJ is well defined with maximum winds in excess of 25 m s$^{-1}$ at 1100 m MSL ($\sim$500 m AGL) that extend across the track. The atmosphere has stabilized as seen by the horizontal isentrope orientation. At the lowest levels the temperature gradient has reversed although the gradient is quite weak. Above the level of the LLJ maximum, the temperature gradient maintains the daytime gradient with warmer temperatures to the west. The $D$ values show a similar nose-shaped profile seen in the composite LLJ case (Fig. 3b) with the maximum wind again above the inflection point of the $D$-value curve.

A curious feature in the cross section is the strong temperature gradient seen above the jet at 1400 m MSL. Note the abrupt slope change in the $D$ values that accompany the horizontal temperature gradients, a visual depiction of the thermal wind process. Airborne measurements of potential temperature and $D$ value are independent of each other. That the two fields show a dynamic consistency again provides support for the validity of the $D$-value analysis. The existence of the enhanced temperature gradient also has implications on the PGF at levels beneath through thermal wind forcing. Combined with the enhanced temperature gradient seen in Fig. 11, changes just above the jet level are apparent. Note that the UWKA profiles shown in Fig. 7 also indicate subsidence over the western part of the sawtooth leg at this time.

The most significant feature is the profile of the PGF (Fig. 12). In the composite case, the PGF showed a progressive decrease with time during the night at low levels; little change occurs at levels near the top of the sawtooth profiles (e.g., Fig. 4). From the UWKA observations, a maximum in the $y$ component of the PGF from the $D$-value analysis is depicted near the level of the LLJ at about 0900 UTC. The $y$ component of the wind roughly follows the $y$-component geostrophic wind profile. Such an enhancement in the nighttime forcing of the LLJ does not fit the classic picture as suggested in the composite grids. Yet, there are suggestions that such an increase has taken place. In addition to the strong horizontal temperature gradient that has developed just above the jet core in Fig. 11, reciprocal 875-hPa isobaric legs flown from 0738 to 0819 UTC and from 0951 to 1032 UTC indicated a $y$ component of the geostrophic wind of 18.0 and 19.2 m s$^{-1}$, respectively. Such values are in good agreement with geostrophic wind components inferred from the $D$-value profiles that represent an increase from that seen earlier in the evening. This enhancement of the PGF near the level of the LLJ is a fundamental departure of the 3 June 2015 case from the mean LLJ as determined from the composite grids and illustrated in Fig. 4.
Grids from the NAM for 1200 UTC 3 June 2015 were also examined but the 600-hPa heights showed little to support these observations. If, for example, the PGF reversed during the night and, hence, the geostrophic wind switched from northerly as shown in Fig. 5 to southerly, such a pressure gradient could lead to an enhancement at lower levels as well. No such process could be observed from the operational forecast model grids.

Evidence from earlier sawtooth profiling leave little doubt that once the residual layer has been decoupled from the surface, an inertial oscillation of the unbalanced wind commences. Likewise, the vertical profile of the $y$ component of the ageostrophic wind (Fig. 12c) is consistent with what is expected of a classic LLJ as demonstrated by the composite grids. Note that the decrease of the ageostrophic wind with height shown in the profile is similar to that shown in the composite LLJ case, consistent with an inertial oscillation given the $v_{ag}$ profile earlier in the evening (i.e., Fig. 4d).

A summary of the UWKA observations from both flights on 3 June at 875 hPa (1200 m MSL) near the level of the maximum winds associated with the LLJ is shown in Fig. 13. The evolution of the wind (Fig. 13a) during the first flight matches that of a classic LLJ such as seen in the composite grids. Wind speeds show a rapid increase with time during the first few nighttime hours and the jet speeds reach 25 m s$^{-1}$ by 0400 UTC on 3 June.

After local midnight, wind speeds remain at approximately the same magnitude for the second flight commencing around 0700 UTC with a suggestion of finally decreasing near the end of the flight at 1000 UTC.

Also shown in Fig. 13a is the evolution of the LLJ winds from the composite grids. Wind speeds increase rapidly during the early nighttime hours and then remain steady until about 1000 UTC after which speeds decrease. UWKA measurements of wind directions (Fig. 13b) show a progressive clockwise turning with time, again consistent with the composite grids and the classic inertial rotation of the ageostrophic components.

Most significant cooling of the atmosphere during the nighttime hours occurs at levels below about 875 hPa (~1200 m MSL, see Fig. 4b). Thus, oscillations in the geostrophic wind are smaller than those at the surface as shown in Fig. 4c. The $y$ components of the geostrophic wind $v_y$ as determined from the sawtooth profiling and isobaric legs (Fig. 13c) show general agreement throughout the both flights. Differences of 3 m s$^{-1}$ are present between the precise isobaric measurements and those inferred from the sawtooth profiles during the middle and later times during the second flight. Both the sawtooth profiling and isobaric legs show a strengthening of the PGF at levels near the LLJ maximum after 0800 UTC, again in contrast to that shown from the composite grids for LLJ development.
Given the general agreement in the $y$ component of the geostrophic wind measurements from the UWKA, it follows that the $y$ component of the ageostrophic winds $v_{ag}$ must also show consistency between the sawtooth profiles and isobaric legs (Fig. 13c). From such measurements, the actual winds are seen to be super-geostrophic throughout most of the night. The pattern of the $y$ component of the ageostrophic winds follows that of the classic LLJ cycle with a rapid increase during the early nighttime hours to $v_{ag}$ values in excess of 10 m s$^{-1}$ by 0500 UTC. UWKA measurements indicate a sharp decrease in the $y$ component of the ageostrophic wind from about 0730 UTC concurrent with the increase in the geostrophic wind component $v_g$. While the trends in the UWKA ageostrophic wind components match that expected as shown in the composite case, the late night sudden decrease from 0900 UTC is larger than that expected for a classic LLJ. Although the increasing geostrophic wind differs from that seen in the composite grids, the overall picture from the UWKA observations offers support for the ageostrophic acceleration suggested by Blackadar (1957).

4. Summary

Airborne measurements taken by the UWKA allow direct assessment of the dynamics that accompany the development and evolution of the LLJ. To understand the significance of such measurements, the structure and dynamics of the mean LLJ were obtained by constructing a composite gridded output set from June and July of 2008 and 2009 for days during which the LLJ was active.

The climatological frequency of the LLJ over the Great Plains is the result of the strong summertime heating of the sloping terrain. A pronounced east–west temperature gradient develops across the Great Plains states in the lower atmosphere from the surface to about 600 hPa. Isobaric temperature gradients dictate that a thermal wind is present throughout the lower atmosphere
that results in an enhancement of the PGF near the surface with lower pressures over the elevated heated terrain. While the horizontal temperature gradient is maximized during late afternoon and early evening, it persists throughout the night at levels above about 400 m AGL. The net effect of such a summertime heating pattern is that the average southerly geostrophic wind within the boundary layer throughout the day is significantly greater over the southern Great Plains than that found to the east. Many have noted the importance of a strong mean flow on the development of the nocturnal LLJ.

As part of the PECAN field study, an LLJ mission was launched on 3 June 2015 to directly measure the forcing associated with the LLJ. A series of vertical sawtooth profiles were conducted along a fixed east–west track that enable D-value cross sections to be constructed and profiles of the PGF. Airborne wind measurements reveal rapid stabilization of the near-surface layers of the atmosphere just prior to sunset at about 0200 UTC; winds accelerate from about 17 m s\(^{-1}\) near the eventual jet core level at 400–500 m AGL to 25 m s\(^{-1}\) by 0400 UTC. The D-value cross sections suggest that y components of the ageostrophic winds at the level of the LLJ increase significantly from 0230 to 0600 UTC, consistent with that from the composite grids.

Winds associated with the LLJ are supergeostrophic throughout nearly the entire night, indicating that the frictional decoupling and attendant ageostrophic acceleration are fundamental to the jet forcing. Airborne observations show that ageostrophic y components evolve similar to that expected from an inertial oscillation, increasing and reaching a maximum as the LLJ reaches peak strength. The y component of the ageostrophic wind decreases after about 0700 UTC, again similar although larger in magnitude than the composite grids suggest.

At the level of the maximum jet wind, diurnal variation of the PGF is small and not critical for the development of the LLJ. In the composite LLJ case, the PGF decreases slightly over the course of the evening. It is concluded that the diurnal evolution of the PGF that is prominent near the surface is not responsible for the development of the nocturnal wind maximum. Evolution of the wind maximum associated with the LLJ is anticorrelated to changes in the surface PGF. For the case of 3 June 2015, the PGF at the level of the LLJ increases after about 0700 UTC, in contrast to that expected from the composite grids. Notwithstanding the increase in the geostrophic wind during the night, the evolution of the LLJ on 3 June 2015 from the UWKA observations remains similar to that of a classic LLJ.

Discussion from previous work has compared and contrasted the Blackadar (1957) and Holton (1967) theories on the forcing of the LLJ. Stabilization of the near-surface layer is apparent in the observations obtained from the airborne sawtooth profiling. Direct measurements of the PGF show the evolution of the LLJ to be supergeostrophic following processes described by Blackadar (1957). It is offered here that oscillations in the PGF owing to the diurnal pattern of heating and cooling of the sloping terrain such as described in Holton (1967) are not tied to the oscillations of the LLJ. The largest magnitude of the geostrophic wind oscillation is found at the surface; minimal changes are seen at the level of the LLJ. At the same time, the physical processes described in the Holton (1967) paper are responsible for the climatological persistence of the LLJ to the lower atmosphere over the Great Plains. Observations and composite grids demonstrate that the heating of the sloping terrain is responsible for the enhancement of the PGF over the Great Plains, which is maintained throughout the day during the summer. The strong low-level pressure gradient is vital to the development of the LLJ.

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