On the Forcing of the Summertime Great Plains Low-Level Jet

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ABSTRACT

The low-level jet (LLJ) is a ubiquitous feature of the lower atmosphere over the Great Plains during summer. The LLJ is a nocturnal phenomenon, developing during the 6–9-h period after sunset. Forcing of the LLJ has been debated for over 60 years, the focus being on two processes: decoupling of the residual layer from the surface owing to nighttime cooling and diurnal heating and cooling of the sloping Great Plains topography.

To examine characteristics and forcing mechanisms for the LLJ, composite grids were compiled from the North American Mesoscale Forecast System for the summertime months of June and July over a 5-yr period (2008–12). One composite set was assembled from well-developed LLJ episodes during which the maximum nocturnal jet magnitude at 0900 UTC over northwestern Oklahoma exceeded 20 m s\(^{-1}\). A second set consists of nonjet conditions for which the maximum nighttime wind magnitude in the lowest 3 km did not exceed 10 m s\(^{-1}\).

The intensity of the horizontal pressure gradient and hence background geostrophic flow at jet level was the dominant difference between composite cases. The horizontal pressure gradient forms in response to the thermal wind above jet level that results primarily from seasonal heating of the sloping Great Plains. Thermal wind forcing is thus the key link between the Great Plains and the high frequency of LLJ occurrence. The nocturnal wind maximum develops primarily because of the inertial oscillation of the ageostrophic wind occurring after decoupling of the lower atmosphere from the surface owing to radiational cooling in the early evening.

1. Introduction

The Great Plains Turbulence Research Program was a major field experiment conducted near the town of O’Neill in north-central Nebraska from August through September of 1953 (Lettau and Davidson 1957). A recurring lower-atmospheric feature documented during that study was what is now known as the Great Plains low-level jet (LLJ). Lettau (1967) noted that the LLJ occurred regularly at an elevation about 450 m above the ground, commencing after sunset and reaching a peak speed of over 20 m s\(^{-1}\) after midnight. Although Means (1954) previously described an individual LLJ event, the Great Plains Turbulence Research Program provided the first extended period of observations of the LLJ.

In the six decades since that field study, the LLJ has been documented extensively through measurements from pilot balloons, radiosondes/rawinsondes, aircraft, wind profilers, and lidar (e.g., Hoecker 1963; Bonner 1968; Parish et al. 1988; Frisch et al. 1992; Mitchell et al. 1995; Whitman et al. 1997; Banta et al. 2002; Song et al. 2005). Analytic models have been developed that describe the evolution of the LLJ (Blackadar 1957; Holton 1967; Shapiro and Fedorovich 2010; Du and Rotunno 2014; Shapiro et al. 2016). In addition, an extensive numerical modeling effort has taken place to study the LLJ (e.g., McNider and Pielke 1981; Fast and McCircle 1990; Zhong et al. 1996; Pan et al. 2004; Ting and Wang 2006, Jiang et al. 2007; Fedorovich et al. 2017).

It is appropriate to acknowledge the wide variety of local wind maxima described in the literature that have been identified by the term low-level jet (e.g., Stensrud 1996). In the discussion that follows, the focus is on the summertime Great Plains nocturnal maximum. Jet profiles linked to transient synoptic disturbances and those tied to the lower branch of a transverse circulation associated with an upper-level jet stream (Uccellini 1980) are not considered.
Forcing mechanisms for the Great Plains low-level jet have been debated since the earliest observations were made. Blackadar (1957) suggested that the nocturnal wind maximum was the result of an inertial oscillation of the ageostrophic wind. Acceleration occurs as the result of frictional decoupling of the lower atmosphere from the surface owing to radiational cooling near sunset that disrupts the daytime force balance. Blackadar (1957) indicated that wind maxima are not related to diurnal oscillations in the pressure field. The Blackadar theory has received widespread support (e.g., Parish et al. 1988; Zhong et al. 1996; Shapiro and Fedorovich 2009; Parish 2016), and it is acknowledged that such a frictional decoupling process must be operating as part of the LLJ development (Shapiro et al. 2016; Fedorovich et al. 2017).

The theory of Blackadar (1957) has also faced some criticism. Foremost, the construct provides no explanation as to why the LLJ displays such a high frequency of occurrence over the Great Plains (Wexler 1961; Holton 1967; Lettau 1967; Bonner 1968). Analyses of 47 radiosonde records for a 2-yr period by Bonner (1968) show the highest summertime LLJ frequency to be centered over northern Oklahoma and southern Kansas. Wexler (1961) proposes that the mountains of Central America deflect trade wind currents northward, thereby intensifying the southerly flow along the sloping Great Plains similar to western oceanic boundary currents. According to the paper, the highly concentrated summertime flows over the Great Plains result from larger-scale inertial effects similar to those that drive the Gulf Stream. Wexler (1961) argued that the strength of the southerly flow east of the Rocky Mountains is a prerequisite to the nighttime LLJ development (see also Ting and Wang 2006).

Letttau (1967) offered that the diurnal heating and cooling of the sloping terrain must be considered in any explanation of the LLJ. Sangster (1967) demonstrated that a diurnal oscillation of the geostrophic wind must occur in response to the heating and cooling of sloping terrain. Bonner and Paegle (1970) have also addressed diurnal heating and cooling of the Great Plains using 3-hourly analyses of geostrophic winds. Acceleration occurs as the result of the diurnal cycle of heating and cooling over sloping terrain. They note that the Holton mechanism alone produces only weak jetlike structures. It is only when the Blackadar mechanism is coupled with the Holton mechanism that strong jet structures emerge.

Discussion continues here regarding the roles of the Blackadar and Holton mechanisms in the forcing of the LLJ. In addition, analyses focus on the development of the requisite background geostrophic wind that is critical to the formation of the nocturnal wind maximum such as discussed by Wexler (1961). In this study, the LLJ environment is viewed through the lens of the North American Mesoscale Forecast System. Composite grids are assembled for a 5-yr period for cases of strong LLJs and non-LLJ episodes. Comparison of the gridded datasets enables key differences to be identified and offers another view into the Blackadar–Holton debate.

2. Composite grids for cases of the LLJ

As part of the Plains Elevated Convection at Night (PECAN; Geerts et al. 2017), a series of LLJ studies were conducted (e.g., Parish 2016; Parish and Clark 2017). Good LLJ cases were observed on some but not all evenings. A question was raised from the PECAN study: What regional or synoptic setup makes for a good summertime LLJ case? For this study, use is made of a 5-yr record (2008–12) from the North American Mesoscale Forecast System (NAM) for the summer months of June and July. Grids are taken from the 0000 UTC run and are available every 3 h. For analyses presented in this paper, gridded output on standard isobaric surfaces are used. Interpolation of model variables on native grids to isobaric grids can be problematic near sloping terrain such as seen over the Great Plains, and some care must be observed as isobaric surfaces intersect terrain adjacent to the Rocky Mountains.

Model output was stratified into two cases: strong LLJs and no LLJs. Output from the NAM for a grid point at 36.4°N, 98.6°W, near the center of the highest frequency of LLJ occurrences [see Fig. 2 in Bonner (1968)], was used to sort grids into respective cases. A “strong LLJ” (SJ) case was identified as one in which a maximum wind speed of 20 m s$^{-1}$ or more was present in the lowest 1000 m at 0900 UTC and with a 600-hPa wind less than 10 m s$^{-1}$. A “no jet” (NJ) case was selected.
when the 0900 UTC wind profile showed no wind speed in excess of 10 m s\(^{-1}\) at levels below 600 hPa. For the 5-yr record incorporating 10 months of NAM output, 64 cases were identified as a strong-jet case, corresponding to about 21% of the days. In contrast, a total of 53 days were identified as no-jet days.

Here, it is assumed that the NAM is capable of simulating the LLJ with some degree of fidelity. Parish and Oolman (2010) and Parish (2016) have used the NAM output grids previously in studies of the LLJ. The Advanced Research Weather Research and Forecasting Model (WRF-ARW) (NAM uses the Nonhydrostatic Mesoscale Model core for WRF) has been used in support of a number of PECAN cases. Parish and Clark (2017) showed that WRF-ARW can replicate the basic jet features and that the model output was consistent with available airborne observations for the 20 June 2015 PECAN case. Recent modeling studies of the LLJ using WRF-ARW [e.g., Storm et al. (2009), Werth et al. (2011), and Mirocha et al. (2016); see also the concise summary in Shapiro et al. (2016)] have shown that key characteristics of the LLJ are replicated.

To identify differences between SJ and NJ cases, averages have been computed for each 3-h time period for all days meeting the respective selection criteria. Individual SJ cases utilized grids from the previous day beginning at 1500 UTC to capture the appropriate afternoon conditions that presaged the LLJ formation. Grids from 1500, 1800, and 2100 UTC are thus taken from the preceding day; the 0000, 0300, 0600, 0900, and 1200 UTC grids are taken from the day of the LLJ case. NAM grids are available at 25-hPa increments.

Figure 1 illustrates the vertical profiles of key variables at 3-h intervals associated with the LLJ development from the composite grids at 36.4°N, 98.6°W during the 10 summertime months for SJ and NJ days. Surface pressure at this grid point is roughly 960 hPa; the first vertical level shown by profiles in Fig. 1 is at 950 hPa, about 90 m or so above ground level (AGL). This corresponds to a height of about 540 m above mean sea level (MSL).

Wind speeds (Fig. 1a) show jet maxima developing on the SJ days at a height roughly 600 m AGL or 1050 m MSL (900-hPa level). SJ wind speeds undergo a rapid increase of greater than 1 m s\(^{-1}\) h\(^{-1}\) commencing at about 0000 UTC and reach a maximum at 0900 UTC. The jet profile becomes well developed during the nighttime hours. Wind speeds decrease with height above the jet nose such that by 4000 m MSL, corresponding to the 600-hPa level, only weak winds are present. The 600-hPa level has been used previously (Parish 2016) as the approximate height above which terrain-induced forcing was minimal; conditions at that level are thus thought to reflect the synoptic setting. Wind speed profiles during NJ conditions in Fig. 1 indicate that weak winds are present below 4000 m MSL with no obvious diurnal oscillation.

Potential temperature profiles at grid point 36.4°N, 98.6°W (Fig. 1b) indicate that the SJ cases consist of temperatures that are 3–4 K higher than those of corresponding NJ cases throughout the atmospheric column from 600 hPa to the surface. Close inspection reveals a significant diurnal signal at lower levels of the atmosphere for both SJ and NJ cases. The largest oscillation occurs, not surprisingly, near the surface. The diurnal oscillation is restricted to the lowest 1500 m MSL (about 1000 AGL) or so for both SJ and NJ cases. SJ cases indicate a 0000–1200 UTC change of potential temperature at the first level shown in Fig. 1b of 9.5 K, about 2 K more than that of the NJ case.

To examine the forcing conditions for both SJ and NJ cases, vertical profiles of the geostrophic wind and hence the horizontal pressure gradient force (PGF) were examined (Figs. 1c and 1d). Profound differences are apparent between SJ and NJ cases. Geostrophic wind magnitudes in the lowest 1500 m AGL are about 15 m s\(^{-1}\) for SJ cases as compared to less than 5 m s\(^{-1}\) for the NJ cases. It can be concluded, not surprisingly, that the magnitude of the low-level PGF is the key factor that differentiates SJ and NJ cases. A similar conclusion was reached in the work of Wexler (1961), who noted the importance of a strong “basic flow” west of the Rocky Mountains on the LLJ formation. The key role of the background flow on the intensity of the LLJ has been acknowledged by others, including the recent work of Shapiro et al. (2016). Also evident in Fig. 1c is the monotonic decrease in the magnitude of the geostrophic wind under SJ conditions from about 15 m s\(^{-1}\) at the level of the LLJ to less than 5 m s\(^{-1}\) in the free atmosphere above about 4000 m or 600 hPa. Geostrophic wind magnitudes are small in the lowest 2000 m for NJ cases. Note that the geostrophic wind magnitudes at 4000 m MSL (600 hPa) are comparable for SJ and NJ composite cases.

Geostrophic wind directions for SJ and NJ cases (Fig. 1d) also display significant differences. Directions for SJ cases show little diurnal directional change. SJ geostrophic winds veer from about 210° at the lowest levels to about 250° at about 600 hPa. In contrast, NJ cases show more variation at the lowest levels with evidence of diurnal changes from southerly in late afternoon to easterly by early morning. NJ geostrophic wind directions by 4000 m MSL are predominantly from the west. A subtle difference between the SJ and NJ cases can be seen above 4000 m MSL. SJ cases show geostrophic wind directions that
retain a southerly component, whereas the NJ cases are associated with geostrophic winds with a northerly component.

To understand the forcing of the LLJ, it is necessary to examine the process by which the PGF, and hence background geostrophic flow, develops at the level of the jet. To start, an examination is made of the synoptic environment in which these SJ and NJ cases are embedded. Figure 2a illustrates the mean 500-hPa heights averaged over the entire 24-h period for the SJ cases. A broad anticyclonic circulation is present over the southern Great Plains with only weak southerly geostrophic flow across the Great Plains states from Texas northward to the Dakotas. Such a pattern is representative of summertime quiescent conditions. Given the minimal height gradient and hence weak geostrophic flow over the southern Great Plains as shown in Fig. 2a, it is apparent that no significant synoptic forcing at 500 hPa is present that drives the strong nocturnal jet.

Mean 500-hPa geopotential heights for the NJ case are shown in Fig. 2b. Weak troughing with northwesterly flow is present across the Great Plains. From an examination of the gradient of the 500-hPa height field, it is apparent that no significant synoptic forcing at 500 hPa is present that drives the strong nocturnal jet. The more-active weather pattern for the NJ composite is also borne out by inspection of cloud coverage and temperature in the composite grids. Mean daily fractional cloud cover in the NAM over a grid box defined by a lower-left corner at 35.4°N, 99.6°W and an upper-right corner at 37.4°N, 97.6°W (see shaded area in Fig. 2a) is 50.1% for NJ cases and 24.8% for SJ cases. Consequently, shortwave radiation reaching the surface is reduced during the NJ cases. Mean 1800 UTC shortwave radiation received at the surface in the NAM composite is 702 W m⁻² for NJ cases and 940 W m⁻² for SJ cases. Summertime heating of the sloping Great Plains terrain is hence more pronounced during the SJ cases, and thus, temperatures as shown in Fig. 1b are a few kelvins higher than seen for NJ cases.

Large changes are seen in both the magnitude and direction of the PGF moving downward from 500 hPa to the surface for SJ cases. Figure 2c shows the mean sea level pressure field for the SJ cases. A pronounced maximum in the magnitude of the PGF is present from Texas northward through Kansas. Orientation of the sea level isobars over the southern Great Plains indicates a
mean geostrophic wind direction from approximately 210°. Near-surface geostrophic winds across the Great Plains during SJ cases are typically 15 m s$^{-1}$, matching roughly that shown in the profiles in Fig. 1c. In contrast, the sea level pressure field associated with NJ cases (Fig. 2d) supports only weak horizontal pressure gradients (geostrophic winds less than 5 m s$^{-1}$) over the Great Plains states of Oklahoma and Kansas.

A central issue for the SJ cases concerns the transition of the tranquil synoptic pattern over the Great Plains at 500-hPa shown in Fig. 2a to the favorable background geostrophic wind for the LLJ shown in Fig. 2c. Pronounced vertical changes in the mean PGF over the Great Plains underscore the importance of the thermal wind in establishing the low-level pressure field. The impact of heating of the Great Plains terrain has been known for many years (e.g., Lettau 1967). Thermal wind forcing occurs in response to heating of the sloping Great Plains terrain. Strongest heating occurs at the surface. Given the topographic slope of the Great Plains, the warmest air is found to the west for a particular isobaric surface. Temperature gradients are thus generated since air to the west is closer to the surface than air at the same isobaric surface but situated to the east and thus farther removed from ground level. Given the large differences in the intensity of the PGF over the southern Great Plains shown in Figs. 2c and 2d, it is concluded that the thermal wind forcing is considerably larger for the SJ cases than NJ cases. In each case, the PGF at jet level occurs as a response to thermal wind forcing from above.

A transition from weak zonal flow at 600 hPa to the relatively intense southerly flow at 950 hPa over the Great Plains for the SJ cases is shown in Fig. 3. The weak anticyclonic circulation present at 500 hPa over the Great Plains in Fig. 2a is also seen at 600 hPa (Fig. 3a). Isobaric temperature gradients are present across the southern Great Plains in Fig. 3a. Warmest air is found over the elevated terrain to the west; isotherms over the southern Great Plains run parallel to contours of the sloping terrain and hence are likely the result of topographic heating. The magnitude of such a temperature

![Composite means from the NAM 5-yr (2008–12) summertime record of (top) 500-hPa heights (m) for (a) SJ and (b) NJ cases and (bottom) sea level pressure (hPa) for (c) SJ and (d) NJ cases. Shaded area in Fig. 2a represents zone for averages discussed later. Thick gray line is location of cross sections presented later.](image-url)
gradient over central Oklahoma is roughly 0.3 K (100 km)$^{-1}$. Assuming such a temperature gradient is representative of the atmosphere between 550 and 650 hPa, thermal wind constraints require that the southerly geostrophic component from 550 to 650 hPa must increase by about 1.7 m s$^{-1}$.

The anticyclonic circulation in the SJ case migrates eastward, moving downward to 700 hPa, consistent with a temperature field with warmest air to the west. The mean temperature gradient in the lower atmosphere increases significantly moving toward the surface. At 700 hPa (Fig. 3b), the temperature gradient over central Oklahoma is about twice as large as that seen at 600 hPa. If, again, the 700-hPa temperature gradient is representative of the layer from 650 to 750 hPa, an increase in the southerly geostrophic wind of about 3.1 m s$^{-1}$ must occur over that same interval.

Continued eastward movement of the anticyclonic circulation is apparent from 700 to 850 hPa. Isobaric temperature gradients in the SJ case continue to increase moving down to 850 (Fig. 3c) and 950 hPa (Fig. 3d). The impact of the enhanced isobaric temperature gradients is to increase the thermal wind forcing. As a result, the southerly component of the geostrophic wind must increase by approximately 8.5 m s$^{-1}$ from 750 to 950 hPa. Based on thermal wind constraints, the southerly component of the geostrophic wind must increase by about 14 m s$^{-1}$ from 600 hPa to the surface for the composite SJ case. Recognizing that a weak southwest component is present at 500 hPa, the geostrophic wind at the surface must contain a southerly component that exceeds 15 m s$^{-1}$. The strong background flow that is prerequisite to the LLJ thus originates because of the thermal wind forcing resulting from summertime heating of the sloping Great Plains terrain.

The above view emphasizes the critical role of isobaric temperature gradients at levels above the LLJ in establishment of the prerequisite background PGF at jet level. Heating of the atmosphere above the sloping terrain is thus vital in providing the necessary base state about which the LLJ forms. It is this thermal wind forcing that provides the key link between the Great Plains topography and the high frequency of occurrence of the LLJ.

As noted previously and illustrated in Fig. 1b, diurnal oscillations in temperature are minimal above about 1000 m AGL. Hence, the isobaric temperature gradients that develop above jet level cannot be the result of
diurnal heating. This implies that longer-term heating of the sloping Great Plains terrain is necessary in the development of isobaric temperature gradients above about 1000 m AGL. Here, it is offered that the diurnal cycle of heating and cooling alone is insufficient to explain LLJ development and an incomplete indicator of the probability of LLJ occurrence. As shown in Fig. 1b, SJ and NJ cases exhibit comparable amplitudes of the diurnal oscillation in potential temperature in the lowest 1000 m AGL yet with dramatic differences in the intensity of the nocturnal wind maximum. The PGF at the level of the LLJ, thus, must arise as a consequence of heating of the terrain on a time scale that is more seasonal than diurnal.

Importance of the longer-term heating of the Great Plains sloping terrain in establishing the low-level pressure field can be seen by inspection of mean monthly maps of isobaric heights and mean temperature gradients. As an example, Fig. 4 illustrates the mean 850-hPa heights and mean temperature gradients between 850 and 700 hPa from NAM output during the summer months of June and July (Fig. 4a) and winter months of December and January (Fig. 4b) for the 3-yr period 2008–10. Height fields show significant seasonal differences. Southerly geostrophic winds are present during summertime months, while zonal flow prevails during the winter period. The summertime 850-hPa height field arises as a consequence of thermal wind forcing through topographically induced seasonal heating that is especially apparent over the southern Great Plains. A pronounced mean east–west 850–700-hPa temperature gradient is present during summer across Texas, Oklahoma, and Kansas that helps shape the 850-hPa height field. Terrain-induced thermal wind influences between 850 and 700 hPa are essentially absent during winter. Mean meridional temperature gradients are found over the Great Plains during winter, similar to those seen elsewhere around the country. As a result, the mean southerly background flows required for development of the LLJ are not climatologically prevalent during colder-season months, and hence, the LLJ is primarily a warm-season event.

Mean NJ conditions (Fig. 5) reveal that similar thermal wind processes are at work but with significantly reduced isobaric temperature gradients and hence a smaller effect. Note that at 600 hPa (Fig. 5a), northwest geostrophic winds are present. Such a pattern results in a northerly component of the geostrophic wind of 3–4 m s$^{-1}$ over the southern Great Plains. Although relatively weak, such a midlevel geostrophic wind component works against development of a southerly geostrophic wind at lower levels. The horizontal pressure gradient force at 600 hPa that supports a northerly component of the geostrophic wind needs to be overcome by thermal wind processes in the lower atmosphere in order for southerly geostrophic winds to become established.

Isotherms at 600 hPa for the NJ case (Fig. 5a) are oriented in a mostly zonal direction and hence show little influence of the underlying sloping Great Plains terrain. Minor changes in the orientation and intensity of the geostrophic wind thus occur between 600 and 700 hPa. Isobaric temperatures begin to show some influence of the heating of the underlying terrain at about 700 hPa (Fig. 5b). Effects of insolation of the sloping terrain become more apparent closer to the surface for the NJ cases; significant isobaric temperature gradients become established at 850 (Fig. 5c) and 950 hPa (Fig. 5d). NJ temperature gradients across Oklahoma as shown in Figs. 5c and 5d are roughly half the magnitude of corresponding gradients found during the SJ cases (e.g., Figs. 3c and 3d). As a result, significantly weaker geostrophic winds are found at low levels for the NJ case.

Two factors are thus at play in the development of the PGF in the lower atmosphere as shown in the mean
composite grids for SJ and NJ cases: 1) the forcing in the free atmosphere and 2) the magnitude of the thermal wind contribution from 600 hPa to the surface. The 500-hPa geostrophic winds across, for example, northern Oklahoma and southern Kansas contain a weak (<2 m s⁻¹) southerly component for the SJ cases and a northerly component of 2–4 m s⁻¹ for the NJ cases. SJ cases thus have southerly flow already established in the free atmosphere. The thermal wind forcing is stronger for SJ cases than NJ cases and throughout a deeper column of the atmosphere (e.g., cf. Fig. 3c with Fig. 5c). The net effect is that the magnitude of the east–west component of the PGF in the lower atmosphere is significantly larger for the SJ cases.

Discussion thus far has focused on time-averaged composite means from the gridded output. Significant diurnal oscillations are, of course, present. Given the debate regarding the importance of the diurnal heating and cooling of the sloping terrain, it is appropriate to examine the daily cycle. Figure 6 illustrates wind speed and streamlines of the wind at 900 hPa, the approximate level of the maximum LLJ speeds over northern Oklahoma and southern Kansas for the SJ case, for selected times during the early evening and night. At 0000 UTC, 900-hPa wind speeds (Fig. 6a) show a maximum of about 11 m s⁻¹ over northern Oklahoma. The wind speed maximum matches the location of the maximum PGF as shown in, for example, Fig. 3c.

As the sun sets and nighttime conditions become established, 900-hPa wind speeds increase significantly. Wind speeds associated with the LLJ reach near 20 m s⁻¹ by 0600 UTC (Fig. 6b) and retain that speed throughout the evening (Figs. 6c and 6d). The zone of strongest LLJ winds extends northward from central Texas to southern Nebraska. The wind speed pattern does not change significantly through the night. Streamlines show a veering of the LLJ core, shifting over northern Oklahoma from roughly 180° at 0000 UTC (Fig. 6a) to about 210° by 1200 UTC.

Forcing of the LLJ can be inferred from the 900-hPa height and temperature fields (Fig. 7). At 0000 UTC (Fig. 7a), a strong isobaric height gradient is present over the Great Plains states. Note that during the subsequent 12 h (Figs. 7b–d), the magnitude of the horizontal pressure gradient over Oklahoma and southern Kansas changes little. The 900-hPa geostrophic wind remains fairly uniform throughout the night over the Great Plains. It is apparent, as noted many years ago...
(e.g., Lettau 1967), that the increase in the wind speed is not accompanied by a concurrent increase in the horizontal pressure gradient force.

Isobaric temperature gradients at 900 hPa display a more significant decrease during the nighttime hours. Gradients weaken rapidly from 0000 (Fig. 7a) to 0600 UTC (Fig. 7b). That the horizontal pressure gradient over the southern Great Plains remains strong throughout the nighttime hours implies that the mean isobaric temperature gradients above 900 hPa must remain intact throughout the entire diurnal cycle.

It follows that at levels above the LLJ, the mean isobaric temperature gradients from 900 to 600 hPa must not exhibit significant diurnal change.

3. Cross sections and forcing of the LLJ

To demonstrate the vertical variation of the geostrophic wind and hence the thermal wind in the development of the LLJ, $D$ values are used. Bellamy (1945) first discussed the concept of a $D$ value, which is simply the difference between the height of a pressure surface and the height of that same pressure surface in the U.S. Standard Atmosphere. Variation of $D$ values along an isobaric surface is proportional to the geostrophic wind. As noted in Parish et al. (2016), a significant advantage of using cross sections of $D$ values is that depiction of the vertical variation of the PGF is facilitated since the hydrostatic height changes are effectively removed. Hence, the interplay between the PGF and thermal wind is easily visualized.

In analyses shown here, the reference atmosphere used in the $D$-value calculations is taken from the mean daytime atmosphere over the Great Plains during summer. A mean lapse rate of 9.0°C km$^{-1}$ and a sea level pressure and temperature of 1006 hPa and 308 K, respectively, are used. Such values enable isobaric heights at the standard levels to be directly computed. The $D$ values are then calculated for each grid point by simply subtracting the height of the reference isobaric surface from the actual composite geopotential height at the same isobaric level.

Figure 8 illustrates the mean 0000 UTC cross sections of wind speed, potential temperature, and $D$ values for SJ and NJ cases along a line from 37.5°N, 105°W to 33°N, 90°W (see thick gray line in Fig. 2a). Note that for the SJ case (Fig. 8a), maximum wind speeds exceed 10 m s$^{-1}$ over a broad region of the cross section from near the

Fig. 6. Composite SJ means of 900-hPa wind speed (black lines; m s$^{-1}$) and temperatures (red lines; °C) at (a) 0000, (b) 0600, (c) 0900, and (d) 1200 UTC.
surface to about 2000 m AGL. Isentropes are oriented in a vertical direction over the lowest roughly 2000 m, indicating the well-mixed nature of the atmosphere at 0000 UTC.

Pronounced 0000 UTC isobaric temperature gradients are also present in Fig. 8a, a consequence of both daily and longer-term heating of the sloping terrain. Note that although the isobaric temperature gradients for the SJ cases are largest at the surface, they extend upward to 600 hPa. Sloping isentropes in Fig. 8a imply thermal wind forcing. Isobaric slopes of the $D$-value contours are modest at 600 hPa. Moving downward, $D$-value contours develop an increasing slope in response to the isobaric temperature gradients and hence the thermal wind. A large PGF becomes established in the lower atmosphere as evidenced by $D$-value slopes.

Figure 9 illustrates cross sections at 0900 UTC, near the time of the maximum LLJ for the SJ case. The most obvious difference between SJ and NJ cases is, of course, wind speed. A maximum wind speed in excess of 20 m s$^{-1}$ is seen for the SJ case (Fig. 9a) and less than 5 m s$^{-1}$ for the NJ case (Fig. 9b). Isentropes also show dramatic differences. The horizontal temperature gradient from jet level to 600 hPa for the SJ cases remains large at 0900 UTC. Comparison with the isentropic slopes from the 0000 UTC SJ cross section (Fig. 8a) shows that, if anything, slopes have steepened. This again implies an enhancement of the thermal wind forcing, ensuring a strong PGF at jet level. By contrast, the NJ case reveals only gently sloping isentropes and hence relatively minor thermal wind enhancements.

FIG. 7. Composite SJ means of 900-hPa geopotential heights (black lines; m) and temperatures (red lines; °C) at (a) 0000, (b) 0600, (c) 0900, and (d) 1200 UTC.
from 600 hPa down to the surface. As a result, only a weak PGF becomes established in the lower atmosphere for the NJ composite grids.

A significant feature in the SJ case is the subsidence along the western edge of the wind maximum as evidenced by the local isobaric maximum in potential temperature in Fig. 9a. Subsidence was observed on a number of LLJ cases in PECAN (e.g., Parish 2016) and is a recurring feature for strong-jet cases. Subsidence occurs in response to the horizontal divergence associated with the strong gradient in wind speed observed along the western edge of the LLJ. Although the magnitude of the wind is illustrated in Fig. 9a, wind directions (not shown) have become more westerly by 0900 UTC, and hence, a significant wind component is parallel to the plane of this cross section in a general eastward sense. Analyses (not shown) demonstrate that the subsidence is associated with the horizontal divergence. Evidence of this is suggested in Fig. 9a since the isentropic maxima align with the wind speed gradients found along the western edge of the jet.

As shown in Fig. 6, wind speed fields associated with the LLJ are asymmetric. Wind speed gradients to the west of the LLJ are far stronger than those seen to the east. Cyclonic vorticity occurs on the western side of the LLJ. The magnitude of cyclonic shear is not subject to constraints imposed on anticyclonic shear. Inertial instability conditions limit the anticyclonic shear such that the absolute vorticity must remain positive. It is offered here that the observed asymmetry of the LLJ core in Fig. 6 reflects similar dynamics that are present for upper-level jet streaks. As a result of the increased gradients associated with cyclonic shear, enhanced subsidence occurs in response to the divergence field.

Consequences of this cross-jet circulation are profound. Subsidence to the west of the LLJ core intensifies the horizontal temperature gradient in the atmosphere immediately above the jet, as can be seen by comparing isobaric temperature gradients above the LLJ at 700 hPa in Fig. 9a with those in Fig. 8a. As a result, thermal wind forcing is augmented and the PGF increases during the nighttime hours. The $D$-value slopes above the jet for the SJ case are slightly larger during the nighttime hours than those seen at 0000 UTC (Fig. 8a).

To quantify the forcing mechanisms at work in the dynamics of SJ and NJ cases, averages are computed over the area described previously and shown with the
gray shaded box in Fig. 2a. Wind speed profiles (Fig. 10a) follow a traditional evolution similar to that shown in Fig. 1a. Wind directions (Fig. 10a) reveal a veering throughout the night that matches observations discussed by others. Potential temperature profiles (Fig. 10b) show the effects of the diurnal heating in the lower atmosphere with an 8-K decrease between 0000 and 0900 UTC at the lowest level shown that corresponds to 950 hPa. At the level of the LLJ, the temperature change between 0000 and 0900 UTC is reduced to about 3 K. Note that potential temperatures increase slightly (about 1 K) above the level of the jet during the nighttime hours, a consequence of the subsidence and the development of the cross-jet circulation discussed above.

Figure 10b also displays the NAM profile of turbulent kinetic energy (TKE) during the early evening and nighttime hours. No direct assessment of turbulent friction is provided from the NAM grids, and the TKE profiles are used as a proxy. The rapid decrease in TKE shown from 2100 to 0000 UTC suggests a significantly reduced level of turbulent exchange. This supports the Blackadar (1957) contention regarding the decoupling of the surface layer from the residual layer during the early part of the evening.

Actual forcing mechanisms of the jet are provided in terms of the $x$ and $y$ components of the geostrophic wind, $u_g$ and $v_g$, respectively (Fig. 10c). Since the Great Plains slopes downward from west to east, only the $y$ component of the geostrophic wind reflects terrain-induced forcing. Effects of diurnal heating are apparent in the PGF at the lowest levels of the atmosphere. Changes in $v_g$ from 2100 to 0900 UTC mirror diurnal signals in the potential temperature profile. The $y$ component of the geostrophic wind at the lowest level (950 hPa) decreases from about 17 m s$^{-1}$ at 0000 UTC to about 11 m s$^{-1}$ at 0900 UTC. The amplitude of the change decreases with height such that at the level of the LLJ, the change in $v_g$ from 0000 to 0900 UTC has been reduced to about 2 m s$^{-1}$. The $y$ component of the geostrophic wind above the level of the jet increases by 1–2 m s$^{-1}$ during the nighttime hours, consistent with subsidence along the western edge of the LLJ and intensification of the thermal wind forcing discussed above.

Thermal wind forcing is evident in the profile of the $y$ component of the geostrophic wind. A pronounced decrease in the magnitude of the 0000 UTC geostrophic wind from 17 to about 2 m s$^{-1}$ occurs between 500 m ($\sim$950 hPa) and 4400 m ($\sim$550 hPa). Note that temporal
changes in the magnitude of the y component of the geostrophic wind above the level of the jet are small.

Profiles of the x component of the geostrophic wind display a near-uniform geostrophic wind throughout the night.

Ageostrophic wind magnitude and direction profiles are illustrated in Fig. 10d. The magnitude of the ageostrophic wind shows a general decrease with height. Largest ageostrophic wind magnitudes are found in the lower atmosphere near the level of the LLJ. Values at jet level remain roughly constant during the nighttime hours, ranging between 7 and 10 m s$^{-1}$. Ageostrophic wind directions at the level of the LLJ display classic veering with time. Angular changes average around 10$^\circ$ h$^{-1}$, less than the idealized frictionless case discussed by, for example, Markowski and Richardson (2010) or Parish and Oolman (2010) of about 18$^\circ$ h$^{-1}$ for the latitude considered. Such a result indicates that while the Blackadar mechanism is fundamental to the jet development, other factors such as turbulent exchange processes must be acting in the atmosphere below the jet maximum and influence details of the wind profile (e.g., Fedorovich et al. 2017).

In contrast, Fig. 11 shows the same terms as in Fig. 10, but for the NJ composite case. Profiles of wind speed (Fig. 11a) depict weak flow throughout the entire atmospheric column. Wind directions shift from southerly flow at lower levels to a northwesterly flow above about 3000 m MSL that corresponds to roughly the 700-hPa level. Diurnal changes are apparent in the potential temperature profiles (Fig. 11b); a decrease of about 7°C at the lowest level is seen from 0000 to 0900 UTC. As in the SJ case, the TKE profiles reveal a dramatic decrease from 2100 to 0000 UTC. From such a profile, the Blackadar (1957) mechanism could appear to operate.

The major difference between the SJ and NJ cases as noted above is the magnitude of the PGF. Note that both the x and y components of the NJ geostrophic wind are near zero at about 1100 m MSL in Fig. 11c, the level of the LLJ in the SJ case. Thermal wind forcing is present for the y component of the geostrophic wind that decreases from about 7 m s$^{-1}$ at the lowest level (950 hPa) to -1 m s$^{-1}$ at 4000 m (near 600 hPa), roughly half that seen for the SJ case. Note that at 600 hPa, weak northerly flow exists. Owing to the relatively weak thermal wind forcing for the NJ case (i.e., Fig. 8b), only minor southerly flow components become established at lower levels. The x component of the geostrophic wind increases with height, consistent with the thermal wind associated with the meridional temperature gradients shown in Fig. 5.

Ageostrophic wind magnitudes (Fig. 11d) are less than 5 m s$^{-1}$, roughly half that of the composite SJ case. Interestingly, ageostrophic wind directions show a
classic veering pattern that is characteristic of that ex-
pected from the Blackadar (1957) mechanism. Such a
profile suggests that the weak ageostrophic components
undergo an inertial oscillation similar to their SJ coun-
terparts. For the NJ case, such an oscillation is small, and
given that the geostrophic wind components are near
zero at a typical jet level, little noticeable effect is seen in
the wind.

4. The Blackadar (1957) and Holton (1967) debate

Nearly universal acknowledgment exists that evening
decoupling of the residual layer from the surface
(Blackadar 1957) is critical in the development of the
LLJ (e.g., Shapiro et al. 2016). Holton (1967) recognized
that time and height variations in the eddy viscosity such
as proposed by Blackadar (1957) are required to match
the timing and vertical structure of the LLJ. Shapiro and
Fedorovich (2009) note that based on results from a
large number of LLJ cases, a strong argument can be
made for the importance of the Blackadar (1957)
mekanism.

A schematic of the LLJ development from the SJ
cases is summarized in Fig. 12 to show the applicability
of the Blackadar construct. LLJ wind speeds increase in
response to a pseudoinertial oscillation of the ageo-
strophic wind vector as seen in Fig. 12a. Rotation of the
ageostrophic wind in a clockwise manner is apparent.
Figure 12b is offered as a comparison of results from the
composite grids with an idealized inertial oscillation
(e.g., Markowski and Richardson 2010; Parish and
Oolman 2010). For the idealized case, the initial geo-
strophic and ageostrophic winds are taken from the 0000
UTC SJ analyses. In addition, it is assumed that the
decoupling has occurred at 2300 UTC based on airborne
observations from PECAN discussed in Parish (2016).
Figure 12b shows that the observed trends in wind speed
and wind direction roughly match that expected from an
idealized case. Further, the evolution of the ageo-
strophic components (Fig. 12c) matches the general
pattern from the idealized solution. Evolution of the
wind speed, wind direction, and ageostrophic wind
components from the composite SJ grids roughly match
that expected from an inertial oscillation of the un-
balanced wind components following decoupling in the
early evening. It is concluded therefore that develop-
ment of the LLJ maximum occurs fundamentally
through the Blackadar (1957) mechanism.

Diurnal variations owing to the heating and cooling of
the sloping terrain (Holton mechanism) are present, yet
it is offered here that such forcing is of lesser importance
in the development of the LLJ. Two characteristics of
the lower atmosphere over the Great Plains during
summer are relevant to this claim. First, the amplitude of
the diurnal cycle of the horizontal pressure gradient
force from the SJ composite grids is small at the level of
the jet. From Fig. 10c, the decrease in the y component
of the geostrophic wind from 0000 to 0900 UTC at the
level of the LLJ maximum is only 2 m s⁻¹. Second, as

![Figure 12](#)
noted by many, including the early works of Sangster (1967) and Bonner and Paegle (1970), the time of the maximum horizontal pressure gradient force arising from diurnal heating of the terrain slopes is not in phase with the maximum nocturnal wind. Figure 10c shows that the x component of the horizontal pressure gradient force, and hence the y component of the geostrophic wind in the lower levels of the atmosphere, undergoes a reduction during the nighttime hours during which the LLJ develops.

In terms of the forcing of the LLJ, a somewhat different role for the Holton mechanism is proposed here. Investigators such as Lettau (1967) and Holton (1967) suggested that the diurnal heating provided a link between the sloping Great Plains terrain and the high frequency of LLJ occurrence. Here, longer-term heating is suggested as critical with emphasis on thermal wind forcing between about 600 hPa and the level of the jet. Establishment of a strong geostrophic wind in the lower atmosphere, prerequisite to the LLJ development, is the response of such heating. In terms of heating of the Great Plains, it is proposed here that the strong background PGF is more important to the development of the LLJ than diurnal oscillations.

The quarrel with the mechanism originally proposed in Holton (1967) is fundamentally about time scale. From analyses shown in Fig. 10c, diurnal heating and cooling is restricted to levels primarily below the LLJ. The thermal wind forcing that gives rise to the prerequisite strong PGF at jet level originates from the layer of atmosphere primarily above the LLJ and thus must reflect heating over a longer time scale. It is hardly surprising that the strong southerly geostrophic flow in Fig. 2c is a warm-season phenomenon.

Recent work by Fedorovich et al. (2017) has noted that the actual structure of the nighttime LLJ is sensitive to the remnant turbulent exchange in the atmosphere below the wind maximum. Actual jet profiles and the height of the LLJ maximum are influenced by the turbulent structure in the lower atmosphere, induced by the diurnal forcing. In addition, along-slope advection of potential temperature can facilitate additional turbulent exchange processes that can serve to modify details of the wind profile. Such processes are represented in a limited manner in the NAM composite grids owing to the relatively coarse grid and averaging used in this study.

5. Summary

Forcing mechanisms for the Great Plains LLJ have been explored using composite grids computed from the NAM 3-hourly grids for summer months of June and July over the 5-yr period 2008–12. Composite grids were assembled for two contrasting cases: strong-jet (SJ) and no-jet (NJ) conditions. The singular and fundamental difference between the SJ and NJ conditions is the magnitude of the PGF at the level of the jet (roughly 900 hPa over western Oklahoma). The PGF for SJ cases develops in response to thermal wind forcing from about 600 hPa to jet level. Here, it is proposed that such thermal wind forcing is a consequence of longer-term heating of the sloping Great Plains terrain. Warming of air closest to the surface results in pronounced east–west isobaric temperature gradients. Isobaric temperature gradients for SJ cases extend in the vertical to above 600 hPa. A northerly thermal wind is present as a result of the isobaric temperature gradients. Hence, the southerly component of the geostrophic wind progressively increases from about 600 hPa downward to the surface. SJ cases require a strong background PGF, a characteristic others (e.g., Wexler 1961) have pointed to as critical for the development of an LLJ. Thermal wind forcing from the free atmosphere down to the level of the LLJ establishes a strong PGF. The prerequisite background flow for the LLJ thus arises from isobaric temperature gradients resulting from summertime insolation of the Great Plains. Such thermal wind forcing is the fundamental reason why the LLJ is climatologically situated over the Great Plains (Parish and Oolman 2010) and is a summertime feature.

Analyses of the composite grids show diurnal variations are present and are largest at levels primarily below the nocturnal jet. Turbulent exchange processes that result from the diurnal heating cycle can modify the LLJ profile and height of the wind maximum as shown by Fedorovich et al. (2017). Minor changes in the magnitude of the PGF occur at the level of the LLJ; diurnal changes become less significant at levels above the wind maximum. Development of pronounced isobaric temperature gradients above jet level require prolonged periods of heating of the sloping Great Plains topography, reflecting more of a seasonal scale. That the atmosphere over the Great Plains in north-central Oklahoma does not display significant diurnal variability above about 850 hPa indicates that the thermal wind forcing from the atmosphere down to the level of the LLJ is relatively insensitive to diurnal processes.

This view of the LLJ development emphasizes the importance of decoupling of the residual layer from the surface in the early evening as a result of radiational cooling, the so-called Blackadar mechanism. Such a process sets up an inertial oscillation of the unbalanced wind component that results in a wind maximum shortly after midnight. It is proposed that the role of longer-term summertime diabatic heating of the atmosphere...
above the level of the jet over the Great Plains in the establishment of the prerequisite background PGF is more important than any enhancements due to diurnal oscillations. The critical role of heating at levels above the jet core requires the LLJ to be a warm-weather phenomenon. The above view recognizes that diurnal variations in the PGF (the Holton mechanism) do occur and can be of importance in the actual wind profile and height of the jet but are small at the level of the LLJ.

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