Importance of Low-Level Jets to Climate: A Review

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ABSTRACT

Low-level jets (LLJs) occur frequently in many parts of the world. These low-level wind speed maxima are important for both the horizontal and vertical fluxes of temperature and moisture and have been found to be associated with the development and evolution of deep convection. Since deep convective activity produces a significant amount of upper-level cloudiness and is responsible for a large fraction of the warm season rainfall in the United States, the relationship between LLJs and deep convection suggests that LLJs are important contributors to regional climate. Results from a number of past studies are reviewed, and the potential for data from the Atmospheric Radiation Measurement program to augment our understanding of low-level jets is discussed.

1. Introduction

The sky over the southern Great Plains Clouds and Atmospheric Radiation Testbed (CART) site of the Atmospheric Radiation Measurement (ARM) program during the predawn and early morning hours often is partially obstructed by stratocumulus, stratus fractus, or cumulus fractus that are moving rapidly to the north, even though the surface winds are weak. This cloud movement is evidence of the low-level jet (LLJ), a wind speed maximum that occurs in the lowest few kilometers of the atmosphere. Low-level jets have been observed over every continent (Fig. 1), although more frequent LLJs are known or suspected to occur over North America (Bonner 1968; Douglas 1993), South America (Virji 1981, 1982), Africa (Findlater 1969; Ardanuy 1979; Kelbe 1988; Jury and Spencer-Smith 1988; Jury and Tosen 1989), Australia (Wilson 1975; Brook 1985; Keenan et al. 1989), Asia (Findlater 1969; Tao and Chen 1987), and Antarctica (Schwerdtfeger 1975; Chiba and Kobayashi 1986). These regions of frequent LLJ occurrence typically are located either to the east of a large mountain range or where large land–sea temperature gradients exist. In the middle latitudes, LLJs typically are more frequent during the summer months.

In any discussion of the LLJ, some distinction must be made between jets that appear to be related to synoptic-scale forcing and have narrow zones of high speed flow that extend for hundreds of kilometers and jets that have either significant diurnal cycles or occur only in localized regions. Reiter (1963, 1969) argues that jets that have significant diurnal cycles associated with the development of nocturnal inversions should be named inversion wind maxima since they have small horizontal shears and therefore are not true laterally confined jet streams. Only those jets with appreciable vertical and horizontal shear should be named LLJs. In addition, he states that small-scale processes that create jetlike vertical profiles, but in which the Coriolis force is not important, also are not to be considered LLJs. However, the term inversion wind maxima has never been widely used in the literature, whereas the term LLJ has been used extensively for lower-tropospheric jets of all types. For example, a LLJ typically is defined by examining the vertical profile of the horizontal wind to determine whether or not a low-level wind speed maximum occurs, with no consideration given to the horizontal shear. Bonner (1968) classified LLJs into three overlapping groups based upon the magnitude of the maximum wind speed, while also requiring that the wind speed decreases by a specified amount above the height of the maximum in order to give a jetlike profile. Macklin et al. (1990) describe a mountain-gap wind produced LLJ over Cook Inlet, Alaska, that is a significant threat to mariners and aviators, even though it occurs over a relatively small region. The jets in both of these studies do not conform to the definition proposed by Reiter (1963, 1969) and illustrate the wide variety of phenomena that are described as LLJs.

These examples provide ample evidence that the use of the term LLJ is almost solely based upon the vertical profile of the horizontal wind. Yet there is good cause to try to distinguish between the many types of LLJs since the temporal and spatial scales associated with them can be vastly different, as can the forcing mech-

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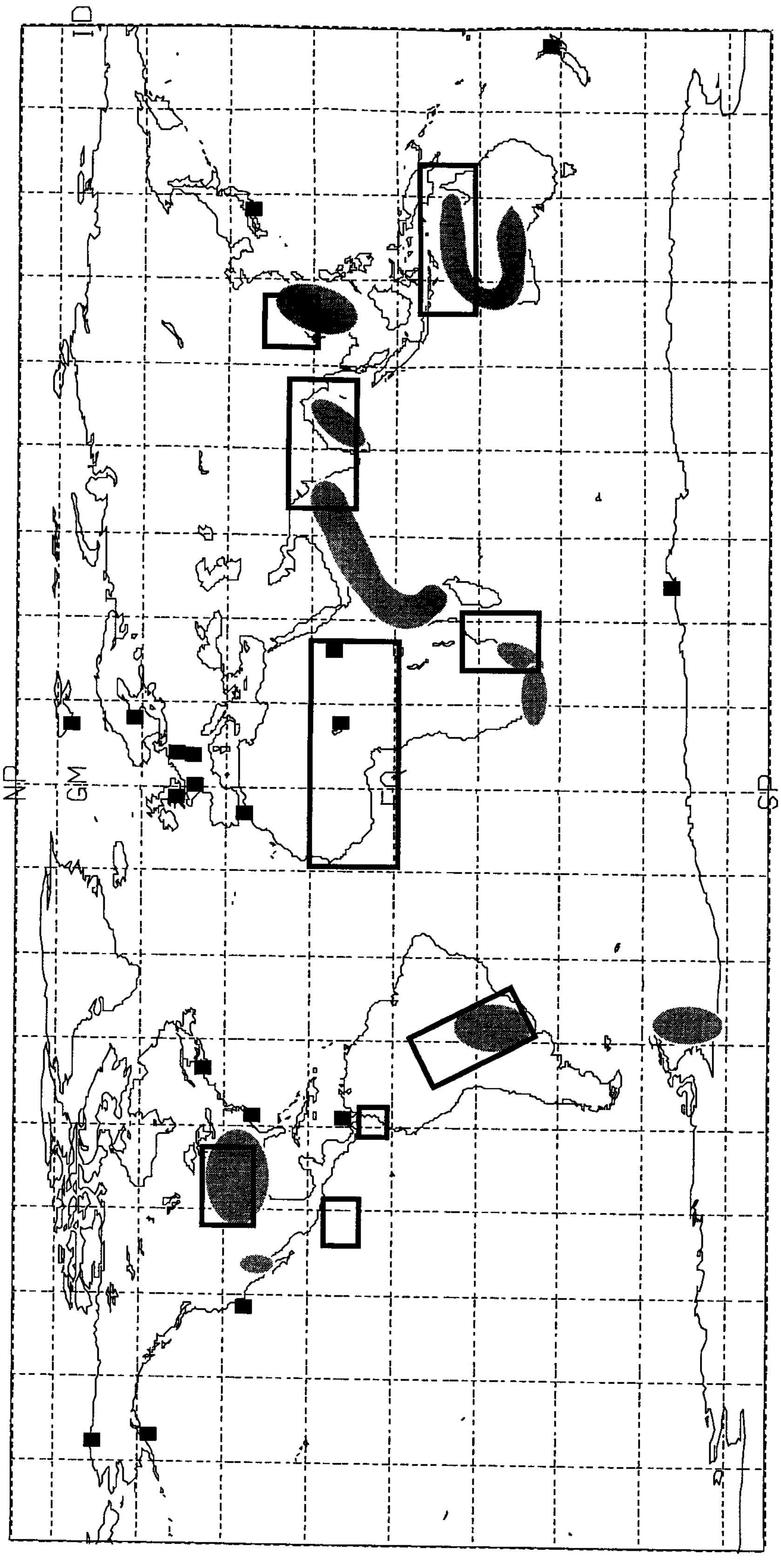

FIG. 1. Regions where low-level jets are known or suspected to occur with some regularity (shaded) and where mesoscale convective complexes are known to occur frequently during the summer (open boxes). Squares denote locations where low-level jets have been observed.

anisms. As an attempt to make this distinction, in this paper the term low-level jet stream (LLJS) is used to define a narrow horizontal zone of high-speed flow that extends for some considerable horizontal distance.[1] Often a LLJS does not have a significant diurnal cycle, is usually coupled to an upper-tropospheric jet stream in midlatitudes, and is analogous to an upper-tropospheric jet stream located in the low levels. In contrast, a LLJ is defined by examining the vertical profile of the horizontal wind to determine whether or not a low-level wind speed maximum occurs. This is the definition of LLJs that is commonly used. Therefore, it is possible to have a LLJ that is not a LLJS since a LLJ could be horizontally restricted to a small geographic region or could occur over such a large area that there are no significant large-scale horizontal wind shears. However, LLJSs also fit the less restrictive definition for LLJs, and can be viewed as a subset of LLJs. The distinction between LLJs and LLJSs is highlighted wherever appropriate.

Although LLJs were first described in the 1930s over Africa (Goualt 1938; Farquharson 1939), it was not until the 1950s that interest in the LLJ blossomed. In one of the earlier studies, Blackadar (1957) noted that while LLJs occur during the daytime, they are almost always better developed at night. He also found that over the Great Plains the wind speed at the level of jet maximum can be considerably supergeostrophic, as did Means (1952), and the height of the wind speed maximum usually coincides with the height of the top of the nocturnal inversion. A special field program to sample the LLJ with high horizontal and temporal resolution was conducted during 1961 in which hourly pilot balloons were launched from 13 sites stretching east to west from Little Rock, Arkansas, to Amarillo, Texas. These observations indicated that the height of the level of maximum wind speed within the LLJ was variable across the network, maximum winds within the LLJ were supergeostrophic at night, the level of maximum wind was located between 300 and 700 m above ground level (AGL) and did not always correspond to the height of the inversion (in contrast to the results of Blackadar), and that two separate wind maxima were sometimes observed at the same time and location but at different heights (Hoecker 1963).

A climatological study of the LLJ was conducted by Bonner (1968) in which two years of four times per day rawinsonde data over the entire United States were examined for the presence of LLJs. Bonner found that LLJs are fairly common occurrences over the central and eastern United States and are more frequent during summer over the Great Plains than during winter. He also confirmed that LLJs are primarily a nighttime feature, with 75% of the jet events sampled in the early morning (1200 UTC or roughly 0600 local time). The level of maximum wind speed seen in the two years of data can either increase or decrease with height during the night, and the correlation coefficient between the level of maximum wind speed and the inversion top is only 0.53, explaining roughly 25% of the variance in jet altitude. Uccellini (1980) examined these cases and others that were reported in the literature and found a strong link in the evolution of the LLJ and the propagation of upper-level jet streaks over the region of jet development in 12 of the 15 cases he examined. He clearly shows the importance of leeside cyclogenesis or leeside troughing on the eastern slopes of the Rocky Mountains to the production of the low-level pressure gradients needed for the development of LLJs in the Great Plains.

Nine years worth of rawinsonde data from three stations in the southeastern United States were used to produce average profiles of the diurnal variations in low-level winds at three hourly intervals (Hoxit 1975). Results show that the diurnal variation in wind speed is greatest during summer, with the mean wind maximum located near 500 m AGL at 0300 local time. The diurnal changes of the angle between the wind direction and the orientation of the surface isobars shows that the winds at all heights veer during the night, with the largest rate of veering occurring at the time of maximum low-level stability (0600 local time).

Wind data over an entire year from six different observation levels on a 447 m tall tower located in central Oklahoma were analyzed by Crawford and Hudson (1970). Their analyses show that the annual mean wind speeds below the 90-m observation height are lowest at night and highest during the day. In contrast, winds above the 90-m observation height are highest at night and lowest during the day, a wind cycle representative of a LLJ. The wind direction also was found to have a distinct diurnal cycle in which the annual mean winds back during the day and veer at night. Veering of the wind during the night is a typical LLJ behavior. However, most of the changes in the annual mean wind speed occur in a relatively short period just after sunrise and just before sunset and are attributed to the onset and decay of convective mixing within the boundary layer.

Numerical models have been used extensively to study LLJ development and evolution and have reproduced the basic features of many observed LLJs. Typically these studies have been conducted using either a case study approach (Brill et al. 1985; Uccellini et al. 1987; Lapenta and Seaman 1990; Doyle and Warner 1993) or simplified analytic initial conditions to examine LLJ sensitivities to various model parameters (Paegle and Rasch 1973; Krishnamurti et al. 1976;

[1] This approach agrees with that proposed by Hoecker (1963), who states that "the author would like to see the term 'low-level jet' applied only to boundary-layer winds such as conceived by Blackadar and Wexler and as described in this paper, and not to include horizontal wind maxima in the 850-mb level which may only be a reflection of higher-level tropospheric wind maxima."

McNider and Pielke 1981; Fast and McCorcle 1993; Savajarvi 1991). The advantage of using numerical models is the ability to separate the effects of various physical processes on the jet evolution.

All these studies indicate that LLJs occur across the globe in a wide variety of large-scale environments and in all seasons, even though the United States has been a focus for many of the studies. In order to place the LLJ in a clearer perspective relative to global climate studies, the relationship between the LLJ and regional climate is discussed in section 2 and the mechanisms of LLJ formation are reviewed in section 3. Observations of LLJs from near the ARM CART site are illustrated in section 4, followed by a discussion in section 5.

2. The LLJ and regional climate

One of the reasons why interest in the LLJ increased rapidly during the 1950s, and why interest continues today, is that LLJs have been shown to be related to deep convective activity. Means (1954) concluded that most of the moisture transported into a region of persistent, heavy convection was brought in by the LLJ and that over a 48-h period this moisture transport was large enough to produce a region of rainfall covering the entire state of Kansas with 4–7 cm of water. In a case study of a severe weather event, Uccellini and Johnson (1979) computed moisture and sensible heat transports and found that the moisture transport increased by a factor of 3 and the sensible heat transport increased by a factor of 2 owing to the development of a LLJS. On a larger-scale, Rasmussen (1967) showed that the mean water balance for northern North America is determined mainly by low-level flux across the Pacific coast, the Atlantic coast, and the southern United States border. Across the southern border, the low-level eddy flux is strongest during the summer months and accounts for a large portion of the mean annual inflow. In addition, there is a pronounced diurnal flux difference that is a maximum during the summer, with the low-level northward moisture flux larger in the morning than in the early evening. Rasmussen (1967) determined that much of this behavior is due to the diurnal character of the LLJ. Roads et al. (1994) documented that there is a good correlation between moisture flux convergence determined from models and observed precipitation over much of the United States. However, many controversies on the hydrologic cycle over North America remain that may be alleviated by improved models and analyses (Roads et al. 1994).

However, LLJs do more than just transport moisture. The proper superposition of LLJs with upper-level jets may enhance upward motion throughout much of the troposphere and assist in the development of convection (Beebe and Bates 1955). An examination of calculated vertical velocity fields from ten LLJ events indicates that the greatest ascent occurs downstream of the horizontal jet maximum (Bonner et al. 1968). Thunderstorm activity in the central United States, which has a maximum at night (Wallace 1975; Balling 1986), has been closely related to the production of these regions of ascending motion associated with LLJs (Pitchford and London 1962). Porter et al. (1955) found that in 171 squall line events, a LLJ was present in over 75% of the cases. While it is likely that LLJs by themselves do not cause the development of convective activity, since they produce broad regions of ascending motion, they help to produce a favorable thermodynamic environment for deep convection (Beebe and Bates 1955) and may be a mechanism for prolonging the lifetimes of regions of convective activity as well (Bonner 1966). Along a similar line of thought, Carbone et al. (1990) suggest that the nocturnal maximum in thunderstorm occurrence over the Great Plains may be due to an interaction between convection disturbances that formed over the Rocky Mountains and unstable conditions in the vicinity of the LLJ.

Several studies have described instabilities that can be produced by a LLJ wind profile and influence convective development. Numerical solutions to the linear, inviscid, Boussinesq equations shows that internal gravity waves can be generated by LLJs (Mastrantonio et al. 1976). Raymond (1978) showed that LLJs within the planetary boundary layer can cause dynamic instabilities, similar to symmetric and parallel instabilities, that produce boundary-layer rolls and may influence the development of deep convective activity. Kuo and Seitter (1985) show that a geostrophic LLJ can produce an unstable mesoscale disturbance that resembles frontal cloud bands and squall lines. In addition, Lemaitre and Brovelli (1990) show that the basic flow associated with a LLJ produces a baroclinic symmetric instability that leads to a narrow mesoscale line of sloping motions.

The relationship between LLJs and convective activity also has been highlighted in studies of mesoscale convective complexes (MCCs) (Maddox 1980). A MCC can be described generally as a very large cloud and precipitation system that includes a group of cumulonimbus clouds during most of its lifetime and, when viewed from satellite, produces a nearly circular cloud shield [see Maddox (1980) for the specific definition]. The life cycle of a typical MCC has three parts: initiation during the late evening, maximum extent during the night, and dissipation during the morning. In a composite of ten MCC cases, Maddox (1983) found that a strong LLJ is a recurrent feature of the precursor environment in which MCCs form and also is present in the environment of mature systems. Strong low-level convergence, warm advection, and ascending motion are present within the LLJ region in the environment prior to MCC development. However, the development of MCCs usually can be traced back to thunderstorms that initiate in the late afternoon (Wetzel et al. 1983; Maddox et al. 1986), suggesting that the important role of the LLJ in these systems is the production of a favorable environment in which convection can organize and persist. This hypothesis is supported by noting that strong LLJs are not present during the dissipation stages of MCCs when divergence, cold advection, and descending motion are prevalent.

In a suite of studies that have examined the global distribution of MCCs, it has been found that MCCs occur seasonally in many regions of the world, including North America (Rodgers et al. 1985; Augustine and Howard 1988), South America (Velasco and Fritsch 1987), Asia (Miller and Fritsch 1991; Laing and Fritsch 1993a), Australia and the western Pacific (Miller and Fritsch 1991), and Africa (Laing and Fritsch 1993b) (see Fig. 1). Most of these regions of significant MCC activity also are regions of frequent LLJ activity as well. Since the MCC life cycle is very similar to that of a LLJ, both having a significant diurnal component, this consistent colocation of regions of high MCC and LLJ activity provides strong circumstantial evidence that the LLJ may be a key ingredient in MCC development and evolution.

It is this relationship between the LLJ and convective activity that is one of the main reasons why LLJs are an important consideration in studies of global climate. It is well known that MCCs account for a large portion of warm season rainfall over the United States (Fritsch et al. 1986; Heideman and Fritsch 1988) and produce significant amounts of high-level cloudiness as seen in monthly frequency counts of cloud-top temperatures (Maddox et al. 1992). It also is well known that the largest uncertainties in the prediction of global climate are cloud–radiation interactions (Ramanathan et al. 1989), including the effects of upper-tropospheric stratiform clouds associated with deep convection (Randall et al. 1989), and the treatment of the surface energy budget (Gutowski et al. 1991; Randall et al. 1992; Stamm et al. 1994). Since both of these uncertainties are influenced strongly by the presence of deep convection and its effect on soil moisture, the LLJ becomes an important mesoscale weather phenomenon in modeling climate on regional and global spatial scales and seasonal timescales.

Additionally, LLJs can affect climate in ways that are not related to convective activity. Schwerdtfeger (1975) discusses the frequent occurrence of high southerly and southeasterly surface winds along the eastern coast of the Antarctic Penninsula during winter. The surface stress produced by these high winds, a reflection of a LLJ, are responsible for a significant transport of pack ice into the northern Weddell Sea, where the ice comes under the influence of more westerly winds. The pack ice is then transported eastward and slowly melts, producing a large region of sea surface temperatures (SSTs) roughly 6°C cooler than are observed either to the west or east of this region. In addition, LLJs have been observed along sea–ice boundaries and may influence the movement of the marginal ice zone (Chu 1986; Langland et al. 1989). Low-level jets associated with drainage flows in the Antarctic appear to be essential in prescribing the larger-scale circulations near the South Pole (Parish and Bromwich 1991).

3. Mechanisms of low-level jet formation

There are a number of physical mechanisms that have been shown to explain many aspects of the development and evolution of LLJs in a wide variety of environments. These are important to understand since, once the mechanisms are known, the influence of model grid spacing and physical parameterization schemes can be evaluated for the likelihood that a given model can simulate LLJs.

a. *Inertial oscillation*

Since vertical turbulent mixing through a jet is not favorable for jet maintainance and observations indicate that LLJs are stronger at night, Blackadar (1957) hypothesized that it is the diurnally varying eddy viscosity that leads to LLJ formation. During the daytime, the planetary boundary layer (PBL) tends to be coupled strongly with the surface layer and frictional effects cause the winds to be subgeostrophic. Once the effects of turbulent mixing cease late in the day, the frictional effects are reduced significantly and the winds above the shallow nocturnal inversion, and within the residual layer from the PBL formed earlier in the day, are decoupled from the surface layer and are no longer in balance. The imbalance between the pressure gradient and Coriolis forces induces an inertial oscillation of the wind. This oscillation has a period of one-half pendulum day, near 17 h in midlatitudes. Thus, the LLJ formed through this mechanism would produce a wind speed maximum on the order of 8 h after the cessation of turbulent mixing (see Hoxit 1975).

b. *Shallow baroclinicity*

In regions where there is a significant change in surface characteristics, such as occur in coastal regions or near marginal ice zones, the horizontal differences in sensible and latent heat fluxes produce strong low-level baroclinicity within the PBL. This shallow region of baroclinicity produces LLJs through strong geostrophic forcing, where the LLJs are oriented parallel to the low-level horizontal temperature gradient (Krishnamurti et al. 1976; Enfield 1981; Mizzi and Pielke 1984; Zemba and Friehe 1987; Langland et al. 1989; Doyle and Warner 1993). Low-level jets produced through this mechanism can be relatively constant throughout the day if associated with the nearly constant temperature gradients found near sea–ice boundaries or gradients in sea surface temperature. However, in regions where the surface fluxes have a diurnal component, such as near sea–land boundaries, significant diurnal changes in the strength of the LLJ occur.

A LLJ also can occur owing to the effects of baroclinicity produced by sloping terrain as first proposed by Bleeker and Andre (1951) and analyzed by Holton (1967) and Lettau (1967). Assuming that the surface temperature along the sloping terrain is everywhere the same and heats at the same rate, then by midday horizontal temperature gradients are produced through the effects of diurnal heating. The air near the surface over the high terrain is warmer than the air well above the surface over the lower terrain. This situation persists until the surface begins to cool as the sun sets. Radiational cooling occurs at the surface, such that the air near the surface over high terrain is cooler than the air well above the surface over lower terrain. This diurnal cycle in the horizontal temperature gradient leads to a diurnal cycle in the geostrophic wind as can be seen through the thermal wind relationship. As is often observed over the central United States, if a surface southerly geostrophic wind is assumed, then, if the air is warmer to the west over higher terrain, the southerly geostrophic wind decreases with height. This situation is reversed at night with the southerly geostrophic wind increasing with height and leading to a jetlike vertical wind profile. Calculations of the diurnal changes in geostrophic wind along the eastern slopes of the Rocky Mountains indicate magnitudes approaching 10 m s^{-1} (Sangster 1967; Bonner and Paegle 1970).

The development and evolution of extratropical cyclones produces large regions of significant low-level baroclinicity, and LLJSs are a common feature in these systems (Newton 1967; Kotroni and Lagouvardos 1993). Djuric and Damiani (1980) found that LLJSs occur frequently in association with the development of extratropical cyclones on the lee side of the Rocky Mountains. Temporal variations in the synoptic-scale pressure field can increase the wind speeds within LLJSs during cyclone intensification and movement (Young 1973; Mahrt 1974). In addition, diabatic effects have been shown in numerical simulations to have a significant influence on LLJS development (Uccellini et al. 1987; Nicolini et al. 1993).

The presence of a mountain range can block the low-level flow of a cold, stable air mass and channel this air mass along the mountain slopes. This process is called cold-air damming and is a frequent concern, for example, in the eastern United States along the Appalachian Mountains (Stauffer and Warner 1987). The low-level stable air trapped against the mountains produces a mesoscale temperature gradient directed normal to the orientation of the mountains. Assuming that this feature persists sufficiently long for the flow to become geostrophically balanced, a LLJ oriented parallel to the mountain can develop (Schwerdtfeger 1974). These barrier jets have been observed in the United States (Forbes et al. 1987; Parish 1982) and Antarctica (Schwerdtfeger 1975), and are likely to occur in other parts of the world as well.

c. Terrain effects

Diurnal heating in regions of complex terrain produces both slope and valley wind systems that can cause LLJs. Pamperin and Stilke (1985) show a LLJ that developed in the mouth of Austria's Inn Valley owing to a down-valley wind system. This LLJ attained its maximum speed of 14 m s^{-1} at 200 m AGL by sunrise. A stronger LLJ was observed over Cook Inlet, Alaska, where wind speeds in excess of 20 m s^{-1} at 80 m AGL were observed from aircraft (Macklin et al. 1990). The presence of this gap wind LLJ was detected over a region 35 km wide and 200 km long. Gap winds occur as the flow is accelerated through a channel owing to the horizontal pressure gradient oriented along the direction of the air flow and have been observed in the United States, Canada, Japan, Africa, and Europe (see Overland and Walter 1981 and references therein). Paegle et al. (1984) show that LLJs also can be produced through the effects of terrain blocking due to increased stratification at night.

Terrain features also can create boundary currents as described by Wexler (1961) in association with the Great Plains LLJS. As the typical summertime pattern of a surface anticyclone over the western Atlantic forms, low-level air is advected westward around the high in the lower latitudes. This air runs up against the Rocky Mountains and must be deflected northward. Assuming that potential vorticity is conserved, an acceleration occurs as the parcels move northward. This accelleration can produce a LLJS, although it provides no mechanism for any diurnal changes in jet structure as are often observed. This mechanism also is important for the formation of the East African LLJS (Krishnamurti et al. 1976).

d. Isallobaric forcing

In the late 1960s it became apparent that there is a relationship between upper-level jet streaks and the formation of LLJSs (Reiter 1969). Some LLJSs appear to develop in association with synoptic-scale forcing, have a minimal diurnal oscillation, and often extend above the PBL depth. Uccellini and Johnson (1979) document a case where a LLJS develops within the lower branch of a transverse, ageostrophic circulation associated with the exit region of an upper-level jet streak. They found that the LLJS intensified owing to an increased isallobaric wind component in the low levels. Uccellini (1980) examined 15 LLJS cases reported in the literature and found that in 12 of these cases the LLJSs are located in the exit region of the upper-level jet, are well-defined by the afternoon, and extend above the PBL. However, even these apparently synoptically forced LLJSs had a tendency for the maximum winds to be observed during the morning.

e. Vertical parcel displacement

In a model simulation of a secondary coastal cyclone, Uccellini et al. (1987) found that a LLJ with

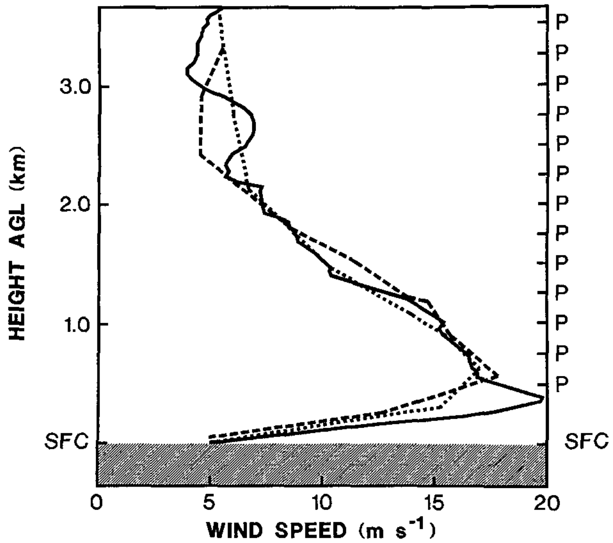

FIG. 2. Profiles of wind speed vs height from a high-resolution research rawinsonde (solid), a nearby National Weather Service rawinsonde (dashed), and the high-resolution rawinsonde data altered using an approximation to the averaging routine used by the National Weather Service (dotted). Letter *P* indicates the 404-MHz radar wind profiler range gate heights. From Stensrud et al. (1990).

wind speeds in excess of 30 m s^{-1} developed in response to the vertical displacement of air parcels within a baroclinic environment. As parcels approached the developing cyclone from the northeast, a change in pressure gradient force was experienced. While this change was modest in the horizontal direction, it was quite large in the vertical as parcels moved through the baroclinic region associated with the coastal front. As parcels were displaced vertically, a rapid increase in the ageostrophic wind component occurred that subsequently led to parcel accelleration and the rapid development of a LLJ. This is a manifestation of the three-dimensional adjustment process that occurs within the transverse ageostrophic circulations associated with an upper-level jet streak as the jet is modified by diabatic processes and frontogenesis within a strongly baroclinic environment.

From the above discussion it is apparent that a number of different mechanisms can be used to explain the formation of a LLJ. Typically, none of these mechanisms alone can explain the observations, as found by Buajitti and Blackadar (1957), who illustrated that diurnally varying eddy viscosity alone cannot give a satisfactory explanation of the diurnal wind structure. Bonner and Paegle (1970) show that a combination of the baroclinicity over sloping terrain plus a diurnally varying eddy viscosity yields the best reproduction of the observed wind variations associated with the LLJ over the southern plains of the United States. This work is extended by Paegle and Rasch (1973), who indicate that horizontal variations of the flow are important to include also since they contribute substantially to the accurate modeling of jet development. Newton (1981) showed that the East African LLJS is dominated by a frictionally damped inertial oscillation superimposed upon a eastward-decreasing geostrophic zonal current. Similarly, both baroclinicity and inertial oscillations are found to produce a LLJ that was observed over the Carolinas (Doyle and Warner 1993). These studies highlight the importance of terrain effects, the diurnal cycle of sensible and latent heat fluxes, large-scale forcing, and PBL evolution to LLJ structure. However, much of the research on LLJs has been accomplished either in a case study approach using high-resolution observations from field campaigns and model data or in a climatological approach using observations with fairly coarse temporal resolution. One of the advantages of the ARM program, in conjunction with the new sensors being deployed by the National Weather Service, is the continuous operation of remote sensing systems across the United States. Data from two of these systems, a 915-MHz radar wind profiler at the ARM CART site and the Weather Surveillance Radar—1988 Doppler (WSR-88D), are used to illustrate the potential that long-term, continuous datasets have to improve climate predictions.

4. LLJ observations from the ARM CART site

There are numerous challenges to observing the low-level jet accurately using present observational systems. Stensrud et al. (1990) report that the heights of the wind maxima for three LLJs observed during summer 1988 were below 500 m, the height of the first range gate of the National Oceanic and Atmospheric Administration (NOAA) 404-MHz radar wind profiler used to provide wind information along the boundaries of the southern Great Plains CART domain. The presence of this low-level height interval void of measurement capability clearly indicates that the standard operational mode of the 404-MHz radar wind profiler is unable to sample the summertime LLJ routinely with a high degree of accuracy (Fig. 2). Stensrud et al. (1990) also show that, although the National Weather Service rawinsondes provide observations every six seconds, the operational wind algorithm is designed to smooth the wind profile over a two-minute interval. This smoothing produces a very different jet structure than indicated by the raw data (Fig. 2), and only the smoothed data are archived.

The WSR-88Ds, currently being deployed across the United States, produce wind profiles every half-hour using the velocity azimuth display (VAD) technique (Browning and Wexler 1968; Rabin and Zrnić 1980), a potentially very valuable dataset for observations of the LLJ. Unfortunately, the winds are calculated at intervals of 304 m, which may not be sufficient to observe all important aspects of the LLJ structure (see Sisterson and Frenzen 1978). While Stensrud et al. (1990) in-

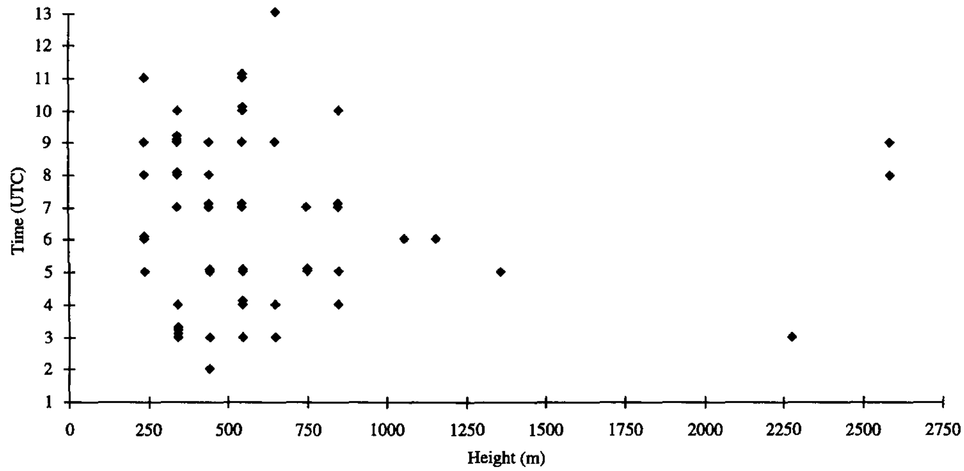

FIG. 3. Plot of the time of maximum wind speed (UTC) from the low-level jets observed using the ARM program 915-MHz radar wind profiler vs the height of this wind speed maximum (m).

dicate good agreement between rawinsonde and Doppler radar-derived wind profiles, more recent studies have shown significant differences between the remotely sensed and in situ observations owing to migrating point targets (Mclaughlin 1993; Wilczak et al. 1995). These differences require that a high degree of quality control be used with the remotely sensed data. Therefore, it appears that while the ability to observe the LLJ is available, numerous problems can arise that make accurate measurement of the LLJ difficult to ascertain without corroborating in situ synoptic observations.

In spite of the difficulties involved in remote wind sensing, a continuous, long-term record of the low-level winds at a point can be very valuable, particularly once the influences of bird and insect contamination are removed. Beginning in the spring of 1994, a 915-MHz radar wind profiler became operational at the ARM CART site near Lamont, Oklahoma (Stokes and Schwartz 1994). Data from this wind profiler are available at hourly intervals. In one sampling mode, the lowest range gate is located at 138 m AGL, the highest gate is located at nearly 2.6 km, and the gate spacing is 102 m. In the second sampling mode, the lowest range gate is located at 320 m AGL and the highest gate is located at above 5 km. This combined data sampling strategy provides sufficient vertical resolution to capture the LLJ in many instances. In addition, the network of WSR-88D radars continues to expand, providing half-hourly vertical profiles of the horizontal wind from across the nation.

An examination of the ARM profiler data from June 1994 shows that the data coverage during this month is very good, with data missing from only one day. Of the 29 days with data, LLJs were sampled clearly on 28 days [the contamination of the June profiler data by migratory birds is less than the contamination that occurs later in the summer (R. L. Coulter, personal communication, 1994)]. After June the data coverage is significantly less complete. However, a total of 56 LLJs are indicated in the available data between 1 June and 2 October 1994. These data show that the level of maximum wind speed within the LLJ occurs throughout a large range of heights (Fig. 3). The level of maximum wind speed is below 500 m AGL for over 46% of the LLJs, indicating that summertime LLJs frequently develop very close to the ground surface and are impossible to observe accurately with the NOAA 404-MHz radar wind profilers as presently configured. These LLJ wind profiles also would be altered significantly if the National Weather Service 2-min wind smoothing algorithm was applied. In addition, only 17% of these summertime LLJs produced maximum wind speeds within ± 2 h of the synoptic observation times, whereas 57% of the wind speed maxima occurred within ± 2 h of 0600 UTC.

The ARM profiler and WSR-88D VAD wind data also show that for some LLJs wind speed maxima can occur at several different times during the night and that these maxima can be located at different vertical heights. In particular, the LLJ on 14 June 1994 has significant mesoscale variability, as can be seen by examining the ARM profiler data and VAD wind data from the WSR-88D sites at Twin Lakes (TLX), Oklahoma, and Wichita (ICT), Kansas (Fig. 4). These three locations fall along a line very nearly north–south, with TLX located on the southern end and ICT located 240 km farther north. The ARM site is roughly equidistant from the two WSR-88D sites. The winds at all three locations veer throughout the night, with the highest wind speeds occurring near 1200 UTC at the ARM and TLX sites. The LLJ at both of these sites appears to ride on top of the developing convective boundary layer after sunrise for heights up to 1000 m

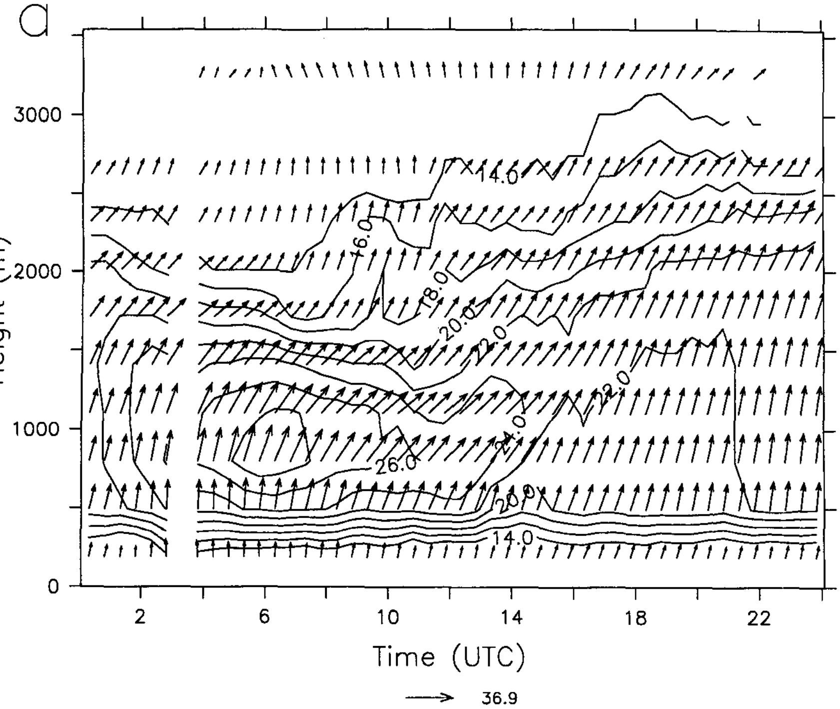

FIG. 4. Wind speed vs height contoured at 2 m s^{-1} intervals from (a) the ICT WSR-88D, (b) the ARM program 915-MHz radar wind profiler, and (c) the TLX WSR-88D from 14 June 1994. Times are in UTC or JJJHH where JJJ is Julian day. Vectors indicate wind direction at each data point (standard compass). Wind speeds are not contoured below 14 m s^{-1}.

above the original level of maximum wind speed. In contrast, the winds farther to the north at ICT show a maximum near 0600 UTC, in agreement with a local wind maximum observed in the ARM profiler data. This earlier wind speed maximum is barely detectable in the data from TLX. Thus, significant changes in LLJ structure and evolution can occur over relatively small horizontal distances and can be sampled using the new network of remote sensors. Such detailed data on both temporal and spatial evolution of the LLJ provides a significant opportunity to evaluate the ability of models to reproduce these important weather phenomena.

The times of the maximum wind speed within the LLJs sampled using the profiler vary between 0200 and 1300 UTC (Fig. 3). Maddox (1985) found that the LLJ evolved differently during the night in different air masses, with the LLJ attaining its maximum wind speed earlier within a dry air mass than within a moist air mass. If the ARM program LLJ data are scrutinized for only those LLJs that have a strong diurnal signal, as would be expected if the LLJ is produced by a combination of diurnal oscillations of heating and cooling over sloping terrain and an intertial oscillation, then there is some indication that a relationship exists between the surface value of relative humidity during late afternoon and the time of the wind maximum within the LLJ (Fig. 5). Although this relationship is far from perfect, it does suggest that the effects of higher moisture levels in the PBL act to alter the effects of radiational cooling such that the timing of the LLJ is affected (Maddox 1985). This type of analysis is but one of the potential benefits of a continuous, long-term sampling of the low-level wind profile and may lead to an improved understanding of the atmospheric and surface processes that influence, and are influenced by, the development and evolution of LLJs.

5. Discussion

Numerous studies have documented the importance of LLJs to moisture transport and deep convection, including indirectly the effects of convection on surface fluxes and cloudiness, illustrating clearly that the LLJ is a phenomenon of importance to the simulation of climate on global and regional spatial scales and on seasonal timescales. Unfortunately, current routine observing systems typically sample the lower troposphere either at 12-h intervals, at vertical levels beginning above 500 m AGL, or at 304-m vertical intervals. These temporal and spatial sampling schemes miss much of the LLJ structure, making it difficult to examine the skill of present numerical weather prediction models to simulate the development and evolution of the LLJ. This previous lack of routine observations of the LLJ

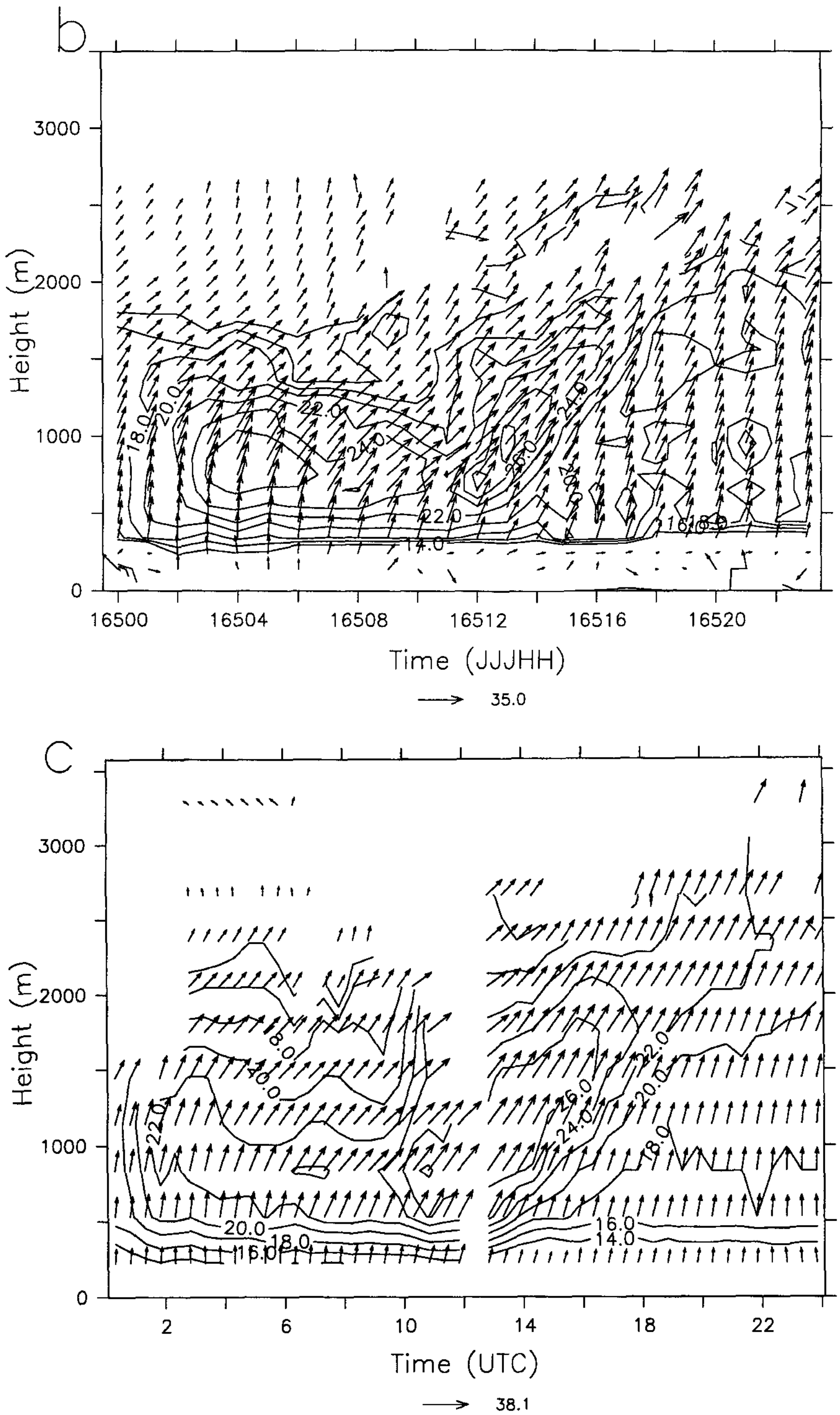

FIG. 4. (*Continued*)

enhances the importance of the continuous, long-term record of low-level winds that is being created at the ARM CART site and with the national network of WSR-88Ds and that will be created at the ARM sites in the tropical western Pacific and on the north slope of Alaska (Stokes and Schwartz 1994). These data provide an opportunity to assess the ability of present and future numerical models to simulate the development

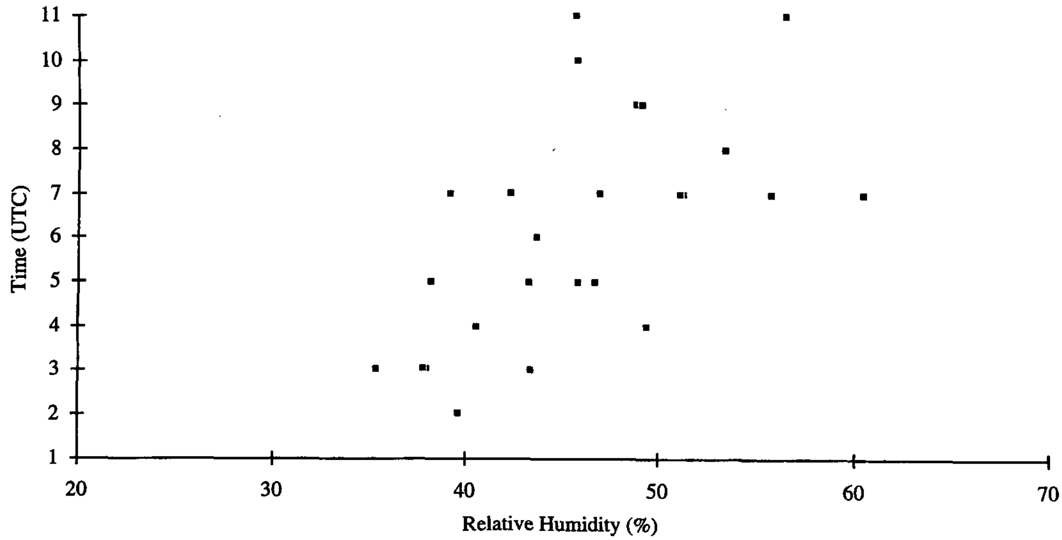

FIG. 5. Plot of the time of maximum wind speed (UTC) from a subset of the low-level jets observed using the ARM program 915-MHz radar wind profiler vs the relative humidity (%) from 1800 UTC the previous day.

and evolution of the LLJ under a wide variety of synoptic conditions at very high temporal and vertical spatial resolution. By combining the wind observations with numerical model output, it is possible to examine quantitatively the mechanisms that produce LLJs and the effects of various model parameterization schemes and model grid spacings on LLJ development over large regions.

These continuous wind observations also may be useful for evaluating the ability of a model to simulate the PBL. Since the remote sensing of temperature and relative humidity is not possible directly, the LLJ evolution can be used as a proxy for the proper PBL structure. Izumi and Barad (1963) show that mixing in the nocturnal PBL is closely tied to the LLJ structure when a jet is present, such that the PBL temperature profile is influenced greatly by mixing associated with the LLJ. This suggests that, if a model can simulate the correct timing, placement, and magnitude of the LLJ, then it is likely (although not certain) that the correct PBL structure also is simulated, including the correct horizontal variation of daytime surface sensible and latent heat fluxes that influence LLJ development (McCorcle 1988; Fast and McCorcle 1993). This is particularly important since the nocturnal evolution of the PBL, beginning with the transition from a deep to a shallow PBL in the late afternoon or early evening, is not as well understood as the daytime evolution of the convective PBL. Yet the signal of climate change appears most clearly in the increase of nighttime minimum temperatures over land (Karl et al. 1991). Therefore, one can argue that, until the nocturnal boundary-layer evolution can be simulated accurately, the utility of regional climate predictions is limited.

The LLJ has been documented to be a very important feature in the development of deep convection. While the grid spacing of present general circulation models (GCMs) inhibits the modeling of mesoscale processes, the spatial extent of many LLJs may allow them to be reproduced to some degree in GCMs. This model information on the LLJ could then be used to assist in the development of a parameterization scheme for deep convection during summertime when large-scale forcing typically is weaker and convection is associated more clearly with mesoscale processes. The ARM program low-level wind data also can assist in model sensitivity studies focused upon an examination of the grid spacing needed to capture the correct LLJ evolution, thereby highlighting regions where nested domains are needed to capture the important horizontal fluxes of temperature and moisture associated with the LLJ.

Lastly, the LLJ on a seasonal basis is important to the migration of birds and insects. Although large gaps exist in our understanding of these biological phenomena, it is clear that both birds and insects within North America, and likely worldwide, use LLJs to assist in their migrations (Drake 1985). This in turn affects farmers, veterinarians, health officials, and consumers who are concerned with, or influenced by, the influx of certain pests and disease agents. Seed dispersion also may be influenced by the LLJ. Therefore, in future long-term climate simulations where the vegetation is allowed to respond to the changing environmental conditions, the interactions between the LLJ and biological organisms may be important considerations in describing the vegetation patterns that would result and in attempting to assess the economic effects of climate change on agriculture, forestry, and human health.

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