Role of Topography in Forcing Low-Level Jets in the Central United States during the 1993 Flood-Altered Terrain Simulations

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ABSTRACT

Regional model sensitivity simulations in which the height of elevated terrain was reduced to explore simulated changes in features of the low-level jet (LLJ) are presented. Such an approach has not been reported, and it provides complementary insight to the previous LLJ studies. The simulations were carried out for a 45-day period during the 1993 summer flood in the central United States, when strong LLJs were frequent. The simulations illustrate directly the significance of topographical blocking, leeside cyclogenesis, and terrain thermal effects exerted by the Rocky Mountains in support of LLJ formation. In particular, it is shown that in the absence of topography the ridging from the Bermuda high extended considerably westward with weaker southerly flow over the High Plains, thus diminishing the potential for LLJ development. The slope-induced nocturnal horizontal thermal gradient was indicated to have a significant role in the formation of the LLJ.

1. Introduction

Motivated primarily by the linkage between the nocturnal low-level jet (LLJ) and warm-season precipitation over the Great Plains, the LLJ has been the focus of various observational studies (e.g., Pitchford and London 1962; Bonner 1968; Augustine and Caracena 1994, among others) and theoretical studies (e.g., Blackadar 1957; Holton 1967). Hypotheses for the formation of the LLJ tend to be grouped into several broad classes: (i) boundary layer mechanisms such as the diurnal variation of eddy viscosity or diurnally varying thermally direct circulations (Blackadar 1957; Holton 1967; McNider and Pielke 1981), (ii) interactions with upper-level dynamical processes such as jet streaks or lee cyclogenesis (e.g., Uccellini and Johnson 1979; Chen and Kpaeyeh 1993; Byerle and Paegle 2003), (iii) topographical blocking of the easterly flow associated with the Bermuda high (Wexler 1961; Zhong et al. 1996), and (iv) the development of the summer monsoon circulation over the western U.S. elevated terrain, which supports the LLJ in the southern High Plains (Tang and Reiter 1984).

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As implied by the aforementioned studies, the present consensus is that the most important forcing mechanisms for the LLJ are the response to diurnal boundary layer evolution over slopes and elevated terrain and the dynamical relationship to upper-level flow. It has to be emphasized that these two types of processes must not be viewed as competing or mutually exclusive explanations for the LLJ. Rather, the characteristics of the LLJ are established by their *combined* influences. General discussions of the combined forcing mechanisms have been given by Mitchell et al. (1995) and Zhong et al. (1996), among others.

Although both theory and observation have extensively documented the Rocky Mountain slope effects on the LLJ [see Strensrud (1996) for a review], direct quantification of the contribution of the elevated terrain to the LLJ forcing, through elimination or modification of the topography in regional model simulations, has not been reported to our knowledge. A hypothetical modification of the topography in the continental United States in a model simulation will affect the characteristics of the LLJ that are topographically forced. Adopting such an approach would provide complementary insight into the knowledge obtained in previous LLJ studies. In particular, it would outline a reference meteorological state absent of the topographical forced processes. Alteration of the topography in a model simulation is likely to modify the following LLJ forcing mechanisms: (i) the background southerly flow in the south-central United States that prevails during summer as a result of modification of the summer meteorological systems over the region, (ii) leeside cyclogenesis in the eastern slopes of the Rocky Mountains, and (iii) the along-slope induced nocturnal thermal gradient and the related mesoscale ageostrophic flow component that contributes to LLJ formation.

Modifying the topography in a model simulation will allow the combined effects of (i)–(iii) above to be examined. Such simulations are likely to provide also some insight into the relative role of the topography in contribution to individual processes that force the LLJ in the central and southern United States. In this note we selected the 1993 summer flood period in the central United States in which LLJs were frequent and intense. In sensitivity simulations we reduced the topography in the domain and examined the impact on the climatological patterns of the meteorological systems and the LLJ related characteristics and forcings. Evaluations are made in terms of time-averaged properties during the simulation period.

2. Meteorological conditions and methodology of evaluation

a. The meteorological conditions during the 1993 flood period

The flood of 1993 occurred at the maturity of an El Niño event with moderate SST anomalies over the equatorial Pacific. The above-normal convective heating in the intertropical convergence zone (ITCZ) shifted toward the equator (Trenberth and Guillemot 1996). This positive phase of the North Pacific teleconnection pattern persisted over the Pacific for 4 months prior to the flood. Then the ridge that prevailed over the western United States diminished and thus the zonal flow in the western Pacific was intensified. This zonal flow provided a channel for cyclones propagating directly to the central United States.

Synoptic patterns over the United States during the flood were characterized by a weaker Bermuda high, deeper leeside troughs east of the Rocky Mountains, and more frequent occurrence and stronger intensity of the LLJ (Mo et al. 1995; Arritt et al. 1997). The stronger upper-level jet streams and associated storm tracks pushed southward while the subtropical high retreated along the ITCZ.

b. Numerical modeling approach

Three options can be considered for modeling of modified topography effects on the 1993 LLJ in the central United States.

1) *Global model simulations.* Global predictability vanishes after a couple of weeks, so such an approach is not well suited to the study of the details of seasonal regional processes relevant to a given anomalous large-scale environment such as the one that prevailed during the summer of 1993.

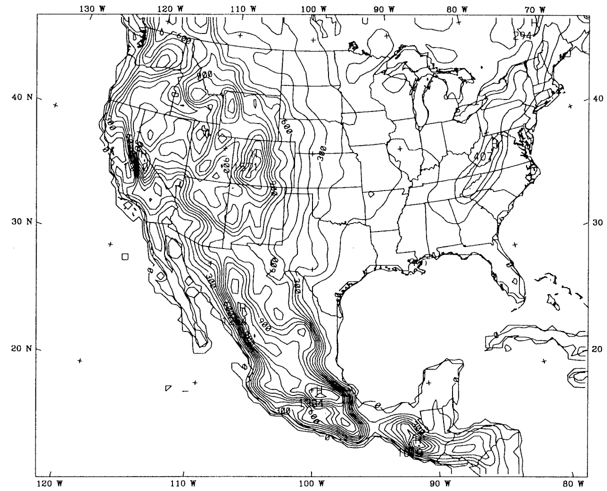

FIG. 1. Terrain heights (m) as resolved by a 52 km × 52 km model grid in the CTRL simulation. Contour interval is 100 m.

2) *Regional model in a short-term prediction mode (e.g., 24–48 h), forced by observed 1993 initial conditions.* A sequence of such simulations covering the period of interest would establish the control simulations. However, in the modified topography simulations, because of the short simulation period, this approach is likely to fail in maintaining an appropriate mass-flow adjustment. Furthermore, the nocturnal LLJ is associated with a large ageostrophic component that evolves on a time scale of about 1 day.
3) *Regional climate model forced continuously in time by observed meteorological lateral boundary conditions, so as to maintain the large-scale atmospheric conditions of 1993 in the modified topography simulations.* In sensitivity simulations we found that if lateral boundaries are not too close or too far from the center of the domain this approach would provide a reasonable compromise for the modeling evaluation. Specifically, we carried out simulations while extending the domain farther to the east, south, west, and north to evaluate the final placement of the lateral boundaries.

This last modeling approach, with appropriate specification of lateral boundaries, is therefore the most suitable considering the objective of this study and was adopted here.

The fifth-generation Pennsylvania State University–National Center for Atmospheric Research (Penn State–NCAR) Mesoscale Model (MM5) is used in this study. The Oregon State University (OSU) land surface model (Chen and Dudhia 2001) was adopted. The model domain was centered at 32°N, 100°W (Fig. 1), covering 101 × 85 grid points, with a horizontal grid spacing of 52 km. The model was configured in this study with 23 layers in the vertical centered at $\sigma = 0.995$, 0.985, 0.97, 0.945, 0.91, 0.87, 0.825, 0.775, 0.725, 0.675, 0.625, 0.575, 0.525, 0.475, 0.425, 0.375, 0.325, 0.275, 0.225, 0.175, 0.125, 0.075, and 0.025, with the top at 100 hPa. We use the Grell convection scheme (Grell et al. 1995), which is a simplified version of Arakawa–Schubert scheme with a single updraft/downdraft plume.

The lateral boundary conditions were imposed in an 11-gridpoint nudging zone adjacent to each lateral boundary where the weighting of the observations was reduced linearly away from the boundaries to the interior of the model domain. The initial and lateral boundary conditions were derived from the National Centers for Environmental Prediction (NCEP)–NCAR reanalysis (Kalnay et al. 1996) and updated every 6 h.

c. *Experimental setting*

We performed a 45-day continuous model integration from 1 June–15 July 1993, the peak period of LLJ activity. The full-terrain experiment (CTRL) uses the actual topography as resolved by the model 52-km grid (Fig. 1). The second experiment is flat terrain (FLAT), where the terrain heights are set to zero. Three intermediate altered-topography experiments were also carried out with 0.75, 0.50, or 0.25 of actual terrain heights.

In locations where topography was reduced, the initial conditions and lateral boundary conditions in the resulting void were extrapolated as follows: standard atmosphere lapse rate for the temperature and constant (same as at the surface) for wind and relative humidity. In the present simulations the lateral boundaries were located far away from the area of interest (the LLJ region east of the Rocky Mountains), minimizing spurious effects of altered terrain. Most parts of the lateral boundaries are at sea level or associated with very shallow topography; only small portions of the lateral boundaries are over high terrain (in Central America, and northwest Canada). Thus, overall the prescribing of lateral boundary conditions when topography is removed is likely to have only secondary spurious effects.

3. Results

a. *Averaged geopotential heights and flow*

In this subsection we present simulated lower-atmosphere patterns implying potential LLJ characteristics. It would be difficult to present the average flow during LLJ events for the simulated period, because of the spatial and temporal variation of the LLJ from one case to another. Therefore we present the simulated 45-day averaged geopotential and flow, which imply the potential for LLJ development.

The geopotential height is presented at 850 hPa (Fig. 2), which corresponds approximately to the elevation where LLJ peak intensity is commonly observed in the High Plains. Note that in CTRL, for locations where the topography is greater than ~1500 m, the geopotential height was extrapolated to 850 hPa. (We compared the geopotential field at 850 hPa in CTRL with both 700 hPa in CTRL and 850 hPa in the 0.5-terrain-height simulations at these locations. The comparison implied that the 850-hPa geopotential field presented in Fig. 2a was a realistic feature at these elevated locations.) In CTRL results for 0600 UTC (around local midnight), a ridging from the Bermuda high is simulated east of the Rocky Mountains, accompanied by a troughing that is centered mostly over the leeside upper slopes. This troughing is a split from a more extensive trough that dominated the western United States. The simulated results are in general agreement with the observed pattern reported in Arritt et al. (1997). The corresponding flow is presented at the level $\sigma = 0.91$ (~700 m above the surface, reflecting the typical height of LLJ maximum). Intense flow is most pronounced in western Texas, Oklahoma, and Kansas, with a maximum speed of ~16 m s^{-1} (Fig. 2a). The core of the strong flow extended eastward into Missouri. In FLAT, lack of topographical blocking resulted in a significant westward expansion of the ridge from the Bermuda high (Fig. 2b). Orientation of the geopotential contours turned more zonal, while the geopotential gradient over the central United States was weakened, suggesting reduced geostrophic flow. In FLAT cyclonic systems crossing the northern United States acquired stronger geopotential gradients and reduced southward expansion compared with those in CTRL (as evident from Figs. 2a,b). Flow in the central United States is much less intense in FLAT than shown previously in CTRL. The flow in northern Texas, Oklahoma, and Kansas weakened noticeably. However, in the absence of mountain blocking noticeable penetration and strengthening of the easterly flow from the Gulf of Mexico in eastern Mexico is evident.

Relatively strong daytime southerly flow is a prerequisite for the development of a pronounced nocturnal LLJ. At 1800 UTC (around local noon), the difference in the flow field east of the Rocky Mountains between CTRL (Fig. 2c) and FLAT (Fig. 2d) is noticeable. This result illustrates that the effects of elevated topography on the intensification of the daytime flow potentially are strongly conducive to nocturnal LLJ formation.

The 45-day average difference in wind velocity (CTRL minus FLAT) at $\sigma = 0.91$ at 0600 UTC showed a pronounced cyclonic flow perturbation over the western United States (Fig. 3). At its eastern side, this perturbation is forced by the combined effects of topographical flow blocking and leeside cyclogenesis. The flow perturbation is further supported by thermally induced pressure perturbation over the elevated terrain in the control simulation [see observational evaluation in Reiter and Tang (1984) and Tucker (1999)]. The southerly flow component difference was enhanced east of the Rocky Mountains in the area where the LLJ was simulated (see Fig. 5 later). The presented relative vorticity at 850 hPa (smoothed) corresponding to the ve-

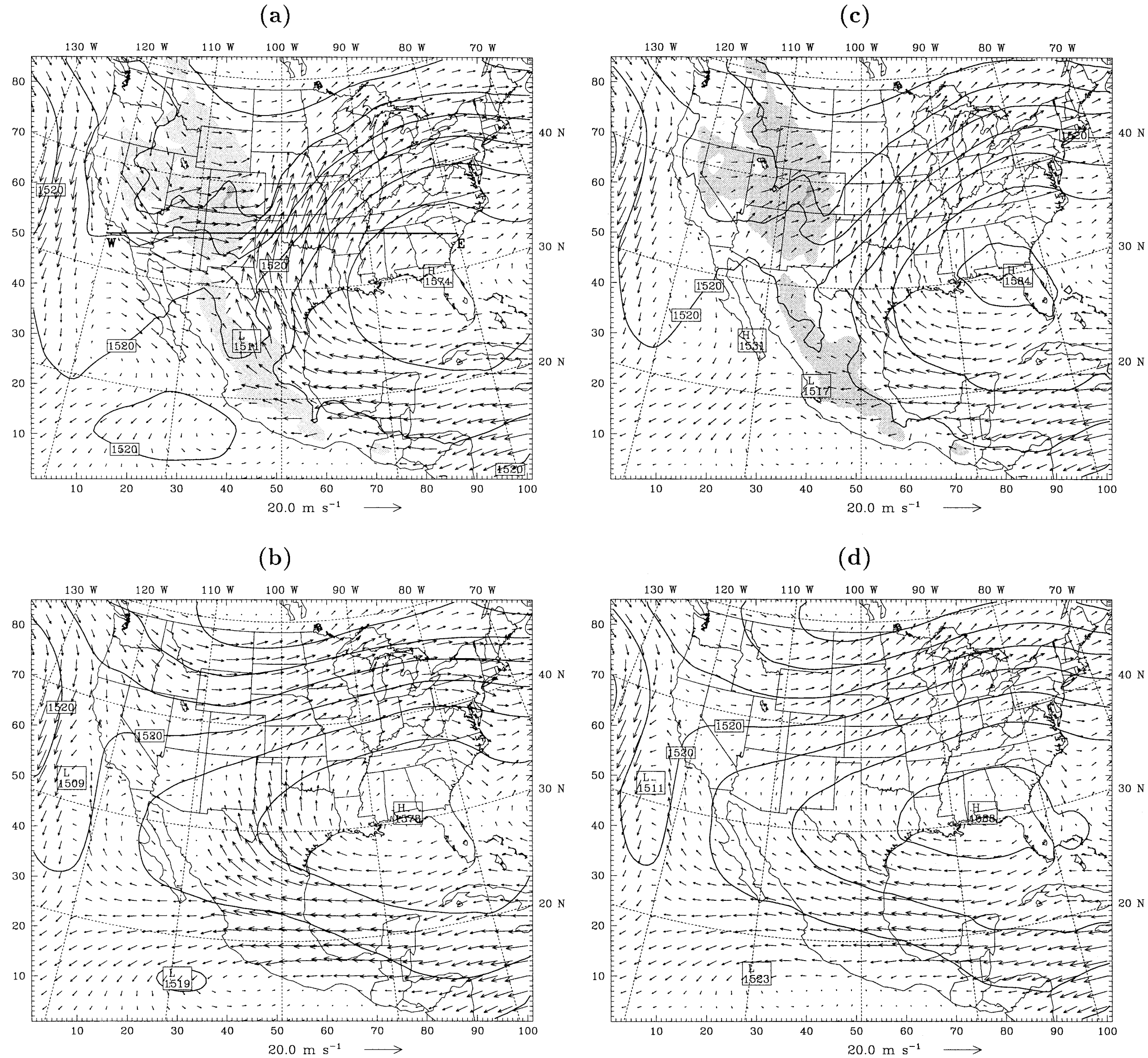

FIG. 2. Simulated 45-day averaged geopotential height (m) at 850 hPa and wind velocity at $\sigma = 0.91$ (~700 m AGL): (a) CTRL and (b) FLAT at 0600 UTC; (c) CTRL and (d) FLAT at 1800 UTC. Shading indicates terrain elevation higher than 1500 m.

locity difference between the CTRL and FLAT simulations provides another perspective on the characteristics of the cyclonic perturbation.

b. Cyclonic and anticyclonic lower-atmosphere averaged geopotential and flow

In an attempt to gain additional perspective of the LLJ forcing we established, based on observations, two dominant synoptic classes over the Rocky Mountains and their eastern side (following T.-C. Chen 2003, personal communication): (i) days in which cyclonic/trough systems (denoted C) prevailed in the lower atmosphere over the region (20 days) and (ii) days in which anticyclonic/ridge systems (denoted A) prevailed over the region (15 days). The remaining 10 days were not sufficiently well defined to be included in either class. Average simulated geopotential height at 850 hPa and flow at $\sigma = 0.91$ at 0600 UTC are presented in Fig. 4. For class C in CTRL (Fig. 4a), a trough was centered over a large segment of the eastern upper slopes of the Rocky Mountains with a strong south-southwesterly flow east of the trough line (the flow was stronger than the CTRL averaged for all simulated days; see Fig. 2a). When topography was removed (Fig. 4b), the troughing was weaker (as forcing for the leeside cyclogenesis is absent). The geopotential gradients and flow at the typical location of observed maximum LLJ in Oklahoma and northern Texas also were weakened. The corresponding simulated patterns for class A in the CTRL

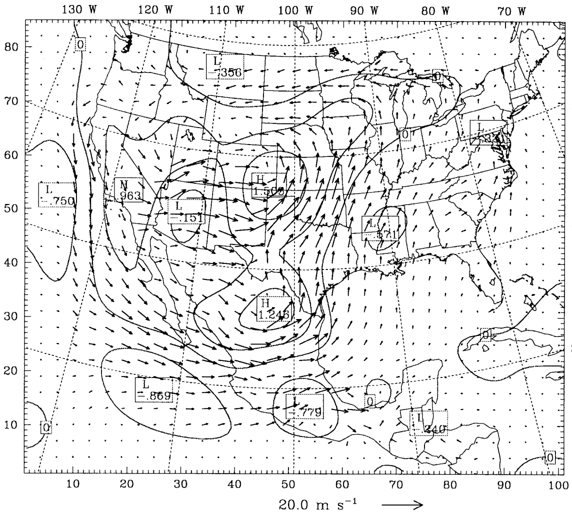

FIG. 3. The 0600 UTC 45-day averaged wind velocity difference (CTRL minus FLAT) field at $\sigma = 0.91$ ($\sim$700 m AGL). The contours depict the 850-hPa relative vorticity (smoothed) corresponding to the velocity difference between the CTRL and FLAT simulations (contour interval is 0.5×10^{-5} s^{-1})

simulation indicated ridging from the Bermuda high was well defined to the east of Rocky Mountains and troughing over the northwest United States (not shown). The flow at the LLJ location was southerly, but weakened compared to the C simulations. In FLAT, the absence of blocking effects of the Rocky Mountains caused the Bermuda high to expand considerably westward and northward and the flow weakened significantly over the observed LLJ maximum location.

c. *Low-level jet frequency*

The LLJ frequency during the simulated period was computed according to Bonner's (1968) classification, which is based on the vertical wind shear and maximum wind speed. Figure 5 presents the frequency of LLJ occurrence for Bonner's classes 2 and 3 (excluding the weaker LLJ class 1), corresponding to 16 and 20 m s^{-1} peak wind speed, respectively. Bonner's class 1 was not considered since the model-simulated wind speeds were somewhat overpredicted. The region of high LLJ frequency in CTRL at 0600 UTC east of the Rocky Mountains occurred in a wide swath extending from northern Mexico to Iowa (Fig. 5a). The LLJ frequency had a primary peak at the United States–Mexico border in southern Texas (60%) and a secondary peak in Oklahoma (45%). The primary LLJ frequency maximum occurred near the Big Bend region of Texas at 30°N. At this latitude, the plateau thermal circulation effects on the LLJ development should peak based on the theoretical study of Sang and Reiter (1982). The FLAT simulation has much lower LLJ frequency in this region (Fig. 5b). Thus various topographic forcings in CTRL appear to explain most of the occurrence of LLJs.

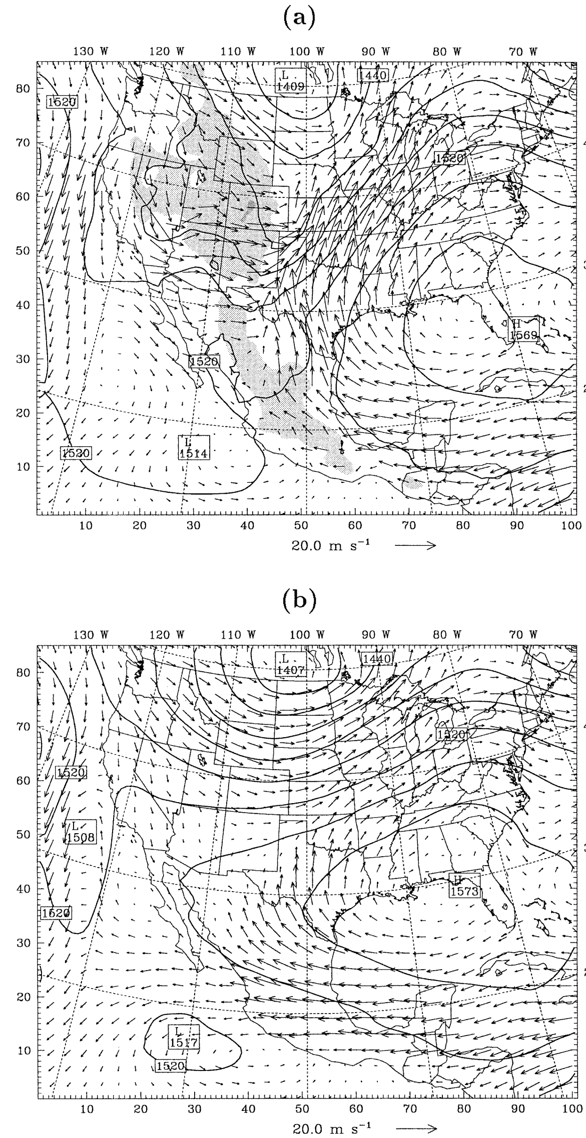

FIG. 4. Simulated 0600 UTC 45-day average geopotential height (m) at 850 hPa and wind velocity at $\sigma = 0.91$ ($\sim$700 m AGL) for cyclonic cases (class C): (a) CTRL and (b) FLAT. Shading indicates terrain elevation higher than 1500 m.

The local LLJ peak over Montana in FLAT is explained by the intensification of geopotential gradients of eastward moving cyclonic systems. The resulting increased baroclinicity likely induced the LLJ in this location. However, it should be noted that the proximity to the domain's lateral boundary may also have some influence.

d. *Vertical structure of the wind and potential temperature*

Evaluating the topographical impact on the LLJ, we present also east–west vertical cross sections (located as indicated in Fig. 2a) of the 45-day averaged potential

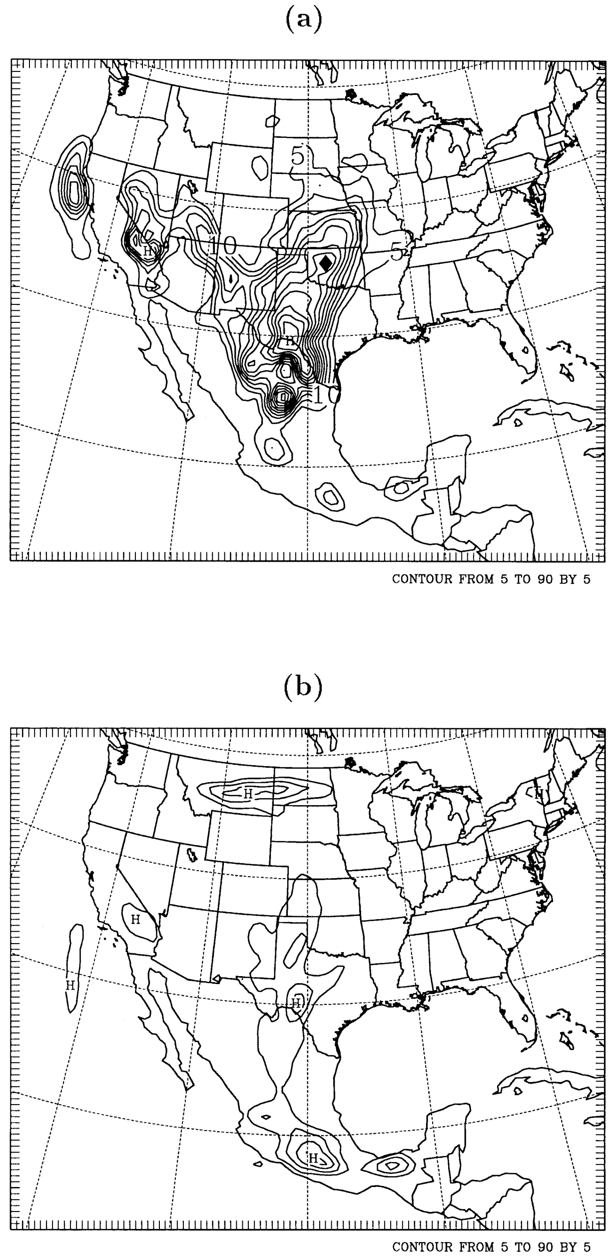

FIG. 5. Low-level jet frequency (%) for Bonner's classes 2 and 3 at 0600 UTC for the simulated 45-day period: (a) CTRL and (b) FLAT. The symbol ♦ indicates local peak frequency.

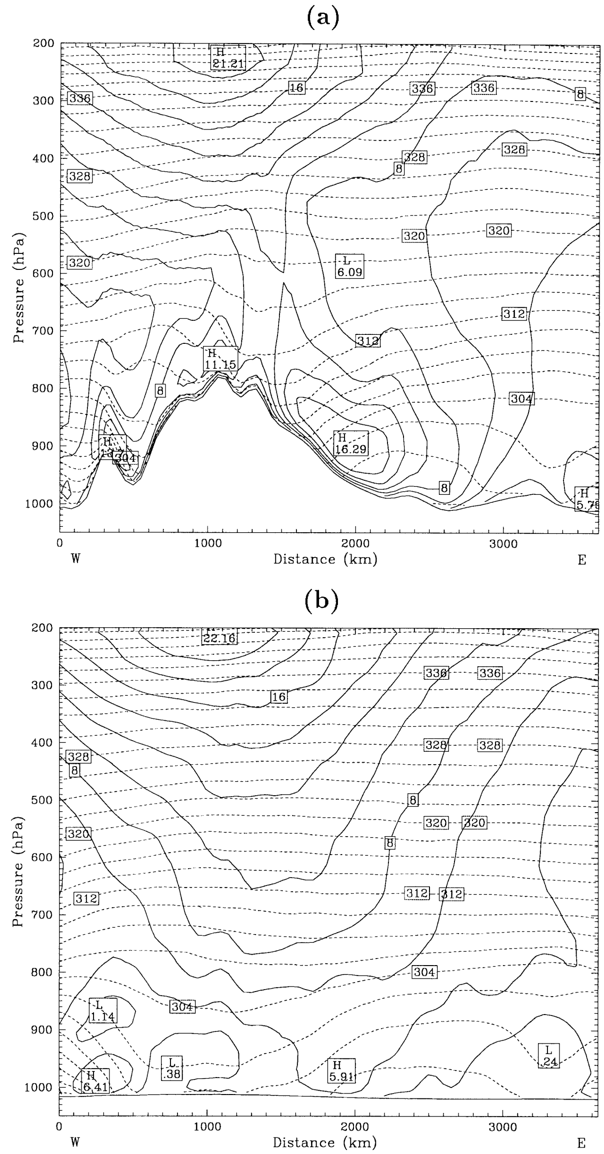

FIG. 6. Vertical cross section of the 45-day averaged wind speed (solid contours in m s^{-1}) and potential temperature (dashed contours in K) along the dark line (W–E) marked in Fig. 2a at 0600 UTC: (a) CTRL and (b) FLAT.

temperature and wind speed for the CTRL and FLAT simulations at 0600 UTC (Figs. 6a,b). The vertical cross section passes through the location of the peak LLJ frequency in Oklahoma simulated in CTRL. The CTRL shows a well-defined LLJ vertical structure (i.e., local wind maximum) over the slopes with wind speed maximum of ~16 m s^{-1}. However, in FLAT there was no evidence of LLJ vertical structure, and the peak wind only reached ~6 m s^{-1}. Comparing Figs. 6a and 6b, it is evident that to the east of the Rocky Mountains the LLJ structure is associated strongly with the along-slope nocturnal thermal gradient, in agreement with McNider and Pielke (1981). However, in the 2D simulations by McNider and Pielke (1981) the background flow was southerly, so that the slope thermal gradient was forced solely by nocturnal diabatic cooling; in the present simulation the slope thermal gradient was additionally supported by westerly advection of warm air from the elevated terrain. In locations where horizontal potential temperature gradients are mild or absent there is no vertical structure resembling the LLJ (see Figs. 6a,b). Thus it appears that the slope-induced thermal gradient contributes significantly to the formation of a pronounced LLJ, in agreement with previous studies. It is

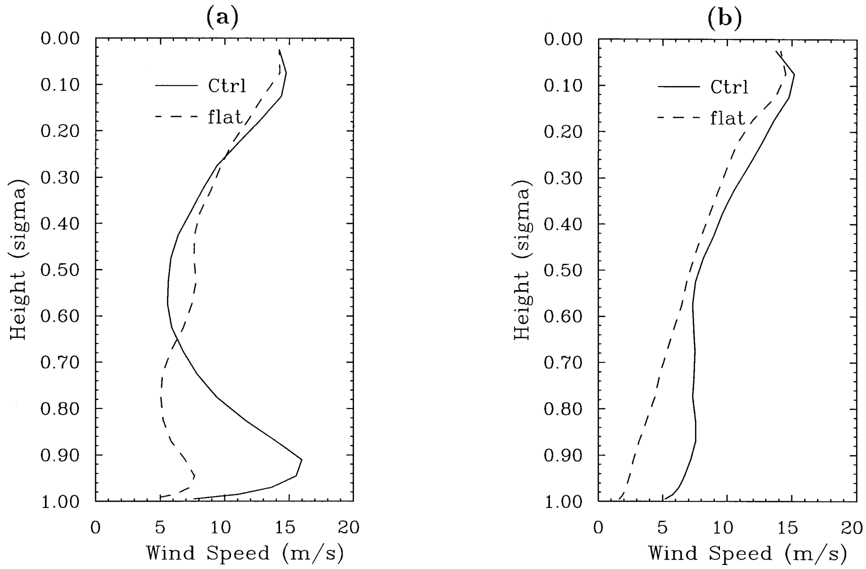

FIG. 7. Vertical soundings of 45-day averaged wind speed for CTRL and FLAT at the point marked by the symbol ♦ in Fig. 5a: (a) 0600 and (b) 1800 UTC.

worth pointing out that a mild horizontal thermal gradient was simulated in FLAT at a location similar to the LLJ peak in CTRL (Fig. 6b). It results from eastward advection of warm air generated over the southwestern United States. The decay of nocturnal boundary layer turbulence, though conducive to intensification of the flow, is insufficient by itself to establish the LLJ, as evident from examining the FLAT results. Also it may be suggested that in CTRL when an easterly flow component exists on the slopes following sunset, development of a nocturnal temperature inversion decreases the Froude number and, thereby enhances topographical blocking of the flow.

In an additional comparison illustrating the difference between CTRL and FLAT, the 45-day averaged vertical profile of wind speed is presented at the location where peak LLJ frequency was simulated in Oklahoma (marked by the symbol ♦ in Fig. 5a). In CTRL at 0600 UTC (Fig. 7a) the profile is typical of observed nocturnal LLJs (e.g., Whiteman et al. 1997). In comparison the FLAT simulation had only a minor wind speed peak. The daytime (1800 UTC) difference in wind speed between CTRL and FLAT in the lower atmosphere is attributed mostly to topographically induced dynamical effects (Fig. 7b).

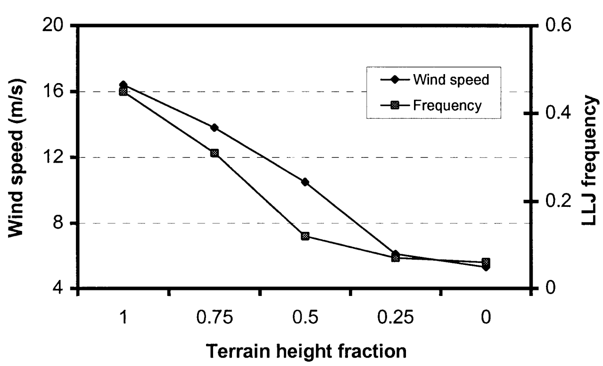

FIG. 8. The peak wind speed of the 45-day average (east of the Rocky Mountains in the lower atmosphere) along the vertical cross section shown in Fig. 6 and the maximum LLJ frequency (for Bonner's classes 2 and 3) at the point marked by the symbol ♦ in Fig. 5a for selected simulations where 0.75, 0.5, or 0.25 of the actual terrain height was considered.

e. *Sensitivity of LLJ features to intermediate modified topography*

Figure 8 presents the peak wind speed of the 45-day average (east of the Rocky Mountains within the lower atmosphere) along the vertical cross section shown in Fig. 6 and the LLJ frequency (Bonner's class 2 and 3) at the point marked by the symbol ♦ in Fig. 5a for simulations where 0.25, 0.50, or 0.75 of the actual terrain height was specified. Both peak wind speed and LLJ frequency decrease with decreasing terrain height, but at faster rates when terrain is high (higher than 0.5). Once the terrain in the simulation was as low as 0.25 times the original height, further reduction in terrain height only slightly affected the LLJ peak wind speed and frequency. It is worth pointing out that coarse-resolution general circulation models underrepresent actual terrain height. The ratio of model-resolved versus actual terrain heights in such models typically is near or less than 0.75, suggesting that LLJ strength and occurrence in the central United States is likely to be underestimated. Similar underestimations of LLJ strength and frequency may also occur in atmospheric reanalyses that make use of coarse-resolution models (see Anderson and Arritt 2001).

4. Conclusions

In this note we presented sensitivity simulations that explored simulated changes in LLJ features in the central United States when terrain elevation was reduced. Such an approach has not been reported, and it provides complementary insight to prior studies of LLJ. The integration period was a 45-day window selected from the 1993 flood when strong LLJs were frequent. Evaluations were made in terms of time-averaged properties during the simulation period.

Potential vorticity theory explains the formation of leeside cyclones for westerly flow and anticyclonic systems for easterly flow interacting with the Rocky Mountains. Relatively strong daytime southerly flow generated by these and other mechanisms is a prerequisite for development of a pronounced nocturnal LLJ. Quantification of these processes in the real world, considering the implications to the development of LLJ, requires sensitivity simulations such as those carried out in the present study.

Simulations with different terrain heights enabled quantification of the profound impacts that the Rocky Mountains have on LLJ strength, frequency, and location through flow blocking, leeside cyclogenesis, along-slope nocturnal thermal gradient induced by diabatic cooling, and thermal effects of the elevated terrain. In particular, the results illustrated the impacts of topographical blocking on constraining the westward expansion of the Bermuda high. Also illustrated quantitatively is the intensification of the southerly flow east of the Rocky Mountains because of topographical blocking and leeside cyclogenesis. In sensitivity simulations to terrain-height modifications, it was found that the topography potential effects on LLJ intensity and frequency decreased for low terrain.

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