Thermally Forced Surface Flow and Convergence Patterns over Northeast Colorado

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ABSTRACT
Numerical model simulations have been performed with the Colorado State University mesoscale model to determine the regions of most likely occurrence of first cumulonimbus activity. It is shown that during the day, convergence along the eastern slopes of the Continental Divide and along the Cheyenne Ridge and Palmer Lake Divide coincides with the regions of most moist and unstable air. During the evening the flow reverses and the main convergence is located in the Platte River Valley. The interaction of these features with the nocturnal jet and Denver convergence/vorticity zone is also discussed.

1. Introduction
Many investigators have studied the repetitive, diurnal convective cycle characteristic of the summertime climate of many mountainous regions. In these regions the topography induces a circulation in which the initial convective activity grows to thunderstorm proportions along the mountain ranges during the morning before moving out onto the plains during the afternoon. This cycle is particularly noticeable along the eastern slopes and plains of Colorado and has been the basis for a considerable amount of research. These studies range from the earlier radar climatologies of Wetzel (1973) and Henz (1974) to the satellite climatology of Klitch et al. (1985) and the surface-flow climatology of Toth and Johnson (1985).

Wetzel's calculations of daily moisture budgets showed that local convergence, induced by the mountain–plains circulation, was important in the production of summer convection. His radar climatology showed the Cheyenne Ridge and Palmer Lake Divide (Fig. 1) to be preferred genesis and propagation regions for this convection. Henz carried this study further and identified ten "hot spots" (regions of repeated convective system genesis) along the Front Range of the Colorado Rockies. He found that these genesis sites produce 40% of all thunderstorm systems and 73% of severe weather occurrences.

Klitch et al. (1985) briefly discuss some of the drawbacks of radar climatologies. These drawbacks are related to the calibration of the radar such that only precipitation-size particles, and not the initial congestous

formation, are detected. In mountainous regions convection may go undetected as ground clutter interferes with radar beams. Klitch et al. used interactive computer techniques to composite satellite imagery and hence show the areal/temporal variation of the diurnal convective cycle over the Rockies and adjacent plains. They reestablished the general features mentioned above, but also showed that convection is relatively infrequent along the Cheyenne Ridge during the general convective cycle. However, on severe weather days there is strong development along the Cheyenne Ridge and a more rapid eastward movement of convective activity.

Toth and Johnson (1985) used surface wind data from the PROFS (Program for Regional Observing and Forecasting Services) mesonet to investigate the diurnal wind-flow pattern over the South Platte River Basin in northeast Colorado. They showed that during the summer months, local confluence is found at midday along the Cheyenne Ridge and Palmer Lake Divide. By superimposing their results and those of Wetzel (1973), they found that there is a strong suggestion that the preferred regions of development coincide with zones of maximum confluence 1 to 2 hours earlier. In addition, they suggested that the late afternoon transition to downslope flow is associated with propagation of thunderstorms from the mountainous regions eastward to the plains.

The summer climate in Colorado is generally characterized by weak flow aloft. Under these conditions synoptic-scale forcing weakens, and mesoscale factors begin to play a more important role in the diurnal
weather cycle. Such a situation is particularly suitable for a climatological study using a mesoscale numerical model.

The study discussed below used the CSU Mesoscale Model (Pielke, 1974; Mahrer and Pielke, 1977; McNider and Pielke, 1981) to predict regions of convergence, and hence preferred regions for thunderstorm genesis, under conditions of light synoptic forcing. The main advantage of this approach over the earlier studies is related to the fact that the character of the surface flow under different synoptic conditions is represented. Also, it is possible to investigate the effects that the upper-level westerlies and various atmospheric stabilities have on the flow. In addition, aspects of the mesoscale climate such as the nocturnal jet (Hahn, 1981) and the Denver surface convergence–vorticity zone (Szoke et al., 1984) may be investigated and their effects on the surface flow determined. A number of these facets will be discussed below.

2. The model

The model physics are unchanged from those reported earlier by Pielke and his colleagues. The parameterization of the surface energy budget is detailed in Mahrer and Pielke (1977), and the PBL parameterization is discussed in Mahrer and Pielke (1977) and McNider and Pielke (1981).

The model simulations commenced at sunrise (0445 LST) and were preceded by 6 hours of dynamic initialization. During the 6 hour initialization period, the model was run without thermal forcing in order to dynamically adjust the mass and velocity fields.

The model atmosphere had 26 vertical levels with greatest resolution near the surface. An absorbing layer (Klemp and Lilly, 1978; Mahrer and Pielke, 1978) was included in the upper levels to control the reflection of vertically propagating wave energy. The bottom of the absorbing layer was at a height of 5.5 km and the top at 13.5 km. There were eight levels in the absorbing layer. In all simulations moist processes in the atmosphere were neglected; however, soil moisture effects were included. The initial conditions for these simulations are presented in Table 1. The orography (Fig. 1) used in this study was 10 minute averages calculated from the Defense Mapping Agency’s 30 second gridpoint data. The horizontal grid interval was approximately 15 km on a 32 × 29 domain. At the lateral boundaries the grid spacing was expanded. At the boundary a zero gradient condition on the orography was applied.

The results from the model are presented in the form of horizontal cross sections, in which the vectors are scaled such that the distance between tick marks is equivalent to a wind speed of 10 m s⁻¹.

In the case study, observation times were converted from Daylight Saving to Standard time which is approximately equal to local solar time for the period under study. The times shown on the figures and quoted in the text are Local Solar Time (LST), which equals GMT minus 7 hours.

3. Case study: 31 July 1982

A case study is presented below to examine the skill of the model over the complex terrain of the Front

<table>
<thead>
<tr>
<th>Table 1. Input parameters used to initialize the model.</th>
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<tr>
<td>Parameter</td>
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</tr>
<tr>
<td>Horizontal grid interval</td>
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<td>Horizontal grid size (x) × (y)</td>
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<td>Vertical levels</td>
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<td>Mean latitude</td>
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<td>Land surface:</td>
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<td>Albedo</td>
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<td>Roughness length</td>
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<td>Initial PBL height</td>
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<td>Initial surface pressure</td>
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<td>Free atmospheric lapse rate of potential temperature</td>
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<td>Soil characteristics:</td>
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<td>Density</td>
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<td>Model levels for u, v and w (the Exner function):</td>
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<tr>
<td>10, 20, 30, 50, 90, 150, 250, 500, 750, 1000, 1350, 1750, 2250, 2750, 3250, 3750, 4250, 4750, 5500, 6500, 7500, 8500, 9500, 10500, 11500, 13500</td>
</tr>
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<td>θ and q are staggered halfway between these levels.</td>
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Range of northern Colorado. A realistic simulation will suggest that the diagnostic studies presented in section 4 are representative of the flow occurring under conditions of light synoptic forcing.

The 12Z (0500 LST) rawinsonde sounding from Denver, used to initialize the model, is shown in Fig. 2a. Other variables used to initialize the model were representative of the average July conditions for this region and are given in Table 1. The surface and 850 mb synoptic analyses are shown in Figs. 2b and 2c. This day was selected because the synoptic forcing was extremely light and little convective activity developed during the day. The model results are compared with observations from the PROFS mesonet to illustrate and

![Diagram of Denver sounding at 12Z](image)

**Fig. 2.** (a) Denver sounding at 12Z used to initialize the model for the 31 July 1982 simulation. Short barbs: 1 m s$^{-1}$; long barbs: 5 m s$^{-1}$. (b) Synoptic scale surface analysis for 12Z 31 July 1982. Frontal positions, pressure centers and isobars at 2 mb intervals (1016 = 16) are shown. Colorado is indicated on the map. (c) 850 mb analysis for 12Z 31 July 1982. Solid lines are height contours (drawn for every 30 m, 153 = 1530) and dashed lines are temperatures at 5°C intervals. Colorado is indicated on the map.
Table 2. List of PROFS and NWS stations and their abbreviations as shown on Figs. 3-5.

<table>
<thead>
<tr>
<th>Station</th>
<th>Abbreviation</th>
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<td>Akron</td>
<td>AKO</td>
<td>Idaho Springs</td>
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<td>Arvada</td>
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<td>Keenesburg</td>
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<td>RB3</td>
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<td>COS</td>
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<td>Elbert</td>
<td>ELB</td>
<td>Nunn</td>
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<td>Estes Park</td>
<td>EPK</td>
<td>Platteville</td>
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<td>Fort Collins</td>
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<td>Rollinsville</td>
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<tr>
<td>Fort Morgan</td>
<td>FTM</td>
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<td>Greeley</td>
<td>GLY</td>
<td>Ward</td>
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compare the evolution of the surface flow during the day. A list of PROFS and National Weather Service stations and their abbreviations, as shown on Figs. 3-5, is given in Table 2.

At 0600 LST (Fig. 3) the agreement between the model results and observations is good. Both show downslope flow along the eastern slopes of the Continental Divide and from the Cheyenne Ridge and the Palmer Lake Divide. In the simulations, the onset of surface heating has caused the winds along the Front Range to become more slope-parallel as they begin to rotate to upslope flow. The model has been unable to reproduce the weak cyclonic vortex, evident in the observations, in the vicinity of Denver. This vortex was not evident in the observations at 0500 LST and appears to have formed in response to the surface heating, as the low-level flow turns from downslope to upslope flow. In both cases the flow in this region is very weak (1 m s\(^{-1}\)). The model has also been unable to reproduce the southeasterly flow at Laramie, possibly because the model is unable to adequately resolve the ridge on which Laramie is located.

Overall, however, the model has adequately simulated the surface wind field for 0600 LST. The main features at this time are downslope flow from the Continental Divide, the Cheyenne Ridge and the Palmer Lake Divide, and weak, variable flow in the Platte River Valley. The flow on the northern slopes of the east-west ridges is stronger than that on the southern, while on the plains the wind is approximately 4 m s\(^{-1}\) from the southwest.

Continued heating of the sloping terrain has produced a differential temperature field so that at 1200 LST (Fig. 4) the wind is strongly influenced by terrain effects. This is evident as upslope flow along the eastern slope of the Continental Divide and along the Cheyenne Ridge and the Palmer Lake Divide. It should be

Fig. 3. Model results and observations for 0600 LST 31 July 1982. Short barbs: 1 m s\(^{-1}\); long barbs: 5 m s\(^{-1}\).
noted, however, that the surface flow is not independent of the synoptic flow, and this is illustrated by the winds observed over the Cheyenne Ridge. The observations show that at 1200 LST the winds on the Cheyenne Ridge are from the northwest and that the Cheyenne Ridge convergence zone occurs farther south than predicted by the model. Figures 2b and 2c indicate that the northern portion of the model domain (in Wyoming to the north) is under westerly or northwesterly geostrophic flow. Indeed, a supplementary simulation using a northwesterly geostrophic flow predicted the Cheyenne Ridge convergence zone to occur close to its observed location. To the south, where the geostrophic wind was from the southwest, the agreement between the observations and model results was much better with an imposed southwesterly geostrophic flow.

The observed and modeled surface flows for 2100 LST are shown in Fig. 5. By this time the surface heating has ceased and downslope flow has commenced along the eastern slope of the Continental Divide. To the east of this, both the observations and model results indicate a flow from the south. This southerly flow is due to both downslope flow and the nocturnal jet. A convergence zone has now developed in the Platte River valley where the downslope flow from the Continental Divide converges with the southerly flow of the nocturnal jet.

This case study has successfully simulated many of the features of the low-level wind field evident on this day. Consequently, it can be assumed that the diagnostic studies presented in section 4 are representative of the flow that occurs when terrain is the dominant forcing mechanism.

4. Diagnostic studies

In the following studies, the vertical sounding used to initialize the model was the average 12Z (0900 LST) July sounding for Denver. Other variables used to initialize the model were representative of the average July conditions for this region.

a. Heating effects

In the first experiment to be discussed, the effect of the diurnal heating cycle only will be outlined. In this experiment the effects of the synoptic flow (0.5 m s⁻¹ from the south) are negligible. Three-hourly horizontal cross sections of the 10 m winds, the vertical velocity at 1350 m and the lifted index are shown in Fig. 6.

The effects of heating became obvious quite early in the day, with upslope flow becoming established along the eastern slopes of the Continental Divide and the southern slopes of the Cheyenne Ridge and the Palmer Lake Divide by 0900 LST. Over the remainder of the domain the low-level winds were light and variable. At
this stage the main regions of ascent were over the eastern slopes of the Continental Divide and over Pikes Peak. Subsidence was evident in the Platte River Valley.

By 1200 LST the upslope flow had become well established, with the low-level flow over the plains now having a strong easterly component. This flow is associated with weak ascent over the east–west ridges and stronger ascent over the eastern slopes of the Continental Divide and over Pikes Peak. It is interesting to note that the upslope flow on the northern slopes of the east–west ridges had not become well established, presumably because these slopes had received less heating than the southern slopes. It is evident that the areas of strongest convergence coincide with the regions of the most moist and unstable air.

As the day progressed the air became more unstable, with the lowest lifted index values of −4.5 occurring west of Pikes Peak at approximately 1400 LST. The low-level wind field showed little change until late afternoon. During the afternoon, the fairly strong southerly upshepe flow on the Cheyenne Ridge and the Palmer Lake Divide propagated to the lee side of these ridges. This resulted in the northward movement of the tongues of unstable air which were coincident with the ridge lines earlier. By 1800 LST the region of strongest instability which was present over the eastern slopes of the Continental Divide had disappeared, and the most unstable air lay along the Palmer Lake Divide.

At 2100 LST the surface heat flux had reversed and downslope flow had commenced along the Front Range. Along the east–west ridges there was little reversal of the flow, although the upshepe flow on the southern slopes had decelerated. Along the northern slope of the Palmer Lake Divide there was an acceleration and veering of the southerly flow. These changes produced a convergence zone along the Front Range and a weak cyclonic vortex. The Palmer Lake Divide convergence zone propagated to the north due to the acceleration of the southerly downslope flow. At this stage it was oriented SW–NE in the Platte River Valley. The most unstable air was located in the Platte River Valley, northeast of Denver and coincident with the Palmer Lake convergence zone.

Wilczak (personal communication, 1985) has used a mixed-layer model to simulate the low-level wind field over this region, both as a function of stability and synoptic-scale wind speed and direction. He, too, found that a convergence zone forms in this region; however, he required a very stable atmosphere and south to southeasterly winds. In addition, he has shown that this convergence zone does not form when the orography of the Palmer Lake Divide is removed. Szoke et al. (1984) have observed a convergence–vorticity zone in the vicinity of Denver. This zone was present early in the day, and they suggest either terrain-blocking by the Palmer Lake Divide or synoptic-scale cyclogen-
Fig. 6. Horizontal cross sections of the (a) 10 m winds, (b) vertical velocity at 1350 m and (c) lifted index for 0900, 1200, 1800 and 2100 LST for the simulation with negligible synoptic forcing. Regions of $w > 0$ and LI $< -2$ are stippled.
FIG. 7. Ten-meter winds at 0600, 0900, 1200, 1500, 1800 and 2100 LST for low-level flows from the north, south, east and west.
Fig. 7. (Continued)
esis as possible mechanisms for its formation. The modeling results presented above suggest that, in addition, convergence of downslope flows from the surrounding terrain can produce a convergence-vorticity zone in this region which may exist until the following morning.

In an analysis of Boulder Atmospheric Observatory (BAO) Tower data, Hahn (1981) found that a southerly nocturnal jet was often present on synoptically undisturbed days. A similar jet (no figure presented) was evident in the model winds at a height of 250 m. It is interesting to note, however, that the modeled jet turned from a southerly jet over the Palmer Lake Divide to a southeasterly jet in the Platte River Valley and was not present west of the Front Range. As the evening progressed the Front Range downslope flow intensified and deepened. This had the effect of eliminating the nocturnal jet to the east of the Front Range and replacing it with downslope flow. This mechanism may account for the midnight decay of the nocturnal jet, as observed by Hahn.

b. Synoptic effects

The effects of the synoptic situation were investigated by performing a number of simulations in which geostrophic winds from different directions were imposed. In these simulations the effects of vertical shear were not included. By imposing various geostrophic winds, it is possible to note how the surface flow and the positions of the convergence zones change.

Simulations were performed for situations in which the geostrophic winds across the domain were light (2.5 m s$^{-1}$) and from (a) the north, (b) the south, (c) the east, (d) the southeast and (e) the west. These simulations produced many of the features mentioned above. The southeasterly case resulted in a surface flow very similar to the southerly case. For this reason it will not be included in the following discussion. It was found, however, that the resultant surface flow for an imposed southerly geostrophic wind was different from that produced in the other cases. These differences did not become obvious until approximately 1500 LST, by which time significant heating had allowed the thermal flows to develop. The evolution of these flows, for the various synoptic conditions, is presented in Figs. 7, 8 and 9. These figures present the winds at 10 m, the vertical velocity at 1350 m and the lifted index, respectively.

At 0600 LST, heating of the eastern face of the Continental Divide had caused the flow along the Front Range to begin turning to upslope flow. In all cases there was a southerly downslope flow on the northern slope of the Palmer Lake Divide. With imposed southeasterly to westerly geostrophic winds, blocking of the flow by the higher terrain resulted in the formation of a cyclonic vortex in the vicinity of Denver.

This vortex had dissipated by 0900 LST with the continued heating of the surrounding slopes. By this stage the upslopes along the Front Range were well established, and upslope flow was beginning to develop along the east–west ridges.

At 1500 LST the differences between the surface flows for the various synoptic situations were becoming apparent. In all cases, except that of a southerly geostrophic wind, there was a well-marked confluent zone along the ridge lines of the Cheyenne Ridge and the Palmer Lake Divide and a diffusional zone in the Platte River Valley. In each of these cases the most unstable air, as measured by the lifted index, lay along the Palmer Lake Divide. Unlike the case with negligible synoptic forcing, however, there was no tendency for the instability to be concentrated over the eastern slopes of the Continental Divide.

With a southerly geostrophic wind, the upslopes on the northern slopes of the east–west ridges were unable to form, and no convergence zone formed. The upslope flow did not form in this case as it was overwhelmed by the synoptic flow. In addition, the southerly upslope flow on the southern slopes was stronger. This mechanism does not apply with a northerly geostrophic wind, as stronger heating of the southern slopes of these ridges produces a stronger north–south temperature gradient on this slope than the northern slope. Consequently, the upslope flow that develops is strong enough to overcome the northerly geostrophic flow.

By 1800 LST, further heating of the southern faces of the Cheyenne Ridge and Palmer Lake Divide had resulted in the acceleration of the southerly upslope flow on these ridges. This situation did not apply to the case with a northerly geostrophic wind. In this case there had been little change since 1500 LST, and presumably the synoptic flow had retarded further development of the southerly upslopes. The most notable change in the lifted index values was the stabilization of the atmosphere over the Palmer Lake Divide in the cases of the westerly and northerly geostrophic flows. With a westerly geostrophic flow, this stabilization coincided with a destabilization of the atmosphere along the Cheyenne Ridge. In the satellite climatology of Klitch et al. (1985), convection generally occurred along the Cheyenne Ridge on severe weather days. In their results it also appeared that this convection lagged that along the Palmer Lake Divide by approximately two hours. They suggested that severe weather may be related to the strengthening of the westerly flow aloft, but from these results it seems possible that low-level westerly flow may be sufficient to enhance convection along the Cheyenne Ridge.

As the evening progressed, downslope flow commenced, first along the Front Range and later along the east–west ridges. This resulted in the formation of a convergence zone in the Platte River Valley. With a southerly geostrophic flow, however, the downslope flow developed later. In all cases a southerly nocturnal
Fig. 8. The vertical velocity at 1350 m at 1200, 1500 and 1800 LST, for low-level flows from the north, south, east and west. Regions of $w > 0$ are stippled.
FIG. 9. The lifted index at 1200, 1500 and 1800 LST for low-level flows from the north, south, east and west. Regions of LI < -2 are stippled.
jet had developed by 2100 LST, although this was not as strong in the northerly case. The flow at 250 m at 2300 LST is shown in Fig. 10. By this time the downslope flow was well developed and a calm region had evolved in the Platte River Valley under southerly and westerly geostrophic flow. Turning of the flow in this region may give the appearance of a vortex in surface observations. As with the case of no synoptic forcing, the nocturnal jet veered during the evening, and a jet-free region developed along the Front Range. This jet-free region had not developed by 2300 LST in the southerly case.

c. Effect of the westerlies aloft

The westerlies aloft have been considered important to the surface flow as reported by Banta (1984) and Toth and Johnson (1985). Banta found that a leeside convergence zone forms when upslope winds meet with convectively-mixed winds blowing down from the direction of the ridges. He considered this zone to be important in cloud formation and its continued development. Toth and Johnson (1985) found that the only effect of stronger westerly winds aloft was to cause the transition from upslope to downslope flow to occur 1–2 hours earlier.

In the present study, constant shear was applied above the planetary boundary layer, such that at 7150 m (ASL) the winds were 7.0 m s⁻¹ from 263°. This value is slightly less than the July average of approximately 9.0 m s⁻¹ at this elevation. Simulations were performed for cases in which the low-level flows were from either the east or the west. From these results it was found that the effect of the upper-level westerlies depends upon the direction of the low-level flow. The effect of the upper-level westerlies was most noticeable when the low-level flow was from the west also. The effect of the westerlies aloft with a low-level westerly flow is illustrated in Fig. 11b.

In this case the upslope flow that developed was slightly weaker. By 1800 LST the upslope flow along the Front Range had begun to weaken and at 1900 LST downslope flow had commenced along the Front Range. Upslope flow was still present to the east and west of the newly developed downslope flow at this time. This flow reversal occurred approximately 1 hour earlier than in the case without upper-level westerlies. Also, these downslope flows were stronger than those produced without westerlies aloft. The confluence of the downslope flow and remaining upslope flow resulted in the formation of convergence zones in the lee of the Continental Divide and Pikes Peak and at the junction of the Cheyenne Ridge and the Continental Divide. By 2000 LST the main convergence zone had propagated farther east and was located on the plains, approximately 20 km east of its position when the westerly flow aloft was neglected. As in the previous cases, a southerly low-level jet began to form at approximately 2000 LST. In this case the low-level jet was weaker, and there was a relatively large jet-free region along the Front Range. In this region the flow was due to downslope flow.

With a low-level easterly flow the effects of the upper-level westerlies were somewhat more subtle. These effects were not noticeable until approximately 2000 LST and were mainly related to the strength of the downslope flow. With the effects of upper-level westerlies included along with low-level synoptic easterly flow, the southerly downslope flow forced by the Palmer Lake Divide was stronger and helped contribute to a stronger convergence zone along the Front Range. In this case the low-level jet was stronger, compared with a weakening of the jet when the low-level flow was from the west. The major effect of the upper-level westerlies in this case, however, was to produce a cyclonic vortex in the vicinity of the Front Range. This vortex was especially obvious at a height of 250 m.

d. Cloud shading effects

In any region where convection develops during the day the effects of cloud shading, as well as the dynamical effects of the convective system, are going to be important to the low-level flow experienced. For example, in a region with significant cloud cover it could be expected that the land surface would begin to cool earlier than in cloud-free regions, due to the attenuation of solar radiation by the cloud. In mountainous regions such a cycle may result in the earlier transition to downslope flow.

A simple routine to represent the effects of cloud cover on the shortwave radiation through a specification of albedo was included. The assumption underlying this approach was that the primary effect of cloud cover was to reduce the incoming solar radiation at the surface. The influence of cloud cover on the heating of the atmosphere is ignored.

A cloud is assumed to be present if the relative humidity exceeds 60%. This cutoff value was chosen as it is the average relative humidity for the eastern slopes of the Continental Divide at 1200 LST, the time at which convection begins to develop over this region climatologically. The maximum relative humidity in a column was determined and this value used to determine a new albedo value. The variation of albedo with relative humidity is shown in Fig. 12. The albedo was not allowed to exceed 0.9.

During the day, up to 1600 LST, the effects of this scheme on surface temperature compared to the experiment in section 4c were negligible, with a virtually identical surface temperature distribution to the case in which cloud shading effects were neglected. The effects of the scheme became evident later in the day when the eastern slopes of the Continental Divide began to cool. As expected, the eastern slopes cooled earlier when cloud shading was represented. At 1800 LST the eastern slopes were 1° to 2°C cooler when cloud
FIG. 10. The winds at 250 m at 2300 LST for low-level flows from (a) the north, (b) the east, (c) the south and (d) the west.
shading was included. The effect of cloud shading on the low-level flow is presented in Fig. 11c. This earlier surface cooling caused the transition to downslope flow to occur approximately an hour earlier than the case when cloud shading effects were neglected. The transition to downslope flows occurred at approximately 1800 LST when the cloud shading scheme was included.

5. Conclusions

Mesoscale numerical model simulations of the flow over the Front Range of the Rockies in northeastern Colorado have been presented. These simulations showed that specific features of the flow are reasonably consistent, regardless of the direction of the synoptic flow and strength of the westerlies aloft. This supports the climatological superposition of surface flows for all days during a summer month as applied by Toth and Johnson (1985).

Upslope flow along the eastern slopes of the Continental Divide and along the Cheyenne Ridge and the Palmer Lake Divide results in zones of convergence along these terrain features. These convergence zones also coincide with the regions of most moist and un-
Fig. 12. Variation of albedo with relative humidity for the cloud shading studies.

stable air and hence are preferred regions for the development of convection. This is particularly true along the eastern slopes of the Continental Divide and along the Palmer Lake Divide; however, with a westerly geostrophic flow, convection is also likely to develop along the Cheyenne Ridge.

As these slopes begin to cool around sunset, the flow reverses and the main convergence zone is then located in the Platte River Valley. The development of the downslope flow occurs up to 2 hours earlier when both westerlies aloft and cloud shadowing occur. Pielke and Segal (1986) suggest that such a process may be a possible mechanism for MCC genesis. At the same time, acceleration of southerly downslope flow from the Palmer Lake Divide results in the northward advection of the weaker, preexisting Palmer Lake Divide convergence zone, so that it is oriented NE–SW in the Platte River Valley. This also coincides with the development of a southerly to southwesterly nocturnal jet.

It appears that as the Front Range convergence zone propagates eastward, the nocturnal jet is destroyed and replaced by light flow, the direction of which depends on location in the Platte River Valley. In some cases the low-level flow may develop a cyclonic vortex in the valley, while in the remaining cases there is a cyclonic turning of the wind but no vortex formation. The model vortex is due to heating effects coupled with convergence of the low-level flow. Similar vortices have been observed by Abbs (1986) and Harada (1981) and modeled by Abbs (1986) and Kimura (1986). This contrasts with the vortex formed by terrain blocking when the geostrophic flow is between southeast and west. Usually, these vortices will dissipate during the morning, once the terrain heating into a deep boundary layer becomes the dominant forcing mechanism.

Obviously, these simulations are unable to explain many of the local variations which are attributable to cumulus dynamics on severe weather days. However, they do show the general surface flow that is experienced over this region when the synoptic forcing is weak and should be of value in predicting regions of most likely occurrence of first cumulonimbus convection.

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