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2A.2

LOW-LEVEL STABLE LAYERS, BLOCKING, AND LOCALLY HEAVY SNOWFALL

D.A. Wesley[1], R.A. Pielke[1], J.F. Weaver[2], R.L. McAnelly[1]

[1]Department of Atmospheric Science
[2]NOAA/NESDIS/RAMM Branch
Colorado State University
Fort Collins, Colorado 80523

1. INTRODUCTION

This study focuses on the dynamical structure of two types of precipitating easterly ("upslope") flows along the Colorado Front Range. Frequently the heaviest snowfall does not coincide with the steepest terrain gradients. One type of snow-producing circulation requires the establishment of an arctic air mass in the low levels. In this situation the arctic air mass is firmly entrenched along the east slope. It also may be significantly deeper close to the foothills due to blocking processes. The inversion is of sufficient strength and depth to prevent downslope winds from penetrating the lower atmosphere to the lee of the barrier. Subsequently, moist westerly flow aloft generates heavy precipitation over the Front Range despite surface temperatures often as low as -25 to -30°C.

A quite different scenario also occurs in the same region and involves blocking. The barrier jet/cold air damming process (e.g. 30-31 March 1988) concentrates snowfall in the foothills and adjacent areas to the east. A shallow stable layer near the ground characterized by weak northwesterly flow is created as moist easterly flow is blocked by the terrain. Analysis of soundings, along with mesonetwork data and low-level Doppler radar reflectivitv and velocity scans, indicate that the stable layer causes enhanced overrunning in a north-south band near the foothills. Resulting snowfall distributions reveal a corresponding band of heavier accumulation in the same area. This paper examines the structure of the blocked flow regime and presents evidence of the effects on precipitation development.

The observations and precipitation mechanisms explained in this paper specifically emphasize snowfall in Fort Collins. However, many of the results can generally be applied to any location where gently sloping plains abruptly meet foothills on the lee side of a major mountain barrier. This would include, for example, the Front Range from southern Colorado northward through southeastern Wyoming.

2. CASE STUDIES

2.1 2-5 February 1989 – An Example of Mesoscale Overrunning of a Blocked Arctic Air Mass

In this case, a unique combination of topographical effects along the Colorado Front Range produced significant precipitation despite extremely cold surface temperatures (−20 to −30°C). Complete details of this case study can be found in Wesley and Pielke (1990). Table 1 presents a chronology of daily surface observations at Fort Collins (FCL) during the episode. In the study we attempted to distinguish the various types of precipitation mechanisms as a function of time during the storm. This was accomplished using a large variety of data sources. Even so, for some of the periods it was difficult to distinguish between synoptically-forced events (see Dunn, 1988), and mesoscale overrunning (i.e. mid-level moist flow precipitating as it advects over the cold dome, as opposed to warm frontal overrunning).

Table 1: Daily surface weather observations for Fort Collins, Colorado, during the arctic outbreak of February 1989.

Date (1989)	Max Temp (°C)	Min Temp (°C)	Snowfall (cm)
Feb. 1	−1	−16	1.0
Feb. 2	−16	−26	6.6
Feb. 3	−24	−28	16.5
Feb. 4	−23	−28	12.4
Feb. 5	−15	−28	1.3
Feb. 6	−8	−27	T
Feb. 7	−4	−25	0

The event began on the evening of 31 January when the leading edge of the arctic air mass reached the Front Range. Shallow upslope cloudiness followed the frontal passage as a result of strong easterly winds encountering the eastern Colorado sloping plains and foothills. Low clouds and light snow showers are the typical weather scenario during anticyclonic, shallow postfrontal easterly flow in this region (see Boatman and Reinking, 1984), and light snowfall did occur on 1 February.

Over the next several days the arctic air mass deepened significantly over the area as the upslope flow persisted. Figure 1 presents the Denver (DEN) NWS sounding for 1200 GMT 4 February. While the inversion depth was only about 450 m several hours after frontal passage, the cold air had deepened to about 1500 m by 1200 GMT 4 February. However, mixing ratios within the cold air mass were consistently 0.4 to 0.6 g/kg during the period, which could not account for the heavy precipitation observed in FCL on the 3rd and 4th even if strong surface convergence had been present. As a possible alternative precipitation mechanism, it was noted that westerly flow aloft became more moist with time, a situation conducive to deeper cloud development in and above the cold air (Boatman and Reinking, 1984). Mixing ratios in this air mass were consistently 1.5 g/kg. As larger-scale disturbances moved through this westerly flow, significant amounts of precipitation-sized ice crystals fell through the cold air to the ground (see Fig. 2 for a representative NGM analyses during the storm).

It was also noted that during several of the periods of heaviest precipitation along the Front Range, rising surface pressures, deepening of the cold pool, and formation of deep cloudiness above the cold pool all occurred nearly simultaneously. These observations indicate mesoscale overrunning as the primary precipitation mechanism.

During the early stages of the event, while the upslope cloudiness was visible on GOES images, a persistent bright band was observed along the western edge of the low clouds, with cloud-top temperatures 3 to 5°C colder than in the upslope cloudiness to the east. The bulge in the cold air mass was probably formed by terrain blocking as the easterly winds were unable to rise with the topography and cold air accumulated along the western edge of the cold air mass. During the period of 1700-2359 MST 2 February (when light to moderate snowfall occurred in FCL), a bright band could again be seen along the Front Range in the IR images which indicated that another bulge in the cold air had developed. This bulge was caused by a secondary surge of cold low-level air. DEN wind profiler data during this period indicated a noticeable increase in the depth of the upslope flow beginning at 1600 and lasting for about three hours. The cold surge was also confirmed in the profiler data through a marked increase in strength of the low-level upslope flow. The PROFS mesonet station (FOR) in northwestern Fort Collins measured rapidly rising surface pressure (about 4.5 mb/ 3 hr) beginning at 1430, accompanied by a temperature drop of 4.5°C in 3 hr and strengthening northeasterly surface winds. The precipitation during this time period was attributed to mesoscale overrunning.

The snow intensity at FCL increased gradually during the afternoon of 3 February (beginning at 1300) with visibility decreasing to 0.8 km or less and weather remarks at both the 1600 and 1700 times reporting heavy snow (S+). At FOR, where precipitation rates dramatically increased at 1515 (from trace amounts to 0.3 – 0.5 mm/hr), pressure began to rise at 1400 after a rapid fall, indicating some deepening of the cold pool. The dramatic increase in snowfall intensity at FCL beginning at 1500 was accompanied by a noted increase in high cloud coverage over the Front Range at approximately the same time. The cloud increase was especially notable between FCL and Leadville (LXV). DEN, however, reported little or no increase in snowfall rate and the deeper cloudiness did not arrive until about 2330.

At 1700 3 February, the southwest-northeast band of very cold cloud tops (–35°C) through northeastern CO was clearly identified. At this time, very low visibilities existed along the northern Front Range, caused by moderate to heavy snowfall. Cheyenne (CYS) reported heavy snow at 2100 and 2200, with liquid precipitation rates of about 0.5 mm/hr. At 0131, the band of deep cloud covered the FCL-CYS region and was well-correlated with areas of moderate to heavy snowfall. DEN reported moderate to heavy snow from 0300 to 0500 just as the deep clouds had begun to spread over this area. Until this time, only trace snow amounts were recorded there.

Beginning at 1900, DEN profiler data again indicated significant heightening of the shear region at the top of the cold pool. This lasted until about 0400 after which snow intensity decreased at FCL. The city received a total of 9.1 cm of snow overnight attributed to a combination of mesoscale overrunning and synoptic-scale ascent, accompanied by deep cloudiness.

Observations of snow crystal types during the snowstorm provided additional information regarding the location of the moisture source for the snowfall. The authors observed primarily dendritic and stellar crystals, predominantly aggregated and frequently heavily-rimed, during several observations of moderate to heavy snowfall on the days of 2-4 February. Several meteorologists in the Boulder area also reported primarily dendritic aggregates during moderate snowfall on 4 February. Significant graupel accumulation was observed in FCL during two episodes of the evening hours of 3 February and the early morning hours of 4 February.

These crystals could not have originated within the –20 to –30°C conditions observed within the cold air throughout the storm (Fig. 1). The moisture source apparently resided in the moist westerly flow aloft, which exhibited –10 to –20°C temperatures, a range favorable for dendritic crystal growth. Such evidence is strongly supportive of the mesoscale overrunning scenario described previously.

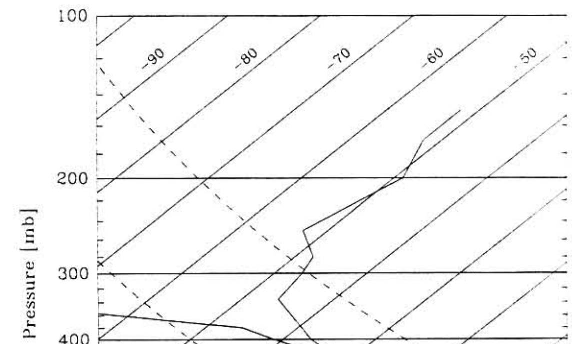

Figure 1: Skew-T diagram for DEN radiosonde data. Isotherms (°C) are the diagonal solid lines running upward from left to right. Dotted lines are selected dry adiabats. 1200 GMT 4 February 1989.

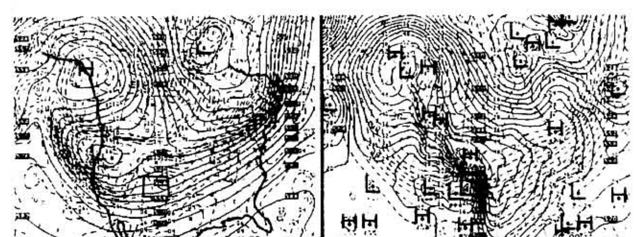

Figure 2: NGM analyses for 1200 GMT 4 February 1989. (a) 500 mb and (b) MSL pressure.

2.2 Simulations of the Cold Pool Evolution

In order to further investigate the dynamics of the arctic outbreak, and since the observational data set was somewhat limited, several two-dimensional experiments were performed using the Colorado State University Regional Atmospheric Modeling System (CSU RAMS; see Schmidt and Cotton, 1990). This numerical model allows some preliminary tests of the mesoscale overrunning hypotheses discussed in Section 2. This model has been utilized previously in studies of Colorado winter storms (Wesley and Pielke, 1988).

Table 2 presents the basic options employed in this mesoscale model for the February 1989 storm. The model was initialized horizontally homogeneously, using the temperature and wind profiles measured at DEN during the initial stages of the storm after the cold air mass was established. Figure 3 presents the temperature field in the x-z cross section after three hours of simulation. Note the strong inversion present at the top of the arctic air mass at approximately 2.8 km MSL. Of particular interest is the increased elevation of the temperature surfaces over the sloping region east of the barrier at heights of 3.1 to 3.3 km MSL. As previously discussed in Section 2.1, a persistent bright band was observed along the foothills region for many hours during the storm and may have influenced the dynamics of the overrunning westerly flow. This band exhibited cloud-top temperatures approximately 3 to 5°C colder than the upslope cloud tops to the east. The similarity in the model features is striking.

The vertical motion field predicted by the model at this time is shown in Fig. 4. Downward motion associated with westerly flow over the barrier is confined to regions above the inversion and is much smaller in magnitude than that usually associated with strong cross-barrier flow, as reported in Lee *et al.*, 1989. Only weak vertical motion characterizes the regions below the inversion. This is also consistent with the blocking scenario described previously. Strong downward motion in the low levels over the plains was found in sensitivity simulations without the cold pool (not shown).

Table 2: Model options used in the Regional Atmospheric Modeling System (RAMS) for the numerical experiments.

Model Category	Option
initialization	horizontal homogeneous, DEN sounding 0000 GMT 3 Feb.
dimensions	2-dimensional east-west
top boundary condition	Rayleigh friction
height of model top	16 km
lateral boundary conditions	radiative, with mesoscale compensation region
thermodynamics	dry
radiation	longwave and shortwave parameterizations
horizontal grid, resolution, size	1 grid, 5 km, 50 grid points
vertical grid, resolution	1 grid, 50 m near surface stretched to 500 m above 10 km MSL
topography	silhouette-averaged from 30 sec. data
time step	30 s

These experimental and simplified simulations provide further confirmation that the cold pool indeed may extend deeper into the atmosphere along the eastern foothills of the Rocky Mountains during cold air outbreaks. This vertical extension provides additional retardation of downslope flow associated with strong westerly winds aloft and probably sets the stage for additional condensate production above this region.

2.3 30-31 March 1988 - An Example of Low-Level Convergence Associated With Cold Air Damming

Meso-fronts associated with cold air damming often form in place just east of the foothills and are sensitive to the nature of the low-level synoptic easterly flow. Measured vertical profiles associated with the blocked surface patterns reveal a distinctly layered temperature and wind structure. Predominance of heavily-rimed, dendritic aggregates implies lifting associated with the layered vertical structure. Bands of enhanced radar reflectivity, apparently initiated by overrunning, exhibit significant correlation with snowfall intensity.

Damming of upslope flow led to a localized region of heavy snowfall in the 30-31 March 1988 storm (>80 cm; see Fig. 5). Note the extension of heavy snowfall onto the plains

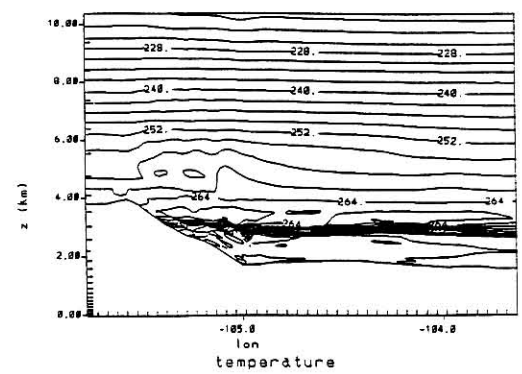

Figure 3: Model-predicted temperature field (°K) over the Front Range region at three hours of simulation. Latitude of east-west domain is that of DEN. Region of close packing in isotherms is the inversion at the top of the arctic air mass. Temperatures within the cold pool are 240 to 250° K.

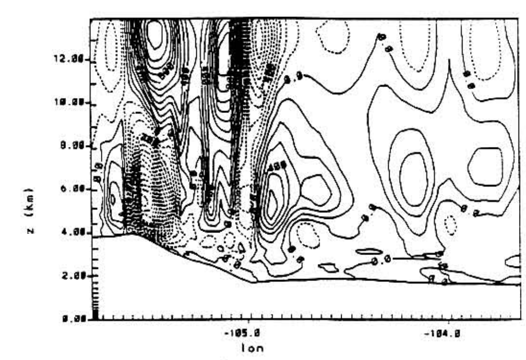

Figure 4: Model-predicted w field (m/s) over the Front Range region at three hours of simulation. Dotted lines are contours of negative w. Strong downslope is confined to heights above the arctic air mass. Maximum updraft speed for this portion of the domain is about 1.2 m/s, while maximum downdraft is 1.6 m/s.

just east of the foothills near the FCL area. At 1200 GMT 31 March (see Fig. 6), when snowfall at FCL was heaviest, the 500 mb analysis indicated a strong, negatively tilted cutoff low over north-central Arizona with weak positive vorticity advection in northeastern Colorado. The corresponding 700 mb low center was located just to the east with east-southeasterly moist flow over the area of interest. The 850 mb analysis indicated a strong low just south of the Four Corners with very strong easterly flow over northeastern Colorado. A weak wedge or nose of low-level high pressure existed over the Front Range at this time (dotted line in Fig. 6b) according to the mean-sea level pressure NMC analysis. This ridging is a result of the development of a cold pool of air against the foothills. This cooling creates slightly higher surface pressures along the Front Range than in surrounding regions.

Figure 7 presents the surface streamline analysis for the PROFS mesonetwork during the early stages of the storm (on the evening of 30 March). This indicates the initiation of a quasi-stationary convergence line due to the inability of the stably stratified, low-level easterly flow to ascend with the terrain. This line developed as a cold pool formed over and next to the foothills. At 0100 GMT 31 March, the northeasterly flow (speeds approximately 2.5 to 7.5 m/s) on the plains converged with north-northwesterly flow (2.5 to 5.0 m/s) along a northwest-southeast line indicated by the heavy dashes. Maximum convergence occurred along the northernmost portion of the meso-front (near FCL). Thus the cold pool dammed the northeasterly low-level flow. The presence of colder air to the west of this line was readily apparent as was a tendency for the line to move southeastward during the latter portion of the storm.

A more detailed look at the vertical structure in this storm is shown in the CLASS (Cross-chain Loran Atmospheric Sounding System) soundings taken just northwest of FCL. Examples of temperature profiles are in Fig. 8. The vertical profile at 0427 GMT was measured during the first few hours of moderate to heavy snowfall in FCL. As shown, three distinct layers of the atmosphere are apparent:

1. a shallow low-level layer of cool air (west of the convergence line shown in Fig. 7) exhibiting weak northwesterly surface winds and stable conditions (approximately 840 to 810 mb),

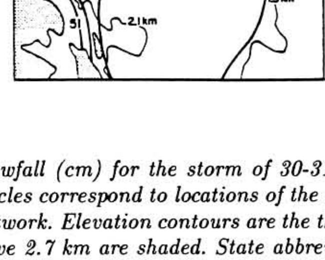

Figure 5: Snowfall (cm) for the storm of 30-31 March 1988. Heavy circles correspond to locations of the snowstorm observer network. Elevation contours are the thin black lines; areas above 2.7 km are shaded. State abbreviations are as follows: Wyoming (WY), Nebraska (NE), and Colorado (CO).

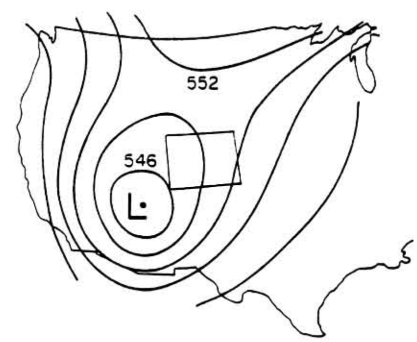

Figure 6a: National Meteorological Center (NMC) height analysis for the 500 mb surface at 1200 GMT 31 March 1988. Heights are in dm. The state border of Colorado is outlined.

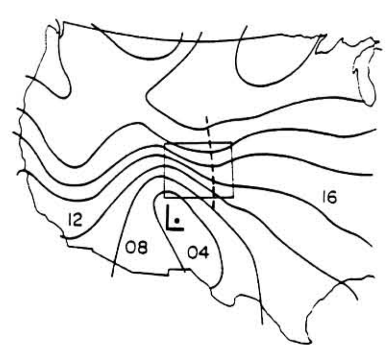

Figure 6b: NMC mean-sea level pressure (mb) analysis for 1200 GMT 31 March 1988. Heavy dashed line is a convergence line (see text).

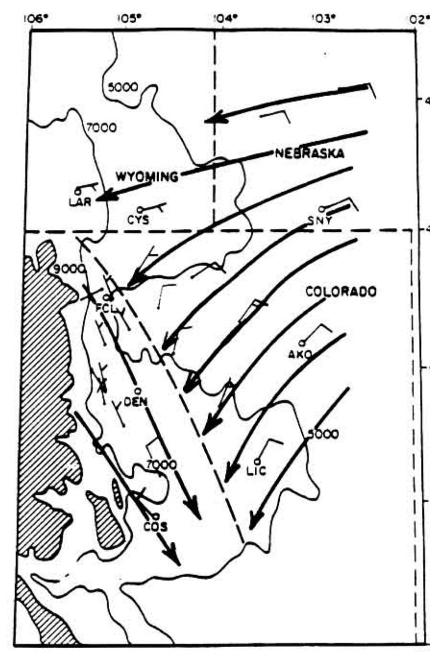

Figure 7: Surface streamline analysis for 0100 GMT 31 March 1988.

2. a moist mid-level easterly upslope flow (810 to 570 mb) which includes a potentially unstable layer from 810 to 770 mb,

3. a fairly moist upper layer consisting of strong southerly winds (above 570 mb); this represents the synoptic flow around the cutoff low pressure system to the southwest.

Thus layer 2 overrides layer 1, creating enhanced vertical motions in a moisture-rich region. Visual observations of cloud base height just northwest of FCL (at 2305 GMT 30 March) indicated an approximate value of 1.1 km or 740 mb, which corresponds with the lower portion of layer 2 at this time as shown in Fig. 8. Similarly, layer 3 overrides layer 2. The moist conditions observed in layer 3 indicates a potential seeder mechanism by this region for the lower layers (Boatman and Reinking, 1984). Importantly, crystals nucleated in layer 3 do not evaporate while descending into lower layers, and are able to grow to large (2 to 4 mm diameter according to surface observations) sizes due to the presence of moist easterly flow from 700 to 650 mb. Temperatures in layer 2 are conducive to dendritic crystal growth; these temperatures, ranging from −12 to −16°C, occurred in relatively strong (10 to 15 m/s), moist easterly flow. The snowfall distribution for the storm, which does not imply simple topographic lifting (Fig. 5), indicates that intense production and growth of precipitation particles occurred as the layer 2 ascended over layer 1, before the easterly flow reached the barrier. The surface convergence line separates layer 1 from layer 2, and the demarcation between the layers apparently slopes upward to the west, yielding a cold pool approximately 30 mb deep over FCL.

During the early portions of the storm on the evening of 30 March, surface observations taken at FCL indicated oscillations of wind direction from easterly to northwesterly; these seemed to have an approximate period of a few hours. The convergence line may have been crossing the area during this time producing changing wind directions. Examination of the surface analyses revealed some movement of the convergence line with time. Snow intensity observations taken at the same times also indicated some oscillation from moderate to heavy, although not clearly in phase with the wind direction. Generally, the heaviest snow was observed at FCL during the periods of northwest winds. This observation is not surprising, considering the vertical storm structure described previously, where lifting along the upper portion of the cold pool, was suggested to create the heavy snowfall.

According to CP-2 Doppler radar measurements, a persistent wide high-reflectivity band was located along and to the west of the convergence line, thus confirming the enhanced ascent in layer 2. This band is clearly evident in the composite in Fig. 9, where reflectivities above 25 dBz are located just east of the foothills region between FCL and DEN. The second band in this composite, located approximately 60 km southeast of DEN, may have been produced by blocking processes since observations of wind direction at Limon and Elbert implied strong convergence in this region. However, lack of detailed surface data prevented a comprehensive surface wind analysis. Stronger reflectivities were observed near the upper portion of the cold pool at FCL and were related to condensate production as the upslope layer rose over the dammed cold pool. Some propagating bands of enhanced reflectivity aloft were observed in regions well east of the convergence line, where heavy snowfall did not occur.

Observations of snow crystal types provide additional insight into the precipitation mechanisms discussed previously. During the 30-31 March storm, a spotter network reported predominantly aggregated, heavily-rimed spatial dendrites, as well as several 15 to 30 minute graupel episodes. Many of the dendrites making up the large aggregated snowflakes contained heavily rimed single crystals in their cores. This tendency for riming to be concentrated near the centers of the individual dendrites rather than the branches implies that liquid water was present in layer 3 (see Fig. 8) with subsequent dendritic crystal growth occurring in layer 2. The observation may be the result of liquid water accumulation near cloud top, as in the

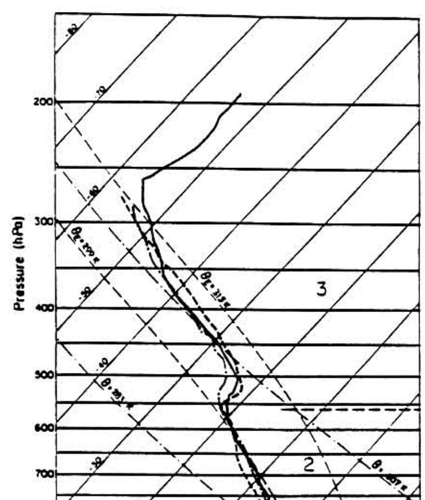

Figure 8: CLASS temperature profiles for the first three balloon releases at FCL, for the 30-31 March 1988 storm. Horizontal solid lines are pressure (mb). Temperature (°C) values are the solid diagonal lines. Two labeled adiabats (dashed-dotted lines) are shown, as well as two moist adiabats (dashed lines). Horizontal dashed lines divide the atmosphere into three layers, denoted 1,2 and 3 (see text for discussion of the layered structure).

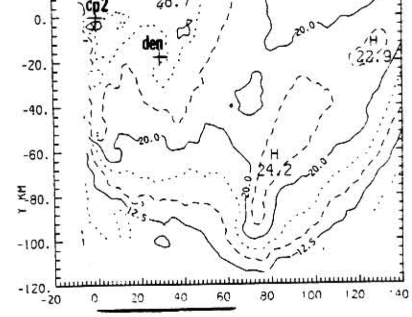

Figure 9: Composite CP-2 radar reflectivity analysis (dBz) for a PPI scan at 0.7°. Average is computed for 19 observations for three hours between 0641 and 0941 GMT 31 March 1988. Ground clutter is removed from analysis.

northern Colorado mountain cloud systems reported in Rauber *et al.* (1986). Unrimed and unaggregated crystals comprise only about 9 percent of the observations. These results indicate that liquid water layers played a significant role in precipitation production although the exact location of these layers is not readily determinable from these data.

Ongoing RAMS mesoscale simulations of this event utilize detailed silhouette-averaged topography and two-way interactively nested grids. An example of the model topography for grids 2 and 3 is shown in Fig. 10, where the major features in terrain are evident, including the east-west ridges over the eastern plains of Colorado. Results of these simulations pertaining to blocking will be presented at the conference.

3. Conclusions

In this paper we have looked at two types of winter precipitation events with distinct precipitation mechanisms. Overrunning characterizes both circulations. From the February 1989 case, and others where satellite and other data were available to establish the cold pool depth, we believe that the following sequence can occur in place of, or as a supplement to, more common precipitation scenarios (i.e., simple synoptic forcing or upslope):

- The magnitude of the geostrophic easterly component increases with time.
- Within the cold pool, mass increases along the foothills.
- The cold pool, as a result, deepens locally on its western edge due to the upslope flow and blocking processes. Surface pressure increases in this region.
- Simultaneously, enhanced precipitation occurs, as the moist westerly flow aloft overruns the deepening cold dome.

Effectively, the cold air acts as a barrier to the westerly flow. Preliminary mesoscale model simulations capture the development of overrunning and a bulge on the western edge of the cold pool. Strong downslope flow east of the mountains is prevented by the arctic air mass.

From the March 1988 case, we have found the following:

- The vertical structure to the lee of the terrain barrier includes a shallow layer of cold air which dams and induces uplift of deep, moist upslope flow. The eastern edge of this layer formed a well-defined convergence zone at the surface east of the foothills.
- Analysis and comparison of snowfall data and locations of the blocking-induced convergence line indicate that the moisture content of the low-level upslope flow overrunning the dammed cold air is critical to the location of heavy snowfall.
- Rapid dendritic growth occurred in the upslope flow. Precipitation termination occurred as upslope flow above the surface weakened and the upper air mass (above the frontal inversion aloft) became drier.
- Overrunning produced banded radar reflectivity structures.

Both of the blocking mechanisms discussed in this paper must be considered by the wintertime forecaster in this region as contradictions to traditional theories regarding heavy snowfall production. The problems with forecasting precipitation amounts during these situations are well-known. The damming scenario has been documented in two case studies (Dunn, 1987 and Wesley and Pielke, 1990) in eastern Colorado, but these

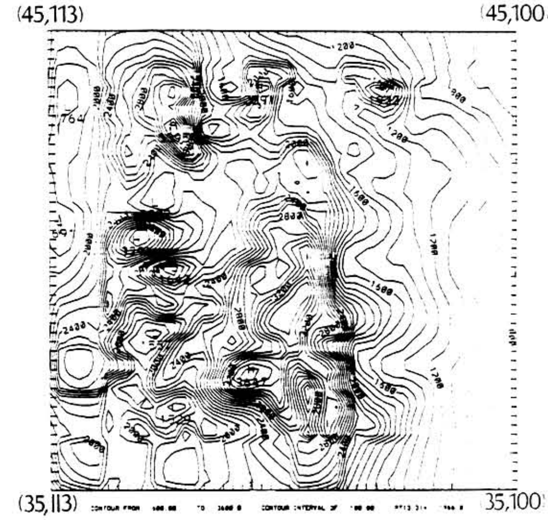

Figure 10: Model topography (grids 2 and 3) for simulations of the 30-31 March 1988 storm. Elevations are in m.

studies were not meant to be comprehensive or applicable to a more generalized dynamical structure. The role of the arctic air mass in preventing downslope flow has been investigated by Lee *et al.* (1989), but the applications to wintertime precipitation in this region were not discussed in that paper. This study has documented mesoscale overrunning as the cause of enhanced snow accumulation over the Front Range during an arctic outbreak.

On a more general note, this paper has demonstrated that low-level cold air masses can act as obstructions to large-scale wind flow, leading to overrunning and precipitation. The topography of the region is responsible for maintaining the depth of the cold air near the eastern edge of the mountains in *both* types of circulations described in this paper.

4. Acknowledgements

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