

HEAVY SNOWFALL DURING AN ARCTIC OUTBREAK ALONG THE COLORADO FRONT RANGE

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Abstract

During the period 1–5 February 1989, a record-breaking arctic air mass invaded the western United States, abruptly ending a period of unusually warm weather that had characterized the region during most of January. The postfrontal upslope flow along the eastern slope of Colorado and Wyoming, along with relatively warm, moist westerly flow aloft, produced periods of heavy snow despite surface temperatures averaging -5 to -20°F (-21 to -29°C). Traditional wisdom would suggest that air at these temperatures is too cold to hold much water, and therefore, that heavy snowfall is normally precluded.

This paper utilizes satellite, sounding and surface observations, as well as mesoscale model simulations, to examine the role of the low-level arctic air mass in preventing downslope, subsiding westerly winds over the region in this storm. One important precipitation enhancement mechanism, terrain blocking, is examined by analyzing the roles of the moisture source above the cold air, the depth of the cold air and the snow intensity. These findings are of immediate benefit to the forecaster, who must recognize the possibility of heavy snow over the Colorado/Wyoming Front Range despite extremely cold temperatures at the surface and strong mid-level westerly flow.

1. Introduction

Heavy snowfall is associated most often with relatively warm surface temperatures (i.e., 15 to 30°F ; -1 to -9°C) since water content of the air becomes diminishingly small as the air mass becomes colder. Auer and White (1982) found that for the vast majority of heavy snowfall events in the United States, the region of most favorable dendritic ice crystal growth (-12 to -16°C) is found between the heights of 3.4 and 5.2 km MSL (or, for the present case, approximately 620 to 550 mb) in the atmosphere, a situation not generally associated with 850 mb temperatures colder than -10°C (assuming typical low-level lapse rates). Indeed, the occasional deep cyclonic system, producing large snow amounts along the Front Range of the Rocky Mountains (refer to Fig. 1 for all geographic references), is almost always associated with surface (near 850 mb) temperatures in the 15 to 30°F (-1 to -9°C) range during the periods of heavy precipitation (Wesley and Pielke, 1990; Schlatter *et al.*, 1983). However, in this particular region, a unique combination of topographical effects can occasionally produce significant wintertime precipitation despite much colder surface temperatures (i.e., -5 to -20°F ; -21 to -29°C). This paper will show how moist westerly flow over the Rocky Mountains, combined with a cold air mass dammed against the east slopes of the mountains, resulted in heavy snowfall in just such a case. This occurred during the record-breaking 1–5 February 1989 arctic outbreak, which affected much of

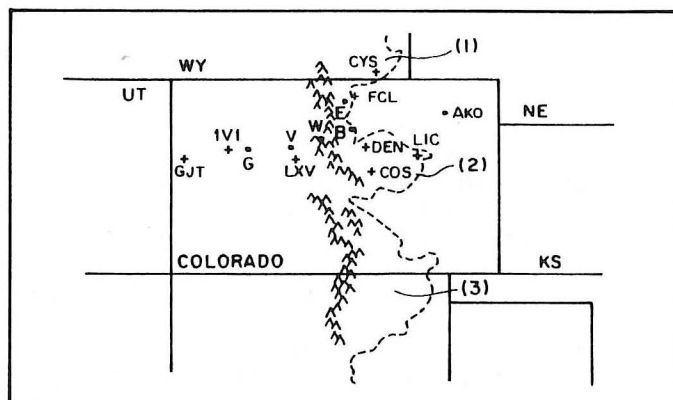


Fig. 1. Topography of the Front Range region. Dotted line is the approximate location of the 5000 foot elevation contour. Primary station identifiers are as follows: Cheyenne (CYS), Fort Collins (FCL), Grand Junction (GJT), Colorado Springs (COS), and Denver (DEN). Secondary locations mentioned in the text are Akron (AKO), Boulder (B), Limon (LIC), Vail (V), Estes Park (E), Glenwood Springs (G), Rifle (1V1), Leadville (LXV), and Winter Park (W). The numbered locations 1, 2, and 3 refer to the Cheyenne Ridge, Palmer Divide and Raton Mesa, respectively. Major mountainous zones (along the Front Range only) are denoted by inverted V's.

central North America, including the Colorado Front Range. Table 1 presents a chronology of daily surface observations at Fort Collins (FCL in Fig. 1) during this episode. Note the occurrence of heaviest snowfall on the coldest days.

The schematic in Fig. 2 depicts the blocking-induced circulation observed over the region for several days during the arctic outbreak. In such cases, the moist flow aloft rises over the Rocky Mountains, generating precipitation particles. With the arctic air entrenched over the eastern foothills, downslope flow and drying are significantly reduced. In fact,

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

Date (1989)	maximum temperature ($^{\circ}\text{F}$)	minimum temperature ($^{\circ}\text{F}$)	snowfall (inches)
Feb. 1	30	4	0.4
Feb. 2	4	-14	2.6
Feb. 3	-11	-18	6.5
Feb. 4	-10	-18	4.9
Feb. 5	5	-19	0.5
Feb. 6	17	-17	T
Feb. 7	24	-13	0

¹NOAA/NESDIS/RAMM Branch

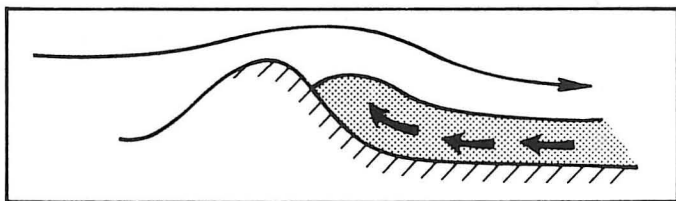


Fig. 2. Schematic of the dynamical circulations present over and east of the Rocky Mountains during an arctic outbreak. Shaded area is the cold pool, characterized by weak to moderate easterly flow. Overrunning westerly flow is strong and extends from just above the cold pool into the stratosphere.

the western edge of the cold pool is actually deeper due to blocking of the upslope flow. Precipitation can thus be heavier on the western edge of the cold pool. The precipitation enhancement over the foothills and immediate areas to the east is especially important as synoptic-scale disturbances imbedded in the westerly flow aloft advect over the cold pool, generating solid and liquid hydrometeors as the mountain barrier is met. The resulting precipitation distribution for the entire February 1–5 event, shown in Fig. 3, demonstrates the extension of significant precipitation amounts 10 to 20 mi east of the foothills along Colorado's northern Front Range. The bulge in the cold pool may even induce additional upward motion and condensation in the westerly flow, although additional research is required to confirm this possibility. Effect-

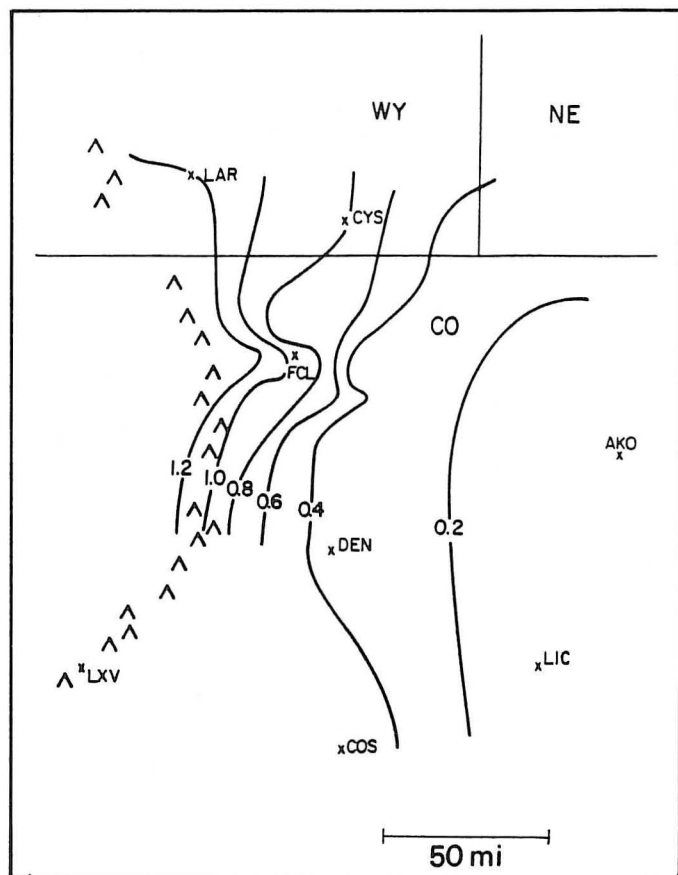


Fig. 3. Precipitation distribution (inches of liquid equivalent) for the Front Range and surrounding area, 1–5 February 1989. Measurements are from NWS stations and trained cooperative observers participating in a snow spotter network.

tively, the cold pool acts as a terrain feature to the overlying air mass (Lee *et al.*, 1989). Meanwhile, further to the east over the Wyoming and Colorado plains, stronger downslope flow reduces precipitation.

Damming scenarios have been described for other storm types in this region by Wesley and Pielke (1990), Dunn (1987), and Weaver and Toth (1990). The precipitation enhancement described here is partially the result of a seeder-feeder mechanism (Juisto, 1967 and Reinking and Boatman, 1986), but depends inherently on a vertical extension of the western edge of the cold pool. Precipitation particles generated in the westerly flow continue to grow as they fall through the nearly saturated (with respect to water) cold air mass, although the latter growth is restricted by the low mixing ratios. Sections 2 and 3 further address the moisture content of the two air masses depicted in Fig. 2. Section 4 presents corroborating evidence for the precipitation mechanisms discussed in this paper, utilizing snow crystal observations, and Section 5 discusses some numerical model experiments which use both actual and modified soundings as initializations.

The observations and precipitation mechanisms explained in this paper specifically emphasize snowfall in Fort Collins. However, 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 the region from southern Colorado northward through southeastern Wyoming and much of central Montana.

The precipitation intensities mentioned in the next two sections are based on two surface measurements: visibility and liquid precipitation rate. In order to assess the role of fog in reducing visibility during this storm, the two measurements at FCL were plotted against each other for the five-day period. The result was a very high correlation between low visibility and precipitation rate. Generally, all of the 6-hour heaviest snow periods exhibited visibilities much less than one mile, while those periods producing only a trace to 0.01" of snow (usually with fog present) corresponded to visibilities greater than 3 mi. Furthermore, surface wind speeds were too low (5 to 15 knots) to produce any significant amounts of blowing snow.

2. Synoptic overview and the moisture source

During the latter half of January 1989, temperatures in northeastern Colorado were remarkably warm. At the same time, a record-breaking anticyclone was strengthening over Alaska and western Canada and was responsible for new all-time minimum temperature records in that region (see Tanaka and Milkovich, 1990). This air mass began to move rapidly southward during the last few days of January. The leading edge of the arctic air mass reached the Front Range on the evening of 31 January. Shallow upslope cloudiness followed the frontal passage, as strong easterly winds encountered the eastern Colorado sloping plains and foothills. Low clouds and light snow showers are the typical weather scenario in this region if a southward-propagating surface high pressure system is present in the central Great Plains (see Boatman and Reinking, 1984), and light snowfall did accompany the upslope clouds on 1 February in northeastern Colorado (See Table 1). Figure 4 shows the National Meteorological Center's (NMC) Nested Grid Model (NGM) analysis at 1200 UTC 1 February.

Over the next several days the arctic air mass deepened significantly over the Front Range area as the upslope flow persisted. Figure 5 presents a series of the DEN National

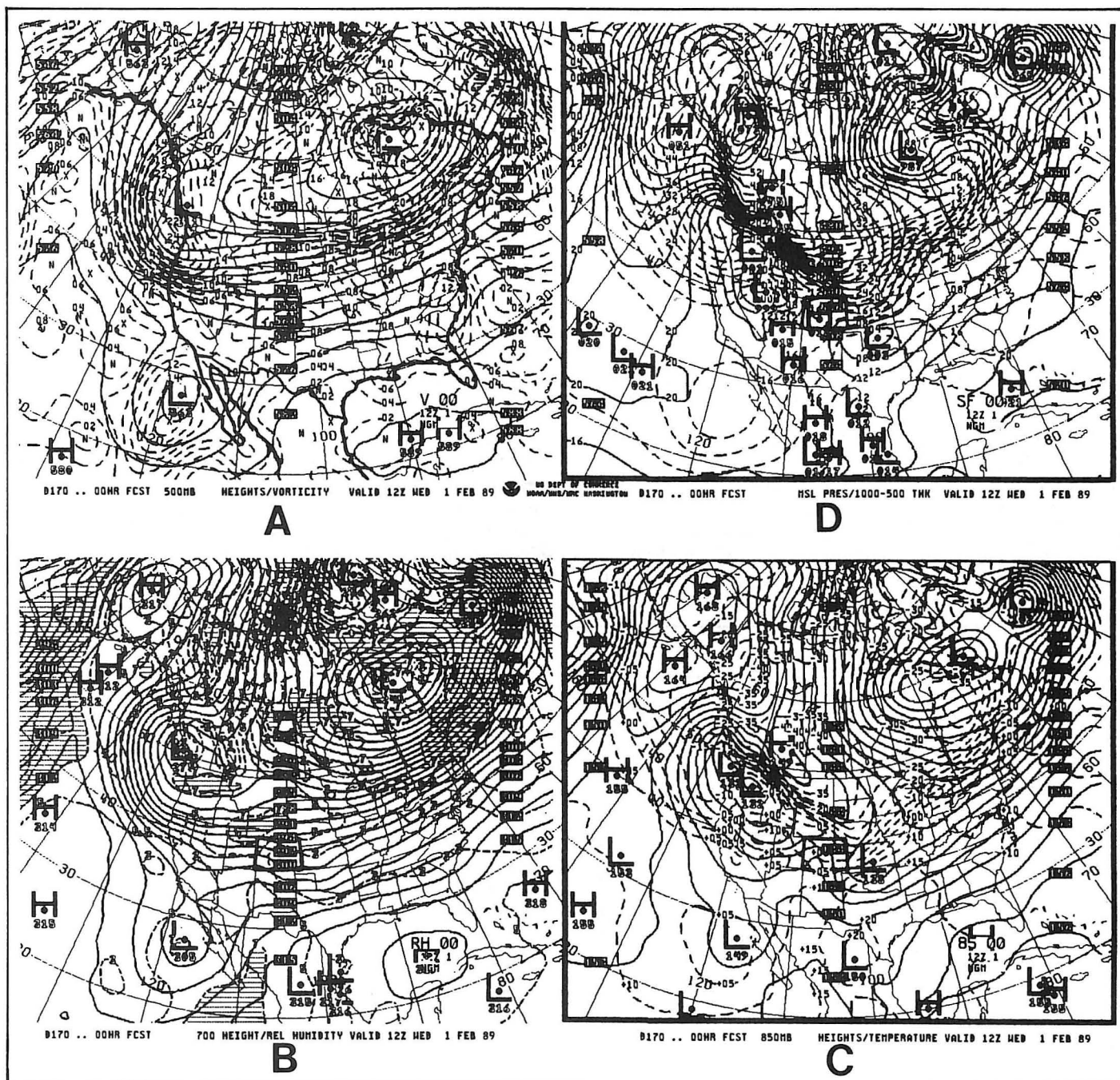


Fig. 4. National Meteorological Center Nested Grid Model analyses for 1200 UTC 1 Feb. 1989. (a) 500 mb (b) 700 mb (c) 850 mb (d) Sea-level pressure. Heights are labelled in dam. The state border of Colorado is heavily outlined.

Weather Service (NWS) soundings for 0000 UTC 3 February through 1200 UTC 4 February. While the inversion depth was only about 450 m (about 50 mb) several hours after frontal passage, the cold air deepened to about 1500 m (150 mb) by 1200 UTC 4 February (also see Fig. 6 for the wind profile at that time). However, mixing ratios within the cold air mass were consistently 0.4 to 0.5 g kg^{-1} during the period, which could not account for the moderate to heavy precipitation observed in FCL on 2–4 February, even if strong surface convergence had been present.

To demonstrate this fact, representative precipitable water values were calculated for the cold air mass during the storm

evolution, using the DEN soundings (See Fig. 5). The approximate precipitable water value is 0.02" for the three-day period shown. Consider an upslope flow velocity of 10 kt in the cold air (this value is overestimated for the present case, considering the surface and sounding wind observations, but is used here for demonstration purposes). Thus a parcel in the cold air moves approximately 60 mi in 6 hours. Assuming that during a six-hour period, all of the water vapor in a 60 mi wide swath condenses and precipitates over a 20 mi width (see the precipitation distribution in Fig. 3), approximately 0.06" accumulates. This value underestimates observed precipitation rates during several portions of the

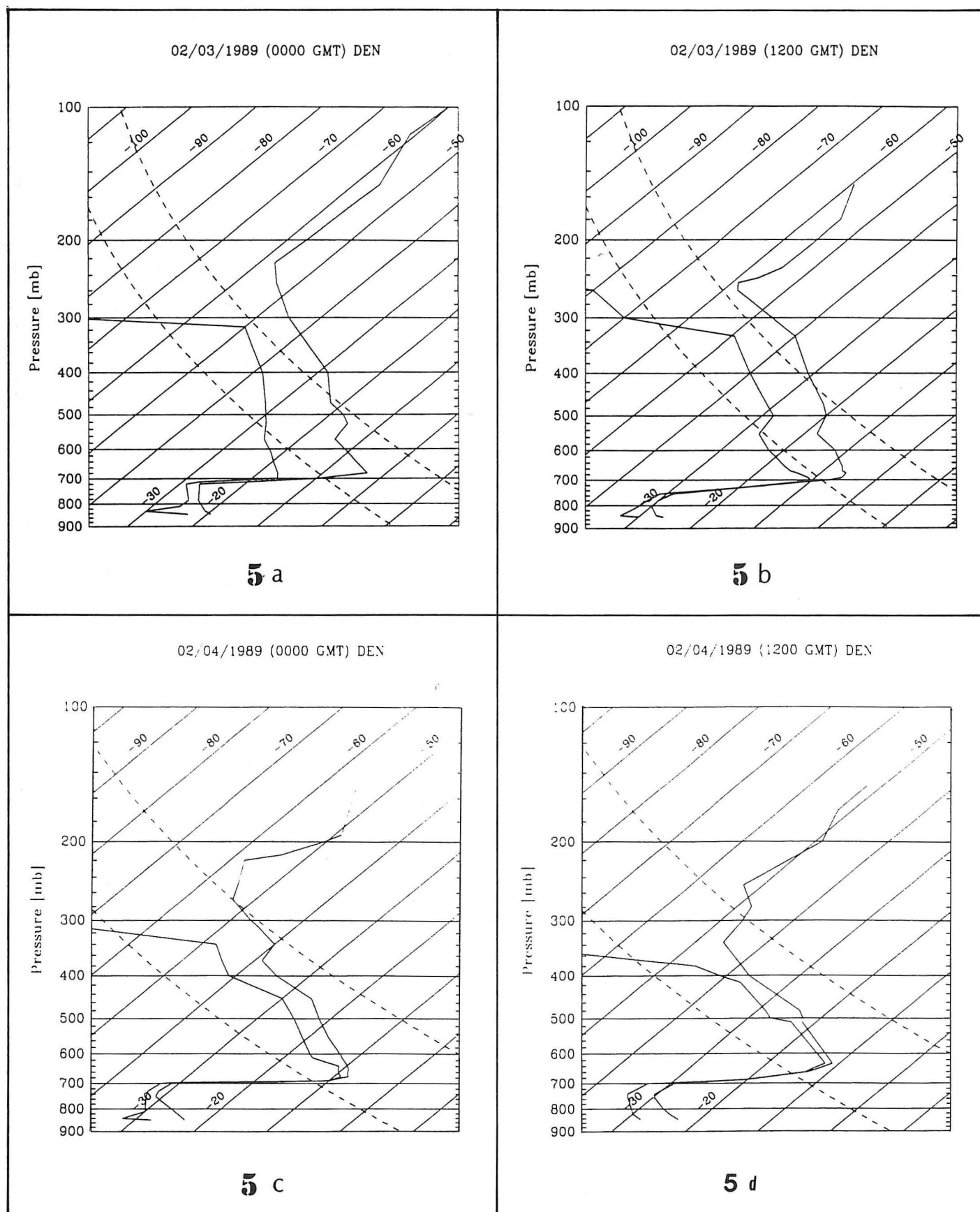


Fig. 5. Skew-T diagram for DEN radiosonde data. Isotherms ($^{\circ}\text{C}$) are the diagonal solid lines running upward from left to right. Dotted lines are selected dry adiabats. (a) 0000 UTC 3 February 1989 (b) 1200 UTC 3 February 1989 (c) 0000 UTC 4 February 1989 (d) 1200 UTC 4 February 1989.

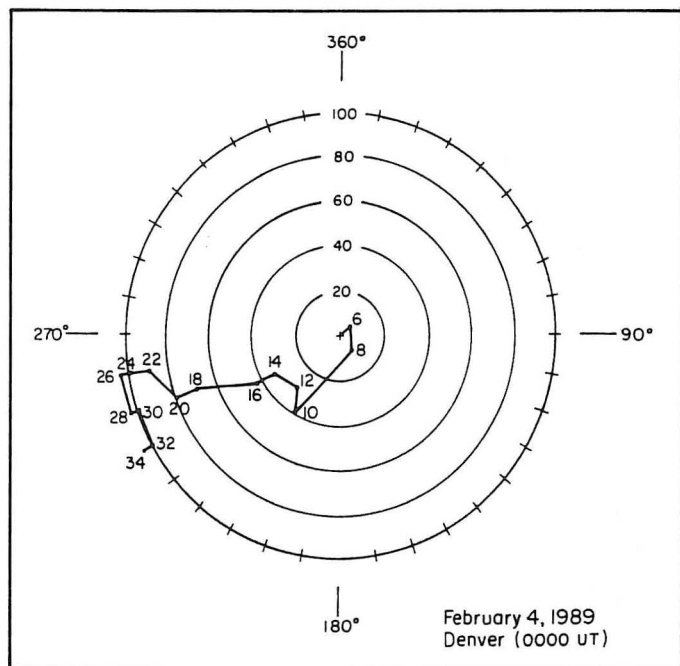


Fig. 6. DEN hodograph for 0000 UTC 4 February 1989. Wind speeds, thousands of feet, MSL. Points plotted are directions from which the winds are blowing.

storm along the northern Front Range by more than a factor of five, assuming a precipitation efficiency of unity. Of course, this approach also assumes that only two-dimensional convergence is occurring, and it ignores the contribution of preexisting cloud liquid or ice particles. Nevertheless, it underscores the inability of strictly low-level moisture to account for the precipitation received.

By utilizing sounding and satellite data, the authors noted that westerly flow aloft became more moist with time both in a relative and absolute sense (see Fig. 5), a situation conducive to deeper cloud development in and above the cold air (Boatman and Reinking, 1984). Larger-scale disturbances moving through this moistening westerly flow produced significant amounts of precipitation-sized ice crystals which fell through the cold air to the ground. The hourly visibility and temperature observations in FCL are shown in Fig. 7. As mentioned previously, heaviest snowfall occurred during the periods of very low visibility.

During some of the periods of heavy precipitation along the Front Range, rapidly rising surface pressures, deepening of the cold pool, and the advection of deep cloudiness over the region all occurred nearly simultaneously. Surface pressure rises were the result of the accumulation of cold air over and near the foothills. These observations indicate that the damming-induced bulge along the western edge of the cold pool played a critical role in the precipitation dynamics in this storm. More evidence for this feature is described in

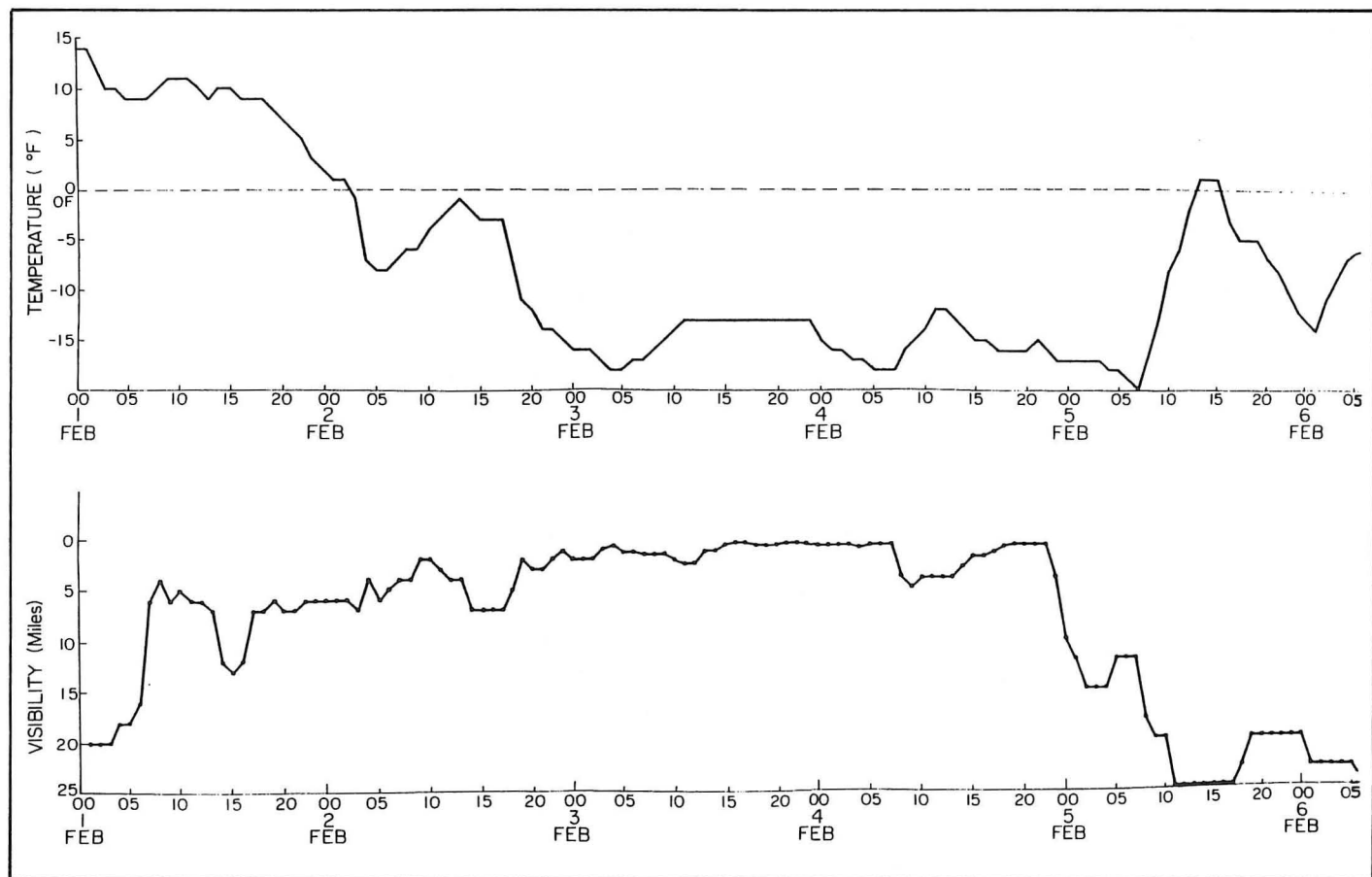


Fig. 7. Selected surface observations for Fort Collins. These data were obtained by averaging those observed at the FCL campus weather station and at MSWS (Mountain States Weather Services) (a) temperature (°F) (b) visibility (mi).

the next three sections. The precipitation gradient near the foothills (Fig. 3) is similar to that of the March 30–31 1988 storm (Wesley and Pielke, 1990), in which cold air damming was a major contributor to the snow production. (Note: in the 1988 case, overrunning occurred in *easterly* flow above the surface).

Some regional snowfall totals for the 1–5 February 1989 storm included FCL at nearly 15" (0.8 liquid), 20" to 30" in Estes Park and 13" (0.9), 11" and 15" (0.9) at Cheyenne (CYS), Boulder and Colorado Springs (COS), respectively. The following sections describe a sequence of events which characterize the evolution of the storm, and explain most of the observations.

3. Storm chronology

Geostationary Operational Environmental Satellite (GOES) visible and infrared data, along with DEN soundings, wind profiler and NMC surface and upper-air analyses, enabled a detailed assessment of the cloud systems responsible for the heavy snowfall during 1–5 February 1989 along the Colorado Front Range. Observations from the PROFS (Program for Regional Observing and Forecasting Services; see Schultz *et al.*, 1985) mesonetwork were also utilized. The data are discussed in relation to standard surface observations at FCL, CYS and DEN.

a. 0700–1300 UTC 1 FEB 1989

Frontal passage occurred several hours prior to midnight LST, and light snow began (trace amounts) during this period in FCL. DEN recorded frontal passage between 0700 and 0900 UTC. The 1200 UTC NGM analyses shown in Fig. 4 suggests that the cold air mass was very shallow (confined to below approximately 750 to 800 mb). Surface observations during this period confirm that the western edge of the cold

pool was located below an elevation of about 2 km MSL (or 800 mb in the standard atmosphere). For example, just west of Denver in the foothills (at this approximate elevation) temperatures reached 45°F (7°C) during the daytime hours of 1 February (Doesken). The NGM analyses also revealed that a strong sea-level pressure (SLP) gradient, directed north-east-southwest, had developed over the region. This gradient is overestimated to some degree due to errors produced by the standard pressure reduction algorithm when a shallow arctic air mass is present over complex terrain (e.g., Pielke and Cram, 1987). It was not clear during this period whether the light snowfall was due to the shallow upslope flow, and/or the weak shortwave seen in Fig. 4a. The presence of a cold-cloud streak over the region on satellite (not shown) also supported the possibility of synoptic forcing.

b. 1300–2000 UTC 1 FEB 1989

The cirrus streak moved off early in the period (Fig. 8). At the same time, the snow intensity *increased* somewhat, as implied by a decrease in visibility (Fig. 7b) and a slight increase in the 6-hour snowfall total reported at 1800 (0.3"). Northeastern Colorado, including the Front Range, was engulfed in low-level cloudiness, with GOES infrared (IR) 11.2 μm cloud-top temperatures (CTTs) ranging from -15 to -20°C . At about 1500, slightly colder (-18 to -25°C) cloud tops were evident in a narrow strip over the northeasternmost foothills of the Front Range on the western edge of the cold pool (See feature B, Fig. 8). It was apparent from the satellite and upper-air analyses that the light precipitation during this time period was probably driven by upslope flow (see surface analysis in Fig. 4d).

Nearby, apparently cloud-free areas (implied by surface and visible satellite data) were in the -5°C to -10°C temperature range. These values were consistent with shelter tem-

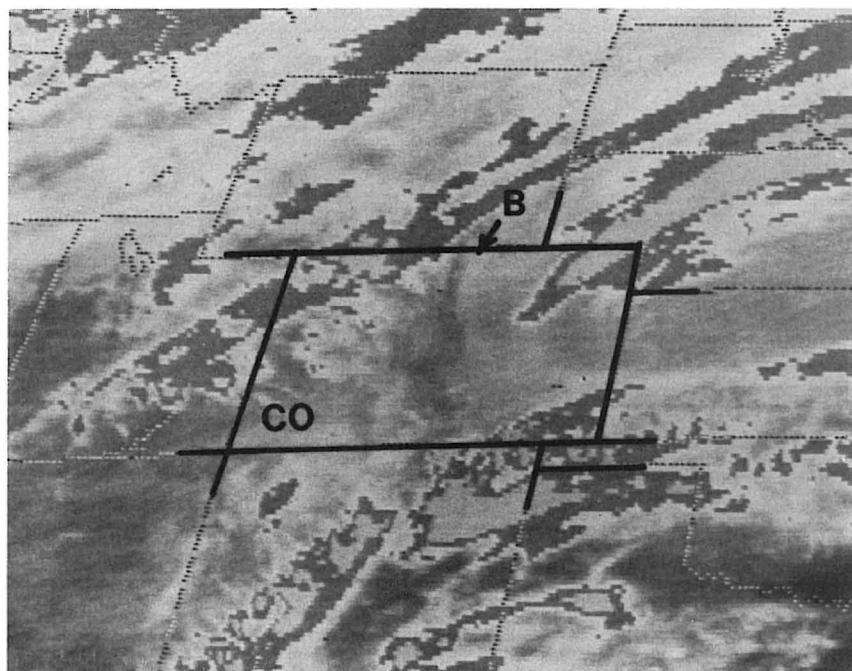


Fig. 8. GOES-IR image (from GOES east), 1501 UTC 1 February 1989. Note the cold streak of upper-level clouds to the east of the Front Range. Also, notice the bright band (B) on the western edge of the upslope cloudiness along the Colorado Front Range.

perature measurements, and illustrated that the GOES IR data were consistent with other data sources in this case.

c. 2000 1 FEB—0600 UTC 2 FEB 1989

The narrow north-south GOES IR bright band remained along the foothills (Fig. 9). Over the FCL region, CTT was approximately -20°C while the shelter temperature was -13°C ($+9^{\circ}\text{F}$). The bright band extended from CYS to DEN and followed the approximate shape of the 5000 foot elevation contour along the northern Front Range. As mentioned previously, this colder band was probably due to a bulge in the cold air resulting from damming. In the present case, easterly flow within the arctic air over the plains encountered both rising topography and a strong inversion as it moved toward the Front Range. This, in turn, led to blocking, a buildup of cold air over the foothills, and a heightened inversion on the western edge of the cold pool. Comparison of DEN and Dodge City, Kansas soundings during later periods revealed that the inversion and wind shift at the top of the arctic air mass consistently occurred at a slightly lower pressure at DEN. The inversion was also much more coherent (i.e. more abrupt at its base) at the DEN site. The satellite observations suggested that the eastward extent of the cold air bulge was only a few tens of mi.

Snow continued during this period, but intensity in FCL decreased mostly to the "trace" variety, and shelter temperatures remained about $+10^{\circ}\text{F}$ (-12.2°C). At 0000 UTC, the surface wind shifted from southeast to east at FCL and the snowfall intensified slightly at this time. NGM analyses at 0000 UTC (not shown) indicated an increasingly intense SLP gradient over the region (again, exaggerated by the SLP reduction). Deep cloudiness, however, was confined to the north and west of the northern Front Range. All of these factors, taken together, suggest that "upslope" was still the primary precipitation mechanism at work.

d. 0700–1200 2 FEB 1989

Light snow ($0.5'' (6\text{hr})^{-1}$) continued at FCL through the period. The area was situated along the southern or southeastern edge of a band of higher clouds (Fig. 10) which stretched from extreme southwestern Colorado into eastern Wyoming (where heavy snow was falling). While the primary precipitation mechanism was probably upslope, the presence of higher cloudiness indicated that some of the snowfall may have originated at higher levels. The NGM analyses in Fig. 11 continued to show the intense SLP gradient and strong westerly flow aloft in northeastern Colorado.

e. 1200–1800 UTC 2 FEB 1989

This period was generally marked by snowfall at a rate of $0.6'' \text{hr}^{-1}$ and warming shelter temperatures (from -8°F to -2°F (-22°C to -19°C) during the 6 hours). Snow intensity at FCL was heaviest from 1600–1800. NGM analyses (Fig. 11) suggested weak positive vorticity advection (PVA) at 500 mb beginning after 1200.

Visible (VIS) satellite data (available after 1430) confirmed that a band of higher clouds stretched from northwest Colorado, across the southeastern half of Wyoming, into South Dakota (Fig. 12). The cloud band appeared to be thick, and CYS received moderate to heavy snowfall during this time ($0.17''$ liquid). The southern edge of this cloud band was over FCL.

Estimates of satellite IR temperatures just southeast of the cloud band at 1700 yielded CTT values of approximately -20°C . Since shelter temperatures at 1700 were also approximately -20°C , the satellite was probably seeing surface-based radiation and, therefore, less upslope cloudiness existed during this time period. Weather and obstruction to visibility listed "light snow showers" at several reporting times, with an occasional "BINOVC" comment. The report-

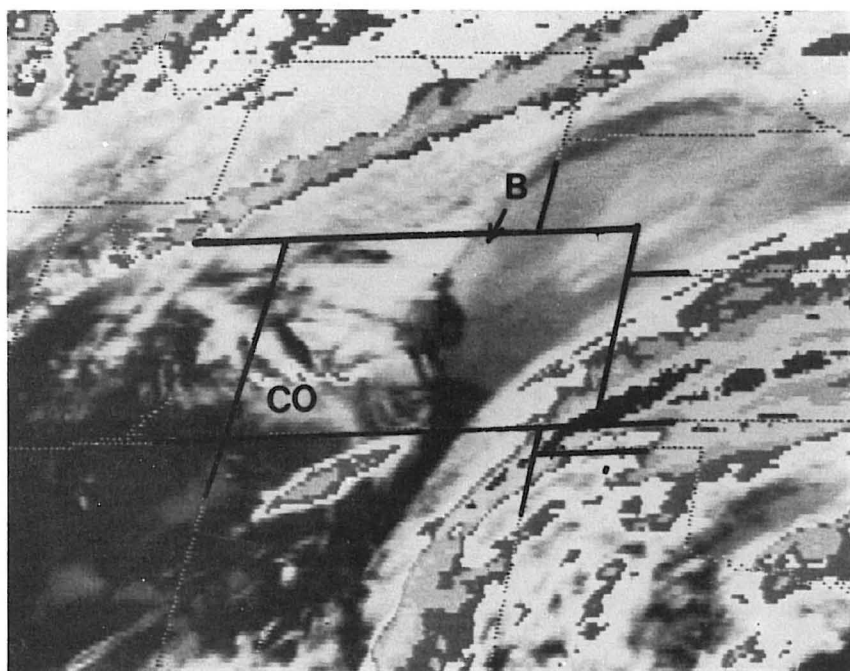


Fig. 9. As in Fig. 8, for 0531 UTC 2 February. The band of colder tops over extreme northeastern Colorado was moving rapidly eastward. The bright band (B) remained along foothills. (Note: difference in contrast between this IR image and that of Fig. 8 is due to the use of separate look-up tables.)

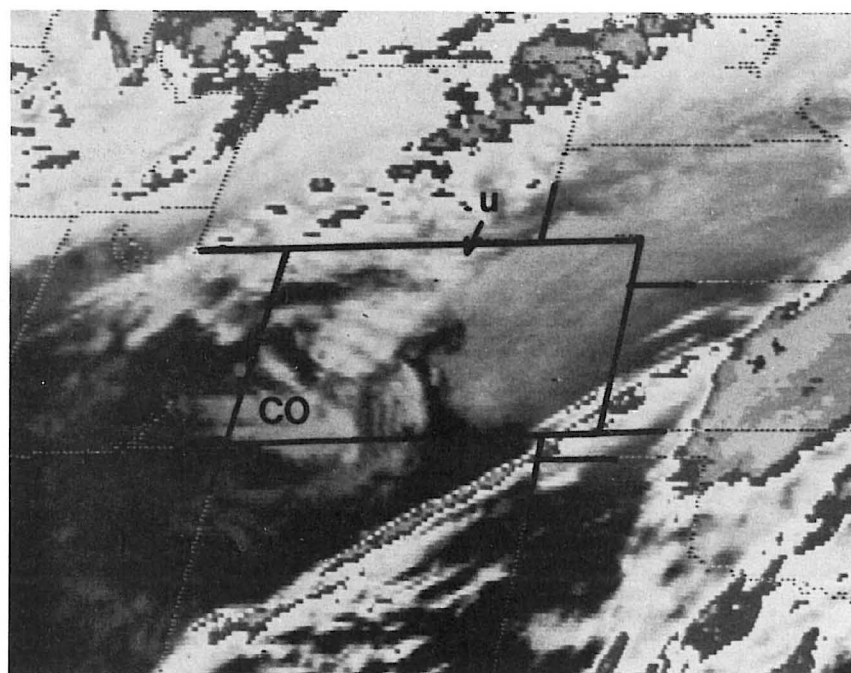


Fig. 10. As in Fig. 8, for 1031 2 February. Note band of higher clouds from southwestern Colorado into southeastern Wyoming (u).

ing station at Mountain States Weather Services (MSWS) in eastern Fort Collins recorded thin spots in the low overcast at 1700 and 1800. The observations for this period, taken together (particularly the decreased low-level cloudiness), imply that most or all of the light snowfall was occurring due to synoptic-scale ascent in the mid- and upper-levels of the atmosphere, (e.g., Donn, 1988).

f. 1800 UTC 2 FEB—0000 UTC 3 FEB 1989

Snowfall during this period amounted to only a trace in FCL. After 2200, the larger band of deep cloudiness began to move off slightly to the north of the FCL area, and the bright band along the western edge of the upslope cloudiness became less distinct. The visibility increased at FCL, while remarks include reports of "SWU SW-NW". Satellite data (not shown) suggested that a few scattered upper clouds remained over most of the region, but were thinner to the south. This was confirmed by the fact that DEN, Limon and Akron reported only fog (not snow). At 2330, the northern Front Range was not entirely free of higher clouds. However, the main deep cloud was again confined to regions north and northwest of FCL.

g. 0000–0700 UTC 3 FEB 1989

While the upper cloud cover increased only slightly during this period, the snow intensified substantially. The precipitation rate increased to $2.3'' (6\text{hr})^{-1}$ of dry, powdery snow, with much more snow reported along the western fringe of FCL. The liquid equivalent at FCL was reported only $0.05''$. However, this value may not have been representative of the precipitation pattern, since several observers in and near the western portions of the city reported more than $0.20''$ during this period. Again, precipitable water values in a column between the surface and the top of the cold air were approximately $0.02''$, based on the DEN sounding (Fig. 5a). Also, a faint 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 probably associated with a secondary surge of cold low-level air. The DEN wind profiler data during this period indicated an increase in the depth of the upslope flow, beginning at 2300 and lasting for about three hours (not shown). The cold surge was measured in the profiler data as a marked increase in strength of low-level upslope winds. The PROFS mesonet station (FOR) in northwestern Fort Collins measured rapidly rising surface pressure (about $4.5\text{ mb } (3\text{hr})^{-1}$) beginning at 2130. This rise suggested accumulation of cold air near the foothills, which was accompanied by a temperature drop of $8^{\circ}\text{F } (4.4^{\circ}\text{C})$ in 3 hr and strengthening northeasterly surface winds (note: FOR is approximately 3.8 mi west-northwest of FCL, and about 100' higher in elevation).

Deepening of the arctic air was also accompanied by westward propagation of the front to higher elevations. At a volunteer observer site (Coal Creek) in the foothills to the west of DEN (about 730 mb), frontal passage occurred at around 2200 UTC and the temperature fell more than $40^{\circ}\text{F } (22^{\circ}\text{C})$ in a few hours.

As a result of the secondary cold surge, FCL shelter temperature fell from -2°F to $-17^{\circ}\text{F } (-19\text{ to } -27^{\circ}\text{C})$ in just a few hours. Some of these factors (cold low-level surge, little increase in upper cloudiness, increased snowfall rate) implied that the dynamics for the snow was created by upslope flow within the cold pool. However, sounding data indicated that the source of moisture must have been located above the cold air mass, and profiler and satellite measurements supported this possibility. NGM analyses at 0000 UTC did not indicate any significant synoptic forcing of upward motion (i.e. PVA or warm advection above the cold pool), but did specify relative humidities above 50% in the surface-to-500mb layer across northern Colorado and Utah. The evidence supports the role of the deepening cold air mass in enhancing snowfall along the foothills during this period.

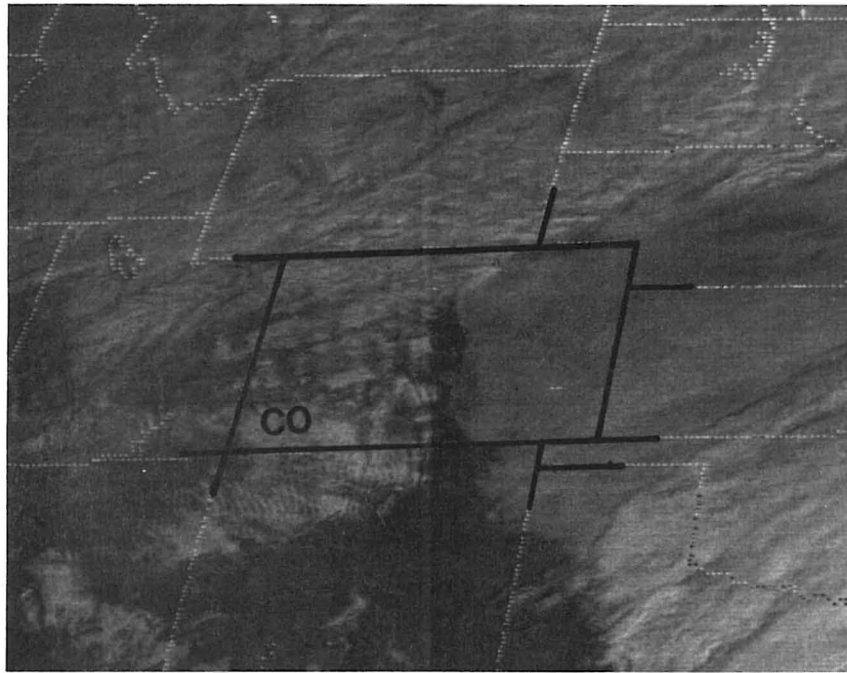


Fig. 12. GOES-VIS, 1631 UTC 2 February. Deep cloudiness was located to the north and west of the Colorado Front Range at this time.

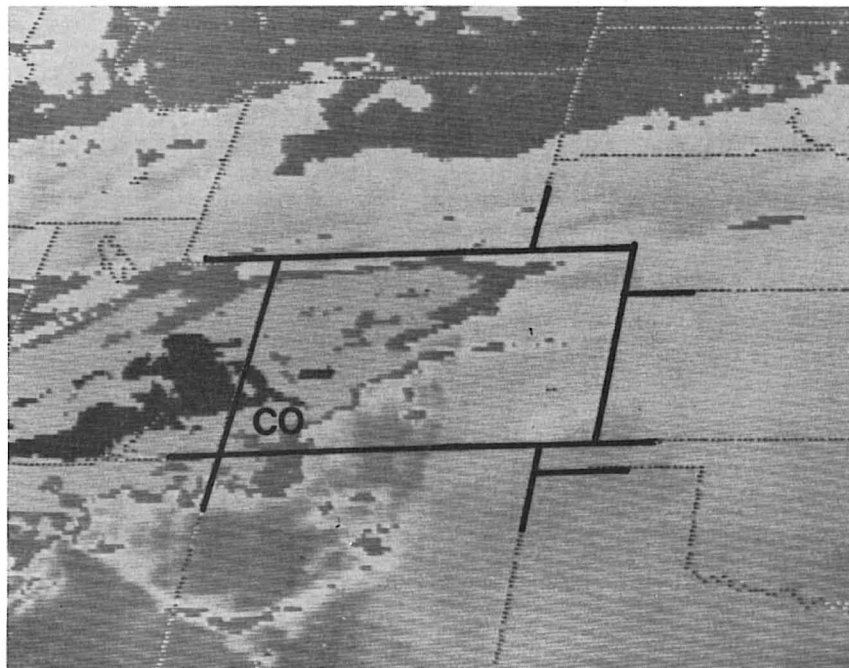


Fig. 13. As in Fig. 8, for 1201 UTC 3 February. Very cold cloud tops were located from southern Utah into western Colorado.

Utah and into western Colorado. In northeastern Colorado, the intense SLP gradient persisted, and relative humidity in the 850-500 mb layer had begun to increase. This is also readily apparent in Fig. 5a-c for the 700-400 mb layer.

i. 1200–1800 UTC 3 FEB 1989

This segment was characterized by light snow at accumulation rates of $1\text{--}2'' (6\text{ hr})^{-1}$. Overall during this period, a large

mass of very cold clouds (associated with the shortwave) began to organize over southern UT, then move into Colorado (Fig. 13). However, as the cloud mass moved into the Colorado Rockies it appeared to begin to break up again. This was accompanied by cloud-top warming. Over FCL this breakup was quite evident in both VIS and IR satellite data (not shown). At 1200, FCL reported moderate snow. CTT's of approximately -30°C were found overhead. The snow

was light at 1500 under a break in the cold cloud. At 1600, the breaks were more apparent, and the snowfall was even lighter. CYS reported about 0.02 hr^{-1} precipitation (liquid equivalent) from 1300 to 1700, with moderate snow and deep cloud coverage at 1630. At 1700, a marked split in upper level cloud was evident along the Wyoming/Colorado border north of FCL. The last two hours of the period found the visibility increasing. FCL reported "BINOVC E" and "PCPN VRY LGT" at 1800.

j. 1800 UTC 3 FEB—0000 UTC 4 FEB 1989

The upper cloud deck over western and north-central Colorado continued to appear fragmented on both VIS and IR

imagery. NGM analyses indicated a correlation of the breakup in Colorado with negative vorticity advection (NVA) (Fig. 14a). This lasted until about 2000, at which time the cloud mass began to fill in and expand again toward the Front Range. The apparent intensification in northeastern Colorado may have been in response to a weak-to-moderate shortwave ($12 \times 10^{-5} \text{ sec}^{-1}$ absolute vorticity) which was entering west-central Colorado by 0000 UTC. It also coincided with the arrival of a very moist, cloudy air mass in the 700–500 mb layer, which had exhibited lower relative humidity and dew points until this time (Fig. 5b–c). At 2100, the coldest IR tops were just barely west of FCL (Fig. 15). Over and south of DEN, where few high clouds were evident,

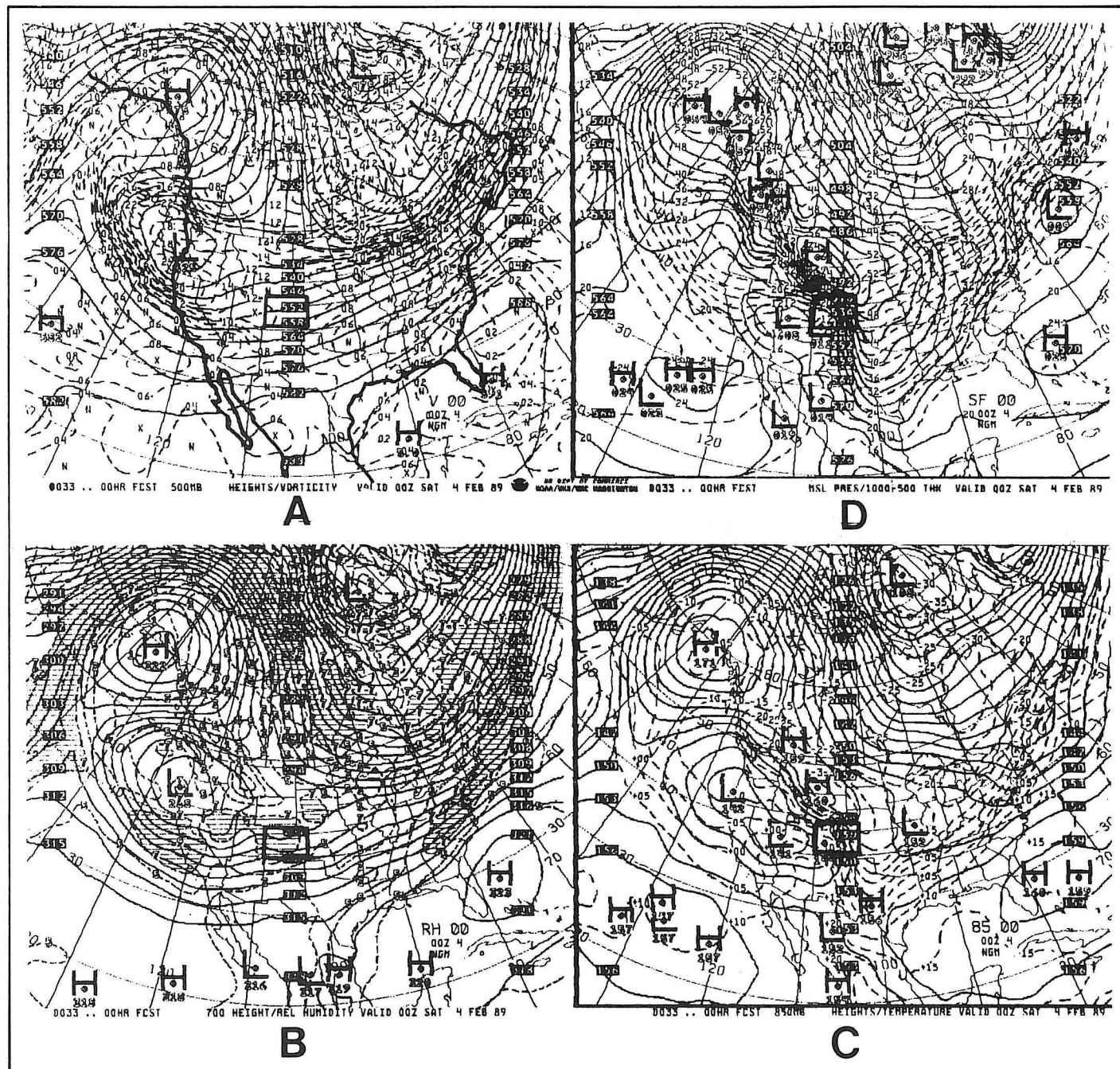


Fig. 14. As in Fig. 4, for 0000 UTC 4 February.

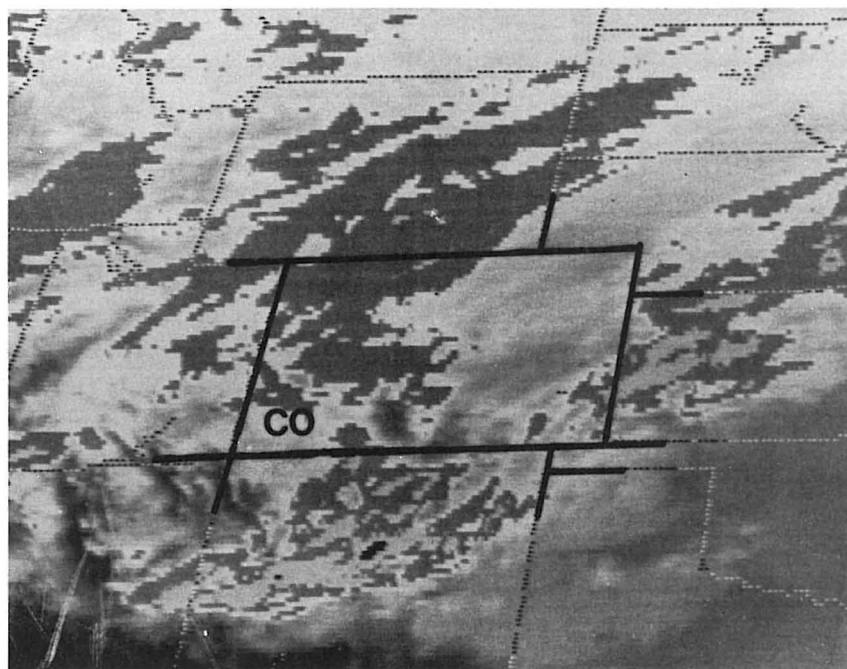


Fig. 15. As in Fig. 8, for 2101 UTC 3 February.

snow occurred in only trace amounts. At 2200, a marked boundary between deep, cold cloud to the northwest and low cloud to the southeast was apparent from approximately CYS to LXV.

The snow intensity at FCL increased dramatically during the latter part of the period (beginning at 2000), with visibility decreasing to 0.5 mi or less and weather remarks at both the 2300 and 0000 times reporting heavy snow (S+). At FOR, where precipitation rates dramatically increased at 2215 (from trace amounts to $0.01\text{--}0.02''\text{ hr}^{-1}$), pressure began to rise at 2100 after a rapid fall, possibly indicating some deepening of the cold pool. The dramatic increase in snowfall intensity at FCL beginning at 2200 was accompanied by a marked increase in high cloud coverage over the Front Range at approximately the same time. The cloud increase was especially notable between FCL and LXV. Much lower precipitation amounts were measured to the east and northeast of FCL, however. DEN reported little or no increase in snowfall rate, and the deeper cloudiness did not arrive until about 0630. The strong southeasterly SLP gradient continued at 0000 UTC on the NGM analyses (Fig. 14d).

k. 0000 UTC—1200 UTC 4 FEB 1989

During this period the area of cold cloud tops over Colorado expanded to cover most of the state. The coldest tops were in the west and southwest, and it was there where maximum snow accumulations occurred. The following 24-hour liquid precipitation totals were recorded: Glenwood Springs 1.1", Rifle 0.74", Winter Park 0.67", Grand Junction 0.59", and Estes Park 0.51". Interstate 70 at Vail Pass closed periodically overnight and numerous vehicles were stranded to the west. There were 24" of snowfall reported at Beaver Creek ski resort (just west of Vail) for the 24 hours ending at 1200 UTC. A significant shortwave crossed the area during the period, with NVA not evident until the very end of the period along the Front Range (see Fig. 16).

At 0000 UTC 4 February, a southwest-northeast band of very cold tops (-35°C) through northeastern Colorado was

clearly identified. By 0100, FCL was again under deep cloud coverage (Fig. 17). At this time, very low visibilities existed along the northern Front Range, caused by moderate to heavy snowfall. CYS reported heavy snow at 0400 and 0500, with liquid precipitation rates at about $0.02''\text{ hr}^{-1}$. At 0831, 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 1000 to 1200, just as the deep clouds had begun to spread over this area. Until this time, only trace snow amounts were recorded there.

Beginning at 0200, DEN profiler data indicated significant heightening of the highly sheared region at the top of the cold pool. As shown in Fig. 18, the depth of the upslope flow increased by several hundred m in about 9 hours. After about 1100, snow intensity decreased at FCL. The city received a total of 3.6" of snow overnight ($0.25''$ liquid); about 0.7" fell at DEN. In this case, FOR surface pressure rises did not occur during heavy snowfall. DEN soundings at 0000 and 1200 UTC 4 February (Fig. 5c-d) revealed a mixing ratio of about 0.3 g kg^{-1} in the cold pool. However, the layer between 700 mb and 500 mb (above the inversion) contained about 1.5 g kg^{-1} . This fact, combined with the observation that heavy precipitation began as colder IR CTT's developed and spread eastward, was further evidence for the moisture source being advection of Pacific air from the west, over the cold air. The satellite, precipitation, pressure and sounding data along with the NGM analyses indicated that synoptic forcing was providing widespread light snowfall east of the Continental Divide during this period, with a continued enhancement in and near the foothills along the northern Front Range.

l. 1200–1800 UTC 4 FEB 1989

Coldest tops during this period were to the south and southeast of FCL (Fig. 19). Upper clouds continued to dominate most of Colorado, but a few breaks developed in the coldest cloud top regions. At about 1500 UTC, the visibility

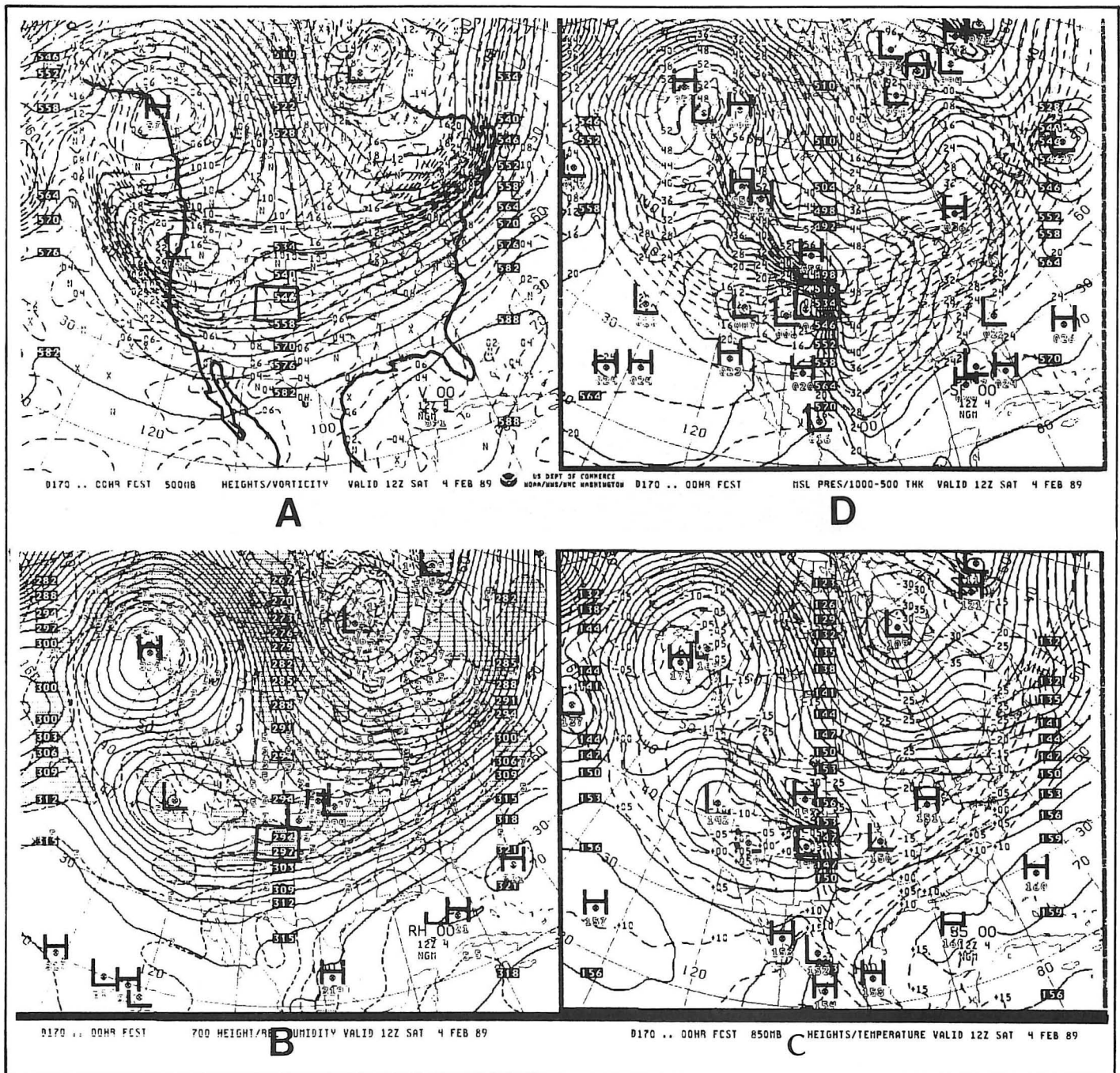


Fig. 16. As in Fig. 4, for 1200 UTC 4 February.

at FCL rapidly increased from less than 1 mi to 4 mi or greater. The snow total for this 6 hr period was 0.6" in FCL, with about 0.01" hr^{-1} liquid rate until around 1500 and trace amounts afterwards. CYS reported 0.03" between 1100 and 1200 with some broken deep cloudiness overhead. Precipitation rates thereafter decreased to only a T hr^{-1} , as the upper cloud dissipated.

The NGM initial analysis at 1200 UTC (Fig. 16) showed that one shortwave trough was exiting the northeast Colorado region, while another, moderately strong wave was just entering eastern Utah. At the same time, the 700 mb analysis showed lower relative humidity in the surface-to-500 mb

layer behind the first wave. There were warmer cloud tops over FCL at 1500 than at any time since before 0700. This was correlated well with the decrease in snowfall. From 1500 to 1800, the high cloud coverage remained about the same (i.e. broken).

m. 1800 UTC 4 FEB—0000 UTC 5 FEB 1989

Generally, snowfall was light in FCL during most of this period, with no deep cloud coverage. The small snowfall amounts (0.2") were probably the result of weak upslope flow. However, the upper clouds associated with the Utah wave began to fill-in over northeastern Utah and northwest-

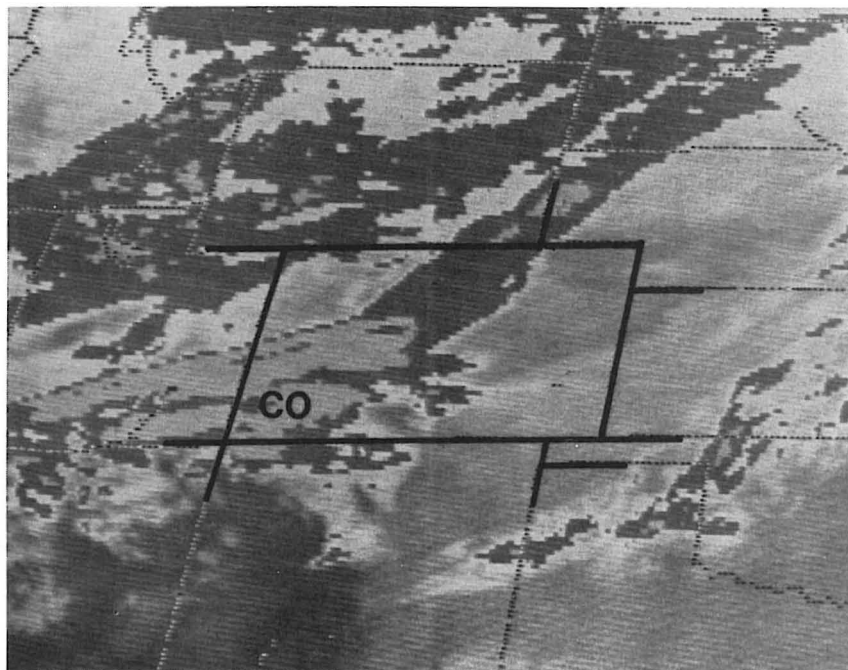


Fig. 17. As in Fig. 8, for 0101 UTC 5 February. The northern Front Range was under deep cloud coverage at this time.

ern Colorado. With time, the mass of cold clouds grew larger and moved east-northeastward. By about 2200, the effects had just reached the FCL area. Visibility began to drop again, and snowfall intensity increased significantly very late in the period.

The evening NGM analyses (not shown) indicated that the large mass of cold clouds was, indeed, associated with the

Utah shortwave. The 850–500 mb relative humidity increased again southwest of FCL. At this time, the deep, cold cloud mass associated with the strong shortwave was poised just west of FCL, stretching from south-central Wyoming to west-central Colorado. It was expanding and deepening, most notably in the northwestern portion of the state.

n. 0000–0600 UTC 5 FEB 1989

During this 6 hr period the FCL snow intensity increased dramatically, apparently due to large-scale dynamical forcing. The cold cloud mass associated with the Utah shortwave persisted and continued to move east-northeastward across northern Colorado and southern Wyoming. Once again, DEN profiler data (not shown) indicated some deepening of the sheared zone at the top of the cold pool beginning at 2300. Surface pressure at FOR was approximately steady as the snowfall rate increased. Visibility was fairly constant (1 mi. or less) until 0600 when observations at FCL indicated both a sudden increase in visibility and the beginnings of a sharp decrease in the snowfall rate (to about $\frac{1}{3}$ of what it had been). However, the only apparent change in the upper cloud field at this time was slight warming at cloud top. The coldest tops were found over southeastern Wyoming and exhibited a comma-shaped appearance. CYS reported moderate to heavy snowfall from 0200 to 0600, while DEN received light to moderate snowfall from 2300 to 0800. Measurable snow continued at both locations until approximately 1200.

According to the DEN sounding at 0000 UTC, easterlies in the lowest levels were no longer present. The 0000 UTC NGM analysis indicated that the SLP gradient strength was rapidly decreasing over northeastern Colorado, although PVA aloft was still fairly strong.

o. 0600 UTC–2100 UTC 5 FEB 1989

The upper cold cloud mass moved out of the area, visibility increased to 15 mi or greater and the upslope ended (surface winds backed from easterly to northerly by 0600, then

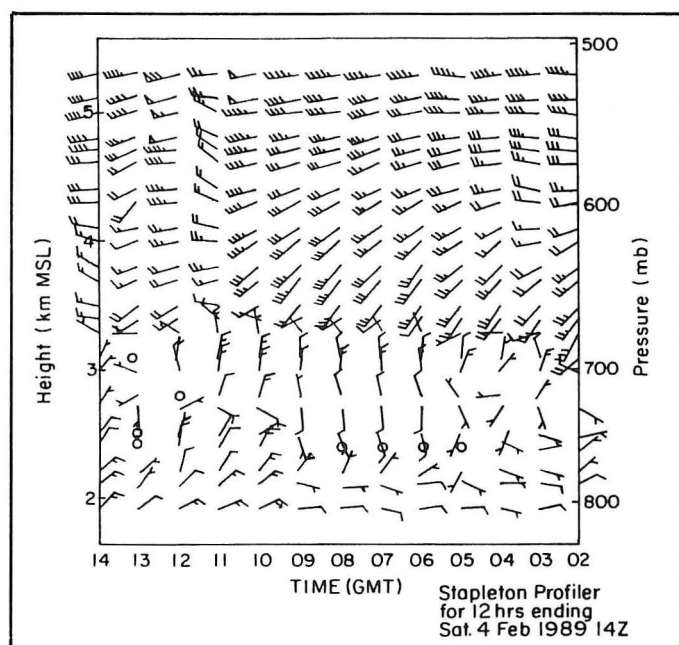


Fig. 18. A series of profiler-derived winds at DEN for 12 hours beginning at 0200 UTC 4 February and ending at 1400 UTC 4 February. Units for wind barbs are standard notation in knots. Circles denote calm winds. Note that time increases from right to left.

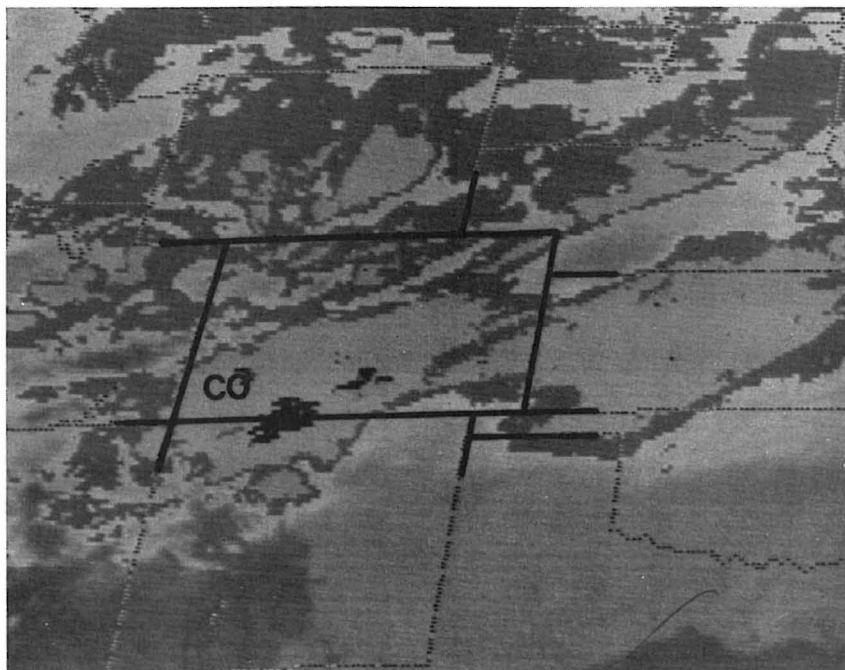


Fig. 19. As in Fig. 8, for 1201 UTC 4 February. Most of Colorado was covered by deep cloudiness, with some breakup evident over the northern portions of the state.

became light and variable). By midnight, cloud top temperatures were increasing rapidly over the area of interest. Correspondingly, this time period marked the last of the snowfall for the storm along the northern Front Range. There was only very light snowfall reported at FCL, CYS and DEN. The FCL temperature began a steady climb from -20°F to $+1^{\circ}\text{F}$ (-29 to -17°C), the first time the mercury had been above 0°F (-18°C) in 83 hours. By 1930, cloud cover was negligible in most of the mountainous regions to the west of FCL. NGM SLP analyses indicated that the intense pressure gradient over the region was no longer present. Upper levels were apparently subsident, and the wind direction at 700 mb had switched to northeasterly.

4. Microphysical Considerations

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 (during the periods g, h, k and n described in the previous section). Several meteorologists in the Boulder area also reported primarily dendritic aggregates during moderate snowfall on both 3 and 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 low-level cold air throughout the storm (Fig. 5; also see Pruppacher and Klett, 1980). The moisture source apparently resided in the moist westerly flow aloft, which exhibited -10 to -20°C temperatures, a regime favorable for dendritic crystal growth. Such

evidence is strongly supportive of the precipitation scenario described previously (see Fig. 2).

5. Simulations of the cold pool evolution

In order to further investigate the dynamics of this storm, 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). The numerical model allowed some preliminary tests of the dynamical hypotheses discussed in Sections 1–3. 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 meso-scale 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 (see Fig. 5a). Using this initialization, the cold air is confined to elevations below about 2.8 km MSL east of the barrier. The location of the east-west domain is approximately the latitude of DEN.

Figure 20 shows 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 heightening of the temperature surfaces over the sloping terrain east of the mountain barrier at elevations of 3.1 to 3.3 km MSL. As previously discussed in Section 3, a persistent bright band in the upslope cloud tops was observed along the foothills region for many hours during the storm. 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 associated vertical motion field predicted by the model

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

Model category	Option
initialization	horiz. homogeneous, DEN sounding 0000 UTC 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, 100 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 sec

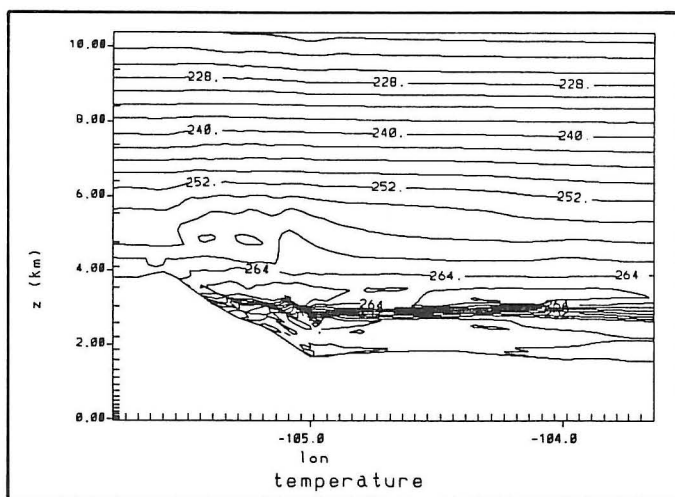


Fig. 20. Model-predicted temperature field ($^{\circ}\text{K}$) over the Front Range region at three hours of simulation. Latitude of east-west domain is that of DEN. Region of close packing of isotherms is the inversion at the top of the arctic air mass. Temperatures within the cold pool are 240 to 250 $^{\circ}\text{K}$.

is shown in Fig. 21. Immediately evident are the deep, terrain-induced gravity waves produced as the strong westerly flow crosses the mountain barrier. However, downward motion over the sloping foothills to the east, created by the 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 region below the inversion. This is also consistent with the blocking scenario described previously.

In a separate sensitivity experiment, the cold pool was removed from the initialization. The original sounding (Fig. 5a) was modified to exhibit a moist adiabatic lapse rate and moderate southwesterly flow between the surface and 700 mb. Fig. 22 shows the resulting w field at three hours of simulation. The area of strongest downward motion just east of the barrier crest is larger in horizontal spatial extent and

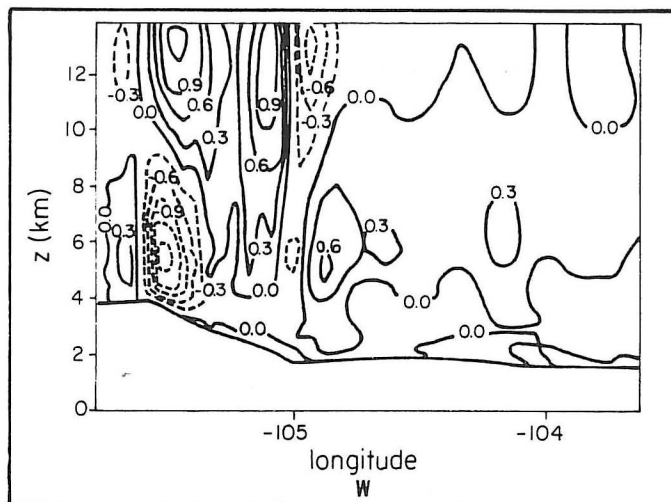


Fig. 21. Model-predicted w field (m sec^{-1}) over the Front Range region at three hours of simulation. Dotted lines are contours of negative w . Strong downslope is confined to elevations above the arctic air mass. Maximum updraft speed for the cross section shown is about 1.2 m sec^{-1} , while maximum downdraft is 1.6 m sec^{-1} . Contour interval is 0.3 m sec^{-1} .

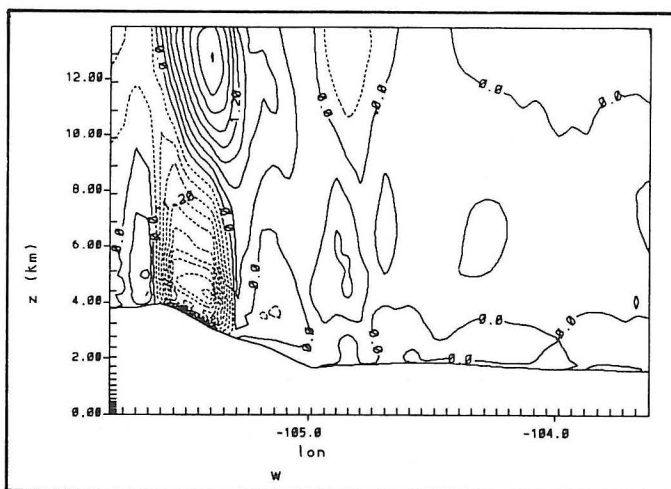


Fig. 22. As in Fig. 21, for sensitivity simulation. Model initialization does not include the cold pool. Maximum updraft speed for the cross section shown is about 2.4 m sec^{-1} , while maximum downdraft is 3.0 m sec^{-1} . Note the absence of gravity wave structure aloft well east of the barrier.

more intense (maximum -3 m sec^{-1} , compared to -1.5 m sec^{-1} in Fig. 21). Downslope flow has propagated well down the east slope, into the foothills. Over the easternmost foothills, an area of moderate downward motion at 3.5 km MSL is evident; this is not present in Fig. 21.

Further comparison of Figs. 21 and 22 reveals a large discrepancy in the w structure over the area of interest in this study (i.e. the extreme eastern portion of the steep terrain). The simulation with the cold pool (Fig. 21) exhibits deep gravity wave structure above the bulge on the western edge of the cold pool, while the sensitivity run (Fig. 22) produces only weak vertical motion in this region. In fact, an ascending portion of the gravity wave in Fig. 21 is located directly over the western edge of the arctic air mass, with

upward motion exceeding 1 m sec^{-1} . The location of the strongest ascent agrees qualitatively with the location of the snowfall enhancement discussed in Sections 1-4.

These simulations provide further confirmation that the cold pool indeed may extend deeper into the atmosphere along the eastern foothills of the Rocky Mountains during strong cold air outbreaks. This vertical extension provides additional retardation of downslope flow associated with strong westerly winds aloft, and apparently generates a deep region of ascent over the western portion of the cold pool.

6. Conclusions

The case study of the 1-5 February 1989 arctic outbreak has provided evidence of a unique precipitation enhancement process induced by blocking along the east slope of the Rocky Mountains. The combination of topography with an extremely cold low-level air mass produced heavy precipitation along the Front Range despite strong mid-level westerly winds. From this 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 low-level geostrophic easterly component increases with time.
- Within the cold pool, mass (and, therefore, surface pressure) increases along the foothills as the cold air is unable to ascend the steep topography.
- The cold pool, as a result, deepens locally on its western edge due to the upslope flow and blocking processes.
- Enhanced condensation (or reduced evaporation or sublimation) occurs above the cold pool, as the moist westerly flow aloft overruns the deepening cold dome. These precipitation particles fall through the western portion of the cold air to the ground.

The deeper cold air acts as a barrier to the westerly flow and prevents downslope. In addition, enhanced ascent may occur over the western edge of the cold pool and extend into the upper troposphere. Observations of crystal habits in this storm confirmed that the moisture source during significant snowfall was the relatively warm westerly flow aloft.

Two-dimensional mesoscale model simulations capture the development of overrunning and a bulge on the western edge of the cold pool. Generally, strong downslope flow east of the mountains in the low levels is retarded by the arctic air mass. The presence of the heightened inversion also leads to the formation of deep tropospheric ascent above this region.

The implications of this study to the wintertime forecaster are important. The dynamical interaction of the upslope flow with moisture aloft can be critical to understanding the snowfall distributions associated with arctic outbreaks along the east slopes of the Rocky Mountains.

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