Contrasting Meteorological Conditions Associated with Winter Storms at Denver and Colorado Springs

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ABSTRACT

Case studies of heavy snowstorms at Denver and Colorado Springs, Colorado, indicate that they occur under different meteorological conditions. The authors examine the hypothesis that there are in fact fundamental differences between the synoptic evolution of events in these two storm types by compositing a total of 28 cases, 17 (11) of which are defined as heavy snowstorms (at least 20 cm of snowfall) at Denver (Colorado Springs). These composited fields were constructed using data at three times in the history of each case. Results show distinct differences in the composited synoptic evolution of the two groups. At low levels the Denver composite shows low static stabilities, warm advection, and high values of potential temperature in the lee of the Rockies. The Colorado Springs composite, on the other hand, shows cold, stable air and cold advection in the lee. At upper levels an eastward-progressing short-wave trough is found at different longitudes in the two composites.

The implied interaction between lower and upper levels of the two composites is also very different. For the Denver composite, the trajectory of the upper-level trough brings it close to the area of low static stability and high surface potential temperature at low levels. This implies strong interaction between the upper-level system and the warm unstable air at low levels and dramatic cyclogenesis east of the Rocky Mountains, typically in southeast Colorado. In contrast, the upper short-wave trough in the Colorado Springs composite is farther north, and a layer of cool stable air is found on the High Plains of Colorado. Not surprisingly, surface cyclogenesis is notably weaker in this composite. These conclusions, substantiated by inspection of the individual cases, have obvious implications for predicting the location of heavy snow along the Front Range of Colorado.

1. Introduction

On 8 March 1992 the Denver area and the northern Colorado Front Range experienced a paralyzing winter storm that closed all roads leading into and out of the northern part of the state. The event started at Denver on the afternoon of 8 March with rain, hail, and lightning and was immediately followed by moderate to heavy blowing and drifting snow. Strong winds of 15–20 m s^{-1} (30–40 kt) with gusts to 25 m s^{-1} (48 kt) produced blizzard conditions that closed highways and created numerous power outages. These severe conditions persisted for 13 h and finally produced 31.5 cm of snow at Denver (50 cm in many areas to the north and west of the city). At Colorado Springs, 100 km to the south, only 3 cm of snow fell during the entire period.

The 26 February 1987 winter storm, on the other hand, immobilized Colorado Springs and the southern part of the state. Early in the day, light snow and weak southerly winds were reported. However, these light surface winds soon gave way to 10 m s^{-1} (20 kt) northwesterly winds accompanied by a strong surge of cold air and moderate snowfall. These severe conditions at Colorado Springs continued for 8 h producing 38 cm of snow. During the same period, only 8 cm accumulated at Denver.

These winter storms are two examples of events that produce very heavy snowfall and severe conditions typically once or twice a year along the Front Range of Colorado. Although these storms are synoptic in scale, the snowfall patterns within them are controlled by the interaction of the synoptic-scale flow with the terrain features of eastern Colorado. For this reason forecasting the distribution of precipitation in these storms poses a major challenge for the operational meteorologist.

The principal motivation for this study is to examine forecasting implications of these two storm types. The intent here is to demonstrate, first, that specific storm characteristics are produced by the interaction between the different meteorological synoptic-scale flows and the mesoscale terrain at Denver and Colorado Springs and, second, that these characteristics are, in principle, predictable.

Section 2 consists of a review of relevant concepts on extratropical cyclones, lee cyclogenesis, and the role terrain plays in the development of these cyclones. In section 3, we describe the data and methodology used in the case studies and composite analyses. Systematic and repeatable differences in the characteristics of the

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individual cases are described in section 4. The composite analyses of 17 Denver storms and 11 Colorado Springs storms are presented in section 5, and in section 6 suggestions are made on how these systematic features can be used by the operational community to help forecast where Front Range winter storms will produce the heaviest snowfall.

2. Background

A variety of conceptual models are commonly used to predict heavy snowstorms in Colorado (e.g., Barnes and Colman 1993). Weismueller (1984) describes three conceptual models associated with heavy snow somewhere along the Front Range: the California Cutoff, Arctic Trough, and Eastern Pacific Developmental Trough (EPDT), subjectively identified by the evolution of the flow at 500 mb. The EPDT and the Arctic Trough are similar to the fully developed cyclone and the shallow anticyclone models, respectively, described by Boatman and Reinking (1984) and Dunn (1987). These authors define their models by the depth of upslope produced along the Front Range of Colorado. In their view, the fully developed cyclone accompanied by deep upslope often produces blizzard conditions at Denver, while the shallow anticyclone associated with shallow upslope provides Denver with only light snow.

The Denver storm example (8–9 March 1992, to be discussed later) developed from a closed upper cyclone that was similar to the EPDT and displayed deep upslope comparable to the fully developed cyclone model of Boatman and Reinking (1984). The Colorado Springs case (26 February 1987), however, developed from a slow-moving upper-level short-wave trough, and surface development was strongly influenced by the presence of stable arctic air at low levels.

Cyclone development in the lee of major mountain ranges has received much attention over the years. Whittaker and Horn (1981, 1984), for instance, show that there exists a strong climatological tendency for lee cyclogenesis along the eastern slopes of the Rocky Mountains in Colorado, a tendency they relate to the large change in elevation between the mountains and the plains to the east. They invoke the traditional explanation using the conservative nature of potential vorticity. Although the potential vorticity of the parent cyclone is conserved as it crosses the ridge, the relative vorticity is temporarily reduced or eliminated by vortex shortening while the cyclone is over the mountains. Upon leaving the mountains, air parcels in the cyclone are stretched to their original length and the cyclone reappears (Smith 1979; Holton 1979).

The formation of lee cyclones has been separated into two phases by many theoreticians: the initial appearance and formation of the cyclone and its subsequent growth. In the rapid development phase, as referred to by Buzzi and Tibaldi (1978) and McGinley (1982), the lee cyclone formation is strongly dependent on orographic influences, whereas in the subsequent baroclinic development phase the cyclone taps into the synoptic-scale potential energy reservoir.

These arguments can be put into a "potential-vorticity (PV) thinking" context as follows. The basic viewpoint of PV thinking is that cyclogenesis occurs as a result of the superposition of a mobile PV maximum over a surface warm anomaly. In the case of lee cyclogenesis the warm anomaly is associated with so-called lee troughing. To the lee of the Rocky Mountains, lee troughing results when cross-mountain flow is partially blocked, that is, when the potentially coolest air, near the surface, is unable to cross the Rockies. The air arriving at the surface on the lee side is thus warmer than at the same elevation west of the mountains, resulting hydrostatically in lower pressure in the lee. This is the first phase of lee cyclogenesis. The second, or baroclinic-development phase, then occurs as the upper PV center moves toward this surface warm anomaly. The resulting superposition can then lead to mutual amplification of the upper and lower systems as is typical in cyclogenesis away from the mountains.

Using idealized model simulations, Snook (1993) finds that lee vortices do develop near the flank of a mountain range like the Colorado Rockies. In particular, he finds that with westerly flow across a north–south ridge, a cyclonic vortex forms downstream from the south end. These results may explain the greater frequency of lee cyclogenesis in southeast Colorado near the southern end of the highest terrain.

On a more local scale, the terrain east of the Continental Divide in Colorado plays a major role in the evolution of the synoptic-scale regimes. There are three prominent east–west ridges in eastern Colorado (Fig. 1): Cheyenne Ridge, Palmer Divide, and Raton Mesa (on the Colorado–New Mexico border). Denver and Colorado Springs are situated on opposite sides of the Palmer Divide. Thus, flow at low levels from the northeast is upslope for Denver but at least weakly downslope for Colorado Springs, whereas flow from the southeast is upslope for Colorado Springs but downslope for Denver. These ridges also affect advancing cold fronts. Young and Johnson (1984) discuss the influence of the Cheyenne Ridge on a southward-progressing shallow cold front. They observe that the front east of the ridge proceeds southward unimpeded, while the front to the west is delayed. As a result, the cold air mass enters the South Platte River drainage basin (to the north and east of Denver on Fig. 1), around the end of the ridge, producing a frontal passage from the east and southeast at stations near the foothills of the Front Range immediately to the south of the Cheyenne Ridge. This sequence of events is also observed at Colorado Springs, where shallow fronts are blocked by the Palmer Divide. The cold air associated with the frontal passage enters Colorado Springs either as northerlies over the Palmer Divide (if deep enough) or as more southeasterly flow.

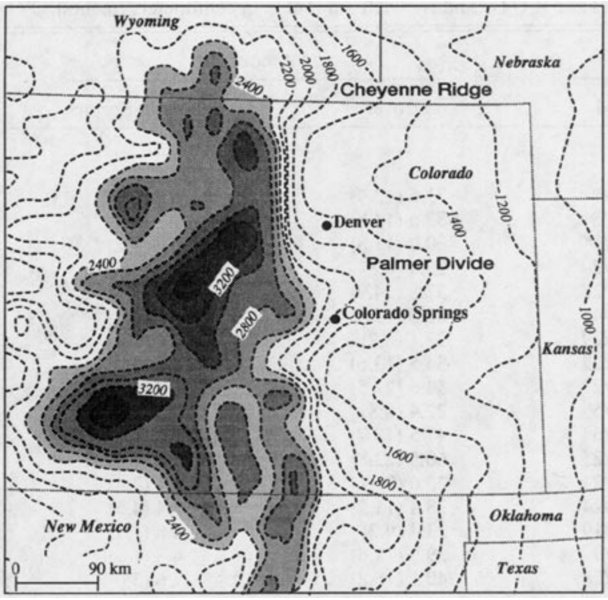

FIG. 1. Terrain features that influence snowstorms at Denver and Colorado Springs. Smoothed elevation contours are at 200-m intervals.

Many aspects of the evolution of the Denver storms are consistent with the more current theories concerning lee cyclogenesis presented above. The following sections will show that the location of low-level warm, unstable air in the initial stages of development, the intrusion of low-level cold air at the time of development, and the superposition of the lower- and upper-level cyclonic features are crucial for cyclogenesis. On the other hand, for the Colorado Springs storms, upper-air features are weak and less clearly defined, and the interaction between lower and upper levels is less evident.

3. Data and methodology

We have taken both a compositing and a case study approach in the analysis of a set of heavy snow events at Denver and Colorado Springs. To be considered a heavy snowfall event, storms were required to produce at least 20 cm of snow at Denver or Colorado Springs. Additionally, the location experiencing the heavy snow must have received at least twice as much snowfall as the other; this restriction eliminates cases that produced heavy snow at both locations. During the period between 1972 and 1992, 28 storms met these criteria, 17 of them producing heaviest snow at Denver and 11 at Colorado Springs. Nine events produced heavy snow at both cities but were excluded from this study since they did not meet the doubled snowfall criterion. Several of these "overlap" cases contain features similar to those found in either the Denver or the Colorado Springs composite, while some cases display characteristics of both. All cases were determined by inspection of MF10A/B hourly observation forms at Denver and Colorado Springs. In addition, three-hourly surface maps and synoptic charts analyzed by the National Meteorological Center were obtained from the National Climatic Data Center. These maps were subjectively analyzed for pertinent features, including fronts, surface winds, and stability patterns.

a. Case studies

Quantitative analyses were produced for one Denver and one Colorado Springs case using upper-air data for all stations in the contiguous United States. These data, obtained from the Forecast Systems Laboratory's rawinsonde archive, have passed quality control standards and hydrostatic consistency checks (Schwartz and Govett 1992). Analyses were performed on both the case studies and the composite results using the GEMPAK (General Meteorology Package) routines (des Jardins et al. 1991) to depict heights of mandatory pressure surfaces, temperatures, winds, and 850–700-mb thermodynamic stability. The objective analysis applied in GEMPAK consists of a Barnes scheme wherein each pass of the analysis interpolates data from stations to a 2.4° latitude and longitude grid spacing using a distance-related weighting function (Koch et al. 1983). The actual winds were objectively analyzed at 50-mb intervals and plotted on a GEMPAK grid. Surface station information, fronts, and pressure centers, unavailable from the GEMPAK analyses, were transferred from the NMC analyses to the GEMPAK-generated figures to illustrate important features in each case study.

b. Composite techniques

Two composites were constructed, one for Denver cases and one for Colorado Springs cases. The selection of storm cases was based solely on snowfall records. Although sample sizes are relatively small, we believe that the differences in composites shown in this paper are not primarily due to sampling effects. Salient information pertaining to these storms is provided in Table 1. Each storm was subjectively inspected to determine the primary characteristics involved in its development. Although the majority of cases displayed similar features, a few cases did not fit our conceptual model. For example, the Christmas 1987 storm (described by Barnes and Colman 1993) followed neither the developing baroclinic wave model nor the typical pattern of lee cyclogenesis. However, it is included as one of the Denver storm composite since it met the previously established snowfall criteria. For additional information about these cases see Mahoney (1992).

Time periods or stages were defined to illustrate essential events during the life cycles of these storms, characteristic features that developed during the storm's intense phase, the location of these features prior to

TABLE 1. Heavy snowfall cases for Denver and Colorado Springs. Time is UTC and snowfall amount is in centimeters (inches).

Snowfall cases	Timing		Amount	
	Start	End	Denver	Colorado Springs
Denver cases				
9 Mar 1992	0055/9	1725/9	31.5 (12.4)	3.0 (1.2)
7–8 Jan 1992	0950/7[a]	0114/8	37.6 (14.8)	1.3 (0.5)
6–7 Mar 1990	0000/6[a]	1200/7[a]	30.0 (11.8)	10.7 (4.2)
28–29 Jan 1989	0445/28	0000/29[a]	22.4 (8.8)	0.8 (0.3)
26–28 Dec 1987	1733/26	1615/28	37.8 (14.9)	14.0 (5.5)
3–4 Apr 1986	0306/3	0740/4	32.0 (12.6)	10.2 (4.0)
28–29 Sep 1985	1534/28	1346/29	22.1 (8.7)	4.8 (1.9)
20–22 Apr 1984	1603/20	0529/22	34.5 (13.6)	13.7 (5.4)
26–27 Nov 1983	0734/26	2041/27	54.6 (21.5)	12.2 (4.8)
3–5 Apr 1983	1243/3	0104/5	22.4 (8.8)	10.2 (4.0)
5–6 Mar 1983	0812/5	1232/6	47.5 (18.7)	4.6 (1.8)
24–25 Dec 1982	0625/24	0835/25	60.5 (23.8)	18.5 (7.3)
31 Mar–2 Apr 1975	2148/31	0715/2	23.6 (9.3)	6.1 (2.4)
23–24 Dec 1973	1805/23	1544/24	28.4 (11.2)	11.4 (4.5)
18–19 Dec 1973	1202/18	0835/19	23.4 (9.2)	4.8 (1.9)
7–9 Apr 1973	0440/7	0715/9	29.5 (11.6)	4.3 (1.7)
26–28 Apr 1972	0750/26	0715/28	40.1 (15.8)	7.6 (3.0)
Colorado Springs cases				
29–30 Apr 1990	0000/30[a]	1643/30	0.0 (0.0)	38.6 (15.2)
2–4 Feb 1989	1200/2[a]	0120/4	5.8 (2.3)	31.2 (12.3)
16–17 Mar 1987	0616/16	2010/17	4.8 (1.9)	32.0 (12.6)
26–27 Feb 1987	1005/26	0314/27	9.1 (3.6)	37.6 (14.8)
15–17 Jan 1987	0030/15	0525/17	25.7 (10.1)	59.7 (23.5)
28–30 Mar 1985	0619/28	1115/30	10.1 (4.0)	41.7 (16.4)
20–24 Dec 1983	1200/20	1512/24	13.7 (5.4)	29.5 (11.6)
3–5 Mar 1982	1601/3	1858/5	5.3 (2.1)	21.3 (8.4)
26–28 Dec 1979	1805/26	1610/28	15.2 (6.0)	30.5 (12.0)
3–4 Apr 1979	0010/3	0248/4	8.1 (3.2)	32.0 (12.6)
19–23 Mar 1979	2139/19	0740/23	9.9 (3.9)	36.1 (14.2)

[a] Times were approximated based on local climatological summaries.

cyclonic development, and the role that they played in subsequent evolution of the synoptic-scale pattern. The definition of the time of the three stages was keyed to the time at which snow was first observed. Stage 2 is generally defined as the 12-h synoptic time just prior to the time of first snow except for cases when the start of snow was more than 9 h later than this synoptic time. In this case, the synoptic time immediately following the start of snow was used. Stage 1 is defined as the synoptic time 12 h previous to stage 2. Stage 3, defined as the synoptic time immediately following the greatest 12-h snow accumulation, often occurs after all significant precipitation has fallen. The stage breakdown of each storm is shown in Table 2.

Rawinsonde observations selected for inclusion in the composite analysis were limited to the geographical area of the conterminous United States west of the Mississippi River, the intention being to highlight features over the intermountain West. These soundings were interpolated vertically (linearly in log P coordinates) to 50-mb levels between the surface and 200 mb. The composited wind vectors were computed as vector averages for each level and station. To compute 850-mb temperatures for stations with elevations near or above the height of the 850-mb pressure level [Lander, Wyoming; Denver, Colorado (DEN; now called DNR, site unchanged); Grand Junction, Colorado; Winslow, Arizona; Albuquerque, New Mexico (ABQ); and Ely, Nevada], a procedure adopted at NMC, which applies the 700-mb height and temperature and the 850-mb height to calculate the 850-mb temperature, was used (Rieck 1976; details are given in the appendix). The motivation for this procedure is to compute an 850-mb temperature for these stations that is representative of the synoptic-scale temperature pattern.

The interpolated data were averaged for each stage and processed using GEMPAK routines. Because of boundary effects, the GEMPAK analyses should be considered reliable only away from the boundary of the analysis domain. Composite fields chosen for analysis included temperature, height, horizontal wind components, and 300-mb PV. Winds at 850 mb produced by GEMPAK analysis over mountainous terrain where elevation is above 850 mb (including western

TABLE 2. Winter storm cases used to construct the Denver and Colorado Springs composites for the three stages. Time is UTC.

	Stage 1	Stage 2	Stage 3
Denver			
9 Mar 1992	1200/08	0000/09	1200/09
7 Jan 1992	1200/06	0000/07	0000/08
6 Mar 1990	1200/05	0000/06	0000/07
28 Jan 1989	1200/27	0000/28	1200/28
26 Dec 1987	0000/26	1200/26	0000/28
3 Apr 1986	1200/02	0000/03	0000/04
28 Sep 1985	0000/28	1200/28	1200/29
20 Apr 1984	0000/20	1200/20	0000/21
26 Nov 1983	1200/25	0000/26	0000/28
3 Apr 1983	0000/03	1200/03	0000/05
5 Mar 1983	1200/04	0000/05	0000/06
24 Dec 1982	1200/23	0000/24	0000/25
31 Mar 1975	1200/31	0000/01	1200/01
23 Dec 1973	1200/23	0000/24	1200/24
18 Dec 1973	0000/18	1200/18	1200/19
7 Apr 1973	1200/06	0000/07	0000/08
26 Apr 1972	1200/25	0000/26	0000/27
Colorado Springs			
30 Apr 1990	1200/29	0000/30	1200/30
2 Feb 1989	0000/02	1200/02	0000/04
16 Mar 1987	1200/15	0000/16	0000/17
26 Feb 1987	1200/25	0000/26	0000/27
15 Jan 1987	1200/14	0000/15	0000/16
28 Mar 1985	1200/27	0000/28	1200/29
20 Dec 1983	0000/20	1200/20	0000/24
3 Mar 1982	0000/03	1200/03	0000/05
26 Dec 1979	0000/26	1200/26	1200/28
3 Apr 1979	1200/02	0000/03	1200/03
19 Mar 1979	0000/19	1200/19	1200/20

Colorado, northeastern Utah, western Wyoming, and parts of northwestern New Mexico) should be interpreted with care.

4. Case studies

In this section, we discuss two typical winter storms alluded to previously; one produced heavy snow at Denver and the other produced heavy snow at Colorado Springs. The purpose is not to present a detailed case study of each but rather to demonstrate that the synoptic-scale features developed differently in the two cases. Consequently, the discussion is confined to a few relevant aspects of each case.

a. Denver storm

The Denver storm on 8–9 March 1992 produced blizzard conditions along the Front Range of Colorado and 30 cm of snow at Denver. Colorado Springs received only 3 cm of snow from this event. Figure 2a shows the low-level meteorological conditions at 1200 UTC 8 March, 13 h before snow started at Denver. At 850 mb a cyclone over southern Utah is drawing a wedge of warm air into its circulation, as evidenced by high potential temperatures over the High Plains of southeast Colorado, New Mexico, and western Texas. A developing warm front with attendant warm advection is located over Kansas. Generally, static stability is low over the southern Rockies and Plains south of the warm front in Kansas. North of the cyclone a southward-progressing surface cold front extends through Wyoming. Temperatures behind the front are up to 8.0°C (15°F) colder than at Denver and surface winds reach 10 m s^{-1} (20 kt). These meteorological ingredients are typical during the early stages of strong cyclogenesis in eastern Colorado.

Meanwhile, a 500-mb cutoff low centered over southern Nevada, southern California, and Arizona is moving toward Colorado (Fig. 2b) preceded by a 31 m s^{-1} (60 kt) jet maximum. As a consequence, the Four Corners area (western New Mexico, eastern Arizona, southern Utah, and southwestern Colorado) have experienced 60-m height falls during the previous 12 h.

In response to the upper-level jet-streak forcing, low-level warm advection, low static stability east of the southern Rockies, and northeast movement of the upper trough, the 850-mb cyclone intensifies in the lee of the southern Rockies between 1200 UTC 8 March and 0000 UTC 9 March (Fig. 2c). Cold air continues to progress southward along the eastern slopes of the Rockies behind the surface cold front (cf. the trough of potential temperature values). As shown in the time series of these events in Fig. 3, the cold air arrives at Denver just before 0000 UTC 9 March, accompanied by rising pressure. In response to these atmospheric changes, the winds become north-northwesterly and increase to 18 m s^{-1} (35 kt) at Denver, with wind gusts exceeding 23 m s^{-1} (45 kt) reported in many areas along the Front Range. Two hours prior to the surge of cold air at Denver, the temperature drops 8°C (14°F) at Colorado Springs, accompanied by thunder and a wind shift to westerly. This cooling is attributed to the passage of the eastward-progressing Pacific front indicated on Figs. 2a and 2c.

As the 500-mb cyclone progresses eastward into Colorado (Fig. 2d), the air in contact with the Rocky Mountains is transformed by either vertical stretching as it proceeds over the mountains or by blocking and leeside vortex formation as it is forced around the barrier. Whatever the cause, leeside troughing is induced. Another result is further reduction in the static stability of the lower troposphere induced by the warm advection at low levels that accompanies this leeside troughing (Hovanec and Horn 1975). In addition to these effects, deepening of this cyclonic system can also be partially attributed to the injection of high PV air into the midtroposphere and its subsequent movement toward and superposition over the terrain-induced leeside trough. The analysis from the Mesoscale Analysis and Prediction System (MAPS) (Benjamin et al. 1991) of PV on the 312 K isentropic surface at stage 2 (Fig. 4)

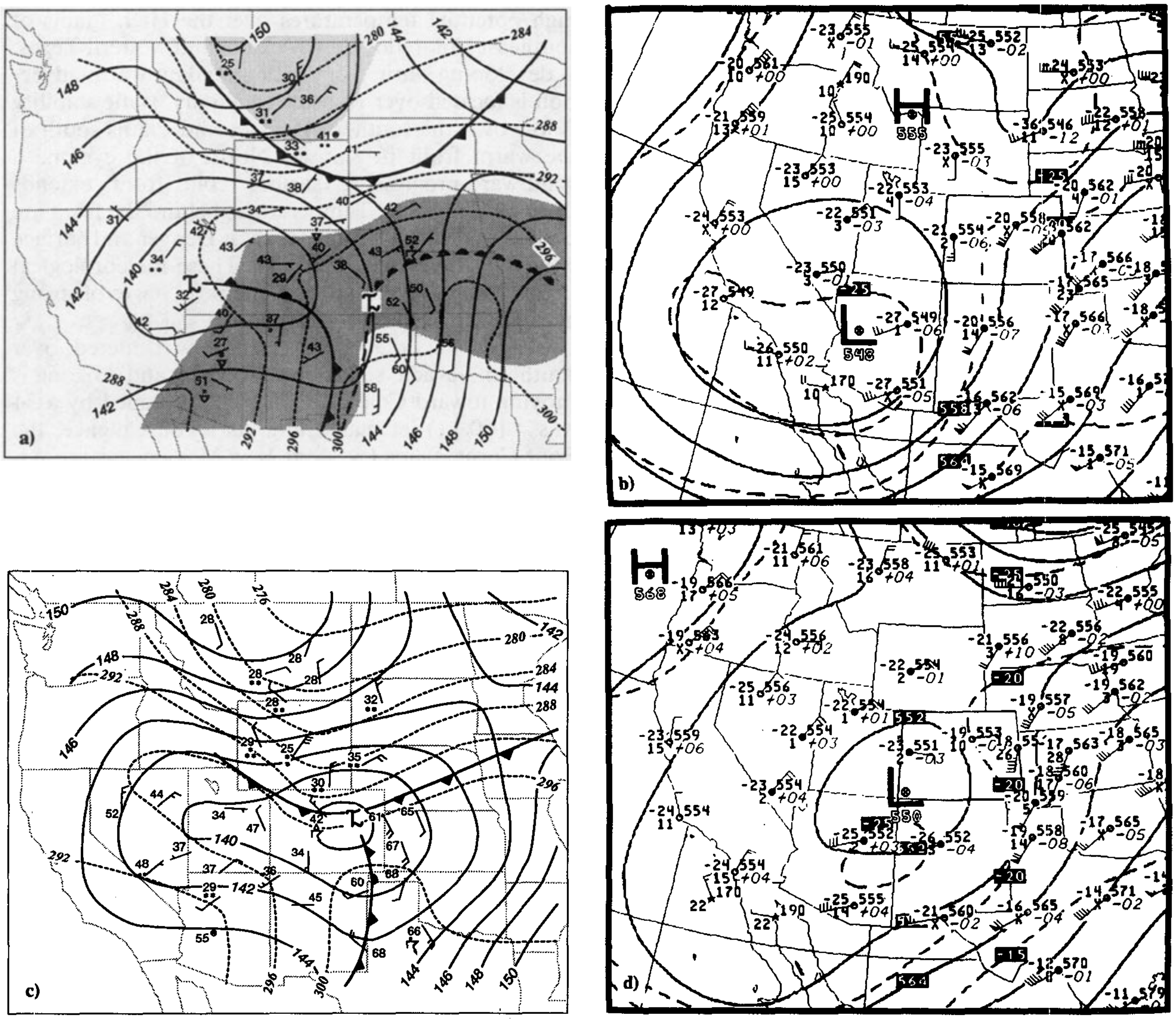

FIG. 2. Synoptic charts for 8–9 March 1992: (a) 1200 UTC 8 March, 850-mb heights (solid lines, 2-dam intervals), potential temperature (dotted lines, 4 K intervals), surface fronts (dark lines, adapted from NMC charts, broken line with half circles attached represents warm frontogenesis), and 850–700-mb lapse rate (darkest stippling greater than 6°C km^{-1} and light stippling less than 3°C km^{-1}). Surface observations follow standard National Weather Service practice [temperatures are in °F, full wind barb 5 m s^{-1} (10 kt), double circle calm winds]; (b) 1200 UTC 8 March, 500-mb plot [standard format, full barb 5 m s^{-1} (10 kt)], height (solid lines, 6-dam intervals), temperature (dashed lines, 5°C intervals); (c) same as (a) excluding stability for 0000 UTC 9 March; and (d) same as (b) except for 0000 UTC 9 March.

indicates a region of high PV over northern New Mexico and southern Colorado where this isentropic surface is near 350–400 mb. Thus, the deepening between 1200 and 0000 UTC results from a synergism of both terrain effects and purely baroclinic processes (the latter being described, for example, by Hoskins et al. 1985). We will see in section 5 (cf. also Mahoney et al. 1994) that this behavior is typical for storms that produce heavy snow at Denver; that is, a well-defined upper-level trough (i.e., PV center) advects eastward so that coupling between upper- and lower-level circulation intensifies an incipient lee cyclone.

By 1200 UTC 9 March (not shown), the system reaches its maximum intensity at both levels. At this time, the 500-mb low is over eastern Colorado and the surface cyclone is in western Kansas. Snowfall at Denver has begun to taper off as the low-level cyclone moves eastward.

b. Colorado Springs storm

The storm of 26 February 1987 produced 38 cm of snow at Colorado Springs in 12 h. Denver received 8 cm of snow. Low-level atmospheric conditions at 0000

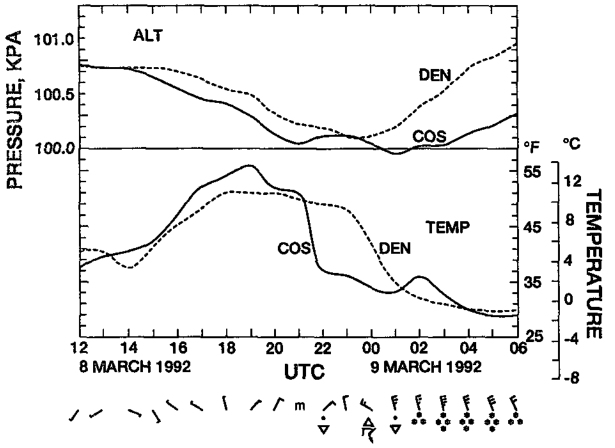

FIG. 3. Temperature and altimeter time series on 8–9 March 1992 for Denver (dotted lines) and Colorado Springs (solid lines). Hourly surface winds (kt, m indicates missing) and precipitation type (symbols follow standard Weather Service definition) at Denver are plotted at bottom.

UTC 26 February (corresponding to stage 1) are shown in Fig. 5a. Lowest heights at 850 mb are located over southern California, significantly farther southwest than in the Denver case. Warm air and low static stability are collocated west of the Rocky Mountains; in the Denver case both are east of the Rockies. In this case, in fact, the air mass east of the Rockies is very stable, with lapse rates less than 3°C km^{-1} in a region extending from southern North Dakota into northeastern Colorado. This area of high static stability is in part associated with the return flow around an arctic high pressure system over the Mississippi River Valley. This feature is not present in the Denver case. The stationary front indicated from northeastern Colorado into the Dakotas is really an "inverted trough" (IT) (Keshishian et al. 1994) entirely within the cold air but along which surface frontogenesis has occurred. The stationary front in New Mexico is of Pacific origin.

Aloft, the initial stages of the Colorado Springs and Denver storms appear similar (Figs. 2b and 5b). However, a short-wave trough over eastern Montana associated with the southward moving arctic front in the Denver case is not present at all in the Colorado Springs case. Further, height changes with the upper low in the Colorado Springs case are slight, suggesting only slow eastward movement. In contrast to the Denver case, only weak low-level intensification occurs, and this is not surprisingly over and west of the Continental Divide where low-level static stability is lowest (Fig. 5c). Meanwhile, the 500-mb cyclone (Fig. 5d) remains virtually stationary over southern California, and only weak height falls occur to its east.

What, then, caused the heavy snow at Colorado Springs? Although we cannot definitively pinpoint the cause, evidence available suggests that interaction of a shallow front with the ambient southeasterly flow was a crucial ingredient. The following scenario describes this process.

The surface IT noted above formed in situ on the High Plains north of Colorado on 23 and 24 February 1987. This suggests that its formation was the result of blocking of stable upslope flow as proposed by Keshishian et al. (1994). The northerly flow to the west of the IT advected colder air southward, resulting in strengthening of the temperature gradient across it, slow deepening of the cold air, and commencement of southeastward movement of the IT in western Nebraska and southeast Colorado. Inspection of standard surface hourly observations indicates that at least one weak surge of enhanced northerlies at the surface propagated southward west of the IT and seemed to reinforce and deepen the cold air north of the Palmer Divide. This weak surge can be seen in Fig. 6 at Denver as the small temperature fall beginning about 1100 UTC. The deepening of the cold air is also apparent through comparison of the 0000 and 1200 UTC Denver rawindsonde soundings (not shown); northerly winds extended to about 2100 m MSL at 0000 UTC and to 2700 m at 1200 UTC.

As the cold air deepened it finally spilled over the Palmer Ridge into Colorado Springs. With south-southeast flow continuing aloft, orographically forced upslope flow south of the Palmer Ridge was enhanced by overrunning this cold air, leading to the increase in snowfall at Colorado Springs seen at 1500 UTC in Fig. 6 after frontal passage there. Once the front had passed the Palmer Divide there was also a mean-state critical level in the flow across the Palmer Divide (low-level northerlies with southerlies above). We speculate that this may have further modulated and locally enhanced the precipitation in the area, as has been demonstrated in a similar context by Snook (1993). In any case, the continued existence of moist southerlies above the surface seems to have been a crucial ingredient for the

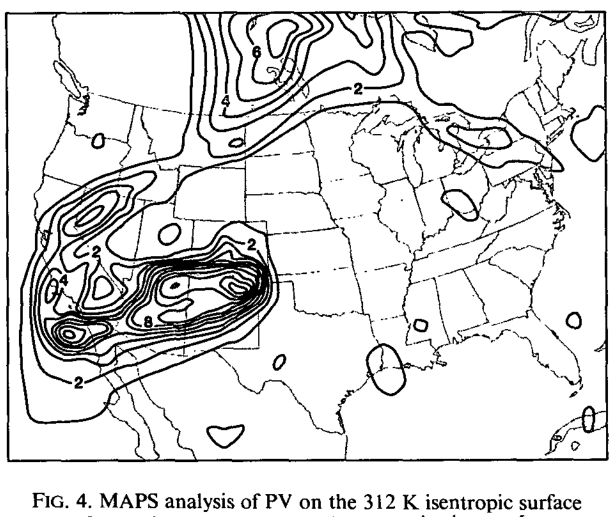

FIG. 4. MAPS analysis of PV on the 312 K isentropic surface for 0000 UTC 9 March, units K mb^{-1} s^{-1} × 10^{-5}.

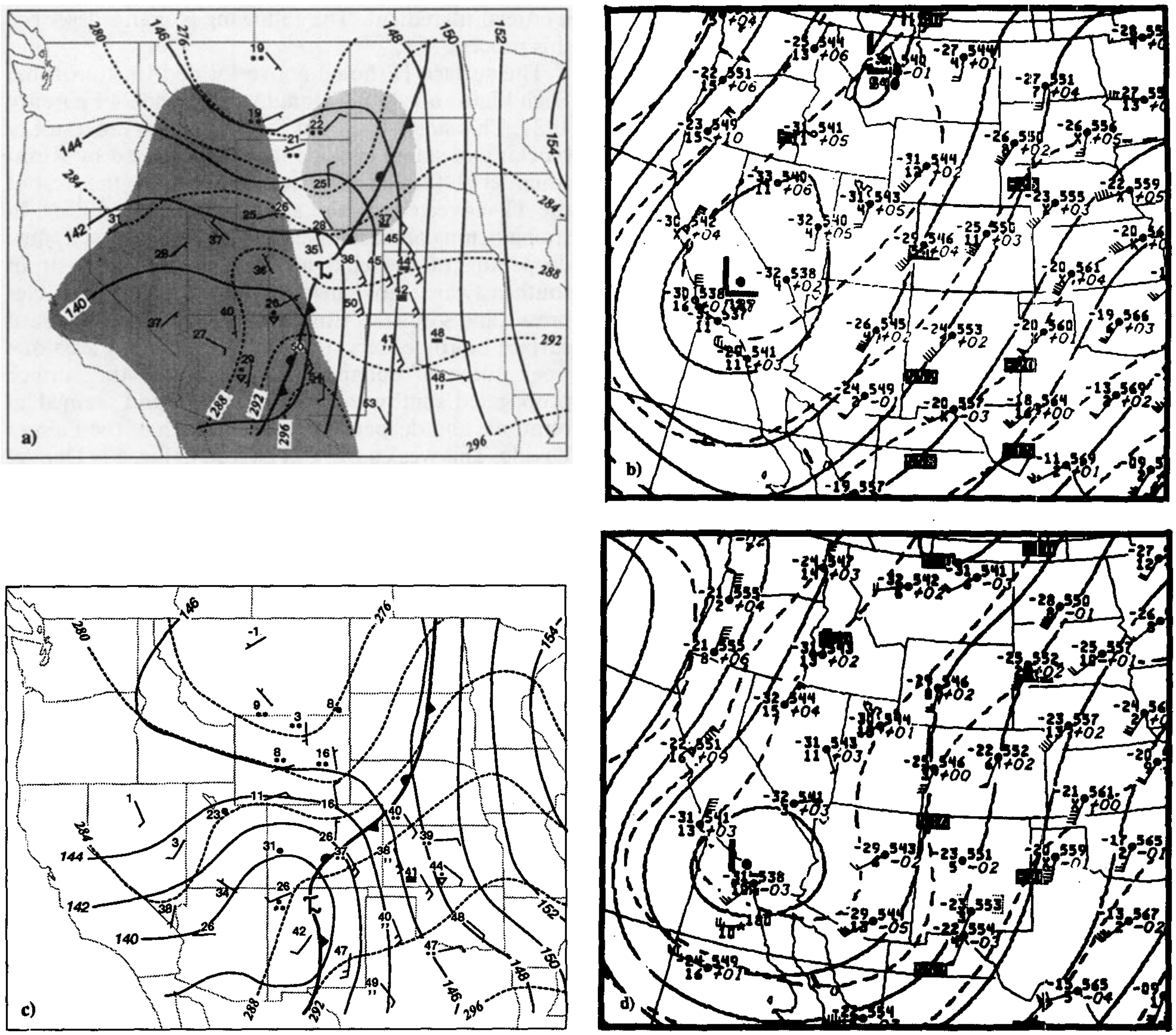

FIG. 5. Synoptic charts for 26 February 1987: (a) same as Fig. 2a except for 0000 UTC 26 February, (b) same as Fig. 2b except for 0000 UTC 26 February, (c) same as Fig. 2c except for 1200 UTC 26 February, (d) same as Fig. 2d except for 1200 UTC 26 February. The 80-kt southeast wind reported by ABQ at 1200 UTC is suspect and thus was deleted from (d).

heavy snow at Colorado Springs; these southerlies were provided by weak cyclogenesis over and west of the Continental Divide and the absence of cyclogenesis to the east.

5. Composite results

On the strength of these case studies and the results of compositing, five principal characteristics of synoptic setting that appear to effectively discriminate between Denver and Colorado Springs storms were identified. They are

1) the location of the low-level warm air relative to the Rocky Mountains;

2) the location of low- or high-static stability east or west of the Rocky Mountains;

3) the character of the leading edge of any low-level southward-moving cold air;

4) the ability of the upper-level flow to induce surface cyclogenesis; and

5) location of the 300-mb PV maximum.

The compositing techniques were devised to assess how useful these five characteristics actually are in distinguishing between the two types of Colorado Front Range winter storms. Not coincidentally, the characteristics are described in terms that are readily determinable from forecast model output or from observations available at Weather Service offices. Thus, the

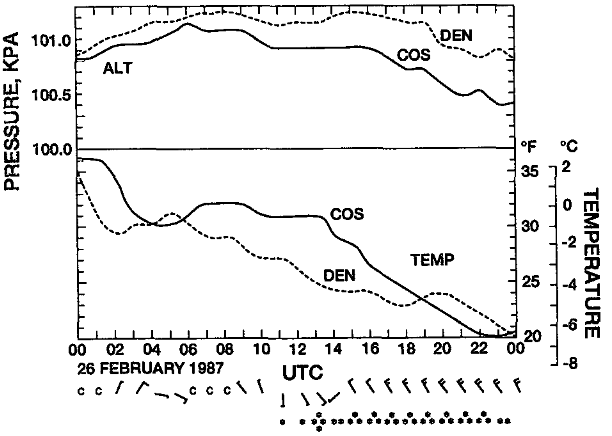

FIG. 6. Temperature and altimeter time series on 26 February 1987 for Denver (dotted lines) and Colorado Springs (solid lines). Hourly surface winds (kt, C indicates calm) and precipitation type (symbols follow standard Weather Service definition) at Colorado Springs are plotted along the bottom.

degree of their generality (as indicated by the composite fields) is also a measure of their potential usefulness in distinguishing between forecasts of heavy snow in Denver and heavy snow in Colorado Springs.

a. Stage 1

Composite 850- and 500-mb fields for the Denver and Colorado Springs storms are shown in Figs. 7 and 8, respectively. The Denver composite at stage 1 (Fig. 7a) indicates a pronounced thermal ridge at 850 mb along the eastern slopes of the Rocky Mountains from central New Mexico to Montana. This ridge of warm air is coincident with warm advection and a lee trough that extends from Colorado northward. These characteristics were key to the development of lee cyclogenesis in the Denver case study.

The temperature advection pattern in the Denver composite is in sharp contrast to that observed in the Colorado Springs composite (Fig. 8a), where cold advection extends from eastern Montana and North Dakota to northeast Colorado. Particularly noteworthy is an area of high static stability stretching from northeastern Colorado to the Midwest and the Great Lakes area and associated with this cold advection. This area of statically stable air reflects the presence of an arctic high pressure system positioned over the Great Lakes to the northeast of the map domain. The location of very stable air to the northeast of Colorado and the axis of warm air and low static stabilities west of the Rocky Mountains are consistent with the case study pattern in Fig. 5a.

Denver and Colorado Springs 500-mb stage 1 composite fields are shown in Figs. 7b and 8b. Both show a short-wave trough over Nevada. Cold advection evident along and west of the trough axis over the northwestern states and Nevada in the Denver composite suggests subsequent amplification of the wave (Fig. 7d shows that amplification does indeed occur). The analogous feature in the Denver case study was a closed 500-mb low over southern California and Nevada (Fig. 2b). In part this difference in location results from the inevitable smoothing produced by compositing procedures. In any case, both the composite and case study analyses agree that a baroclinic wave exists at 500 mb at this stage and later moves eastward to southwestern Colorado. The Colorado Springs 500-mb composite on Fig. 8b, in contrast, exhibits southwesterly flow over most of the analysis domain, with a stronger zonal component and little or no temperature advection associated with what appears to be a more diffuse shortwave trough. Together with the colder and more stable conditions at low levels east of the Continental Divide, this indicates a less favorable environment for leeside development in the Colorado Springs composite than in the Denver composite.

b. Stage 2

Near the time of the first snowfall (stage 2, 12 h later), both composites show low-level cyclogenesis (Figs. 7c and 8c). In the Denver composite a well-defined cyclone has formed just to the lee of the Rockies. This lee cyclogenesis, apparently triggered by an eastward-progressing short-wave trough at 500 mb (Fig. 7d), is consistent with the case study scenario shown in Fig. 2. The approach of the 500-mb short-wave trough now over the Four Corners region has resulted in 50-m height falls over that area between stages 1 and 2. Eastern Colorado is now on the cyclonic shear side of the 500-mb jet.

Quite a different picture is suggested by the Colorado Springs 850-mb stage 2 composite shown in Fig. 8c. Cold air east of the Rocky Mountains, already a dominant feature, is about to be reinforced by a fresh push of north-northwest flow across the northern High Plains. This well-entrenched air mass appears to prevent lee cyclogenesis; weak baroclinic development occurs over and west of the Rockies. At 500 mb the upper-level trough (Fig. 8d) has shown little movement or development since stage 1.

This dramatic difference between Denver and Colorado Springs composites storms is also evident in the evolution of the composite wind fields (Figs. 9 and 10). For example, the preferred wind direction for Denver during stage 1 is southwesterly at both 700 and 500 mb, which reflects the position of the upper-level trough. As this trough starts to develop during stage 2, the preferred wind direction becomes southerly at 700 mb. (Given that rapid changes occur during stage 2, the constraint of 12-h sampling introduces variability that shows up as the scatter of the wind directions on Fig. 9. For example, the clustering of surface winds into two primary directions, north-northwesterly and southeasterly, possibly indicates that several of the

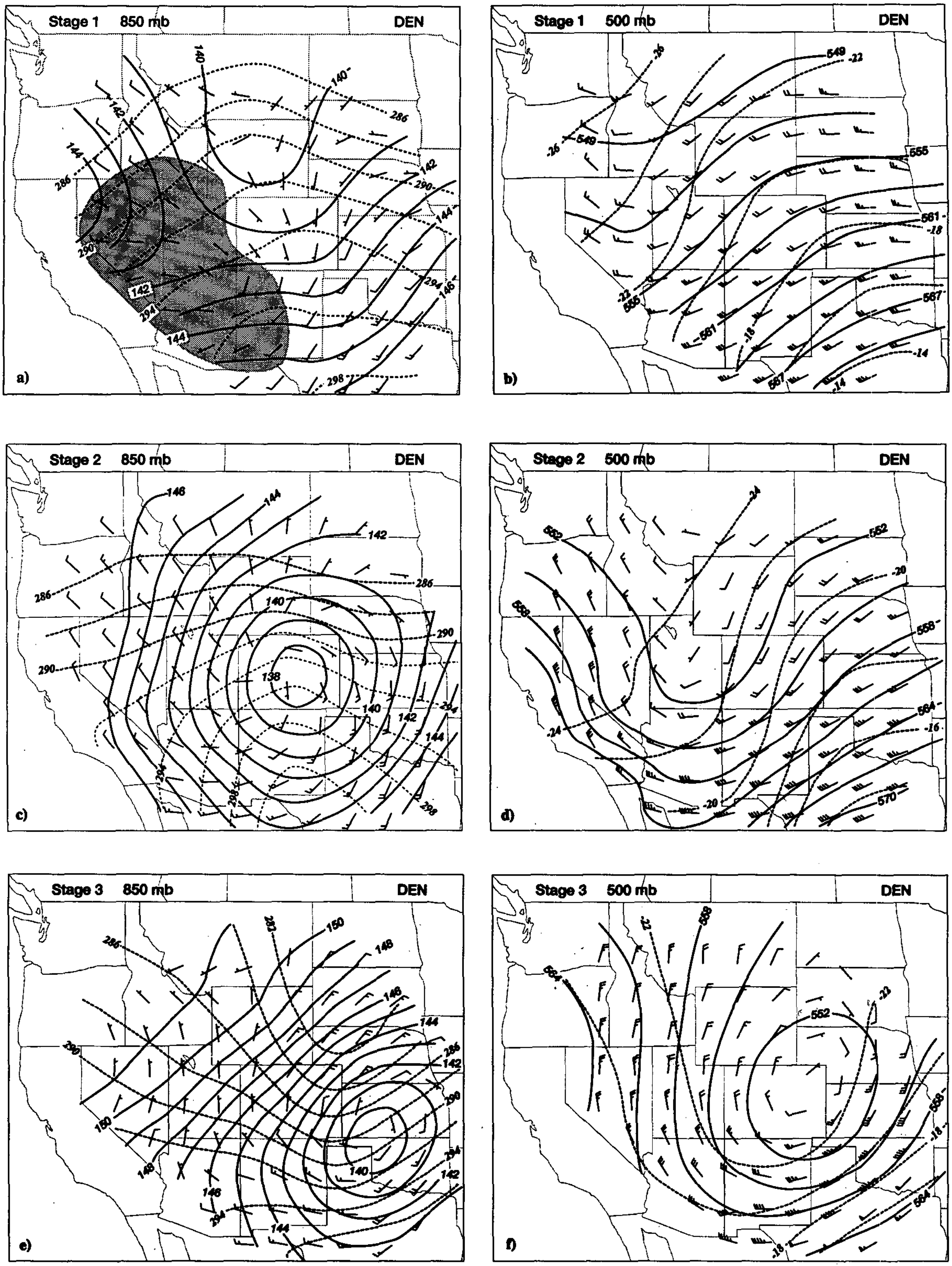

FIG. 7. Denver composite for (a) 850-mb during stage 1. Plotted are heights (solid lines, 1-dam intervals); potential temperature (dotted lines, 2 K interval), 850–700-mb lapse rate (darkest stippling greater than 7°C km^{-1} and no stability less than 3°C km^{-1}), and winds [full barb, 5 m s^{-1} (10 kt)]. (b) For 500 mb during stage 1. Plotted are heights (solid lines, 3-dam intervals), temperatures (dotted lines, 2°C intervals), and winds [full barb, 5 m s^{-1} (10 kt)]. (c) Same as (a) without stability except for stage 2. (d) Same as (b) except for stage 2. (e) Same as (c) except for stage 3. (f) Same as (d) except for stage 3.

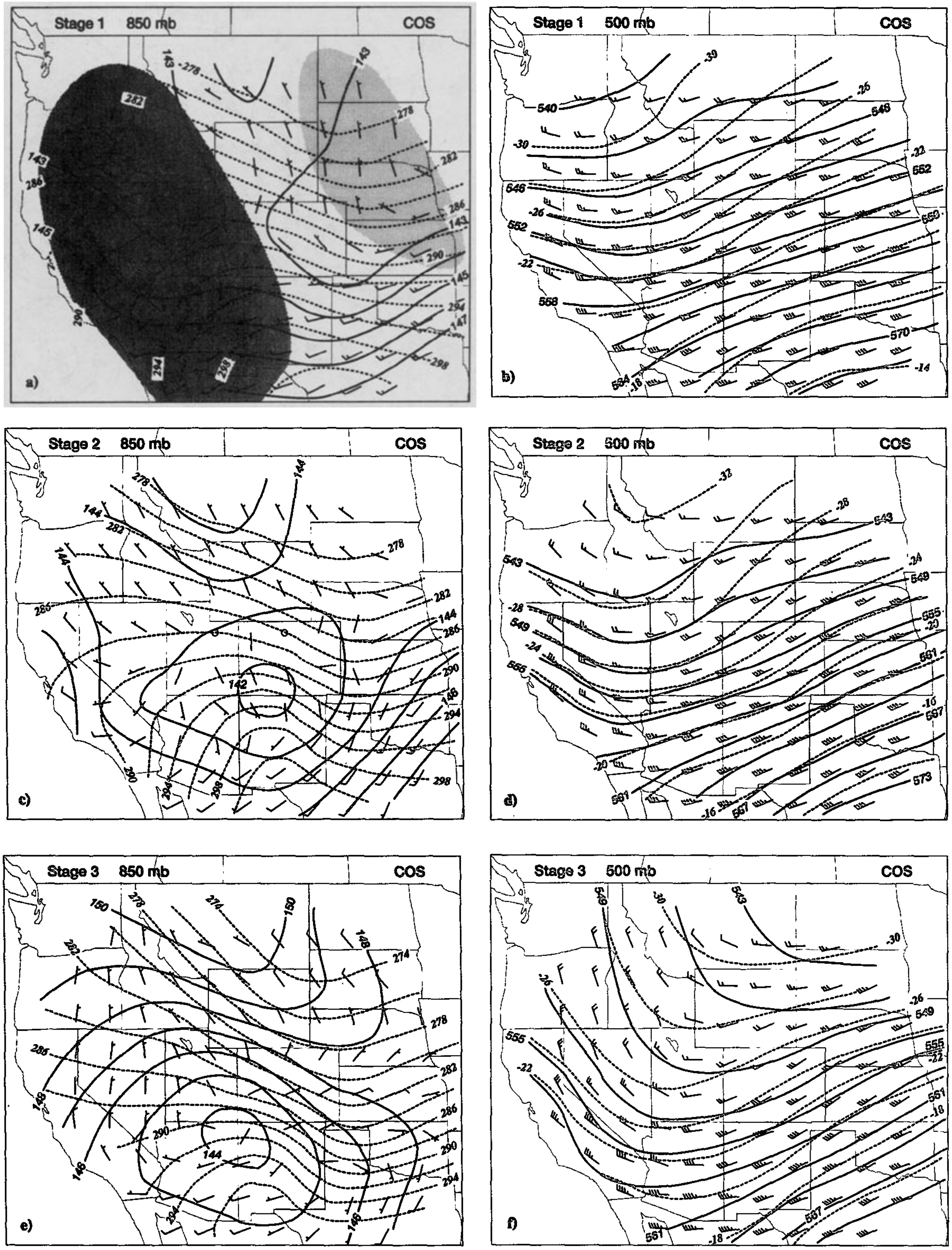

FIG. 8. Same as Fig. 7 for Colorado Springs, except 850–700-mb lapse rate (darkest stippling greater than 6°C km^{-1} and lightest stippling less than 3°C km^{-1}).

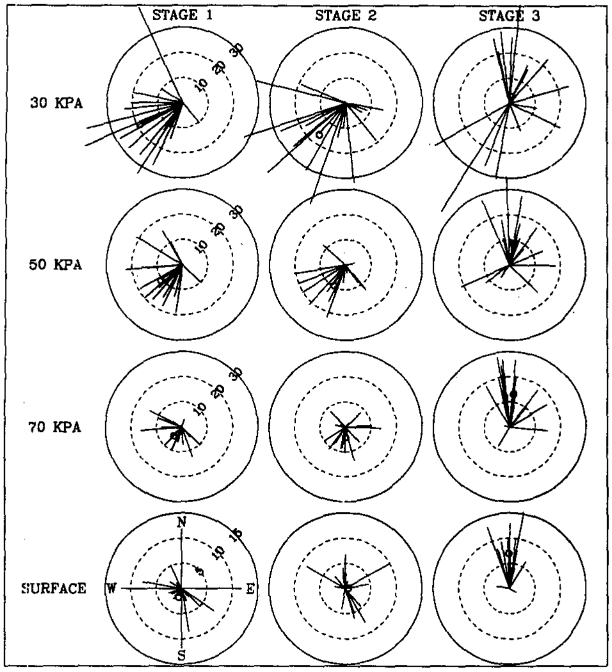

FIG. 9. Wind roses constructed from rawinsonde observations of individual Denver cases. Surface winds are taken from Denver MF10A/B hourly observation forms. All winds are in meters per second. The symbol "*O*" indicates the resultant wind formed as an average of the *u* and *v* components of the individual cases.

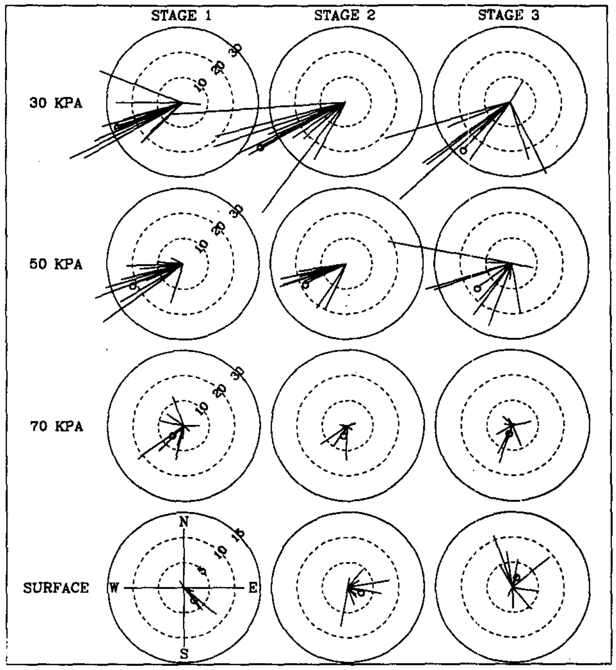

FIG. 10. Same as Fig. 9, except for Colorado Springs. Upper-level winds are taken from Denver soundings and surface winds are taken from Colorado Springs MF10A/B hourly observation forms.

synoptic times sampled during stage 2 occur before winds have shifted to the north at Denver. Once snow starts at Denver, surface winds are typically from the north.) By stage 3 northerlies dominate all levels, suggesting a well-developed cyclone extending to 300 mb. The preferred wind direction for Colorado Springs (Fig. 10) remains generally southwesterly during all stages at 700 and 500 mb with little scatter. This substantiates the impression gained from Fig. 8—that of a major long-wave trough with only weak short-wave features embedded.

The appearance of the 300-mb composite PV fields in Fig. 11 provides another clue that the Denver storms evolve differently from the Colorado Springs storms. For these figures PV has been computed by GEMPAK on pressure surfaces as

$$\left[(\bar{\zeta}_p + \mathbf{k} \times \frac{\partial \bar{\mathbf{V}}}{\partial \theta} \cdot \nabla_p \bar{\theta}) + f\right] \frac{\partial \bar{\theta}}{\partial p} = \mathrm{PV},$$

where all symbols are defined in their usual way and the overbar indicates an average across the individual cases in each set. (The choice to present these computations on pressure surfaces was strictly one of convenience; because advection is primarily along isentropic surfaces, PV should really be viewed on surfaces of constant potential temperature.) The upper-level PV maximum in the Denver composite is located over southern Utah and northern Arizona and is moving eastward toward the low-level warm air located in the lee of the Rocky Mountains. This injection of high PV air into the midtroposphere and the movement toward a low-level warm anomaly are often associated with

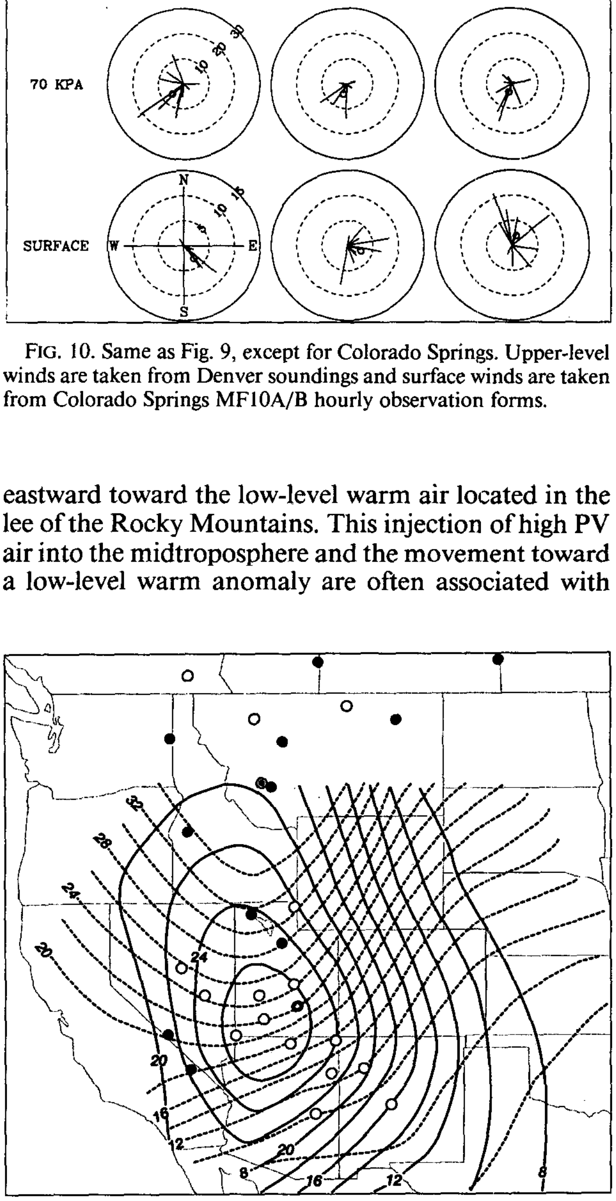

FIG. 11. Stage 2 composite 300-mb PV for Denver (solid lines) and Colorado Springs (dotted lines), units are 2 K s^{-1} Pa^{-1} intervals. Centroid locations of predominant PV maxima for Denver cases (open circles) and Colorado Springs cases (black circles) and median positions for both sets (double circles).

the first stages of deep baroclinic development (Hoskins et al. 1985). In the Colorado Springs composite, on the other hand, a broader PV pattern, centered farther north, is advecting eastward with little or no apparent interaction with low levels, simply because the warm anomaly is absent or lies to the west of the Rocky Mountains.

c. Stage 3

By stage 3 cold air from the north and northeast has displaced the warm air associated with lee troughing and cyclogenesis in the Denver composite, in effect "wedging" the cyclone away from the eastern slopes of the Rockies (Fig. 7e). During this stage the wind is consistently from the north at the surface and 700 mb (Fig. 9). At 500 mb, the trough has now closed off over eastern Colorado and the adjoining states (Fig. 7f). This intense cyclonic development has likely been enhanced by dynamic feedback mechanisms between the lower- and upper-level cyclones.

As previously seen in the Colorado Springs case study, the composite Colorado Springs 850-mb cyclone and anticyclone are oriented in such a way as to produce strong southeasterly geostrophic flow into eastern Colorado (Fig. 8e). Figure 10 shows, however, that the resultant surface winds in Colorado Springs (somewhat above 850 mb) are really the sum of several cases with north-northwesterly flow, including the Colorado Springs case study, and other cases with southeasterly flow. Evidently, in many (but not all) Colorado Springs storms, low-level terrain-induced upslope from the southeast is enhanced by overrunning of shallow north-northwesterlies.

At upper levels (Fig. 8f), cyclonic development in the Colorado Springs composite is considerably less than in the corresponding Denver composite. The combination of a persistent long-wave trough over the western United States, entrenched cold air over the High Plains, and a favorable position on the cyclonic side of the jet help maintain the low-level cyclone west of the Rocky Mountains.

d. Composite assessment

The credibility of these (and any similar) composited fields depends on how well they summarize the individual cases used in their construction. Ideally, the composited features will closely reflect features present in most or all of the individual storms. In the worst case, individual input fields will be so highly variable that the composites constructed from them will either be flat and featureless or wildly different from any of the input fields. It is thus important to demonstrate whenever possible that there is some degree of consistency within the set of cases selected for compositing.

Consistency of the wind fields in the two composite sets analyzed here can be judged qualitatively from the wind roses in Figs. 9 and 10. Quantitative assessment is possible using the measure of persistence given by Panofsky and Brier (1963) as the ratio of the vector average wind speed to the scalar average wind speed. Values of this measure applied to the winds are presented in Table 3. When approaching unity, these values indicate considerable uniformity in direction and speed of the input winds. Conversely, values near zero indicate a high degree of intercase variability and, as a consequence, significant ambiguity in the interpretation of the composited wind constructed from these cases.

Examination of the wind roses and the table of persistence values shows that at most pressure levels and during most stages the composite wind is a good representation of the winds that prevail in the individual cases. The 500- and 300-mb winds in the first two stages of the Colorado Springs composite, for instance, show persistence values well over 0.9, a result that is clearly illustrated by the corresponding wind roses in Fig. 10. When the persistence values fall below 0.4, however, the wind roses show increasingly scattered wind directions. During stage 2 in Denver, for instance, (Fig. 9) persistence of the surface winds falls to 0.09. (An explanation for this low value is suggested in section 5b.) We thus view the composite winds here with skepticism. By stage 3, winds have come around to northerly in almost all cases and persistence values have recovered to values approaching unity. Beyond these general comments, no distinct patterns in height or trends with time are revealed in Table 3.

Another qualitative measure of the meaningfulness of the two composites is also illustrated in Fig. 11. Within the set of Denver storms there is a clear indication of clustering of centroids around the median location in southern Utah. Although the greater scatter of centroids in the set of Colorado Springs storms renders the median location in Montana less meaningful, there still appears to be a distinct difference between the preferred regions of the two sets. Furthermore, these preferred regions correspond well to the contoured composite PV patterns in the figure. Given the heavily

TABLE 3. Mean wind and persistence values.

Station / Level	Stage 1	Stage 2	Stage 3
Denver			
Surface	0.39	0.09	0.90
700 mb	0.55	0.44	0.81
500 mb	0.71	0.72	0.55
300 mb	0.78	0.69	0.14
Colorado Springs			
Surface	0.69	0.40	0.35
700 mb	0.53	0.65	0.50
500 mb	0.94	0.95	0.76
300 mb	0.91	0.95	0.81

computed nature of PV, this is in fact a fairly rigorous test of the composite results.

6. Discussion and conclusions

Observational evidence for distinct differences in the synoptic and mesoscale patterns associated with two sets of storms (those that produce heavy snowfall in Denver and those that produce heavy snowfall in Colorado Springs) has been presented. This evidence is based on case studies of individual storms and on composite analyses combining atmospheric fields from several storms of each type. The case studies were discussed to demonstrate how these different storm environments evolved in specific instances, while the composite analysis, based on snowfall differences alone, was designed to assess the generality of the two environments and hence the importance of the observed differences between them.

a. Differences in the storm environments

The set of Denver storms display characteristics of a strongly developing baroclinic cyclone: a deepening mid- to upper-tropospheric shortwave associated with a locally lowered tropopause or PV maximum, warm advection and unstable air at lower levels ahead of this shortwave, surface cyclogenesis in the lee of the Rockies, and apparent interaction between upper- and lower-level circulations. The Colorado Springs storms, on the other hand, exhibit significantly less dramatic development in both the upper- and lower-level circulations and minimal interaction between levels, despite the considerable amount of snow produced at Colorado Springs.

Part of the difference between the two storm environments is simply one of placement, and, in particular, placement relative to large-scale and regional-scale terrain features. The location of the low in southeastern Colorado at the time of heaviest snow in the Denver storms provides strong north to northeasterly upslope flow near Denver. Development of the Denver surface cyclone benefited from the cyclogenetic properties of High Plains air with a history of descent off the Rockies. In the Colorado Springs storms, on the other hand, the location of the surface low farther west and south, in combination with a cold anticyclone over the central and northern Plains, produces a low level east to southeasterly geostrophic wind. Also, the more western location of the upper-level short-wave trough in the Colorado Springs storms contains surface development west (that is, upstream) of the Rockies.

The situation is not as simple, however, as the preceding discussion might suggest. For instance, characteristics of the coldest air and the timing of its arrival play roles substantially different but equally important in both storms. At the time of heaviest snowfall in the Denver storms, the southward-advancing cold air behind a cold front or surge has become a wedge of strong northerly flow extending eastward out into the High Plains from the mountain barrier to the west. As discussed previously, this forces the low-level flow from the east to ascend over this wedge of cold air and affects future development by displacing the surface cyclone southeastward. [However, the 26 December 1987 (Barnes and Colman 1993) and 28 September 1985 and 3 April 1983 cases were different in that they had little if any cyclone development, being primarily upper-level systems with the cold air already in place.] In the Colorado Springs storms, a surface cold air mass exists over northern Colorado well before the approach of the upper-level disturbance and subsequent surface development. Further intrusion of this cold anticyclone into eastern Colorado during the period of snowfall in Colorado Springs produces a very stable atmosphere east of the Rockies, inhibiting cyclogenesis there and effectively confining the surface low west of the mountains. An inspection of low positions in these cases (see Tollerud and Mahoney 1995) indicates that by stage 3, only two or maybe three cases have lows that move into the lee of the Rockies, but none of them are strongly cyclogenetic as compared to the Denver cases. In essence, the cold air in the Denver storm is characterized as a fresh outbreak that becomes available to the system at the critical stage in its development and leads to further development. This result is supported by Buzzi and Tibaldi (1978) and McGinley (1982), who determined that the onset of rapid cyclonic growth occurs when the cold air arrives in the lee. Conversely, cold air in the Colorado Springs storm is present before cyclonic development begins.

b. Implications for operational forecasting

Since the storm characteristics described above are for the most part easily seen in existing observational fields and readily identified in numerical guidance, they may also prove useful to forecasters faced with a decision between heavy snowfall in Denver or Colorado Springs. Therefore, the following guidelines are suggested. A summary of these is given in Table 4.

1) Given the possibility that heavy snow may occur, conditions necessary for determining which location has the highest potential for heavy snow can be identified from observations 6–18 h before snow is expected.

2) At that time, the presence of warmest air and steepest lapse rates to the east (west) of the Continental Divide indicates subsequent heaviest snow in Denver (Colorado Springs).

3) Preexisting cold air over the Great Plains favors heaviest snowfall in Colorado Springs; an approaching cold front or surge timed to reach central Colorado just prior to the upper-level trough favors Denver.

4) Low-level warm advection in the lee of the Front Range ahead of any approaching cold front favors

TABLE 4. Conditions favoring heavy snow.

Denver	Colorado Springs
1. Warmest air and steepest lapse rates at low-levels are or are anticipated to be east of the Continental Divide.	1. Warmest air and steepest lapse rates at low-levels are west of the Continental Divide.
2. Cold front or surge is timed to reach central Colorado just prior to upper-level trough.	2. Preexisting cold air is positioned over the Great Plains.
3. Warm advection is present in the lee of the Front Range.	3. High static stability and cold advection are present in the lee of the Front Range.
4. A well-defined 500-mb short-wave trough moves across the Four Corners region.	4. A slow moving 500-mb long-wave trough is positioned to the southwest of the Four Corners.
5. Injection of high PV air into middle troposphere and subsequent superposition with low-level leeside warm air.	5. Advecting upper-level PV maximum having minimal interaction with low levels.

heaviest snow in Denver; high static stability and cold advection in the lee favors Colorado Springs.

5) A well-defined, mobile, 500-mb short-wave trough or upper-tropospheric maximum of isentropic PV moving into the Four Corners region indicates heaviest snow in Denver; a quasi-stationary 500-mb low or long-wave trough centered farther to the southwest, with an advecting PV maximum having a track projected to keep it remote from leeside warm air, favors Colorado Springs.

It is well known (e.g., Brown and Szoke 1994) that the operational numerical models have notable idiosyncrasies in their performance over the Rocky Mountains. In particular, the Nested Grid Model and the Global Spectral Model tend to be too far north and too fast in predictions of surface features to the lee of the mountains. Further, these models will often forecast too much leeside deepening in situations where shallow cold air is entrenched east of the Rockies. [Based on one winter's (1993–1994) limited experience, the eta model performs somewhat better in both respects.] Forecasters are well advised, then, to evaluate the current data as well as model forecasts when applying these guidelines.

Since "PV thinking" concepts seem to encapsulate many of the pivotal characteristics described here (e.g., the role of near-surface stability, lee cyclogenesis, and coupling between upper- and lower-level circulations), forecasters are encouraged to apply these concepts in operations, especially as gridded analysis and model forecast output and the ability to manipulate them become commonplace.

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APPENDIX

850-mb Temperatures over High Terrain

The 850-mb temperature for all rawinsonde stations reporting surface pressure less than 850 mb are computed using the following relationship (Rieck 1976):

$$T_{850} = 0.352(Z_{700} - Z_{850}) - T_{700} - 546.3.$$

Heights Z are in meters and temperatures T are in degrees Celsius. It is assumed that temperature varies linearly with the logarithm of pressure and that the difference between virtual and actual temperatures is negligible.

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