Mesoscale Weather Effects of Variable Snow Cover over Northeast Colorado

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ABSTRACT

Data from the PROFS (Program for Regional Observing and Forecasting Services) surface mesonetwork have been used to document the effect of variable snow cover on atmospheric boundary layer properties, cloudiness and weather conditions over northeast Colorado on 15 April 1983. On this day an oval-shaped $\sim 10^4$ km^2 area of snow-free ground surrounded by snow-covered ground existed along the Colorado Front Range. While sky conditions on the morning of this day were everywhere clear, cloudiness developed by midday over the snow-free region as a result of the more rapid boundary-layer heating and mixed-layer growth there. During midafternoon snow showers occurred over the snow-free ground whereas skies remained mostly clear over the snow-covered area.

Our analysis suggests that snow boundaries in the region may have acted through the development of a weak solenoidal field to enhance low-level inflow into the snow-free area, thereby assisting with cloud development in the region. Analogous to the sea breeze, this phenomenon might be termed a "snow breeze." Even without such an enhancement to the circulation, the variable snow cover through its impact on the surface energy budget had a profound effect on the regional weather conditions on that day. This situation represents just one example of a class of complex interactions and feedback processes involving variable surface properties and the large-scale flow.

1. Introduction

Forecasters in colder climates are well aware of the problems snow and ice cover can create for local weather forecasts. During the cold season, for example, forecasters in the United States frequently modify numerical guidance for maximum and minimum temperatures [e.g., those given by Model Output Statistics (MOS) based on conventional synoptic data; Glahn and Lowry, 1972] using satellite and surface station information on regional snow cover. Particularly noteworthy are variations in weather conditions in the vicinity of snow/no snow boundaries. Wash *et al.* (1981) have noted important effects of a rapidly-melting snowband in the central and upper Mississippi River valley on the diurnal temperature cycle and boundary-layer development in that region. Bluestein (1982) has recently documented a case where variable snow cover over Oklahoma and Kansas had a significant impact on maximum temperatures for the day in a mesoscale ($\sim$100 km) area in those states.

With the advent of mesoscale networks of surface stations in the United States, the opportunity has arisen for more detailed (in the horizontal scale) investigation of the forecast problem described above. In this study, data from the Program for Regional Observing and Forecasting Services (PROFS; Reynolds, 1983) surface mesonetwork and other sources in northeast Colorado will be used to document the impact of snow cover variability in this region on the evolution of temperature, wind, cloud cover and precipitation on 15 April 1983. On this day an $\sim 10^4$ km^2 area of snow-free ground surrounded by snow-covered ground led to strikingly different weather conditions during the afternoon as a result of differential surface heating. By afternoon, overcast skies with snow showers occurred over the snow-free region, whereas mostly clear conditions prevailed over the snow-covered areas. The authors' ability to monitor the developing weather conditions on this day through access of real-time data contributed immeasurably to our motivation to proceed with this study as well as to our success in acquiring sufficient PROFS data to carry out a relatively complete analysis. An understanding of the processes leading to such important differences in weather conditions on the mesoscale is a vital factor in improving short term weather forecasting in these types of situations.

2. Data sources

a. PROFS mesonetwork

Surface observations from a mesometeorological network of 22 stations over northeast Colorado have been used in our analysis. A map of these stations (small letters) and surrounding National Weather Service (NWS) stations (large letters) is shown in Fig. 1. The separation between PROFS stations is variable, but averages about 40 km. Standard meteorological variables as well as pyranometer measurements of total solar radiation (direct and diffuse) on a plane surface are recorded and transmitted in the form of 5-minute averaged data. A listing of the station identifiers and elevations of all PROFS stations in Fig. 1 is given in Table 1. Major topographic features of the region are also indicated in Fig. 1. The PROFS stations are centered over the upper reaches of the South Platte River drainage basin with the Continental Divide to the west, the Cheyenne Ridge to the north and the Palmer Lake Divide to the south. The variation in elevation over the region is large, with stations ranging from 1372 m (4500 feet) at FTM to 3505 m (11 500 feet) at ISG. This study will concentrate on analyses over the high plains where the variation in elevation is considerably less.

b. Geostationary satellite and NWS platforms

Visible satellite imagery from the National Oceanic and Atmospheric Administration (NOAA) Geostationary Operational Environmental Satellite (GOES) system having 2 km resolution has been used to determine the horizontal extent of snow cover. These data have been mapped onto a regional geographic grid using the Colorado State University (CSU) Interactive Research Imaging System (IRIS, Green and Kruidenier, 1982).

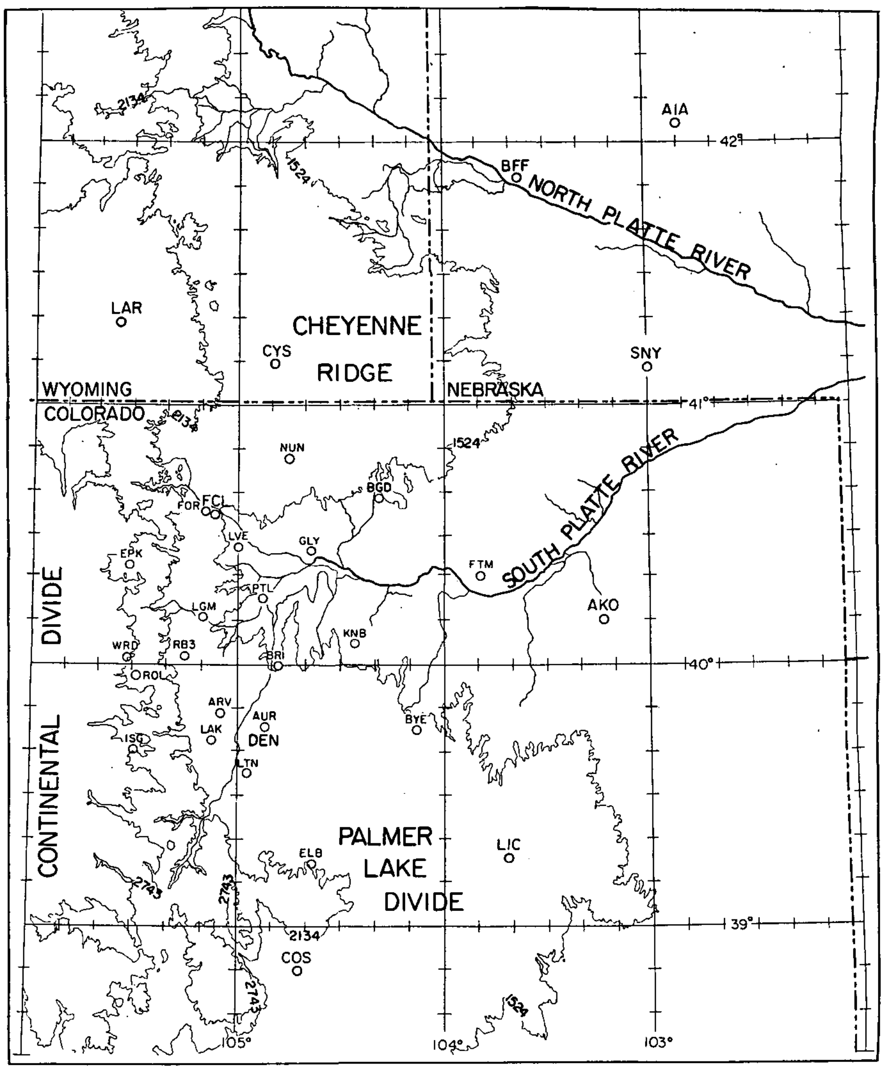

FIG. 1. PROFS mesonetwork stations (small letters) and surrounding National Weather Service stations (large letters). Major geographical features are identified. Elevation contours are in meters [1524 m (5000 feet), 2134 m (7000 feet), 2743 m (9000 feet)].

TABLE 1. PROFS mesonetwork stations and their elevations.

Identifier	Name	Elevation (m)
ARV	Arvada	1643
AUR	Aurora	1625
RB3	Boulder	1609
BGD*	Briggsdale	1483
BRI	Brighton	1518
BYE	Byers	1554
ELB	Elbert	2146
EPK	Estes Park	2377
FOR	Fort Collins	1609
FTM	Fort Morgan	1372
GLY	Greeley	1414
ISG	Idaho Springs	3505
KNB	Keenesburg	1521
LAK	Lakewood	1832
LTN	Littleton	1750
LGM	Longmont	1533
LVE	Loveland	1512
NUN	Nunn	1634
PTL	Platteville	1457
ROL	Rollinsville	2749
WRD	Ward	3048

* On several charts Briggsdale is indicated by BRG.

NWS surface station data and Denver (DEN) sounding data have been obtained both through real-time acquisition in the CSU Department of Atmospheric Science and from the National Climatic Center, Asheville, North Carolina.

3. General meteorological conditions

a. *Snow cover*

A major snowstorm developed in northern Colorado on 12 April 1983, as a result of strong cyclogenesis in the southeastern part of the state. Near-blizzard conditions ensued across much of the northeast Colorado, leaving a continuous snow cover of variable depth over the region. By the morning of 15 April, melting had left only a portion of northeast Colorado covered by snow. The distribution of snow cover is best revealed by visible satellite data. Visible imagery over north central and northeast Colorado at 0802 MST (Mountain Standard Time) from GOES-East is shown in Fig. 2. Several Colorado cities and county boundaries are depicted in the picture. Surface observations from stations within the region at the time of this picture early in the morning show clear skies; thus, we can interpret the bright portions of the image as corresponding to snow cover. To the west (left) of a Fort Collins–Boulder–Denver–Colorado Springs line, snow cover exists virtually everywhere despite the appearance of darker regions on the satellite image. The areas of darkness are attributable to the forest canopy prevalent throughout this region. This problem of interpretation does not exist to the east of this line over the plains where no forests exist.

The satellite-determined snow cover has been transferred to a basemap encompassing the PROFS mesonetwork in Fig. 3. Reported snow depths (cm) at stations in the region within ~1 h of the time of the satellite picture are entered in the figure. These rather sparse measurements are in general agreement with the areal extent of snow cover given by satellite. In several locations snow is reported by observers (3 cm or ~1 inch) where none is evident from the satellite picture. We regard these discrepancies as minor and a consequence of slightly variable station observing times and the difficulty in measuring the depth of shallow, patchy snow cover. The snow-covered area as depicted by Fig. 3 is regarded as a reasonable approximation to the average for the day, since the rather cool daytime temperatures prevented significant melting (as supported by snow depth observations on the following day).

b. *Synoptic setting*

At the surface at 0500 MST (1200 GMT) 15 April a ridge of high pressure extends across the west-central United States with a weak pressure field providing for a weak northwesterly gradient wind over northeast Colorado (Fig. 4). The storm which provided the snow several days earlier is located over the Great Lakes area. Twelve hours later at 1700 MST (henceforth all times are MST) 16 April (Fig. 5) the same general synoptic-scale conditions exist with a surface high pressure center along the Colorado–Wyoming border. At 700 mb (Fig. 6, 0500) and at 500 mb (not shown), cold northwesterly flow exists over the central United States. There is a general tendency for rising heights over the Rocky Mountain states as a major upper-level trough over the east-central United States moves

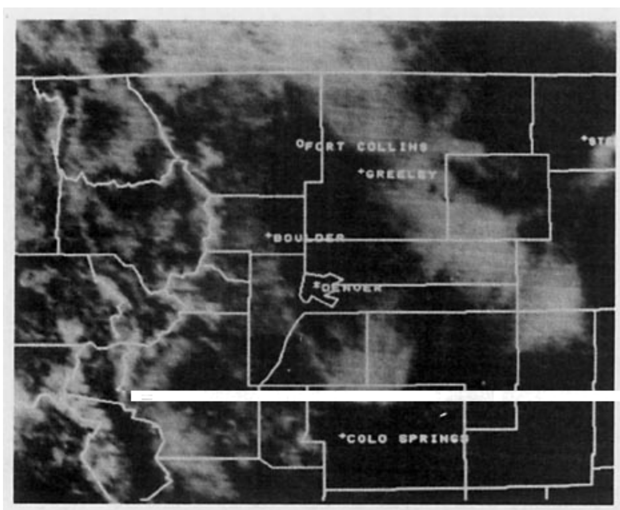

FIG. 2. Visible GOES-East imagery (2 km resolution) at 0802 MST 15 April 1983 depicting snow cover over northeast Colorado. Several Colorado cities and county boundaries are indicated.

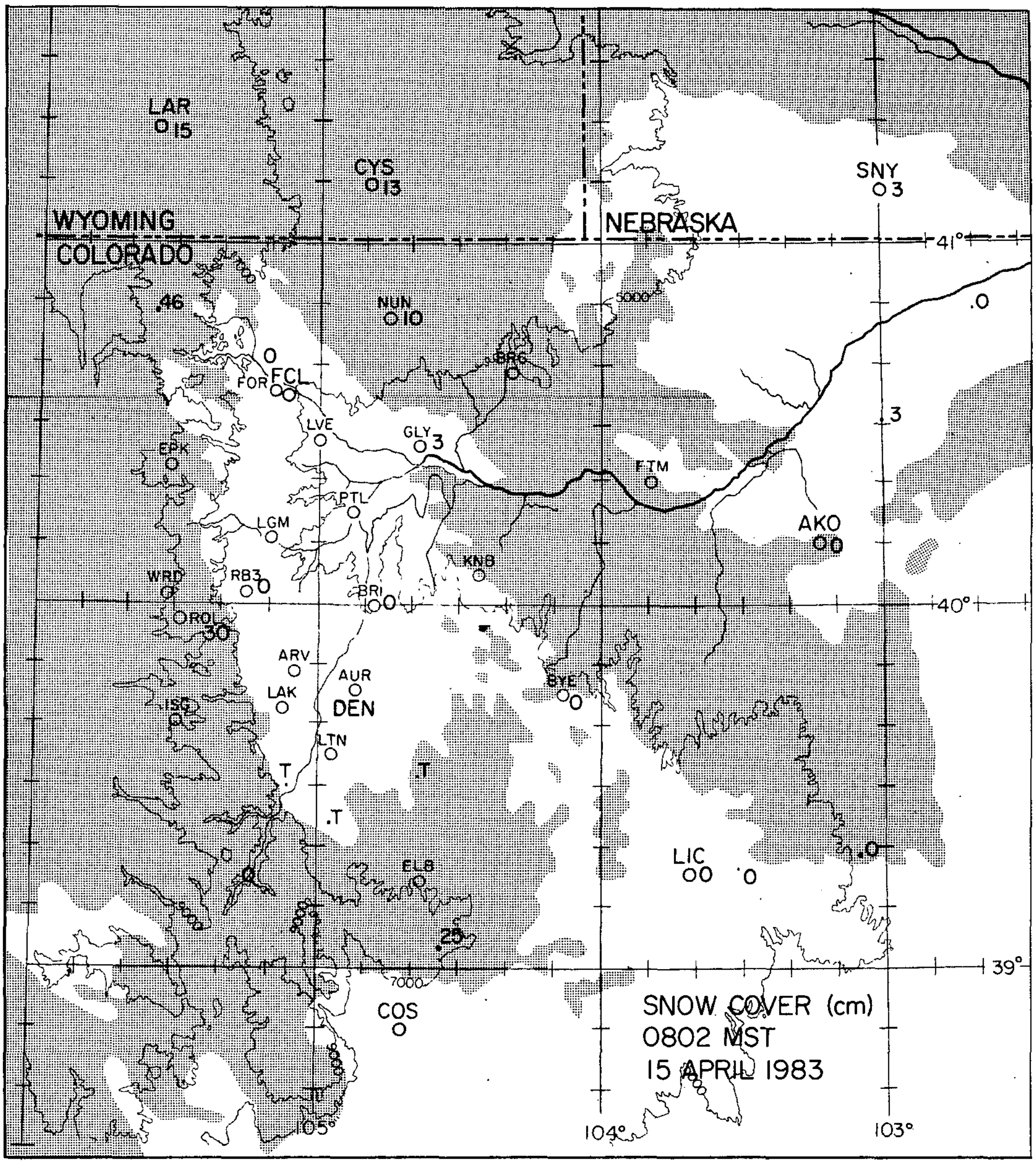

FIG. 3. Snow cover at 0802 MST (shaded area) estimated from satellite picture (Fig. 2). Reported depths (cm) at National Weather Service stations are indicated.

eastward. No significant short-wave disturbances are evident in the northwesterly flow; however, an area of enhanced moisture in the north-central United States (shaded area in Fig. 6) can be tracked in this flow southeastward to the central United States twelve hours later (Fig. 7, 1700). The moistening evident at Denver through the day (shown later in Fig. 11) may be partly attributable to the large-scale moisture advection illustrated here. Curiously, the 700 mb wind at Denver changes from northwesterly at 5 m s^{-1} at 0500 to southeasterly at 5 m s^{-1} at 1700, with the direction at 1700 being exactly opposite to the geostrophic wind direction (Fig. 7). The magnitude of the ageostrophic wind component at 1700 is approximately 15 m s^{-1}. More will be said about this feature later.

c. *Mesoscale analysis*

Surface potential temperature (referenced to Fort Collins pressure, 840 mb) and streamline analyses over northeast Colorado at 0655, 0855, 1255 and 1655 are shown in Fig. 8. The use of potential temperature in this mountainous region assists in distinguishing that part of the surface air temperature contrast due to air mass differences from that due to elevation differences. Over the plains, temperature adjustments due to ele-

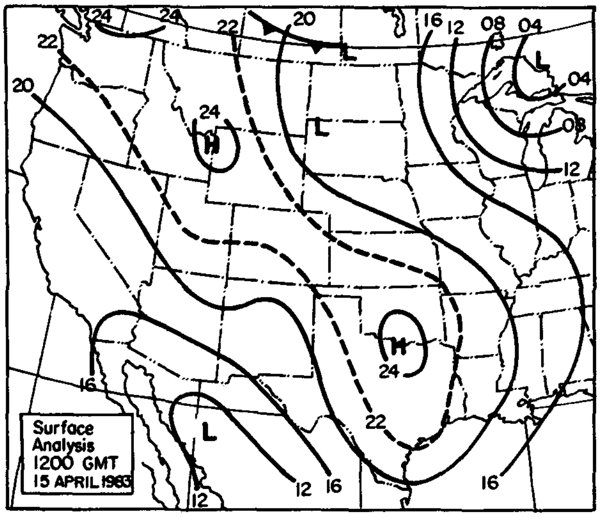

FIG. 4. Surface isobaric analysis at 0500 MST (1200 GMT) 15 April 1983 (24 = 1024 mb, etc.).

vation are small (≤2°C, except at COS, ELB and SNY where they are 3°C), while at the Rocky Mountain stations they are large. Interpretation of analyses of surface potential temperature is least complicated when the boundary layer is well mixed through a deep layer, as it is during the afternoon of the day studied here (to be shown later). The snow cover is indicated in Fig. 8 by shading and the surface water vapor mixing ratio is plotted at each station.

The analysis at 0655 (approximately 1 h 20 min after sunrise) shows a rather dramatic impact of the snow on surface temperature with a region of cold air over most of the snow fields in the South Platte River drainage basin (Fig. 8a). Relatively warm air exists in the western portion of the snow-free region which can probably be partly attributed to a weak early morning downslope flow along the foothills of the Rocky Mountains. The lower mixing ratios at LAK, RB3 and FOR support such an argument. Surface flow over the plains is characterized by drainage flow off the ridges and down the river valley, a situation typical of that expected shortly after sunrise (e.g., Defant, 1951; Johnson and Toth, 1982). At 0855 (Fig. 8b) the same temperature pattern exists, although some warming has occurred, with temperature differences as great as 5°C over a 60 km distance at the same elevation, [e.g., LVE to BGD (BRG)] due to snow cover effects. By this time the warming along the east-facing slopes of the Rocky Mountains has generated an upslope flow over the snow-free area as far east as 105°W. The timing (relative to sunrise) and pattern of the reversal to upslope flow closely follow the climatological behavior over this region on clear days (Johnson and Toth, 1982), except that on this day downslope flow off the Cheyenne and Palmer ridges persists longer than normal, presumably due to the snow cover there.

Clear skies are reported at all stations in the region at this time.

Four hours later (1255, Fig. 8c) temperatures have risen everywhere by approximately the same amount, leaving a temperature anomaly pattern similar to that before. Sky cover (in quarters) based on estimates of solar radiation depletion over clear-sky values is shown at the individual PROFS stations. It can be seen that by 1255 considerable cloudiness has developed in the west and northwest portions of the PROFS mesonet area. At the same time skies are mostly clear over the surrounding snow-covered ground. As will be discussed in more detail later (Section 4), the clouds that develop in the snow-free region are, at least initially, low clouds having bases at the top of the boundary layer. It will be shown that these clouds are a consequence of surface heating to and above the convective condensation temperature.

The surface flow at 1255 over the PROFS mesonetwork is characterized by an anticyclonic eddy which exists over the snow-free area with northwesterly flow to the north and east. This surface anticyclonic circulation has been noticed by the authors to develop frequently downstream of the Cheyenne Ridge when a northwesterly gradient wind in the lower troposphere exists over the area. It appears to be the counterpart of a cyclonic circulation centered near Denver which develops when southeasterly flow occurs over the area (Szoke *et al.*, 1984). These circulation features are very likely induced by the topography of the region. The surface circulation in the region at this time departs considerably from that normally observed on clear, snow-free days with weak lower-tropospheric winds (Johnson and Toth, 1982). On such occasions upslope flow occurs in all areas by early afternoon with flow up the ridges and toward the Continental Divide. The

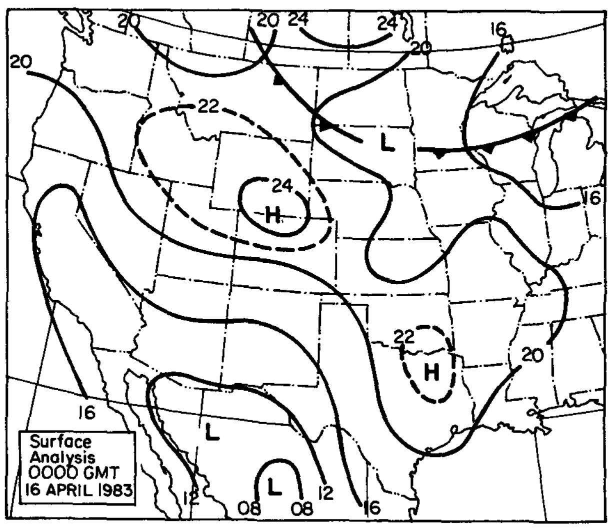

FIG. 5. Surface isobaric analysis at 1700 MST 15 April 1983 (0000 GMT 16 April).

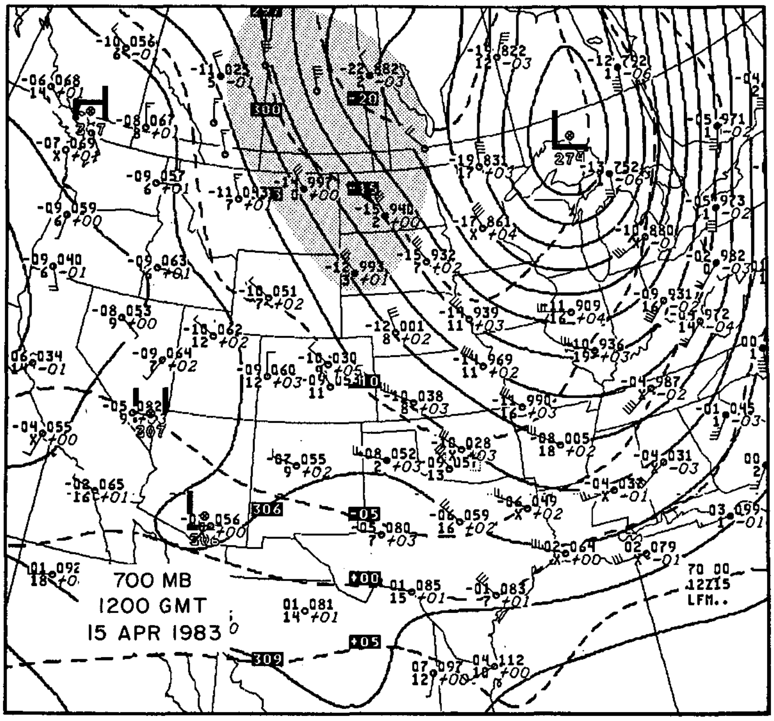

FIG. 6. 700 mb analysis at 0500 MST (1200 GMT) 15 April 1983. Shaded area denotes dew point depressions ≤ 5°C.

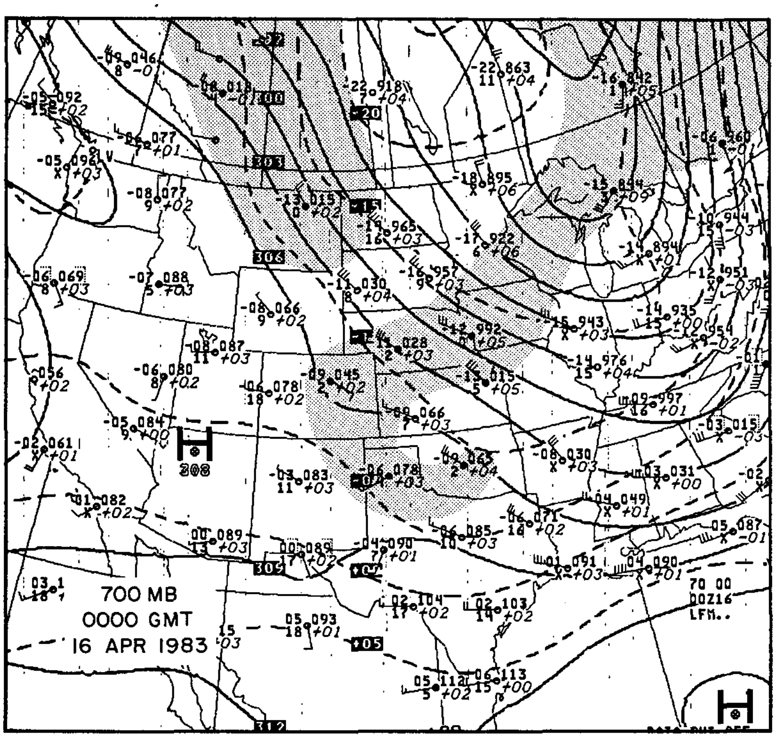

FIG. 7. As in Fig. 6 but for 1700 MST 15 April 1983 (0000 GMT 16 April).

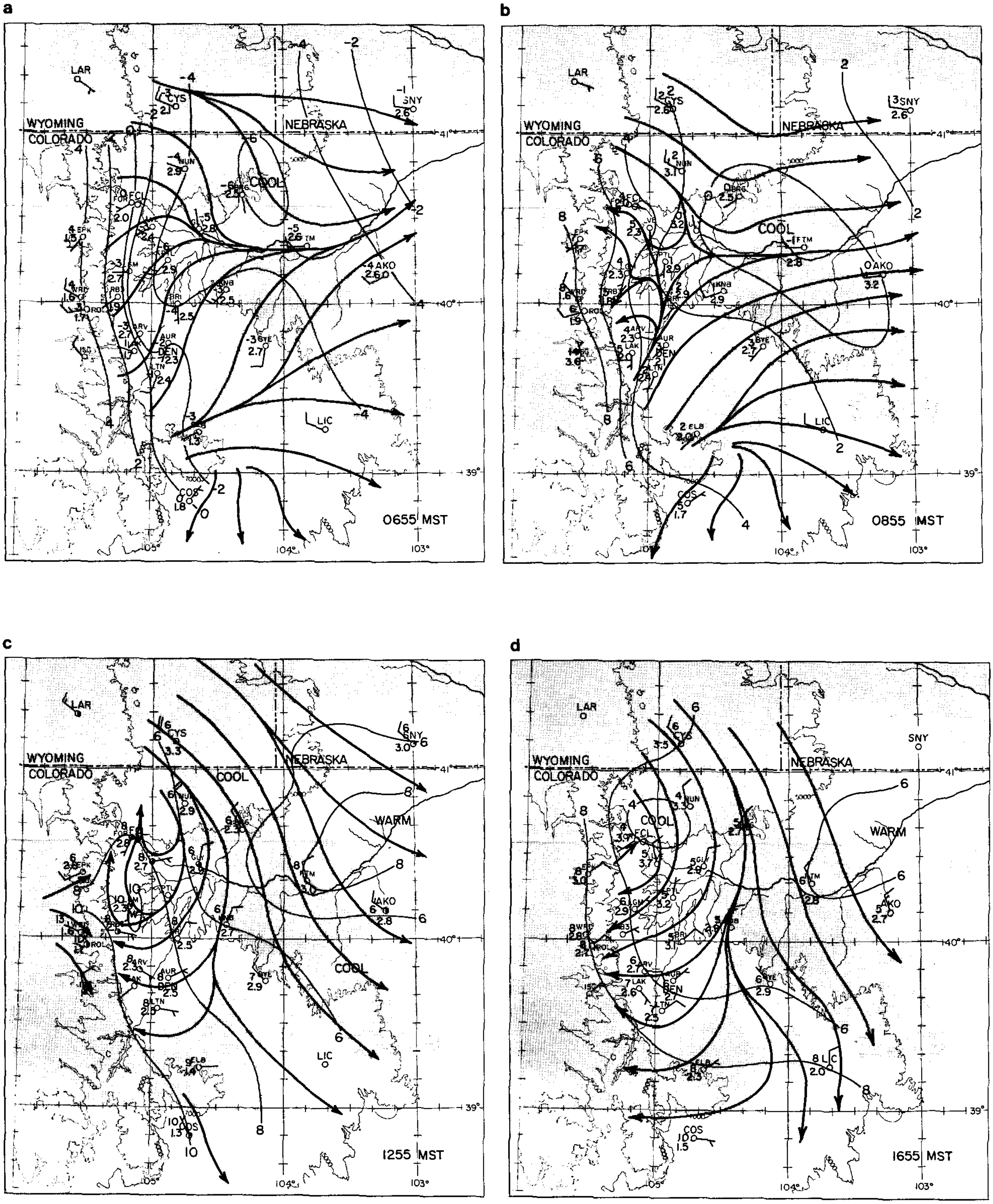

FIG. 8. Surface potential temperature and streamline analyses at (a) 0655, (b) 0855, (c) 1255 and (d) 1655 MST. Potential temperature (referenced to 840 mb) in °C. Surface mixing ratio (g kg^{-1}) is plotted below the potential temperature. One full wind barb = 5 m s^{-1}, one-half barb = 2.5 m s^{-1}. Cloud cover at PROFS stations (in quarters) is estimated from radiation traces. Cloud estimates at PROFS stations are not available at 1655 MST due to low sun angle. Shading indicates snow cover.

extension of the cool anomaly southeastward from Cheyenne (CYS) is undoubtedly aided by cold advection in the northwesterly flow there. The actual daily temperature trace at each station is a complicated function of cloudiness, local snow cover and the fetch of the air upstream over snow-covered ground.

By late afternoon (1655 or 1 h 45 min prior to sunset) the same general surface flow and temperature anomaly patterns exist; however, the amplitudes of the temperature anomalies are somewhat reduced (Fig. 8d). The reduction in the relative warmth of the snow-free region is largely accomplished by a considerable increase in cloud cover and the development of scattered snow showers over this area. The reduced surface temperatures and increased surface mixing ratios in the northern part of the snow-free area are probably due to evaporation of precipitation in that region. It can also be noted that by 1655 the upslope flow has extended to the level of the highest PROFS stations. The climatological study of Johnson and Toth (1982) indicates that a reversal to downslope normally occurs at the high elevation stations by this time in the late afternoon. Thus, it appears that the anticyclonic circulation over this region was relatively strong and deep on this day.

4. Analysis of the development of the cloud field

Within the enclosed snow-free region only two regularly-reporting National Weather Service stations exist: Denver (DEN), a first-order hourly-reporting station, and Fort Collins (FCL) a secondary two-hourly-reporting station. A list of cloud and weather portions of the aviation weather observations at these stations during the course of the day is given in Table 2. Low cloudiness develops by midmorning at both DEN and FCL with rain and snow showers occurring at FCL during midafternoon. The low clouds disappear shortly after sunset (1840).

TABLE 2. Surface observations of clouds (heights of bases in km), weather and remarks for Denver (DEN) and Fort Collins (FCL) on 15 April 1983.*

Time (MST)	DEN	FCL
0700	CLR	CLR
0800	CLR	
0900	CLR	CLR (FEW CU W)
1000	CLR (CU SW-NW)	
1100	CLR (FEW CU 1.8 KM)	2.0 SCT 5.8 BKN
1200	CLR (CU SW-NW)	
1300	CLR (CU ALQDS)	1.2 SCT E1.8 BKN 5.8 BKN (VIRGA E-SE, SWU SW-WSW)
1400	1.4 SCT	
1500	1.4 SCT 7.3 SCT	E1.2 BKN 1.8 OVC RW-SW- (RWSWB10, BINOVC N)
1600	1.4 SCT 7.3-SCT	
1700	1.4 SCT 7.6-SCT	1.2 SCT E1.8 OVC (RWSWE1520)
1800	1.4 SCT	
1900	1.4 SCT	1.2 SCT E1.5 OVC (SWU W-E, BINOVC S-N)
2000	CLR (SC SW-NW)	
2100	CLR (SC W)	3.0 SCT
2200	CLR	
2300	CLR	CLR

* CLR—Clear; CU—Cumulus; SCT—Scattered; BKN—Broken; ALQDS—All quadrants; RWSWB—Rain/snow shower began; BINOVC—Breaks in overcast; RWSWE—Rain/snow shower ended; SWU—Snow shower of unknown intensity.

A view of the cloud cover development over the PROFS mesonetwork area is possible from an examination of station solar radiation data. In Fig. 9 solar radiation traces are depicted in three divisions: 1) the snow-free area, 2) the mountain area and 3) the snow-covered area. Five-minute average values are indicated by crosses and the range of values by horizontal bars. Over the snow-free and mountain stations, cloud cover develops in most cases within two hours of local noon and continues until sunset. In sharp contrast, PROFS stations in the snow-covered region show very little, if any, cloud cover throughout the day.

From the curves in Fig. 9, the time of initial cloud cover has been estimated at each station and the results are presented in the form of isochrones of first cloud cover in Fig. 10. Low clouds first occur at about 1000 in the foothills of the Rocky Mountains near EPK and in the snow-free area near FOR/FCL. Cloud cover gradually expands eastward in the snow-free region with an easternmost extension along the South Platte River. The correlation between the area of the first low cloud development and the snow-free area is significant.

To understand why the cloudiness progressed as it did, it is instructive to examine the 0500 and 1700 soundings at DEN (Fig. 11). The sounding at 1700 terminates at 528 mb. In the morning a temperature inversion exists in the lowest 400 m with a slightly stable lapse rate above extending to 630 mb (2.2 km). By late afternoon (1700) surface heating has established (above a surface superadiabatic layer) a layer approximately well-mixed in potential temperature and specific humidity to 670 mb (1.8 km). There is a considerable increase in lower tropospheric moisture ($\sim$1 g kg^{-1} below 700 mb) from morning to evening, probably as a result of the combined effects of large-scale advection (Figs. 6 and 7) and evaporation and sublimation at the surface. During the same time a slight warming ($\sim$1–2°C) occurs above the mixed layer.

Keeping in mind the moistening which has occurred, the morning (0500) Denver sounding has been used as the environmental sounding to estimate a convective

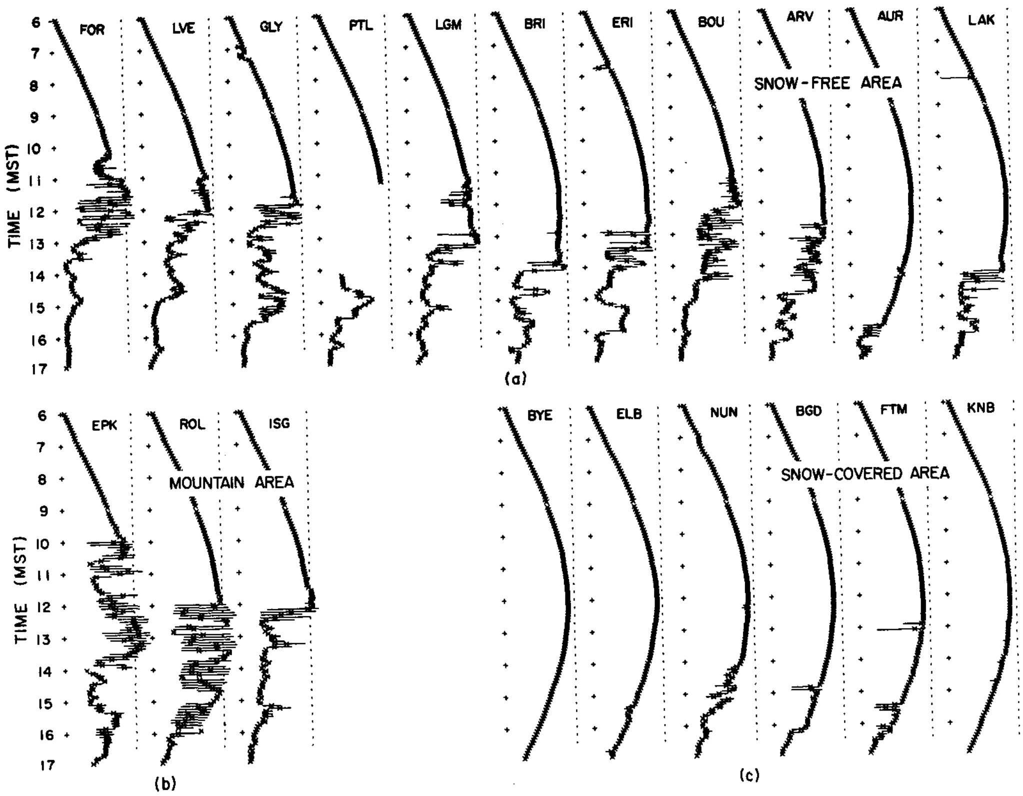

FIG. 9. Solar radiation traces (direct and diffuse on a plane surface) at PROFS stations in the (a) snow-free area, (b) mountain area and (c) snow-covered area. Pluses mark zero and vertical dashed lines 1000 W m^{-2}. Mean values at 5 min intervals are indicated by crosses and the ranges between high and low values during 5 min periods are denoted by horizontal lines.

condensation temperature (CCT) and convective condensation level (CCL; Iribarne and Godson, 1981) which should give some indication of the timing and altitude of the first cumulus clouds at each PROFS station. In the analysis, the initial parcel temperature and dewpoint used to determine the CCL at each station includes adjustments based on observed lapse rates (2 and 1°C, respectively) in the superadiabatic layer of the 1700 DEN sounding. A map of isochrones of the calculated time of first clouds using the PROFS surface data and DEN sounding data is shown in Fig. 12. The general character of the isochrone pattern agrees favorably with the observed first cloudiness shown in Fig. 10. The greatest discrepancies occur in the southern and eastern parts of the region near the border of the snow field. There were several sources for error in this analysis, including 1) the use of the morning sounding as the environment, particularly at stations to the east with later first-cloud times (recall that 1–2°C warming above the mixed layer occurred during a 12 h period), and 2) the fact that the lapse rates of temperature and dew point near the surface over snow-free ground as given by the DEN sounding may be different from those over the snow surface. The agreement suggests that the development of clouds in the snow-free area is simply a consequence of the faster rise of the surface temperature and growth of the mixed layer in the region such that mixing to the CCL was more quickly achieved there. Development of the mixed layer to the CCL was retarded in the snow-covered regions.

The explanation given above is supported by cloud-base height estimates obtained from surface temperature and dew point measurements (e.g., Iribarne and Godson, 1981, pp. 140–141). These values, determined at the time of the first low clouds (inferred from radiation traces) and shown in Fig. 13 in the region of low cloud cover, range from a maximum of 2.3 km centered in the snow-free area to a minimum of 1.3 km over the snow-covered ground (near NUN). The deeper subcloud layer in the snow-free area is likely a consequence of higher surface temperatures and a slightly reduced surface moisture source there, whereas the shallower depths occur over the cooler, slightly moister, snow-covered surface (Fig. 8c). The agreement between these cloud base estimates and those reported at DEN and FCL (in parentheses) is reasonably good.

A plot of the surface air saturation points (Betts,

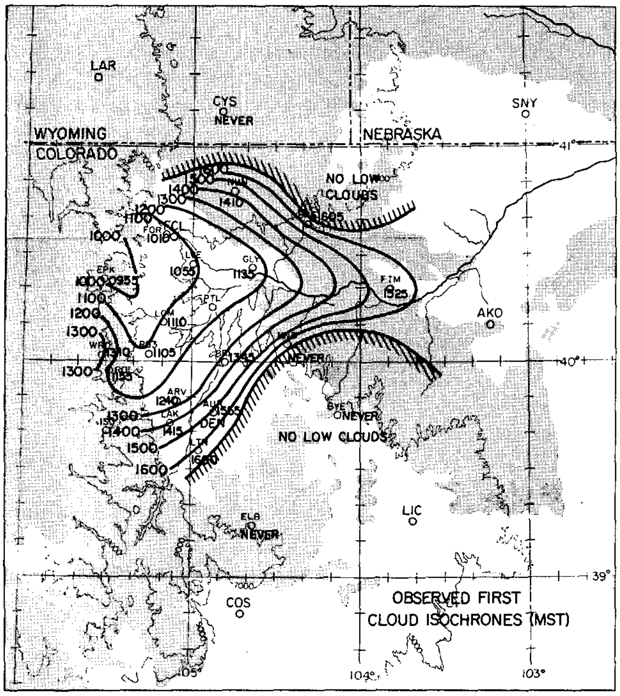

FIG. 10. Isochrones (MST) of first low clouds determined from PROFS radiation traces (Fig. 9). Shaded area is snow cover.

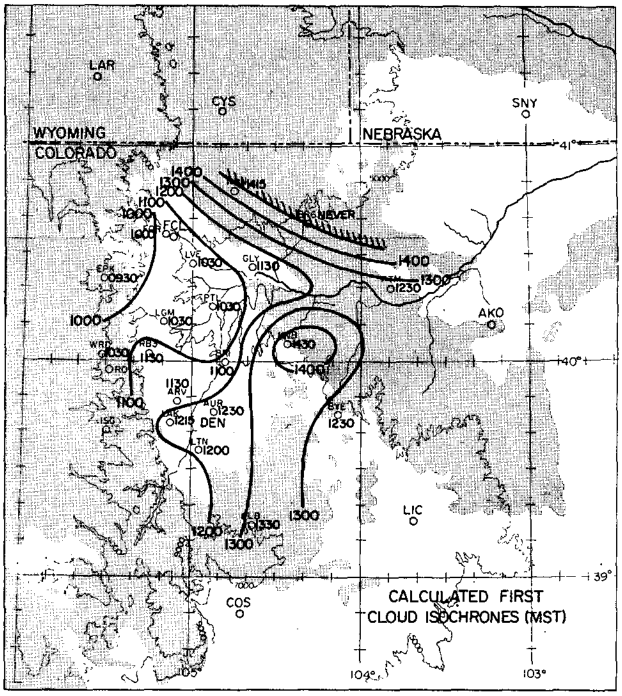

FIG. 12. Isochrones (MST) of first low clouds calculated from PROFS surface temperature and dew point measurements and Denver 0500 MST sounding data. Shaded area is snow cover.

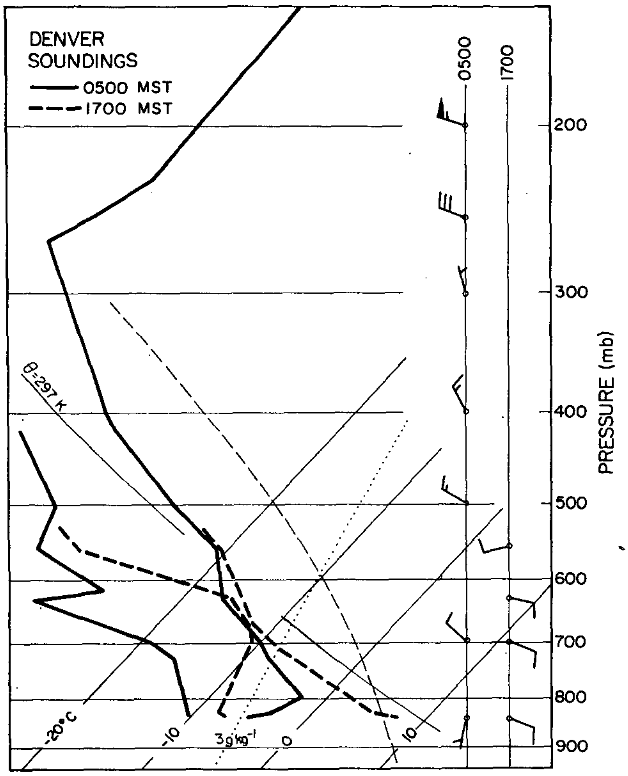

FIG. 11. Denver soundings at 0500 (solid) and 1700 MST (dashed) 15 April 1983. For plotted winds one full barb = 5 m s^{-1}.

1982) or lifting condensation level (LCL) of the surface air as a function of the LCL temperature at the time of first cloud is shown in Fig. 14. Most of the saturation points are within 1°C of the afternoon DEN sounding curve and all except one (RB3) lie to the right of this curve. The slight displacement to the right of the sounding curve is a consequence of the superadiabatic surface layer, which is not considered in the method to determine the LCL. The results here suggest that if parcels rose from the surface into clouds without mixing, they would be slightly warmer (~1°C) than their environment at cloud base. It is likely that 1) mixing occurs and 2) much of the air rising through cloud base has its origins above the surface superadiabatic layer such that cloud base environment temperature differences are small (e.g., Betts *et al.*, 1974).

The actual surface temperatures, when plotted on this graph, show a considerable departure from the DEN sounding (they are warmer) at the higher-elevation stations. This surface-to-free-atmosphere temperature difference along the slopes of the Continental Divide drives the upslope flow, which eventually leads to a wind reversal to upslope at the highest stations at 1655 (Fig. 8d).

5. A possible circulation enhancement mechanism and feedback to larger scales

The establishment of a horizontal temperature gradient in the mixed layer along the boundary of the

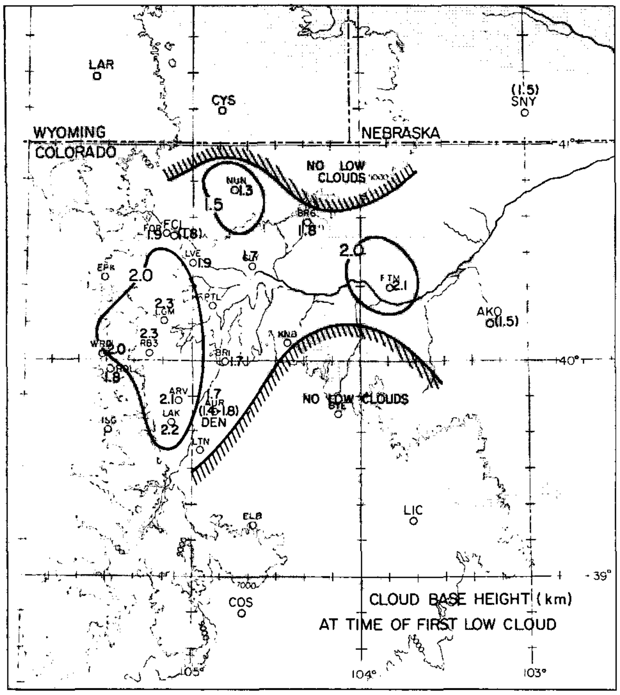

FIG. 13. Cloud base height (km) at PROFS stations calculated from surface temperature and dew point measurements at time of first low clouds (inferred from radiation traces). Numbers in parentheses are reported heights from aviation weather reports. Shaded area is snow cover.

snow-free region could conceivably contribute to a solenoidal field of sufficient magnitude to influence the local circulation of the region. (The effect of change in surface roughness from snow-covered to bare ground for this problem should not be significant.) The ensuing circulation would be analogous to a sea breeze (Haurwitz, 1947) or inland sea breeze (Sun and Ogura, 1979) and might be termed a "snow breeze." The sense of the circulation would be to enhance the low-level inflow into the snow-free region (and consequently convergence in this oval-shaped area), thus leading to lifting on the mesoscale. Mesoscale lifting over the snow-free region, likely superimposed on weak synoptic-scale subsidence, may have aided the more rapid growth of the mixed layer and the earlier onset of cloudiness there.

The cloud development in the snow-free region was sufficiently deep to produce snow showers and, in a larger sense, could be considered a natural mechanism to equalize the distribution of snow cover over a region. The snow showers that developed in this case were light, however, and did not contribute to any significant new snow cover.

It is conceivable that the convection in the snow-free region, accompanied by the formation of clouds and precipitation and coupled with the mesoscale circulation feature described above, could have contributed to an enhancement of the upslope flow over the eastern half of the snow-free region during the afternoon. While a snow boundary also existed along the western portion of this region, the effects there of the steep-sloped topography and elevated heat source provided by the Rocky Mountains on local solenoidal circulations should have dominated those produced by contrasts in surface properties. An additional important contributor to this enhancement appears to have been the development of the topographically-induced, low-level anticyclonic circulation over the region (Fig. 8c). Little is known about the depth of such circulations under normal circumstances; however, we postulate that on this day the local forcing mechanisms described above may have led to the deepening of the upslope flow to near the 600 mb level at 1700 at DEN (Fig. 11). Normally, on undisturbed days with weak gradient winds in the lower troposphere, the amplitude of the modulation of the wind at 600 mb by the upslope circulation does not exceed 1 m s^{-1} (Modahl, 1979), whereas in this case an ageostrophic wind component of $\sim$15 m s^{-1} has been noted (see discussion of Fig. 7). This topographically-induced mesoscale circulation feature modified by the "snow breeze" effects may be one possible explanation for the unusual Denver observation at 700 mb at 1700 (Fig. 7) which indicates a wind direction opposite that of the gradient wind. Following careful study of all synoptic data, we have been unable to find any evidence that this wind reversal may have been accounted for

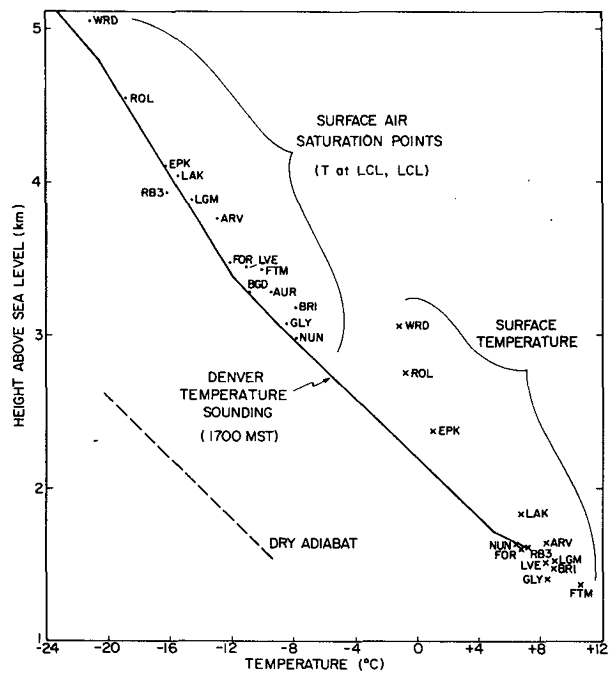

FIG. 14. Surface air saturation points (LCL of surface air and temperature at LCL, dots) and surface temperatures (crosses) as a function of height above sea level at the time of first cloud at each PROFS station. Denver sounding at 1700 MST is indicated.

by the influence of a disturbance at upper levels, e.g., a short-wave trough from the west. However, even if a subsynoptic disturbance may have played some role in this circulation, we propose that the snow cover and topographic effects were necessary to bring about the observed enhancement of the ageostrophic circulation. The process described here represents just one of the ways in which variable underlying surface properties, originally established by large-scale circulation features, may, in turn, feed back to and alter the large-scale flow.

6. Summary and conclusions

In this study, data from the PROFS (Program for Regional Observing and Forecasting Services) surface mesonetwork have been used to document the effect of variable snow cover on boundary-layer properties, cloudiness and weather conditions over northeast Colorado. On 15 April 1983, an oval-shaped $\sim 10^4$ km^2 area of snow-free ground surrounded by snow-covered ground existed along the Colorado Front Range. While sky conditions on the morning of this day were everywhere clear, cloudiness developed by midday over the snow-free region as a result of the more rapid boundary-layer heating and mixed-layer growth there. During mid-afternoon, snow showers occurred over the snow-free ground whereas skies remained mostly clear over the snow-covered area. While other winter days with surface temperature contrasts due to snow cover greater than those observed on 15 April can be readily found, the events on this day are worthy of note because the stability was sufficiently weak to permit the effects of the variable snow cover to be manifested in the form of cloudiness and precipitation.

It has been suggested in this study that snow boundaries in the region (which probably moved slowly with time as melting occurred) may have acted through the development of a weak solenoidal field to enhance low-level inflow into the snow-free area, thereby assisting with cloud development in the region. This phenomenon, which is analogous to the sea breeze, may be appropriately termed a "snow breeze." Even without such an enhancement to the circulation, the variable snow cover, through its impact on the surface energy budget, had a profound effect on the regional weather conditions on this day. This situation represents just one example of a class of complex interactions and feedback processes involving variable surface properties and the large-scale flow. This study points to possible improvements in short-term (0–12 h) weather forecasts in regions of variable snow cover when conventional synoptic, satellite and mesoscale station network data are required, integrated and evaluated in a timely manner.

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