Mesoscale Anticyclonic Circulations in the Lee of the Central Rocky Mountains

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ABSTRACT

Composite analyses of terrain-forced, mesoscale anticyclonic circulations over southern Wyoming and northern Colorado are constructed. These suggest two different types of circulations, based on upstream flow direction. The air within type 1 circulations originates to the west of the Continental Divide and contains little moisture. Type 2 circulations form in more moist, northerly flow and are sometimes associated with snowbands. Both types tend to form during the afternoon and dissipate after sunset, although type 2 events may follow frontal passages and occur at night.

Case studies of one event of each type suggest that anticyclonic vorticity generation occurs within the lowest kilometer above ground level when that layer is nearly vertically mixed in both potential temperature and velocity. An analogy with vorticity generation in mixed-layer models is considered, and it is shown that the conditions for generating negative vorticity in those models are satisfied in each observed case. The mixed-layer mechanism is also favored as it naturally explains the diurnal tendency of the circulations and may therefore explain the observed, late-day snowfall maximum along the Front Range of Colorado.

1. Introduction

At the core of an emerging picture of the mesoscale response to orography is the occurrence of lee vortices and wakes. Lee vortices have at least two important consequences on sensible weather. The first is transport. Local flow reversal implies a radical difference in the motion of atmospheric constituents from what one would infer from the larger-scale motion. Such flow reversal is common in the lee of mountainous islands such as Hawaii (Nickerson and Dias 1981). Perhaps best observed in the lee of tropical islands (Etling 1989; Smith et al. 1997), numerous examples of terrain-induced vortices over continental regions have recently been reported with the aid of improved surface observing networks (Szoke et al. 1984) and mesoscale model simulations (Sun and Chern 1993; Georgelin et al. 1996).

The second effect is on clouds and precipitation resulting from low-level convergence and or "upslope" flow forced by lee circulations. This can completely change the distribution of precipitation from what one would infer based on the large-scale wind flowing up or down the local terrain slope. For instance, Wesley et al. (1995) documented a case where snowfall was locally enhanced associated with low-level convergence zones within a topographically induced anticyclone. Convergence zones induced by orography have been documented in many other areas, such as Puget Sound in Washington (Ferber and Mass 1990), the Denver metropolitan area (Szoke et al. 1984), and others.

Considerable progress in understanding lee circulations has been made through laboratory studies (e.g., Hunt and Snyder 1980) and nonlinear simulations (Smolarkiewicz et al. 1988; Smolarkiewicz and Rotunno 1989a, SR hereafter) of low–Froude number flow. The Froude number is defined Fr = U/Nh, where U is the flow speed, N the Brunt–Väisälä frequency, and h the height of the topography. The nonlinear simulations help distinguish two mechanisms of vorticity generation that can produce similar vortices: viscous boundary layer effects and inviscid tilting of horizontal vorticity. The numerical simulations showed that vortices with reasonable correspondence to observations in the lee of the Hawaiian mountains (Grubisic and Smith 1993) can be produced in inviscid, low–Froude number flow.

Lee vortices can also form when Fr is large. Wilczak and Glendening (1988, hereafter WG) performed simulations of a terrain-induced cyclonic vortex (the Denver cyclone) using a mixed-layer atmospheric model. In such a model, velocity and potential temperature are assumed well mixed within a layer adjacent to the surface, capped by an inversion. The Froude number Fr is undefined in the mixed layer. However, there is a related parameter that characterizes the behavior of a mixedlayer model. It is defined as the ratio of flow speed to

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the phase speed of an interfacial gravity wave $Fr_m = U/(g'H)^{1/2}$, where g' is reduced gravity and H is the undisturbed mixed-layer depth. Simulations with small Fr_m produce similar results to stratified flow models with small Fr, but the mechanism of vorticity generation is different. In a mixed-layer model, the stress terms generate vorticity, as opposed to the inviscid generation demonstrated by SR in stratified flows. Even viscous generation of vorticity in stratified flows proceeds quite differently from that in mixed-layer models, as discussed further in section 3.

Without background rotation, uniform upstream flow characterized by small Fr in the stratified case, or small Fr., in the mixed-layer case, will engender an antisymmetric vortex pair in the lee of a circular mountain. The addition of background rotation breaks this symmetry and imposes another nondimensional parameter necessary to describe the system, the Rossby number, Ro = U/fL, where f is the Coriolis parameter and L the topographic length scale. Decreasing the Rossby number from large values through one and then to small values, at low Froude number, one sees the transition from cyclonic and anticyclonic lee vortices of comparable strength to an anticyclonic vortex over the mountain that is geostrophically balanced. At moderate Rossby number the lee response is complicated, even in idealized simulations, consisting of strong inertia-gravity waves and the emergence of a relatively dominant anticyclonic lee vortex (Boyer et al. 1987; Peng et al. 1995).

The bulk of our knowledge about lee-vortex formation has come from numerical simulation and laboratory experiments. It is important to document the occurrence of such circulations in the real atmosphere with observations suitable for diagnosing the dynamical mechanisms likely at work. In addition to possibly confirming theoretical results, and thereby improving our understanding of this complicated phenomenon, such work is important for better interpretation of operational forecast model output and improved precipitation forecasts near complex terrain. Current forecast models resolve only partially the important terrain features that generate these lee circulations. Such features are often less than 100 km in scale. Toward this goal, the present paper documents the composite structure of anticyclonic circulations observed in the lee of the central Rocky Mountains. Such circulations have been termed "Longmont anticyclones," presumably because the center of anticyclonic circulation tends to lie near Longmont, Colorado (Figs. 1 and 2). Prediction of snowfall resulting from these circulations is a particularly difficult forecast problem because the prevailing flow in vortex cases is often northwesterly. Such flow ordinarily produces downslope motion, suppressing precipitation. Invariably, operational forecast models indicate little or no snowfall for the area in these situations.

Data for our study were collected during the Winter Icing and Storms Project (WISP94), which occurred be-



FIG. 1. Reference map for the study area. Terrain, contoured in 400-m intervals, is dashed and represents about 13-km horizontal resolution. Gray, dashed line indicates the Continental Divide. Three-letter station identifiers are printed for stations referred to in text; MBW—Medicine Bow, Wyoming; CGW—Chugwater, Wyoming; CYS—Cheyenne, Wyoming; DEN—Denver, Colorado; CRG—Craig, Colorado; PLT—Platteville, Colorado; KIM—Kimball, Nebraska. MC1 and MC2 are mobile CLASS vehicles (here, MC1 is at Glendo, Wyoming, highway) are denoted by asterisks, CLASS soundings by squares (filled for mobile CLASS), NWS soundings by crosses, and profilers by triangles.

tween 25 January and 25 March 1994. Scientific objectives of the experiment (Rasmussen and Cooper 1994) primarily focused on ice initiation in both lenticular and upslope clouds over northeastern Colorado, southeastern Wyoming, and southwestern Nebraska (see Fig. 1). Intimately related to the production of snow and supercooled liquid water are mesoscale frontal and terrain-induced upward motion in the region.

Facilitating our study is the addition of special sounding data, especially upstream from important topographic features; aircraft data; radar; and supplementary surface data. These data allow us to distinguish various mechanisms of vorticity generation associated with flow over/around complex terrain. Our results suggest that these shallow, topographically induced circulations are produced by a mechanism similar to that operating in mixed-layer models of vorticity generation by flow around obstacles. This result, as discussed in section 4, possibly explains the noted late afternoon snowfall maximum in Denver (Weismueller 1982). It also points to the importance of realistic boundary layer treatment in mesoscale models in order to correctly predict orographically induced low-level winds and precipitation, especially in the presence of weak, synoptic-scale systems.

2. Composite structure

a. Data

Conventional data include the NOAA Program for Regional Observing and Forecasting Services (PROFS) mesonet, standard surface airways observations (SAOs), and National Weather Service (NWS) soundings from Denver as well as the 404-MHz profilers from Platteville, Colorado, and Medicine Bow, Wyoming. Special data obtained for WISP94 include the Wyoming Highway Department surface stations, Cross-chain Loran Atmospheric Sounding System (CLASS) soundings (five fixed sites and two mobile sites), 5-min surface observations at each CLASS site, and enhanced Doppler radar coverage. In addition, aircraft data are available from both the NCAR Electra and Wyoming King Air, which flew missions to examine the microphysics within lenticular clouds.

To better understand the mean structure of the anticyclonic circulations observed during WISP94, we will first present composite analyses, then in section 3, we will review individual cases. In compositing surface data, we will restrict ourselves to the 5-min PROFS data and the Wyoming highway network (see Fig. 1). These two surface networks are able to resolve important features of the anticyclonic circulations and greatly enhance the coverage of SAOs.

It is well known that the Longmont anticyclone forms when the synoptic flow is northerly or northwesterly at low levels (Wesley et al. 1995). South of the Cheyenne Ridge (Fig. 1) the flow often acquires an easterly component as it turns anticyclonically and slows. Based on these characteristics we define a Longmont anticyclone as a circulation where the average winds over the northern part of the mesonet have a northerly component and winds over the southwestern part of the mesonet have an easterly component (Fig. 2). We also require at least a 45° clockwise turning of the wind between the northern and southern stations. No criteria demand a closed circulation be present. In fact, many cases do not feature a closed circulation, but strongly affect local weather nonetheless.

It is also known that these anticyclonic circulations can persist for several hours. However, when employing arbitrary thresholds to define events, small fluctuations in winds may prevent the anticyclone criteria from being satisfied continuously over the lifetime of an event. We therefore do not require that the criteria be satisfied continuously during an event; rather, we place a restriction on the minimum time that an anticyclonic circulation is continuously *not* present. Thus, periods of time during which conditions are fluctuating about the threshold are considered events.

In practice, the criteria are usually satisfied continuously over a few hours, especially for stronger events. Using 5-min data, we require that the anticyclone criteria are satisfied 20 times for event designation and inclusion of that event in our composite. The minimum length of time between events is arbitrarily assigned to 3 h, but, in practice, events are all separated by 12 h or more. A composite of each event is formed by averaging those time periods that satisfy the anticyclone criteria.



FIG. 2. Mesonet stations used to determine the existence of an anticyclonic circulation. Stations in gray lettering were required to have a northerly wind component; stations with black lettering were required to have an easterly component. For northern stations, CGW—Chugwater, Wyoming; KIM—Kimball, Nebraska; NUN—Nunn, Colorado; BGD—Briggsdale; FTM—Fort Morgan; AKR—Akron. For southern stations, BOU—Boulder; LGM—Longmont; ERI—Erie; BRI—Brighton; KNB—Keensburg; AUR—Aurora; ARV—Arvada; LAK—Lakewood; LTN—Littleton. Terrain plotted as in Fig. 1.

The grand composite of all events is simply the average of the composites.

There are a few subtle aspects of compositing the surface mesonet data. First, we wish the composites to accurately represent the horizontal variation of fields such as wind, pressure, and temperature. Because not all stations report data at all times, simply averaging individual stations, independently, may produce spurious gradients of fields in the composite. For this reason, we identify the station that reports most often, call it station "X," and composite the *difference* of variables relative to that station. When the average of X is added to the composited differences, a representation of the total fields, with gradients preserved, results.

Another subtlety is the analysis of horizontal pressure gradients. The significant variations of elevation across the WISP94 region make reduction of surface pressures to a common altitude a challenging task. Furthermore, the biases of pressure reports from individual mesonet stations often swamp the signal of the true horizontal pressure variations, usually less than 1 mb across the domain. To mitigate these problems, we compute a mean surface pressure for each station for each hour of the day over the 2-month period of WISP94. We then subtract this pressure field from the total and reduce the



FIG. 3. Winds and 1600-m perturbation pressure analysis for the composite of all 19 anticyclone cases. Plotting convention for winds uses a long barb to represent 2 kt and a pennant to represent 10 kt. Contour interval for pressure is 0.2 mb. Terrain (dashed) is plotted as in Fig. 1. Pressures from stations above 2500 m were not analyzed.

resulting pressure perturbations to a common reference altitude, 1600 m MSL. This is done using the hydrostatic equation $\Delta p' = -\rho' g \Delta z$, where the density perturbation ρ' is computed as the deviation from a similar 2-month mean of density for each hour of the day. Here, g is gravity and Δz is the difference between the surface station altitude and 1600 m. This treatment of pressure removes most of the systematic station pressure errors and the average diurnal cycle of pressure as well.

b. Results

During WISP94, we found 19 anticyclone cases. Their composite surface wind and perturbation pressure fields are displayed in Fig. 3. These indicate strong northwesterly flow over the Cheyenne Ridge, which turns abruptly southward and then westward over the southwestern part of the domain. The pressure perturbation field reveals mainly enhanced geostrophic northwesterlies over the northern and eastern part of the mesonet. The pressure gradient of about 1 mb per 200 km corresponds to a geostrophic wind perturbation of about 5 m s⁻¹. The actual wind in this area flows mainly up the gradient toward higher pressure and exhibits notable slowing. The appearance of easterlies over the western part of the network is qualitatively consistent



FIG. 4. Timing of the 19 anticyclone events observed during WISP94. For each case, a horizontal bar indicates its duration. Above the bar is the date. Solid bars indicate type 1 events; dashed bars indicate type 2 events.

with acceleration down the pressure gradient represented in the analysis. There is no evidence of any kind of gradient wind balance between the mass and wind fields.

These anticyclonic events show a marked preference for occurring in the afternoon (Fig. 4). Only 3 out of 19 cases occurred at other times. This suggests a profound influence of the diurnal cycle on these circulations. Events were typically 5–6 h in duration, and were dependent on the synoptic-scale flow, as evinced by the clustering of events in time. During WISP94, there were four time periods when anticyclones formed on three or more successive days. One could also consider most of the anticyclone events as occurring in two regimes, late January/early February and early March.

A representation of the mean vertical structure of the anticyclones is given by measurements from the 404-MHz profiler at Platteville, Colorado (Fig. 5). We composited hourly consensus winds for all times during events (shown in Fig. 4; about 15% of the profiler data were missing). Northwesterly flow is evident aloft, with a significant weakening and directional veering into a northerly direction over the lowest 2 km AGL. This implies that the anticyclonic disturbance resides entirely below the altitude of the Continental Divide of northern Colorado.

There appear to be two qualitatively different largescale flow patterns associated with anticyclone events. Examples from each will be discussed in detail in the subsequent case studies. The wind direction at Chug-



FIG. 5. Average of hourly consensus winds from Platteville profiler during all 19 events. Winds are plotted in the standard convention, with a long barb equal to 10 kt.

water, Wyoming, is useful for distinguishing the two types of circulations, one with predominantly westerly flow upstream from the WISP region, the other with mainly northerly flow. We identify each of our 19 cases as a westerly, type 1, event or a northerly, type 2, event, based on whether the mean wind direction at Chugwater, Wyoming, during the case is 315° or less (type 1) or more than 315° (type 2).

The wind and pressure patterns over northeast Colorado from types 1 and 2 (Fig. 6) reveal that the type 2 circulation is confined closer to the Front Range. In type 1 cases, anticyclonic shear vorticity appears to extend southeastward beyond the mesonet in the form of a shear line. The orientation of this line parallels the direction of the upstream flow. In type 2 cases, the vorticity is not so obviously elongated along the upstream flow direction (i.e., north–south). Aside from the obvious difference in the orientation of the pressure gradient across the northeast part of the mesonet, consistent with the partitioning of the wind direction at Chugwater, the highest pressure appears adjacent to the foothills south of Denver in type 2, but found much farther out on the plains in type 1.

The winds measured by the 404-MHz profiler at Medicine Bow, Wyoming, show a pronounced difference between type 1 and type 2 events at low levels (Fig. 7). The strong westerlies near the surface are a characteristic feature of the so-called wind corridor, documented by Marwitz and Dawson (1984). This is a region north of the Medicine Bow Mountains, but south of the Laramie Range and other isolated peaks in central Wyoming, through which air is channeled as it flows eastward from the Great Basin. The characteristic synoptic pattern for all events features northwesterly flow aloft, but for type 1 events, a large, quasi-stationary region of high pressure is found over the Great Basin associated with cold, stagnant air. It is the pressure gradient associated with this high that is primarily responsible for the strong winds over southern Wyoming. Given the large-scale flow differences between type 1 and type 2 events, one would expect corresponding differences in moisture and precipitation. Air that flows through the wind corridor in Wyoming reaches northern Colorado with little moisture. Northerly flow regimes can be comparatively moist, and indeed, all of the precipitation associated with anticyclonic events during WISP (five total precipitation events) occurred in the northerly, type 2, flows.

3. Case studies

While the composites summarize the mean structure of surface features for the anticyclones, aspects of the vertical structure and temporal evolution cannot be represented by composites because upstream and in situ aircraft and sounding data were not collected for the majority of events. To elucidate the dynamics that produce the anticyclonic circulations, we will now investigate individual cases that each represent type 1 and type 2, respectively. The cases were selected for their representativeness of the composites and the availability of supplementary observations.

a. A type 1 case: 1 February 1994

1) **DESCRIPTION**

The synoptic-scale weather pattern on 1 February featured northwesterly flow aloft and the commonly seen surface anticyclone over the Great Basin, with cold, low-level air (Fig. 8). Winds at low levels resembled the flows documented by Maurwitz and Dawson (1984), except that there was notable northwesterly flow extending well out onto the High Plains at 1800 UTC 1 February (Fig. 9a). By 2200 UTC, a mesoscale, anticyclonic circulation, with a scale not more than 100 km, has formed over northeast Colorado, to the south of the Cheyenne Ridge (Fig. 9b). The vortex formed just after 1800 UTC and began dissipating about 5 h later. Elements of the anticyclonic circulation persisted until around 0100 UTC 2 February.

The surface potential temperature field at 1800 UTC exhibits strong baroclinicity over northeast Colorado (note the gradient along the topography contours). As the vortex circulation intensifies, it advects potential temperature, eventually transporting colder air westward and northward. The mesonet site at Nunn, Colorado, recorded a sharp drop in temperature as this cooler air arrived from the south and southwest around 2300 UTC (Fig. 10). For all practical purposes, this feature is a cold front. Using a simple time–space conversion, we estimate the potential gradient just behind this front to







be about 4 K per 10 km, or about five times larger than the mesoscale gradient analyzed at 1800 UTC. Thus active frontogenesis occurred, initiated by the vortex circulation.

The surface wind pattern clearly suggests blocking of flow impinging on the Continental Divide from the west. A sounding from Craig, Colorado, upstream from the high terrain over north-central Colorado, at 1900 UTC confirms that the stratification is large over a deep layer (Fig. 11). Using the mean wind and stratification between 2200 and 3000 m MSL, we estimate a Froude number of 0.4, consistent with blocking. Here a mountain height of 3500 m is assumed, representative of the Continental Divide in northern Colorado. Being blocked, the flow is forced around the high terrain of northern Colorado and accelerates down the large-scale pressure gradient (see Fig. 8). The resulting, strong lowlevel winds are evident from a time series of 404-MHz profiler winds at Medicine Bow, Wyoming (Fig. 12). The low-level winds are remarkably steady despite significant changes in the flow above 3 km MSL during the day. The vertical extent of this strong, steady westerly flow at Medicine Bow corresponds well to the layer in the Craig sounding that is blocked.

Soundings available at Denver, Colorado (Fig. 13), show strong northwesterly flow at all but the lowest levels at 1200 UTC 1 February. By 0000 UTC 2 February, in addition to significant warming throughout the lower troposphere, the winds below 3 km MSL weaken considerably and even reverse near the surface in association with vortex formation (Denver is located south of the axis of the anticyclonic circulation). From this we infer that the depth of the anticyclonic perturbation is on the order of 1 to 1.5 km. Above this (3 km MSL), the winds do not change much during the day.

Aircraft data, from both the NCAR Electra and Wyoming King Air were available on this day. Both planes flew missions in the late afternoon to examine lenticularcloud microphysics. Such clouds were not near our domain of interest. Surface stations over northeast Colorado and southeastern Wyoming reported clear skies. Data most useful for our study were collected during the Electra's ferry from Boulder–Jefferson County airport (BJC) to Cheyenne during the period 2200 to 2230 UTC. The flight track and corresponding time series are displayed in Figs. 14 and 15, respectively. Much of the flight occurred near 2550 m MSL (about 1 km AGL). During this portion of the flight, the aircraft flew nearly orthogonally to the wind direction. If we assume that the variation of the flow across the flight path is rela-

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FIG. 6. (a) Winds and 1600-m perturbation pressure analysis for the type 1 composite. Plotting convention for winds is as in Fig. 3. Contour interval for pressure is 0.2 mb. (b) As in (a) but for the type 2 composite. Terrain is as in Fig. 1. Pressures from stations above 2500 m were not analyzed.



FIG. 7. Composite wind profiles from the Medicine Bow, Wyoming, 404-MHz profiler for type 1 (heavy lines) and type 2 (thin lines) events.

tively small, we can infer the vertical component of vorticity from a time series of aircraft-measured wind.

A prominent region of negative relative (and absolute) vorticity is noted beginning about halfway into the flight, with a mean value of absolute vorticity along a 30-km segment (2222–2227 UTC) of at least -2f and locally even more negative. Within this shear region, the temperature falls steadily to the north, with a gradient averaging about 1 K per 20 km. It is also apparent that the mesoscale region of anticyclonic shear encountered by the aircraft is directly above the horizontal shear



FIG. 8. The 500-mb winds and potential temperature (black lines) and sea level pressure (gray lines) for 0000 UTC 2 February 1994. Contour interval for θ is 3 K, for sea level pressure is 4 mb. Labels for pressure contours omit hundreds digit(s).



2200 UTC 1 FEBRUARY

FIG. 9. Surface winds and potential temperature for (a) 1800 UTC and (b) 2200 UTC 1 February. Contour interval for θ is 2 K. Terrain is as in Fig. 1. Shaded region in (a) corresponds to area of possible mixed-layer vorticity generation (see section on dynamic mechanisms for details). Boxed region in (b) is inset for Fig. 14.

observed at the surface (Fig. 14). The shear region features significant nonuniformity in vorticity with localized jets of 5–7 m s⁻¹ about 15 km apart. These have no apparent correspondence with changes in wind direction. Turbulence, as indicated by the magnitudes of the vertical velocity (not shown), was more intense in the mesoscale region of large anticyclonic vorticity than elsewhere.

The King Air flight took place over southern Wyoming at nearly the same time as the Electra flight. The



FIG. 10. Time series of potential temperature (solid) and perturbation pressure (dashed) from Nunn, Colorado (see Fig. 2 for location).

significant part of the flight, for our purposes, was a descent near Medicine Bow at about 2250 UTC. This location is about 100–200 km upstream from where the vorticity generation appeared to occur. As such, it is about the only representative upstream sounding in this case. A scatterplot of θ during the descent reveals a notable inversion at 3 km MSL (Fig. 16). Thus, the inversion height at Medicine Bow corresponds to the top of the layer of blocked flow based on the Craig



FIG. 11. Profiles of wind and θ from Craig, Colorado, at 1900 UTC 1 February (local noon). Here, Fr is Froude number, and a curly bracket indicates the layer over which the Froude number is estimated.



FIG. 12. Time series of hourly consensus winds at Medicine Bow, Wyoming, during 1 February. Time is UTC.

sounding and the vertical extent of the anticyclonic disturbance over northeast Colorado. Beneath the inversion, the lapse rate is only slightly stable, averaging about 1.5 K km⁻¹, and wind speed is nearly uniform, with only a subtle backing from westerly to southwesterly. The speed close to the ground is still greater than 20 m s⁻¹.

2) Dynamic mechanisms

Openly debated in the literature are the mechanisms by which vertical vorticity produced as a flow, initially



FIG. 13. Profiles of potential temperature (K) and wind from Denver, Colorado, at 1200 UTC 1 February and 0000 UTC 2 February.



FIG. 14. Local surface winds and θ analysis at 2200 UTC 1 February (see Fig. 1 for area) along with NCAR Electra measurements of wind and θ . Potential temperature is plotted as $\theta - 280$ K. Heavy dashed line indicates roughly constant altitude portion of flight (2500 m MSL), whose time series appears in Fig. 11. For this plot, aircraft data were filtered to remove variations less than 120 s.

with no vorticity, impinges on a mountain. We consider here three mechanisms that are distinct at a fundamental level: boundary layer mixing, gravity wave breaking, and vortex tilting. Vortex compression, a fourth possible mechanism, is an unlikely source of circulation. This assertion is based on scaling arguments (see Dempsey and Rotunno 1988) and the fact that downslope flow, prevalent where vorticity appears to originate, cannot create negative absolute vorticity when it is initially positive. The negative absolute vorticity must come from one of the other mechanisms.

Wave breaking and inviscid baroclinic generation can both create vertical vorticity in flows characterized by



FIG. 15. Time series of wind (direction and speed) and θ along NCAR Electra flight leg at 2500 m MSL. Data were filtered to remove variations with timescale less than 20 s (about 2.2-km spatial scale). Time is in minutes after 2200 UTC.

a small Froude number. Considerable controversy surrounds the relative importance of these two mechanisms (see Smith 1989; Smolarkiewicz and Rotunno 1989b) and even the issue of whether they are distinct. It would seem that these mechanisms would not be able to explain the occurrence of a diurnally forced circulation with maximum strength during the afternoon when the Froude number is systematically largest.

On the larger scale, the presence of flow blocking is undeniable as low-level air over the Great Basin cannot cross the Continental Divide in Colorado. This leads to the anticyclonic gyre on a scale of 300–400 km as southwesterly flow becomes westerly over southern Wyoming and then northwesterly as it enters Colorado. This larger-



FIG. 16. Wyoming King Air descent sounding near Medicine Bow, Wyoming, around 2300 UTC 1 February.

scale flow actually persists for a few days, nearly continuously. The lee vortex, on the other hand is intermittent, present only during the afternoon of 1 February, (and, similarly, during the afternoons of 2–4 February). It is difficult to argue that the formation of the vortex results from a change in the conditions upstream from the Continental Divide. For instance, the low-level stratification and wind in the sounding from Craig, Colorado, hardly change during the morning of 1 February. It seems that whatever the cause of the lee circulation, it is driven by factors on or near the eastern slopes of the Continental Divide.

Perhaps the most useful observations to characterize the flow immediately upstream from the region of vorticity generation are contained in the Wyoming King Air sounding near Medicine Bow, Wyoming (Fig. 16). This sounding differs markedly from that at Craig in that it features weak stratification and weak shear below 3 km MSL. This type of flow, impinging on the downward-sloping terrain east of the Continental Divide, would be characterized by a high Froude number rather than a low Froude number. This is another piece of evidence suggesting that the dynamical paradigm of vortex formation in low–Froude number flow probably does not apply to our anticyclonic circulations.

Mechanisms involving heating or mixing are naturally suited to explain circulations that tend to develop during the afternoon or in the presence of strong lowlevel flow. Diurnal surface heating in initially low-Froude number flow was modeled by Reisner and Smolarkiewicz (1994). Their simulations showed a clear enhancement in the strength of lee vortices by the end of the period of heating. Though detailed diagnosis of the vorticity generation was not performed, it likely resulted from the differential nature of the heating, which was proportional to the terrain height. However, as these authors point out, when the scale of the heating becomes much larger than the horizontal scale of the topography, the response is flow that is uniformly accelerated, not vortical in nature. For islands, it is logical to assume that the scale of the heating and topography are closely related, but not necessarily so in our region of study.

The change in surface potential temperature during the late morning only slightly favors higher elevations (Fig. 17). Much of the horizontal variation of potential temperature changes may be explained by advection. From Fig. 10a, it is evident that cold advection (in terms of θ) is occurring over northeastern portions of the mesonet, while warm advection associated with downslope flow takes place over the southwestern part of the mesonet. As skies were clear, and visibility good at all stations on this day, there is little reason to think that significant horizontal variations of heating occurred.

Uniform heating, even if weak, along with strong surface winds will still lead to boundary layer mixing. Thorpe et al. (1993) and WG discuss different types of boundary layer mixing, each able to produce significant vertical vorticity as air flows past complex terrain. In



FIG. 17. Difference in surface θ between 1900 and 1600 UTC 1 February. Contour interval is 1 K. Plus signs indicate locations of stations used in analysis.

the Thorpe et al. mechanism, as the stress reduces the velocity near the ground, vorticity is created if there is a slope to the surface that is orthogonal to the flow direction. This mechanism depends primarily on surface friction and the stress associated with it above the surface layer. This mechanism can operate in a barotropic flow, in a boundary layer that is stratified or neutral. An important signature of this mechanism would be the presence of vertical shear, not in thermal wind balance, within a layer extending several hundred meters above the ground.

The other mixing-driven mechanism occurs in boundary layers with uniform profiles of velocity and potential temperature. In this situation, vorticity can likewise be produced by the Reynolds stress divergence, but one requires baroclinicity in the direction orthogonal to the boundary layer slope. The assumption of a baroclinic boundary layer that is well mixed in velocity implies that the stress divergence varies with height (Fig. 18). This vertical variation is required to cancel the vertical variation of the pressure gradient force, which would otherwise differentially accelerate the flow into a solenoidal circulation. The pressure gradient force has no curl, by definition, and cannot generate vorticity. However, if there is a slope to the mixed layer (actually to the mean height of the layer above a fixed, reference height), the divergence of the Reynolds stress will have a nonzero curl and a vertical component of vorticity will be generated (Dempsey and Rotunno 1988).

The reason that the stress divergence can generate vorticity results from the fact that the vertical average of the stress divergence depends only on the boundary stresses (at the top and bottom of the mixed layer). Therefore, the part of the stress divergence that cancels

MIXED-LAYER VORTICITY GENERATION



PGF tries to create a solenoidal circulation; stress divergence cancels this to maintain zero vertical shear.



With a uniform north–south gradient of potential temperature, PGF does not vary with x, but as z₄ varies with x, so does the stress divergence. Thus, a vertical component of vorticity is generated.

FIG. 18. Schematic of mixed-layer generation of vorticity. Top figure indicates sense of pressure gradient and stress forces in a baroclinic mixed layer, with colder air to the north. PGF is the pressure gradient force, and p' is the pressure at y_1 relative to y_2 . Middle figures are adapted from Dempsey and Rotunno (1988) and show a cross section, parallel to the gradient of boundary layer midheight (z_a), of the stress terms, assuming colder air to the north. Circles with centered dots indicate a stress force out of the page (directed to the south); circles with crosses indicate a northward-directed stress. Gray circles with centered dots represent a southward-directed pressure gradient force. Bottom figures indicate the gradients of z_a and potential temperature corresponding to the middle plots.

the pressure gradient force must average to zero over the mixed-layer depth at every horizontal location. As Fig. 18 depicts, even if the stress divergence has the same vertical profile everywhere, relative to the ground, a sloping boundary layer will lead to a horizontal variation in the stress divergence, and a nonzero curl.

Following Dempsey and Rotunno, the mixed-layer form of the equation for the vertical component of vorticity can be written, excluding stretching terms and assuming that the boundary stresses have no curl,

$$\frac{d\zeta}{dt} = -\frac{g}{\theta_0} \mathbf{k} \cdot (\nabla z_a \times \nabla \theta). \tag{1}$$

Here, θ_0 is a constant, reference potential temperature;

 $\theta(x, y)$ is potential temperature in the boundary layer; $z_a = (\frac{1}{2})(h + z_s)$, where *h* is the height of the boundary layer top (above mean sea level); and z_s is the surface elevation. We refer to the quantity z_a as the mean boundary layer height. The generation of vorticity is maximized if the horizontal gradient of boundary layer potential temperature is orthogonal to the gradient of mean boundary layer height (Fig. 18). If, following the gradient of z_a , the potential temperature gradient points toward the left, anticyclonic vorticity is produced. Cyclonic vorticity is produced if the potential temperature gradient points to the right.

One possible way to distinguish which mixing mechanism best applies to the vorticity production in this case is to examine upstream soundings. If the low levels are strongly sheared and/or strongly stratified, mixedlayer dynamics are not relevant. An appropriate sounding was taken during the King Air descent near Medicine Bow. This sounding shows weak stratification and shear over the lowest kilometer, favoring a mixed-layer interpretation. Without knowing what departures from perfectly mixed profiles are allowed within the context of mixed-layer dynamics, we cannot say more from a single sounding. However, it is suggestive that the layer of weak shear and stratification in the Medicine Bow sounding corresponds well to the depth of the anticyclonic disturbance downstream.

An alternate way to assess the applicability of mixing mechanisms is to examine analyses of surface θ , wind, and terrain elevation (see Fig. 9a). At 1800 UTC, there is a mesoscale region characterized by a significant gradient of surface θ orthogonal to the gradient of topographic height. Though we do not know the height of the boundary layer in detail, it is likely that the gradient of $h + z_s$ will look similar to the gradient of z_s , given that the relevant variation of z_s is at least 1 km. This region therefore meets the criterion for mixed-layer generation of vorticity, and $-\mathbf{k} \cdot \nabla z_a \times \nabla \theta$ points downward, consistent with anticyclonic vorticity generation. The mechanism of Thorpe et al. (1993) is likely not operating here simply because the flow is primarily across, not along, terrain contours.

To be more quantitative, the rate of vorticity generation can be estimated from $\zeta \sim -(g/\Theta_0)(\Delta z_s/L_x)(\Delta \theta/L_y)$, where L_x and L_y are length scales along the gradients of θ and z_s , respectively. Here, Δz_s and $\Delta \theta$ refer to variations in topographic height and potential temperature over length scales of L_x and L_y , respectively. If $L_x = L_y$ = 100 km and $\Delta z_s = -1$ km and $\Delta \theta = -8$ K (estimated from Fig. 9a), the generation rate is -2.7×10^{-8} s⁻², almost -f h⁻¹. This is rapid enough to account for the observed development of anticyclonic vorticity of about -2 to -3 times the Coriolis parameter in just a few hours.

Two items possibly contribute to the timing of vorticity production. The most obvious is the onset of mixing, presumably driven by daytime heating. The other is the increase in the gradient of θ along the terrain contours. In the lee of the highest terrain, downslope flow increases θ locally, but farther north and east, such warming does not occur. Thus the northeast-southwest gradient of θ is increased. The synoptic-scale gradient of θ is also in the same direction, further conditioning the flow for vorticity generation. As the vortex intensified in this case, it advected θ in the boundary layer to the north and west, helping to reduce the gradient of θ established earlier. This, coupled with sunset and the reduction of mixing, probably led to the demise of the circulation.

b. A type 2 case: 29 January 1994

As type 2 cases feature more northerly large-scale surface flow, as opposed to northwesterly type 1 cases,



FIG. 19. Potential temperature (black, 2-K interval) and winds at 500 mb and sea level pressure (gray, 4-mb interval) for 0000 UTC 30 January.

there is less downslope flow on the synoptic scale, and therefore greater moisture and precipitation are possible. Such was the case during the event of 29 January. Synoptic analyses for 0000 UTC 30 January (Fig. 19) indicate that the core of a cold trough aloft was located over western South Dakota and northeastern Wyoming, with the coldest 500-mb temperatures about -38° C. Over the WISP region, a jet of 50 kt is evident. The sea level pressure analysis reveals an inverted trough to the east of the Rocky Mountains, with arctic air at the surface farther east. On the west side of this trough, the prevailing surface flow was from the north-northwest. An important difference between this surface pattern and that of 1 February is the absence in this case of surface cold air over the intermountain region and a strong pressure gradient over western Colorado and Wyoming. Thus, in this case, boundary layer air probably does not originate from the cold, dry Great Basin, but rather from east of the Continental Divide.

Figure 20 summarizes the behavior of the primary snowband that occurred on this day over northeastern Colorado as seen by the Front Range WSR-88D radar (located by an " \times " in Fig. 20a). Some light snow was present during the late morning, but around noon local time (1900 UTC) enhanced snowfall became evident along the northern periphery of the radar scan. This area grew slightly but generally remained stationary until about 2000 UTC when it rapidly surged southward. A well-defined banded structure developed with a cross band width of perhaps 40 km and length of over 100 km. Within the band, heavy snow was reported.

As the snowband approached the higher terrain to the south, its orientation became more north–south instead of northwest–southeast, and it slowed, gradually evolving into an area of light to moderate snow adjacent to the foothills. This area of snow persisted for several



FIG. 20. WSR-88D CAPPIs at 3 km MSL from Front Range radar on 29 January 1994 at hourly intervals from 2000 to 2300 UTC 29 January. Reflectivity scale (dBZ) is shown at bottom of plot. Heavy black lines are terrain-height contours (see Fig. 1).

hours, resulting in snow accumulations of several inches in many locations to the west and south of Denver.

Mesoscale surface analyses (Fig. 21) indicate the movement of a wind-shift line southward over southeastern Wyoming, southwestern Nebraska, and northeast Colorado around local noon on 29 January. Potential temperatures do not indicate a strong front, but there is likely considerable masking from the nocturnal cooling over the Platte River valley of northeast Colorado. By 1900 UTC, we note the strengthening of northerlies over the Cheyenne Ridge. Winds in the lee of the ridge begin turning northeasterly and, by 2100 UTC, strong northeasterlies have pushed all the way to the Denver metropolitan area.





2100 UTC 29 JANUARY

FIG. 21. Surface analyses of potential temperature (solid, 2-K interval wind station winds (kt) at (a) 1900 UTC and (b) 2100 UTC 29 January.



FIG. 22. Mesonet time series of potential temperature (solid) and perturbation pressure (dashed) at Fort Morgan, Colorado (top), and Boulder, Colorado (bottom), during the afternoon of 29 January.

A time series of mesonet observations from Fort Morgan, Colorado, about 100 km east of the Front Range, details the passage of some type of surface trough around 1930 UTC (Fig. 22). As the wind shifted into the north, the potential temperature leveled off and then fell as snowfall commenced. At this time, the perturbation pressure (and the total pressure) showed a rapid pressure fall. This pressure fall was also evinced at Boulder, farther west, without a notable shift in wind direction. Later, Boulder experienced what seemed to be a frontal passage, accompanied shortly thereafter by heavy snow around 2130 UTC. A gradual pressure rise occurred at both Fort Morgan and Boulder following 2000 UTC, seemingly independent of the passage of a trough or frontlike feature. Note that these pressures exclude the mean diurnal cycle.

A time series of Cross-chain Loran Atmospheric Sounding System (CLASS) profiles from Chugwater, Wyoming (Fig. 23), indicates the passage of a weak frontal feature, perhaps, but the more obvious theme is the presence of strong north-northwesterly flow at low levels, weak stratification in the afternoon, and baroclinicity with colder air to the north (and east). The snowband may have had a fairly long history, given that Casper, Wyoming, reported a period of moderate to heavy snow around 1500 UTC. Later, around 1800 UTC, Cheyenne reported a similar occurrence. The band must have passed the Chugwater location just prior to the 1800 UTC sounding, but it is difficult to point to a strong frontlike feature.



FIG. 23. Time series of CLASS soundings from Chugwater, Wyoming, on 29 January. Contoured is potential temperature (1-K intervals).

If we accept that a weak frontlike feature may have produced the snowband north of the Cheyenne Ridge, there is still the question of what caused its strengthening and changing of orientation in the lee. Closely related are questions about the low-level anticyclonic circulation in this case, namely, what causes it and how does it modify the snowfall?

Northwesterly flow prevails at all stations over, north, and east of the Cheyenne Ridge, but strong northeast flow is evident at Denver over about the lowest several hundred meters (Fig. 24). The shift from northwesterly to northeasterly flow occurs quickly, in a span of 2 h, as captured by the 404-MHz profiler at Platteville, Colorado (Fig. 25). Absolute vorticity within the anticyclonic region can be estimated from the numerous CLASS soundings that were launched on this day. Using the triangle method (Bellamy 1949), we estimate absolute vorticity to be about zero within the triangle bounded by Chugwater, Wyoming; Kimball, Nebraska; and Denver, Colorado, at 2200 m MSL, the altitude where the anticyclonic circulation was strongest.

Concerning mechanisms of anticyclonic vorticity formation, we look to an upstream sounding taken at Glendo, Wyoming, on the afternoon of 29 January (Fig. 24). This CLASS profile shows a boundary layer that is relatively well mixed in both potential temperature and momentum over the lowest several hundred meters. The low-level winds, from the north-northwest at this location, are mainly parallel to the local terrain contours;



FIG. 24. Profiles of θ and win 0000 UTC 30 January from the Mobile Class (MC1 on Fig. 1) at Glendo, Wyoming (solid), and the NWS sounding at Denver, Colorado (dashed).

therefore the principal topographic feature for this flow to encounter is the Cheyenne Ridge downstream. With respect to this ridge, the Froude number is large (3–5). Thus, the incident flow is in a regime where we expect neither the tilting mechanism of SR nor gravity wave breaking to generate a vertical component of vorticity.

With a relative absence of low-level shear in the upstream sounding, it is difficult to support the frictional



FIG. 25. Time series of winds from Platteville, Colorado, from 1700 UTC 29 January to 0200 UTC 29 January. Standard plotting convention for winds.

generation mechanism of Thorpe et al. (1993). The most likely mechanism would appear to be similar to the one we hypothesized for type 1 cases, namely, mixed-layer generation of vorticity. Is this consistent? Circumstantial evidence exists in the fact that the circulation forms during the afternoon and dissipates during the evening, when mixing presumably weakens. The well-mixed profile at Glendo, Wyoming, is additional evidence. For the mixed-layer mechanism to operate, there is also the requirement of baroclinicity. The only appreciable gradient of potential temperature at low levels is the southwestward-directed gradient within the region of cold advection evident, for instance, in the Chugwater soundings (Fig. 23). To produce anticyclonic vorticity, the midheight of the mixed layer must slope upward to the west or northwest, thus ensuring that $-\mathbf{k} \cdot \nabla z_a \times \nabla \theta$ is negative. We note that there exists a difference in mixedlayer midheight between the RAP mobile sounding (RP2, near Cheyenne, Wyoming) at 2100 UTC and Kimball, Nebraska, of about 250 m (not shown). Given the observed gradient of θ , about 3 K over 150 km, the mixed-layer mechanism would produce a local vorticity tendency of about 0.2f in 3 h. This is probably insufficient to account for the observed vorticity. However, we note that the terrain slope and, presumably, mixedlayer mean depth increase more abruptly to the west of Cheyenne. If the gradient of z_a attains a value of about 1 km per 100 km, true of the terrain slope, the vorticity generation over 3 h would be about -f. It is, in fact, from Chevenne west, in the lee of the Chevenne Ridge, that the most pronounced anticyclonic turning of the flow occurs.

Concerning the effect of the anticyclone on snowfall, Wesley et al. (1995) documented a case where snowfall was enhanced near the foothills in association with boundary layer convergence of wind produced by the Longmont anticyclone. The clockwise rotation of the snowband in our case evinces the effect of the anticyclonic circulation. Boundary layer convergence within this circulation is apparent in the surface analyses in Fig. 26. Additional low-level convergence is expected as the northeasterly flow encounters the foothills. Although the stratification over the lowest few hundred meters is small, stable conditions at higher levels are sufficient to help block this flow. The pressure field is consistent with an upstream blocking effect, as we noted the existence of a "stagnation high" in both the composite and this case (Fig. 26).

4. Conclusions

Based on data collected during WISP94, we have identified two types of quasi-stationary, terrain-induced mesoscale anticyclonic circulations over northeast Colorado. Traditionally, these have broadly been termed "Longmont anticyclones," but the two types of circulation discussed herein differ systematically in structure, intensity, and associated weather. Despite these differences, they appear to be linked by similar dynamics.

Type 1 events feature a somewhat stronger circulation and are strongly diurnally forced. They occur under synoptic-scale northwesterly flow and are aided when potentially colder air exists to the northeast of the region. High pressure over the Great Basin is often seen in these events, probably because the favorable large-scale pattern exhibits ridging to the west and troughing to the east in the upper troposphere. This synoptic situation lends itself to strong westerlies just north of the Medicine Bow mountains in southern Wyoming, which subsequently turn clockwise to become northwesterlies over southeastern Wyoming. Northwesterly, downslope flow results over extreme north-central Colorado and southeastern Wyoming. In the 1 February case, it was observed that the lowest kilometer over the wind corridor was characterized by strong winds, weak stability, and weak shear.

Type 2 events, those forming in more northerly flow, feature circulations that are, on average, weaker than those of type 1 events, but because of the additional moisture, type 2 events can significantly affect snowfall. The general characteristics of type 2 events are a north–south temperature gradient and a well-mixed boundary layer. These conditions are often present after frontal passages in the region. Indeed, at least four out of the nine type 2 events in the WISP94 composite formed following a frontal passage.

The dynamical mechanism believed responsible for generating vertical vorticity is boundary layer mixing similar to that which takes place in a mixed-layer model when the gradient of potential temperature is orthogonal (to the left) of the gradient of the midheight of the mixed layer. In both type 1 and type 2 cases, the gradient of θ is usually directed toward the southwest, the gradient of mixed-layer midheight (terrain height) to the northwest. One could imagine that type 2 events would occur quite often following the passage of strong fronts as cold advection occurs on northerly or northwesterly flow adjacent to the foothills. In this case, we always expect some gradient of θ directed to the south, orthogonal to the terrain height gradient. It is even possible that many of the so-called shallow upslope snowstorms along the Front Range are caused by this mechanism.

The sign of the implied vorticity generation agrees with the observations, and the magnitude is qualitatively correct, roughly -f in 1–2 h, perhaps locally larger. The mixed-layer mechanism also can explain the finding that the circulation is often diurnally forced, occurring during the afternoon when well-mixed conditions are most likely. Because there are systematic convergence zones expected within the anticyclonic circulation, namely, where it interacts with terrain, snowfall events around Denver should and do exhibit an afternoon and evening maximum frequency.

Some important similarities between the anticyclonic circulations studied here and the Denver cyclone may ex-



FIG. 26. Analyses of perturbation pressure at 2-h intervals from 1900 UTC 29 January to 0100 UTC 30 January. Contour interval is 0.4 mb.

ist. There is also much that is unclear about the relationship between the two phenomena. For instance, Crook et al. (1990) suggest that the Denver cyclone often forms in low–Froude number flow on many occasions, yet WG were able to simulate a morning occurrence of a particular case with a mixed-layer model. We have found no evidence that the low–Froude number (Fr) paradigm applies to the Longmont anticyclone.

In the simulation of WG, low-level convergence was found to maximize immediately in the lee of the Palmer Lake Divide and extend downstream with time. The low-level convergence pattern of the Denver cyclone bears qualitative resemblance to convergence seen in surface analyses for anticyclonic circulations, especially in type 2 cases. For cases studied here, strong convergence is favored immediately in the lee of the Cheyenne Ridge as the anticyclonic circulation develops, and then this convergence zone may move downstream. It is possible that such convergence in the 29 January case may have accounted for the northwest–southeast orientation of the snowband south of the Cheyenne Ridge.

Our results suggest that, despite the apparently restrictive nature of the mixed-layer assumption, models based on that assumption may have considerable relevance to the atmosphere, even in cases where small, but finite, stratification and shear are present. The action of daytime heating, combined with the strong low-level winds observed upstream of the alleged region of vorticity production, appears sufficient to mix the boundary layer even in winter. There are, however, cases that occur at night. For these it is unlikely that mechanical mixing alone could account for a deep mixed layer.

Another possible contributor to mixing is related to baroclinicity, which we have already suggested is necessary for vorticity production. C. Bishop (1995, personal communication) has suggested that symmetric instability subsequently followed by buoyant convection can occur in a baroclinic boundary layer when the surface pressure and potential temperature gradients are parallel. The frictionally induced ageostrophic surface wind produces warm advection near the surface, leading to continued destabilization. Most of the anticyclone cases in our sample formed in baroclinic environments in which large-scale (over several hundred kilometers) gradients of potential temperature and pressure were parallel (toward the southwest). The process of destabilizing a baroclinic boundary layer may be particularly important for the observed nocturnal cases of anticyclonic vorticity generation.

This paper has not addressed a practical forecast question related to the Longmont anticyclone. Given that northwesterly flow and baroclinicity are common features over northeast Colorado in the cool season, why does a Longmont anticyclone not form nearly all the time? As shown by Dempsey and Rotunno, formation of lee vortices in mixed layers depends critically on Fr_m. It is possible that sensitivity to Fr_m may explain the frequency of anticyclone events, roughly one-third of the days in our sample, but it is unclear how to apply quantitatively the idealized modeling results to the physiography and complex flow structure of the central Rockies. Another difficulty in applying theoretical results, which pervades most problems in flow over complex terrain, is a lack of atmospheric profiles in key regions near and just upstream of where vorticity is produced. The present study was helped a great deal by even the relatively small amount of upstream data taken during WISP94, which suggests that a modest, but carefully placed, enhancement of the observations could help greatly in forecasting these type of events.

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