Radio Acoustic Sounding System (RASS) and Wind Profiler Observations of Lower- and Midtropospheric Weather Systems

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ABSTRACT

The National Oceanic and Atmospheric Administration (NOAA) Wave Propagation Laboratory (WPL) wind profilers and accompanying radio acoustic sounding system (RASS) temperature profilers in eastern Colorado jointly measure nearly continuous (<1 h), high vertical resolution (<300 m) wind-velocity and virtual-temperature profiles. This study presents NOAA/WPL wind profiler and RASS observations and diagnostics of propagating lower- and midtropospheric weather systems over Colorado. The wind and temperature remote-sensing systems observed wind-velocity and virtual-temperature structures associated with a synoptic-scale trough and embedded fronts, and a propagating short-wave trough and trailing midtropospheric jet-stream–frontal-zone (jet–front) system. Single-station hourly diagnostic calculations of geopotential heights, horizontal virtual potential temperature gradients, thermal advections, vertical velocities, gradient Richardson numbers, and cross-frontal isentropic potential vorticity demonstrate that dynamically consistent synoptic-scale and mesoscale signals can be obtained by combining wind profiler and RASS observations. The wind profilers and RASS documented mesoscale wind velocity and thermal features up to 400 mb that were unresolved temporally and spatially by the synoptic-scale rawinsonde network. Results demonstrate the potential for multisystem (network) applications of this revolutionary technology for describing the temporal and spatial evolution of synoptic-scale and mesoscale weather systems.

1. Introduction

Since the 1930s, meteorologists have relied upon synoptically spaced (~400 km), 12-h rawinsonde soundings to resolve the synoptic-scale structure of the atmosphere’s wind and thermal fields. During the past two decades, selected research-oriented field studies [e.g., Severe Environmental Storms and Mesoscale Experiment (SESAME) (Hill et al. 1979); Experiment on Rapidly Intensifying Cyclones over the Atlantic (ERICa) (Hadlock and Kreitzberg 1988)] have utilized regional rawinsonde networks with enhanced spatial (~200 km) and temporal (3–6 h) resolution to observe mesoscale weather systems. These experimental networks, however, did not provide extended periods of enhanced upper-air data. Recent advances in remote upper-air wind-observing capabilities with radar wind profilers have resulted in unsurpassed detailed measurements of synoptic-scale and mesoscale wind fields and offer promise for observing these wind fields on a regular basis in an operational setting. Radar wind profilers measure temporally continuous (<1 h), high vertical resolution (<300 m) wind-velocity profiles through the troposphere and into the lower stratosphere (up to ~18 km). Single-station and triangle-network research applications have demonstrated the versatility and accuracy of these active remote observing systems in describing synoptic-scale and mesoscale weather systems (e.g., Gage and Clark 1978; Larsen and Röttger 1982; Shapiro et al. 1984; Nastrom et al. 1985; Sweezy and Westwater 1986; Zamora et al. 1987; Neiman and Shapiro 1989). The installation of the National Oceanic and Atmospheric Administration (NOAA) 30-station wind profiler demonstration network throughout the central United States in 1991 (Chadwick and Hassel 1987; Weber et al. 1990) marks yet another advancement in wind-observing technology and capability. This new profiler network is considerably larger in areal coverage than previous local profiler networks and provides both improved horizontal and

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temporal resolution over the current rawinsonde network.

Despite these ongoing advancements, there remains the task of obtaining remote temperature profiles with comparable vertical sampling resolution and vertical range. Recent technological advances in the active remote sensing of temperature have been combined with radar wind profiler technology to produce a radio acoustic sounding system (RASS). The RASS technique (first explored by Fetter 1961; Smith 1961; and Marshall et al. 1972) is used to obtain vertical profiles of the speed of sound, from which virtual temperature profiles are calculated. An important feature of the RASS is its ability to measure virtual-temperature profiles with the same vertical and temporal sampling resolution (<300 m and <1 h, respectively) as its collocated wind profiler (Currier et al. 1988; May et al. 1988). The vertical range of early RASS systems was limited to <1000 m (Bonino et al. 1986), but present-day wind profilers have improved the height coverage of the RASS to 2–6 km under all wind conditions (May et al. 1988). Although the RASS coupled with a wind profiler cannot measure virtual temperatures to heights attained by the rawinsonde (>15 km), it can resolve the vertical structure and rapid local changes of lower- and midtropospheric (<6 km above ground level (AGL)) temperature features such as those associated with stable layers and frontal zones. The coupled RASS and wind profiler observing system represents one component in a stand-alone ground- and space-based active remote observing system under development.

During the last several years, the NOAA Wave Propagation Laboratory (NOAA/WPL) wind profilers and accompanying RASS temperature profilers in eastern Colorado have taken detailed wind and virtual-temperature measurements of lower- and midtropospheric weather systems. In a recent study, Neiman et al. (1991) presented single-station NOAA/WPL wind profiler and RASS observations of an arctic front over Colorado in the lowest 2 km AGL. The current study presents single-station NOAA/WPL wind profiler and RASS observations of migrating weather systems and associated fronts extending to 5–6 km AGL. These observations are used to deduce hourly diagnostic estimates of geopotential heights, horizontal virtual potential temperature gradients, thermal advections, vertical velocities, gradient Richardson numbers Ri, and cross-frontal isentropic potential vorticity IPV. Analyses of the high temporal resolution wind profiler and RASS observations and diagnostics reproduce many aspects of, and in some ways improve upon, the synoptic-scale and mesoscale analyses obtained from the rawinsonde network. In addition, the wind profiler and RASS analyses contain jet-stream–frontal-zone (jetfront) structures that are quite similar to those of “classic” mid- and upper-tropospheric jet–front systems and associated tropopause folds documented by numerous investigators including Newton (1954), Reed (1955), Reed and Danielsen (1959), Shapiro (1981), and Keyser and Shapiro (1986). These analyses of wind profiler and RASS observations and diagnostics demonstrate the potential for multisystem (network) applications of this revolutionary technology for describing the spatial and temporal evolution of synoptic-scale and mesoscale weather systems.

2. Observing systems

Since mid-1989, the NOAA/WPL has operated and maintained a coupled 50-MHz wind profiler and RASS temperature-observing system at Platteville, Colorado [1524 m above mean sea level (MSL)], and a 915-MHz wind profilers–RASS system at Stapleton International Airport (1611 m MSL) in Denver, Colorado (Fig. 1). These systems are separated by 50 km.

The 50-MHz wind profiler measures one composite wind-velocity profile per hour (an average of 12 individual profiles per hour) with high vertical resolution and one with low vertical resolution. The high vertical resolution profiles contain wind-velocity information from 1.5 to 8.2 km AGL, inclusive, at 290-m intervals.
whereas the low vertical resolution profiles contain information from 4.2 to 17.3 km AGL, inclusive, at 870-m intervals. The accompanying RASS provides hourly virtual-temperature profiles with vertical resolution and minimum range equal to those of the 50-MHz high-resolution wind-velocity profiles. The maximum range of this RASS can exceed 7 km AGL in low wind (<10 m s\(^{-1}\)) conditions but is typically 5–6 km AGL.

The 915-MHz wind profiler provides 20-min composite wind-velocity profiles (also an average of 12 wind profiles) with 150-m vertical resolution from 0.25 to 5.5 km AGL. The companion RASS provides virtual-temperature profiles of equal resolution, but its maximum range does not exceed 2 km AGL.

The primary factor limiting the vertical range of RASS virtual-temperature measurements is acoustic absorption by the atmosphere. Detailed discussions of coupled RASS–wind profiler systems and their applications and limitations were described by Peters et al. (1983), May et al. (1989, 1990), and Neiman et al. (1991). A comprehensive description of wind profiler performance characteristics was provided by Strauch et al. (1984).

Additional measurements of temperature, moisture, and wind velocity used in this study were taken by the NOAA/WPL six-channel radiometric profiler at Stapleton and the North American operational rawinsonde and surface observation networks. The radiometric profiler remotely sensed column-integrated water vapor and liquid water every 2 min from the brightness temperatures obtained in the 20.6- and 31.6-GHz wavelength bands using statistical retrieval methods (Guiraud et al. 1979; Westwater and Guiraud 1980; Hogg et al. 1983). The rawinsonde network (~400 km between sites) provided 12-h profiles of wind velocity, temperature, and moisture. The surface observing network provided hourly surface measurements of wind velocity, temperature, and moisture.

3. Data processing

Each hourly high and low vertical resolution wind-velocity profile from Platteville was combined with the corresponding wind-velocity profile from Stapleton into one “single-station” profile with 300-m vertical resolution by vertical linear interpolation. The RASS virtual-temperature profiles from Platteville and Stapleton were similarly combined. These pseudo–single-station (combined) wind-velocity and virtual-temperature profiles contained Stapleton data up to the maximum range of the Stapleton RASS (~2 km AGL) and contained the Platteville data above. The combined profiles exhibited a minimum in mismatch error for the observational periods hereinafter presented. The time series of wind velocity and virtual temperature at each range gate were filtered with one pass of a three-point Shuman filter (Shuman 1957) to eliminate temporal features with periods less than 2 h.

The wind-velocity and virtual-temperature profiles were measured as a function of height (km) but converted to pressure coordinates so that 1) the RASS virtual potential temperature \( \theta_v \) profiles could be calculated and 2) the single-station diagnostic techniques discussed in section 5 could be implemented. The conversion from height to pressure coordinates was implemented by hydrostatic integration using the surface pressure at Stapleton as a lower boundary and RASS mean-layer virtual temperatures.

The single-station wind profiler and RASS diagnostics of horizontal virtual potential temperature gradients, thermal advections, vertical velocities, gradient Richardson numbers, and cross-frontal isentropic potential vorticity discussed in section 5 employed first-order vertical and temporal finite-centered differencing.

In order to compare vertical-velocity profiles of the single-station diagnostic technique with independently derived vertical-velocity profiles, vertical profiles of horizontal divergence \( \nabla \cdot V \) and kinematic vertical motion \( \omega = \omega_0 - \int_0^\infty (\nabla \cdot V) d\phi \) were calculated at 25-mb intervals from a triangular array of rawinsonde wind-velocity profiles (Denver, Colorado; North Platte, Nebraska; and Dodge City, Kansas). Vertical velocity at the surface boundary was assumed to be nonexistent. This technique was first discussed by Bellamy (1949). The vertical profiles of divergence were adjusted to satisfy mass-balance requirements (O'Brien 1970).

4. Observations of a synoptic-scale trough passage

a. Synoptic overview

Between 9 and 12 December 1989, a strong synoptic-scale (~2500 km) trough containing frontal and jet-stream structure propagated southeastward across Colorado. The 500-mb geopotential height and virtual-temperature analysis for 1200 UTC 11 December (Fig. 2) shows the geopotential trough and cold (~ -35°C) polar air (thermal trough) over Colorado and an associated polar frontal zone circumscribing Colorado to the south and west. The segment of the front to the south and southeast of Colorado is referred to as the leading branch of the polar front, and the segment to the west of the state is referred to as the trailing branch. A 500-mb ridge in geopotential height and temperature was situated along the Pacific coast of the United States upstream from the trough and trailing branch of the polar front.

Figure 3 presents a cross-section analysis of virtual potential temperature \( \theta_v \) and the front-parallel wind-velocity component \( u \) along line AA' in Fig. 2. The analysis shows the extensive depth (>6 km) of the polar air over Colorado. The associated leading (southeastern) and trailing (western) branches of the polar front (shown in Fig. 2) extended downward from the tropopause to near the surface and contained moderate vertical wind shear (~20–30 m s\(^{-1}\) km\(^{-1}\)) beneath the 60 to 70 m s\(^{-1}\) jet-stream cores. The cross-section
b. Wind profiler and RASS observations

A time–height analysis of virtual potential temperature was prepared from hourly RASS measurements taken between 9 and 12 December, and was accompanied by hourly wind profiler vectors (Fig. 4). The analysis clearly shows the leading and trailing branches of the polar frontal zone surrounding the deep (~6 km) polar air mass, and also shows the shallow (below 650 mb) arctic air mass beneath. A zone of enhanced static stability $\partial \theta_v / \partial p$ and local cooling near the surface at 0600 UTC 10 December represented the passage of the leading polar frontal branch, which temporally ascended to 400 mb by 1600 UTC. The frontal passage below 550 mb was characterized by a wind-direction shift from westerly to northerly. Assuming the applicability of Taylor’s hypothesis (i.e., that nonevolving weather systems propagate at a fixed velocity) during the observing periods discussed in this study, the wind-direction shift marked a concentrated region of cyclonic vorticity (cyclonic wind-direction shift). The frontal passage above 550 mb was characterized by a decrease in the westerly wind speed followed by little postfrontal change in $\theta_v$ within the polar air mass. Below 600 mb, postfrontal cooling occurred until 1600 UTC 11 December in association with a southward influx of arctic air. An easterly (upslope) wind-velocity component was confined to the shallow arctic air mass. The arctic air migrated to the east of Colorado between 1600 and 2300 UTC 11 December, during which enhanced warming (3 K h$^{-1}$) occurred with the onset of adiabatically warmed downslope westerly winds below 700 mb.
mb. The eastward departure of arctic air in the lower troposphere occurred together with the eastward departure of polar air in the middle troposphere. Between 1500 UTC 11 December and 0400 UTC 12 December, the passage of the trailing polar frontal branch was characterized by a temporally descending zone of enhanced static stability and warming, during which the wind direction shifted cyclonically from northwesterly to northerly.

Figure 5 presents a temporally compressed version of the RASS $\Phi$ analysis in Fig. 4, together with an analysis of the wind profiler front-parallel wind-velocity component. The analysis shows the midtropospheric extension of the jet-stream cores on the warm side of each polar frontal branch, and moderate vertical wind shear within the transition zone of each frontal branch (also see Fig. 3). Weak wind speeds ($<10$ m s$^{-1}$) were confined to the polar and arctic air masses. It is instructive to note the striking similarity between the spatial structure of the cross-sectional analysis in Fig. 3 and the temporal structure of the time-height analyses in Figs. 4 and 5.

Time series of geopotential heights $\Phi$ at designated pressure levels were calculated from hourly RASS virtual-temperature profiles by hypsometric integration of RASS mean-layer virtual temperatures. Figure 6 presents time series of RASS $\Phi$ and profiler wind velocities at 700, 500, and 400 mb, and corresponding time series of the 12-h Stapleton rawinsonde $\Phi$ and wind velocities. The RASS time series were in close agreement with the corresponding rawinsonde time series with regard to documenting the propagating synoptic-scale ($\approx 2500$ km) weather systems. The RASS and rawinsonde time series at 500 and 400 mb each show decreasing heights associated with the approach of the synoptic-scale geopotential trough between 1200 UTC 9 and 11 December, followed by increasing heights associated with the approach of the upstream geopotential ridge. Likewise, comparative RASS and rawinsonde time series at 700 mb each show a minimum in $\Phi$ associated with the passage of the leading polar frontal branch at $\approx 1200$ UTC 10 December, followed by a 24–36-h period of increasing geopotential...
heights associated with the southward intrusion of a shallow arctic anticyclone. The comparative RASS and rawinsonde time series also show a decrease in the thickness between the isobaric surfaces associated with the approach of the cold core of polar air, followed by an increase in the thickness after the passage of the cold polar core to the east.

The RASS and wind profilers also provided unique observations of lower- and midtropospheric geopotential height and wind-velocity structures (Fig. 6) associated with propagating mesoscale (≤1000 km) weather systems embedded within the synoptic-scale (∼2500 km) cyclonic circulation. Focusing on mesoscale structures at 400 mb, the RASS and wind profilers observed two mesoscale troughs and three mesoscale ridges (unresolved temporally by Stapleton’s rawinsonde time series and spatially by the rawinsonde network) propagating eastward across eastern Colorado ahead of the synoptic-scale trough axis passage (trough 3) at 1200 UTC 11 December. The passage of each short-wave ridge and trough was accompanied by an anticyclonic and a cyclonic wind-direction shift, respectively (assuming the applicability of Taylor’s hypothesis). After 1200 UTC 11 December, the amplitude and inferred half-wavelength between the synoptic-scale trough axis (trough 3) and the upstream ridge axis (ridge 4) were larger and smaller, respectively, than what was resolved temporally by the rawinsondes. A cyclonic wind-direction shift occurred during the synoptic-scale trough-axis passage, whereas an anticyclonic wind-direction shift accompanied the upstream ridge-axis passage.

Correlating the remote observations of the embedded mesoscale short waves with local weather conditions is not attempted here, because the remote observations were from a single station and their proximity to high terrain and associated topographically forced circulations make it difficult to determine if the changing weather conditions were caused by migratory, evolving, or topographically forced atmospheric features. Considerable insight can be gained regarding the propagation and life cycle of mesoscale short waves embedded in synoptic-scale flow, and the influence of these short waves on the local weather, with a network of RASS and wind profilers located far from complex terrain.
5. Observations and diagnostics of a midtropospheric frontal zone

a. Synoptic overview

On 7 and 8 December 1989, an amplifying 500-mb short-wave (≈2000 km) trough in geopotential height propagated east-southeastward across Colorado and was accompanied by a thermal trough of maritime polar origin (Figs. 7a and 7b, respectively). A thermal ridge of subtropical origin was situated upstream from the short-wave trough. An eastward-propagating, north–south-oriented midtropospheric frontal zone (Figs. 7a and 7b) marked the transition between these two air masses; an accompanying 300-mb 70 m s⁻¹ northerly jet was situated above (Fig. 8). The vertical structure of the jet–front system is illustrated in a cross-section analysis of virtual potential temperature θ_v and the front-parallel wind-velocity component u (Fig. 9a) at 1200 UTC 7 December (along line BB' in Fig. 7a). The analysis shows the front extending downward from the tropopause over North Platte, Nebraska (LBF), to near the surface at Desert Rock, Nevada (DRA). The frontal layer contained large vertical wind shear (≈30 m s⁻¹ km⁻¹) and cross-frontal vorticity (2.7 × 10⁻⁴ s⁻¹) beneath the 70 m s⁻¹ jet-stream core. An axis of cold virtual potential temperatures associated with the thermal trough was situated at the eastern boundary of the frontal zone. A companion cross-frontal isentropic potential vorticity (IPV) analysis (Fig. 9b) shows an intrusion of stratospheric IPV (>1 × 10⁻⁵ K s⁻¹ mb⁻¹) into the frontal zone extending downward to ≈700 mb during a tropopause-folding episode. In the following 12 h (not shown), the front propagated eastward and descended more than 200 mb over Denver.

b. Wind profiler and RASS observations

The eastward-propagating short-wave trough and trailing jet–front system were clearly observed by the wind profilers and RASS during the period 0000 UTC 7 to 1200 UTC 8 December. The wind profiler time–height analysis (Fig. 10) shows a cyclonic wind-direction shift associated with the passage of the geopotential short-wave trough axis, temporally ascending from near the surface at 0300 UTC 7 December to 400 mb at 1100 UTC. The wind-direction shift was most evident near 400 mb. Following the wind-direction shift, a weak northerly jet (≈15 m s⁻¹) developed at ≈700 mb after 0800 UTC 7 December in response to a strengthening cyclonic circulation associated with lower-tropospheric cyclogenesis southeast of Colorado (not shown). The much stronger upper-tropospheric (≈300 mb) northerly jet (56 m s⁻¹) and its associated frontal wind shear (30 m s⁻¹ km⁻¹) beneath were observed after 1200 UTC 7 December. By 0800 UTC 8 December, the frontal shear layer temporally descended to ≈700 mb. Above the 300-mb jet-stream core, a temporally ascending zone of decreasing winds with height coincided with the temporally ascending tropopause in advance of the subtropical ridge.

![Map of 500-mb virtual temperature (°C, dark contours) and geopotential height (dam, shaded contours)](image-url)

**FIG. 7.** 500-mb virtual temperature (°C, dark contours) and geopotential height (dam, shaded contours) analyses at (a) 1200 UTC 7 December and (b) 0000 UTC 8 December 1989. Dotted line BB' in (a) is the projection line for Fig. 9. Wind vectors are as in Fig. 2. The dashed line denotes frontal boundaries.
Hourly RASS $\theta_e$ profiles extended to almost 400 mb during this 36-h observing period, but did not reach the jet-stream level (~300 mb). Therefore, subsequent time–height analyses of RASS and wind profiler data will extend only to 400 mb. The time–height analysis of RASS $\theta_e$, with companion hourly wind profiles (Fig. 11) shows the wind-direction shift associated with the passage of the geopotential short-wave trough axis (between 0300 and 1100 UTC 7 December) accompanied by a temporally ascending zone of enhanced cooling, static stability, and vertical and temporal wind shear. After 1300 UTC, the RASS and wind profiler observations (Fig. 11) documented the descent of the jet-front system and its associated rapid warming (~3 K h$^{-1}$), large static stability [~16 K (100 mb)$^{-1}$], and large vertical and temporal wind shear (~30 m s$^{-1}$ km$^{-1}$ and ~6 m s$^{-1}$ h$^{-1}$, respectively). The collocation of the vertical wind shear and the local changes in $\theta_e$ indicates that the trough axis and jet-front system contained baroclinic structure. The front-parallel (northerly) wind maximum on the warm side of the jet–front system was consistent with thermal-wind constraints. Following the jet–front descent, the warming and static stability decreased along with the weakening wind-velocity gradients.

Time series of RASS geopotential heights $\Phi$ at 700, 550, and 420 mb (Fig. 12) were calculated as before by hypsometric integration. Between 0200 and 1100 UTC 7 December, the local minimum in $\Phi$ associated with the geopotential short-wave trough axis exhibited a temporal (westward) phase tilt with height. A characteristic cyclonic wind-direction shift was observed across the trough on the three isobaric surfaces. Following the passage of the geopotential short-wave trough axis, the thickness between the isobaric surfaces decreased in response to the approaching thermal-trough axis, indicating that the trough in the temperature field was situated upstream from the trough in the geopotential height field. The upstream displacement (phase lag) of the thermal trough in conjunction with the westward phase tilt with height of the geopotential trough are characteristic signatures of an amplifying baroclinic wave (cyclogenesis was observed southeast of Colorado). After 1300 UTC 7 December, the 420-mb geopotential height rapidly increased and the 420–550-mb thickness correspondingly increased in response to the approaching upstream geopotential ridge and associated warm air. Similar geopotential height and thickness tendencies were observed after 2000 UTC 7 December at 550 mb and in the layer between 550 and 700 mb, respectively. A cyclonic wind-shear zone associated with the jet–front system accompanied the initial, large geopotential height increase at 420 and 550 mb. The measurements of $\Phi$ and wind velocity from the RASS and wind profiler were similar to those from the Stapleton rawinsondes, although only the RASS and wind profiler clearly documented the transient weather systems over eastern Colorado on 7 and 8 December.

c. **Horizontal potential temperature gradient and thermal-advection diagnostics**

Synoptic-scale and mesoscale horizontal virtual potential temperature gradient and thermal-advection profiles were diagnosed from the hourly, “single-station” wind-velocity profiles, using the geostrophic thermal-wind equation (technique as applied to wind profiler data described by Neiman and Shapiro 1989). The single-station thermal retrieval technique has been utilized for many years using twice-daily rawinsonde wind measurements (e.g., Oliver and Oliver 1945; Saucier 1955), but not with the vertical and temporal resolution offered by wind profilers. These approximated thermal gradient and advection diagnostics are most applicable to cases of nearly straight flow, small changes in trajectory curve with height, and minimal accelerations. Because the north–south-oriented jet–front system closely satisfied these constraints, the diagnosed thermal advections are combined in section 5d with RASS measurements of static stability $\partial \theta_e / \partial p$ and virtual potential temperature tendency $\partial \theta_v / \partial t$ to diagnose hourly vertical-velocity profiles.

Previous single-station wind profiler studies using
FIG. 9. Cross-section analyses of (a) virtual potential temperature (K, solid contours) and the front-parallel wind velocity component (m s⁻¹, dashed contours), and (b) cross-horizontal isentropic potential vorticity (x10⁻² K mb⁻¹ s⁻¹, dark contours) and virtual potential temperature (K, shaded contours) along line BB' of Fig. 7a at 1200 UTC 7 December 1989. Rawinsonde sites are, from west to east, Desert Rock, Nevada (DRA); Ely, Nevada (ELY); Salt Lake City, Utah (SLC); Lander, Wyoming (LND); Denver, Colorado (DEN); and North Platte, Nebraska (LBF). Wind vectors are as in Fig. 2.
the geostrophic thermal-wind equation (e.g., Neiman and Shapiro 1989; Crochet et al. 1990) were carried out in the absence of comparably resolved, remote temperature observations. The present study is the first in which the single-station diagnosis of horizontal potential temperature gradients and thermal advections is accompanied by hourly remote \( \theta_e \) observations with comparable vertical and temporal resolution. Taylor's hypothesis is invoked to compare the diagnosed horizontal virtual potential temperature gradients \( \nabla \theta_v \) and thermal advections \( \mathbf{V} \cdot \nabla \theta_v \) with temporal changes of \( \theta_v \) and static stability measured by the RASS.

Figure 13 presents a time–height analysis of \( \nabla \theta_v \) and of RASS \( \theta_v \) for 0000 UTC 7 December to 1200 UTC 8 December. The time–height analysis clearly defines the passage of the jet–front system, showing a temporally descending layer of large baroclinity [3–5 K (100 km)\(^{-1}\)] diagnosed from the hourly wind profiles coinciding with the temporally descending layer of enhanced warming and enhanced static stability observed by the RASS. Within this frontal layer, the diagnosis of warmer air to the west of the profiler is consistent with 1) the RASS \( \theta_v \) analysis of enhanced warming associated, in part, with the eastward advance of warmer air across Colorado, and 2) the rawinsonde-derived horizontal and cross-section analyses (Figs. 7 and 9) showing the east–west orientation of the virtual temperature gradient within the jet–front system. The analyses of the \( \nabla \theta_v \) diagnostics, the RASS \( \theta_v \) observations, and the rawinsonde observations were similarly well correlated following the passage of the jet–front system; the diagnosis of small baroclinity \( (\sim 1 \text{ K} (100 \text{ km})^{-1}) \) and the diagnosis of warmer air to the west and southwest of the profiler were accompanied by decreased warming and weaker static stability during the eastward advance of the weakly baroclinic thermal ridge. The time–height analysis of \( \nabla \theta_v \) and of RASS \( \theta_v \) (Fig. 13) proved equally successful in documenting the passage of the thermal-trough axis at the eastern boundary of the jet–front system. The RASS \( \theta_v \) analysis of the period prior to the jet–front descent shows an axis of cold virtual potential temperatures extending upward from near the surface to at least 400 mb. Above 700 mb, the \( \nabla \theta_v \) vectors across this cold axis shifted in direction from southeasterly to westerly (coldest air to the northwest and east, respectively) in response to the passage of the eastward-propagating thermal trough, thus indicating that the minimum in RASS \( \theta_v \) above 700 mb represented the cold axis of the thermal trough (shown in Figs. 7 and 9). The thermal trough was sharpest in both the \( \nabla \theta_v \) and \( \theta_v \) analyses at \( \sim 450 \text{ mb} \) between 0900 and 1500 UTC 7 December. Below 700 mb, the \( \nabla \theta_v \) vectors across the cold axis remained southwesterly (coldest air to the northeast), during which a shallow arctic air mass migrated southward into northeastern Colorado and Nebraska (not shown). The lower-tropospheric cold axis observed by the RASS was the result of the southward transport of arctic air to the lee of the Continental Divide, rather than the eastward advance of the midtropospheric thermal-trough axis.

An analysis of \( \mathbf{V} \cdot \nabla \theta_v \) and RASS \( \theta_v \) for the same time period as in Fig. 13 is presented in Fig. 14; it shows consistency between the diagnostics and observations. A temporally descending layer of strong warm advection \( (\sim 40 \text{ K day}^{-1}) \) coincided with the temporally descending layer of enhanced warming and static

![Figure 10](image1.png)

**Fig. 10.** Time–height analysis of wind profiler horizontal wind speed (m s\(^{-1}\), solid contours) between 0000 UTC 7 December and 1200 UTC 8 December 1989. Wind vectors are as in Fig. 2. The dotted line marks the wind-direction shift associated with the passage of the geopotential trough axis.

![Figure 11](image2.png)

**Fig. 11.** Time–height analyses of RASS virtual potential temperature (K, dark solid contours) and the wind profiler front-parallel wind-velocity component (m s\(^{-1}\), shaded dashed contours) between 0000 UTC 7 December and 1200 UTC 8 December 1989. Wind vectors are as in Fig. 2. The dotted line marks the wind-direction shift associated with the passage of the geopotential trough axis.
stability associated with the passage of the jet–front system and the advance of the warm air from the west. Prior to the jet–front descent, regions of weak cold advection (\( \sim -5 \text{ K day}^{-1} \)) coincided with cooling observed by the RASS and the approach of the thermal–trough axis. Neutral thermal advection was found in the vicinity of the cold axis.

d. Vertical velocity diagnostics

RASS observations of virtual potential temperature tendency \( \frac{\partial \theta_v}{\partial t} \) and static stability \( \frac{\partial \theta_v}{\partial p} \), together with the diagnosed thermal-advection profiles \( \mathbf{V} \cdot \nabla \theta_v \), were used to diagnose hourly profiles of vertical velocity \( \omega \) through the adiabatic vertical motion equation,

\[
\omega = - \left( \frac{\partial \theta_v}{\partial t} + \mathbf{V} \cdot \nabla \theta_v \right) \left( \frac{\partial \theta_v}{\partial p} \right)^{-1} \quad .
\]

This single-station diagnostic technique is especially suitable for studying nonprecipitating midtropospheric circulations and adds to the growing list of methods by which vertical velocities can be measured or deduced.

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Fig. 12. Time series of RASS and rawinsonde geopotential heights (m, dark and shaded lines, respectively) at 700, 550, and 420 mb, and corresponding wind profiler and rawinsonde wind velocity vectors (dark and shaded vectors, respectively), between 0000 UTC 7 December and 1200 UTC 8 December 1989. Wind vectors are as in Fig. 2.

Fig. 13. Time-height analyses of diagnosed horizontal virtual potential temperature gradients \( \nabla \theta_v \) [K (100 km)]^{-1}, dark contours] and RASS virtual potential temperature (K, shaded contours), between 0000 UTC 7 December and 1200 UTC 8 December 1989. Flags = 5 K (100 km)]^{-1}, full barbs = 1 K (100 km)]^{-1}, half-barbs = 0.5 K (100 km)]^{-1}. The plotted vector shafts point in the direction of the coldest air. The dotted line marks the \( \nabla \theta_v \) vector shift associated with the passage of the thermal–trough axis.
Fig. 14. Time–height analyses of diagnosed horizontal virtual potential temperature advection (K day$^{-1}$, dark contours) and RASS virtual potential temperature (K, shaded contours) between 0000 UTC 7 December and 1200 UTC 8 December 1989. Dark solid contours, warm potential temperature advection; dark dashed contours, neutral potential temperature advection.

by radar wind profiler technology on time scales relevant to mesoscale phenomena. Two other methods consist of kinematic calculations based on a set of at least three wind profilers, and suitably averaged direct measurements using a vertically pointing antenna from a single station. The first approach is the same as that used for rawinsonde data but has the advantage of providing vertical profiles of $\omega$ as frequently as every hour (e.g., Carlson and Forbes 1989). Although the application of this kinematic method has been limited by a questionable lower boundary condition resulting from a lack of wind data between the surface and the lowest radar range gate, future hybrid profiler systems could reduce this problem. The second approach consists of direct measurements of $\omega$. While direct observations are preferable to calculations based on horizontal motions, this method is potentially limited by falling hydrometeors (e.g., Wakasugi et al. 1987; Gossard 1988) and measurement errors due to finite beamwidth effects. In addition, directly measured $\omega$ can be dominated by internal gravity waves that usually have periods from several minutes to a couple of hours, and amplitudes much greater than those expected for large-scale motions. Hence, in studies that wish to reduce gravity-wave effects, suitable time averages must be chosen depending on the specific application, as, for example, in Nastrom et al. (1985). Although the adiabatic method presented here necessarily assumes no diabatic effects are present, the possible limitations of the kinematic and the direct methods previously discussed are reduced or eliminated.

The adiabatic $\omega$ diagnostics are compared here to vertical velocities simulated by the National Meteo-

logical Center’s Nested Grid Model (NGM) and diagnosed from rawinsonde wind-velocity profiles. The diagnosed adiabatic $\omega$ analysis (Fig. 15) shows a characteristic ascent–descent dipole structure associated with the propagating geopotential short wave, in which ascent is diagnosed during most of the period prior to 1200 UTC 7 December and subsidence diagnosed thereafter. This dipole signature extended into the midtroposphere. The NGM 12-h vertical velocity forecast (at 700 mb) valid at 1200 UTC 7 December 1989 (Fig. 16a) also shows a vertical-velocity dipole signature with the short wave. Ascent ($\sim 3$ cm s$^{-1}$ $\approx -3$ mb s$^{-1}$) preceded the trough axis over northeastern Colorado and descent ($\sim -3$ cm s$^{-1}$) followed the trough axis over southwestern Colorado. Twelve hours later (Fig. 16b), the vertical velocity couplet migrated eastward with the short wave, with 700-mb subsidence over Colorado. Kinematic divergence and vertical velocity calculations (Bellamy 1949) from the Denver–North Platte–Dodge City rawinsonde “triangle” (Fig. 17a) gave tropospheric ascent ($\leq -5$ mb s$^{-1}$) ahead of the short-wave trough axis at 1200 UTC 7 December 1989, and 12 h later (Fig. 17b) showed subsidence ($\leq 8$ mb s$^{-1}$) behind the trough axis up to $\sim 375$ mb.

The hourly $\omega$ diagnostics revealed details not resolved by either the NGM forecasts or by the 12-h kinematic calculations. Upward motion of $-4$ to $-8$ mb s$^{-1}$ (Fig. 15) accompanied the passage of the short-wave trough axis between 0300 and 0700 UTC 7 December, followed at 0800 UTC by a brief period of subsidence in colder air centered above 700 mb, suggesting the presence of a kinetic energy–generating circulation (i.e., transverse circulation containing ascent.

Fig. 15. Time–height analyses of diagnosed adiabatic vertical velocity (mb s$^{-1}$, dark contours) and RASS virtual potential temperature (K, shaded contours) between 0000 UTC 7 December and 1200 UTC 8 December 1989. Dark solid contours, downward motion; dark thin-dashed contours, upward motion; dark thick-dashed contours, no vertical motion.
within warm air and descent within cold air). Integrated atmospheric water vapor measured by the NOAA/WPL six-channel radiometer in Denver (Fig. 18) attained a maximum value of nearly 1 cm at the onset of the upward motion between 0200 and 0300 UTC. After 1200 UTC, adiabatic downward-motion maxima (Fig. 15) temporally descended at a rate similar to that of the jet–front system, at which time the integrated water vapor decreased to \( \sim 0.35 \) cm. These results are consistent with the IPV cross-section analysis (Fig. 9b) showing stratospheric IPV within the descending jet–front system, suggesting that dry air parcels originating in the stratosphere descended to mid-tropospheric levels.

e. Gradient Richardson number diagnostics

Observations of RASS static stability \( \partial \theta_e / \partial z \) and wind profiler vertical wind shear \( \partial V / \partial z \) were used to identify preferred regions of vertical-shear instability and associated turbulence through the hourly diagnosis of the gradient Richardson number,

\[
R_i = \left( \frac{g}{\theta_e} \frac{\partial \theta_e}{\partial z} \right) \left( \frac{\partial V}{\partial z} \right)^{-2},
\]

where \( g = 9.8 \text{ m s}^{-2} \) is the gravitational acceleration. A Richardson number of 0.25 is typically cited as the critical value below which a statically stable atmosphere will transform from laminar to turbulent flow as the result of vertical-shear instability.

Figure 19 presents the analysis of hourly Ri diagnostics and shows a layer of small Ri (\( \sim 1.0–2.5 \)) within the temporally descending jet–front layer. Commercial aircraft encountered moderate and severe turbulence (Fig. 19) within the frontal layer, where Ri was minimized although greater than the critical value of 0.25. The diagnosed Ri values within the turbulent frontal layer were subcritical (\( >0.25 \)) because the finescale (\( <100 \) m), turbulence-inducing shear layers internal to fronts are not resolved by the hourly averaged, 300-m vertical resolution wind-velocity profiles measured by the wind profiler. We would expect smaller values of Ri if the vertical and temporal resolution of the observations were increased. Nevertheless, the RASS and wind profiler observations did identify a preferred region of turbulence through the diagnosis of small Ri (\( <<2.5 \)) within the frontal vertical-shear layer.

Because of the present limited vertical range of RASS, the hourly Ri diagnostics can only be applied within the lower and middle troposphere to identify turbulence layers. In addition to alerting low-flying commercial and private aircraft to the potential occurrence of turbulence, the Ri diagnostics can also serve to estimate the severity of air-pollution episodes. The potential for turbulent exchange of chemical contaminants across lower-tropospheric trapping stable layers can be estimated through the diagnosis of Ri. The likelihood of pollution increases when turbulent exchange is inhibited (Ri is large).

f. Isentropic potential vorticity diagnostics

Isentropic potential vorticity \(- (\zeta_s + f)(\partial \theta_e / \partial \rho)\) is the product of the absolute vorticity measured on is-
entropic surfaces \((\zeta + f)\) and the ambient static stability. An estimate of cross-frontal IPV and its temporal evolution was obtained from single-station wind profiler wind-velocity profiles and RASS \(\theta_e\) profiles by invoking Taylor's hypothesis. During the passage of a quasi-two-dimensional weather system that is not rapidly evolving, time-to-space conversion of the front-parallel wind-velocity component \(u\) and RASS \(\theta_e\) can be employed such that IPV can be expressed as

\[
IPV = -\left[ (\frac{-1}{C}) \left( \frac{\partial u}{\partial t} + f \frac{\partial \theta_e}{\partial p} \right) \right],
\]

where \(f \approx 10^{-4} \text{ s}^{-1}\) is the Coriolis parameter, \(C = 10 \text{ m s}^{-1}\) is the phase speed of the weather system, and \(n = 0\) for advancing warm air while \(n = 1\) for advancing cold air.

**Fig. 17.** Kinematic divergence \((\times 10^{-5} \text{ s}^{-1}, \text{ thick line})\) and vertical velocity \((\text{mb} \text{ s}^{-1}, \text{ thin line})\) profiles from the Denver–North Platte–Dodge City rawinsonde triangle at (a) 1200 UTC 7 December, and (b) 0000 UTC 8 December 1989.

**Fig. 18.** Time series of integrated atmospheric water vapor (cm) measured by the radiometric profiler at Stapleton International Airport in Denver between 1200 UTC 6 December and 1200 UTC 8 December 1989.

**Fig. 19.** Time–height analyses of diagnosed gradient Richardson numbers (<2.5, dark contours) and RASS virtual potential temperature (K, shaded contours) between 0000 UTC 7 December and 1200 UTC 8 December 1989. Pilot turbulence reports over Colorado, Wyoming, and Utah were time-to-space adjusted and are shown: MDT = moderate, MSR = moderate to severe, SVR = severe.
The descending jet–front system presented in this study lends itself to single-station IPV diagnostic calculations because the front was quasi–two-dimensional (small alongfront variability) and not rapidly evolving. We stress that observations from a network of RASS and wind profilers would require fewer flow constraints for networkwide IPV calculations. The IPV analysis (Fig. 20) diagnosed from the \( \theta_e \) and \( u \) analyses (Fig. 11) shows the midtropospheric frontal extension of stratospheric IPV associated with the tropopause fold. This IPV anomaly approached \( 3 \times 10^{-3} \, \text{K \, mb}^{-1} \, \text{s}^{-1} \) above 500 mb at 1800 UTC 7 December. These IPV diagnostic results, together with observations of adiabatically diagnosed subsidence (Fig. 15) and decreasing integrated water vapor (Fig. 18) during the passage of the jet–front system, suggest that air parcels with stratospheric origin were transported down the frontal zone into the midtroposphere during the tropopause-fold event. The IPV analysis (Fig. 20) has comparable structure to the IPV cross-section analysis shown in Fig. 9b.

6. Summary and conclusions

This study demonstrates the ability of the RASS and companion wind profilers, possessing high vertical resolution (\( \sim 300 \, \text{m} \)) and high temporal resolution (\( \leq 1 \, \text{h} \)), to continuously observe migrating lower- and midtropospheric weather systems. The NOAA/WPL combined wind and temperature remote-sensing systems clearly documented wind velocity and virtual temperature structures associated with a synoptic-scale trough and embedded fronts between 9 and 12 December 1989, and a propagating short-wave trough and a trailing midtropospheric jet–front system on 7 and 8 December 1989.

Single-station, hourly diagnostic calculations of geopotential heights, horizontal virtual potential temperature gradients, thermal advections, vertical velocities, gradient Richardson numbers, and cross-frontal isentropic potential vorticity demonstrate that dynamically consistent synoptic-scale and mesoscale signals can be derived by combining RASS and wind profiler observations, recognizing the limited scope of single-station diagnostic calculations. The true diagnostic and prognostic potential of the coupled RASS–wind profiler system will be realized only through networkwide assessments in conjunction with wind profiler networks, including the forthcoming 404-MHz NOAA Wind Profiler Demonstration Network in the central United States. Given that the vertical range of RASS with these new 404-MHz wind profilers should extend up to 5 km AGL (Moran et al. 1990), the results presented are compelling for the installation of RASS to network sites to assess the future role of this exciting new technology in research and operational meteorology.

Research and operational network assessment studies will very likely document the synoptic-scale and mesoscale aspects of lower- and midtropospheric processes and structures including land-based cyclogenesis, propagating mesoscale waves and their effect on severe thunderstorm development, frontal circulations, IPV intrusions into the middle troposphere, pollution-trapping inversions, and nocturnal boundary-layer inversions and associated low-level jets. In addition, assessment studies will include initializing and updating short-range (<24 h) mesoscale numerical models with real-time network data (four-dimensional data assimilation—4DDA). Kuo et al. (1987) and Kuo and Guo (1989) have already demonstrated the potential effectiveness of assimilating time-continuous profiler data into a mesoscale model by performing observing-system simulation experiments (OSSEs) with a hypothetical network of profilers. The 404-MHz NOAA Wind Profiler Demonstration Network assessment studies such as described here will be used for guidance in the design and implementation of a next-generation, North American, remote upper-air network.

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