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ABSTRACT

Data from airborne W-band radar, thermodynamic fields from the Weather Research and Forecasting (WRF) Model, and air parcel back trajectories from the Hybrid Single-Particle Lagrangian Integrated Trajectory (HYSPLIT) model are used to investigate the finescale reflectivity, vertical motion, and airmass structure of the comma head of a winter cyclone that produced 15–25 cm of snow across the U.S. Midwest on 29–30 January 2010.

The comma head consisted of three vertically stacked air masses: from bottom to top, an arctic air mass of Canadian origin, a moist cloud-bearing air mass of Gulf of Mexico origin, and a drier air mass originating mostly at low altitudes over Baja California and the Mexican Plateau. The drier air mass capped the entire comma head and significantly influenced precipitation distribution and type across the storm, limiting cloud depth on the warm side, and creating instability with respect to ice-saturated ascent, cloud-top generating cells, and a seeder–feeder process on the cold side. Convective generating cells with depths of 1.5–3.0 km and vertical air velocities of 1–3 m s\(^{-1}\) were ubiquitous atop the cold side of the comma head.

The airmass boundaries within the comma head lacked the thermal contrast commonly observed along fronts in other sectors of extratropical cyclones. The boundary between the Gulf and Canadian air masses, although quite distinct in terms of precipitation distribution, wind, and moisture, was marked by almost no horizontal thermal contrast at the time of observation. The higher-altitude airmass boundary between the Gulf of Mexico and Baja air masses also lacked thermal contrast, with the less-stable Baja air mass overriding the stable Gulf of Mexico air.

1. Introduction

The comma head of wintertime extratropical cyclones is a common locus of heavy snow, blizzards, ice storms, and other hazards as these storms move eastward across the North American continent. In mature cyclones, the comma head is often characterized by a warm occluded frontal structure (e.g., Kuo et al. 1992; Schultz and Mass 1993; Martin 1998a, 1999; Stoelinga et al. 2002; Schultz and Vaughan 2011). In the classical warm occlusion process (Bjerknes and Solberg 1922), air behind the cyclone’s cold front ascends the warm-frontal surface, isolating a wedge of warm air aloft. This wedge, first investigated by Crocker et al. (1947), Godson (1951), and Penner (1955), is sometimes referred to as the trough of warm air aloft, or trowal (e.g., Penner 1955; Galloway 1958, 1960; Martin 1999) and is associated with the comma head clouds and precipitation evident in satellite and radar imagery. Airflow into and through the comma head has also been described in storm-relative coordinates as the westward-flowing branch of a warm
conveyor belt (Schultz 2001). Madonna et al. (2014) provide a recent review of research related to warm conveyor belt flows, Schultz and Mass (1993), Stoelinga et al. (2002), and Schultz and Vaughan (2011) provide thorough reviews of the literature on occlusions, and Martin (1999), Grim et al. (2007), and Han et al. (2007) provide analyses of thermodynamic structure and dynamic forcing of comma head cloud systems.

When viewed with scanning radars, precipitation within the comma head is typically organized in linear, banded features of enhanced reflectivity (e.g., Nicosia and Grumm 1999; Novak et al. 2004, 2009, 2010). The propensity of banding to occur within the comma head has been thoroughly documented (e.g., Martin 1998b; Novak et al. 2004; Browning 2005; Moore et al. 2005; Nicosia and Grumm 1999; Novak et al. 2008, 2009, 2010). These banded precipitation structures are normally identified from low-level radar scans typical of operational S-band (10-cm wavelength) radars such as the National Weather Service Weather Surveillance Radar-1988 Doppler (WSR-88D). Because of their scanning strategy, operational radars provide little insight into the vertical structure of precipitation, and its relationship to the air mass and frontal structures that characterize the comma head. Information about the finescale structure of precipitation comes primarily from vertically pointing radars, which view precipitation as storms pass over a fixed location. Vertically pointing radar measurements within the comma head have consistently shown the presence of generating cells within extratropical cyclones (e.g., Marshall 1953; Gunn et al. 1954; Wexler 1955; Douglas et al. 1957; Wexler and Atlas 1959; Carbone and Bohne 1975; Hobbs and Locatelli 1978; Syrett et al. 1995; Stark et al. 2013; Rosenow et al. 2014). The cells form near cloud top well above fronts, are 1–2 km in horizontal extent, have updrafts with magnitudes of 1–3 m s\(^{-1}\), and produce streamers of precipitation that merge during descent into the stratiform radar echo. Microphysical characteristics of generating cells are described in Plummer et al. (2014).

The advent of high-resolution, airborne Doppler W-band radar (Wang et al. 2012) provides a new opportunity to investigate the finescale structure of precipitation across the comma head of winter storms and its relationship to fronts and other airmass boundaries. The advantage of airborne radar measurements is that they provide spatial information about structure, rather than temporal evolution over a single location. On 30 January 2010, the National Science Foundation/National Center for Atmospheric Research (NSF/NCAR) C-130 aircraft carrying the University of Wyoming Cloud Radar (WCR) flew two transects across the comma head of a winter cyclone that produced 15–25 cm (6–10 in.) of snow in a swath from Kansas to Indiana. The data from this flight provide a unique perspective on structural features of comma head clouds, precipitation, and their relationship to vertical motion, flow fields, and airmass structure that are not easily deduced with conventional and vertically pointing radars.

In this paper, an analysis of two high-resolution radar cross sections derived from WCR measurements made during the 30 January 2010 storm is presented. The analysis uses data from the WCR in conjunction with thermodynamic fields from the Weather Research and Forecasting (WRF) Model and air parcel back trajectories calculated using the National Oceanic and Atmospheric Administration (NOAA)/Air Resources Laboratory Hybrid Single-Particle Lagrangian Integrated Trajectory (HYSPLIT) model (Draxler and Hess 1998; Draxler and Rolph 2014), to elucidate the finescale features of the comma head clouds and precipitation and their relationship to the air masses in which they are embedded. The analyses are unique in that they 1) represent very high-resolution (15 m) cross-sectional radar depictions of comma head cloud and precipitation structure from the elevated cloud shield on the cold side of the storm to the dry slot on the equatorward side; 2) relate flow field, airmass structure, and stability to finescale vertical motion fields; 3) show how air masses from diverse source regions juxtapose to create the structural features characterizing comma head precipitation; and 4) demonstrate important roles of three airstreams in the formation and distribution of precipitation across the comma head.

2. Data sources and analysis methods

The measurements reported here were made during the 2009–10 Profiling of Winter Storms (PLOWS) field campaign (Rauber et al. 2014; Rosenow et al. 2014). This paper focuses on data from the WCR collected on 30 January 2010. The WCR is described in Wang et al. (2012) and its use in PLOWS is described in Rosenow et al. (2014). Summarizing key points from Rosenow et al. (2014), the WCR had two downward-looking beams, one at nadir, and one 34.3° aft of nadir, and a single, upward beam, with the width of all beams <1°. The WCR data are available at a horizontal scale of 4–7.5 m at C-130 nominal airspeeds between 100 and 150 m s\(^{-1}\). The WCR typically transmitted a 250 ns (37.5 m) pulse, sampled at 15-m resolution. The unambiguous range was 9 km, and the unambiguous velocity width was 26.3 m s\(^{-1}\). The orientation of the WCR beams in ground-relative coordinates was computed using the aircraft roll, heading, and pitch angles and the ground-relative aircraft velocity. These were used to identify and correct for the component of aircraft motion along each of the beams, and to
determine the vertical component of the radial velocity from the upward- and downward-pointed beams. The variance of the computed radial velocities was \(<0.04 \text{ m}^2 \text{s}^{-2}\) at reflectivity factors \(>\sim 23 \text{ dBZ}_e\). Rosenow et al. (2014) describe the procedure to derive the vertical component of the radial velocity \(W\).

The equivalent radar reflectivity factor \((Z_e, \text{ hereafter reflectivity})\) and \(W\) measurements were regridded from aircraft-relative to ground-relative coordinates. The \(W\) values are shown positive upward in all figures depicting the WCR vertical radial velocity measurements. Rosenow et al. (2014) showed that in stratiform regions of the comma heads of winter storms observed during PLOWS, the reflectivity-weighted terminal velocity of ice crystals and snowflakes \(V_T\), after accounting for the decrease of atmospheric density with altitude, ranges from about \(-0.7 to -1.0 \text{ m s}^{-1}\) over most of the cloud depth. Therefore, a rough estimate of \(w\), the vertical air motion, can be made by adding 1 m s\(^{-1}\) to the \(W\) values.

The 34.3° aft beam simultaneously sampled the vertical motion and horizontal motion of hydrometeors. The horizontal motion of hydrometeors sampled by the radar is related to the horizontal wind component in the direction of flight. The vertical motion of hydrometeors is related to the reflectivity-weighted fall speed of the precipitation plus any vertical air motion. As a result of the geometry, the 34.3° slant beam radial velocity measures 56% of the horizontal and 83% of the vertical component.

A 3-h forecast with the WRF Model (version 3.5.1) was used to obtain fields to overlay upon and aid in interpretation of the WCR data. Forecast fields included potential temperature \((\theta)\), equivalent potential temperature derived with respect to ice \((\theta_e)\), relative humidity with respect to water \((\text{RH}_w)\) and ice \((\text{RH}_i)\), convective available potential energy with respect to ice-saturated ascent \((\text{CAPE})\), and large-scale vertical air velocity \((\pi)\), all in planes corresponding to the WCR cross sections. The WRF simulation was initialized with National Centers for Environmental Prediction (NCEP) North American Mesoscale Model (NAM) analyses (218 grid, 12-km spacing) for a domain centered near 34°N, 89°W with 9-km grid spacing and 220 vertical layers. The simulation employed positive-definite advection, a gravity wave damping layer near the domain top, Thompson microphysics (2013 version, 300 cm\(^{-3}\) cloud droplet concentration used for continental settings), Rapid Radiative Transfer Model for Global Climate Models (RRTMG) shortwave/longwave radiation, the Noah land surface model, the Yonsei University boundary layer treatment, the Kain–Fritsch cumulus parameterization with the default (Kain 2004) trigger function, and 2D Smagorinsky (diagnosed from deformation) diffusion. The model was initialized at 0000 UTC 30 January. Forecast fields for 0300 UTC, the central time of the flight legs across the comma head, were interpolated to the WCR cross sections.

The 3-h WRF forecast was used rather than the 0000 UTC NAM initialization because 1) the nearest NAM initialization was several hours prior to aircraft data collected between 0225 and 0422 UTC, 2) the vertical resolution of the NAM analyses was insufficient in the layer of the generating cells and in the vicinity of airmass boundaries to properly interpret the structure and stability of these features, 3) the 3-h forecast allowed spin up and development of vertical motions within the model, and 4) the 3-h WRF forecast closely matched the structures revealed by the WCR airborne observations.

As a test, we also examined WRF simulations with the same parameters but without cumulus parameterization, and a 9-h simulation initialized at 1800 UTC 29 January and valid at 0300 UTC. There were minor differences in the vertical velocity distribution at the generating cell level between these simulations. The differences were sufficiently small to have no impact on the findings of this paper. The utilized 3-h forecast was found to best match the WCR structures and was therefore selected for analyses overlaid on the WCR figures to provide larger-scale context for the detailed WCR observations.

Air parcel back trajectories were calculated using the HYSPLIT model, August 2013 revision (515) version (Draxler and Hess 1998; Draxler and Rolph 2014). The meteorological dataset used to initialize each HYSPLIT trajectory was the NCEP NAM Data Assimilation System (NDAS; Rogers et al. 2009). The archived data have a grid spacing of 12 km and are available every 3 h. HYSPLIT has been used in past studies of winter storms (Grim et al. 2007; Fuhrmann and Konrad 2013) to determine airmass source regions. In this paper, the HYSPLIT model is used to calculate 48-h back trajectories for air parcels located at every 1 km in the vertical and \(\sim 40\) km in the horizontal (5-min flight time) along the WCR cross sections.

3. Storm overview

The 29–30 January 2010 cyclone was a significant winter storm that produced between 15–25 cm (6–10 in.) of snow across Iowa, Missouri, Illinois, and Indiana. The surface low pressure center associated with the cyclone formed east of the Sierra Madre Occidental in Mexico near 0000 UTC 29 January, moved over the Gulf of Mexico just east of the southern tip of Texas, progressed northeastward over water into southern Louisiana, and was located in east-central Mississippi at the time that
the C-130 aircraft was crossing the cyclone’s comma head in Missouri and Illinois on 0300 UTC 30 January (Fig. 1d). During the 27-h transit across Mexico and the Gulf of Mexico, the central sea level pressure of the cyclone decreased 3 hPa, from 1010 to 1007 hPa. During the event, subfreezing surface temperatures extended southward into Texas, Louisiana, and Alabama.

The cyclone was associated with a wide upper-tropospheric trough that formed within the southern branch of the jet stream. By 0000 UTC 30 January, just before the C-130 flight, the trough was located east of the southern Rocky Mountains and over the southern Great Plains (Fig. 1a). A westward-tilted, closed circulation was present between the surface and 500 hPa within the trough (Figs. 1b–d). The comma-shaped precipitation shield of the cyclone was located north and east of the center of circulation at 500 hPa (Fig. 1b).

Figure 2 shows a Geostationary Operational Environmental Satellite (GOES) enhanced water vapor satellite image of the cyclone during the flight at 0300 UTC. Superimposed on the image are the surface fronts and low pressure center, the 500-hPa relative humidity with respect to water, and surface isobars. Surface isotherms are shown in the vicinity of the comma head. The heavy white line shows the track and direction of flight of the C-130. The western flight leg (WFL) was flown between 0204 and 0324 UTC, and the eastern flight leg (EFL) between 0324 and 0422 UTC. Three features of importance in Fig. 2 are relevant to the subsequent discussion. First, from the water vapor imagery and the 500-hPa relative humidity distribution, note that the drier air associated with the cyclone’s dry slot appears to override moist air along the southern side of the EFL in southeast Missouri. Farther to the west, the dry airstream appears to split over Oklahoma, flow along and override moist air at lower levels on either side of the comma head, and cross the WFL on both its northern and southern ends. Second, note the moisture located along the central axis of the comma head. As will be shown later, this region corresponded to the location of the deepest clouds and was coincident with the heaviest snowfall accumulation at the ground. Finally, note the surface temperatures.
During this cyclone, arctic air had intruded far enough south so that temperatures were well below freezing across the comma head region in the vicinity of both the WFL and EFL. Consequently, most precipitation beneath both flight legs was in the form of snow. There were some reports of freezing drizzle where the WFL and EFL join.

Figure 3 shows a composite WSR-88D image at 0300 UTC with the WFL and EFL and surface frontal analyses superimposed. This image confirms that the locally strong precipitation radar echo extended across the central part of both flight legs. Based on the radar echo intensity, only light to trace precipitation fell on the southern side of the comma head beneath the regions where the WFL and EFL legs join in northern Arkansas, and no precipitation fell at all along the northern third of the EFL. The reason for this snowfall distribution will become apparent when we examine the finescale WCR and WRF analyses in the following section.

4. Finescale structure of the comma head

a. Eastern flight leg

1) AIRM ASS STRUCTURE AND AIR PARCEL TRAJECTORIES

The radar reflectivity measured by the WCR along the EFL of the C-130 appears in Figs. 4a, b and 5a. Superimposed in Fig. 4a are analyses of \( \theta_{ci} \) and \( RH_i \), and in Fig. 4b, analyses of \( \theta \) and \( RH_w \), all at 0300 UTC 30 January. Superimposed in Fig. 5a are points from which 48-h back trajectories were calculated using HYSPLIT. Although these are backward trajectories, the points in Fig. 5a are referred to as the endpoints of the trajectories, since, in time, air parcels moved from some other location at 0400 UTC 28 January to their position on the cross section at 0400 UTC 30 January. The starting points of the trajectories at 0400 UTC 28 January are shown in Fig. 6. Best estimates of airmass boundaries, denoted by frontal symbols, were determined by considering the \( \theta_{ci} \) field, air parcel back trajectories initiated from points on the cross section (Fig. 5a), and the WCR 34° slant beam radial velocity (Fig. 5b).

The comma head of the cyclone comprised three vertically stacked air masses. For convenience, we will refer to these air masses (from bottom to top) as the Canadian, Gulf, and Baja air masses, although some trajectories did not originate directly over Canada, the Gulf of Mexico, and the Baja Peninsula. Air parcels located within the Canadian air mass originated 48 h earlier at locations across central Canada and the northern United States (cf. square symbols in Figs. 5a and 6). The Canadian air was 1–2 km deep along the southern half of the EFL. Example trajectories in Fig. 7a show that air parcels arriving at the 1-km level (all heights above mean sea level) underwent subsidence on
their path to the cross section, typically descending 2–4 km during the transit from Canada to the EFL (see trajectories $\kappa$, $\varepsilon$, and $\beta$ in Fig. 7a, and the inset in Fig. 6). The trajectories did not exhibit much cyclonic curvature, as exhibited in cold conveyor belt type flows in stronger cyclones [e.g., Mass and Schultz (1993, their Fig. 16), Martin (1998b, his Fig. 16)]. The slope of the airmass boundary changed near the center of the cross section, with the cold air mass deepening to 5 km at the northern end of the cross section. Unlike the near-surface air, air within the Canadian air mass arriving at 3–4-km altitude maintained a near-constant altitude along its trajectory (see trajectories $\Psi$ and $\rho$). Air at intermediate altitudes within the Canadian air subsided ~1 km (e.g., trajectory $\theta$). As indicated by the RH$_w$ and RH$_I$ fields, the Canadian air was sufficiently dry that precipitation falling from aloft sublimated after entering the Canadian air, particularly on the northern side of the cross section where the Canadian air was deeper.

The boundary between air originating over Canada and air originating over the Gulf of Mexico (depicted as a cold front in Figs. 4 and 5) did not have characteristics of a true front. From a dynamic perspective, fronts are associated with density gradients and marked by steep isentropes and a discontinuity in the gradient of potential temperature (e.g., Bluestein 1993, 245–248; Martin 2006, 189–193). The airmass boundary is evident in the slant-beam radial velocity in Fig. 5b, the boundary marked by the transition between air with a northerly flow component and air with a southerly component. The boundary also coincides with sharp vertical and horizontal gradients of $\theta_e$ (Figs. 4a and 5b) and closely matches the results of the trajectory analyses. However, the boundary is not obvious in the potential temperature field and thus cannot be characterized as a true front. Note that the descending Canadian air was unsaturated, while the rising Gulf air was supersaturated with respect to ice (Fig. 4a). The trajectories suggest that over the 48-h period before the aircraft sampled the storm, dry adiabatic warming within the descending Canadian air and moist adiabatic cooling within the rising Gulf air largely homogenized the temperature field across the comma head at the level of the airmass boundary at the time the aircraft sampled the storm.

The central midtropospheric air mass depicted in Figs. 4 and 5 originated over or near the Gulf of Mexico (see Fig. 7b, cf. the circle symbols in Figs. 5a and 6). Along the EFL, air within the Gulf air mass was supersaturated with respect to ice, providing an environment for ice particle growth (Fig. 4a). RH$_w$ values within the
air mass approached or exceeded 90%, and even reached 100% in the shallow clouds on the southern end of the EFL where freezing drizzle was observed at the surface (Fig. 4b). Example trajectories of the Gulf air are shown in Fig. 7b. The trajectory analyses indicate that Gulf air arriving at the cross section ascended 2–4 km to its position along the EFL (e.g., trajectories 1, I, L, R, M and E in Fig. 7b, also compare circle symbols in Figs. 5a and 6). In nearly all trajectories, Gulf air initially flowed northwestward, curving back northeastward to arrive at the EFL cross section. These trajectories differ from those described in Martin et al. (1998a, his Fig. 19) in that they do not turn cyclonically westward into the comma head.

The upper-tropospheric air mass depicted in Figs. 4–5 originated near Baja California and the Mexican Plateau. The boundary between the Gulf and Baja air masses appears on satellite water vapor images (e.g., Fig. 2) as an extension of the leading edge of the cyclone’s dry slot boundary westward along (and over) the comma head. Again, from a dynamic perspective, the boundary between the Gulf and Baja air did not have characteristics of a true front—the isentropes are nearly horizontal (Fig. 4b). However, the boundary is evident by a transition to much drier air aloft and a marked change in stability. As will be shown in the next section, cloud-top generating cells with vertical air velocity of 1–4 m s$^{-1}$ were ubiquitous near the cloud top along the central part of the EFL cross section within the Baja air mass. The exact position of the Gulf–Baja air boundary was not as distinct as the Canadian–Gulf boundary. In Fig. 4 and subsequent figures, the boundary (marked with upper-level frontal symbols) was placed at the base of the generating cell layer, which corresponds with the change in stability and most trajectories. A few trajectories near the northern edge of the cloud shield below the marked boundary were from the Baja region, while some above the marked boundary in the generating cell region tracked back to the Gulf. These differences may be related to differences between the WRF simulation and the NAM/NDAS data (on which the HYSPLIT trajectories were based). Trajectory analyses of the Baja air (see Fig. 7c and star symbols in Figs. 5a and 6) indicate that most of the air arriving at the EFL cross section originated at lower altitudes over the Mexican Plateau and Baja California 48 h earlier and
ascended to its position on the EFL cross section. For example, trajectories I and b in Fig. 7c ascended 6 and 4.7 km, respectively, from just west of Baja California to the EFL cross section. Some Baja trajectories loop near the New Mexico–Texas border. The reason for the looping behavior is discussed below. Some trajectories arriving in the upper troposphere on the EFL cross section originated at higher altitudes and underwent only a small amount of ascent. All of these were similar to trajectory o, progressing southward from California and Nevada and then northeastward to the EFL cross section. The Baja air approached the EFL cross section, effectively capping and secluding the moist Gulf air on its north and south sides (see also Fig. 2). Although the Baja air was drier than the Gulf air mass (Fig. 4b), a result of its history of ascent, as well as the cold temperatures in the upper troposphere. As will be discussed later, this was important to both generating cell formation and the formation of precipitation particles in the cloud-top region of the storm.

There were several trajectories (denoted by hexagons and rectangles in Figs. 5a and 6) that did not conform to the three primary groups. The endpoints of these trajectories in all cases were very close to an airmass boundary (Fig. 5a). The starting points of these trajectories were in geographic locations located between the starting points of air arriving on the EFL cross section from air masses on either side of the boundary (Fig. 6). The evolution of trajectories in Fig. 7 can be better understood by considering their positions in the context of evolving tropospheric flow patterns. Figure 8 shows a series of four panels at 1200 UTC 28 January (Fig. 8a),

Fig. 5. (a) WCR equivalent radar reflectivity factor (dBZ) for the eastern flight leg. Each symbol in (a) denotes the 0400 UTC 30 Jan location of a HYSPLIT trajectory. The opposite end of each trajectory 48 h earlier at 0400 28 Jan 2010 appears in Fig. 6. Select trajectories are shown in Figs. 7, 8, and 9. The squares, circles, and stars had trajectories that originated 48 h earlier over Canada, the Gulf of Mexico, and the Baja California region, respectively. The hexagons and rectangles very close to the airmass boundaries had transitional trajectories that ended between the three primary locations. (b) Radial velocity from the 34° slant beam of the WCR. The boundary between red and blue has been adjusted by 1 m s⁻¹ to account for the contribution of terminal velocity of snow to the radial velocity, so that red represents a horizontal wind from south to north (left to right) and blue from north to south. The θ_e field is superimposed (thin black lines, K). The C-130 flight track is shown in red. Airmass boundaries are denoted with frontal symbols.
0000 UTC 29 January (Fig. 8b), 1200 UTC 29 January (Fig. 8c), and 0000 UTC 30 January (Fig. 8d), each showing 500- and 975-hPa height fields with a few key trajectories superimposed. The highlighted portions of the trajectories are within ±3 h of the time of the maps. The numbers show the starting and ending pressure–altitudes of the trajectories, and the pressure–altitude of each trajectory at the time of each map.

The Canadian air flowed southward from central Canada to the EFL cross section under the influence of a deep cutoff low centered between Hudson Bay and the Great Lakes. Beginning about 1800 UTC 30 January, the Canadian trajectories then turned westward under the influence of easterly flow north of the cyclone which, at 0000 UTC 30 January, was positioned over northeast Louisiana. Gulf air that arrived at the EFL first flowed northward and then ascended to 623 hPa as it moved up the slopes of the Sierra Madre Occidental in western Mexico, and then flowed northeastward, farther ascending on the east side of the mid- and upper-tropospheric trough as it approached the EFL. Air associated with trajectory w started at a higher pressure–altitude (745 hPa), descended to 794 hPa as it rounded the west side of the trough, and then ascended to 352 hPa as it flowed on the east side of the trough to its position on the EFL cross section. Air associated with trajectory c, one of the looping trajectories, started at 456 hPa, first ascended, and then descended to a position just northeast of the closed low circulation at the center of the trough. From that position, the air flowed westward within the easterly flow north of the low center. The air then turned south to the west of the low, rounding the base of the upper-level low and proceeding eastward toward the EFL. The net ascent along this trajectory was not as great as along trajectories originating at other locations over the Baja region.
2) PRECIPITATION, STABILITY, AND VERTICAL MOTION

Precipitation along the EFL can be divided into three zones. In the southern zone (Fig. 4, 0325–0345 UTC), Baja air overran the Gulf air mass capping the clouds at 4–6 km. The cloud top sloped upward to the north within this zone. Based on surface reports at the time of the flight, precipitation was very light beneath this zone, and was a mix of light snow and freezing drizzle. Beneath the

FIG. 7. Example trajectories from the Canadian (squares in Figs. 5a and 6), Gulf of Mexico (circles in Figs. 5a and 6), and Baja air masses (stars in Figs. 5a and 6). The altitudes (km, MSL) of the trajectory start and end points are indicated. The letters for each trajectory correspond to Figs. 5a and 6.
central zone (0345–0355 UTC), reports of light to moderate snowfall were common. Clouds in the central region were 8–9 km deep and extended well into the Baja air mass. In the northern zone of the comma head (0355–0423 UTC), snow formed aloft, but fell into the dry Canadian air mass where it completely sublimated, producing virgalike radar echoes at the base of the precipitation shield (Fig. 4). No precipitation reached the ground in this zone.

Figure 9b shows $W$ measured by the WCR along the EFL. Figure 9c shows $\pi$ from the 3-h WRF forecast. Superimposed in Figs. 9b,c are analyses of $\theta_{ei}$ at 0300 UTC. The light blue line in Fig. 9b shows the level of free convection, and the light red line the equilibrium level for the most unstable parcel for ice saturated ascent in areas along the EFL cross section where CAPE$_i$ exists. The values of CAPE$_i$ for the most unstable parcel are plotted above Fig. 9a. Note from Fig. 4a that the entire atmosphere encompassing the radar echoes in the cloud-top region, including the areas where CAPE$_i$ exists, were supersaturated with respect to ice.

Based on the vertical gradient of $\theta_{ei}$ in Fig. 9b, the Gulf and Canadian air masses were stable, except in the (shallow) boundary layer. Based on the WRF solution in Fig. 9c, and the HYPSPLIT trajectories in Figs. 7 and 8, both the Gulf and Baja air masses were rising as a result of the larger-scale forcing associated with the cyclone and advancing trough (Fig. 8). The convective-scale vertical motions at cloud top (and cloud base in the northern zone), and associated structure of clouds and precipitation in each zone are considered separately below.

The vertical air motions along the EFL are best understood by examining together Figs. 9b,c and 10, which shows contour frequency by altitude diagrams (CFADs; Yuter and Houze 1995) of $W$ in regions A–D in Fig. 9b. In general, because of the contribution of hydrometeor terminal velocity to $W$, $W$ will underestimate true...
convective-scale updraft intensity by $\sim0.5$–$1.0$ m s$^{-1}$ and overestimate downdraft intensity by the same amount (Rosenow et al. 2014). This is important to keep in mind when examining Fig. 9b and the CFADs in Fig. 10.

In the southern zone, the clouds were capped at 4.5–6.0 km by the Baja air. The $\theta_{ei}$ distribution near the cloud top, and the ice-saturated conditions in the presence of 0–40 J kg$^{-1}$ of CAPE$_i$, together indicate that air in the cloud-top region in the southern zone was unstable. We note that the flight occurred at night, and cloud-top radiative cooling may also have contributed to the instability, although its role, if any, is difficult to assess. The instability manifested in a shallow, layer near the cloud top, where 0.7-km-deep generating cells were ubiquitous (Figs. 10a and 11). The aircraft passed through the tops of these cells at $\sim12^\circ$C and encountered turbulence and aircraft icing. As a result of turbulence, errors were introduced into the corrected Doppler velocities due to both flexing of the airframe and motions that are not well resolved by the aircraft navigation system. These led to oscillations in vertical radar beam position, and thus in radial velocity (see vertical striping in Fig. 11b). For this reason, the magnitudes of $W$ appearing in Fig. 11b and region A of Fig. 9b are less certain in this zone. With these greater uncertainties in mind, stronger updrafts in the generating cell region ranged from $\sim0.5$ to $1.5$ m s$^{-1}$, based on the CFAD for region A in Fig. 10.

Each cell produced a fall streak of precipitation that subsequently merged with neighboring fall streaks in the clouds below. Below the generating cell level, modal values of $W$ (Fig. 10, region A) suggest that the clouds were stratiform and that snow was falling at or near its terminal velocity.

The WRF $\theta_{ei}$ distribution and the CAPE$_i$ values in the Baja air mass in the central and northern regions suggests that Baja air in these zones was stable except in some regions where CAPE$_i$ reached values of 20 J kg$^{-1}$. The distribution of $W$ in the Baja air mass suggests that...
the instability was more widespread than the WRF CAPE values suggest, particularly north of about 0347 UTC (~180 km) on the cross section in Fig. 9b. In this region, 1.5–2-km-deep cloud-top convective generating cells were ubiquitous. Figure 12 shows a 1:1 view of reflectivity and W associated with the generating cells in the central zone. The cells were 1–2 km wide, and were within the Baja air mass. The CFADs in Fig. 10 together with Fig. 9b shows that the updrafts originated very near the boundary between the Baja and Gulf air masses so that Gulf air may have been transported upward into the convective updrafts. From the CFADs the convective vertical air velocity can be estimated. Within the Baja air mass, the stronger updrafts approached values of 3–5 m s\(^{-1}\) in region B and 3–4 m s\(^{-1}\) in region C in Fig. 9b. Downdrafts of 1–2 m s\(^{-1}\) were also present between the updrafts. The slightly stronger updrafts in region B were associated with the cell at the interface between the southern and central precipitation zone. Below the Baja air mass, based on the CFADs, snow fell to the surface at speeds near terminal velocity. The values of vertical motion observed in generating cells in this storm are consistent with measurements in other PLOWS cyclones reported by Rosenow et al. (2014).

The upper part of the northern zone of the EFL contained generating cells with similar updraft and downdraft structure as the central zone (cf. CFADs from region D with region B and C in Fig. 10). The difference between these zones was the sublimation just below the boundary between the Canadian and Gulf air masses and its effect on downdraft generation. Figure 13 shows the reflectivity and W field at a 1:1 ratio in a small

![Fig. 10. CFAD of W measured by the WCR for regions A–D in Fig. 9b. Color shaded values indicate percentage of observations at that altitude falling in each 0.2 m s\(^{-1}\) velocity bin. The break in the diagram near 6 km is the aircraft altitude. The numbers on the contours denote the percent of observations with values to the left of the contour.](image-url)
sector of the zone. Sharp 1–3 m s\(^{-1}\) downdrafts developed as snow sublimated into, and cooled air within, the top of the Canadian air. These coincided with the virgalike streaks in the reflectivity field. The overall impact on the distribution of vertical motions at the base of the cloud shield is evident from the CFAD for region D in Fig. 10. Schultz and Trapp (2003) review how sublimation into dry air can impact cold frontal structure. In this case, the impacts were small, reducing the potential temperature over an approximately 1-km-deep layer at the base of the cloud shield (Fig. 4b).

b. Western flight leg

Radar analyses from the WCR along the WFL of the C-130, similar to Figs. 4–5 for the EFL appear in Figs. 14–15. The starting points of the trajectories at 0200 UTC 28 January (for regions sampled by the aircraft before 0230 UTC) and 0300 UTC 28 January (for regions sampled by the aircraft after 0230 UTC) are shown in Fig. 16. Some aspects of the comma head were common to both flight legs. In particular, the comma head in both locations comprised three vertically stacked air masses with trajectory starting points in the lowest to highest air mass originating, respectively, over Canada, the Gulf of Mexico, and Baja California (cf. Figs. 15a and 16). The Gulf and Canadian air masses were stable. Based on the potential temperature field, airmass boundaries did not have characteristics of fronts. The evolution of the trajectories in the three air masses was the same (i.e., ascending from the Baja and Gulf regions and descending from Canada to their positions on the cross sections). In this section, we limit our discussion to aspects of airmass structure, trajectories, and stability unique to this region of the comma head.
1) AIRMASS STRUCTURE AND AIR PARCEL TRAJECTORIES

Focusing first on the Gulf air mass, Figs. 15a and 16 show two groups of trajectories denoted by rectangles and circles. Trajectories ending at the circles in Fig. 15a originated over or near the Gulf of Mexico and followed paths similar to Fig. 7b for the EFL. However, air arriving just above the Canadian air mass, denoted by rectangles, followed different trajectories. The starting points for these trajectories were near the WFL cross section (Fig. 16). During the ensuing 48 h, the air parcels migrated eastward and then westward within a few hundred kilometers radius before returning to their endpoints at the WFL.

The most substantial difference between the WFL and EFL were some trajectories within the Baja air mass. As noted earlier in Fig. 2, dry air on the water vapor image appeared to flow around both the north and south sides of the comma head. On the WFL cross section in Fig. 15a, this air (indicated by stars) is evident on both sides of the WFL and indeed is deepest on the northwest and southeast ends of the cross section. As with the EFL, the Baja air originated primarily at altitudes between the surface and 4 km over Baja California and ascended to higher altitudes in the vicinity of the comma head, although some trajectories again originated at higher altitudes.

One significant difference on the WFL, based on the trajectories, was that Baja air apparently mixed in some regions with air of other origins. On the north-west and southeast end of the cross sections, air flowing around the comma head region apparently mixed with air ascending from low altitudes over central Mexico and southern Texas (see hexagons in Figs. 15a and 16), and air originating at low altitudes over northern Texas, Kansas, and Missouri (see upright and inverted triangles).

2) PRECIPITATION, STABILITY, AND VERTICAL MOTION

Precipitation ranged from light snow and freezing drizzle on the southeast end of the WFL to light to moderate snow in the central region, to scattered snow showers at the northwest end. The most significant difference between the WFL and EFL was the clear presence of instability with respect to ice-saturated ascent within the Baja air mass across the WFL. Note from Fig. 14a that the entire Baja air mass encompassing the generating cells across the WFL was supersaturated with respect to ice. Based on Fig. 17c and the HYSPLIT trajectory analysis, both the Gulf and Baja air masses were ascending in the vicinity of the WFL cross section. The CAPE values in the Baja air mass ranged as high 20 J kg$^{-1}$ across much of the comma head (Fig. 17a). CAPE values of 10–20 J kg$^{-1}$ correspond to maximum adiabatic updrafts of 4.5–6.3 m s$^{-1}$. Figure 18 shows two CFADs, one in the region of the deep generating cells (region E), and the second farther to the southeast (region F). Convective generating cells were present in both regions, but the cells in the northwestern region were deeper, and the vertical air motion somewhat greater. The stronger updrafts, after accounting for particle terminal velocities, ranged from 2 to 3 m s$^{-1}$, about half the adiabatic values. Figure 19 shows a 1:1 ratio depiction of one of the generating cells. In this figure, the plume of precipitation forming within the cell can be seen to be sheared to the southeast as it descends through the 7-km level. The strongest vertical motions within these generating cells were primarily between the

FIG. 13. (a) WCR reflectivity and (b) vertical radial velocity W, shown in a 1:1 aspect ratio, for a small section of region D in Fig. 9b. The red line is the C-130 flight track.
7.5- and 9-km altitude, entirely in the Baja air mass, and centered within the region between the level of free convection and the equilibrium level for ice-saturated ascent (Fig. 16b).

5. Discussion

A significant body of work has been directed at understanding how frontal structures evolve and the occlusion process proceeds at the interface of the warm, cold, and dry airstreams within the comma head region of extratropical cyclones. Schultz and Vaughan (2011) provide a thorough historical review of past and current conceptual models of occlusions, clearly demonstrating how occlusion occurs as three air masses—warm, cold, and dry—wrap into the circulation of the extratropical cyclone central vortex. Of particular relevance to this paper are works of Kreitzberg (1964), Kreitzberg (1968), and Kreitzberg and Brown (1970), which were the first papers to demonstrate that complex mesoscale frontal structures can characterize the occluded region of cyclones and modify the distribution of precipitation within the comma head. They noted that potential instability was triggered behind a precold-frontal midtropospheric dry surge of subsiding air moving over a deep moist stable layer associated with a warm front. Mass and Schultz (1993) subsequently showed how dry slot air intrudes over the southern quadrant of the comma head (see their Fig. 10) in their modeling study of a strong cyclone over the midcontinent. The basic structure in the Mass and Schultz (1993) simulation has since been confirmed in observations reported by Grim et al. (2007) and Rauber et al. (2014), the latter paper also demonstrating that the overrunning of dry air can be associated with potential instability and trigger winter thunderstorms. Browning (1997) also discusses how potential instability can be generated at the moist-dry interface at the top of a moist air mass overrun by dry air [we note that the intrusion of dry slot air over the southern quadrant of the comma head does not necessarily create instability (e.g., Novak et al. 2009)].

Trajectory analyses of the dry airstream, first proposed by Danielsen (1964) and subsequently diagnosed in modeling studies of a historically strong coastal cyclone by Whitaker et al. (1988) and a continental cyclone by Schultz and Mass (1993), show that the dry
airstream typically descends to its position in the dry slot from the west side of an upper-tropospheric trough, often in association with a tropopause fold (e.g., Danielsen 1980, his Fig. 12). Kreitzberg and Brown (1970) also noted in their case study that the dry air mass was subsiding.

This case study presented here differs from these previous studies in that the cyclone considered here was weak, forming in association with a broad upper-tropospheric trough (Fig. 1), and having its central sea level pressure decrease only 3 hPa during the 27 h prior to the period of observation. The dry airstream observed in this storm originated primarily at low levels over the Baja Peninsula, the eastern Pacific, and western plateau regions of Mexico 48 h earlier. A key difference between this and previous studies is that this airstream ascended throughout most of its path, with some trajectories ascending as much as 4–6 km. From a satellite perspective (Fig. 2), this airstream appeared as the cyclone’s dry slot, with its northern boundary roughly defining the southern edge of the cloud shield of the comma head. The trajectory analyses, together with the radar and WRF analyses, show that air associated within this airstream capped the entire comma head. The juxtaposition of the Baja air mass atop the Gulf air significantly influenced the type and distribution of precipitation across the comma head region.

On the southern side of the comma head, the Baja airstream limited the cloud depth, thus limiting surface snowfall. Visually, from an infrared satellite perspective, these shallower clouds appeared as part of the cyclone’s dry slot. Within this region, cloud tops were sufficiently warm to limit ice production in the clouds below, leading to aircraft icing at cloud top and freezing drizzle at the surface.

Over and near the more steeply sloped Canadian air mass, the Gulf air mass was deeper and moderate snow fell at the surface. The analyses show that snow falling in
this region originated in convective generating cells that
developed within the Baja airstream capping the Gulf
air mass. A key to understanding the nature of the in-
stability in this region is to recognize that the instability
must be analyzed with respect to ice processes. The
rising trajectories within the Baja air mass led to the
entire atmosphere encompassing the generating cells in
the cloud-top region to be supersaturated with respect to
ice, despite being well subsaturated with respect to wa-
ter. Much of the layer where generating cells were
present along the WFL, and some areas along the EFL,
posessed convective available potential energy for ice-
saturated ascent of the order of 5–20 J kg\(^{-1}\), corre-
sponding to adiabatic updraft intensities of 4.5–6.3 m s\(^{-1}\)
about twice the values measured in the generating cells
by the WCR. Some regions along the EFL possessed no
CAPE\(_i\) according to the WRF simulation, but still
produced generating cells. It may be possible that radi-
active cooling at the cloud top may have contributed to
the development of instability in this region (the flight
was after dark), but evaluating these effects are beyond
the scope of this work. The use of potential, rather than
conditional instability here is appropriate, since the
layers bounding the interface between the Gulf and Baja
air aloft were both rising based on the larger-scale ver-
tical motion field, and the Gulf and Baja air was su-
persaturated with respect to ice, so if any potential
instability existed, it would have been released. The
generating cells in some locations extended from the top
of the Gulf air well into the Baja air mass. In other lo-
cations, the base of the generating cells appeared to
originate in the Baja air mass. Reflectivity streamers
from the generating cells typically merged into the more
uniform echo within the Gulf air mass, although some
streamers could be traced from the cloud top to the surface. These observations point to an important role of the Baja airstream in the central region of the comma head in precipitation production. The generating cells that developed within the Baja air mass seeded the moist Gulf air mass with ice particles. The Gulf air mass subsequently provided moisture that allowed these ice particles to continue growth as they fell. The Baja air mass was thus a critical component of a seeder–feeder process that created ice particles in this storm. Plummer et al. (2014) present microphysical characteristics of ice particle growth within generating cells atop this and other PLOWS storms.

The Canadian air also played an important role in the distribution of precipitation. Because this air was dry, precipitation only reached the ground where the Canadian air was shallow (<1.5 km). On the north side of the comma head, where the Canadian air mass was deeper, precipitation falling from aloft into the Canadian air mass sublimated before reaching the surface. The cloud base in this region conformed roughly to the top of the Canadian air mass. Generating cells continued to be present within the Baja air mass atop this region of the comma head, although the precipitation produced by these cells never made it to the ground.

As noted by Schultz and Vaughan (2011) and summarized in many texts (e.g., Bluestein 1993; Martin 2006), fronts are characterized by near vertical isentropes, with very low static stability at the leading edge, and well-mixed post frontal air behind the front. The potential temperature analyses across the comma head of this storm showed that the airmass boundaries in this case were not fronts. Trajectory analysis showed that the Canadian air had subsided 1–2 km during the 48 h prior to arriving on the cross section, while Gulf air had ascended about 3 km from near the surface in this same period. These circulations, a likely response to frontogenesis during the storm’s 48-h evolution prior to the
arrival of the air on the cross sections, apparently homogenized the temperature gradient across the front due to adiabatic warming on the front’s cold side and cooling on the front’s warm side. By the time air arrived at its position in the comma head, from a dynamic perspective, a front no longer existed. Yet from an air mass, precipitation, wind, and moisture (i.e., equivalent potential temperature) perspective, the airmass boundary was still quite distinct.

The boundary between the Baja air mass and the Gulf air mass also showed little thermal contrast. Over the central United States, along the tail of many extratropical cyclones, the boundary between the dry and moist air masses is bounded by a cold front aloft (e.g., Hobbs et al. 1990), and is coincident with the leading edge of a cyclone’s dry slot. In other cases, the boundary has been characterized as a “humidity front” (e.g., Mass and Schultz 1993; Grim et al. 2007). In the comma head region of the present storm, the Baja air was drier than the Gulf air, suggesting a humidity front characterization, but both air masses were supersaturated with respect to ice, a result of the condition that both air masses had a history of ascent, as well as the cold temperatures in the vicinity of cloud top.

6. Summary

This paper presented a detailed analysis of structure of the comma head of a continental extratropical cyclone that occurred over the central United States on 30 January 2010. Data from the University of Wyoming Cloud radar, flown on board the NSF/NCAR C-130
aircraft, were combined with thermodynamic fields from the Weather Research and Forecasting Model and air parcel back trajectories calculated with the HYSPLIT model to examine finescale precipitation structures within the comma head and their relationship to fronts.

The key findings of the paper are as follows:

1) The comma head of this cyclone comprised three vertically stacked air masses: from bottom to top, an air mass of Canadian origin, a moist cloud-bearing air mass of Gulf of Mexico origin, and an air mass originating primarily at low altitudes over the Pacific just offshore of Mexico, the Baja Peninsula, and the Mexican Plateau.

2) The Baja air mass capped the entire comma head and significantly influenced the precipitation distribution within the comma head. On the comma head’s south side, the Baja air mass limited the cloud depth, reducing surface precipitation intensity and in some locations led to a change in precipitation type from light snow to freezing drizzle. In the central and northern part of the comma head, where the Gulf air mass was deeper, generating cells with depths of 1.5–3.0 km and peak vertical air velocities of 2–4 m s⁻¹ were ubiquitous within the Baja air mass atop the comma head clouds. In some locations, the generating cells extended from the top of the Gulf air mass well into the Baja air mass, while in others, the cells appeared to be entirely within the Baja air mass. The generating cells initiated a seeder–feeder process, whereby reflectivity streamers from the cells merged into the more uniform echo within the Gulf air mass.

3) Precipitation was prevented from reaching the surface on the northern side of the comma head by sublimation as ice fell into the dry surface-based Canadian air mass. Within the zone of sublimation, downdrafts as strong as 1–3 m s⁻¹ were observed.

4) The airmass boundaries in the comma head region of this storm did not have characteristics of fronts. Although the airmass boundaries were quite distinct, from a potential temperature perspective, no fronts were present at the time of observation across the comma head region of this storm.

This is one case study, and the applicability of the results to other cyclones is uncertain. We note, however, that additional evidence from other PLOWS investigations (Rauber et al. 2014; Rosenow et al. 2014), and past case studies derived from research aircraft measurements across the comma head of continental winter cyclones (Grim et al. 2007), together are providing an increasing body of evidence concerning the important influence of the drier air mass aloft in controlling the type and distribution of precipitation across the comma head of continental wintertime cyclones. Rauber et al. (2014), for example, found that in certain cyclones, dry air intruding over Gulf air on the southern side of the comma head possesses sufficient potential instability that elevated convective towers develop. These were shown to have vertical velocities in the range of 6–8 m s⁻¹, contain supercooled water and graupel, and occasionally produce lightning and so-called “thundersnow.” Rosenow et al. (2014) showed that generating cells, with vertical velocities similar to those reported here, were ubiquitous across the tops of the deeper stratiform clouds of the comma heads of several cyclones. Together these studies are providing evidence that features observed in the present case study are not unique to this storm, and may indeed be common across the comma head of continental winter cyclones.

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