

Convective Cells in Altocumulus Observed with a High-Resolution Radar

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ABSTRACT

Very-high-resolution Doppler radar observations are used together with aircraft measurements to document the dynamic and thermodynamic structure of a dissipating altocumulus cloud system associated with a deep virga layer. The cloud layer circulation is shown to consist of shallow vertical velocity couplets near cloud top and a series of subkilometer-scale Rayleigh–Bénard-like cells that extend vertically through the depth of the cloud layer. The subcloud layer was observed to contain a number of narrow virga fall streaks that developed below the more dominant Rayleigh–Bénard updraft circulations in the cloud layer. These features were discovered to be associated with kilometer-scale horizontally orientated rotor circulations that formed along the lateral flanks of the streaks collocated downdraft circulation. The Doppler analysis further reveals that a layer mean descent was present throughout both the cloud and subcloud layers. This characteristic of the circulation is analyzed with regard to the diabatic and radiative forcing on horizontal length scales ranging from the Rayleigh–Bénard circulations to the overall cloud layer width. In particular, linear analytical results indicate that a deep and broad mesoscale region of subsidence is quickly established in middle-level cloud layers of finite width when a layer-wide horizontal gradient in the cloud-top radiative cooling rate is present. A conceptual model summarizing the primary observed and inferred circulation features of the altocumulus layer is presented.

1. Introduction

Recent detailed observations of middle-level altocumuli have served as a basis for increasing the understanding of the importance of these cloud systems on planetary climate (Heymsfield et al. 1991; Houze 1994; Fleishauer et al. 2002; Larson et al. 2006). Well removed from the direct influence of Earth's surface, and typically located within colder low water vapor environments, these cloud layers often consist of a mixed-phase composition (Heymsfield et al. 1991) with liquid or ice

water content values generally less than a few tenths of a gram per cubic meter (Fleishauer et al. 2002). Altocumuli are often less than 300 m in thickness and tend to form in narrow bands less than 50 km in width (Ansmann et al. 2009). Despite this, altocumuli can produce deep virga layers upward of 2 km thick that are formed from evaporating or sublimating precipitation-sized drops or ice crystals that settle into subsaturated conditions below cloud base (Sassen 1991; Hobbs et al. 2001; Wang et al. 2004; Marsham et al. 2006; Carey et al. 2008; Noh et al. 2013). These thin cloud layers and their underlying virga thus remain radiatively active with peak heating rates ranging from 2 to -15 K h^{-1} for the lower and upper regions of the cloud, respectively (Heymsfield et al. 1991; Larson et al. 2007).

The strong radiative forcing associated with these cloud layers has served as a focus of several studies seeking to understand the underlying factors governing

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the turbulent nature of the observed altocumulus roll structure or other identifying cloud properties. In specific case studies conducted by [Ansmann et al. \(2009\)](#), it was determined that the vertical velocity values within individual altocumulus cells or roll circulations were generally limited to $\pm 2 \text{ m s}^{-1}$. The horizontal scale of altocumulus circulations were determined in an early study by [Suring \(1950\)](#) to be $<0.25 \text{ km}$ in 39% of cases, $<0.5 \text{ km}$ in 78% of cases, and $<0.75 \text{ km}$ in 93% of cases. Limited numerical simulations ([Starr and Cox 1985](#); [Liu and Krueger 1998](#)) indicate an aspect ratio of these circulations of approximately 1:1. More recent studies of altocumuli performed in conjunction with the First International Satellite Cloud Climatology Project (ISCCP) Regional Experiment (FIRE) by [Heymsfield et al. \(1990, 1991\)](#) further documented altocumulus cloud structure and provided initial assessments of the relative importance of radiative forcing, entrainment, and cloud microphysical structure on the overall dynamics of the individual cells that were observed to have a horizontal scale on the order of 250 m . Data obtained from the Complex Layered Cloud Experiment (CLEX) as documented by [Fleishauer et al. \(2002\)](#), [Smith et al. \(2009\)](#), and [Larson et al. \(2006, 2007\)](#) also provided great insight on the vertical structure of these mixed-phase cloud layers and motivated additional studies on the governing middle-tropospheric processes, such as larger-scale subsidence and the diffusional growth of ice particles, which were noted to influence the cloud lifetime.

In this study, documentation of a dissipating altocumulus cloud that produced a deep virga layer is provided through a combination of in situ aircraft observations and high-resolution ground-based Doppler radar measurements. In light of the previous work of [Heymsfield et al. \(1991\)](#), [Larson et al. \(2006\)](#), and [Durrán et al. \(2009\)](#), we examine various attributes of the observed circulation within the cloud system in regard to the impact of radiative forcing on length scales associated with the subkilometer-scale cloud-layer eddies and the overall lateral width of the cloud band. A description of the instrumentation and methodology used to develop the analyses is provided in [section 2](#). An observational overview of the cloud and subcloud layers are presented in [section 3](#) and a discussion of the results can be found in [section 4](#). A schematic model depicting the main attributes of the cloud circulation and reflectivity structure is presented in the summary ([section 5](#)).

2. Instrumentation and methodology

The primary observational tool used in this study was the high-resolution U.S. Navy's Doppler Mid-Course Radar (MCR) located near Titusville, Florida. The MCR is

a 3-MW peak-power C-band dual-polarization radar that alternatively transmits two linear frequency modulated wave forms with a 0.22° beamwidth at a pulse repetition frequency (PRF) of 160.1 Hz . The two wave forms provide a choice in the 6-dB width range resolution (R_6) of either 37 m (referred to as the narrowband waveform) or 0.546 m (denoted as the wideband waveform) and were recorded at an oversampled range-gate spacing of 11.25 and 0.1464 m , respectively. The narrowband and wideband waveforms were transmitted with slightly different wavelengths (λ) of 0.0545 m and 0.05306 m , respectively, and each consisted of two independent range windows either 15 km (narrowband) or 120 m (wideband) in length. The two wideband range windows could be positioned contiguously or at independent locations within the cloud layer and, in practice, were often set to slightly overlap each other in order to examine a specific portion of the cloud layer in great detail. The narrowband was then used to help place the detailed wideband observations in context of the surrounding cloud field. The cloud layers examined in this study were observed at altitudes of $4\text{--}7 \text{ km}$ above ground level and at all times the two wideband windows were placed contiguously in order to document the structure near cloud top. At these range limits, the two-way 6-dB cross-width varies from 15.36 to 26.88 m and this leads to corresponding changes in the pulse volumes from 55 to 168 m^3 for the wideband and 3722 to 11400 m^3 for the narrowband waveforms, respectively. These pulse volume values differ from what would be obtained from the more widely quoted radar pulse volume form, such as that of [Probert-Jones \(1962\)](#), owing to the assumption that MCR pulse shape in the along-range direction is considered to have a Gaussian, rather than rectangular, shape (see [appendix A](#) for details). The combination of the MCR wavelength and PRF lead to relatively small values of the Nyquist interval ($\lambda \times \text{PRF}/4$) of $\pm 2.2 \text{ m s}^{-1}$ for both waveforms but this was not found to produce velocity folding issues owing to the fact that we were using vertically pointed radar scans to examine the weak vertical velocity component associated with a thin altocumulus cloud layer.

In an effort to ensure the highest possible data quality, the radar was carefully calibrated each day using the orbiting calibration sphere 5398 LCS4. The MCR data were also subjectively edited to remove radar artifacts and then further processed in order to merge the wideband and narrowband data onto a common time(abscissa)–height(ordinate) grid used for the display and construction of various derived quantities. For the time–height plots shown here, each column along the abscissa represents a temporal average of 128 raw radar pulses or approximately 0.8 s of elapsed time. Values along the ordinate of the time–height grid represent the vertical range from the

radar using either the native range-gate spacing of each waveform or a special 1-m range increment that was used to merge two or more range windows from one or more wave forms onto a common analysis grid. In the latter case, the wideband range windows were mapped to the 1-m vertical grid increments using an average of the surrounding ± 5 range gates within the column (spaced 0.14643 m apart) while the narrowband estimates at a particular range gate were obtained through linear interpolation from the nearest two surrounding range gates. In regions where two range windows overlapped, a linear blend of the data from each range window was used. Any averaging, interpolation, or filtering involving the logarithmic radar reflectivity factor $Z(\text{dBZ}) = 10 \log_{10}[z(1 \text{ mm}^6 \text{ m}^{-3})^{-1}]$ was performed in terms of the linear radar reflectivity factor (z). Hereafter, we refer to $Z(\text{dBZ})$ as the reflectivity. A simple 1–2–1 filter was applied to z and various derived fields in an effort to reduce any residual noise in the merged data set.

The methodology of [Doviak and Zrnić \(1984, 89–97\)](#) was used to construct the Doppler analysis from a FFT power spectra formed from the complex signal voltage using the in-phase (I) and quadrature (Q) signal components derived from the edited amplitude and phase data. In this application, the FFT was also constructed from 128 consecutive raw radar pulses taken at a fixed range. The assessment of the velocity contribution of the droplets V_T to the net measured Doppler velocity (which is a sum of the true vertical velocity of the air w and V_T ; $W_{\text{net}} = w + V_T$) is based on the reflectivity weighted terminal velocity methodology outlined by [Baker and Hodson \(1985\)](#), [Frisch et al. \(1995\)](#), [O'Connor et al. \(2005\)](#), and [Marshall et al. \(2006\)](#). Because of the presence of precipitation-sized particles in the cloud layer, we follow [Frisch et al. \(1995\)](#) and use a lognormal size distribution [Eq. (B2)] where the logarithmic width of the distribution σ is set to a value of 0.35 (see [appendix B](#) for details). This setting was found through experimentation and comparisons with other relationships in the literature ([Atlas 1954](#); [Fox and Illingworth 1997](#); [Baedi et al. 2000](#); [Khain et al. 2008](#)) to lead to reasonable fits with available aircraft measurements of the liquid water content (LWC) and mean drop size. The expression for the reflectivity weighted velocity [Eq. (B1)] is solved numerically using the velocity–diameter relationships provided by [Pruppacher and Klett \(1978, 322–324\)](#) for drop radii falling within their regime I ($0.1 < a < 10 \mu\text{m}$) and regime II ($10 < a < 535 \mu\text{m}$) size categories. As indicated in [appendix B](#), the expression for the reflectivity weighted velocity is viewed to have the greatest accuracy in the cloud layer where observed concentrations were relatively high and varied by less than an order of magnitude. The resulting expression for the Z –LWC

relationship used in this study is given by Eq. (B5) in [appendix B](#). Other derived fields, such as the power spectra (shown in [Fig. 5](#)) and the layer mean fields (shown in [Fig. 6](#) and [Fig. 11](#)) were constructed through a normalization procedure whereby data columns were first vertically realigned so that the analyzed position of the cloud base was assigned the same vertical level in the time–height array used to construct the given derived field. The data lying above cloud base in a given column were then normalized with respect to the maximum cloud layer depth obtained during the course of the roughly 90-min measurement period.

Additional instrumentation used in the experiment included the Sigma Space Micro Pulse lidar, the METEK Micro Rain Radar ([Peters et al. 2005](#)), an in-house developed all-sky camera, the Meteolabor Snow White rawinsonde ([Verver et al. 2006](#)), and a Cessna Cheyenne research aircraft operated by Weather Modification, Inc., that was equipped with a forward scattering spectrometer probe (FSSP) ([Baumgardner 1983](#)), a two-dimensional optical array imaging probe (2D-C) ([Knollenberg 1981](#); [Korolev et al. 2011](#)), a Commonwealth Scientific and Industrial Research Organisation (CSIRO) King liquid water probe for measuring the LWC ([King et al. 1978](#)), as well as additional sensors for measuring the state variables of temperature (Rosemount 102 deiced series), dewpoint temperature, pressure, and the GPS location and altitude ([Schmidt et al. 2012](#)). A video camera was also placed on the aircraft and was found useful in corroborating rapid fluctuations evident in the measured aircraft parameters during its flight into and out of the individual turrets within the cloud layer. The aircraft executed a series of short circular loops over a 30-min period that sampled the cloud in the direct vicinity of the MCR approximately every 5 min. The aircraft measurements were supplemented with special environmental soundings derived from the Snow White rawinsondes released at regular time intervals from a location just south (within 20 km) of the MCR site (at the Cape Canaveral Air Force weather station). Particular use is made of the sounding at 1730 UTC 27 August, which was released approximately 1.5 h prior to the start of the radar analysis. Among other purposes, this sounding was used to identify a positive bias in the reported aircraft temperature field that was subsequently corrected with the methodology outlined by [Inverarity \(2000\)](#) and by accounting for the self-heating and deicing heating error terms associated with the Rosemount 102 deiced temperature probe in subsonic flight conditions ([Stickney et al. 1994](#), p. 22).

The radiative heating rates shown in the text were derived from the [Fu and Liou \(1992\)](#) radiation scheme using environmental sounding data obtained from a merged

analysis of the 1730 UTC Snow White sounding, the aircraft ascent sounding taken from the surface to the top of the altocumulus cloud layer near 6950 m, and the radar-derived liquid water content using Eq. (B5). The sounding was interpolated to a 1D grid column that was 21 km deep using a vertical grid increment of 100 m. This resolution was modified in the region of the analyzed cloud and subcloud layers (taken to lie between 4000 and 7500 m) to a value of 11.25 m in an effort to create greater consistency with the vertical resolution associated with the merged MCR analysis grids. The vapor field in the sounding was modified within the analyzed borders of the cloud layer to produce a saturated state determined by the given sounding temperature and pressure at those levels. The upper cloud boundary was determined objectively by computing the first point from the top of the analysis domain where the reflectivity exceeded the background minimum value of -50 dBZ. Determining cloud base was more subjective but a value of -40 dBZ was found to closely correspond to a strong reflectivity gradient that ran along nearly the entire base of the layer and to the level where aircraft measurements also indicated cloud base to occur. In regions where narrow bands of higher reflectivity values were found to extend into the subcloud layer, cloud base was subjectively set to a level determined by the height of the -40 -dBZ isopleth residing on either side of the bands. The radiative calculations assume a solar hour of 1900 UTC at the longitude of the MCR (88.8°W). The effective radius was determined as the ratio of the third to second moment of the size distribution using the parameterization of Bower and Choularton (1992) and a droplet concentration of 25 cm^{-3} . The surface emissivity was set to 0.96 and the surface temperature was set to that obtained from the lowest sounding level (301.8 K). Finally, details of the drop and crystal evaporation/sublimation models used to calculate the vertical particle displacements (shown in Fig. 7 and Fig. 10a) can be found in appendix B.

3. Results

a. Overview

The altocumulus cloud layer examined in this study was observed by the experimental instrumentation suite between 1830 and 2030 UTC 27 August 2010. The all-sky camera images shown in Fig. 1 reveal a gradual change in the sky conditions from an optically thin and fibrous altostratus first present over the radar site near 1830 UTC to an optically thicker and more cellular altocumulus structure that eventually obscured the entire sun from view some 90 min later. Radar imagery shows that this transition in sky conditions was associated with

two separate altocumulus cloud layers that were separated in elevation by nearly 1500 m (Fig. 2). The upper cloud layer has the more complex structure of the two and it is this layer that will serve as the focus of study for the remainder of this paper. As indicated by the bold white stippling in Fig. 2, the upper cloud system was composed of a 200–400 m thick cloud layer centered near an elevation of 6.7 km and an extensive layer of enhanced reflectivity that extended 500–1500 m below the analyzed cloud base. Several narrow and vertically oriented bands of enhanced reflectivity are evident within the subcloud region and are similar in structure to the virga streamers noted to occur in other altocumulus cases (Sassen 1991; Wang et al. 2004). As noted in both altocumulus and precipitating boundary layer stratus (Pincus and Baker 1994; Feingold et al. 1996; Ansmann et al. 2009; Frisch et al. 1995), the strongest reflectivity in the subcloud layer in many instances appeared to be well correlated with the overall cloud thickness and the presence of individual cells of enhanced reflectivity located within the cloud layer. The deepest penetration of particles into the subcloud layer observed near 1910 UTC occurred during a general upward trend in the overall height of the layer at a time when the in-cloud reflectivity was also reaching a peak value. The gradual oscillation evident in the height of the cloud layer exhibited a period of approximately 40 min and was superimposed with a series of higher frequency fluctuations in the reflectivity values associated with the translation of individual cells across the vertically pointed radar beam. The movement of these cells was determined to be on the order of $4\text{--}6\text{ m s}^{-1}$ from a southwesterly direction based on the environmental flow conditions at this level obtained from the 1730 UTC sounding (Fig. 3a).

The radar-suggested presence of hydrometeors in the subcloud layer between 5.0 and 6.5 km is supported by measurements taken by the aircraft as it ascended toward cloud base in subzero temperatures between 1940 and 1944 UTC (Fig. 3b). As indicated by the bold colored circles in Fig. 2, the aircraft first recorded an increase in the LWC between an elevation of 5 and 5.8 km. The FSSP probe at this time indicated that this LWC was associated with a low concentration ($<0.02\text{ cm}^{-3}$) of supercooled liquid water drops less than $5\text{ }\mu\text{m}$ in diameter. The range of relative humidity values measured by aircraft over this layer reveals that these drops would have experienced subsaturated conditions along their entire descent from cloud base (Fig. 3c), indicating they originated as much larger precipitation-sized drops in the cloud layer aloft (Li and Srivastava 2001). Notably, there was no visual sign of any cloud particles from the onboard video camera until the visibility became completely obscured as the aircraft penetrated the visually distinct cloud base near an elevation of 6.68 km. It was

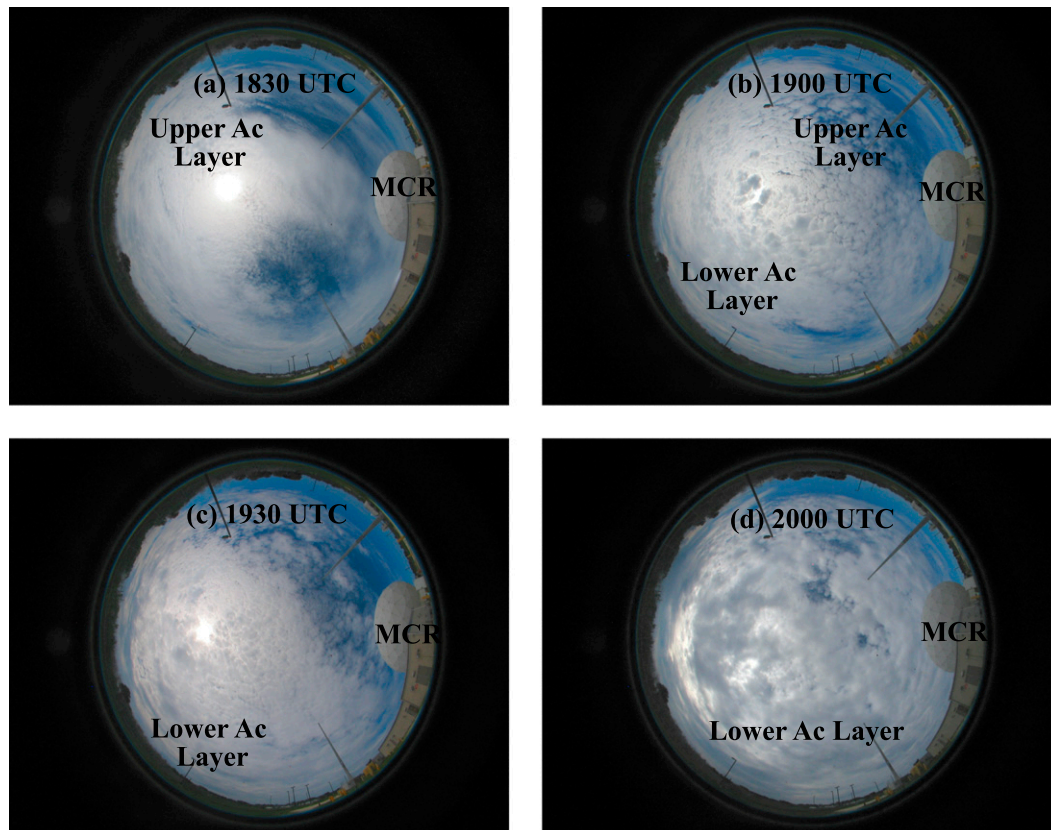


FIG. 1. All-sky camera images at 30-min intervals between 1830 and 2000 UTC 27 Aug 2010 showing the cellular nature of the altocumulus layer: (a) 1830, (b) 1900, (c) 1930, and (d) 2000 UTC. The MCR radar is labeled on the eastern (right) portion of each image. The images reveal a change in the cloud optical thickness near 1930 UTC as a lower altocumulus layer advected from the west (left) at an altitude of 5 km and undercut a thin 300-m-thick upper cloud layer near 6.7 km. A movie of this case based on the whole-sky camera images is available at <http://www.youtube.com/watch?v=Bymq1lb40nY>.

just prior to this time that the FSSP and the King probes first began to record noticeable increases in the droplet number concentration and LWC, suggesting the aircraft had a brief encounter with an isolated pocket of hydrometeors within 100 m of the observed cloud base (Figs. 4a–c).

The higher amplitude fluctuations in the microphysical measurements made above cloud base were found to occur at times when the onboard video revealed partial or nearly complete clearing above the altitude of the aircraft. Their presence is thus linked to the passage of the aircraft through a series of individual turrets within the cloud layer. The structure of a particularly large amplitude fluctuation in the thermodynamic variables measured just below the upper temperature inversion (denoted by the arrow labeled S in Fig. 3b) suggests that the aircraft encountered a narrow downdraft circulation that possibly originated 20–50 m above the observed cloud top (Figs. 3b,d,e and Fig. 4c). Use of the known temporal spacing of the gap in the LWC across this feature, together

with knowledge of the aircraft's true airspeed of 98 m s^{-1} , leads to estimates of the downdraft size on the order of nearly 500 m (Fig. 4c). Using this same scaling, other eddies encountered by aircraft are judged to range in width from a few hundred meters near cloud top to approximately 1200 m within the central and lower portions of the cloud layer (Fig. 4c). Overall, the measured LWC was consistent with that found in other altocumulus studies with peak values on the order of 0.2 g m^{-3} obtained within approximately 50 m of the cloud top (Fig. 4c). The FSSP drop concentration within each of the primary eddies intercepted by the aircraft consistently reached maximum values of 20 to 28 cm^{-3} while the peak drop diameters were on the order of $35 \mu\text{m}$ (Figs. 4a,b). The largest values for the mode of the size distribution were consistently less than $20 \mu\text{m}$ (Fig. 4a).

Images from the 2D-C probe were available for a total of 34 s of flight time. These measurements occurred over a series of 0.1–6-s intervals as the aircraft periodically

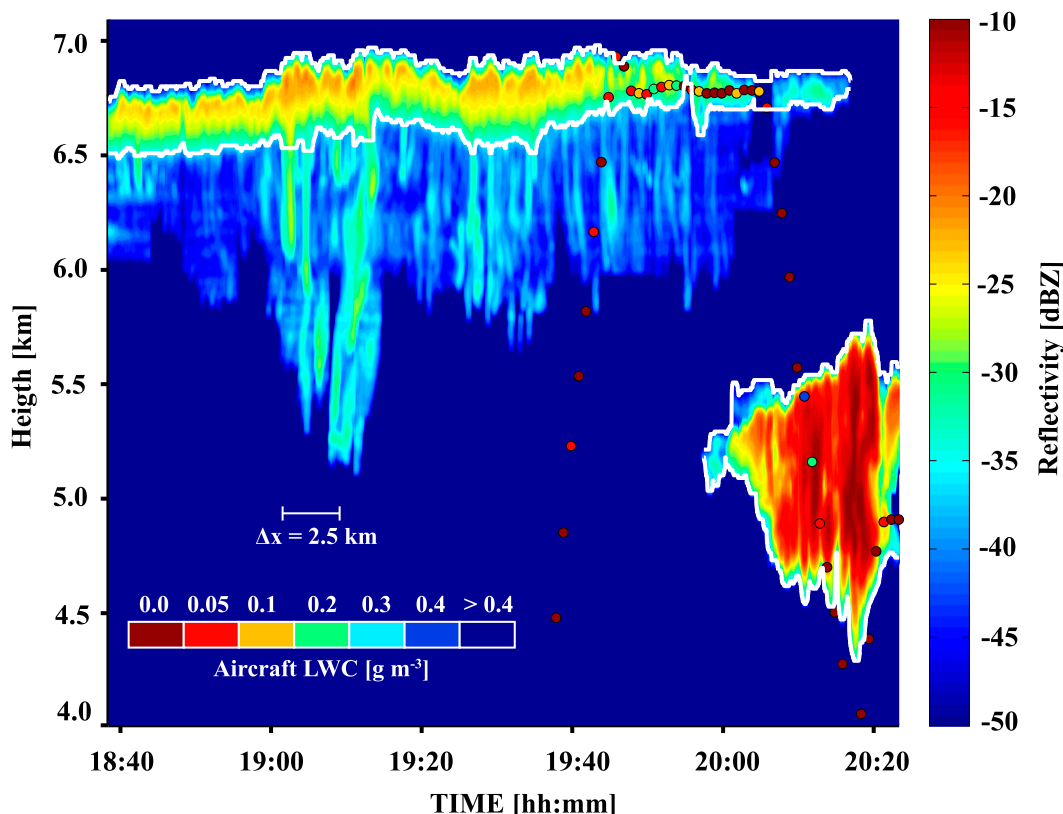


FIG. 2. Time–height plot of MCR narrowband radar reflectivity (dBZ) from 1840 to 2020 UTC 27 Aug 2010. Reflectivity is color shaded and the bold white lines denote the derived cloud boundaries. Colored dots represent aircraft-measured liquid water content (g m^{-3}) at 1-min intervals. The value of the liquid water content is given by the color table in the lower-left portion of the plot. The displayed horizontal scale was derived assuming that the upper cloud layer was moving at 5.5 m s^{-1} . The aircraft was performing a circular flight pattern that passed over the MCR approximately every 5 min and thus the plotted dots do not lie in the time–height plane of the radar observations at all times as depicted.

sampled larger particles over the middle-to-lower half of the dissipating cloud layer between 1945 and 2005 UTC (see flight track in Fig. 2). These images were analyzed based on the methodology of Korolev (2007) in order to determine the size and phase of the imaged particles. Of the total of 305 particles sampled, nearly 35% were classified to be of indeterminate type or size because only a portion of the particle was imaged by the sensor. Another 5% were identified as platelike ice crystals likely viewed on edge because of their low width-to-length aspect ratios apparent in the imagery. The remaining particles had an aspect ratio closer to 1.0 and were determined to be either spherical liquid water drops (60 particles) ranging in radius from 25 to $300 \mu\text{m}$ (solid rectangles in Fig. 4d) or a combination of spherical or nearly circular particles (126 total) of unknown phase (plates, frozen drops, or liquid drops) that exhibited a slightly greater range in size (dashed rectangles in Fig. 4d). While the total number of particles analyzed 2D-C images is relatively small, they nevertheless provide

a hint that the upper cloud layer was of mixed-phase origin and that a precipitation mode was present at a number of locations within the cloud layer.

b. General circulation features and scale analysis

The overview provided thus far has revealed the cellular and mixed-phase microphysical nature of the thin upper altocumulus cloud layer that generated a deep layer of enhanced reflectivity within the subsaturated air mass residing below the aircraft measured cloud base. Additional aspects of the general cloud and subcloud circulation structures are examined in this section using results derived from the analyses of the condensate-corrected Doppler-derived vertical velocity field (referred to as simply the vertical velocity; see appendix B for details on the velocity correction method used in this study). We begin with a more quantitative overview of the cloud and subcloud eddy structure revealed from the power spectra analysis of the perturbation vertical velocity field shown in Fig. 5. This plot was based on the

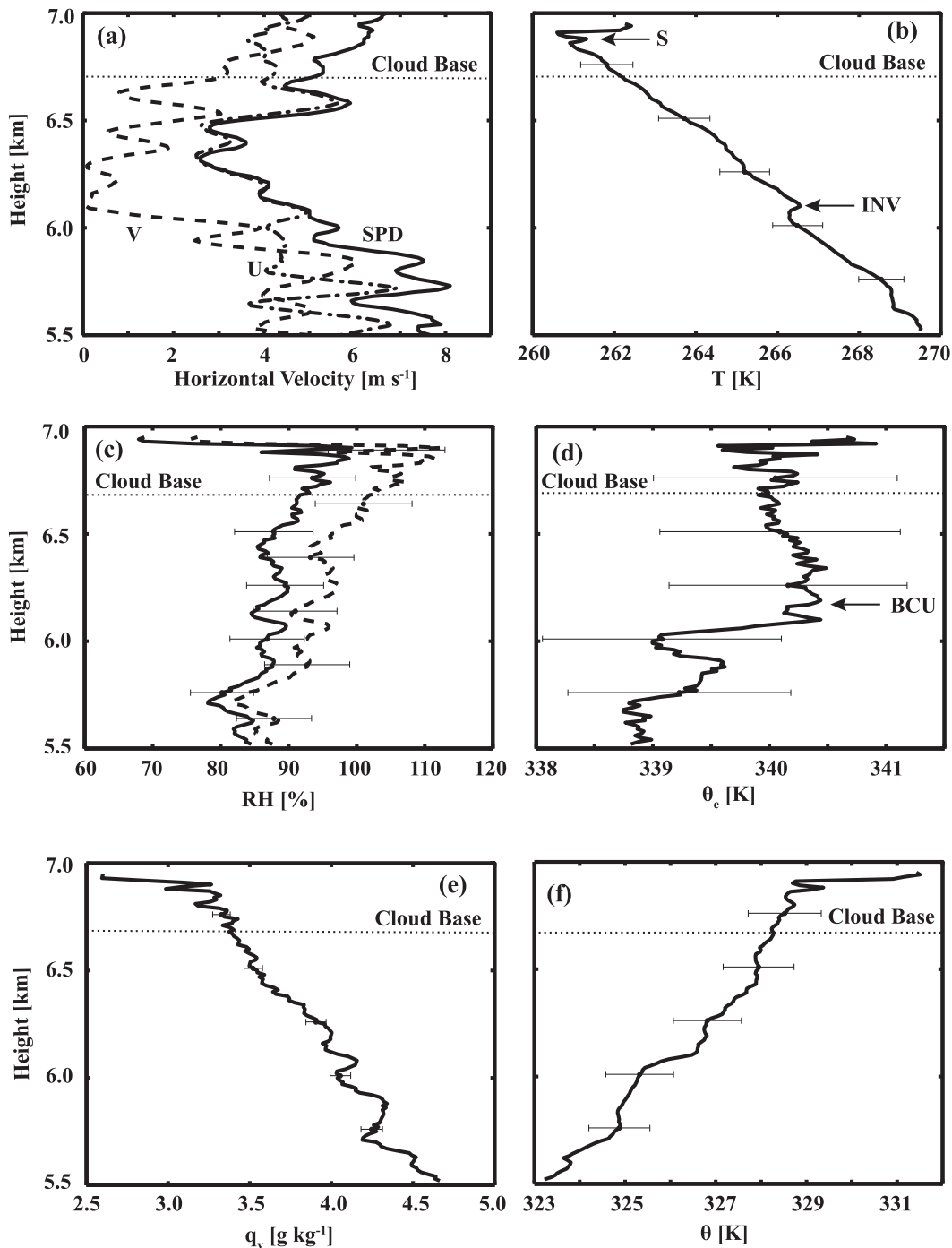


FIG. 3. Profiles from 27 Aug 2010 derived from (a) the Snow White rawinsonde released at 1730 UTC and (b)–(f) aircraft observations taken between 1940 and 1946 UTC. The panels show (a) total wind speed (solid), u (dotted–dashed) and v component of the flow (dashed), and the total wind speed (m s^{-1}); (b) temperature (K); (c) derived relative humidity (%) with respect to ice (dashed) and liquid water (solid); (d) the derived equivalent potential temperature (K) using Bolton (1980); (e) water vapor mixing ratio (g kg^{-1}); and (f) the derived potential temperature (K). The horizontal dashed line at 6.68 km denotes the cloud base determined from the aircraft video camera. The arrows in (b) labeled S and INV denote the downdraft encountered near cloud top and temperature inversion discussed in the text. The arrow in (d) at 6.2 km labeled BCU denotes the base of the conditionally unstable layer.

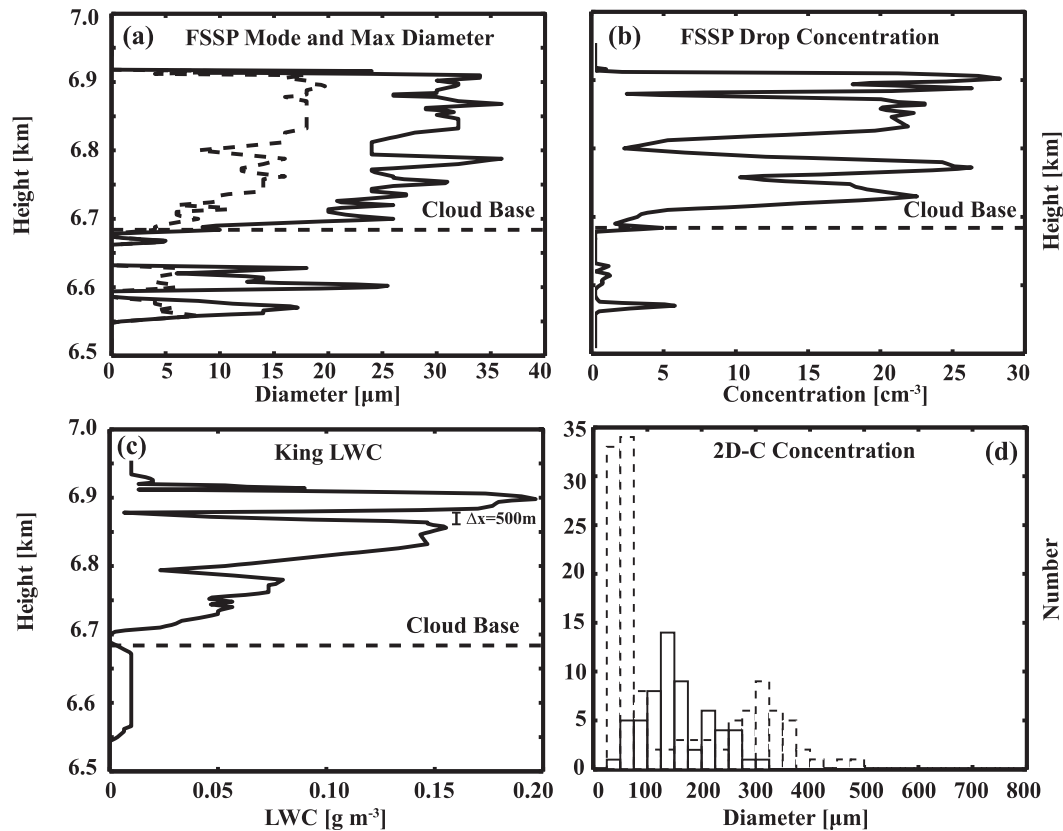


FIG. 4. Aircraft-measured microphysical profiles from 27 Aug 2010 taken of the upper cloud layer between 1943 and 2005 UTC. The panels show (a) FSSP maximum drop diameter (solid) and the mode of the size distribution (dashed) (μm), (b) FSSP drop concentration (cm^{-3}), (c) King liquid water content (g m^{-3}), and (d) the 2D-C particle count for each of the thirty-two $25\text{-}\mu\text{m}$ -size bins obtained from all available 2D-C images recorded during a total of 34 s of flight time. The particles in (d) were classified as either spherical liquid drops (solid rectangles) or a mixture of spherical drops, frozen drops, or nearly circular ice crystals (dashed rectangles). The bold dashed horizontal line in (a)–(c) denotes the cloud-base level from the aircraft video camera. The short vertical line in (c) near 6.88 km labeled $\Delta x = 500\text{ m}$ depicts the horizontal scale of the eddies determined from the aircraft speed. The aircraft observations were collected at 1-Hz intervals and interpolated to a vertical grid consisting of a uniform grid spacing of 2 m .

cloud-base normalization procedure discussed in [section 2](#) and we further follow the methodology outlined by [Kollias and Albrecht \(2000\)](#) to perform a conversion from frequency space to a given length scale using a mean cloud layer motion of 5.5 m s^{-1} determined from the environmental flow shown in [Fig. 3a](#).

The general trend evident in the power spectra is a shift in power from shorter to longer length scales as the vertical displacement from cloud top increases. This trend may reflect actual changes in the length scale of specific circulation features arising within either the cloud or subcloud layers or from changes in the space–time conversion factor as the condensate settles in an environment exhibiting vertical wind shear ([Fig. 3a](#)). While several of the spectral peaks appear largely confined to the cloud layer itself, others exhibit a much deeper

vertical coherency and extend well into the subcloud layer (see the spectral peaks denoted by bold asterisks in [Fig. 5](#)). Such structure is suggestive of a strong linkage in the circulation features between the two layers at specific length scales. Several secondary peaks in the subcloud layer occur at length scales within the $600\text{--}1500\text{-m}$ portion of the spectrum in the 300-m -deep layer that resides just below cloud base. This portion of the subcloud layer was previously shown to exhibit a weak conditionally unstable lapse rate ([Fig. 3d](#)) and also contains the bulk of the virga layer that extends along the entire base of the cloud layer ([Fig. 2](#)). The origin of these spectral peaks are discussed in terms of the subcloud rotor circulations observed in association with individual virga shafts that are presented in greater detail in [section 3d](#).

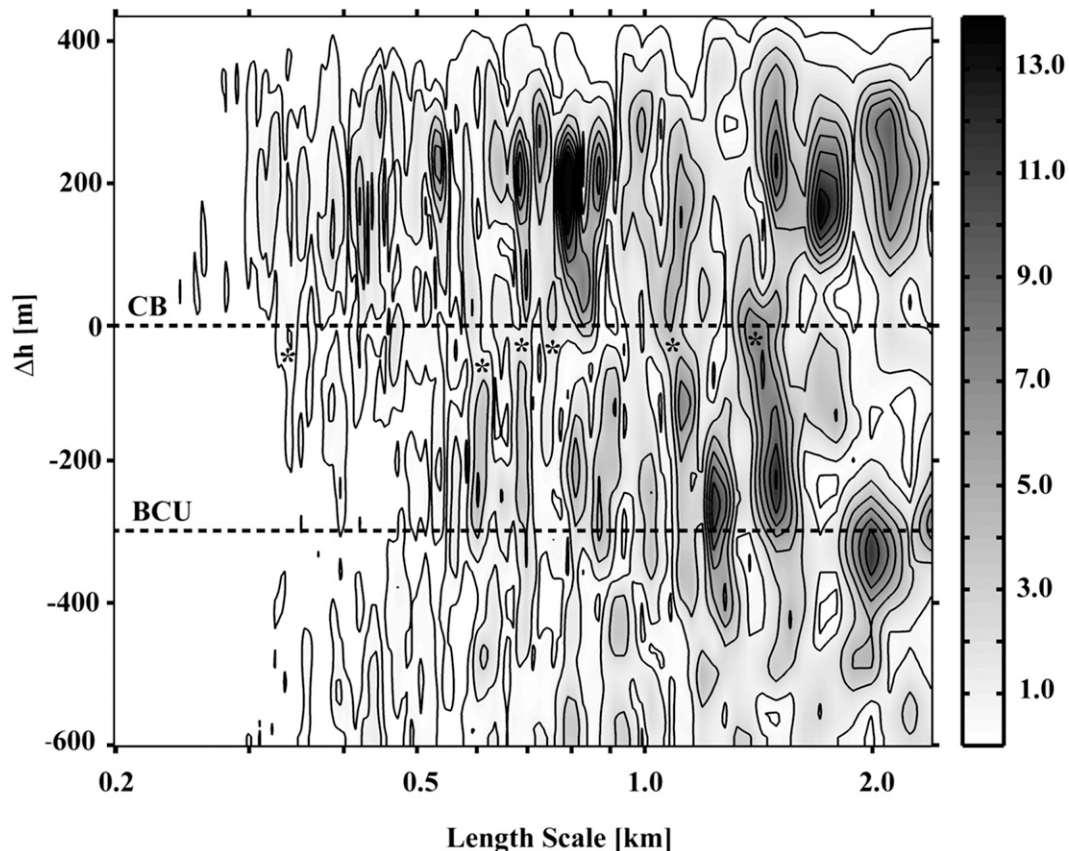


FIG. 5. Doppler-derived power spectra (m s^{-2}) of the condensate-corrected vertical velocity for the upper cloud layer using the MCR narrowband waveform measurements taken between 1840 and 2005 UTC 27 Aug. Selected contours are shown at intervals of 0.25, 1.0, 3.0, 7.0, 9.0, 11.0, and 13.0 m s^{-2} . The ordinate represents the displacement in meters from the normalized cloud-base position (assigned a value $\Delta h = 0.0 \text{ m}$). The horizontal dashed line labeled BCU denotes the base of the conditionally unstable subcloud layer shown previously in Fig. 3d relative to the position of cloud base (dash line labeled CB). The six bold asterisks just below the CB line denote regions where deep vertically coherent peaks exist between the two layers.

Additional properties of the cloud and subcloud vertical velocity structure are revealed through the layer mean fields presented in Fig. 6. As in Fig. 5, these plots were constructed using the cloud-base normalization procedure discussed in section 2. The primary feature in these plots is the dominance of the negative component of the vertical velocity that arises in both layers. This trend becomes most evident over the lower third of the cloud layer and extends through the depth of the subcloud layer where negative velocity values reach their peak magnitude and fractional coverage (Figs. 6a,b). The mean subcloud vertical velocity reaches a local maximum of -0.45 m s^{-1} near the level denoting the base of the conditionally unstable layer (dashed line labeled BCU near $\Delta h = -300 \text{ m}$ in Fig. 6a). This level also serves as a marker where the positive component of the subcloud vertical velocity exhibits a gradual increase in magnitude as well. With an exception of the local

minimum observed with 100 m of cloud base, the upward trend in the positive vertical velocity component continues to a level 200 m above cloud base before declining again near cloud top. The positive and negative components of the vertical velocity have a nearly equal magnitude and fractional coverage in the center portion of the cloud layer, providing an indication that there is a dominant symmetric component of the circulation within this region of the cloud (Figs. 6a,b). The negative component of the circulation is slightly greater in magnitude and has a higher fractional coverage throughout most of the cloud layer leading to a net velocity that is also less than zero.

The dominance of the downward component of the circulation is further indicated by the layer mean skewness [$S_w = \overline{w}^3 / (\overline{w}^2)^{1.5}$] shown in Fig. 6c. The vertical profile of the skewness exhibits a sharp minimum near cloud top and remains negative throughout the cloud and

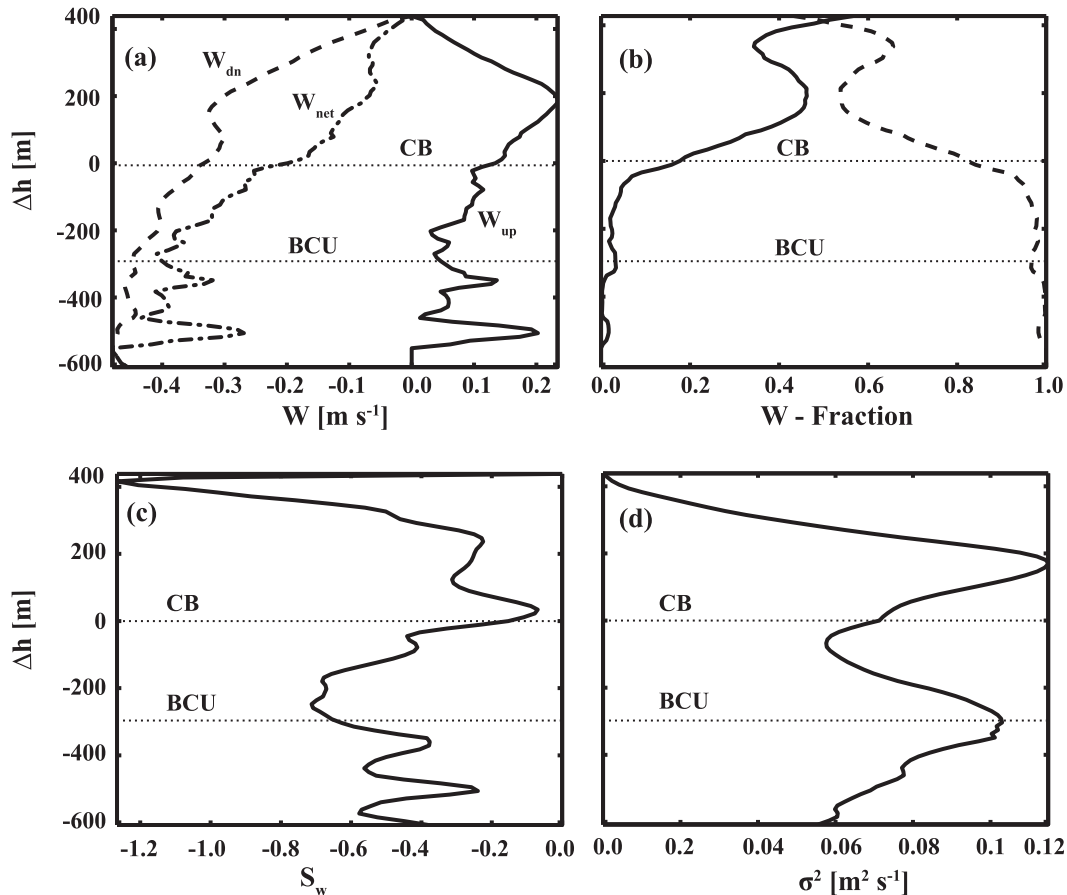


FIG. 6. Layer mean averages of the MCR narrowband data for a 1000-m-deep portion of the cloud and subcloud layers: (a) the layer mean Doppler-derived condensate adjusted vertical velocity (m s^{-1}) showing the positive (solid), negative (dashed), and net (dotted-dashed) components of the circulation derived at each range gate; (b) the fractional coverage of positive (solid) and negative (dashed) velocity components computed from the total number of each component obtained at a given range gate; (c) the vertical velocity skewness [$S_w = \overline{w^3}/(\overline{w^2})^{1.5}$]; and (d) the vertical velocity variance ($\text{m}^2 \text{s}^{-2}$). The ordinate and labels CB and BCU are as defined in Fig. 5.

subcloud layers. Numerical results obtained by Moeng and Rotunno (1990) suggest that negative skewness is indicative of a circulation dominated by strong cooling in the upper portion of the layer and little, if any, forcing near the base of the layer. The structure of the variance shown in Fig. 6d is considerably less complex than that of the skewness and exhibits a bimodal signature that is quite similar to that obtained by aircraft in the precipitating stratus cases examined by Nicholls (1984). The peak values occur within the center of the cloud layer and near the base of the conditionally unstable environment measured by the aircraft where the mean negative velocity profile is seen to reach a local minimum (Fig. 6a). The pronounced minimum in the variance field near cloud base indicates some weakness in the coupling of the circulation between the two layers near this particular level. These aspects of the mean fields will be addressed further in the following subsections as we begin to examine the time–height

structure of circulation fields that arise in the cloud and subcloud layers.

c. Cloud layer entrainment and Rayleigh–Bénard cells

Factors contributing to the derived variability in the cloud and subcloud spectral characteristics and layer mean vertical velocity structure are examined in the next three subsections using high-temporal-resolution time–height plots obtained from the merged analyses of the narrowband and wideband vertical velocity, LWC, and associated radiative heating rates. We begin with the 5-min analysis shown in Fig. 7 as it captures many elements of the circulation features discussed thus far. The cloud layer is seen to be undergoing a series of fluctuations in the overall thickness as both the cloud-top and cloud-base change in association with a series of well-defined velocity perturbations evident throughout

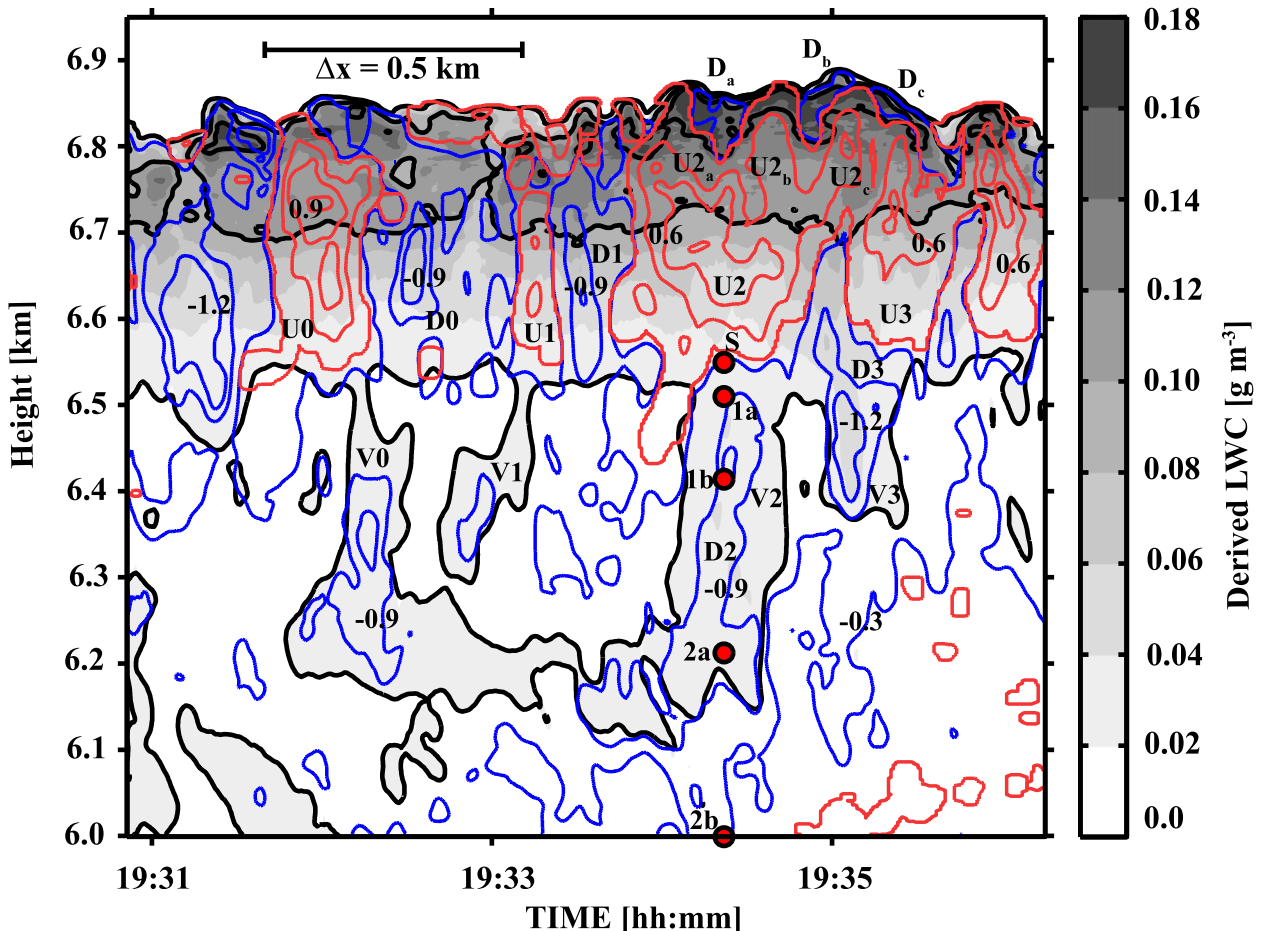


FIG. 7. A time–height analysis of the merged MCR narrowband and wideband data between 1931 and 1936 UTC 27 Aug 2010. Radar-derived liquid water content is shaded at intervals of 0.02 g m^{-3} . Positive (negative) values of the Doppler-derived condensate corrected velocity are shown at 0.3 m s^{-1} intervals by the solid red (blue) contours. The bold black contours depict the derived liquid water content at values of 0.02, 0.1, and 0.14 g m^{-3} , respectively. The circulation features labeled U0–U3, U2_a, U2_b, U2_c, D0–D3, D_a, D_b, and D_c are discussed in the text. The labeling V0–V3 depicts the virga shafts discussed in the text. The bold red dots depict the fall distance from a release point near $z = 6550 \text{ m}$ (labeled S) for drops initialized with radii of either $50 \mu\text{m}$ (1a and 1b) or $100 \mu\text{m}$ (2a and 2b) with the “a” and “b” denoting whether the drops fell in a quiescent environment at their terminal velocity or within a specified mean downdraft of -0.5 m s^{-1} , respectively. The displayed analysis domain has a fixed $\Delta h = 1 \text{ m}$. The displayed length scale is computed as in Fig. 2.

the layer. Similar fluctuations are evident in the LWC as well, particularly near cloud top where a number of local maxima reside. Peak-derived liquid water content values within these maxima reach 0.2 g m^{-3} in good agreement with those measured by aircraft as it ascended through the cloud layer only a few minutes later near 2044 UTC (Fig. 4c), lending credence to the choice and formulation of the Z–LWC conversions provided in appendix B.

Overall, there are three possibly interrelated circulation features of varying depth, horizontal scale, and location within the cloud layer that can be identified in Fig. 7. These include the deeper alternating zones of the positive and negative velocity components indicative of Rayleigh–Bénard circulation cells (Rayleigh 1916) that extend through the entire cloud layer with great regularity (such

as the velocity couplet labeled D0 and U0), the lower-level centers of negative velocity (such as D2 and D3) that tend to coincide with the narrow vertically oriented bands of higher LWC values in the subcloud layer (labeled V0 through V3), and the shallow vertical velocity couplets located near cloud top (such as those labeled U2_a through U2_c and D_a through D_c). The deeper Rayleigh–Bénard circulation cells (referred to as RB cells) are a distinctive feature of the cloud layer both because of their repetitive nature and vertical coherency. These circulation features contribute to the series of spectral peaks representing length scales generally in excess of 200 m (Fig. 5) and help contribute to the overall symmetry of the mean velocity field evident within the middle-to-upper portion of the cloud layer

shown in Fig. 6a. The magnitude of the vertical velocity associated with these cells lies well within $\pm 2.1 \text{ m s}^{-1}$ Nyquist interval of the radar and is found to be in good overall agreement with values obtained for other altocumulus layers (Ansmann et al. 2009).

The presence of the RB cells is interesting as the circulations are forming in the free atmosphere without the benefit of fixed rigid upper or lower plates and in the presence of other forcing mechanisms that can alter or otherwise modify their structure. It is evident from Fig. 7, for example, that the upward component of the RB circulation appears to be primarily confined to the cloud layer while downward components (such as D0 or D1) can at times extend well below the inferred cloud base (determined here to lie between 6.5 and 6.6 km). The “plates” associated with the RB cells in altocumuli are thus more loosely defined and are presumed to consist of the local temperature inversions and the highly variable zones of radiatively driven cloud-top cooling and cloud-base warming rates that freely fluctuate in time owing to vertical displacements in the overall cloud top, cloud base, and liquid water content (Fig. 7). A similar RB analogy was invoked by Mellado et al. (2009), who studied the properties of negatively buoyant plumes generated near the top of a stratiform cloud layer. Given the lack of fixed boundaries, the altocumulus RB circulations cells become susceptible to mass loss within either component of the circulation particularly if environmental conditions favor continued descent of the air parcels deep into the subcloud layer. These observed attributes of the circulation are not dissimilar to that found in stratocumulus where the cloud layer eddies are often found to extend to the lower solid surface (Feingold et al. 1996; Stevens et al. 2003; O’Connor et al. 2005).

The structure of the D0 and D1 components of the circulation provides an initial view of the coupling between the cloud and subcloud layer circulation fields first inferred from the spectra analysis shown in Fig. 5. There is a high degree of confidence that these features are in fact downdrafts that are extending into the subcloud layer as the higher FSSP drop concentrations observed over these regions of the cloud layer would tend to minimize the applied velocity corrections stemming from the use of Eq. (B1) (see appendix B for details). A second region of direct coupling between the cloud and subcloud layers is associated with the negative velocity centers that extend below cloud base within the narrow bands of higher LWC labeled V0–V3 (Fig. 7). The velocity and condensate structure of these features are quite similar to the virga fall streaks observed in precipitating altocumulus and stratocumulus (Wang et al. 2004; Luke and Kollias 2013). The location of these

features beneath the prominent updrafts in cloud layer (such as U2 and U3) is also similar to the measurements obtained in precipitating stratocumulus by O’Connor et al. (2005). The negative velocity signatures noted to arise in the virga streaks they observed were attributed to a combination of subcloud downdrafts and an evaporatively induced reduction in the overall drop concentration below cloud base. These factors were noted to lead to an increase contribution to the Doppler signal from the surviving (and faster falling) precipitation-sized hydrometeors in the subsaturated subcloud layer.

A similar viewpoint of the subcloud virga velocity signatures is adopted here given the presence of a low concentration of larger drops measured by the 2D-C probe in the cloud layer (Fig. 4d) and the observed dropoff in the FSSP concentrations below cloud base (Fig. 4b). The qualitative calculations of the vertical drop displacement from cloud base indicated by the bold red dots in Fig. 7 labeled 2a and 2b, for example, indicate that the larger particles measured by the 2D-C could fall well into the subcloud layer before experiencing complete evaporation. As drops with radii smaller than $50 \mu\text{m}$ are not indicated to travel nearly as far (red dots labeled 1a and 1b), we note that the velocity corrections derived from Eq. (B1) would thus begin to increase in the subcloud layer owing to the overall inferred reduction in the total concentration. Calculations shown in appendix B using a fixed value of $\sigma = 0.35$ and a representative value of the reflectivity within the virga of -30 dBZ indicate, in fact, that the correction begins to approach the values measured in the V0–V3 virga shafts if the drop concentrations within these features happened to fall into the $(0.01\text{--}1) \text{ L}^{-1}$ range. Given the sensitivity of the velocity retrieval to the drop concentration, and the lack of a significant number of measurements of the actual drop concentrations deeper within the subcloud layer, we caution that the computed velocity values in the subcloud layer shown in Fig. 7 are best viewed as an upper limit to the actual downdraft magnitude.

One last noteworthy feature of Fig. 7 is the series of shallow vertical velocity couplets that appear near cloud top (labeled $D_a\text{--}D_c$ and $U_{2a}\text{--}U_{2c}$). Their structure is shown in greater detail in the composite image displayed in Fig. 8, which combines the vertical velocity and the derived LWC and longwave radiative heating rates obtained over the upper 150 m of the cloud layer. The structure of the velocity couplets near the top of the U2 updraft is particularly interesting as we find a number of shallow smaller-scale negative downdrafts (contoured in blue and also labeled D_a , D_b , and D_c) interspersed with series of smaller-scale updrafts labeled U_{2a} , U_{2b} , and U_{2c} . The location of the primary updraft turrets correspond

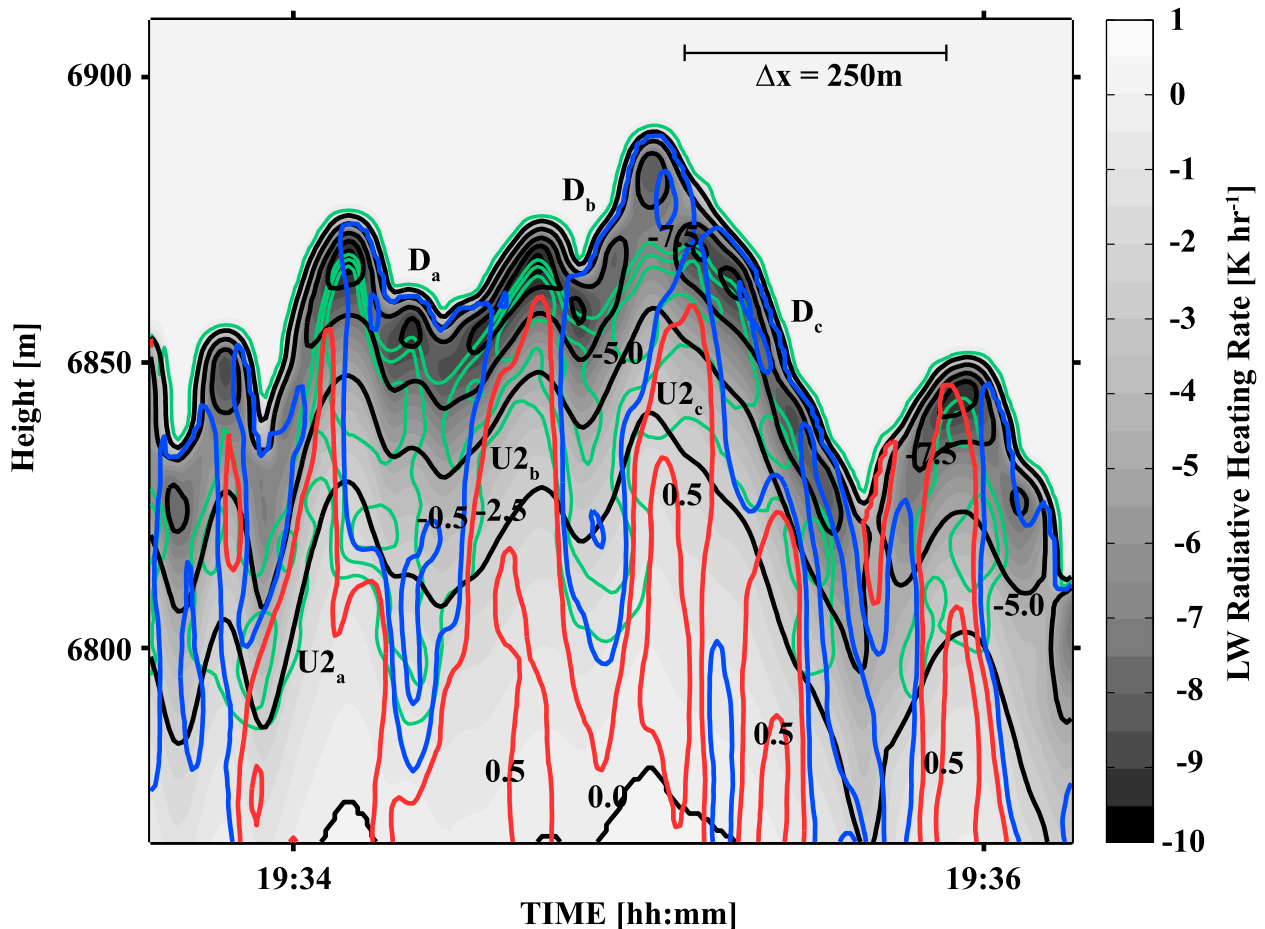


FIG. 8. A time–height analysis of the merged MCR narrowband and wideband velocity, liquid water content, and derived radiative heating rates between 1931 and 1936 UTC 27 Aug 2010 for a subset of the domain shown in Fig. 7. Derived radiative heating rates are shaded and contoured in black at intervals of 2.5 K h^{-1} . Positive velocity values are denoted by the red contours shown at intervals of 0.1 and 0.5 m s^{-1} while the negative velocity values are contoured in blue at intervals of -0.3 and -0.5 m s^{-1} . The velocity features labeled $U2_a$, $U2_b$, $U2_c$, D_a , D_b , and D_c are discussed in the text. The solid green contours represent the derived liquid water content contoured at values of 0.01 , 0.15 , 0.16 , and 0.17 g m^{-3} . The displayed analysis domain has a fixed $\Delta h = 1 \text{ m}$. The displayed length scale is computed as in Fig. 2.

well spatially with the local upward displacements evident in the cloud-top height. They are also seen to be associated with a pronounced upward inflection in the isopleths of the derived LWC within the upper 50 m of the layer leading to a structure where the peak LWC values are located near the top and lateral flanks of each turret. It is interesting that these smaller-scale turrets residing near the top of the primary U2 and U3 updrafts (Fig. 7) first become apparent in the region where the strongest vertical gradient in the cloud-top cooling is analyzed. Overall, the derived radiative heating rates varied from values of -7 to -9 K h^{-1} near cloud top to 0.5 K h^{-1} within the lower portion of the cloud layer. These values are consistent with the radiative heating rates computed in other altocumulus studies (Larson et al. 2007). It is also notable that the smaller-scale D_a ,

D_b , and D_c downdrafts all tend to lie within the flanking LWC maxima in regions where the local radiative cooling rates are also near their peak values (Fig. 8). As noted in an earlier study by Heymsfield et al. (1990), it is in such narrow zones where we see an association between the subkilometer variation in the radiative forcing and condensate loading with the negative velocity component of the observed altocumulus cloud circulation.

d. Subcloud layer virga and circulation structure

A second example of coupling between the cloud and subcloud layer circulation is shown for an isolated and narrow band of enhanced reflectivity that is also observed to extend below cloud base (labeled V1 in Fig. 9a). As in the previous examples shown in Fig. 7, the V1 feature is

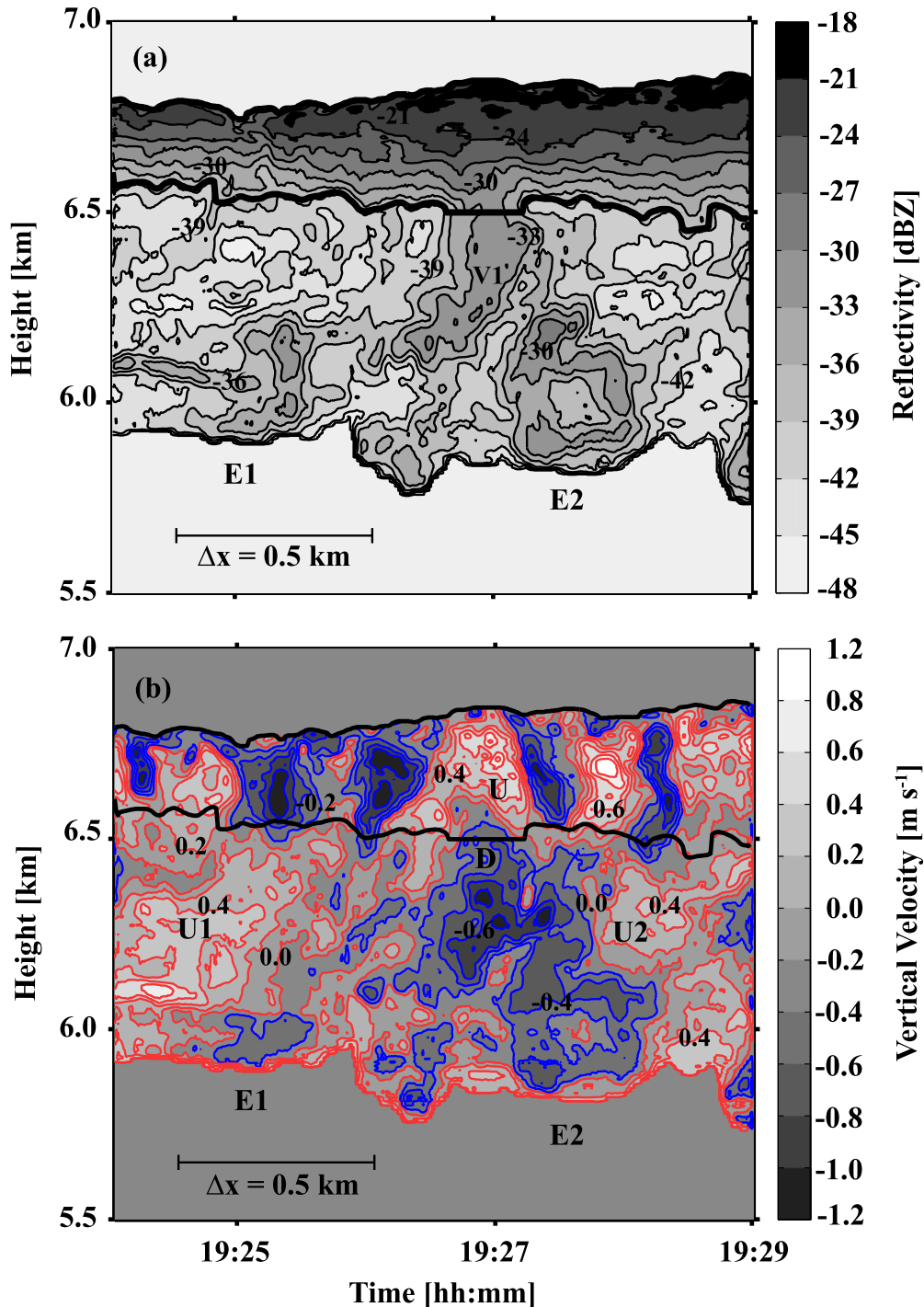


FIG. 9. A time–height analysis of the merged MCR narrowband and wideband data between 1924 and 1929 UTC 27 Aug 2010 showing (a) the radar-derived reflectivity (shaded and contoured at intervals of 3 dBZ) and (b) condensate corrected Doppler-derived perturbation vertical velocity (shaded) and contoured at intervals 0.2 m s^{-1} . The displayed analysis domain has a fixed $\Delta h = 1$ m. The bold solid black curves in both panels represent objectively and subjectively determined estimates of the cloud boundaries; that is, the horizontal segment of the cloud base through the primary drizzle shaft that extends below cloud base near 1927 UTC was subjectively set to a fixed altitude of 6.5 km. The symbols E1 and E2 denote two nearly circular reflectivity features on either side of the primary virga shaft that suggest the presence of horizontally oriented rotor circulations. The symbols U, U1, U2, and D represent positive and negative vertical velocity features discussed in the text. The displayed length scale is computed as in Fig. 2.

regarded as a virga shaft that resides below one of the main updrafts in the cloud layer and is also associated with a narrow band of negative vertical velocities which first appears near cloud base (labeled D in Fig. 9b). The noteworthy feature of this particular plot is the nearly circular reflectivity structure labeled E2 (Figs. 9a,b) that appears on the right lateral flank of V1 near an elevation of 6 km. This closed annular structure is nearly 500 m across and is positioned in a region of lateral shear in the vertical velocity formed between the centrally located feature D that accompanies the V1 virga shaft and the broad updraft labeled U2 evident to the right of E2 (Fig. 9b). A similar, though less well defined, feature labeled E1 is positioned to the left of V1 in a zone of lateral shear in the vertical velocity of the opposite sign that arises between D and the updraft labeled U1. The presence of the circular reflectivity features in the regions of counter rotating horizontal vorticity zones that arise on either flank of V1 bears resemblance to the numerically simulated counterrotating horizontal rotor structures obtained by Mellado et al. (2009). In their case, the rotors were associated with the descent of narrow negatively buoyant plumes from the top of a stratiform cloud layer (see their Fig. 4)—a situation not too dissimilar from that depicted in Fig. 9. The combined length scale of the E1 and E2 rotor circulations in the subcloud layer (as measured by the distance between the bounding U1 and U2 updrafts) extends across nearly the entire 2-km width of the displayed domain. The horizontal scale of this circulation feature is thus considerably greater than those evident in the cloud layer above including that of U from which D and V1 emanate. The development of such circulation features in the subcloud layer may thus account for some of the variability in the scale of the circulation obtained between the cloud and subcloud layers initially indicated in the analysis of Fig. 5. The circulation associated with these features within this portion of the subcloud layer also helps explain the increase in the mean positive vertical velocity component shown to arise just above the base of the conditionally unstable layer (Fig. 6a).

e. Subcloud layer virga and radiative structure

In one final example of the cloud reflectivity and velocity structure, we return to a broader overview of the cloud layer at a time when the depth of the subcloud reflectivity layer was at its greatest vertical extent (Fig. 10). The Doppler analysis reveals a number of RB cells in the cloud layer as well as additional examples of the linkage in the circulation between the cloud and subcloud layer associated with the negative component of the derived velocity. Two prominent examples of this linkage include the downward velocity band labeled D1 that is

associated with a virga shaft that formed beneath a well-defined RB updraft in the cloud layer and the RB-based downdraft labeled D2 that extends from cloud top into the subcloud layer. As discussed previously, we view these features to result from a mixture of subsiding air parcels and the Doppler signature associated with a lower concentration of larger precipitation-sized particles in the subcloud layer. The presence of similar features throughout other portions of this longer time series plot demonstrate the ubiquitous nature of these circulation features that were first evident in the shorter time series plots of Fig. 7 and Fig. 9.

The longer-term time record of the cloud layer also reveals the distinct separation in the scale of the circulation features between the cloud and subcloud layers that tends to occur near cloud base. This was suggested by power spectra analysis shown in Fig. 5 as well as the velocity variance field shown in Fig. 6d, which both indicated well-defined minimums in their respective values at this level. An interesting and surprising aspect of the plot, in particular, is the broad meso- γ -scale (2–20 km) appearance of the negative velocity structure within the subcloud layer. A portion of this signature is attributable to the D1- and D2-type velocity signatures noted above that are embedded throughout the subcloud layer. Two thermodynamic processes that may also act to influence the characteristics of this broad subsidence zone include the diabatic cooling associated with the evaporation and sublimation of the drops and ice crystals in the drier air below cloud base and the net radiative heating and cooling rate structure of both the cloud and subcloud layers.

The bold colored circles in Fig. 10a provide a qualitative assessment of the vertical displacement that drops and crystals of various sizes would undergo prior to their complete evaporation or sublimation in the subcloud layer. In these calculations, use is made of the lower and upper limits of the estimated relative humidity range shown in Fig. 3c and all particle displacements are based on their descent from a common point denoted by the large blue dot labeled S located near an altitude of 6.6 km. A variety of particles falling without the aid of any vertical transport are calculated to evaporate (sublimate) completely within 500 m of cloud base. This displacement corresponds well with the depth found to characterize the shallower regions of the subcloud virga layer. Evaporative cooling over the lower portion of the virga layer is indicated to arise from the displacement of faster falling particles initially having radii in excess of $100\text{ }\mu\text{m}$ embedded within the stronger downdrafts and more favorable thermodynamic conditions stemming from the use of the higher relative humidity range depicted by the error bars shown in Fig. 3c.

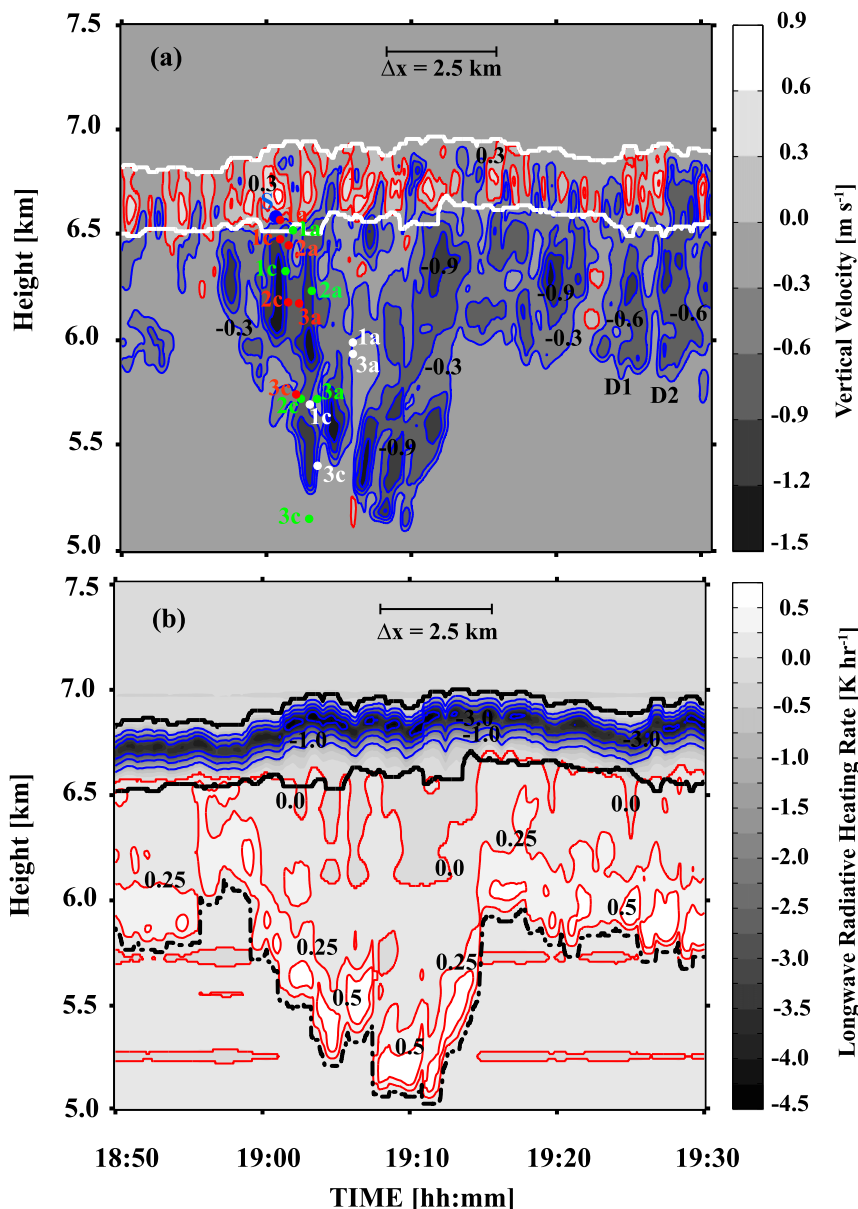


FIG. 10. Time–height plot from 1840–2020 UTC 27 Aug 2010 showing (a) the narrowband condensate corrected Doppler-derived vertical velocity (m s^{-1}) and (b) the derived longwave radiative heating rates (K h^{-1}). Positive vertical velocity values in (a) are denoted with light gray shading and red contours at intervals of 0.01, 0.3, and 0.6 m s^{-1} and negative velocity values are given by the darker gray shading and solid blue contours shown at intervals of -0.3 , -0.6 , -0.9 , and -1.2 m s^{-1} . The solid colored circles represent the terminal points of drops (solid green and red dots) and hexagonal crystals (solid white dots) released from a common point near 6600 m (blue dot labeled S). The numbers 1, 2, and 3 next to each dot represent initial particle radii of 50, 100, and 150 μm , respectively. The additional labels “a” and “c” denote particles falling at their terminal velocity or with the aid of a specified downdraft of -1.0 m s^{-1} , respectively. The red and green dots are drops released using the environmental data shown by the lower and upper limits of the error bars depicted in Fig. 3, respectively. The horizontal displacement for each drop is computed assuming a differential velocity of 1 m s^{-1} between the cloud and subcloud layers. (b) The darker gray shaded areas represent negative longwave radiative heating rates (solid blue contours) at intervals of -1.0 , -2.0 , -3.0 , and -4.0 K h^{-1} . White shaded areas denote positive heating rates contoured at 0.0, 0.25, and 0.5 K h^{-1} (solid red lines). The solid white and bold black curves in (a) and (b), respectively, denote the cloud boundaries as in Fig. 2. The bold dashed–dotted line in (b) depicts the base of the virga layer. The horizontal scale displayed in the plot was derived as in Fig. 2. The negative velocity centers labeled D1 and D2 in (a) highlight regions of coupling between the cloud and subcloud layers.

As indicated in Fig. 10b, the temperature tendencies due to the evaporation in the subcloud layer has the opposite sign to that associated with the derived radiative heating rates obtained below cloud base. This results, in part, from the positive derived radiative heating rates associated with the vertical flux divergence of the upwelling longwave radiation. Like the inferred evaporative cooling, the longwave radiative heating is spread over a deep portion of the subcloud layer with peak values tending to occur near the base of the virga layer rather than at cloud base. The dominant longwave radiative signature in the cloud layer thus becomes the much stronger -5 K h^{-1} band of cooling located near the cloud top and it is this feature that commonly serves as a focus as one of the primary driving mechanisms for the smaller-scale circulation features in stratiform cloud layers (such as the RB cells and associated D1- and D2-type downdrafts shown in Fig. 10a).

The vertical displacement of the peak longwave warming rates from cloud base is greatest near 1910 UTC as the base of the virga layer reaches its maximum depth (Fig. 2). Overall, this vertical shift of the maxima heating rates in time leads to a layer mean warming that extends well below the level of the mean cloud base (Fig. 11a). This warming is further strengthened as a result of the positive contribution arising from the mean shortwave component, which is seen to have nearly the same magnitude as the longwave component in the subcloud layer. Together, these two components lead to a weak secondary heating maximum evident at a vertical displacement of nearly 800 m from cloud top. This level corresponds well with the mean depth of the virga layer over the observational period and also closely corresponds with base of the conditionally unstable layer shown previously in Fig. 3d (see the dashed line labeled BCU in Fig. 11a), suggesting these two features may be linked. Whether this secondary heating peak helps produce or maintain the conditionally unstable subcloud layer merits additional study given the suggested linkage of this thermodynamic feature with the change in the circulation length scale below cloud base. It is this deeper layer of warming that would also tend to reduce the evaporatively driven diabatic cooling and downdraft forcing below cloud base in this low liquid water content environment. This motivates a discussion in the following section of additional factors that could also possibly contribute to the mesoscale structure of the observed subcloud subsidence field.

4. Discussion: Cloud-induced mesoscale circulation

Examples shown thus far illustrate the structure of the derived subkilometer-scale and mesoscale radiative

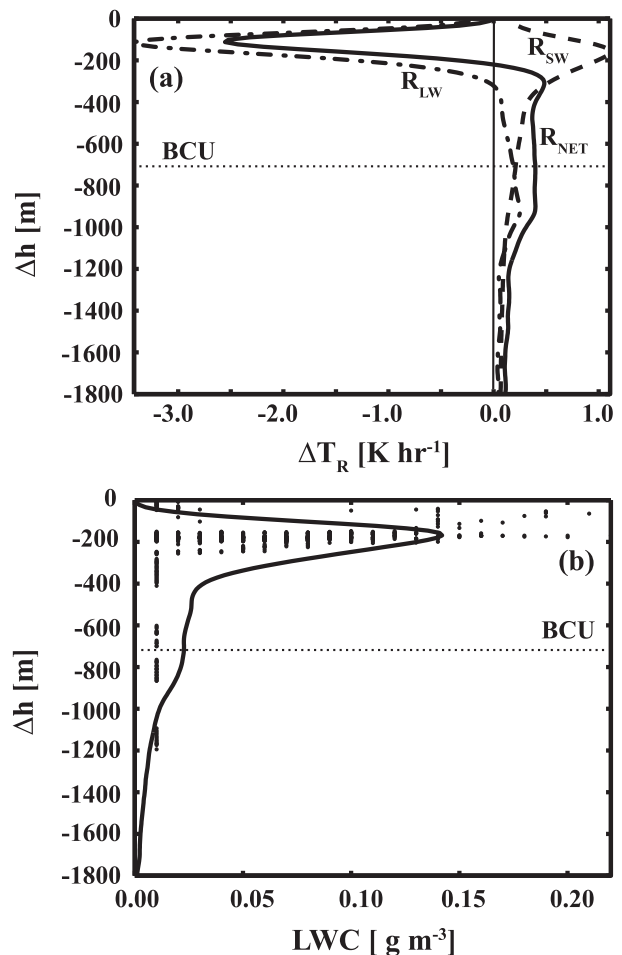


FIG. 11. The MCR-derived mean fields for the upper cloud layer between 1840 and 2005 UTC showing (a) the derived radiative heating rates (K h^{-1}) and (b) the derived liquid water content (g m^{-3}) as compared against the instantaneous aircraft observations (denoted by the bold black dots). The labeled curves in (a) represent the shortwave (dashed curve, R_{SW}), longwave (dotted-dashed curve, R_{LW}), and net (solid curve, R_{NET}) radiative heating components. The two ordinates represent the vertical displacement (Δh) from the normalized cloud top. The dotted line labeled BCU in each panel denotes the base of the conditionally unstable layer relative to cloud top.

heating rate profiles associated with this particular altocumulus cloud layer (Fig. 7 and Fig. 10b). The precise role of radiative heating acting on the overall cloud circulation structure and evolution is a complex issue as it encompasses a number of interrelated processes acting across a broad range of length and time scales (Fiedler 1984; Heymsfield et al. 1991; O'Connor et al. 2005; Larson et al. 2006; Durran et al. 2009). While the nature of the radiative forcing on the smaller-scale eddies within the layer has been well researched, it appears less certain to what extent these subkilometer circulations may be modulated by mesoscale or

synoptic-scale circulation features arising either within the cloud layer or surrounding environment. Larson et al. (2006), for example, notes that their simulated cloud lifetimes changed by a factor of four when the imposed synoptic-scale subsidence was varied from 1 to 6 cm s^{-1} . More recent work by Durran et al. (2009) also indicates that weak positive radiatively heated cloud layers can induce thermally forced mesoscale gravity wave circulations that can act to enhance overall cloud lifetime provided the waves have sufficient time to develop.

The mesoscale circulation forced by the strong radiative cooling associated with thin altocumulus layers in the middle troposphere represents an interesting counterexample to that obtained by Durran et al. (2009) for thin cloud layers in the upper troposphere undergoing weak positive heating. To show this, we examine the nature of their linear solutions to a prescribed forcing representative of altocumuli using a two-dimensional x - z framework that is 8 km deep and 800 km long and that has fixed vertical and horizontal grid increments of 25 and 1000 m, respectively. The solutions on this domain are obtained from their linearized Eqs. (1)–(7), which govern the horizontal u and vertical w velocity, Boussinesq pressure potential, buoyancy, and two-dimensional divergence terms in a Boussinesq fluid. The solutions are obtained in an identical manner wherein their analytical expression for w serves as a basis for numerically computing the remaining diagnostic and prognostic equations for the horizontal flow and buoyancy equations. As in their study, we assume quiescent environmental flow conditions and specify a thermodynamic structure that consists of fixed static stability ($N = 0.011 \text{ s}^{-1}$). The simulation was initialized with the heating function given by Eq. (5) of Durran et al. (2009) except that we specify the center level of the heating function using conditions representative of the observed cloud altitude of 6.5 km. The half-width of the heating profile was set to a value of 10 km in order to mimic the observed cloud layer width obtained both here (Fig. 2) and in a number of other altocumulus cases (Ansmann et al. 2009). To replicate the bimodal nature of the heating and cooling structure evident over the upper 300 m of Fig. 11a, use was made of two half-cosine heating functions of opposing sign with peak cooling and heating rates of -5 and 2 K h^{-1} , respectively. Each half-cosine profile was 240 m thick and slightly offset vertically from the reference level to produce an overall heating layer 300 m thick. The model was applied once for each separate half-cosine heating profile and linear superposition of each solution was then used to obtain the final result. This is a legitimate approach owing to the linear nature of the Durran et al. (2009) analytical

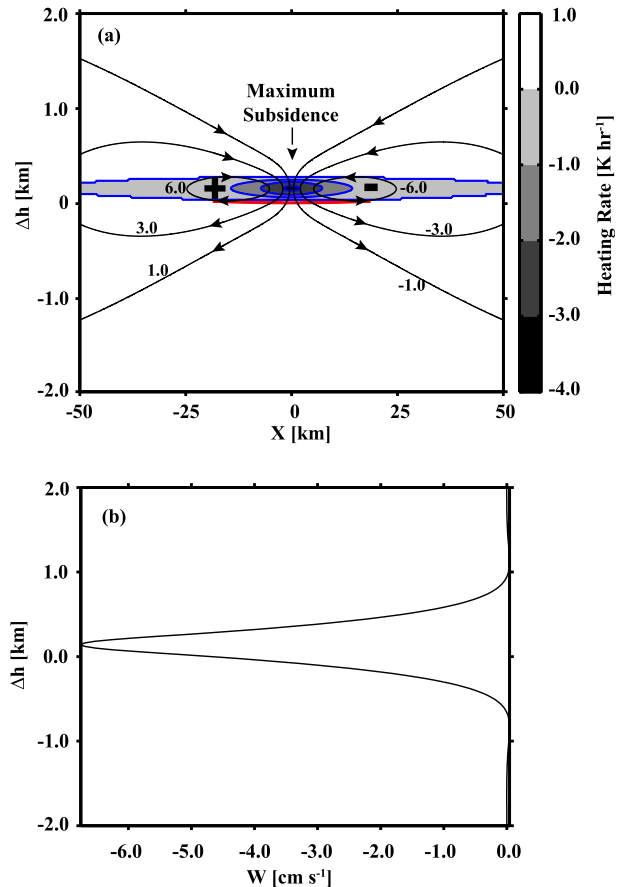


FIG. 12. Linear solution to the imposed heating function shown after 2 h of elapsed time: (a) the streamfunction ($\times 10^2 \text{ m}^2 \text{ s}^{-1}$) and the imposed heating rate shaded and contoured at intervals of 1 K h^{-1} and (b) the mean vertical velocity averaged over the heating half-width. The solid lines in (a) represent positive and negative streamfunction contoured at values of ± 1 , ± 3 , and $\pm 6 \times 10^2 \text{ m}^2 \text{ s}^{-1}$, respectively. The two ordinate values represent the displacement (Δh km) from the center of the vertical domain used to obtain the analytical solutions. The bold black plus and minus signs and black arrows in (a) depict the sense of the horizontal vorticity in the cross-plane direction generated by the applied heat source.

solution for the vertical velocity component of the circulation.

Application of their model under these conditions indicates that thermally forced waves associated with altocumulus cloud layers can induce a deep layer of subsidence over a broad area of the cloud field in as little as 2 h of elapsed integration (Fig. 12). A fascinating aspect of these solutions is that subsidence is induced by the finite attributes of the lateral width of the cloud layer itself and extends vertically for well over a kilometer both above and below the imposed heating (Fig. 12b). As suggested by the nature of the streamfunction displayed Fig. 12a, this subsidence is supported by narrow bands of inflow near the top of the heating zone and

outflow below. This indicates a reversal in the overall flow pattern obtained by Durrán et al. (2009) for the weakly positively heated cloud layer examined in their study. Weaker uplift is found along the primary axis of the imposed heating beginning at distances slightly greater than the specified 10 km half-width. Given the symmetry, the net circulation thus takes the form of two counterrotating horizontal meso- β -scale (20–200 km) vortices that act to preferentially force downward motion over a large area near the geometrical center of the imposed cooling (near $x = 0$ and $z = \Delta h = 0.1$ km). The sign of vorticity couplets can be inferred from forming a horizontal vorticity equation from Eqs. (1) and (2) of Durrán et al. (2009) and through consideration of the resulting sign of the horizontal gradient of the buoyancy term that appears on the right-hand side of their Eq. (2). The rapidity and extent of the derived mesoscale response for typical altocumulus heating rates suggest that this type of forcing could quickly act to modulate or otherwise influence the smaller-scale individual circulation cells or other attributes of the cloud layer. Note, in particular, that the peak magnitude of the mean velocity computed over the 10-km half-width of the cloud layer is on the order of 7 cm s^{-1} (Fig. 12b). This is a value of subsidence in excess of that found by Larson et al. (2006) to significantly alter the cloud lifetime in their simulated altocumulus cases.

The structure shown in Fig. 12 is a simplified representation of the actual cloud layer and most applicable to altocumulus layers that exhibit 2D or circular symmetry. These solutions also neglect the diabatic and turbulence mixing tendencies that will partially offset or otherwise alter the structure of the layer heating as a whole. However, several initial inferences can still be drawn from the derived fields that could be readily tested in a more advanced numerical framework wherein the hydrostatic waves and other smaller-scale cloud circulation features are fully coupled. First, we see that the wave-induced subsidence could lead to preferential drying near the geometrical center of the cloud layer ($x = 0$ and $\Delta h = 0.1$ km in Fig. 12a) that would tend to reduce (or possibly destroy) the very same horizontal liquid water gradient that spawned the circulation in the first place. Second, if the subsidence is sufficiently strong, it could overcome any positive mesoscale forcing, leading the cloud layer in this region to completely evaporate. Third, the wave-induced subsidence could contribute to the development and maintenance of the stable inversion near cloud top—a feature often attributed, in part, to synoptic-scale forcing alone. Finally, subsidence associated with the wave-induced meso- β -scale circulation could combine with that forced by other circulation features, such as those depicted in Fig. 7 through Fig. 10,

to generate a deep layer of descent that possibly extends well beyond that resolved within the radar-detected cloud and subcloud virga boundaries alone. Further investigation of the role of this subsidence in modulating the overall behavior of the cloud layer is left for more advanced numerical model simulations in which the interplay between both features can be directly assessed over a much wider variety of environmental conditions and initial cloud structures.

5. Conceptual model and summary

A conceptual model summarizing the primary cloud and subcloud layer features of this observed altocumulus cloud layer is presented in Fig. 13. The primary attributes of the altocumulus cloud layer include (i) a series of shallow small-scale eddies near cloud top that may represent the incipient circulations associated with entrainment (depicted as the short curved black arrows), (ii) deep and well-defined vertical velocity maxima associated with Rayleigh–Bénard-type convective circulations [denoted by the thin red (updraft) and blue (downdraft) solid lines evident within the bold white cloud boundaries], (iii) vertically coherent downdrafts between the cloud and subcloud layers (depicted by the black arrows of medium thickness), (iv) a deep and broad meso- γ -scale negative velocity signature associated with the primary virga shafts in the subcloud layer (denoted with the thickest black arrows), and (v) subcloud rotor circulations associated with negative buoyancy that arose along the flanks of individual narrow virga shafts (vertically oriented black ellipses). Additional observational details include the aircraft-derived multilayered thermodynamic profile that includes the rapid change in stability near cloud top, the presence of a conditionally unstable lapse rate in both the cloud layer and the portion of the subcloud layer that contained the bulk of the virga layer (labeled subcloud layer 1), and an increase in static stability below the base of the virga layer (labeled subcloud layer 2).

In addition to the observed characteristics, the schematic model includes two inferred features derived from application of an advanced radiation scheme initialized with locally observed atmospheric profiles and the MCR-based LWC. One feature is the strong layer of longwave radiative cloud-top cooling (blue dashed line) and the longwave radiative warming (red dashed line), which dips from a location near cloud base to the lower extremities of the deeper virga shafts. This vertical shift in the maximum radiative heating rates resulted in a weak layer of radiatively induced warming that extended over the entire subcloud layer as indicated in Fig. 11a. A second inferred feature consists of the meso- β -scale rotor circulations

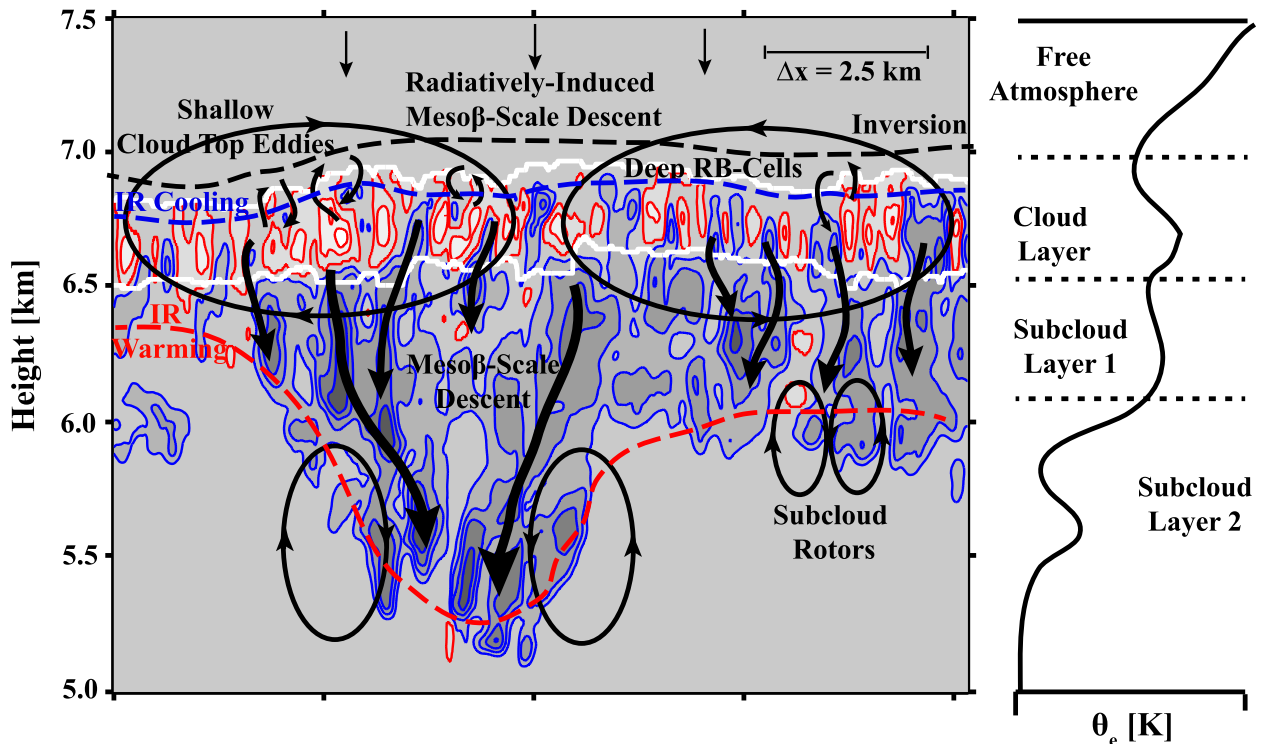


FIG. 13. (left) Schematic diagram summarizing the primary observational features of the observed altocumulus cloud layer. Gray shading represents the MCR narrowband Doppler analysis of the vertical velocity obtained between 1850 and 1930 UTC 27 Aug 2010. Negative vertical velocity values are depicted by the darker gray shaded regions bounded by the solid blue contours (shown at intervals of -0.3 , -0.6 , -0.9 , and -1.2 m s^{-1}). Positive vertical velocity values are depicted by the lighter gray shaded regions bounded by the solid red contours (shown at intervals of 0.0 , 0.1 , and 0.3 m s^{-1}). The alternating updraft and downdraft circulations extending through the depth of the cloud layer (bounded by the bold white curves) depict the deep Rayleigh-Bénard-type convective circulations. The shorter curved thin black arrows suggest the presence of shallow eddies near cloud top. The longer and thicker black arrows depict the parcel movement of individual small-scale downdrafts that originate at various levels within the cloud layer. The thickest and longest black arrows depict the regions of meso- γ -scale negative velocity associated with the Doppler measured terminal velocity values and downdraft circulations present in the subcloud layer. The thin black vertically oriented ellipses denote subcloud rotors associated with the primary virga shafts. The horizontally oriented black ellipses denote the thermally forced meso- β -scale rotor circulations inferred from the analytical solutions. The bold dashed black line depicts the top of the inversion. The downward pointed arrows above the inversion denote subsidence in the free atmosphere. The maximum longwave radiative cooling and heating rates are depicted by the bold dashed blue and red lines, respectively. The horizontal scale shown at the top of the figure was produced by assuming a mean cloud advection of 5.5 m s^{-1} . (right) A simplified static stability profile depicts several prominent thermodynamic layers accompanying this event.

that were derived when the linear model of Durran et al. (2009) was forced by a prescribed heating function that was designed to mimic that produced by a thin cloud layer of midtropospheric origin undergoing strong cloud-top cooling and weak cloud-base warming. This meso-scale circulation is depicted as horizontally orientated black ellipses (thin black ellipses with directional arrows) that produce uplift along the periphery of the imposed heating function and subsidence both above and below the cloud layer. As inferred from the structure shown in Fig. 12, the strongest induced subsidence is found near the geometrical center of the imposed heat source owing to the nature of the horizontal vorticity generated by a layerwide lateral gradient in the radiative heating structure within the cloud layer.

The high-resolution Doppler data reveals detailed information on the updraft structure near cloud top as well as the formation of subcloud rotor circulations tied to the flanks of individual virga shafts. Such features were discussed in respect to the regulating role of the radiative heating on scales ranging from the individual Rayleigh-Bénard cells and the meso- β -scale thermally forced hydrostatic waves of the type studied by Durran et al. (2009). The wave structure obtained for altocumulus heating rates was interesting as the induced circulation led to zone of maximum subsidence near the center of the imposed heating. When applied to a real cloud layer, such a circulation could act to significantly reduce (or possibly destroy) the very same liquid water gradient that spawned the circulation in the first place.

Whether this represents another form of self-regulation that alters the characteristics of longer-lived altocumulus cloud layers, such as maintenance of the stable inversion near cloud top, remain unclear. As such, the dynamics of these features and their importance on the overall cloud layer and surrounding environment await further study within the framework of advanced large-eddy simulation models in which the interplay between the smaller-scale eddies and meso- β -scale radiative forcing over narrow altocumulus cloud layers can be examined in a full three-dimensional setting spanning a wide range of environmental conditions. Whether such radiatively induced meso- β -scale subsidence zones impact the observed structure in more expansive stratiform cloud fields, such as the generation of the pockets of open cells observed in stratocumulus, merits further investigation.

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APPENDIX A

The MCR Radar Pulse Volume

The calculation of the MCR radar pulse volume is performed in the following manner. First, from the definition of the radar reflectivity (η) we have

$$\eta = \frac{\pi^5 |K|^2 z}{\lambda^4} \left(\frac{\text{m}^2}{\text{m}^3} \right), \quad (\text{A1})$$

where the factor $K = (\epsilon - 1)/(\epsilon + 1)$ depends on the dielectric constant (ϵ) of the scatters, z is the radar reflectivity factor ($\text{m}^6 \text{m}^{-3}$) and λ is the wavelength (m). The incremental radar cross-section ($d\sigma$) can be expressed in terms of the incremental volume (dv) as

$$d\sigma = \eta dv. \quad (\text{A2})$$

For scatters located at a range R from the radar, the incremental volume can be expressed as

$$dv = g_T^2 \left(\frac{x}{R} \right) g_E^2 \left(\frac{y}{R} \right) g_r(\zeta) dx dy d\zeta, \quad (\text{A3})$$

where $g_T(\theta)$ is the one-way antenna traverse normalized power response θ in radians, $g_E(\phi)$ is the one-way antenna elevation normalized power response in ϕ radians, and $g_r(\zeta)$ is the matched filter output normalized point source power response (ζ) in meters. Using a Gaussian fit for all power responses and expressing the standard deviations in terms of the one-way 3-dB width in radians, we have

$$g_T \left(\frac{\theta_3}{2} \right) = \exp \left[-\frac{1}{2} \left(\frac{\theta_3}{2\sigma_T} \right)^2 \right] = \frac{1}{2} \quad (\text{A4})$$

and

$$g_E \left(\frac{\phi_3}{2} \right) = \exp \left[-\frac{1}{2} \left(\frac{\phi_3}{2\sigma_E} \right)^2 \right] = \frac{1}{2}, \quad (\text{A5})$$

and for range power response with the standard deviation expressed in terms of the 6-dB width we have

$$g_r \left(\frac{R_6}{2} \right) = \exp \left[-\frac{1}{2} \left(\frac{R_6}{2\sigma_r} \right)^2 \right] = \frac{1}{4}. \quad (\text{A6})$$

It follows that

$$\sigma_T = \frac{\theta_3}{\sqrt{8 \ln(2)}}, \quad (\text{A7})$$

$$\sigma_E = \frac{\phi_3}{\sqrt{8 \ln(2)}}, \quad \text{and} \quad (\text{A8})$$

$$\sigma_r = \frac{R_6}{\sqrt{16 \ln(2)}}, \quad (\text{A9})$$

where σ_T and σ_E are in radians and σ_r is in meters. From Eq. (A2), the total scatter volume may be expressed as

$$V = \iiint g_T^2 \left(\frac{x}{R}\right) g_E^2 \left(\frac{y}{R}\right) g_r(\zeta) dx dy d\zeta, \quad (\text{A10})$$

from which we obtain the final result:

$$V = R^2 \left[\frac{\pi}{8 \ln(2)} \right]^{1.5} \theta_3 \phi_3 R_6. \quad (\text{A11})$$

APPENDIX B

Radar Processing, Conversions, and the Drop Model

This appendix provides a brief description of additional radar analysis techniques, the Z–LWC conversion used in the study, and the formulations used to compute the vertical displacement of the drop and ice crystals described in the text. Additional steps taken to process the radar prior to correcting the velocity or computing any derived fields included first subjectively editing the data by hand to remove unwanted radar artifacts. These include ground clutter, periodic transmitter stoppages, short temporal gaps between consecutive radar scans, and parabolic shaped reflectivity maxima caused by the movement of the research (or other) aircraft passing over the radar site. Small data gaps created as a result of the editing were then filled with a Barnes (1973) objective analysis technique using no more than 12 surrounding points. The objectively analyzed fields were used to create continuous display plots and the spectral analysis shown in Fig. 5. All mean fields were created without the use of the objectively analyzed fields.

The correction for the terminal velocity of the drops within a given radar pulse volume used in this study stems from the methodology of Baker and Hodson (1985), Frisch et al. (1995), O'Connor et al. (2005), and Marsham et al. (2006). In these studies the contribution of the hydrometeors to the net Doppler velocity W_{net} is based on the reflectivity weighted terminal velocity of an assumed drop size distribution written here in the form

$$W_{\text{cor}} = \frac{\int_0^\infty V(D) D^6 n(D) dD}{\int_0^\infty D^6 n(D) dD}, \quad (\text{B1})$$

where D is the particle diameter (m), $V(D)$ is the diameter-based droplet velocity (m s^{-1}), and $n(D)$ represents the assumed particle size distribution (m^{-4}). The aircraft data presented in the text reveal the presence of

a low concentration of precipitation-sized particles as well as a droplet population in the observed cloud layer where maximum drop sizes were less than $40 \mu\text{m}$. We thus follow Frisch et al. (1995) and Flatau et al. (1989) and use a lognormal size distribution written in the form

$$n(D) = \frac{N_t}{\sqrt{2\pi\sigma}D} \int_0^\infty \{-[\ln(D/D_N)/\sqrt{2\sigma}]^2\} dD, \quad (\text{B2})$$

where σ represent logarithmic width of the distribution, D_N is referred to as the characteristic diameter (m), and N_t represents the total number of particles (m^{-3}). We use a value of $\sigma = 0.35$ after Frisch et al. (1995).

Additional expressions of interest are those for the liquid water content (M) and radar reflectivity factor z . Using Eq. (B2) and the traditional definitions of z and M , we obtain after Flatau et al. (1989) and Frisch et al. (1995)

$$M = \frac{\rho_l N_t D_N^3 \exp(9\sigma^2/2)}{6} \quad \text{and} \quad (\text{B3})$$

$$z = N_t D_N^6 \exp(18\sigma^2). \quad (\text{B4})$$

Eliminating D_N in Eqs. (B3) and (B4) leads to an expression for M in terms of z :

$$M = 0.0003 \rho_w N_t^{0.5} z^{0.5}, \quad (\text{B5})$$

where the factor 0.0003 accounts for the conversions when z is expressed in $\text{mm}^6 \text{m}^{-3}$ and M and ρ_w are expressed in units of g m^{-3} . This can be compared to expressions of the form $M = az^b$ such as can be found in the study by Khain et al. (2008), who examined the Z–LWC relationship for both precipitating and non-precipitating boundary layer stratocumulus as well as other cloud liquid water systems. The integral for Eq. (B1) is solved numerically using the velocity–diameter relationships of Pruppacher and Klett (1978, 322–324) for the regime I ($0.1 < a < 10 \mu\text{m}$) and regime II ($10 < a < 535 \mu\text{m}$) particle sizes where a represents the particle radius. The value of the dynamic viscosity (η), free path (λ), and density (ρ) used in their expressions [Eq. (10-106) and Eq. (10-107a); p. 323] was determined using a temperature of 261 K and an observational pressure of 450 hPa.

Representative values of the velocity correction obtained from the application of Eq. (B1) can be estimated for fixed values of the reflectivity and σ . The calculation procedure first involves converting the input

reflectivity to z and then applying Eq. (B5) to obtain the LWC. We can solve Eq. (B3) for the characteristic diameter D_N and use this value in the expression for the velocity correction [Eq. (B1)]. In our case, Eq. (B1) was numerically integrated using the above-mentioned expressions for the velocity–diameter relationships. As an example, using a reflectivity value representative of the virga shafts (-30 dBZ), each order of magnitude change in the total concentration from the peak observed value of $2.8 \times 10^6 \text{ m}^{-3}$ can be shown to produce corresponding values of the correction term obtained from Eq. (B1) of 0.03, 0.06, 0.12, 0.24, 0.45, 0.8, and 1.3 m s^{-1} , respectively.

The detailed cloud particle evaporation and sublimation rates used to determine the particle survival distance below cloud based depicted in Fig. 7 and Fig. 10a were developed from the studies of Hall and Pruppacher (1976) for ice crystals and Pruppacher and Klett (1978) for liquid drops [using their Eqs. (13-61) and (13-62), p. 425; Eq. (13-3), p. 413; Eq. (13-16), p. 418; and Eqs. (13-57) and (13-58), p. 443]. The saturation vapor pressure for liquid and ice was determined from lookup tables based on the study of Goff and Gratch (1946). The environmental temperature and vapor pressure values needed each time step were obtained from the knowledge of the drop altitude at a given time using vertically interpolated input derived from the aircraft sounding shown in Fig. 3. Updated values of the drop radius at each time step were used to compute the vertical displacement using a time step of 1 s. The drop terminal velocities were again determined from the regime I and regime II size ranges discussed in Pruppacher and Klett (1978) using the temperature-dependent dynamic viscosity values determined from their Eq. (10-107), p. 323. The crystal type was assumed to be a hexagonal plate having a length to width aspect ratio indicated in Hall and Pruppacher (1976). The crystal fall speed follows that used for hexagonal plates discussed by Pruppacher and Klett (1978) [their Eqs. (10-134) and (10-136), p. 337]. For the data shown, the vertical displacement was computed using either the terminal velocity of the particle alone or combined with a specified downdraft (set here to a value of either -0.5 or -1.0 m s^{-1}). We did not attempt to account for the thermodynamics of the descent path of a given downdraft parcel in these calculations. The calculations were terminated when the particle reached a radius of $1 \mu\text{m}$ or less during a time step. The drop descents depicted in Fig. 7 used the more favorable relative humidity (RH) environment constructed by taking the upper limit of the error bars depicted in Fig. 3c. Both the upper (green dots) and lower (red dots) limits of the RH where used to calculate the drop descents depicted in Fig. 10a.

REFERENCES

- Ansmann, A., and Coauthors, 2009: Evolution of the ice phase in tropical altocumulus: SAMUM lidar observations over Cape Verde. *J. Geophys. Res.*, **114**, D17208, doi:10.1029/2008JD011659.
- Atlas, D., 1954: The estimation of cloud parameters by radar. *J. Atmos. Sci.*, **11**, 309–317, doi:10.1175/1520-0469(1954)011<0309:TEOCPB>2.0.CO;2.
- Baedi, R., J. de Wit, H. Russchenberg, J. Erkelens, and J. Poiras Baptista, 2000: Estimating effective radius and liquid water content from radar and lidar based on the CLARE98 data-set. *Phys. Chem. Earth*, **25B**, 1057–1062, doi:10.1016/S1464-1909(00)00152-0.
- Baker, P. W., and M. C. Hodson, 1985: Effect of deviations from the Marshall-Palmer drop size distribution on the calculation of vertical air velocity by Rogers's method. *J. Appl. Meteor.*, **24**, 495–498, doi:10.1175/1520-0450(1985)024<0495:EODFTM>2.0.CO;2.
- Barnes, S. L., 1973: Mesoscale objective map analysis using weighted time-series observations. NOAA Tech. Memo. ERLTM-NSSL-62, 60 pp.
- Baumgardner, D., 1983: An analysis and comparison of five water droplet measuring instruments. *J. Appl. Meteor.*, **22**, 891–910, doi:10.1175/1520-0450(1983)022<0891:AAACOF>2.0.CO;2.
- Bolton, D., 1980: The computation of equivalent potential temperature. *Mon. Wea. Rev.*, **108**, 1046–1053, doi:10.1175/1520-0493(1980)108<1046:TCOEPT>2.0.CO;2.
- Bower, K. N., and T. W. Choulaton, 1992: A parameterisation of the effective radius of ice free clouds for use in global climate models. *Atmos. Res.*, **27**, 305–339, doi:10.1016/0169-8095(92)90038-C.
- Carey, L., J. Niu, P. Yang, J. Kankiewicz, V. Larson, and T. Haar, 2008: The vertical profile of liquid and ice water content in midlatitude mixed-phase altocumulus clouds. *J. Appl. Meteor. Climatol.*, **47**, 2487–2495, doi:10.1175/2008JAMC1885.1.
- Doviak, R., and D. S. Zrnić, 1984: *Doppler Radar and Weather Observations*. Academic Press, 458 pp.
- Durran, D., T. Dinh, M. Ammerman, and T. Ackerman, 2009: The mesoscale dynamics of thin tropical tropopause cirrus. *J. Atmos. Sci.*, **66**, 2859–2873, doi:10.1175/2009JAS3046.1.
- Feingold, G., W. R. Cotton, B. Stevens, and A. S. Frisch, 1996: The relationship between drop in-cloud residence time and drizzle production in numerically simulated stratocumulus clouds. *J. Atmos. Sci.*, **53**, 1108–1122, doi:10.1175/1520-0469(1996)053<1108:TRBDIC>2.0.CO;2.
- Fiedler, B., 1984: The mesoscale stability of entrainment into cloud-topped mixed layers. *J. Atmos. Sci.*, **41**, 92–101, doi:10.1175/1520-0469(1984)041<0092:TMSOEI>2.0.CO;2.
- Flatau, P. J., G. J. Tripoli, J. Verlinde, and W. R. Cotton, 1989: The CSU-RAMS cloud microphysics module: General theory and code documentation. Colorado State University Department of Atmospheric Science Tech. Rep., 88 pp.
- Fleishauer, R. P., V. E. Larson, and T. H. Vonder Haar, 2002: Observed microphysical structure of midlevel, mixed-phase clouds. *J. Atmos. Sci.*, **59**, 1779–1804, doi:10.1175/1520-0469(2002)059<1779:OMSOMM>2.0.CO;2.
- Fox, N. I., and A. J. Illingworth, 1997: The retrieval of stratocumulus cloud properties by ground-based cloud radar. *J. Appl. Meteor.*, **36**, 485–492, doi:10.1175/1520-0450(1997)036<0485:TROSCP>2.0.CO;2.
- Frisch, A., C. Fairall, and J. Snider, 1995: Measurement of stratus cloud and drizzle parameters in ASTEX with a K_a -band

- Doppler radar and a microwave radiometer. *J. Atmos. Sci.*, **52**, 2788–2799, doi:[10.1175/1520-0469\(1995\)052<2788:MOSCAD>2.0.CO;2](#).
- Fu, Q., and K. N. Liou, 1992: On the correlated k -distribution method for radiative transfer in nonhomogeneous atmospheres. *J. Atmos. Sci.*, **49**, 2139–2156, doi:[10.1175/1520-0469\(1992\)049<2139:OTCDMF>2.0.CO;2](#).
- Goff, J. A., and S. Gratch, 1946: Low-pressure properties of water from –160 to 212 F. *Trans. Amer. Soc. Heat. Vent. Eng.*, **51**, 125–164.
- Hall, W., and H. Pruppacher, 1976: The survival of ice particles falling from cirrus clouds in subsaturated air. *J. Atmos. Sci.*, **33**, 1995–2006, doi:[10.1175/1520-0469\(1976\)033<1995:TSOIPF>2.0.CO;2](#).
- Heymtsfield, A. J., K. M. Miller, and J. D. Spinhirne, 1990: The 27–28 October 1986 FIRE IFO cirrus case study: Cloud microstructure. *Mon. Wea. Rev.*, **118**, 2313–2328, doi:[10.1175/1520-0493\(1990\)118<2313:TOFICC>2.0.CO;2](#).
- , L. M. Miloshevich, A. Slingo, K. Sassen, and D. O’C. Starr, 1991: An observational and theoretical study of highly supercooled altocumulus. *J. Atmos. Sci.*, **48**, 923–945, doi:[10.1175/1520-0469\(1991\)048<0923:AOATSO>2.0.CO;2](#).
- Hobbs, P. V., A. L. Rangno, M. Shupe, and T. Uttal, 2001: Airborne studies of cloud structures over the Arctic Ocean and comparisons with retrievals from ship-based remote sensing measurements. *J. Geophys. Res.*, **106**, 15 029–15 115, doi:[10.1029/2000JD900323](#).
- Houze, R. A., Jr., 1994: *Cloud Dynamics*. International Geophysics, Vol. 53, Academic Press, 573 pp.
- Inverarity, G., 2000: Correcting airborne temperature data for lags introduced by instruments with two-time-constant responses. *J. Atmos. Oceanic Technol.*, **17**, 176–184, doi:[10.1175/1520-0426\(2000\)017<0176:CATDFL>2.0.CO;2](#).
- Khain, A., M. Pinsky, L. Magaritz, O. Krasnov, and H. W. J. Russchenberg, 2008: Combined observational and model investigations of the Z–LWC relationship in stratocumulus clouds. *J. Appl. Meteor. Climatol.*, **47**, 591–606, doi:[10.1175/2007JAMC1701.1](#).
- King, W. D., D. A. Parkin, and R. J. Handsworth, 1978: A hot-wire liquid water device having fully calculable response characteristics. *J. Appl. Meteor.*, **17**, 1809–1813, doi:[10.1175/1520-0450\(1978\)017<1809:AHWLWD>2.0.CO;2](#).
- Knollenberg, R. G., 1981: Techniques for probing cloud microstructure. *Clouds: Their Formation, Optical Properties, and Effects*, P. Hobbs, Ed., Academic Press, 15–91.
- Kollias, P., and B. Albrecht, 2000: The turbulence structure in a continental stratocumulus cloud from millimeter-wavelength radar observations. *J. Atmos. Sci.*, **57**, 2417–2434, doi:[10.1175/1520-0469\(2000\)057<2417:TTSIAC>2.0.CO;2](#).
- Korolev, A., 2007: Reconstruction of the sizes of spherical particles from their shadow images. Part I: Theoretical considerations. *J. Atmos. Oceanic Technol.*, **24**, 376–389, doi:[10.1175/JTECH1980.1](#).
- , E. F. Emery, J. W. Strapp, S. G. Cober, G. A. Isaac, M. Wasey, and D. Marcotte, 2011: Small ice particles in tropospheric clouds: Fact or artifact? Airborne icing instrumentation evaluation experiment. *Bull. Amer. Meteor. Soc.*, **92**, 967–973, doi:[10.1175/2010BAMS3141.1](#).
- Larson, V. E., A. J. Smith, M. J. Falk, K. E. Kotenberg, and J. C. Golaz, 2006: What determines altocumulus dissipation time? *J. Geophys. Res.*, **111**, D19207, doi:[10.1029/2005JD007002](#).
- , K. E. Kotenberg, and N. B. Wood, 2007: An analytic longwave radiation formula for liquid layer clouds. *Mon. Wea. Rev.*, **135**, 689–699, doi:[10.1175/MWR3315.1](#).
- Li, X., and R. Srivastava, 2001: An analytical solution for raindrop evaporation and its application to radar rainfall measurements. *J. Appl. Meteor.*, **40**, 1607–1616, doi:[10.1175/1520-0450\(2001\)040<1607:AASFRE>2.0.CO;2](#).
- Liu, S., and S. K. Krueger, 1998: Numerical simulations of altocumulus using a cloud resolving model and a mixed layer model. *Atmos. Res.*, **47–48**, 461–474, doi:[10.1016/S0169-8095\(98\)00034-9](#).
- Luke, E. P., and P. Kollias, 2013: Separating cloud and drizzle radar moments during precipitation onset using Doppler spectra. *J. Atmos. Oceanic Technol.*, **30**, 1656–1671, doi:[10.1175/JTECH-D-11-00195.1](#).
- Marshall, J. H., S. Dobbie, and R. J. Hogan, 2006: Evaluation of a large-eddy model simulation of a mixed-phase altocumulus cloud using microwave radiometer, lidar and Doppler radar data. *Quart. J. Roy. Meteor. Soc.*, **132**, 1693–1715, doi:[10.1256/qj.05.145](#).
- Mellado, J. P., B. Stevens, H. Schmidt, and N. Peters, 2009: Buoyancy reversal in cloud-top mixing layers. *Quart. J. Roy. Meteor. Soc.*, **135**, 963–978, doi:[10.1002/qj.417](#).
- Moeng, C.-H., and R. Rotunno, 1990: Vertical-velocity skewness in the buoyancy-driven boundary layer. *J. Atmos. Sci.*, **47**, 1149–1162, doi:[10.1175/1520-0469\(1990\)047<1149:VVSITB>2.0.CO;2](#).
- Nicholls, S., 1984: The dynamics of stratocumulus: Aircraft observations and comparisons with a mixed layer model. *Quart. J. Roy. Meteor. Soc.*, **110**, 783–820, doi:[10.1002/qj.49711046603](#).
- Noh, Y.-J., C. J. Seaman, T. H. Vonder Haar, and G. Liu, 2013: In situ aircraft measurements of the vertical distribution of liquid and ice water content in midlatitude mixed-phase clouds. *J. Appl. Meteor. Climatol.*, **52**, 269–279, doi:[10.1175/JAMC-D-11-0202.1](#).
- O’Connor, E. J., R. J. Hogan, and A. J. Illingworth, 2005: Retrieving stratocumulus drizzle parameters using Doppler radar and lidar. *J. Appl. Meteor.*, **44**, 14–27, doi:[10.1175/JAM-2181.1](#).
- Peters, G., B. Fischer, H. Münster, M. Clemens, and A. Wagner, 2005: Profiles of raindrop size distributions as retrieved by microrain radars. *J. Appl. Meteor.*, **44**, 1930–1949, doi:[10.1175/JAM2316.1](#).
- Pincus, R., and M. B. Baker, 1994: Effect of precipitation on the albedo susceptibility of clouds in the marine boundary layer. *Nature*, **372**, 250–252, doi:[10.1038/372250a0](#).
- Probert-Jones, J., 1962: The radar equation in meteorology. *Quart. J. Roy. Meteor. Soc.*, **88**, 485–495, doi:[10.1002/qj.49708837810](#).
- Pruppacher, H. R., and J. D. Klett, 1978: *Microphysics of Clouds and Precipitation*. Springer, 714 pp.
- Rayleigh, L., 1916: On convection currents in a horizontal layer of fluid, when the higher temperature is on the under side. *Philos. Mag.*, **32**, 529–546.
- Sassen, K., 1991: The polarization lidar technique for cloud research: A review and current assessment. *Bull. Amer. Meteor. Soc.*, **72**, 1848–1866, doi:[10.1175/1520-0477\(1991\)072<1848:TPLTFC>2.0.CO;2](#).
- Schmidt, J., and Coauthors, 2012: Radar observations of individual rain drops in the free atmosphere. *Proc. Natl. Acad. Sci. USA*, **109**, 9293–9298, doi:[10.1073/pnas.111776109](#).
- Smith, A. J., V. E. Larson, J. Niu, J. A. Kankiewicz, and L. D. Carey, 2009: Processes that generate and deplete liquid water and snow in thin midlevel mixed-phase clouds. *J. Geophys. Res.*, **114**, D12203, doi:[10.1029/2008JD011531](#).
- Starr, D. O., and S. K. Cox, 1985: Cirrus clouds. II: Numerical experiments on the formation and maintenance of cirrus. *J. Atmos. Sci.*, **42**, 2682–2694, doi:[10.1175/1520-0469\(1985\)042<2682:CCPINE>2.0.CO;2](#).

- Stevens, B., and Coauthors, 2003: Dynamics and Chemistry of Marine Stratocumulus—DYCOMS-II. *Bull. Amer. Meteor. Soc.*, **84**, 579–594, doi:[10.1175/BAMS-84-5-579](https://doi.org/10.1175/BAMS-84-5-579).
- Stickney, T. M., M. W. Shedlov, and D. I. Thompson, 1994: Total temperature sensors. Rosemount Aerospace/BFGoodrich Tech. Rep. 5755, Revision C, 32 pp. [Available from Rosemount Aerospace/BFGoodrich, 1256 Trapp Road, Eagan, MN 55121.]
- Suring, R., 1950: *Die Wolken*. Akademische Verlagsgesellschaft, 153 pp.
- Verver, G., M. Fujiwara, P. Dolmans, C. Becker, P. Fortuin, and L. Miloshevich, 2006: Performance of the Vaisala RS80A/H and RS90 Humicap sensors and the Meteolabor “Snow White” chilled-mirror hygrometer in Paramaribo, Suriname. *J. Atmos. Oceanic Technol.*, **23**, 1506–1518, doi:[10.1175/JTECH1941.1](https://doi.org/10.1175/JTECH1941.1).
- Wang, Z., K. Sassen, D. N. Whiteman, and B. B. Demoz, 2004: Studying altocumulus with ice virga using ground-based active and passive remote sensors. *J. Appl. Meteor.*, **43**, 449–460, doi:[10.1175/1520-0450\(2004\)043<0449:SAWIVU>2.0.CO;2](https://doi.org/10.1175/1520-0450(2004)043<0449:SAWIVU>2.0.CO;2).