Spatial and Microphysical Characteristics of Low-Ceiling, Temperature-Inverted Clouds in Warm Overrunning and Freezing-Rain Conditions: A Case Study

RICHARD K. JECK
William J. Hughes Technical Center, Federal Aviation Administration, Atlantic City International Airport, Atlantic City, New Jersey

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ABSTRACT
In-flight microphysical measurements in classical freezing-rain conditions were used to study the vertical and horizontal characteristics of the precipitation and associated low-ceiling, stratiform clouds, which are usually present as overcast in freezing-rain conditions. The low overcast is usually based in the surface cold layer but may extend up into the inversion, or transition layer, between the overrunning warm air and the surface cold layer. This gives the cloud an unusual temperature-inverted structure—supercooled in the lower half and warmer than freezing in the upper half. The low cloud is also subject to wind shear and turbulence that is due to the warm overrunning. The apparent effects of this are 1) increased cloud droplet concentrations in clusters up to a few hundred meters wide that occur sporadically in the cloud layer, 2) possible forcing of cloudy air upward from lower levels against the resistance of the temperature inversion and into the transition layer, and 3) highly variable air temperatures during level flight in the inversion layer.

1. Introduction
Aviation weather observations show that a low ceiling is usually present during classical freezing-rain (FZRA) conditions (Guttman and Jeck 1987). So-called classical FZRA occurs when a relatively shallow, surface air mass at temperatures below 0°C is overridden by slantwise advection of air warmer than 0°C. This provides a melting layer for snow falling from cold clouds farther above. Below the melting layer, a notable temperature inversion marks the boundary between the cold, surface air mass and the relatively warmer overriding air mass. The melted snow becomes supercooled rain while falling through the surface air mass and freezes on objects with surface temperatures below 0°C on or above the ground. A good schematic illustration of the temperature T and cloud structure is shown in Fig. 263 of Byers (1944).

a. Low overcast clouds during FZRA
The low ceiling or overcast is due to a stratus cloud that is based in the surface cold layer or in the overrunning warm layer. In fact, cloud layers inferred from the relative humidity data in radiosondes indicate that freezing-rain conditions may be accompanied by a variety of low cloud profiles. Rauber et al. (2000) classified them into six different categories, depending on their depth and cloud-top temperature. Three independent examples from the Washington, D.C., area are shown in Fig. 1. In most FZRA cases, the low-ceiling clouds1 are based in the surface cold layer. The overrunning warm layer may be unsaturated and cloud free, as in Fig. 1a where the low cloud extends part way up into the inversion (transition) layer. On occasion, the surface cold layer is unsaturated but the warm layer is humid, in which case the overcast starts in the warm layer, as in Fig. 1b. More often, both layers have been humidified enough that the overcast starts in the cold layer and extends well into the warm layer or beyond, as in Fig. 1c.

Rauber et al. (2000, their Fig. 4) find that these low-ceiling clouds range in depth from a few hundred meters

1 In Fig. 1, the presence of cloud is inferred by the author’s unpublished version of a method proposed by Appleman (1954) for detecting the presence of clouds from radiosonde relative humidity (RH) data. My version infers cloud wherever the reported RH aloft is equal or greater than the following temperature-dependent threshold value: RH \( \geq 93\% \) for \( T < -5°C \), and RH \( \geq 0.9T + 100\% \) for \( -25° < T \leq -5°C \).
to 8 km, depending on the category. The airplane penetrations in the current study found the low-ceiling overcast to range from about 215 to 900 m (700–3000 ft) deep, with an average depth of about 500 m (1700 ft).

The overcast can form in the surface cold layer as a result of ordinary convection, turbulent mixing, upslope flow, or humidification of the layer by (partial) evaporation of raindrops falling through. Saturation of the coldest part of the layer may be expected first, which is at the base of the inversion, with the cloud base gradually expanding downward from there. Strong temperature inversions ordinarily block the upward progress of clouds, stopping them at the base of the inversion. This means that the top of the convective cloud layer should be coincident with the base of the inversion.

In subsidence and trade-wind inversions, the air immediately above the inversion is usually dry and cloud free. In FZRA conditions, however, the air in the overrunning warm layer may be either dry or humid. If sufficiently humid, clouds may form in the overrunning layer because of gradual cooling in the upslope flow as the warm air slides up over the cold layer. Evaporation of rain can help to saturate the warm layer too if the initial humidity is great enough (Dolezel 1944).

b. Temperature-inverted clouds

Most of these low overcast clouds accompanying FZRA span part or all of the inverted temperature profile such that the cloud layer is supercooled below the inversion and is warmer than freezing in the inversion and melting layer. The inversion barrier implies that the normal bottom-to-top in-cloud circulations do not occur across the barrier. The upper (warmer) half of the cloud may have a different supply of cloud condensation nuclei and water vapor for droplet growth than the lower half. This should make for some interesting microphysics but, except for two known reports (Thomas and Marwitz 1995; Rasmussen et al. 1995), the effects on drop size and liquid water content (LWC) profiles in these low-ceiling clouds are practically unexplored in FZRA conditions. These clouds are therefore of interest because of the different ways they may be formed and because of their unusual inverted temperature profile. When they are present as supercooled clouds they can also add to the icing rate on any aircraft that may happen into the FZRA conditions. One question is whether there are significant differences in the cloud microphysics above and below the base of the inversion. In section 3 we examine this question in detail.

2. Data

The microphysical measurements used here are part of a larger dataset of about 4700 n mi of quality airborne
measurements in FZRA and freezing drizzle (FZDZ) collected from various sources for aircraft-icing studies (Jeck 2010). Several different analyses of these FZRA and FZDZ data may be found in Cober et al. (2009) and Jeck (2010). The current paper shows selected in-flight data from this collection as provided by the University of North Dakota (UND).

For convenience, the LWC is divided into five components: 1) the ordinary cloud water content (CWC) for droplets in the 1–100-μm range of diameters, 2) the drizzle water content (DWC) for drops in the 100–500-μm range, 3) the rainwater content (RWC) for drops that are larger than 500 μm, 4) the precipitation liquid water content (PLWC = DWC + RWC), and 5) the total liquid water content (TLWC = CWC + PLWC).

The droplet data are from conventional forward-scattering spectrometer probes (FSSP), one-dimensional cloud (1D-C) probes, and 1D precipitation (1D-P) probes manufactured by Particle Measuring Systems, Inc, (PMS) in Boulder, Colorado. Their 1D measure of the particle size does not offer the use of particle shape to discriminate between ice particles and water drops. In mixed-phase conditions, this can be a problem. In the current case, however, the atmospheric conditions are ideal for minimizing or eliminating ice particle interference. The measurements of interest are below a melting layer that is 1200 m (4000 ft) or more deep through which no ice particles are likely to survive. Any ice particles that may nucleate in the supercooled layer below the melting layer are expected to be few and small such that they add negligibly to the rainwater content that is computed from the 1D droplet size distribution. Even if some of the FZRA drops refroze as sleet, they would still be legitimate counts, equivalent to FZRA drops, for the purposes of this paper. That is because they are still roughly spherical, were liquid drops higher up, or would still be liquid drops in cases in which the temperature was a bit warmer at their location.

The 1D probe data are more likely to contain artifacts (false, large-particle counts) because of shedding or splashing of water off the probe tips, especially in high-LWC conditions. The author’s method for recognizing and eliminating probable artifacts is explained in the appendix.

3. Case study—Freezing rain at Kansas City, Missouri, on 1 February 1990

This event is one example of low-ceiling clouds in overrunning and FZRA conditions. The event was sampled from the air for 2.5 h by a flight research team from UND. This was a major ice storm with freezing precipitation that lasted 9 h. Some aspects of the storm were described by Prater and Borho (1992). They mention that the FZRA formed a 7–12-mm layer of clear ice on all exposed surfaces in the metropolitan Kansas City area and caused numerous reports of severe aircraft icing in the vicinity of Kansas City Municipal Airport (MKC). The current paper describes some features and characteristics that were not previously analyzed and reported in detail.

Figure 2 shows the state of the atmosphere near Kansas City at the time of the research flight. The figure is from a synoptic rawinsonde launched at Topeka, Kansas. Topeka is only 100 km west of Kansas City, and the conditions there were similar to those in Kansas City. The outside air temperatures (OAT) and wind speeds measured on the UND flight near Kansas City closely match those in Fig. 2. There is no radiosonde station closer to Kansas City, but the Topeka sounding made a good substitute in this case.

a. Observed vertical profiles of low cloud variables in FZRA

Figure 3 shows the profile of temperature and CWC during takeoff and ascent in FZRA at Kansas City on 1 February 1990. The UND research airplane ascended to the coldest (−5°C) level in the cold layer and loitered there at 660 m (2250 ft) AGL in midcloud for 13 min before ascending higher on the initial profile run. This −5°C level was also the base of a sharp temperature inversion [11°C (290 m)−1] marking the boundary between the overlying warm layer and the underlying subfreezing layer. During the loiter, an initial northbound heading out of MKC took them about 20 n mi north, where they turned west for another 20 n mi before continuing the ascent.

In this case, the low-ceiling cloud layer was 600 m (1900 ft) deep. Cloud base was at about 375 m (1250 ft)
AGL at −4°C. The cloud is of the type in Fig. 1a for which the cloud starts in the cold layer and extends up into the inversion layer—in this case to +5°C at about 970 m (3180 ft) AGL.

The surface weather observations at MKC recorded continuous 8/8 sky cover with a ceiling height slowly descending from 410 to 320 m (from 1350 to 1050 ft) AGL in continuous light freezing rain during the 2.5-h flight. The flight data revealed, however, that the overcast cloud was highly nonuniform, and frequently discontinuous, as indicated in Fig. 3 by the CWC excursions to 0 g m$^{-3}$ during the midcloud loiter at constant altitude. The horizontal variability of the stratiform cloud density is clearer in Fig. 4 in which the CWC, OAT, and altitude are plotted as functions of time.

1) CLOUD DROPLET CONCENTRATION

The peaks in the CWC reflect local increases in the cloud droplet concentration (CONC) in the overcast. According to the FSSP (3–51-μm range of diameters), the peaks in CWC were entirely due to momentary increases in the number of cloud droplets in the first six size channels (3–16-μm diameter). Throughout the flight penetration at midcloud level, the cloud droplet mean diameter and mean-volume diameter (MVD) remained approximately constant at about 9 and 11 μm, respectively. Only the droplet number density changed (see Table 1). Pulses of CWC like these, due to increased droplet number densities but unchanged shape of the drop size distributions, were described in detail by Korolev and Mazin (1993). Similar CWC pulses appear in other FZRA cases (not shown) in the present collection, and in Rasmussen et al. (1995, their Fig. 18).

The fluctuations in temperature, shown more clearly in Fig. 4, are mostly uncorrelated with the pulses in CWC. Both they and the CWC pulses are discussed in more detail later in section 3b.

This variability is shown another way in Fig. 5 in which the cloud droplet concentration from the FSSP is plotted second-by-second during ascent through the overcast. The impression is that of a background droplet concentration steadily increasing with height in the first 150 m (from points A to G) above cloud base, but accentuated by random pulses of extra droplets at A, D, and F. The drop concentration appears to decrease over the next 150 m
### Table 1. Cloud drop size distributions and related details for selected cloud parcels indicated in Figs. 3–5.

<table>
<thead>
<tr>
<th>Data point label</th>
<th>Time (LT)</th>
<th>Assigned droplet diameter (μm) at channel center</th>
<th>Droplet concentration (cm$^{-2}$) in each size channel</th>
<th>FSSP mean diam (μm)</th>
<th>FSSP LWC (g m$^{-3}$)</th>
<th>FSSP mean-vol diam</th>
<th>Alt (ft AGL)</th>
<th>Alt (m AGL)</th>
<th>OAT (reverse flow; °C)</th>
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<td></td>
<td></td>
<td>2.8 4.8 8.3 13.1 15.5 19.4 24 27.6 32.8 36.7 40.7 44.3 47.7 51.4</td>
<td></td>
<td></td>
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<tr>
<td>A</td>
<td>1657:39</td>
<td>46 46 22 1 0 0.2 0.2 0.05 0.05 0.03 0 0 0 0 0 0</td>
<td>69 0.03 6 9 1289 393</td>
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<tr>
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<td>7 7 13 3 0 0.3 0.1 0.03 0.03 0 0 0 0 0 0 0 0</td>
<td>202 0.15 9 11 1506 459</td>
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<tr>
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<td>1657:44</td>
<td>97 97 233 52 6 0.7 0.1 0.05 0.14 0.02 0.02 0 0 0 0 0 0 0</td>
<td>389 0.25 8 11 1538 469</td>
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<tr>
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<td>1657:46</td>
<td>36 36 120 36 5 0.8 0.2 0.04 0.02 0.02 0.09 0 0 0 0</td>
<td>198 0.14 9 11 1604 489</td>
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<td>609 0.47 9 11 1692 516</td>
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<td>91 78 78 77 13 1.2 0.3 0.15 0 0.05 0 0 0 0 0 0 0</td>
<td>249 0.19 9 11 1830 558</td>
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(from G to J) up to the base of the inversion, but the intermittency of the cloud parcels there makes this conclusion uncertain. In the inversion layer (from M to U), the droplet concentration decreases with height toward cloud top but is still punctuated by occasional bursts of extra droplets at M, P, and S. Table 1 provides more details on this behavior.

The data records in Table 1 are listed in order of ascending altitude. The first 10 records (A–J) cover the 25-s climb from cloud base to the beginning of level flight in midcloud at the base of the inversion. The last nine records (M–U) cover the 45-s ascent from midcloud to cloud top. The cloud (A–J) in the cold layer, despite the variability of droplet concentration and CWC, maintains a remarkably consistent mean diameter and mean-volume diameter of about 9 and 11 μm, respectively. This includes the spikes (K and L) in droplet numbers encountered during level flight in midcloud. The cloud in the inversion layer (M–U) has a background droplet population that decreases with height (and with increasing temperature). It has a 25%–50%-larger mean diameter and mean-volume diameter of up to 12 and 16 μm, respectively, relative to the cloud at, and below, the base of the inversion. The spikes (M, P, S, and U) in droplet concentration still have about the same mean diameter (8 or 9 μm) and mean-volume diameter (11 μm) as in the cold-layer cloud below, however.

The cloud below the inversion obviously has its origin in the surface cold layer, but what is the origin of the cloud in the highly stable transition layer above the base of the inversion? The cloud does not extend into the overrunning warm layer, and so its origin is not there. One possibility is turbulent transport of cloudy air upward from below that is due to wind shear forces in the transition layer. The farther up the cloudy air is forced, the warmer it becomes because of the inverted temperature lapse in the transition layer. This warming could preferentially evaporate the smaller drops, leaving a lower droplet count and relatively larger residual mean droplet diameters. This may explain the gradual reduction in the background cloud with height as observed in the upper half of Figs. 3 and 5 and in rows M–U in Table 1. It does not explain the spikes in droplet count as at points M, P, and S, however. These will be discussed in more detail in section 3b.

2) RAINWATER CONCENTRATION

Although the density of the low cloud appears to be unevenly distributed and occasionally discontinuous, the freezing rain that is falling through it appears to be more continuous, although variable, as judged from the rain LWC (hereinafter called RWC, as computed from the 1D-P precipitation drop size distribution). Figure 6 shows the variable CWC riding on a background RWC of about 0.1 g m\(^{-3}\), with occasional showery rain intervals like that between about 1700 and 1706 LT.

The measured icing rate is from a calibrated (Jeck 2007) Goodrich (formerly Rosemount) model 0871FA ice detector and is the rate on a 6.35-mm (1/4 in.) diameter cylinder at aircraft speed. The icing rate seems to
respond mostly to the trends in RWC, with the CWC peaks sometimes adding to the icing rate. The icing rate is also reduced because of incomplete freezing of the impinging drops when the OAT rises to $-1^\circ$C or higher, as at 1659 and 1711:30 LT (see OAT trace in Fig. 4). The highest icing rates of the flight were momentarily about 2.7 mm min$^{-1}$ (6 in. h$^{-1}$) in a later pass through the FZRA at this altitude. During that pass, the pilots had to increase fuel consumption on the twin jet by 60%—from 225 to 365 kg h$^{-1}$ (from 500 to 800 lb h$^{-1}$)—to maintain the uniced performance level. If this icing rate were sustained during a 15-min approach and landing, potentially up to 38 mm (1.5 in.) of ice could accumulate on some parts of the aircraft during this time.

3) RWC DISTRIBUTIONS VERSUS DROP SIZE FOR SELECTED FZRA INTERVALS

Table 2 lists the RWC distributions at several positions of interest along the interval shown in Fig. 6. The purpose is to show the differences between the bottom and top of the profile and between the shower and background portions of the FZRA. The columns labeled 1–5 at the top of Table 2 give the LWC contributions as 1-min averages in each of several drop size intervals. The “% of PLWC” values to the right of each main column in Table 2 show the percent of the total large drop LWC that is contained in each drop size interval. These relative percentages make the LWC distributions easier to compare.

Looking at the columns from right to left (from top to bottom of the cloud), there is a gradual decrease in the DWC components (0.1–0.4 mm) from just under cloud top to just below cloud base. These drops may be growing out of the drizzle-size range by collision–coalescence on their way down because, at the same time, the population of the midsize drops (0.5–1.5 mm) is growing from 60% of the PLWC near cloud top to 90% just below cloud base. The overall PLWC remains practically unchanged from top (0.12 g m$^{-3}$) to bottom (0.14 g m$^{-3}$), except for the purposely selected minimum RWC in column 3 and the larger RWC during the rain shower, in column 4.

b. Effects of wind shear, turbulence, and mixing in the transition layer

Figure 2 illustrates a typical wind profile in many freezing-rain situations (this study and Rauber et al. 2001). The surface wind is light, usually with a northeasterly to southeasterly component. With increasing height, the wind speed strengthens and veers to become westerly at all heights above the top of the temperature inversion. Thus there is usually considerable wind shear up through the inversion layer. No attempt has been made here to compute wind shear values, only to illustrate the apparent effects of it as revealed by the in-flight data.

1) EFFECTS ON THE AIR TEMPERATURE PROFILE

The usual simple depiction of the airmass interface is that of a smooth, inclined plane separating the warm overrunning air from the cold layer below (e.g., Fig. 263 in Byers 1944). The temperature profile is also usually visualized as making a smooth and linear transition from the lowest to the highest temperature in the inversion,
like that in Figs. 1 and 2. These are just useful simplifications, however, because the airplane data reveal a far more complicated picture. The temperature data alone can show the effects that wind shear and turbulence apparently have on the interface between the cold and warm air masses, and on the temperature structure in the transition layer. Figures 3 and 4 show the effects at the base of the inversion. There, the airplane intercepted intermittent parcels of air with temperatures up to 5°C warmer than the minimum 2°C at that level. This implies a folded or irregular interface at the airmass boundary, probably similar to that illustrated for a plane shear layer in Baker et al. (1984, their Fig. 1). It is evident that large-scale eddies can force air parcels from the 0°C level down as much as 150 m to the −5°C level at the base of the inversion, in this case.

Later segments of the flight (not shown) at other levels in the transition layer demonstrate the same type of OAT fluctuations. For example, with the airplane in level flight higher up at the average −2°C height in the transition layer, the airplane intercepted air parcels with temperatures as high as 0°C at an altitude of 2500 m. These same features have been observed in other FZRA flights by UND in the Kansas City area on 19 January 1990 and 14 February 1990, and independently in at least one other FZRA flight near Kansas City by the University of Wyoming on 12 February 1992 (Thomas and Marwitz 1995). Their Figs. 1 and 2 show similar sharp temperature excursions in the transition layer and above.

If the smallest bumps or notches in the OAT trace in Fig. 4 represent individual eddies, then the smallest eddies resolvable in this data are about 500 m wide.
They can occur individually like the tall, sharp peaks at 1659 and 1700 LT or in broader clusters like those centered at 1705 and 1710 LT. The approximate widths of these four clusters, measured at their bases in Fig. 4, are about 6.4, 3.6, 6.8, and 9.8 km, respectively.

Temperatures seem to be far less erratic below the base of the inversion, as is implied by the smoother profile here in Fig. 3, and in Coleman and Marwitz (2002, their Fig. 7a). Only one other case (17 March 1987, not shown) exhibited some noticeable deviation from a smooth temperature profile within the surface cold layer. Figure 3 shows that the pulses of denser cloud intercepted below the base of the inversion do not involve temperature changes either. This seems to indicate that the overturning in the transition layer is mostly confined to the transition layer and above.

The air in the transition layer is continually churned by one or more mechanisms described by Coleman and Marwitz (2002), Baker et al. (1984), and Korolev and Mazin (1993), bringing air parcels to lower or higher levels in the transition layer and possibly into the over-running warm layer above. The result is a time-varying inhomogeneous mixture with temperature fluctuating about the homogeneous average, which increases more or less linearly with height throughout the transition layer.

2) Effects on the Cloud Droplet Population

This overturning may be expected to have an effect on the cloud droplet population in the transition layer. If the up or down movements of air parcels are considered to be adiabatic processes, then cloud parcels moving downward, say a distance of 150 m (500 ft) from 0° to −5°C at the base of the inversion, would be warmed 1.5°C by compression of the parcel. This may cause some of the smaller droplets to shrink or evaporate. In any case, the warm parcel, surrounded by colder air in this inverted temperature regime, would tend to rise back up to near its original level in the transition layer until the outside air temperature matches that of the parcel. In the meantime, the parcel is losing its identity because of smaller-scale mixing, with uncertain effect on the cloud droplet population that it contained.

The evidence so far is that the cloud layer below the base of the inversion is continuous but with embedded parcels of denser cloud. These denser parcels, such as at locations D, F, and H in Figs. 3–5 and Table 1, have a markedly greater population of cloud droplets but no noticeable difference in the mean droplet sizes relative to the less dense background cloud. There is also no apparent correlation between the spikes in droplet number density and the fluctuations in air temperature. The spikes appear to fit the description of the increased droplet concentration zones (Korolev and Mazin 1993), otherwise known as droplet clusters or preferential concentrations (Vaillancourt and Yau 2000). Here, they are about 4–7 s (300–600 m) wide and are therefore 2–3 times as large as the widest observed by Korolev and Mazin.

In the base of the inversion and above, the cloud tends to be more intermittent and the air temperature tends to be highly variable, although the changes in cloud parcels and the air temperature do not seem to be correlated. That is, the displaced air parcels in the transition layer do not necessarily contain any (extra) cloud droplets, and the denser cloud parcels, such as at locations S and P in Figs. 3–5 and Table 1, do not necessarily bring noticeable changes in OAT or in RWC. The denser parcels (M, P, S, and U) higher up in the cloud (and the transition layer) have mean droplet diameters that are the same as those below the inversion.

4. Summary and Conclusions

This study details one example of low-ceiling clouds that usually accompany freezing rain, and the apparent effects of wind shear and turbulence in the airmass transition zone above the freezing-rain layer. The main observations are as follows:

- In this example case, the low-ceiling cloud apparently formed under the base of the characteristic temperature inversion where the temperature is the lowest in the surface cold layer. Wind shear and turbulence at this airmass interface may be transporting cloudy air parcels up into the inversion (transition) layer, thus accounting for the extension of the low cloud layer up into the warmer, overrunning air mass.
- The air temperature at a given height and location in the inversion (transition) layer may differ, momentarily, and positively or negatively, from the local average at that height by an amount up to the limiting temperatures in the inversion. These local temperature fluctuations are interpreted as air parcels up to several hundred meters wide that are recently displaced whole by large-scale turbulent eddies generated by wind shear in the inversion layer. This means that an individual sounding (balloon or otherwise) may not represent the true average temperature profile if one or more of these temperature-displaced air parcels are in its ascent path. For example, the actual minimum temperature at the top of the cold layer could be several degrees colder than indicated if, at that level, the sounding happened to intercept a warmer air parcel displaced down from higher up in the inversion layer.
- The temperature fluctuations are easily detected in continuous, high-resolution vertical and horizontal temperature profiles from an in situ, instrumented airplane.
They have been found in all of the FZRA with warm overrunning events examined so far (Kansas City vicinity on 19 January and 1 and 14 February 1990, and near Grand Forks, North Dakota, on 17 March 1987).

- Apparent cloud droplet clustering, presumably due to wind shear and turbulence in the warm overrunning situation, may occur at any height in the accompanying low-ceiling stratiform cloud. Peak number densities of up to 3 times more droplets per unit volume than the background cloud have been recorded in some clusters. The droplet number density does not seem to be correlated with the temperature of the air parcel containing the cluster. Some clusters, or groups of clusters, appear to be up to 600 m wide.

- Cloud droplet clusters may not be as frequent as the temperature anomalies, and they may not be easy to recognize if the cloud layer is more broken than continuous. Of the four FZRA events listed above, the 1 February case featured in this paper seems to offer the most convincing evidence for droplet clustering.

Acknowledgments. The author is indebted to C. A. Grainger at the University of North Dakota who provided the data for this study.

APPENDIX

Method for Recognizing and Minimizing Probable Artifacts in the 1D-C and 1D-P Drop Size Distributions

The particle counts from the PMS 1D-C and 1D-P particle size spectrometers can contain sporadic and anomalous counts that are due to intermittent splashing or shedding of water drops from the probe tips. These “streakers” can be easily recognized in the two-dimensional imagery of the PMS 2D particle size spectrometers. With the 1D probes, however, the suspected artifacts are usually lone counts in one of the higher particle size bins. These outliers can strongly skew the LWC or other quantities computed from the recorded particle size distributions. An example is shown in Fig. A1a in which the RWC computed from the 1D-P is plotted, second by second, during a segment of the flight near Kansas City on 1 February 1990. The suspected artifacts stand out as anomalous 1-s LWC spikes. The actual LWCs are the lower and slower-varying LWC trends indicated by the denser concentration of data points below 0.3 g m\(^{-3}\) in this example. The spikes (artifacts) occur about 1 or 2 times per minute.

The data can be corrected for artifacts by either deleting or trimming (truncating) the spikes in one of several ways. The tall spikes are obvious offenders, but a procedural rule is needed to decide which of the smaller spikes are to be considered as artifacts and which are to be allowed to stay as possibly legitimate RWCs. One way is

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**FIG. A1.** One method for minimizing artifacts among the droplet counts from the 1D-P particle size spectrometer. (a) Originally computed RWCs, where suspected artifacts appear as 1-s spikes. (b) Artifacts mostly eliminated and RWCs adjusted by truncating spikes to 0.05 g m\(^{-3}\) over previous 1-s RWC. (See text. Note change of scale for graph.) (c) RWCs smoothed by using a 20-s running average.
to declare a limiting jump in RWC, which, if exceeded, would indicate a possibly inflated RWC due to an artifact. For example, an increase of more than 0.05 g m\(^{-3}\) from one RWC to the next could be flagged as a possible artifact in the stratiform rain situation of interest here. Then, the suspected 1-s record can be deleted or it can be salvaged by replacing the RWC with a reasonable substitute value. One possibility is to set its RWC to the value of the previous RWC plus the allowable growth rate of 0.05 g m\(^{-3}\). This scheme allows for the possibility that the RWC may really be increasing, but more gradually than the sharp, 1-s spikes. When the data in Fig. A1a are processed this way, the resulting trimmed RWCs are as shown in Fig. A1b. The scheme may still inadvertently reduce a few legitimate 1-s RWCs, but at least it is a conservative correction because it always lowers the otherwise computed RWC rather than raising it.

The RWC trace can be smoothed even further by plotting a running average instead of individual values. Figure A1c shows a 20-s running average that emphasizes the main features while still preserving some of the fine structure. This procedure also helps to minimize or to eliminate large ice particles from the 1D probe data. In the freezing-rain conditions of interest here, few, if any, ice particles are expected to nucleate or otherwise be present. Nevertheless, ice particles showing up as artifacts would be screened out by the procedure described above. Any ice particles that are too small to be caught by the 1D screening process would be too few to add much to the computed RWC anyway.

REFERENCES