

**Cloud structure and crystal growth in nimbostratus clouds.**

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## **Abstract**

Cloud structure and crystal growth in two nimbostratus were examined using in situ and airborne radar observations. In both cases, structure throughout the cloud depth was modulated by generating cells at about 8 km altitude. Large-scale horizontal homogeneity at altitudes below the generating cells was due to the rapid movement of the generating cells relative to the main cloud mass. In addition, significant horizontal variability was evident on the scale of few hundred meters, principally in the radar reflectivity data but also readily detectable in the particle data. The melting layer was clearly defined in the radar images. Thin dry layers just above the melting layer were also observed in both cases.

In agreement with earlier studies, particle spectra in these clouds are adequately described by exponential relationships. There is a strong correlation between the slope ( $\lambda$ ) and intercept ( $N_0$ ) parameters; this relationship is well characterized by a power law with constants varying from case to case. Radar reflectivity is negatively correlated with the slope parameter  $\lambda$ .

It is concluded from the observations that the vapor supply made available by large scale lifting was taken up by depositional growth of the ice crystals. Aggregation of the crystals led to the final shaping of the precipitation size spectra.

*Keywords:* Nimbostratus; Cloud structure; Particle spectra; Radar

## 1. Introduction

Nimbostratus (Ns) clouds occur extensively both in the mid-latitudes and in the tropics (Houze, 1993) and are responsible for long periods of light precipitation in all seasons. Stratiform cloud regions associated with tropical squall lines are believed to contribute about half of the total precipitation associated with those storms and therefore have received considerable attention in the literature (Johnson and Hamilton, 1988; Houze, 1989; Biggerstaff and Houze, 1993). Some aspects of mid-latitude Ns are also being addressed: for example, the parameterization of ice clouds in numerical weather prediction and global climate models (Genio et al., 1996; Ødegaard, 1997), the diagnosis of such clouds with satellite observations (Lau and Crane, 1997) and the effects of nimbostratus on the radiation balance (Poetsch-Heffter et al., 1995).

Nimbostratus usually extend to near the tropopause and contain no strong localized updrafts. Earlier studies (e.g., Lo and Passarelli, 1982; Herzegh and Hobbs, 1985; Gordon and Marwitz, 1986; Bower et al., 1996; Field, 1999) agree that the evolution of ice crystals is dominated by depositional growth in the upper regions of the cloud, and is followed by aggregation as the crystals grow to larger sizes. Evidence for this sequence comes from direct examinations of the ice crystal shapes, and from the size distributions of the crystals. The total number of crystals changes little with altitude, while the size distributions gradually broaden with decreasing altitude. Evaporation, nucleation, riming, secondary ice generation and breakup of aggregates may also play a role, but appear to change the overall pattern only to a minor extent. A common feature of Ns is the 'bright band', the radar signature of melting ice crystals, with moderate rainfall rates below that level. This relative simplicity of the dominant processes in nimbostratus provides a good basis for generalized

descriptions of this class of clouds. Yet, current knowledge is insufficient to account for the specific characteristics of given events. It is also likely that details superimposed on the general patterns hold the key to many of the important aspects of how ice crystals are initiated and grow, and how these processes interact with the dynamics of the storms. Thus, our goal was to explore nimbostratus clouds with the improved detail made possible by the airborne cloud radar, and to thereby gain some additional insights into the relationship between cloud structure and ice crystal evolution.

We present in situ microphysical and airborne radar data for two nimbostratus cases. Both nimbostratus had cirrus generating cells on top. Ice trails from the generating cells were also evident in the near-horizontally stratified regions through most of the cloud depth. Negligibly small amounts of supercooled cloud were present. Both cases had well-defined melting layers and produced light rain at the ground. The main points we address in this paper are documentation of the evolution of the ice crystal spectra, and the relationship between these spectra and radar reflectivities. The results confirm earlier findings regarding the former and provide new information regarding the latter. In addition, the radar images provide ready visualization of the structure of these clouds.

## **2. Data sources**

The University of Wyoming King Air (KA) aircraft was equipped with cloud physics instruments and with a 95 GHz cloud radar. The measurements pertinent to this study are particle shapes and concentrations, aircraft position, horizontal and vertical winds, temperature, moisture, LWC, and radar reflectivity. A description of the Wyoming Cloud Radar can be found in Wolde and Vali (2001) or in Vali et al. (1998).

Data collected with the PMS (Particle Measurement Systems, Inc., Boulder, CO) FSSP, 1D-C, 2D-C, and 2D-P probes were used in analyzing the hydrometeor spectra. Data from the lowest 1-3 bins of the probes were disregarded due to undercounting. With this truncation of the data, the minimum sizes included in the data are: 25  $\mu\text{m}$  for the 1D-C probe, 100  $\mu\text{m}$  for the 2D-C probe and 800  $\mu\text{m}$  for the 2D-P probe. The particle size refers to the maximum crystal dimension along the flight path.

In order to select averaging periods that would yield size spectra of reasonable statistical stability, we required that the coefficient of variation,  $\beta$ , not exceed 0.4, i.e.

$$\beta = \frac{\sigma_D}{\bar{D}} < 0.4 \quad (1)$$

where  $\sigma_D$  is the standard deviation of the mean diameters calculated for 1-s segments within the time interval, and  $\bar{D}$  is the mean diameter for sizes greater than 100  $\mu\text{m}$  over the entire time interval. This criterion lead to averaging intervals of five seconds for the first case to be presented and to fifteen seconds for the second case.

Particle spectra for diameters greater than an assigned minimum size,  $D_m$ , were approximated with exponential distributions of the form

$$N = N_o e^{-\lambda D} \quad (2)$$

where  $N$  is particle concentration per unit size interval,  $\lambda$  is the slope parameter,  $N_o$  is the intercept parameter and  $D$  is crystal diameter. The values of  $\lambda$  and  $N_o$  for the spectra were computed using a least-squares linear fit to the logarithms of the concentration and diameter.

### 3. Case studies

#### 3.1. Case 1: Frontal Ns cloud over SE Wyoming

##### 3.1.1. General description

The Ns was observed on October 31, 1992, near Torrington in SE Wyoming. It formed along a Pacific front associated with an area of low surface pressure centered over Colorado. At mid-troposphere, a low pressure trough extended along a line connecting Idaho, Utah and New Mexico. Sequences of soundings near the study site show a cooling trend throughout the troposphere. Satellite images show relatively uniform cloud cover from eastern Wyoming to Nebraska.

The flight pattern employed in this case is shown Fig. 1. The staircase flight pattern consisted of alternate legs of slowly descending ( $4 - 8 \text{ m s}^{-1}$ ) and level segments. The level legs (labeled with numerals in Fig. 1) were flown in a SE to NW direction while the descent legs were flown in the reverse direction. The straight portions of the flight legs were about 20 km in length, and the turns extended about 5 km further at either end. The total duration of the descent from near 7 km to 1.5 km was nearly one hour, corresponding to an average descent rate of  $1.4 \text{ m s}^{-1}$ . The choice of the staircase pattern was made in order to obtain images from the airborne radar that are vertical cross-sections of the cloud, and to measure the fall velocities of ice particles from the vertically pointing Doppler data.

A composite of vertical-plane radar reflectivity fields (recorded during three different level flight segments) is shown in Fig. 2. The salient features evident in these images are the generating cells near 7 km, a more uniform and layered region between 6.5 and 2.4 km altitudes, and a melting layer (bright band) at 2.4 km. The magnitudes of the observed

reflectivities vary over two orders of magnitude (20 dB) with the maxima in the glaciated regions reaching 15 dBZ.

### *3.1.2. Vertical temperature and wind structure*

Vertical profiles of key parameters from the in situ sensors are shown in Fig. 3. Data points are 10-s averages except for the FSSP and liquid water plots where the points are 1-s averages. Apart from the lowest few hundred meters of the sounding, generally stable temperature stratification prevailed. The region immediately above the melting level (2.4 km) was the least stable and the layer between about 3.5 km and 4.2 km was the most stable. Saturation with respect to water was reached in two layers: 2.1–2.5 km and 3.5–4.2 km. The humidity was somewhat below saturation with respect to ice from about 2.4 to 3 km, i.e. directly above the 0°C level.

Winds below 4.2 km were from the NNW, varying in speed from 8 to 14 m s<sup>-1</sup>. A rapid backing of the wind between 4 and 5 km was accompanied by the lowest wind speeds (4 m s<sup>-1</sup>). Above 5 km winds continued to back, so that by 7 km the wind direction was from the SSW and speeds increased to over 10 m s<sup>-1</sup>. Vertical air velocities were generally small, ranging from -2 to +1 m s<sup>-1</sup>.

### *3.1.3. Hydrometeor profiles*

#### *3.1.3.1. Cloud droplets*

Regions containing cloud droplets were observed near cloud top and at two other levels. Near cloud top (7 km), a relatively large region (8 km in horizontal extent) of supercooled cloud drops was observed at a temperature of -29°C with droplet concentrations up to 40 cm<sup>-3</sup> and with liquid water contents of up to 0.03 g m<sup>-3</sup>. Dominantly positive, but weak vertical velocities (<1 m s<sup>-1</sup>) were recorded in this region. Supercooled drops were also

observed at temperatures close to  $-5.6^{\circ}\text{C}$  and  $-1.7^{\circ}\text{C}$ , in both cases associated with slight temperature inversions. At the  $-5.6^{\circ}\text{C}$  level, droplet concentrations of up to  $30\text{ cm}^{-3}$  and LWC up to  $0.25\text{ g m}^{-3}$  were recorded but the region was only about 1.5 km in extent. The most substantial region of cloud droplets was observed at a temperature of  $-1.7^{\circ}\text{C}$ . That region was nearly 20 km in extent and covered most of flight segment #6. At this level, droplet concentrations of up to  $160\text{ cm}^{-3}$  were measured but the LWC remained  $< 0.1\text{ g m}^{-3}$ .

### *3.1.3.2. Ice particle concentrations*

Between height of 3 km and 7 km ( $-29$  to  $-2^{\circ}\text{C}$ ), ice particle concentrations, when averaged over 1-km horizontal distances, showed no clear trend with altitude. Mean concentrations from the 2D-C probe varied in the range  $20\text{--}30\text{ L}^{-1}$ , and those from the 2D-P probe were in the range  $5\text{--}10\text{ L}^{-1}$ . Within the ranges cited, ice crystal concentrations exhibited variability on many scales, as demonstrated by the time sequences shown in Fig. 4; the flight distance corresponding to 10 s of time is about 1 km.

Maximum ice particle concentrations were recorded in the layer near  $-5^{\circ}\text{C}$  where needle crystals and their aggregates were detected (cf. Fig. 5). As mentioned in the foregoing section, some cloud droplets were also observed in this same layer.

Below 3 km altitude, the ice crystal concentrations decreased steadily and reached very low values ( $< 1\text{ L}^{-1}$ ) just above the melting level.

### *3.1.3.3. Particle types and spectra*

Samples of 2D images observed at various heights are shown in Fig. 5. Crystal types were dominantly irregular, except for a small number of plate crystals and the region of needle crystals. Some aggregation is evident even at temperatures of  $-22^{\circ}\text{C}$ . The relative proportion of aggregates and their sizes increase toward warmer temperatures.



Exponential functions defined by equation (2) were fitted to the observed size spectra for  $D > 100 \mu\text{m}$  over 15-s averages. Above  $-15^\circ\text{C}$  (4 km), the distributions fit the exponential function with a coefficient of determination  $R^2 > 0.92$ . Between  $-15$  to  $-2^\circ\text{C}$  (2.5 to 5.5 km)  $R^2$  values varied from 0.85 to 0.99. Cloud volumes having crystal sizes exceeding 6 mm tend to have an excess number of small crystals in comparison with the value from the exponential fit to the complete size range, and hence also have lower correlation coefficients.

The parameters  $\lambda$  and  $N_o$  from eq. (2) are shown in Figs. 6a and 6b as functions of altitude. As can be seen from the figures, the general trend is for both  $\lambda$  and  $N_o$  to increase with altitude. In a number of short altitude intervals,  $\lambda$  and  $N_o$  appear to increase with decreasing altitude but these are likely to have resulted from horizontal inhomogeneities. Fig. 6c shows a plot of  $\lambda$  versus  $N_o$  on logarithmic scales. It can be seen that there is high correlation between  $\lambda$  and  $N_o$  and that a power law provides the best description of the relationship between these two parameters.

The maximum crystal size increased with decreasing altitude throughout the range of observations until reaching the melting level (not shown). At the melting level (2.4 km), there was a sharp decrease in the maximum particle size, as expected, eventually resulting in raindrops of about 1 mm maximum size.

#### *3.1.4. Relationship between spectral parameters and radar reflectivity*

Figure 7a shows 15-s averages of ice particle concentrations measured by the 2D-C probe, the slope parameter  $\lambda$ , and the radar reflectivity  $Z_e$  at 120 meters from the flight level. For  $\lambda$ , only those points are plotted for which  $R^2 > 0.9$ , and where the coefficient of variation defined by equation 1,  $\beta < 0.3$ .

Restricting attention to the region above 3.5 km, the height below which evaporation began to exert a dominant influence, trends are near-constant concentrations, decreasing  $\lambda$  and increasing  $Z_e$  values with decreasing altitude. The approximate rates of change are, from subjectively fit linear variations:  $d\lambda/dz = 0.5 \text{ m}^{-2}$  and  $dZ_e/dz = -1.2 \text{ dBZ km}^{-1}$ . Superimposed on the overall pattern are some intervals of minor systematic changes. During flight leg #3 at 5.4 km ( $-15^\circ\text{C}$ ), a steady decrease in reflectivity was coupled with an increase in  $\lambda$  and a decrease in crystal concentrations from the southeast to the northwest. Similar systematic variations are found in the region just below 6 km and between 5.3 and 4.5 km. The opposite relationship,  $\lambda$  and concentration increasing together, can be noted just above and just below 4 km.

The rapid decrease in reflectivity (10 dBZ over roughly a km) below 3.5 km is in a region of evaporation, and is accompanied by decreases in both  $\lambda$  and in concentration, although the change in  $\lambda$  is relatively small.

Scatter plots of crystal concentrations, the slope parameter  $\lambda$ , and maximum particle sizes versus radar reflectivity are shown in fig. 7b, for the region above 3.5 km, and for the layer 2.4–3.5 km (dry layer). Particle concentrations are plotted on logarithmic scales to allow inclusion of the very low values in the dry layer. Above 3.5 km, the correlation between  $Z_e$  and particle concentrations is positive but not significant (the linear correlation coefficients between  $Z_e$  and particle concentrations measured by the 2D-C and 2D-P are 0.2 and 0.4, respectively). In contrast,  $Z_e$  and  $\lambda$  are negatively correlated with a linear correlation coefficient of -0.7, indicating that the increase in reflectivity with decreasing height in this altitude range was resulted from a broadening of the size distribution of crystals. This is also borne out by the positive trend between reflectivity and maximum

crystal size. Both facts are consistent with progressively greater degrees of aggregation as the crystals fall. In the dry layer, there are strong positive correlations between  $Z_e$  and particle concentrations ( $r=0.7$  for 2D-C and  $r = 0.65$  for 2D-P) as well as between  $Z_e$  and  $\lambda$  ( $r=0.6$ ). Similarly, maximum sizes decreased along with the other two parameters. This fact is consistent with the assumption that larger aggregates were breaking up, reducing the maximum size, and that the concentrations of smaller crystals decreased at the same time yielding lower values of  $\lambda$ .

### *3.2. Case 2: Ns over the Oregon Coast*

#### *3.2.1. General description*

Data were obtained from a late-summer Ns cloud near the Oregon coast on September 6, 1995. The surface weather map from three hours prior to the King Air flight indicates no major features near the study area. However, a mid-tropospheric trough extended southwards from a cut-off low off the coast of Washington and northern Oregon passed over the study area. Soundings near the study area show significant cooling between 350 and 750 mb, and rapid lowering of the tropopause during the occurrence of the NS. Satellite images (Fig. 8a) show that clouds covered much of western Oregon and Washington and the coastal waters. The small box in Fig 8a indicates the study area. It was located over the NOAA/ERL  $K_a$ -band radar; much of the mission was arranged to provide coincident radar data from the ground-based and the airborne radars.

The flight segment included in the analysis was about twenty minutes long and covered an area of about 6 by 6 km. During this period, the KA first made a spiral ascent from near 1.5 km to 4 km altitude and then a spiral descent (Fig. 8b). The following discussion

concentrates on the descending segment; deviations from uniform circles were made in order to improve the coincidence of sample volumes of the two radars.

The top and middle panels of Fig. 9 show a time sequence and a vertical-plane scan (RHI) of reflectivity recorded with NOAA K<sub>a</sub>-band radar. Maximum cloud height decreased with time as a generating cell drifted to the east. Fall streaks originating from the generating cell produced an echo structure with an east to west tilt. The melting layer is evident near 2.6 km altitude. The plane of the RHI scan is not far off the dominant wind direction. Using a mean velocity of 6 m s<sup>-1</sup> for the cloud motion (cf. Fig. 10), the location of the axis of the descent spiral is about 2.5 km to west of the origin of the polar reflectivity plot. Due to the tilt of the reflectivity field, each loop of the spiral moved from regions of stronger reflectivity on the east end of the loop to lower reflectivities on the west ends of the loops.

The bottom panel of Fig. 9 shows reflectivity data registered with the airborne radar (WCR) during the spiral descent. The conical surface scanned by the radar is projected onto a vertical cylindrical surface which is then mapped onto the plane of the paper for this figure. This projection leads to the oscillating appearance of the tilted bands of high reflectivity. The higher resolution of the WCR reveals even finer details than are visible in the data from the NOAA radar. Layers of 100 to 300 m thickness are evident from this image.

### *3.2.2. Vertical temperature and wind structure*

Vertical profiles for state variables and for particle concentrations are plotted in Fig. 10. The temperature lapse rate was stable (about -5.5°C/km) above the 0°C level. Just below that level, the lapse rate was slightly unstable (about -10°C/km) and there was a small

isothermal layer; both of these factors probably arising from the diabatic cooling associated with melting. Humidity below the 0°C level was below saturation with respect to water. Above the 0°C level, the humidity was (within the accuracy of the dew point instrument) above saturation with respect to ice up to about 3.5 km. No cloud droplets were observed at any level within the cloud, indicating that saturation with respect to water was not reached anywhere.

Winds throughout the cloud depth were from the southwest, with a change from 210°-220° at about 4 m s<sup>-1</sup> near the 0°C level to 250° and near 8 m s<sup>-1</sup> both below and above that level. The sharp decrease in wind speed at and below the 0°C level is similar to that observed in the October 31, 1992 case study.

Both positive (upward) and negative (downward) vertical air velocities are <1 m s<sup>-1</sup> for the 5-s averages shown in Fig. 10. Predominantly negative values are found in the first 300 m layer below the 0°C level, switching to positive values for the next 300 m or so near 2000 m. Other apparently systematic variations from slightly positive to slightly negative values are seen at other levels as well, perhaps related to changes in horizontal position.

### *3.2.3. Hydrometeor profiles*

#### *3.2.3.1. Hydrometeor concentrations*

Raindrop concentrations below the melting layer were < 1 L<sup>-1</sup>. Between the maximum altitude sampled and the 0°C level, there was a gradual decrease in hydrometeor concentrations in all size ranges. The smaller-scale patterns reflect a combination of horizontal and vertical variations within the cloud and are related to the shallow layers of reflectivity maxima observed by the radars (Fig. 9).

#### *3.2.3.2. Particle types and spectra*

Figures 11 show sample 2D images and size spectra recorded during the descent spiral. Each strip of particle images is positioned to correspond to the altitude shown on the right hand scale. As the images illustrate, ice particle types were aggregates of dendrites, single crystals with irregular shapes, planar crystals and capped columns.

Sample size distributions for 5-s averages (500 m horizontally and 20 m vertically) are shown in the middle column of Fig. 11, with lines indicating exponential fits to the data. For spectra extending to 6 mm in size, correlation coefficients to the exponential function are higher if the fit is restricted to  $D > 900 \mu\text{m}$ , however, even using  $D > 100 \mu\text{m}$  the correlation coefficients range from -0.99 to -0.85. Thus, the same threshold diameter of  $100 \mu\text{m}$  was used for all spectra. The slope parameter,  $\lambda$ , is shown as a function of altitude in the right hand panel of Fig 11. As expected, where large aggregates are present  $\lambda$  is smaller, while for single crystals with sizes extending only to about 3-4 mm,  $\lambda$  is larger. In all cases, the concentrations of small particles ( $< 100 \mu\text{m}$ ) are significantly higher than values indicated by the extrapolation of the exponential function to these sizes. The difference is greatest for spectra which extend to the largest sizes due to the shallower slopes of these distribution.

Cloud volumes containing mainly single crystals were consistently observed in the eastern halves of the loops while larger aggregates were observed in the western halves. This led to the oscillations in  $\lambda$  with altitude apparent in Fig. 11. As was pointed out in reference to Fig. 9, this periodic change arises from the intersection of the spiral flight path with the inclined layers of different reflectivities.

To help separate the vertical and horizontal variabilities of the spectra parameters, Fig. 12 shows vertical profiles of  $\lambda$  and  $N_0$  together with a plot of the east-west distance

between the aircraft and the NOAA-K<sub>a</sub> radar site. Changes in  $\lambda$  and  $N_o$  are seen to correlate strongly with position. The correlation is also evident when the data are stratified according to which quarter of the circular flight paths they originate from (right-hand panels on Fig. 12).

Figure 13 shows a scatter plot of  $\lambda$  versus  $N_o$  for both the ascent and descent spirals. As shown in the figure, there is a strong correlation ( $r = 0.94$ ) between these two parameters; the relationship is well described by a power-law expression.

#### *3.2.4. Relationship to the airborne radar data*

The vertical profiles of the ice crystal's concentrations and size distributions, and the observed reflectivities are displayed in Fig. 14. Radar reflectivities are from the nearest usable range gate (roughly 150 m distant and 135 m higher than the aircraft and the in situ probes) but are displayed at the altitude of the aircraft. This manner of displaying the data would be justified if horizontal variations within the 5-s (500 m) averaging intervals were clearly more important than vertical variations over the 135 m difference in location. In fact, the two scales of variations are comparable, as judged from the angle of tilt of  $10^\circ$  of the echo features in the RHI image shown in Fig. 9. This ambiguity renders comparisons between the in situ and the radar data meaningless on scales smaller than about 2.5 km in the horizontal and 0.7 km in the vertical (5 times the data resolution). Those scales are close to the diameter and the vertical spacing of the loops within the spiral flight tracks. Thus, it must be concluded that no significance can be attached to fine-scale features apparent in the relationship between radar reflectivity and in situ data. Nonetheless, it is worth noting that the magnitudes of the small-scale variations in all parameters are quite significant: roughly a factor of two to three (3-5 dB) in magnitude and with large local gradients.

The dominant trends shown in Fig 14 are decreases for all three parameters, relatively slow decreases from 4 to 3.2 km, and more rapid changes below that. The data shown in Fig. 14 cover the altitude range of aircraft sampling. At least for reflectivity, data are available over the entire cloud depth; this is shown in Fig 15 from the NOAA  $K_a$ -band radar and from the airborne radar. Averages derived from data shown in the top panel of Fig. 9 are given in Fig. 15 (left panel). The main features seen in these profiles are (i) the rapid increase in reflectivity with decreasing altitude from the top of the echo to near 5 km altitude, (ii) the nearly constant values between 5 km and the melting level at 2.6 km and (iii) the typical spike in reflectivity (bright band) just below the melting level. The slight decrease in reflectivity noted in Fig 14 is consistent with the profile from the ground based radar.

The right-hand panel in Fig. 15 shows reflectivity data from the airborne radar. Points are averages of five radar profiles (time interval of 0.6 s corresponding to ~60 m flight path) once every 300 m of flight path. The rapid increase in  $Z_e$  with decreasing height at the top of the sounding, and the maximum in reflectivity near 5 km are consistent with data from the ground-based radar. Absolute values of the reflectivities are comparable within the accuracies of the measurements. The region of increasing reflectivities from cloud top down to about 5 km includes the generating cell and the upper parts of the precipitation trails (cf. Fig. 9). Aircraft data obtained in this altitude range (prior to the spiral descent and somewhat further to the southwest) indicated that the increase in  $Z_e$  is caused by an increase in crystal sizes by deposition and aggregation, without significant changes in total concentration. Below 5 km altitude, the airborne radar data indicate a more rapid decrease



in reflectivity than the ground based radar. The likely cause of this difference is the averaging of the ground-based radar data over a period of time.

The scatter of WCR reflectivity values at given altitudes in Fig. 15 is about 5 dBZ (factor 6); this value includes both the spatial (4-6 km) and temporal (8-12 min) variabilities. This variability is in agreement with the changes seen in the NOAA radar data (Fig. 9) on comparable scales.

## **4. Discussion and Summary**

### *4.1. Cloud characteristics*

The two case studies described here occurred at two widely separated locations and at different times of the year. In spite of that, there are substantial similarities between the two storms. In both cases, ice crystals originated from well defined generating cells at cloud top, and neither cloud had significant supercooled liquid regions. The heights of the generating cells were similar (near 7 km) and so was the altitude of the melting level (2.4 vs. 2.6 km). The concentrations of ice crystals in the 2D-C size range were also similar. The concentration of larger crystals and aggregates detected by the 2D-P probe was about five times higher in the October 31, 1992 case than in the September 6, 1995 case, yet maximum radar reflectivities were about 10 dBZ in both cases.

### *4.2. Particle spectra*

In accordance with earlier studies, it was found that ice particle spectra (above some minimum size) are adequately described by exponential relationships in both data sets. Since there is no strong *a priori* reason for that relationship to apply, the generality of the exponential law may be taken as an indication that the shapes of the size distributions are relatively insensitive to the specific conditions influencing growth by deposition and

aggregation. The minimum sizes for exponential spectra ranged from 500 to 1000  $\mu\text{m}$  in earlier studies. In this study, using a minimum size of 900  $\mu\text{m}$  yielded better fits, particularly in segments where the spectra extended to sizes  $>4$  mm, but even with a minimum size of 100  $\mu\text{m}$  the fits were very good (correlation coefficients  $>0.9$ ). It appears that the larger the maximum sizes are in a spectrum, the more pronounced are the excess concentrations above the exponential function at small sizes.

The observed presence of small crystals essentially everywhere within the Ns is not easily reconciled with the common assumption that all crystals originate near the upper cloud boundaries (in the generating cells). With that scenario, one would have to also assume that only a small fraction of the total number of crystals participate in growth by aggregation. Alternatively, the small crystals may be products of fragmentation, or of some other secondary process, but then that process must be effective over a wide range of conditions to account for the observations.

#### *4.3. Evolution*

Generating cells are common features of Ns (e.g., Houze, 1993; pp. 205-206) and are either a consequence, or, at least, are associated with, strong wind shear. The combination of small fall velocities and wind shear leads to near horizontal layers within the clouds below the generating cells leading to the impression of horizontal homogeneity and nearly steady state vertical structure. However, as our radar data shows, the ice trails remain distinguishable throughout the depths of the clouds, and even in the rain below the melting layer. This is a consequence of the horizontal spacing of the generating cells at approximately 2 to 4 times the sizes of the cells.

Since the flight pattern in neither of our cases was designed to follow a set of particles emanating from a generating cell, and since the cells were relatively small (2-3 km cores) we cannot rigorously document the evolution of given sets of ice crystals. Even so, some comments on the growth processes can be made. As Lo and Passarelli (1982) have argued, aggregation results in a simultaneous decrease in both  $\lambda$  and  $N_o$  in the exponential function fitted to the size distributions. Growth by deposition only would tend to leave  $\lambda$  unchanged and increase  $N_o$ . In both data sets described here, a strong correlation was found between  $\lambda$  and  $N_o$  (see Figs. 6 and 13) which is consistent with the pattern expected for aggregation, though the scatter of points makes it unreliable to examine sequences over small intervals of the variables. For the 1992 case, 2D images recorded near cloud top clearly show no aggregates, yet  $\lambda$  and  $N_o$  vary in parallel. It is not clear how much of this is due to local variations or as evidence against the generality of the signature attributed to aggregation. In the other case we described, the flight pattern didn't extend to near cloud top, and aggregation is indeed evident in all of the 2D images.

In the 1992 case, aggregates started to be observed around  $-21^\circ\text{C}$  and became more frequent within 2 km of the melting level. The dominance of aggregation near the melting level agrees with the observations of Stewart et al. (1984), and of others. In both cases, the presence of a dry layer above the melting layer caused evaporation of particles and a reduction of maximum crystal size as well as concentrations.

Table 1 lists the parameters of the power law functions fitted to the  $\lambda$  vs.  $N_o$  data for the two cases described in this paper, and some values reported in the literature. There are evident differences in these parameters for the different cloud cases. This is not surprising in view of differences in temperatures, and other factors that might influence the numbers

of crystals nucleated, and other characteristics like vertical velocities, the presence of supercooled cloud etc., that influence the evolution of the size distributions. There is yet no basis on which to make general interpretations of the parameters listed in Table 1, or to relate variations in these parameters to other cloud characteristics, so the main utility of these data at the moment is to bracket the range of values that may be expected.

#### *4.4. Radar reflectivity*

The reflectivities measured by the airborne radar are negatively correlated with the size distribution parameters. This is consistent with the expectation that increases in crystal size enhance the reflectivity and larger sizes correspond to smaller values of  $\lambda$  and  $N_0$ . However, since larger aggregates often develop in regions of higher overall crystal concentrations, it is not possible to generalize the separate effects of crystal size and of crystal concentrations on  $Z_e$ . Only in the dry layer of the 1992 case is there a strong positive correlation between reflectivity and particle concentration.

### **5. Conclusions**

Observational evidence gathered in two case studies with a combination of in situ and remote sensing instruments support the generally accepted picture of nimbostratus structure. These data further help to delineate the range of variability of parameters like ice concentration and spectral shape. Importantly, the data point to the need to take into account the inhomogeneities that result from the cellular character of the ice generating regions at cloud top. Since so much of the nimbostratus structure derives directly from the generating cells at the top of the storm, it is clear that a better understanding of the dynamic and microphysics mechanisms at work in the generating cells is needed.

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Table 1. Parameters for the relationship  $N_o = a \left( \frac{\lambda}{1000} \right)^b$ . Values are for  $\lambda$  in [ $m^{-1}$ ],  $N_o$  in [ $m^{-3} \mu m$ ]. The listed  $a$  values are the  $N_o$  values at  $\lambda$  of  $1000 m^{-1}$ . The minimum size,  $D_{min}$  [ $\mu m$ ] used to compute  $\lambda$  and  $N_o$  and the linear correlation coefficient  $r$  are also indicated when available.

Case/author	$D_{min}$	$a$	$b$	$r$	$\lambda$ range
September 06, 1995 (Ns over Oregon)	100	3.3	2.74	0.93	367-2962
October 31, 1992 (Ns over Wyoming)	100	1.88	2.32	0.91	1470-9340
October 31, 1992 (Ns over Wyoming)	900	0.68	2.41	0.91	1236-3672
Altostratus cloud (Field 1999)	800	1.54	2.89	0.7	1928-5818
Winter cyclonic storm (Lo and Passarelli 1982)	300	1.07	1.85		1000-8000
Rainband (Gordon and Marwitz 1986)	500	0.13	2.26	0.9	1120-10760