

Aircraft Microphysical and Surface-Based Radar Observations of Summertime Arctic Clouds

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ABSTRACT

Updated analyses of in situ microphysical properties of three Arctic cloud systems sampled by aircraft in July 1998 during the Surface Heat Budget of the Arctic Ocean (SHEBA)/First International Satellite Cloud Climatology Project (ISCCP) Regional Experiment–Arctic Clouds Experiment (FIRE–ACE) are examined in detail and compared with surface-based millimeter Doppler radar. A fourth case is given a cursory examination. The clouds were at 78°N over a melting ice surface, in distinctly different yet typical synoptic conditions. The cases comprise a midlevel all-ice cloud on 8 July; a deep, weakly forced, layered, mixed-phase stratus cloud system with pockets of drizzle, large dendrites, rimed ice and aggregates on 18 July; and a deep, mixed-phase cloud system with embedded convection on 28 July followed by an all-water boundary layer cloud on 29 July. The new observations include measured ice water content exceeding 2 g m^{-3} on 18 and 28 July and 3-cm snowflakes and 5-mm graupel particles on 28 July, unexpected in clouds close to the North Pole. Radar–aircraft agreement in reflectivity and derived microphysical parameters was reasonably good for the all-water and all-ice cases. In contrast, agreement in radar–aircraft reflectivity and derived parameters was generally inconsistent and sometimes poor for the two mixed-phase cases. The inconsistent agreement in radar–aircraft retrievals may be a result of large uncertainties in both instrument platforms and the algorithms used to retrieve derived parameters. The data also suggest that (single-wavelength) radar alone may not be capable of accurately retrieving the microphysical effects of cloud drops and drizzle in mixed-phase clouds, especially radiative properties such as extinction, albedo, and optical depth. However, more research is required before this generalization can be considered conclusive.

1. Introduction

This paper presents new observations on mixed-phase and all-ice Arctic clouds, achieved by applying analysis of two-dimensional particle probe imagery collected by the National Center for Atmospheric Research (NCAR) C-130 research aircraft. The data were collected as part of the Surface Heat Budget of the Arctic (SHEBA)/First International Satellite Cloud Climatology Project (ISCCP) Regional Experiment–Arctic Clouds Experiment (FIRE–ACE; Curry et al. 2000). The in situ measurements are compared with microphysical retrievals from a ship-based millimeter Doppler radar,

with surface observations and radiosondes also providing context to the aircraft. The observations come from July of 1998 and were taken at $\sim 78^\circ\text{N}$ in the Beaufort Sea, within 1200 km from the North Pole, above a melting ice surface. As such, they contrast with clouds sampled over a coastal open ocean during the Mixed-Phase Arctic Cloud Experiment (MPACE; Verlinde et al. 2007) and with the weakly forced clouds above an ice surface discussed in Zuidema et al. (2005) and Pinto (1998). This paper augments the Lawson et al. (2001) and Khvorostyanov et al. (2001) case descriptions, for which information on two-dimensional particle imagery was not yet available, as well as the further analysis of Morrison et al. (2009, manuscript submitted to *J. Atmos. Sci.*, hereafter MZM).

Motivation for this work stems from the need to better characterize Arctic clouds, and mixed-phase clouds in particular, toward improving our understanding of the

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cloud processes that influence Arctic climate and Arctic climate change. At a practical level, analysis of aircraft and surface-based radar data is needed to improve the retrievals used for space-based cloud radars and their applications to Arctic research (e.g., L'Ecuyer et al. 2008; Kay et al. 2008) and to improve the cloud microphysical parameterizations applied within Arctic cloud simulations (e.g., MZM).

The synoptic pattern associated with each of the three cases is shown in Fig. 1. On 8 July a broad anticyclone known as the Beaufort High had its maximum sea level pressure (SLP) located to the south of the SHEBA site (Fig. 1a). By 18 July the anticyclone had moved to the northeast of the ship (Fig. 1b). Both days exemplify more suppressed conditions. The SLP diminished over the ship after the 18th (Fig. 1d), with a cyclonic front passing over the ship on 28 July in response to an SLP minimum located near the North Pole (Fig. 1c). Although the SLP spatial distributions differed significantly among the three cases, they were all characteristic of the western Arctic in July, when large variations in SLP are the norm (Wylie 2001). Typically in July a large-scale climatological anticyclone resides over the Beaufort Sea (between 120° and 150°W, north of east Alaska and west of the Canadian Archipelago, best captured here in Fig. 1a) in July, encouraging cyclone movement from the southwest to the northeast (Maslanik et al. 2001).

The cloud cases documented here help provide insight into the cloud changes that accompany recent changes in Arctic climate. The July 1998 circulation pattern was similar to that documented for 2007, when a record minimum in the sea ice extent coincided with reduced summer cloudiness (Kay et al. 2008; Stroeve et al. 2008; Perovich et al. 2008). Similarly, during the summer of 1998 an above-normal high-pressure cell resided over the eastern Beaufort Sea and the mean cloud cover decreased from its May value (Maslanik et al. 2001). A record reduction for its time was documented for the July 1998 sea ice cover of the Beaufort and Chukchi Seas (Maslanik et al. 1999). The melt pond fraction increased almost linearly from about 0.15 on 1 July 1998 to 0.40 by the end of July, with an areal-mean surface albedo decreasing from ~0.5 on 1 July to ~0.4 by 27 July (Curry et al. 2001; Tschudi et al. 2001). Thus, the clouds sampled during the July cases documented here occurred over a melting ice surface, which is steadily becoming more the Arctic norm.

The reduced cloudiness and above-normal Beaufort SLP from the summers of 1998 and 2007 may reflect the high degree of natural variability possible in the Arctic (e.g., Ogi and Wallace 2007; Zuidema and Joyce 2008), with the 8 July and 18 July cases representing cloud structures associated with such large-scale suppressed conditions. In contrast, other indicators point to in-

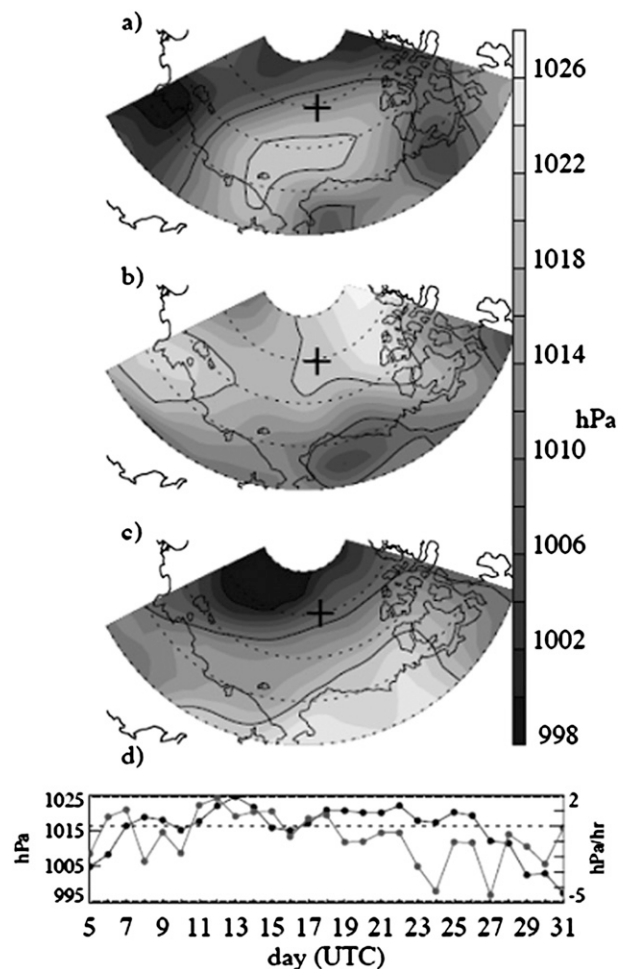


FIG. 1. NCEP-NCAR reanalysis sea level pressure for (a) 8, (b) 18, and (c) 28 July. (d) SHEBA site sea level pressure (black line) and 700-hPa vertical velocity (gray line), with positive values indicating subsidence. Contour lines indicate sea level pressure values of 1000, 1010, and 1020 hPa, and the plus sign indicates the location of the SHEBA/FIRE-ACE ship.

creasing cyclonic activity in the recent past (Zhang et al. 2004) and in future projections (Solomon et al. 2007). Summertime cyclones in the Beaufort/Chukchi Seas with oceanic origins are an important pathway for moisture transport to the polar cap, implying that total moisture transport into the Arctic is also increasing (Sorteberg and Walsh 2008; Bhatt et al. 2008). Arctic cyclones are more numerous and long-lasting during the summer than the winter, if typically less intense (Zhang et al. 2004; Sorteberg and Walsh 2008). The ultimate demise of the cyclones in the central Arctic generates the central Arctic summer SLP minimum (Serreze and Barry 1988; Reed and Kunkel 1960), with ramifications for the ice motion (Ogi and Wallace 2007; Tremblay and Mysak 1998). The 28 July case provides one window into the

TABLE 1. Aircraft instrumentation.

Instrument	Parameter	Range (μm)	Resolution (μm)	Reference
FSSP*-100	Drop, crystal size	2–47	—	Knollenberg (1981)
2D-C OAP** cloud probe	Shape, size	125–1060	25	Strapp et al. (2001)
2D-P OAP** precipitation probe	Shape, size	200–6400	200	Lawson et al. (2006)
Cloud particle imager	Phase, shape, size	5–2000	—	Lawson et al. (2001)
King hot-wire probe	Liquid water content	0.05–3.0 g m ⁻³	—	King et al. (1978)
Rosemount icing detector	Phase	—	—	Mazin et al. (2001)

* Forward scattering spectrometer probe.

** Optical array probe.

cloud processes and microphysics accompanying such a high-latitude Arctic cyclone.

While the 28 July case has not been previously discussed in the literature, preliminary observational findings on the 8 July and 18 July cases are shown in Khvorostyanov et al. (2001) and Lawson et al. (2001), respectively. The analysis of the all-ice 8 July case by Khvorostyanov et al. (2001) relied solely on measurements on the smaller particle sizes and included a finding that radar- and aircraft-derived ice microphysical properties compared poorly. We reevaluate this finding, which is important because this is the only all-ice SHEBA/FIRE-ACE cirrus cloud reported in the literature and it should in principle be best suited for a radar retrieval of ice cloud properties. The deep stratus cloud of 18 July contained regions of water, ice, and mixed-phase conditions. Lawson et al. (2001) previously discussed one cloud ascent, reporting on two regions with high ice particle concentrations. We reevaluate these regions with the addition of the aircraft 2D-C (two-dimensional imaging probe with 25- μm resolution for cloud particles) and 2D-P (two-dimensional imaging probe with 200- μm resolution for precipitation particles) data, cloud radar reflectivities, and new insights on ice crystal shattering on probe inlets, as well as aircraft data from the descent (Knollenberg 1981). The descent included constant altitude legs that can be applied to better discriminate temperature-dependent processes and to identify the layer structure of the cloud system. Comparisons between in situ observations and radar retrievals provide insight into the sampling of each platform.

2. Instrumentation, data processing, and method

a. NCAR C-130 in situ sensors

The capabilities of the NCAR C-130 and instrumentation on the research aircraft are described by Curry et al. (2000) and Lawson et al. (2001); microphysical instruments used for the data analysis presented here are listed in Table 1. Of these, the cloud particle imager (CPI) is noteworthy because it is useful for phase discrimination

within mixed-phase clouds. The high image resolution and 256 gray levels allow spherical particles to be distinguished from nonspherical particles if the focus is good and the particles are larger than about 30 μm in diameter. Because ice particles will typically grow to recognizable nonspherical shapes in less than a minute in a mixed-phase cloud, the separation by shape allows the separation of water drops from ice particles in mixed-phase clouds. A focus algorithm was used to automatically reject out-of-focus images. In-focus images were then classified by another software algorithm measuring the sphericity of the image. Several hundreds of particles were also classified by eye to verify the accuracy of the automated algorithm. The agreement between the automated and manual techniques was very good for images larger than about 30 μm in diameter. In regions where classification of images smaller than about 30 μm was essential, such as in regions with high ice concentration, the particles were manually classified. Particles smaller than about 20 μm that could not be confidently classified were considered to be water drops. Recent insights regarding the mechanisms of ice particles shattering on probe inlets now lead us to suspect that some of the small, quasi-spherical ice particles are artifacts resulting from shattering of large ice. It is also suspected that data from the forward scattering spectrometer probe (FSSP) is contaminated as a result of ice particle shattering. Based on recent studies reported elsewhere, in some regimes investigated here we argue that the effects of shattering are significant, and in other regimes the effects are minimal. This information is incorporated in a qualitative way into discussions included in this paper.

Significant improvements to data processing have occurred since the preliminary results presented in Lawson et al. (2001). One is the inclusion of the 2D-C and 2D-P probe measurements. Another improvement is the use of a new, algorithm for computing ice water content (IWC) from the 2D and CPI images (Baker and Lawson 2006). The CPI algorithm uses particle length, width, area, and perimeter. IWC from the 2D data was processed using particle area. Details are described by

Baker and Lawson (2006), who show that the new algorithms produce a 50% reduction in root-mean-square error when applied to the Mitchell et al. (1990) dataset of ice crystal images collected on glass slides, compared to length-to-mass algorithms from Mitchell et al. (1990). The new algorithm (Lawson and Baker 2006) shows that differences in IWC in SHEBA/FIRE-ACE clouds of up to a factor of 2 can result when using the Baker and Lawson (2006) technique, when compared against the Brown and Francis (1995) and Mitchell et al. (1990) techniques. The characterization of cloud drops was also improved. FSSP liquid water content (LWC) was erroneously overestimated by factors varying from 1.27 to 2.25 (Lawson et al. 2001). These were corrected by multiplying the FSSP particle size distribution (PSD) bin centers and bin widths by the fourth root of the ratio of FSSP to LWC from a King probe (also applied within Zuidema et al. 2005).

The FSSP, CPI, 2D-C, and 2D-P particle data were combined to form complete PSDs (see example in Fig. 2), with “breakpoints” establishing the size range for each probe’s contribution to the combined PSD. The breakpoints can vary from 20 to 40 μm between FSSP and CPI and from 150 to 500 μm between CPI and 2D-C; and they are fixed at 1 mm between 2D-C and 2D-P. 2D-C data were not used for sizes $<100 \mu\text{m}$, based on findings of poor sampling at that size range (Strapp et al. 2001; Lawson et al. 2006). The combined PSD was then used to derive parameters such as total particle concentration, volume extinction coefficient, liquid and ice water content, effective particle size, and radar reflectivity. The volume extinction coefficient was computed as twice the sum of the projected particle areas. The effective radius was computed from dividing particle mass [using the Baker and Lawson (2006) technique] by particle projected area. The reflectivity calculation assumed the density–size relationship of Brown and Francis (1995) for the ice particles. This has been found to apply well to all-ice clouds by Hogan et al. (2006) and Matrosov et al. (2002). In the mixed-phase clouds, FSSP data were classified as water drops, whereas CPI images were classified as water or ice based on their degree of sphericity.

b. Ship-based cloud radar

The cloud radar was located on the SHEBA/FIRE-ACE ship, which was frozen into the Beaufort Sea. The radar operates at 35 GHz (8.66-mm wavelength) with a 0.5° beamwidth and 45-m vertical resolution. Data acquisition reflects a cycling through four separate modes, each reflecting individual tradeoffs (Moran et al. 1998). Its sensitivity is approximately -46 dBZ at 5 km without attenuation (Moran et al. 1998), with signal saturation occurring at about 20 dBZ . The basic mea-

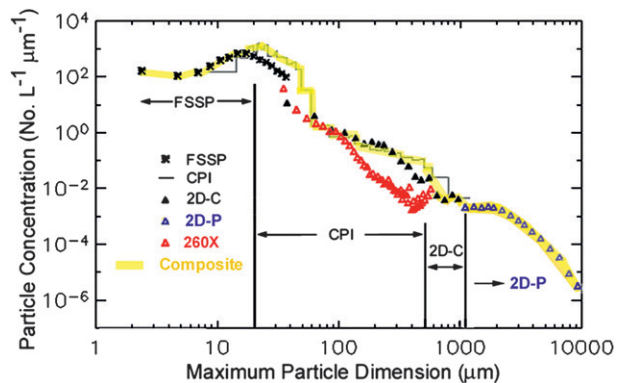


FIG. 2. Example of size distributions from particle probes during the time period from 2310:00–2310:30 UTC on 28 July showing how the PSDs are combined into a composite PSD. Breakpoints between probes are selected depending on relative concentrations. Breakpoints can vary from 20 to 40 μm between FSSP and CPI and from 150 to 500 μm between CPI and 2D-C; they are fixed at 1 mm between 2D-C and 2D-P. PMS 260X PSD, which only provides one-dimensional sizing, is never used and is shown here to emphasize the difference between it and other probe PSDs.

surement of the radar is the power of the backscattered energy in decibels (dB). This returned power is then converted to a reflectivity (dBZ) value assuming Rayleigh scattering from spherical water drops. The Rayleigh scattering assumption degrades at 35 GHz for particles larger than about 1 mm and Mie scattering becomes appropriate. However, Mie scattering approximations for 35 GHz are complex and unproven and therefore are not included in our retrievals. Most of the radar data presented here were averaged over a 1-min time period. Retrievals of ice microphysical parameters were done following Zuidema et al. (2005), which in turn relied on modified versions of relationships presented in Matrosov et al. (2002, 2003). These retrievals account for the differences in the complex dielectric constant and density between the water drops assumed by the operational radar processing code and the actual ice particles of the clouds. Mean particle size (from an assumed exponential ice particle size distribution) is related to the mean reflectivity-weighted Doppler radar velocity averaged over a 20-min time interval to remove turbulent air motion effects. The IWC algorithm assumes the Brown and Francis (1995) ice particle bulk density–mean particle size relationship. The result is combined with a mass–area size relationship slightly modified from that of Heymsfield et al. (2002) to retrieve the volume extinction coefficient. While the radar retrievals were initially developed for all-ice clouds, they were extended with apparent success using 260X probe data to the ice portion of a long-lived, weakly forced, mixed-phase case by Zuidema et al.

(2005). These retrievals receive further evaluation in the present study.

Microphysical values derived from the radar data were compared to those from in situ aircraft PSDs, gathered during radar overpasses within 4–5-km horizontal distance from the ship. Another methodology was employed in an attempt to increase the number of comparative data points and thereby improve statistics. Radar reflectivity values within a time span of typically 2 h are compared with aircraft values in a similar altitude range within the same time frame. In this comparison aircraft data are collected within approximately 80 km of the ship. Aircraft overflights were typically done along the wind in order to optimize radar and aircraft sampling of the same air mass. Note that an advective wind speed of 5 m s^{-1} corresponds to $\sim 35 \text{ km}$ within 2 h, so that the aircraft cloud sampling typically spans a larger cloud area than the radar samples within 2 h.

3. 28 July 1998 (78.60°N, 164.08°W): Mixed-phase cloud with embedded convection

a. Synoptics

The 28 July case study, the penultimate research flight of SHEBA/FIRE-ACE, is unique within the dataset, and perhaps among all airborne investigations conducted north of the Arctic Circle, because of convection that produced 5-mm graupel particles and 1- to 3-cm snowflakes. The synoptic analysis of Wylie (2001) tracked a low pressure system moving eastward along $\sim 75^\circ\text{N}$, with a large front passing the ship on 28 July and the lowest sea level pressure of the SHEBA/FIRE-ACE time period measured at the ship on 29 July. Figure 1c is consistent with a strong southerly flow bringing moisture to the region. During the first half of 28 July above-freezing temperatures prevailed in the boundary layer up to approximately 1.8 km. After midday an upper-level cloud descended, precipitating snow at the end of the day, with a surface fog developing on 29 July.

The research aircraft sampled the cloud after the 28 July frontal passage, between about 2230 and 2330 UTC. The cloud radar showed convective cells with tops reaching 8 to 9 km and precipitation reaching the surface after about 2315 UTC (Fig. 3). A forward-looking camera mounted on the C-130 provided video that documented embedded convection, cloud remnants, and areas of clear air (middle panels of Fig. 3).

b. Microphysics of embedded convective turrets

The C-130 descended from 6150 m (-27°C) at 2234 UTC to 1677 m ($+2^\circ\text{C}$) at 2254 UTC, then as-

cended to 6400 m at 2315 UTC, apparent from the superimposed aircraft track in the top panel of Fig. 3. In situ measurements show that the C-130 penetrated convective turrets during both the descent and ascent, observing small concentrations of supercooled liquid water (typically 0.01 to 0.1 g m^{-3}) in all the turrets, including at the maximum altitude with a temperature of -27°C . Two of the turrets (sampled during 2240:25–2240:52 UTC on descent, and 2310:00–2311:00 UTC on ascent) occurred at temperatures between -17° and -13°C and contained peak updraft velocities of 2 m s^{-1} , a peak liquid water content of 0.2 g m^{-3} but typically lower values (e.g., Fig. 4, bottom panel), and 2- to 5-mm dendrites (Fig. 4). The second turret was sampled soon before precipitation began reaching the surface of the SHEBA site. The composite PSD (Fig. 4) from the turret sampled during the ascent (2240:25–2240:52 UTC) corresponds to the highest average IWC observed on 28 July (2.02 g m^{-3}). The high IWC was composed of relatively high (200 to 400 L^{-1}) concentrations of large single dendrites up to 5 mm and aggregates of dendrites exceeding 1.5 cm, as measured by the 2D-P probe.

Further down near cloud base (2250 to 2253 UTC, ~ 2.2 -km altitude, 0°C), 1–3-cm snowflakes were observed (Fig. 3, bottom right). The air cooled with the frontal passage, with snowflakes within a deep, somewhat isothermal layer spanning 0°C extending from the surface to about 750 hPa ($\sim 2.5 \text{ km}$), or about 450 m above the cloud base height, as seen from the sounding at 2322 UTC (Fig. 5). Further above, from about 560 to 460 hPa (i.e., -13° to -25°C), is a weakly convectively unstable layer; the convective available potential energy (CAPE) was about 57 J kg^{-1} within the 1900 UTC sounding but was depleted by the time of the 2300 UTC sounding. The C-130 also observed the isothermal layer during its descent, with a temperature range of about -2°C at 2.5 km to $+2^\circ\text{C}$ at $\sim 1.7 \text{ km}$.

Lawson et al. (1998a) observed similar conditions of weak midlevel instability associated with very large snowflakes during the second Canadian Atlantic Storms Project (CASP II). During CASP II, research flights were conducted over the North Atlantic open ocean near Newfoundland, Canada. In both the CASP II case and the 28 July SHEBA/FIRE-ACE case, the generation of very large snowflakes was associated with the passage of a synoptic cyclone, with both events taking place in the warm sector of the cyclone. This case shows that Arctic cloud systems north of the Bering Strait and close to the North Pole are also capable of generating graupel particles, large snowflakes, and high IWC. It should be noted, however, that the summertime Arctic frontal passage appears to be less intense than that documented for the winter case near Newfoundland. This

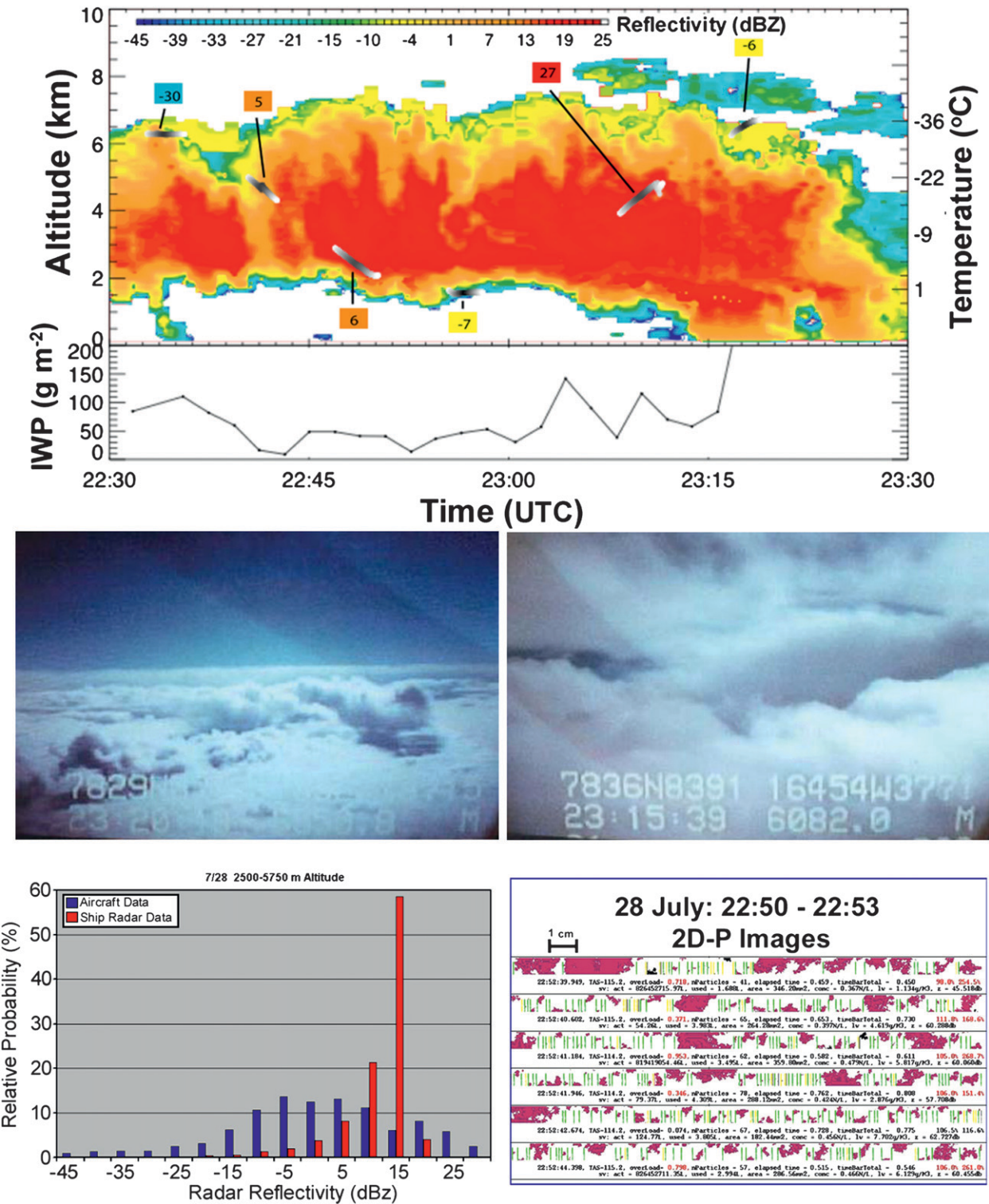


FIG. 3. (top) 28 July cloud radar reflectivities (10-s data), with dashes showing altitude and reflectivities calculated from combined PSDs where aircraft overflow the ship. Microwave-derived liquid water paths are denoted below the cloud radar reflectivities. (middle) Photographs from the aircraft nose video camera taken at (left) 2320 and (right) 2315 UTC. (bottom left) Histograms of radar and aircraft-derived reflectivities sampled from 2230 to 2330 UTC and between 2500- and 5750-m altitude when the aircraft was in clouds (determined when the FSSP or 2D-C measured more than 1 L^{-1}). (bottom right) Examples of 2D-P images showing 1- to 3-cm snowflakes observed near cloud base (at about 0°C).

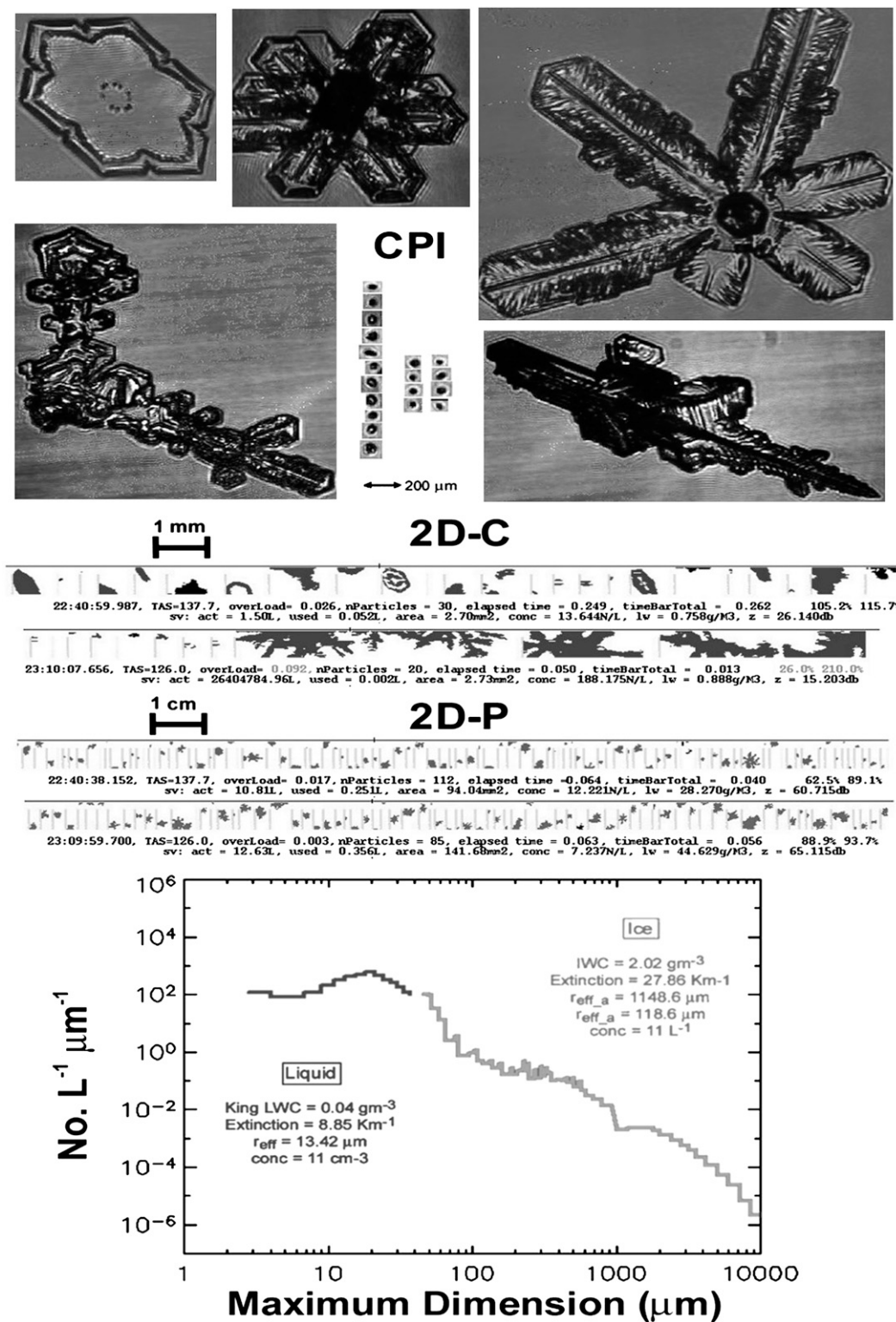


FIG. 4. Example of CPI, 2D-C, and 2D-P images observed in convective turrets at -17° to -13°C (2240 and 2310 UTC 28 July), with water (black) and ice (gray) PSDs shown below images at -13°C (4.5 km). Mean microphysical parameters derived from the water and ice PSDs at 2240 UTC are shown as insets.

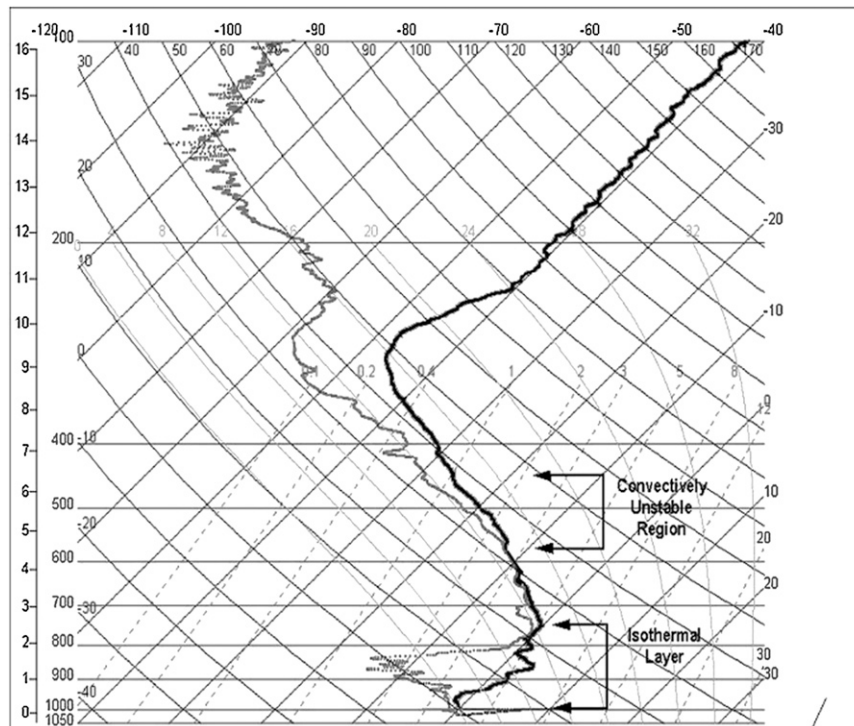


FIG. 5. Skew T -log p sounding taken at the SHEBA ship at 2322 UTC on 28 July, showing the elevated region of convectively unstable air and a deep isothermal layer near 0°C , with temperature and dewpoint temperature shown as black and gray lines, respectively.

is consistent with cyclone climatology (e.g., Sorteberg and Walsh 2008; Zhang et al. 2004) and with the differences in the underlying surfaces: whereas near Newfoundland the Gulf Stream maintained (nonfreezing) near- 0°C temperatures and provided moisture and warmth to the colder atmosphere, in the 28 July FIRE-ACE case the underlying melting sea ice surface helped stabilize the overlying atmosphere.

c. Radar-aircraft comparisons

There is agreement to within 10 dBZ between radar and aircraft-derived reflectivity values (see Fig. 3, top) when the aircraft passed within 5 km of the ship on 28 July. A comparison between two larger samples (Fig. 3, bottom left) does not show equally good agreement. It is of interest, however, that few of the aircraft-calculated reflectivities exceeded the cloud radar's dynamic range maximum of about 25 dBZ. This implies that although the radar reflectivity values in Fig. 3 are high, they are probably not too affected by saturation of the radar. The high radar reflectivity values may also be depressed through Mie scattering, which is not taken into account within the aircraft-calculated reflectivity value.

Figure 6 shows radar- and aircraft-derived microphysical parameters (reflectivity, volume extinction co-

efficient, ice water content, and mean particle size) for the times of the aircraft overpasses. Aircraft data were averaged for 30 s on each side of the overpass. While both radar and in situ reflectivity values generally follow a similar dependence on altitude within the cloud, there is a significant (about 10 dBZ) average difference in the absolute values. One explanation for the points in Fig. 6 where the aircraft value is less than the radar reflectivity is an underestimate of the particle density in the aircraft algorithm. This cloud contained a deep ice-supersaturated embedded convection in the presence of some liquid water. Such conditions favor diffusional growth and riming, both of which favor the formation of high-density ice. In such cases, the Brown and Francis (1995) density-to-particle size relationship used to calculate the reflectivity from the aircraft PSD may underestimate the actual particle density, as noted previously for a mixed-phase case by Hogan et al. (2006). The Brown-Francis relationship was originally developed for lower-density aggregates. Another explanation for the differences in retrievals is the spatial variability observed in both the aircraft and radar observations. This is apparent by noticing that the radar- and aircraft-derived reflectivity values sometimes differ in Figs. 3 and 6; this results from the 60-s averaging in Fig. 6 compared with

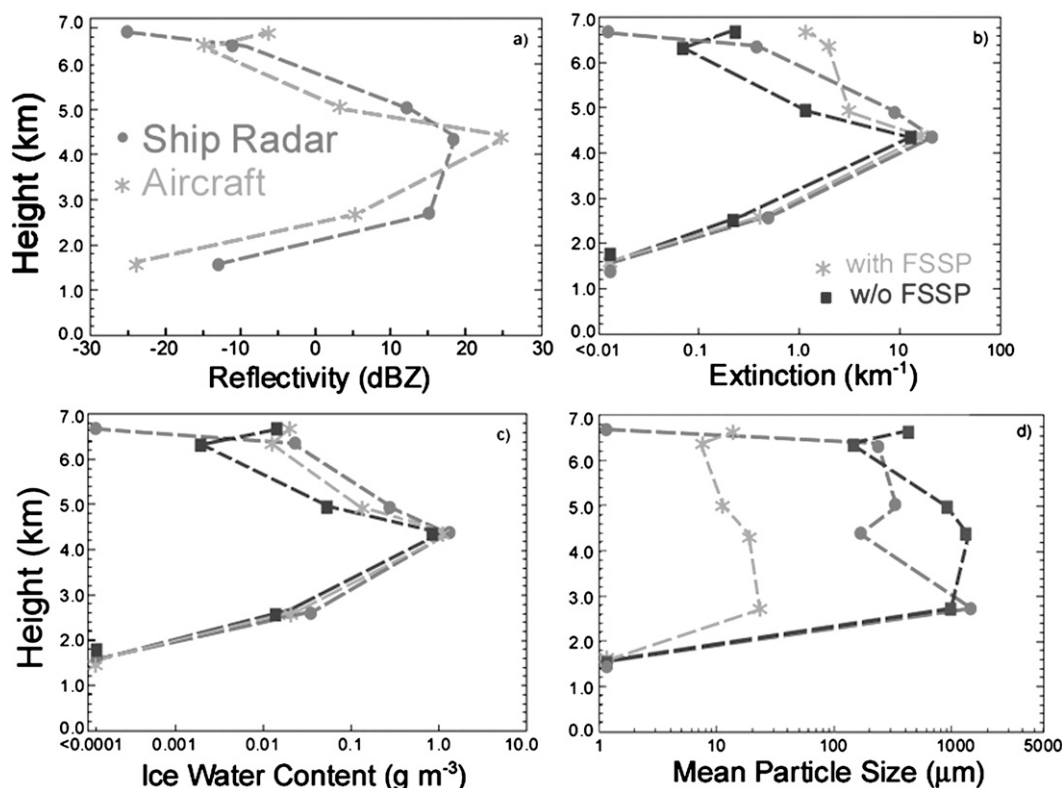


FIG. 6. 28 July comparisons of radar- (dark gray filled circles) and aircraft- (light gray stars with FSSP, black squares without FSSP) derived microphysics from aircraft overpasses at discrete levels. Aircraft data were averaged for 30 s each side of the overpass with dashed lines added for clarity. Note that the radar retrievals only pertain to the ice contribution, and FSSP data are assumed to reflect liquid water drops. Slight differences between the aircraft-calculated reflectivities shown here and those in Fig. 3 reflect differences in averaging time.

averaging of only a few representative 1-Hz points when the aircraft overflow the radar in Fig. 3. Additional explanations are also possible, including the large uncertainties in both measurement platforms.

Despite the radar–aircraft reflectivity differences, the radar microphysical retrievals are mostly within a factor of 2 of the aircraft measurements, except for the point near cloud top where the radar may be sampling some clear air within its 1-min average. The IWC and volume extinction coefficients differ most strongly in the upper 2 km of the cloud. In the upper 2 km there are supercooled liquid cloud drops that dominate extinction and are a significant fraction (about 40%) of the total condensed water content. For example, at -27°C , the concentration of water drops in a turret was 80 cm^{-3} , the water extinction coefficient was 11 km^{-1} , and the LWC was 0.04 g m^{-3} . In comparison, the ice portion of cloud had a concentration of 0.02 cm^{-3} , extinction coefficient of 1.4 km^{-1} , and IWC of 0.06 g m^{-3} . The situation is reversed at and below 4.5 km where the derived microphysics comparisons of extinction and IWC were good (Fig. 6). In this lower cloud region, values of ice

extinction are 2 to 3 times the water values, and the IWC is one to two orders of magnitude greater than the LWC, as demonstrated in the example PSD shown in Fig. 4. This is consistent with a picture of ice particle size increasing via vapor diffusion, accretion, and aggregation as ice falls down through this deep cloud system with embedded convection.

Large ice particles shattering on the inlet of the FSSP and CPI may artificially enhance the small particle concentration throughout the depth of the cloud (Field et al. 2003). However, recent investigations by Jensen et al. (2009) suggest that the contribution of shattering is minimal when the natural concentration of small particles is relatively high, on the order of 5 cm^{-3} and greater, even when IWC values exceed 1 g m^{-3} . On 28 July FSSP concentrations in the presence of supercooled liquid water (identified by the Rosemount icing probe) generally exceeded 5 cm^{-3} in turrets and were generally $>10\text{ cm}^{-3}$ (Fig. 4). Thus, the recent analysis by Jensen et al. (2009) suggest that shattering plays a minor role in this regime.

The agreement in mean particle size shown in Fig. 6 is poor at all altitudes. Mean particle size is a function of

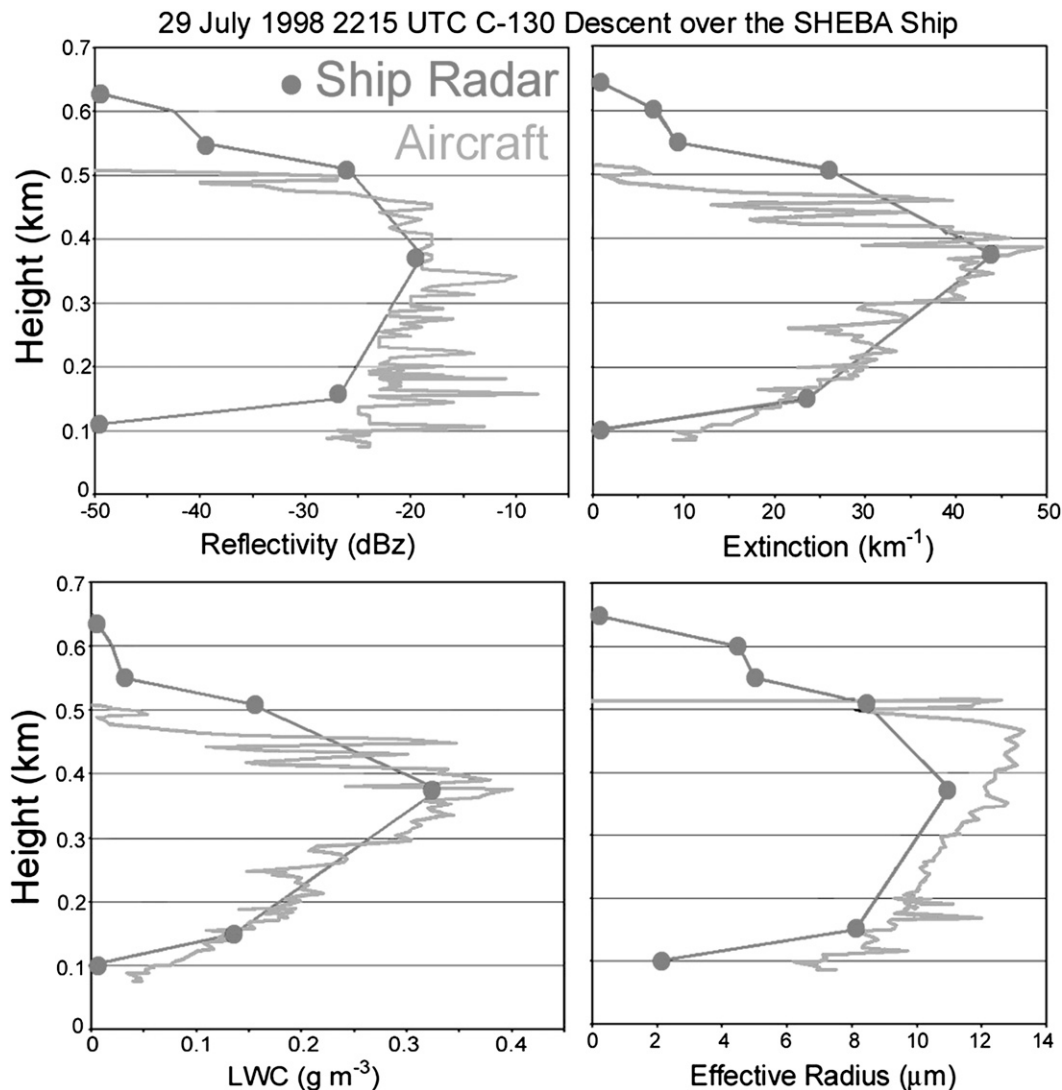


FIG. 7. 29 July comparisons of radar- (dark gray filled circles) and aircraft- (light gray line) derived microphysics from aircraft descent over the SHEBA/FIRE-ACE ship. The radar retrievals follow the methodology of Frisch et al. (1995), using the mean aircraft-derived cloud droplet number concentration.

number concentration and particle linear dimension. Since Fig. 6 suggests that there are about three orders of magnitude more particles in the small mode ($< \sim 50 \mu\text{m}$) compared with the large mode ($> \sim 50 \mu\text{m}$), mean particle size is strongly influenced by the small end of the particle size distribution. The radar algorithm is not recognizing the small particles, which in part may explain the poor agreement; this highlights a fundamental limitation of the radar to retrieve microphysical parameters in mixed phase when the small particle mode dominates the number PSD.

d. Radiative impact

The 28 July precipitation fell on a warm surface, with light rain reported by surface observers and a sub-

sequent short-lived rise in surface albedo noted by Curry et al. (2001). Arguably the more significant radiative impact associated with the frontal passage was the subsequent longer-lasting, surface-coupled low cloud. This fog layer prevailed into August, with melt ponds beginning to freeze and snow beginning to accumulate on the surface again in mid-August (Curry et al. 2001). The fog was sampled on 29 July during the last research flight of SHEBA/FIRE-ACE, with the profiles of corrected FSSP-derived reflectivity, liquid water content, effective radius, and volume extinction coefficient shown in Fig. 7 for one aircraft descent. The 29 July research flight sampled low concentrations of drizzle-sized drops, but the cloud vertical distribution was basically adiabatic except near cloud top, where entrainment

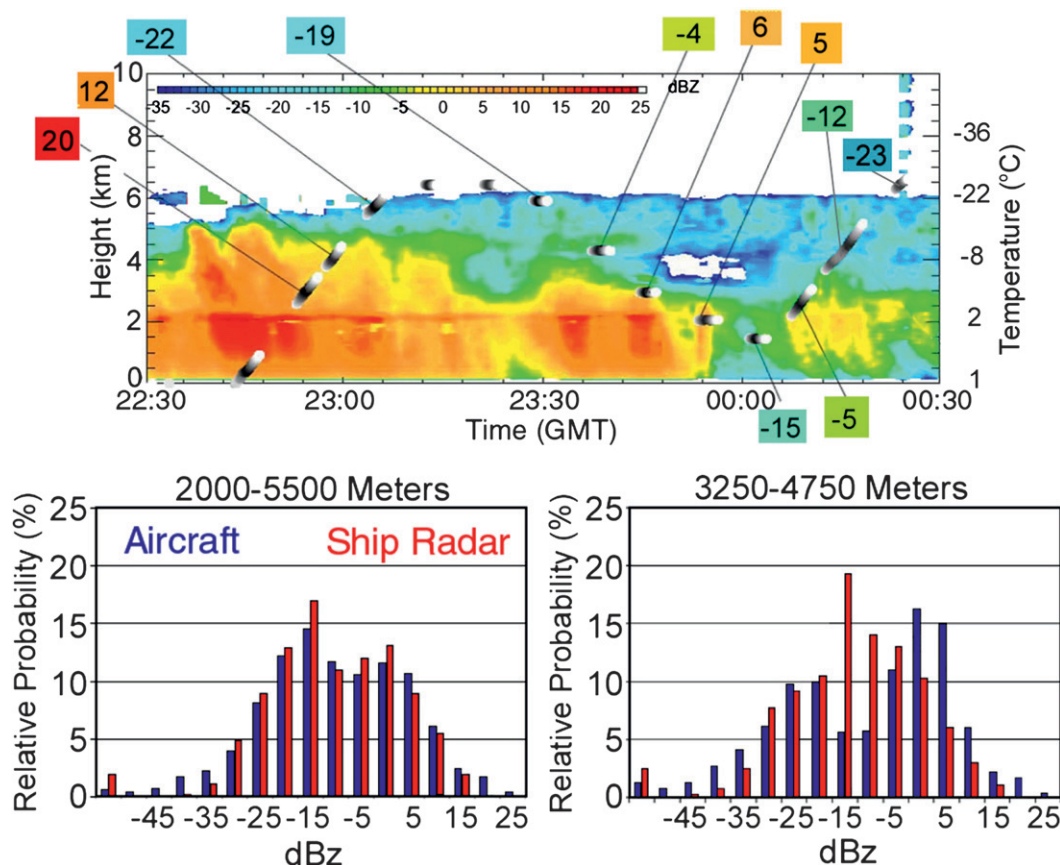


FIG. 8. (top) Cloud radar reflectivities on 18 July with dashes showing altitude and reflectivities calculated from combined PSDs where aircraft overflew the ship. (bottom left) Histograms of radar- and aircraft-derived reflectivities sampled between 2000- and 5500-m altitude from 2230 to 2430 UTC. (bottom right) Histograms of radar- and aircraft-derived reflectivities sampled between 3250- and 4750-m altitude.

reduced the bulk microphysical values (Lawson et al. 2001). The radar-retrieved values invoke the aircraft-determined mean droplet concentration, helping to explain the good radar–aircraft agreement. The obvious discrepancy is near cloud top, where the larger sample volume of the radar again may be averaging in clear air.

The corresponding cloud optical depth was 8–10, or reasonably opaque to the incoming solar radiation. This ensured the cloud radiatively cooled the surface compared to clear sky. The net cloud forcing during July was consistently negative, even more so than June despite the decrease in incoming solar radiation, because of the (usually) lower surface albedo (e.g., Intrieri et al. 2002). The documentation of the 28 July frontal passage and the consequent low cloudiness provides insight into the processes that maintain the summertime cloudiness maximum over the central Arctic (Wang and Key 2005; Hahn and Warren 2007) and the ultimate influence of cyclone activity on the summer surface energy budget near the North Pole.

4. 18 July 1998 (78.23°N, 166.12°W): Mixed-phase deep stratus cloud

This deep mixed-phase cloud evolved within a more synoptically suppressed environment than that of 28 July. An atypical cloud clearing occurred on 17 July (Curry et al. 2001), in accord with the large anticyclone near the ship (Fig. 1b), followed by the development of a thin liquid ($\sim 50 \text{ g m}^{-2}$) cloud increasing in height from approximately 2 km to almost 4 km during 18 July (not shown), consistent with the National Centers for Environmental Prediction (NCEP)–NCAR 700-hPa vertical velocity field (Fig. 1d). The $\sim 500\text{-m}$ -thick liquid cloud precipitated ice at 1800 UTC and then evolved into the deeper, precipitating cloud sampled as part of SHEBA/FIRE–ACE late in the day.

a. Cloud vertical structure and microphysics

Figure 8 shows the 18 July cloud radar reflectivities with the aircraft flight path superimposed. The enhanced cloud radar reflectivity field at 2 km (i.e., bright band)

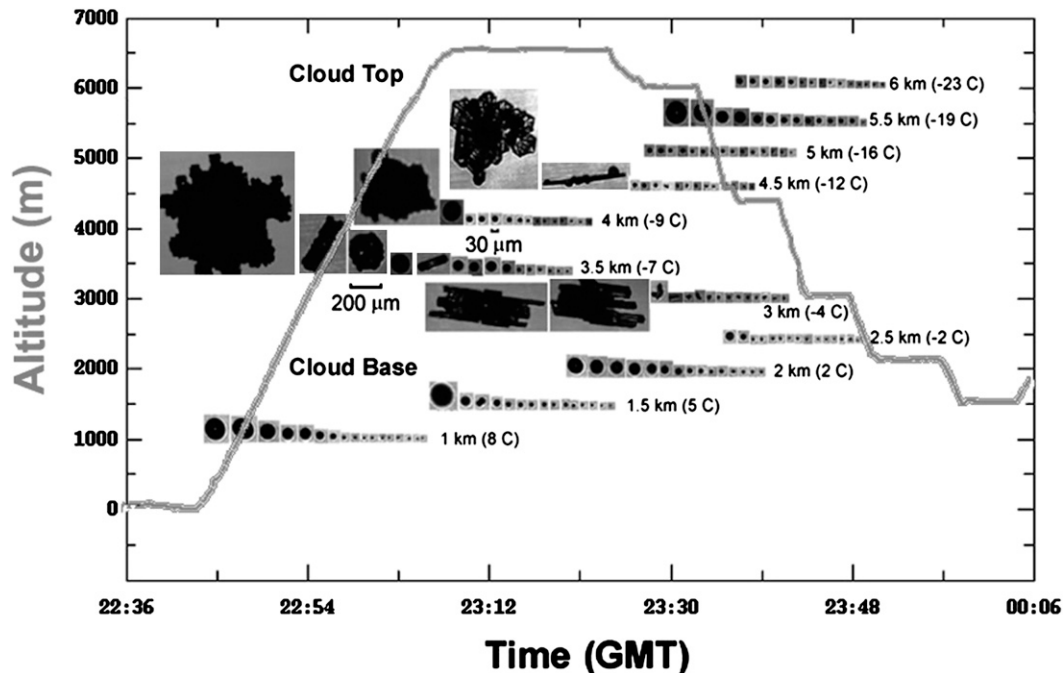


FIG. 9. Plot showing (left axis) C-130 pressure altitude and (right axis) temperature on 18 July 1998 with examples of CPI images observed during ascent from 2247 to 2305 UTC.

indicates melting and implies the presence of ice particles above. The cloud-top temperature was between -23° and -25°C . More detail on the cloud vertical structure is provided in Fig. 9, which shows the ascent and descent profiles of the C-130 along with examples of CPI images as a function of temperature and altitude observed during ascent through the 18 July deep stratus cloud. The images reveal a complex vertical microphysical structure, with all-ice, all-water, drizzle, and mixed-phase layers.

In the upper cloud portion, from about -13°C (4.6 km) up to cloud top at 6 km, the FSSP droplet concentration varied from about 50 to 250 cm^{-3} , averaging about 125 cm^{-3} , and the LWC ranged from 0 to 0.3 g m^{-3} , averaging about 0.15 g m^{-3} . The FSSP drop size distributions were relatively uniform but not especially broad, with drop sizes $>20\text{ }\mu\text{m}$ found in concentrations $>1\text{ cm}^{-3}$ only in the regions with LWC $>\sim 0.1\text{ g m}^{-3}$. CPI images showed a relatively low concentration of drizzle drops up to $125\text{ }\mu\text{m}$ in diameter near -19°C . The aircraft microphysical measurements corresponded with a reduction in radar reflectivities at altitudes above about 4.6 km (Fig. 8), with values remaining $<-10\text{ dBZ}$ up to the top of the cloud. The low radar reflectivity values within the upper half of the cloud (Fig. 8) suggest that large ice particles were sparse, and CPI images (Fig. 9) and the FSSP drop size distribution suggest that the small particles were mostly su-

percooled cloud drops. The cloudy portion associated with the low reflectivity values deepened as time progressed (Fig. 8), but a bright band (from melting ice) is apparent throughout almost the entire time–height section.

As shown in Fig. 9, and to some degree in the radar data (Fig. 8), the middle of the cloud (3 to 4.5 km, -4° to -12°C) contained the most microphysical variability over the course of the sampling, from supercooled small cloud drops to drizzle to large snowflakes. Varying amounts of supercooled liquid water dominated the observations at temperatures $>-12^{\circ}\text{C}$. Pockets of drizzle drops 50 to $250\text{ }\mu\text{m}$ in diameter were sampled at temperatures between -7° and -11°C , with both supercooled and frozen drizzle observed at -7.5°C ($\sim 3.5\text{ km}$, 2256 UTC; Fig. 10). In contrast to the July 28 case, supercooled liquid water was sampled more frequently and at higher liquid water contents. LWC values approaching 0.8 g m^{-3} were recorded at -4°C (2344 UTC) and values of $0.2\text{--}0.3\text{ g m}^{-3}$ at -15°C (2330 UTC and between 2250 and 2300 UTC). Large dendritic ice particles and aggregates, thermodynamically consistent with the measured temperature range of -11° to -12°C , contributed to the highest ice water contents observed on this day, with an 18-s average of 1.3 g m^{-3} (Fig. 11) and peak values exceeding 2 g m^{-3} . The pronounced change in microphysics sampled by the aircraft (e.g., supercooled drizzle at 3.5 km, then

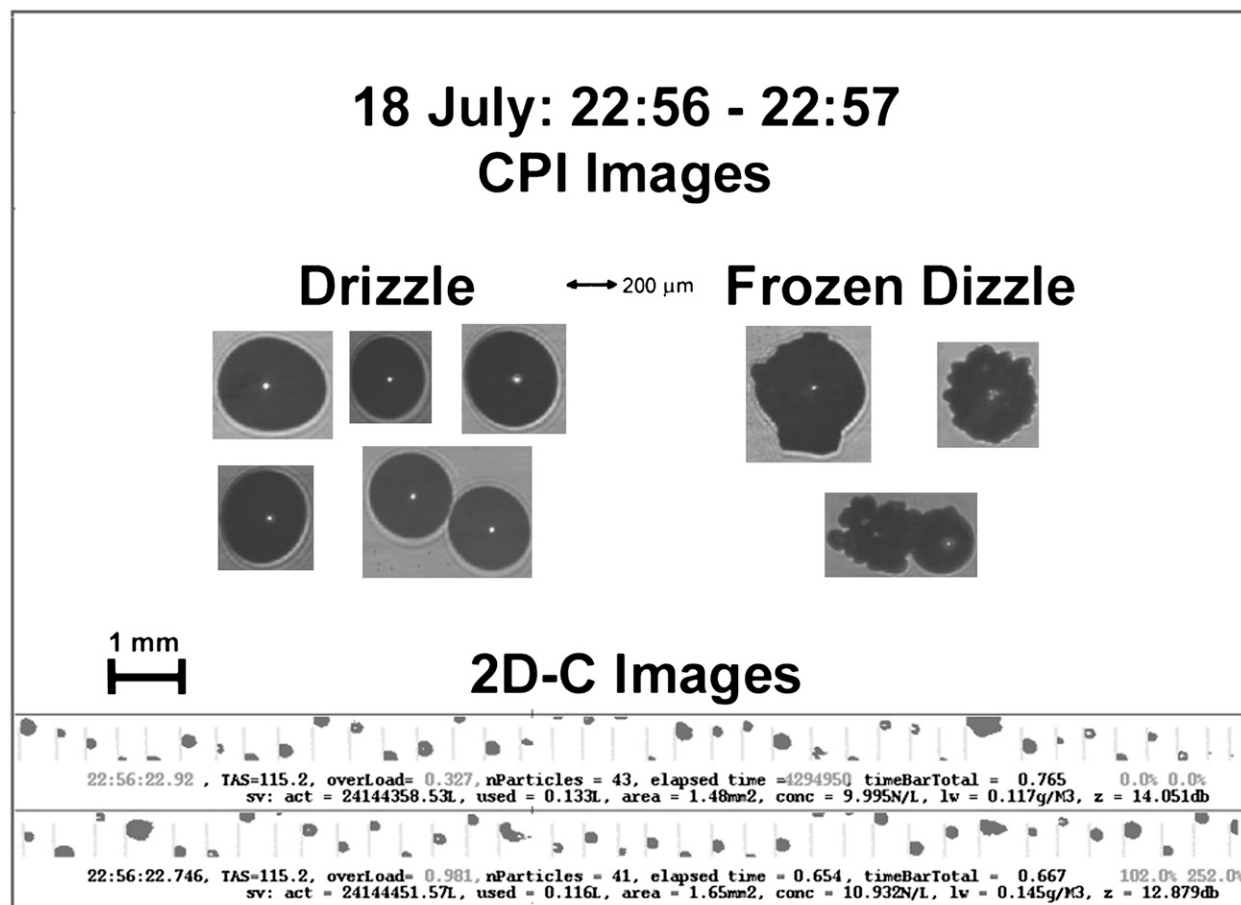


FIG. 10. CPI and 2D-C images showing examples of supercooled drizzle and frozen drizzle observed at -7.5°C during ascent on 18 July.

snowflakes at 4.5 km at around 2300 UTC) is undetected in the cloud radar reflectivity profile (Fig. 8) because large ice particles still dominate the cloud radar reflectivity field at 3.5 km. The drizzle sampled at -19°C (5.5 km) by the aircraft was apparently too small an amount to noticeably increase the cloud radar reflectivity values, even within an all-liquid layer.

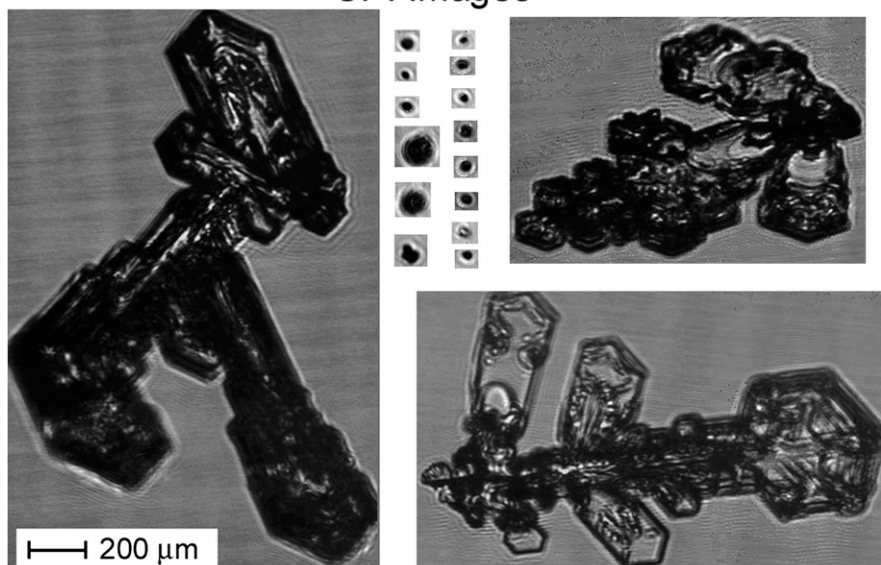
CPI images gathered during the descent from 2310 to 2335 UTC (between -25° and -12°C and above 4.5-km altitude) contained very low concentrations of drizzle drops ($<0.01\text{ m}^{-3}$). Ice crystal observations were similarly rare during this time period ($<0.1\text{ m}^{-3}$), with the few observed consisting of mostly large ($>750\text{ }\mu\text{m}$), heavily rimed particles. Dendrites were present within the horizontal leg occurring at 2338 UTC (4.5 km; not shown but similar to Fig. 11), consistent with the -11°C temperature. Continuing the descent through the -11° to -7°C region, only occasional small ($\sim 50\text{ }\mu\text{m}$) drizzle drops were observed. All of this is consistent with a deep (3–4 km) layer of low radar reflectivities that only occasionally exceeded -15 dBZ (Fig. 8), bearing witness to

a cloud microphysical field during the 2310–2335 UTC time period that contained little if any large ice.

In contrast, the horizontal leg at 2343–2348 UTC ($\sim -5^{\circ}\text{C}$, 3.5 km) took place within a region of slightly higher cloud radar reflectivity, about 0 to 5 dBZ as shown in Fig. 8, and a more complex microphysical population was encountered. The ice particles were mainly columns, many of them rimed (Fig. 12), with supercooled drizzle out to $150\text{ }\mu\text{m}$ in diameter. One of the highest LWC values observed during FIRE-ACE, approaching 0.8 g m^{-3} (Fig. 13), occurred at about 2344 UTC (-4°C), exceeding even the 28 July convection maximum. The maximum vertical velocity in this region was 1 m s^{-1} and there was no evidence of embedded convection. This measurement of relatively high LWC appears to be reliable and is corroborated by the FSSP, which measured 1.1 g m^{-3} in this region. Figure 14 shows the ice and water particle size distributions, demonstrating that in this region the liquid portion of cloud dominates all three moments of the size distribution.

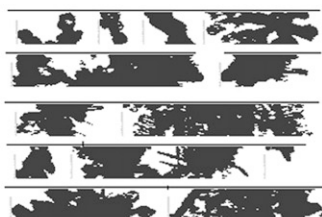
July 18 22:59:12 - 22:59:30

CPI Images



2D-C Images

1 mm



2D-P Images

1 cm

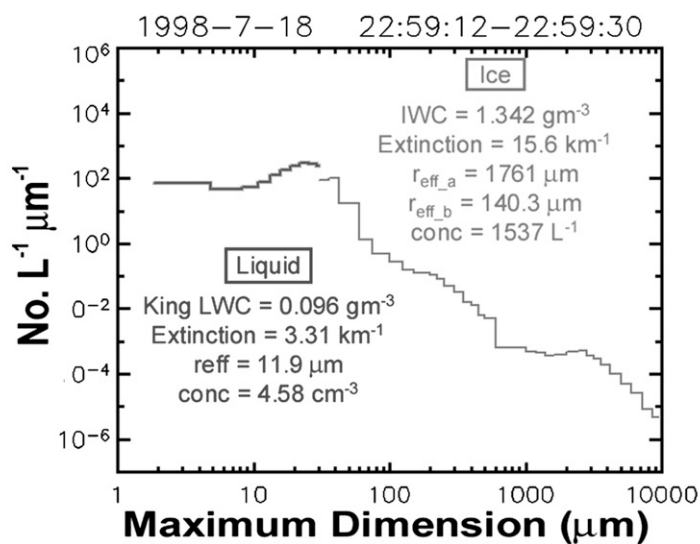
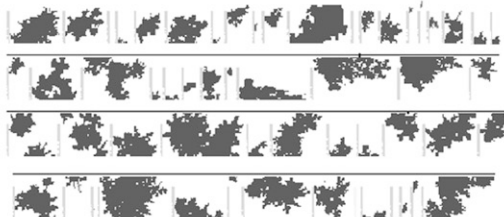


FIG. 11. Examples of (top) CPI and (middle) 2D-C and 2D-P images collected at -11° to -12°C during ascent over the SHEBA ship on 18 July 1998. (bottom) PSDs for ice (light gray) and water (dark gray) with derived microphysical parameters.

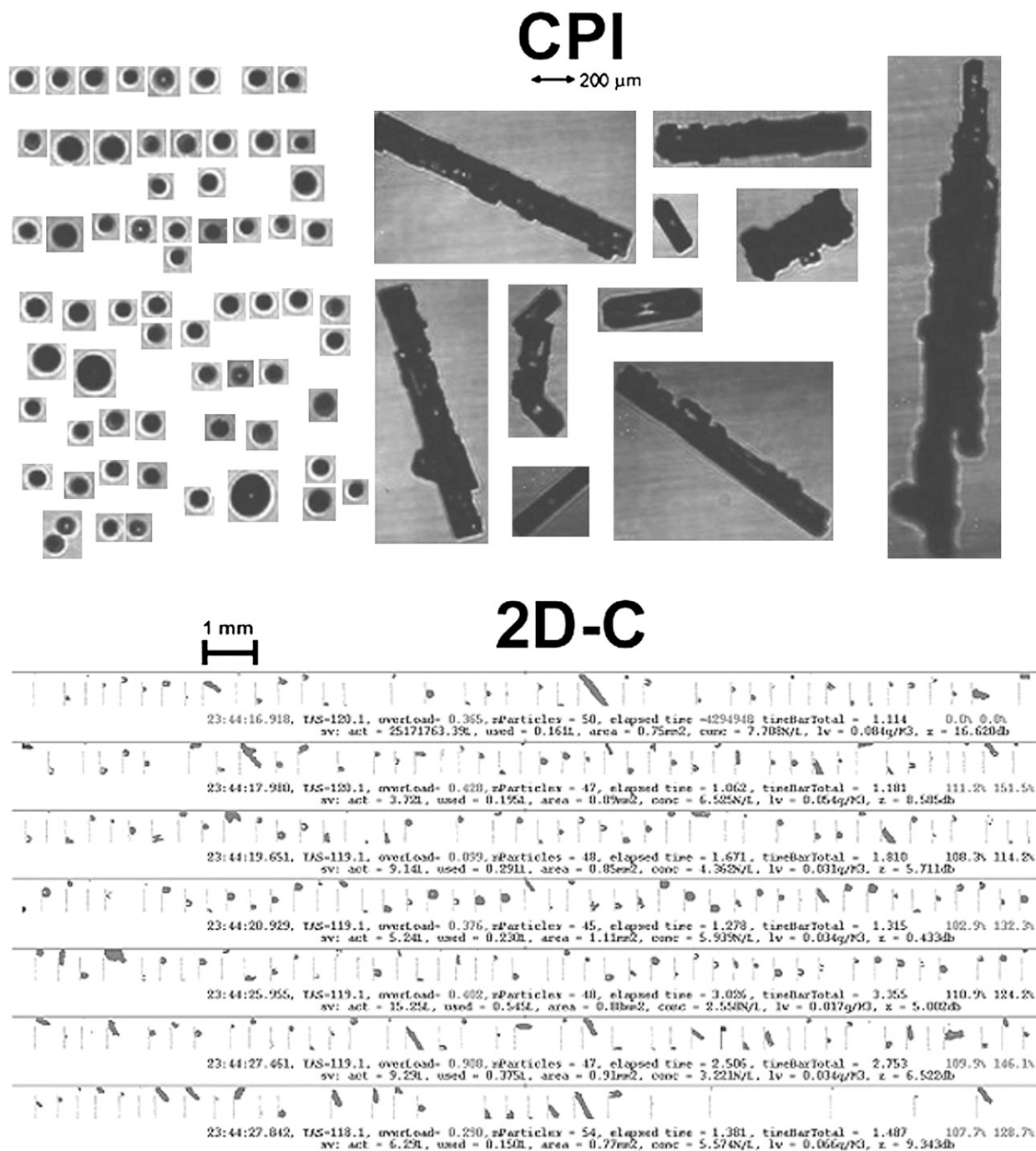


FIG. 12. Examples of CPI and 2D-C images for measurements from 2343–2348 UTC in a region with supercooled drizzle at -4.5°C on 18 July.

The occurrences of freezing drizzle observed at -19°C and between -11° and -4°C (Figs. 10 and 12) appear to be examples of “non-classical freezing drizzle” formation (Cober et al. 1996; Lawson et al. 1998b), where drizzle is formed through coalescence of supercooled drops, usually near the tops of stratus layers. However,

in both the Cober et al. (1996) and Lawson et al. (1998b) measurements, the FSSP drop spectra were generally broader and the 2D-C drizzle concentrations were significantly higher, in excess of 500 L^{-1} , whereas in this case the 2D-C concentration was $<0.1\text{ m}^{-3}$. The presence of freezing drizzle was confirmed using CPI

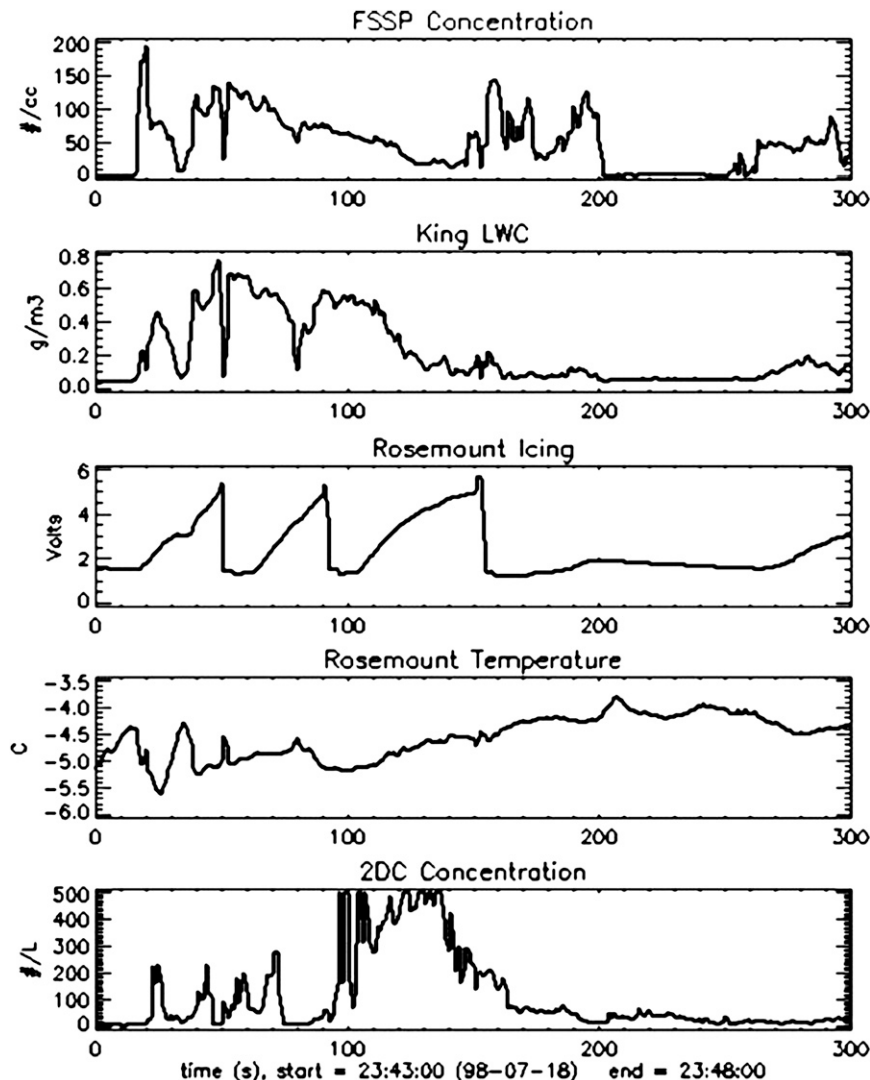


FIG. 13. Time series of microphysical parameters observed in a region with high LWC, supercooled drizzle, and rimed columnar ice particles on 18 July.

imagery when there was no activity on the Rosemount icing probe, presumably because the supercooled LWC was below the detection level of the Rosemount icing probe, which is about 0.001 g m^{-3} (Mazin et al. 2001; Cober et al. 2001). Politovich (1989) and others have pointed out the potential for supercooled large drops to affect aircraft performance. There is not enough data to determine whether these low concentrations of freezing drizzle can significantly affect aircraft performance, but it is worthwhile to note that freezing drizzle existed and was not detected by the icing probe, which is commonly used on commercial airliners.

The early portion of the time–height plot of radar reflectivity (Fig. 8) appears to have a cellular structure that would normally be associated with convection. Also,

there was a 10-km region that contained an average LWC of about 0.5 g m^{-3} , which is generally associated with convection. Nevertheless, the measured updrafts were weak, not exceeding about 1 m s^{-1} , and the thermodynamic vertical profile was convectively stable.

In addition, the 18 July radar time–height plot often contained a region from about 4.5 to 6 km that was mostly devoid of large ice, in contrast to the 28 July radar time–height plot (Fig. 3), which contained significant convection that extended to nearly 8 km. The 28 July radar profile is commensurate with a picture of rising parcels that develop ice in the updrafts. On 18 July, deep regions of supercooled cloud with low ice concentration could have formed by gentle lifting in very clean air devoid of ice nuclei. Pockets with large ice

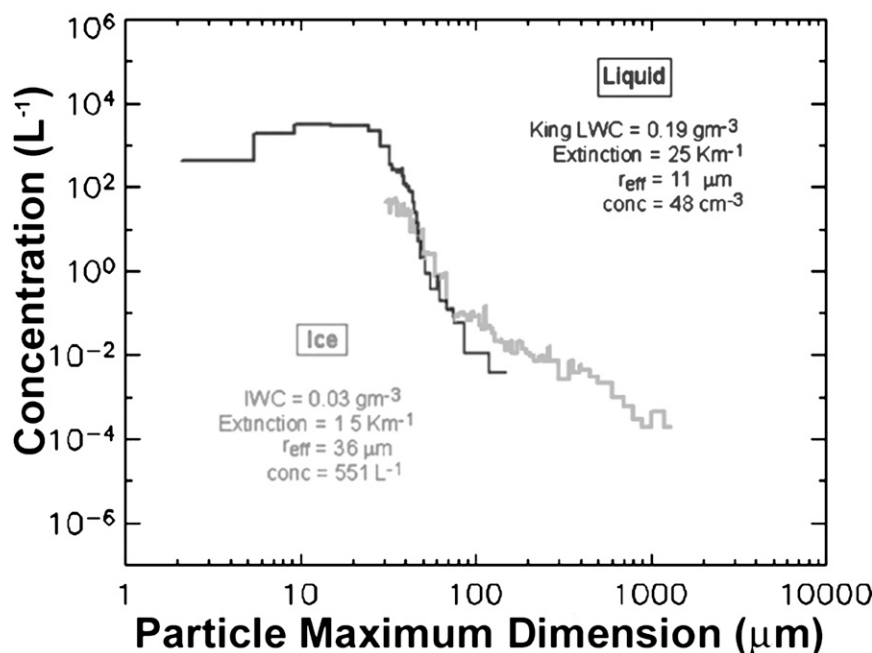


FIG. 14. Liquid (black) and ice (gray) size distributions on 18 July from 2344 to 2348 UTC at -4.5°C . Derived microphysical properties for each phase are shown in insets.

aggregates (Fig. 11) and higher radar reflectivity (Fig. 8) may have formed where localized regions with high ice nuclei concentrations existed. This picture is generally consistent with SHEBA/FIRE-ACE ice nuclei observations reported by Rogers et al. (2001), who measured very low ice nuclei concentrations in general (50% of the measurements were zero), with rare localized pockets of high ice nuclei (hundreds per liter) at a chamber temperature of -25°C . However, the Rogers et al. measurements were collected in May when there was little open water and no ice nuclei measurements were made in July when there were considerable melt ponds. The pockets of high ice nuclei concentrations may have biogenic origin from the melting ice surface (Curry et al. 2000; Leck and Bigg 2005), and the resulting ice development may have been aided by the presence of drizzle (Hobbs and Rangno 1985).

Lawson et al. (2001) reported on two regions with very high ($>2000 \text{ L}^{-1}$) ice particle concentrations sampled during the ascent on 18 July. The ice particles were identified using CPI imagery and about two-thirds were small spheroidal particles. At the time, it was felt that shattering of large ice particles would not result in a preponderance of small spheroidal ice in a uniform size range of 20 to $40 \text{ }\mu\text{m}$. However, more recent evidence of ice particle shattering using high-speed video strongly suggests that this is possible (A. Korolev 2008, personal communication). The new 2D probe imagery shows that large dendrites and aggregates were present

in significant concentrations. Dendrites and aggregates are the particle habits that Korolev and Isaac (2005) found were most prone to particle shattering. It is our opinion that in this regime where the Rosemount icing rod did not detect supercooled liquid water, the relatively high ($>1 \text{ cm}^{-3}$) ice concentrations measured by the FSSP (and confirmed to be ice using CPI imagery) may have been contaminated with spurious small ice artifacts resulting from shattering. However, we cannot quantitatively estimate the extent of the enhancement due to shattering, so the actual ice concentration could still have been relatively high when compared with classical nucleation theory.

In the 0° to -2°C region of cloud, Lawson et al. (2001) were puzzled by high LWC and FSSP droplet concentrations observed during ascent, with no accompanying ice particles, despite observations of large particles higher in the cloud and drizzle below the 0°C level. Lawson et al. (2001) did not analyze data from the descent, which, along with identification of the melting level bright band, did report large ice particles in the layer from 0° to -2°C . CPI and 2D imagery (Fig. 15) show that the ice particles were mostly heavily rimed columns and aggregates. These data suggest that when climbing at a shallow angle during ascent, the aircraft penetrated an isolated region that was devoid of large ice despite the indication of melting within the cloud radar data. This further attests to the spatial inhomogeneity of this deep stratus cloud.

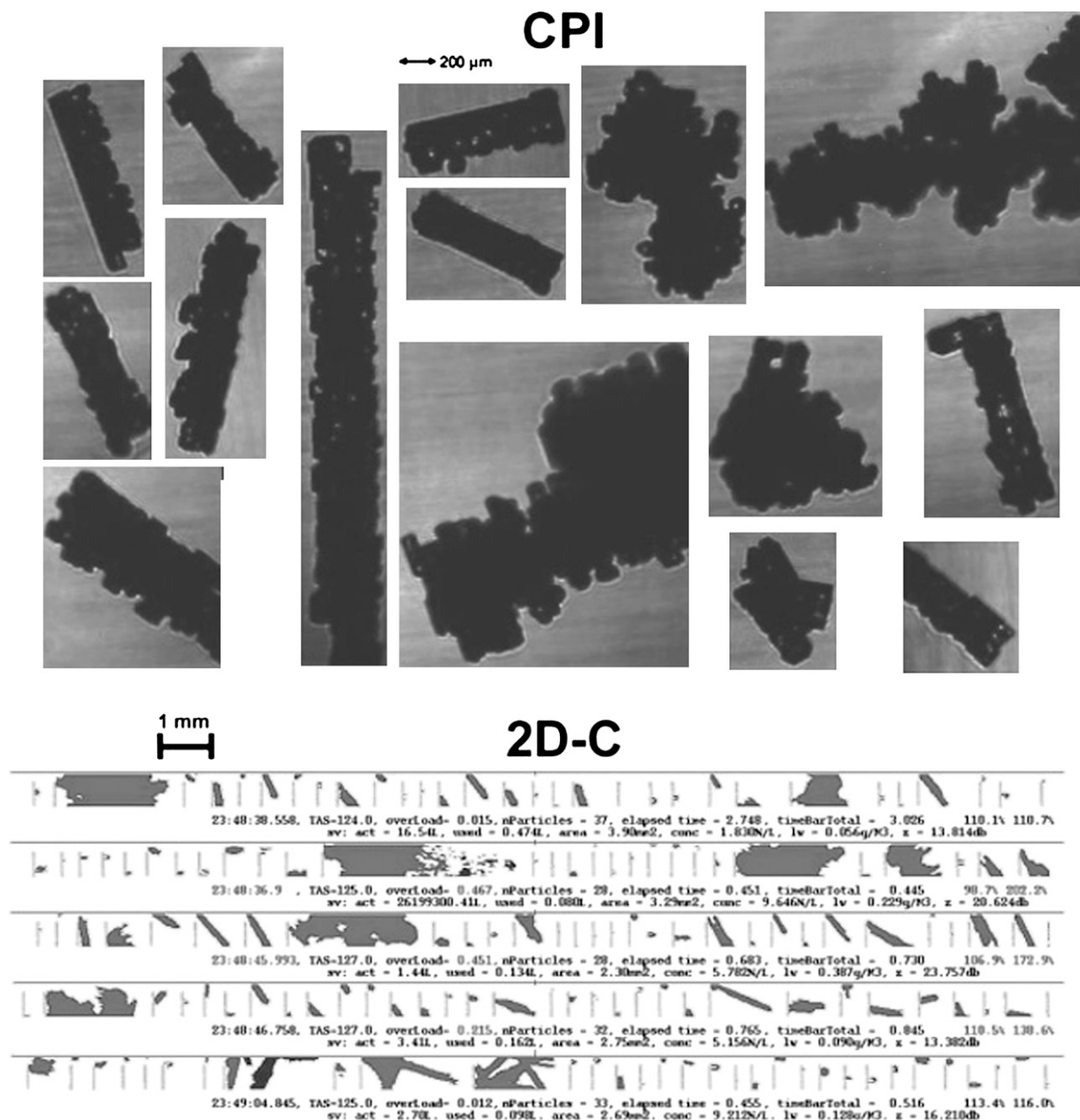


FIG. 15. As in Fig. 12, but for images observed at 2348–2349 UTC at 0° to −2°C during ascent.

b. Radar/in situ microphysical comparisons

There is rough agreement between radar and aircraft-derived reflectivities shown in Fig. 8, with the aircraft values mostly exceeding the radar reflectivity. There is no single explanation for the lack of agreement; however, several factors contribute to uncertainties in both measurements. The aircraft-derived reflectivity is subject to sizing errors in the particle probes and assumptions regarding the effects of nonspherical particles. The

cloud is also spatially inhomogeneous and the aircraft has a much smaller sample volume than the radar. The radar is often sampling particles outside the Rayleigh scattering regime, and the beam is likely to be attenuated by precipitation that is falling with variable intensity from the stratiform cloud.

Thus, there is no consistent explanation for the lack of agreement in radar–aircraft reflectivity and the poor agreement in radar–aircraft IWC shown in Fig. 16. The disagreement in other derived aircraft microphysical

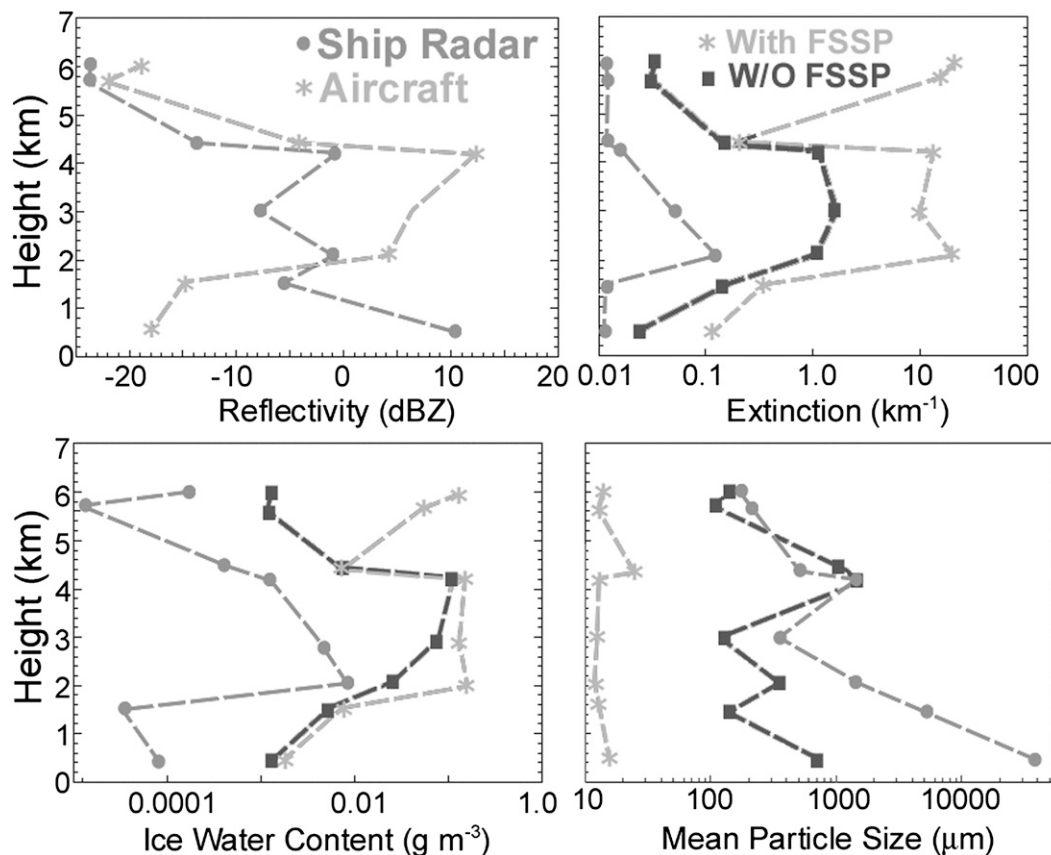


FIG. 16. As in Fig. 6, but for measurements and retrievals from the 18 July case.

parameters appears to be due in part to the presence of cloud drops that contributed significantly to extinction and mean particle size but were not detected by the radar in the presence of large ice. Also, the radar microphysical retrievals are based on a priori assumptions that may not have been representative of actual conditions. The disagreement in radar- and aircraft-derived microphysical parameters is in contrast to the comparison for a SHEBA/FIRE-ACE mixed-phase cloud shown in Zuidema et al. (2005) and suggests that more aircraft-radar comparisons in mixed-phase clouds are needed to determine regimes where radar retrievals of microphysical parameters may be reliable.

5. 8 July 1998 (78.08°N, 166.12°W): All-ice middle- to upper-level cirrus cloud

As the only all-ice cloud sampled during SHEBA/FIRE-ACE by the C-130, the 8 July case is particularly valuable for radar-aircraft microphysical comparison. The Khvorostyanov et al. (2001) analysis relied solely on the FSSP and CPI measurements, reporting aircraft-derived ice water contents an order of magni-

tude smaller than those derived from radar. We revisit the comparison again with the addition of 2D-C measurements and consideration of possible effects of crystal shattering on the inlet of the FSSP (Field et al. 2003; Jensen et al. 2009), which in this case appear to be significant.

The cloud radar reflectivities are shown superimposed with the aircraft flight track (Fig. 17, top), with the in situ aircraft reflectivity values calculated from the aircraft overflights of the ship. The radar- and aircraft-derived reflectivity values were in fairly close agreement during the six overpasses, especially where the cloud radar reflectivity gradients were small. Figure 17 shows histograms of reflectivity from the radar and aircraft measurements for the lower and upper half of the cloud. The reflectivity values in the histograms were derived using the same technique described for Fig. 3. Unlike histograms from the mixed-phase cases (Figs. 3 and 8), the radar measurements consistently exceed the aircraft-derived reflectivities by a mean value of about 5 dBZ.

Recent evaluation of the effect of crystal shattering on the cloud and aerosol spectrometer (Baumgardner et al.

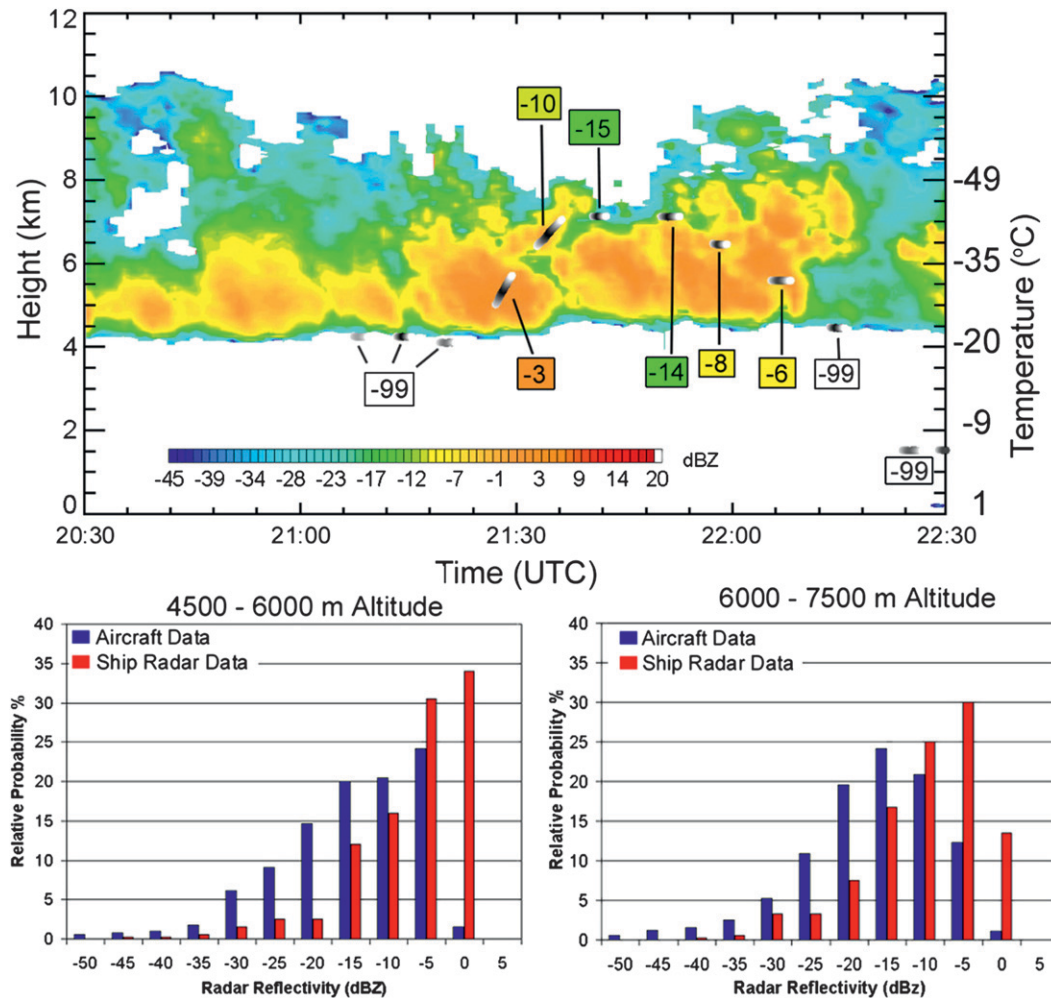


FIG. 17. As in Fig. 8, but data were collected on 8 July in an all-ice cloud.

2001), an instrument with an inlet tube and measurement principle similar to the FSSP, has shown that the number of spurious small particles is proportional to the mass of large crystals in cirrus anvil clouds (Jensen et al. 2009). Additionally, shattering on the tips of the 2D-C probe can enhance the concentration of small to midsize particles (Korolev and Isaac 2005; Field et al. 2006; Jensen et al. 2009). Jensen et al. (2009) show that the effect of shattering on FSSP particle concentration can be negligible when there is a large natural concentration of small particles, either water drops or ice, such as is seen in the mixed-phase clouds investigated here. In contrast, the 8 July ice particle size distributions (Fig. 18) are similar to those for which Jensen et al. (2009) suggest crystal shattering is likely to be playing a role in significantly enhancing the small mode of the size distribution. That is, the total particle concentration in the 8 July case is moderate, 200 to 300 L^{-1} , except for the outlier point shown in Fig. 18. The outlier point at 5.5 km

(2127 UTC) that has a much higher particle concentration is suspicious because compared to the other points the outlier has about twice the ice mass. This is likely to be a situation in which large ice crystals shattering on the inlets and tips of the particle probes have artificially enhanced the total particle concentration.

Figure 19 shows vertical profiles of microphysical cloud properties derived from the radar and aircraft measurements from the overpasses shown in Fig. 17. As expected from Fig. 17, there is good agreement in reflectivity derived from the radar and aircraft measurements. The derived microphysical properties are shown with and without FSSP (3 to 45 μm) measurements; recall that for all-ice clouds the FSSP contribution reflects ice, not water. As discussed previously, in general, large ice particles shattering on the inlets and tips of particle probes may or may not have a significant impact on derived microphysical parameters, depending on the mass of large ice and the natural concentration of small

CPI and 2D-C Images and Composite PSDs of Regions over Sheba Ship on 8 July 1998

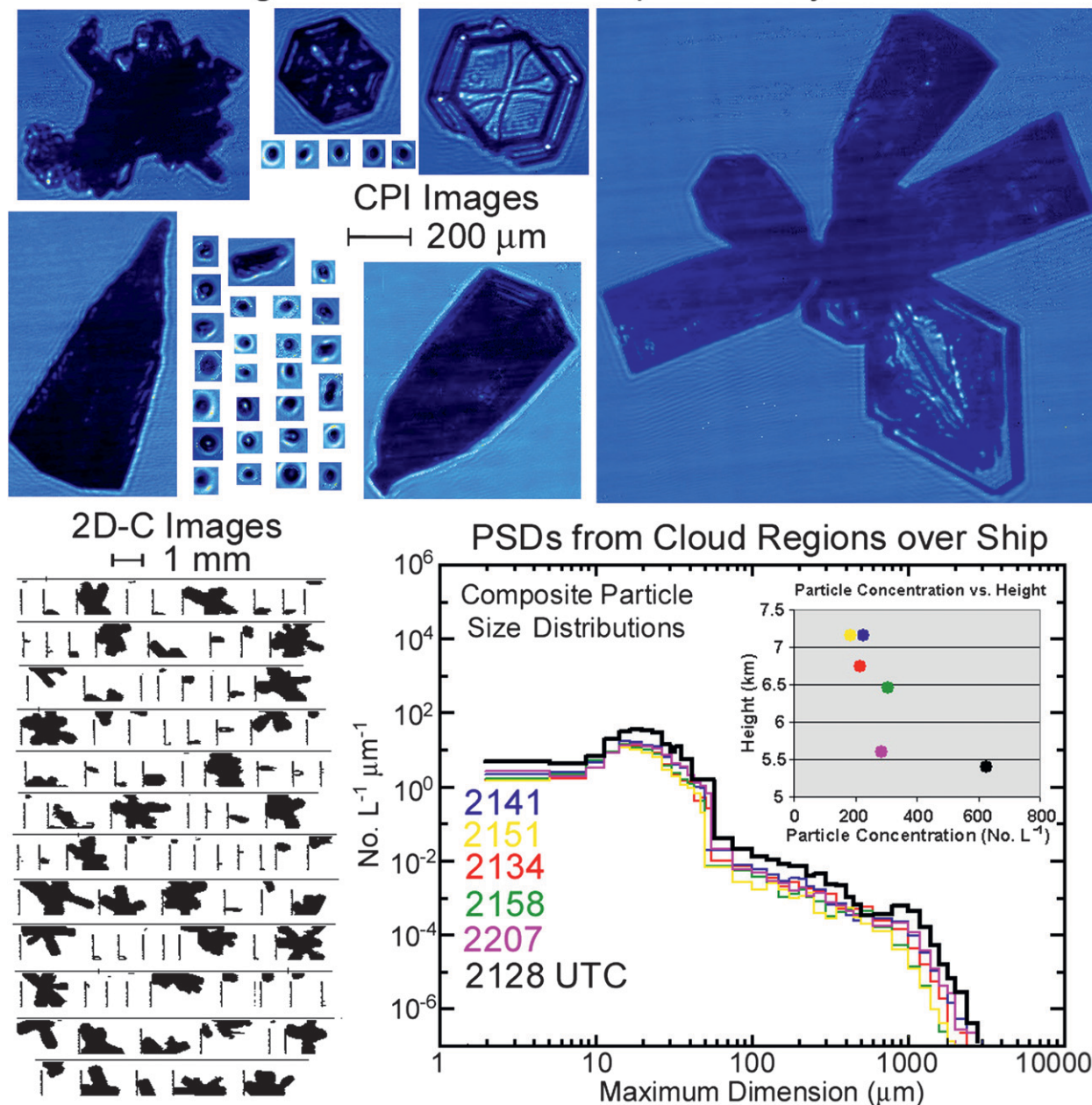


FIG. 18. Examples of (top) CPI and (lower left) 2D-C images of ice particles observed in cirrus cloud on 8 July. Composite size distributions and total particle concentrations vs altitude shown at lower right are color-coded to times when the C-130 flew over the ship.

ice. The volume extinction coefficient agrees better with the radar-derived value without contribution from the FSSP, whereas IWC agrees better with the radar-derived value when FSSP measurements are included. Mean particle size agrees much better with the radar value when the FSSP measurements are removed. Since

both the radar and aircraft contain significant measurement uncertainties, (e.g., shattering on the inlets of the in situ probes, insensitivity of the radar to the small particle mode, and uncertainties in ice particle densities), it is unwise to speculate which retrieval, if either, is correct.

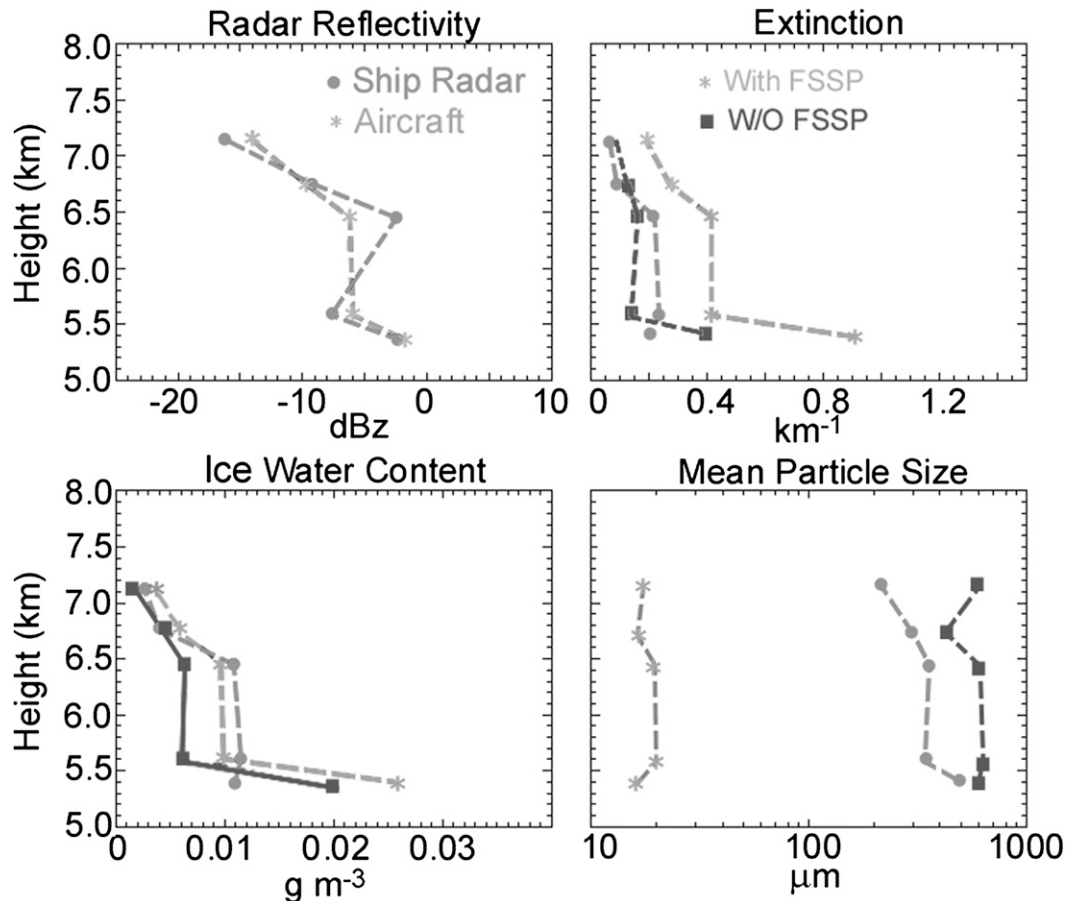


FIG. 19. As in Fig. 6, but for microphysical properties derived from data collected in an all-ice cloud on 8 July. Note that for this case the FSSP contribution is from ice particles and that the radar retrievals of ice microphysical quantities were developed to assume a small-particle contribution.

6. Summary and discussion

Microphysical data from the NCAR C-130 operated during the 1998 SHEBA/FIRE-ACE project were re-analyzed to include PMS 2D-C and 2D-P measurements and compared with corresponding millimeter Doppler radar data analysis. Three case studies were discussed in detail, an all-ice cirrus cloud (8 July); a deep, layered, mixed-phase stratus cloud (18 July); and a deep mixed-phase nimbostratus cloud system with embedded convection (28 July). An all-water boundary layer cloud observed on 29 July was briefly discussed.

The cases occurred within varied yet typical synoptic situations over a melting ice surface 1200 km from the North Pole. The 28 July case is also discussed in MZM; the 8 July case is discussed by Khvorostyanov et al. (2001), and some of the data from 18 July were discussed in Lawson et al. (2001). Large ice particle aggregates (1 to 3 cm) and IWC (up to 2 g m^{-3}) were documented in both mixed-phase (18 and 28 July) clouds, and 5-mm

graupel particles occurred within embedded convection during 28 July. These observations of high IWC, graupel, and large snowflakes within 1200 km of the North Pole have not been previously reported in the literature and as such are useful for further development of model microphysical parameterizations for the Arctic (e.g., MZM).

For the all-ice cirrus case (8 July), reflectivity derived from four of the five aircraft overpasses was within 2 dBZ of the radar value, with the fifth overpass agreeing to within 5 dBZ (Figs. 17 and 19). This suggests either that the Brown and Francis (1995) relationship used in the radar retrieval was appropriate in this case or that the agreement was a fortuitous coincidence. The aircraft-derived volumetric extinction coefficient was about a factor of 2 greater than the radar value when FSSP data were included but improved by about 70% when FSSP data were excluded. In contrast, the agreement in radar- and aircraft-derived IWC degraded when FSSP data were removed. The agreement in aircraft- and

radar-derived mean particle size improved by more than an order of magnitude after FSSP data were removed. These inconsistencies cannot be completely explained by shattering on the FSSP inlet. It is likely in this case that shattering had an effect on the small particle mode measured by the FSSP and is perhaps best reflected in the improved radar–aircraft agreement in mean particle size when FSSP measurements are removed.

The large snowflakes on 28 July were associated with a deep, almost-isothermal layer below a layer of weak convective instability aloft. The conditions that generated the large snowflakes are similar to a North Atlantic case over open ocean off the coast of Newfoundland documented in Lawson et al. (1998a). After the frontal passage a long-lasting low cloud was sampled on 29 July during the last SHEBA/FIRE–ACE research flight. The 28–29 July transition helps elucidate how storm-track cloudiness contributes to the summertime cloud maximum (which is primarily low cloud) over the central Arctic. The case is one example of a summertime cyclone in the Beaufort/Chukchi Seas; cyclones are an important transport mechanism for moisture to the polar cap (e.g., Sorteberg and Walsh 2008). Arctic cyclonic activity has been increasing in recent times and is projected to increase further (Solomon et al. 2007; Zhang et al. 2004).

Radar reflectivity and aircraft reflectivity derived from ship overpasses differed by about 10 dBZ throughout the entire cloud depth on 28 July. Radar- and aircraft-derived microphysical parameters did not compare well in the region from cloud top down to 4.5 km, where there was mostly supercooled cloud drops and low concentrations of ice. In contrast, radar- and aircraft-derived volumetric extinction coefficient and IWC compared very well at and below 4.5 km, in the region where the aircraft often observed high IWC and large ice particles.

The 18 July case is more representative of large-scale suppressed conditions, reflecting above-normal summertime 1998 Beaufort sea level pressures and a circulation pattern resembling that documented for the summer of 2007. The extended analysis of the 18 July cloud documents a layering of mostly all water, all ice, or mixed phase. Regions with supercooled drizzle existed in isolated pockets extending for a few to several kilometers. From cloud top (-25°C) downward to about the -14°C isotherm, the general vertical structure consisted of supercooled cloud drops interspersed with low ice concentrations. Supercooled cloud drops in lower concentrations interspersed with pockets of large ice extended from the -14°C level down to the melting level. Coalescence–collision processes appear to be responsible for creating supercooled drizzle, which was observed in

pockets from -4° to -19°C . Around temperatures of -11° to -12°C , large dendrites and aggregates were observed in the aircraft data, resulting in very high IWC values with an 18-s average of 1.3 g m^{-3} and peak values exceeding 2 g m^{-3} . Radar time–height imagery also revealed pockets with high reflectivity in the -11° to -14°C region that were likely attributable to the large ice particles.

The 18 July case appears to contain regions of gentle uplift, more prevalent after 2300 UTC, that are capable of supporting moderate liquid water contents. Precipitation was apparent as a melting-level radar bright band, which is likely to have attenuated the radar beam and influenced retrievals. Like the 28 July mixed-phase case, there was poor agreement between radar- and aircraft-derived reflectivities, with differences on the order of 10 dBZ. Derivations of cloud properties also did not compare well, with the liquid water contributing substantially to cloud extinction and total water content.

In general, radar- and aircraft-derived reflectivity and microphysical parameters were in poor agreement in the 18 and 28 July mixed-phase cases. The poor agreement may be due to several factors, including the high degree of spatial variability of microphysical properties in these clouds, attenuation of the radar beam in the presence of precipitation, reduced sensitivity of the radar to cloud drops in the presence of larger ice, and uncertainties in both radar retrievals and aircraft measurements. The measurement and retrieval uncertainties are inherent limitations that are difficult to overcome at this time. For example, radar reflectivity is only well defined when cloud particles are spherical water drops. The current radar retrieval algorithms use assumptions on density–size and mass–area relationships that are statistical and developed for all-ice conditions and that further assume an underlying exponential size distribution. Since ice particles vary considerably in size, shape, and density, arguably even more so within mixed-phase clouds, the retrievals’ extension to mixed-phase conditions requires further testing. When the liquid portion of a mixed-phase cloud is a significant fraction of the total water content, the insensitivity of radar to cloud drops in the presence of ice is also a fundamental limitation. In such mixed-phase conditions, one might expect the radar-derived volumetric extinction coefficient (in particular) and (less so) the total water content to be underestimated unless there is an *a priori* assumption or additional information regarding the amount of liquid cloud water. These limitations also apply to space-based cloud–radar retrievals, and especially their application within global radiative flux calculations for these so-named “difficult clouds” (e.g., L’Ecuyer et al. 2008; Kay et al. 2008), lending importance to the need for further

research. A statistical evaluation involving more cases than are examined here may improve radar retrievals within mixed-phase conditions.

The in situ particle probe measurements are also subject to uncertainties that are difficult to overcome. The (older) particle probes used in the SHEBA/FIRE-ACE project contain several measurement errors, including the inability of the 2D-C probe to image particles with sizes less than about 125 μm (Lawson et al. 2006). This limitation can have a significant impact on volumetric extinction coefficient, with a lesser effect on IWC when large ($>125 \mu\text{m}$) ice dominates the IWC. Also, ice particles shattering on the inlet of the FSSP and CPI and the tips of the 2D-C/P probes introduce uncorrectable errors (Field et al. 2003, 2006; Korolev and Isaac 2005; Jensen et al. 2009). The shattering effect may or may not have a significant impact on derived microphysical properties, depending on the ratio of shattered artifacts to natural ice particles in the same size range. Finally, deriving IWC from two-dimensional particle imagery contains inherent uncertainties that are difficult to quantify when ice particle shapes and density are highly variable, which was the case in these mixed-phase clouds.

The radar and in situ comparisons reveal each platform's relative strengths and weaknesses. Some instances occurred where the aircraft sampling missed a microphysical population suggested by the radar, such as large ice particles above the melting level during the 18 July ascent. Alternatively, the aircraft data revealed detailed microphysical properties, such as high concentrations of cloud drops, drizzle, regions with very high IWC, 5-mm graupel particles, and 1- to 3-cm aggregates that cannot be derived from the radar data.

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