A Multisensor Investigation of Rime Splintering in Tropical Maritime Cumuli

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ABSTRACT

Three flights from the Ice in Clouds Experiment–Tropical (ICE-T) field campaign examined the onset of ice near the ascending cloud tops of tropical maritime cumuli as they cooled from 0°C to −14°C. Careful quantitative analysis of ice number concentrations included manual scrutiny of particle images and corrections for possible particle-shattering artifacts. The novel use of the Wyoming Cloud Radar documented the stage of cloud development and tops relative to the aircraft sampling, complemented the manual estimates of graupel concentrations, and provided new clear evidence of graupel movement through the rime-splintering zone. Measurements of ice-nucleating particles (INPs) provided an estimate of primary initiated ice. The data portray a dynamically complex picture of hydrometeor transport contributing to, and likely resulting from, the rime-splintering process. Hundreds per liter of supercooled raindrops ascended within the updrafts as the cloud tops reached 0°C and contributed in part to the 0.1 L−1 graupel detected soon after the cloud tops cooled to −5°C. Rime splintering could thus be initiated upon first ascent of the cloud top through that zone and arguably contributed to the 1 L−1 or more graupel observed above it. Graupel ascending/descending into, or balanced within, the rime-splintering zone were found. In wider, less isolated clouds with dying updrafts and tops near −14°C, ice particle concentrations sometimes reached 100 L−1. Future 3D numerical modeling will be required to evaluate if rime splintering alone can explain the difference of three to four orders of magnitude in the observed INPs and the graupel observed at −5°C and colder.

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1. Introduction

The Ice in Clouds Experiment–Tropical (ICE-T) was a dual-aircraft study (including the NSF/NCAR C-130 and the SPEC, Inc. Lear jet) based out of St. Croix, U.S. Virgin Islands, during July 2011. The objectives pertinent to this study included (i) documenting the evolution of ice in maritime cumulus with cloud-top temperatures warmer than −10°C; (ii) characterizing the aerosol particles potentially acting as ice-nucleating particles (INPs); (iii) determining if primary nucleation can explain the observed ice development, or if any secondary ice formation process(es) is (are) critical to the glaciation of the clouds; and (iv) determining the amount of INPs sufficient to trigger the secondary ice formation process(es).

These objectives grew from the longstanding challenge of reconciling the hundreds per liter of ice particles sometimes observed in maritime clouds at temperatures greater than −10°C with the measured (or expected) amount of INPs active at those temperatures. INP concentrations measured over remote oceans or scaled from laboratory measurements of sea spray aerosols tend to fall in the range of a few thousandths per liter at −10°C (Bigg 1973; DeMott et al. 2016).

Uncertainties regarding both ice and INP observations can obfuscate their relationship. Ice particles remaining from nearby decaying clouds, or falling from clouds overhead, can “seed” newer clouds and be an additional ice source. The sampling aircraft may also seed the cloud through aircraft-produced ice particles (APIPs)—ice nucleated by the local cooling produced by aircraft propellers (e.g., Rangno and Hobbs 1983; Heymsfield et al. 2010). The sampling distance of an aircraft beneath a quickly growing cloud top can be consistent with active rime splintering. A later study of Florida cumuli by Hallett et al. (1978) supported rime splintering as an explanation for ice particles exceeding tens per liter for cloud-top temperatures greater than −9°C, but subsequent passes through clouds were often made (possibly contributing to APIPs), and the clouds had characteristics of ingesting continental air [droplet number concentrations ranged from 400 up to 1000 cm\(^{-3}\), and airborne measurements of INPs were 10 times greater than those of Fletcher (1962), at −20°C].

Studies by Hobbs and Rangno (1985, 1990) and Rangno and Hobbs (1991) found high numbers of ice crystals in shallow polar maritime cumuli off the coast of Washington state and in deeper cumuli over Kwajalein Atoll (Rangno and Hobbs 2005) but were pessimistic about the importance of rime splintering. Hobbs and Rangno (1985) claimed those clouds lacked the smaller droplets necessary for rime splintering; their later studies estimated ice production rates by rime splintering as too low to explain the observations. Rangno and Hobbs (2005) suggested rime splintering and shattering of freezing raindrops combined might explain the high ice concentrations they found in tropical Pacific cumuli. All their studies lacked in situ cloud vertical velocity measurements, however—important for understanding the stage of cloud development and the transport of liquid and ice particles. Huang et al. (2008) observed ice crystal concentrations of 70 L\(^{-1}\) at −10°C and warmer in a cumulus with low drop concentrations sampled multiple times over southwest England, and estimated rime splintering could explain their development. That study also included axisymmetric
numerical modeling that, like some other numerical modeling studies (e.g., Ovtchinnikov et al. 2000; Phillips et al. 2001; Sun et al. 2010), emphasized the importance of early supercooled raindrop formation in producing the graupel needed to initiate rime splintering more quickly. Curiously, no supercooled raindrops were found in Huang et al.’s aircraft measurements. Hallett et al. (1978) did measure supercooled raindrop concentrations of 1 L$^{-1}$ that they argued were critical to graupel formation and, thus, rime splintering. They also postulated that a downward flux of graupel in regions of low vertical velocity, and an upward flux of graupel in regions of high vertical velocity, could have been enhancing the production of splinters and subsequent graupel in their observed cases.

Two more recent studies from the ICE-T field campaign have also reached conflicting conclusions regarding the importance of rime splintering. Heymsfield and Willis (2014, hereafter HW14) conducted a statistical study using probe data from all the C-130 flights with cloud sampling at $-15^\circ$C and warmer and showed that the conditions for rime splintering were often met in the ICE-T clouds: smaller (<12 μm) and larger (>25 μm) cloud drops within the temperature zone from $-3^\circ$ to $-8^\circ$C, and graupel number concentrations around 1 L$^{-1}$ at the top of that zone. They also identified low liquid water content—low-vertical-velocity locations in the clouds with an abundance of needles. They found very limited evidence of shattering from freezing raindrops. Lawson et al. (2015, hereafter L15), on the other hand, focused on the probe data collected by the Lear jet that sampled ICE-T clouds multiple times as their tops cooled below $-20^\circ$C. Within stronger and deeper clouds than those summarized by HW14, they found a “rapid-glaciation zone” between $-12^\circ$ and $-20^\circ$C where hundreds of ice particles per liter were found. They did not attribute these ice particles to rime splintering because the particles were found at lower temperatures (outside the rime-splintering zone), were produced in clouds with strong updrafts, and had habits other than columns. Instead, noting an abundance of irregular ice in the CPI images, they favored the shattering of freezing raindrops as an explanation. Their modeling exercise of a rising parcel allowing freezing raindrops to shatter could produce ice rapidly at these lower temperatures, with rime splintering having little effect, if the number of ice fragments produced per shattering event were allowed to be very high (sometimes over several thousand). Such shattering rates have yet to be demonstrated in laboratory studies.

The current study investigates the rime-splintering process as a candidate for explaining high ice number concentrations in the ICE-T clouds, presenting dynamical information absent in HW14 or L15, by including data from the Wyoming Cloud Radar (WCR). The WCR can document the stage of cloud development (including the height of cloud top) during an aircraft pass, has a larger sample volume than the microphysical probes, and detects particle net velocities so that particle transport can be inferred. For the cloud passes analyzed here, the aircraft sampling strategy was to pass through developing clouds near their ascending tops to assess the early development of ice. These cases tend toward the more vigorous clouds depicted in HW14’s analysis, but are shallower, and have weaker updrafts than those studied by L15. INPs measured during ICE-T are also presented and illustrate the extent to which secondary ice processes may have been active. Manual evaluation of particle images was performed, rather than using automated algorithms, to discriminate between ice and liquid particles, and allows the earliest detectable graupel to be found at smaller number concentrations than in HW14 and L15. Several examples of 100 L$^{-1}$ of ice crystals with cloud-top temperatures at $-14^\circ$C, corrected for possible shattering artifacts on the probes (not performed heretofore in other studies of rime splintering), lend confidence to the extent of secondary ice production occurring in the ICE-T clouds.

2. Data and analysis methods

During the ICE-T field campaign the C-130 aircraft was outfitted with a suite of aerosol and microphysical instrumentation, standard probes to measure winds and state parameters, the WCR, and Wyoming Cloud Lidar. Flights were conducted over the ocean either upstream or far enough downstream of islands to limit land-based aerosol influences. For the last three flights on 27, 28, and 30 July, single passes through the ascending cloud tops of isolated cumulus congestus were given priority to capitalize on the WCR data. (Delay in receiving a replacement part rendered the cloud radar inoperative after the first research flight, until the final third of the project.) The sampling technique did not attempt to track ascending parcels through multiple passes through the same cloud, as performed by L15, nor was it random to generate a population irrespective of cloud stage, as used by HW14. Instead, the strongest cells (as judged by appearance) ascending up to the particular flight level, typically near 0°, $-5^\circ$, or $-10^\circ$C, were chosen for sampling.

A major focus in this study is quantifying the earliest detectable ice particles from imaging probes and radar: those larger than 200-μm diameter and being irregular (i.e., noncircular) in shape. For brevity, we call these particles “graupel,” being indistinct in habit and having substantial size to effectively rime cloud drops and raindrops. They likely consist of rimed frozen raindrops.
and/or higher-density graupel, given the very wet conditions in the sampled clouds as described in section 3.

a. Wyoming Cloud Radar data

The WCR is a 95-GHz, dual-channel Doppler radar designed for studies of cloud microphysics and dynamics and is regularly deployed on board the C-130 or the University of Wyoming King Air (Leon et al. 2006; Wang et al. 2012). For ICE-T, it was mounted on the rear ramp of the C-130 and used a pair of downward-looking antennas (one nadir, the other 34° aft of nadir), supplemented by a single upward-looking antenna mounted approximately 6 m forward of the downward-looking antennas.

When both ice and warm rain processes are active as in the ICE-T clouds, high radar reflectivity can result from large irregular particles (graupel/rimed frozen raindrops), rain or drizzle-sized liquid drops, or a combination of the two. Thus, radar reflectivity alone is not sufficient to distinguish between frozen and liquid hydrometeors. For ICE-T, the downward-slanted antenna was set up to transmit and receive two orthogonal linear polarizations. Nonspherical particles (i.e., ice) scatter some component perpendicular to the polarization of the transmitted signal. In regions of uniform particle type, the ratio of the cross-polarized to the copolarized return, expressed as the linear depolarization ratio (LDR), is governed by the hydrometeor type (e.g., Matrosov et al. 2001). Unfortunately, numerical simulations of scattering from graupel at 95 GHz are not widely available for determining the corresponding value of LDR. Thus, an empirical value of LDR (−18 dB) is used, consistent with that reported by Wolde and Vali (2001a, b) and observations in regions dominated by graupel in the ICE-T clouds. It is possible, however, that recently frozen raindrops do not produce the same LDR as heavily rimed graupel, thereby contributing to uncertainty in the resulting estimates. But as noted for example by Blyth and Latham (1993), recently frozen raindrops will rime quickly in high LWC conditions such as those examined here and become more irregular in shape.

When a mix of hydrometeor types is present, the LDR results from the cross-polarized return from different particles types weighted by their contribution to the total (copolarized) reflectivity. In regions consisting of a mix of ice particles and raindrops, LDR is decreased by the presence of the raindrops, which contribute to the copolarized radar reflectivity but not the cross-polarized return. Thus, the observed LDR can be expressed as

\[ \text{LDR}_{\text{obs}} = \text{LDR}_g + (Z_g - Z_{\text{tot}}), \]

where \( Z_g \) is the reflectivity contribution from graupel, \( Z_{\text{tot}} \) is the total reflectivity (both in dBZ), and LDR\(_g\) is the expected value based on observations of graupel (−18 dB). As \( Z_{\text{tot}} \) will be significantly greater than \( Z_g \) where the mixed-phase cloud is still dominated by supercooled raindrops, the LDR\(_{\text{obs}}\) will be significantly reduced from LDR\(_g\).

The −35 dB cross-channel isolation of the downslant WCR antenna allows detectable LDR down to at least −38 dB, and possibly even to −40 dB, following correction for cross-channel leakage when the cochannel return is sufficiently strong. In cases where the cross-polarized return does not exceed the noise in that channel by at least one standard deviation, LDR cannot be measured directly, but it can be constrained by the LDR corresponding to a cross-polarized return one standard deviation above the noise in that channel. The actual LDR could be lower than the maximum LDR inferred, but not higher, as a higher LDR would produce a detectable signal. This constraint is particularly useful when the cross-polarized return is large, and therefore the maximum LDR is small enough (i.e., −38 dB or below) that it can be used to rule out contributions from graupel.

Only LDR values within a few range gates below the aircraft are used; farther along the radar beam, multiple scattering produces artificially inflated LDR values. The use of values as low as −38 dB to indicate the presence of small amounts of ice within a population of raindrops was developed by comparison with the manual identification of large ice particles in the ICE-T 2D-C aircraft images, which are nearly centered within the WCR dead zone extending approximately 100 m above and below the aircraft. Given the 20-dB difference between the LDR value observed for all graupel (−18 dB) and the approximate minimum detectible LDR of −38 dB, graupel must only contribute 1% of the total reflectivity for the resulting LDR to be detectible.

The translation of this minimum 1% of reflectivity into the fraction of particles that must be frozen to have a detectable LDR depends strongly upon particle size, as a result of the different Mie scattering curves for ice versus water [cf. Fig. 9 of Lhermitte (1990)]. Thus, estimating the percentage of particles frozen requires additional information about the size distribution, here supplied by an optical array probe (2D-P) that senses the largest particles (from 300 µm to 1 cm, ignoring data in the first bin). For the cloud passes where LDR was used to estimate graupel number concentrations, 0.6-mm-diameter raindrops dominated the WCR reflectivity; a sharp decrease in frequency of larger particles was clear from the 2D-P data. For a population of raindrops less than 1.5-mm diameter, the reflectivity is reduced nearly 7 dB upon freezing of the entire population owing.
to the change in refractive index from water to ice. Hence, based on the Mie scattering curves, 5% of the raindrops (1.5 mm or smaller) must be frozen for 1% of the reflectivity to be due to graupel and thus yield a detectable LDR.

b. Microphysical probe data

An assortment of microphysical probes was mounted on the C-130 aircraft to document cloud droplets, raindrops, and small and large ice particles (HW14). Those used here include a FSSP measuring cloud droplets of 2–47-μm in diameter, a 25-μm-resolution “fast” two-dimensional cloud probe (2D-C), and a 150-μm-resolution “fast” two-dimensional precipitation probe (2D-P). The Small Ice Detector-2 High-Performance Instrumented Airborne Platform for Environmental Research (HIAPER) (SID-2H) was intended to detect smaller ice particles (2–50 μm); it counts and sizes these particles and assigns a habit (spherical, plate, column, or irregular) based on the scattering patterns of illuminated particles as they pass through the probe (e.g., Hirst et al. 2001). This probe had performance difficulties in the extremely high LWC of the ICE-T clouds, however, and the related sampling-volume uncertainties prevented the quantification of number concentrations (Johnson et al. 2014). Despite these issues, determining the presence of small ice crystals (as determined by the occurrence of irregular scattering patterns by the probe) was robust and is used here when pertinent.

1) Rain rate and rain water content estimates

Rainwater contents and rain rates for cloud passes at 0°C near the cloud tops were estimated from the raindrop sizes and number concentrations at the flight altitude using the 2D-C or 2D-P number concentrations and diameters (values for sizes between 100 and 500 μm were supplied by the 2D-C, and for all larger raindrops by the 2D-P, which has a larger sample volume). Fall speeds were those from Beard (1976), adjusted for atmospheric pressure. (Since these are fall speeds with respect to still air, the resulting fluxes are with respect to an air parcel.) Only images characterized as “round” were used in these computations to prevent the usage of images affected by splashing.

2) Manual estimates of the graupel

Because the discrimination between supercooled raindrops and graupel can be quite subtle, manual identification of ice particles from the 2D-C shadow images was performed. Each cloud pass was divided into subsections judged to be relatively homogeneous, based on the number, size, and shape of the images, as well as the presence of updraft or downdraft. Particles were then counted within smaller regions of those subsections, with the fraction of those images judged as ice later applied to the entire subsection, using the total particle count for particles greater than 150 μm from the 2D-C.1 Several separate graupel number concentration estimates within a single cloud pass resulted.

This method is designed to provide a conservative quantitative estimate of the graupel in a given cloud pass. Shadow images that were round were assumed to be unfrozen; the high LWC would promote the quick capture of supercooled cloud droplets by any frozen drop and quickly deform its shape from circular. Any nonround particle was counted as ice, unless it was clear from the image that a stuck or malfunctioning bit was making it irregular or if the image was out of focus (large center spot clearly visible). The smallest images scrutinized were 200 μm; images having fewer pixels were too difficult to identify as round. If the image inspector was in doubt, it was not counted as ice.

Because this type of analysis is inherently subjective, a method was devised to “train” the inspectors for consistency. The group of inspectors first was given individually the same time periods and images and asked to count ice particles; they then came together as a group to compare their ice counts and discuss their reasoning. Often the inspectors were in agreement, but for some images they were not, and this is a source of uncertainty quantified in the results. The inspectors were also tested with three different time periods at temperatures much above 0°C (so that all images were guaranteed to be liquid) but were told that the cloud passes were at subfreezing temperatures. None of the inspectors labeled any images as ice, providing more confidence in their ability to discriminate irregular images of frozen particles from circular images of unfrozen particles.

The inspectors then counted ice images over the same time periods for several cloud passes. Each cloud section contained several hundred to over 25,000 images, with the mean being just over 9900. The number of images identified as ice was very small compared to the total number of images for each cloud section, ranging from 0% to 0.5%. In general, the four different inspectors converged upon very similar results, with most estimates lying within 0.3% of each other. For the particular cloud passes where several inspectors scrutinized the exact same time periods and gave independent estimates, maximum and minimum numbers of graupel are presented as uncertainty bars in the results.

1 Ice particle estimates from the 2D-P were made using the WCR data as described in section 2a, as its image resolution is not enough for an inspector to discriminate between round and irregular particles when the number of pixels is small (less than four or five).
3) Estimates of High Number Concentrations of Ice

Multiple cloud passes appeared to contain high numbers of ice particles (exceeding 100 L\(^{-1}\)), as also found by HW14 and L15. Here, to locate samples containing only ice, the 2D-C images were manually inspected to find periods when all images were irregular shaped; these included a range of habit types and rimed particles. Because it is impossible to discriminate between liquid drops and ice particles for the smallest images, the Rosemount Icing Detector was used to eliminate times when supercooled water was present (e.g., Baumgardner and Rodi 1989). Total ice particle counts from the 2D-C probe were then limited to images exceeding 100 \(\mu m\), as the quantification of smaller particles is much less reliable (e.g., Korolev et al. 1998).

It is now widely accepted that optical array probe estimates of ice number concentrations can be contaminated by ice shattering on the leading tips of the probe; particles smaller than 500 \(\mu m\) must be carefully inspected to ensure that they are not artifacts (e.g., Korolev et al. 2013b). Some probes have been fitted with antishattering tips (Korolev et al. 2013a) that partly alleviate the problem (Jackson and McFarquhar 2014; Jackson et al. 2014) but these were not available for ICE-T. However, many of the shattering artifacts can be removed in postanalysis using particle interarrival times (Field et al. 2003, 2006); the artifacts usually have a shorter time difference between successive images than for real ice particles. The cloud passes presented here did not appear to be dominated by shattering effects, but there was a small secondary peak at very short interarrival times possibly indicative of shattering artifacts. Following Korolev and Field (2015), the minimum value at the start of the main (second) peak of values in the interarrival time analysis was chosen as the cutoff time to discriminate between the unshattered particles and the shattering artifacts. Those particles with smaller interarrival times than the cutoff were removed from the total, typically 10% or less, for the cases presented.

c. Ice-nucleating particle data

Multiple INP measurement methods were used during ICE-T, but the only method that provided data at temperatures exceeding \(-16^\circ C\) was an offline immersion freezing method (DeMott et al. 2016; G. R. McMeeking et al. 2016, unpublished manuscript). Filter-collected particles were analyzed for freezing following rinsing from the filters and distribution as aliquots for cooling using the device and methods described by Hill et al. (2014). Filters were exposed for 900–3600 s at a sample flow rate of 10 L min\(^{-1}\) during level C-130 flight legs in clear air at various altitudes (DeMott et al. 2016; G. R. McMeeking et al. 2016, unpublished manuscript). Higher-volume filter samples were collected (typically 4 h, and up to 60 h) at coastal and mountaintop sites in Puerto Rico and during onshore flow at a coastal site in St. Croix.

Figure 1 is a compilation of INPs active at temperatures greater than \(-20^\circ C\) during the entire study period. Samples from 2 of the 3 days of interest here are highlighted; both were collected at 1051 m in Puerto Rico, which is at or above the MBL top at various times. INP concentrations ranged from just slightly greater than the Fletcher (1962) parameterization (solid line) and a curve for one-tenth the standard Fletcher values (dashed line).

3. Analysis results

a. Cloud characteristics for 27, 28, and 30 July

Cloud bases were typically around 700 m, at pressures near 940 hPa and temperatures near 23°C, with FSSP-measured droplet concentrations from 50 to 100 cm\(^{-3}\).
Typical heights of the 0°C, −5°C, and −10°C levels were near 4800, 5600, and 6500 m, respectively, indicative of the large cloud depth over which the warm rain process could act. Vertical wind shear was weak on these days; winds near the cloud bases were easterly at 10 m s\(^{-1}\), decreasing to 7 m s\(^{-1}\) by the −15°C level. Maximum cloud-top heights varied, but clouds analyzed here were limited to −14°C (~7400 m) and warmer. To be included in the analysis, cloud passes had to be within 1 km of an ascending cloud top, characterized as having a broad region of aircraft-measured positive vertical velocity, and appearing “sharp” against the environmental air. The analyzed clouds also had no taller clouds nearby that could introduce ice into the sampled cloud from higher levels. Passes through an individual cloud were made once to prevent possible contamination by APIPS.

b. Abundant raindrops upon cloud-top ascent to 0°C

Given the warm cloud bases and the low droplet number concentrations, a rapid, strong warm rain process (relative to colder-based or continental clouds) was occurring. The adiabatic liquid water content was 7 g m\(^{-3}\) at the 0°C level. An example of measurements at the 0°C level near cloud top is shown in Fig. 2. Cloud droplets 25-μm diameter and larger were prevalent, in this example exceeding 3 cm\(^{-3}\), and were ascending within the updraft (maximum speed of 7 m s\(^{-1}\)). Raindrops of 100-μm diameter and larger exceeded 200 L m\(^{-3}\) in 1-s samples; those greater than 1 mm were present in amounts up to 5 L m\(^{-3}\). A comparison of the aircraft-measured vertical velocity and the radar-sensed particle velocities above and below indicate that most of the raindrops were still ascending (tan pixels) in the stronger parts of the updraft at the flight altitude. The radar signal was completely attenuated in the updraft 1.5 km below the aircraft (not shown, but near 1720:30 UTC). The (parcel relative) rain rate estimated from the 2D probe data was 10 mm h\(^{-1}\), and the maximum rainwater content (RWC) estimated from the 2D probes was near 1 g m\(^{-3}\).

The case shown in Fig. 2 lies toward the lower end of the range of rain development among the 0°C cloud passes\(^2\) analyzed (Table 1). Maximum updraft speeds ranged from 4 to 15 m s\(^{-1}\), most capable of suspending raindrops up to 1 mm in diameter, and often larger. Raindrop number concentrations typically exceeded hundreds per liter and showed no relationship with the distance of sampling from the cloud top. Rainwater contents indicated that over 10% of the adiabatic cloud water had often been converted to rain. The rain rates calculated from these same raindrop size distributions usually exceeded 25 mm h\(^{-1}\) and sometimes even 100 mm h\(^{-1}\).

Thus, by the time the cloud tops had cooled below 0°C, ingredients favoring the rime-splitting mechanism were already present within the updrafts: cloud droplets exceeding 25 μm in diameter, and large raindrops that could initiate riming upon freezing, as also shown by HW14, but the values of LWC, updraft speed, and RWC presented here are at the extremes of their reported ranges, illustrating that the sampling strategy used here was effective in collecting data on the stronger clouds in earlier stages of development.

c. Graupel amounts and locations as cloud tops ascended through −3°C to −8°C and higher

A separate set of clouds than those discussed in the preceding section was used to examine the development of ice. Figure 3 shows a cloud pass at the −6°C level through an ascending turret with a minimum cloud-top temperature of −8°C (based on the sounding, although it was likely warmer because the cloud top was still ascending). Thus, the sampling level and cloud top (CT) were collocated in the rime-splitting zone at this time. The aircraft measured a maximum updraft speed just over 5 m s\(^{-1}\) and a narrow (<60 m) downdraft of −1 m s\(^{-1}\) right before exiting the cloud. The radar-measured particle motion above the aircraft was mostly positive but often less than half this value. Although LDR data were not available above the aircraft, the lower reflectivity in some of these ascending regions (particularly immediately above the aircraft from 1627: 27 to 16:27:28 UTC) may be due to graupel; images of the graupel were identified within the updraft (some are shown) ranging from 0.3 to 0.5 L m\(^{-3}\). The SIT-2H probe detected smaller columns and irregular particles (likely ice) 13–34 μm in diameter at the edges of the updraft. The radar data also showed descending particle motion (green pixels) in some regions immediately above and beneath the aircraft and at the cloud edges. Although the probe and radar data are not strictly collocated, the disparity of the sensed vertical motions likely indicates that large particles were sometimes descending in the weaker parts of the updraft as well as the downdraft. Some pixels above the aircraft in the higher reflectivity regions have near-zero vertical velocity, indicating that (at least temporarily) some large particles are being balanced in the updraft. Two tiny “ribbons” of

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\(^2\) No correction was applied to the cloud temperatures reported here for wetting of the temperature sensor. Limited comparisons with temperatures outside of cloud at equivalent altitudes in the sounding profiles were still 1°–2°C warmer, indicative of the buoyant updrafts sampled by the aircraft. The possibility exists that the in-cloud temperatures reported here are underestimated slightly but would not affect the outcomes of this study.
Fig. 2. Data from cloud pass at 0°C at 1720:25–1720:35 UTC 28 Jul. (top left) The 4-s average of cloud and raindrop size distributions sampled by the FSSP (red), 2D-C (blue), and 2D-P (maroon), with vertical red line marking 25-μm diameter and vertical blue line marking 100-μm diameter. (bottom left) Picture from forward-looking camera. (top right) WCR reflectivity above and below flight track (color scale; dBZ). Flight level shown by dashed line with asterisks, except where aircraft-measured vertical velocities are inset (range from −3 to 7 m s⁻¹ with zero marked by thick black line; labeled in 2-s intervals). (bottom right) Zoomed-in view of corresponding radar-sensed particle velocities, with same inset aircraft-measured vertical velocities.
higher LDR indicative of graupel are also evident immediately beneath the flight level (marked with black boxes). The rightmost box contains falling graupel as indicated by the radar-sensed velocity data. The top of the leftmost box, however, is in a region with near-zero radar-sensed vertical velocity; at this moment some graupel are balanced near the bottom of the rime-splintering zone. The maximum LDR in these ribbons within the first few range gates beneath the aircraft was approximately $-35\text{ dB}$, implying 10% of the larger particles (typically 0.6 mm in size as shown by the 2D-P) were frozen, equivalent to approximately $0.5\text{ L}^{-1}$, consistent with that estimated from the 2D-C images as well. The quick formation of this graupel is likely attributable to the supercooled raindrops produced at altitudes below the rime-splintering zone. The aircraft and radar data combined illustrate that graupel are slowly ascending, balanced within, or falling through different parts of the updraft, as well as descending in a cloud edge downdraft. Thus, the real setup for rime splintering in clouds can be quite complex.

Multiple examples of graupel ascending and descending into the rime-splintering zone were found. A set of clouds sampled within and above this temperature zone are listed in Table 2. Often clouds were connected near the bases but are considered “isolated” in that colder clouds (with potentially much more ice) were not nearby. However, in a few cases, new turrets sampled off to the side of older, taller turrets were included in Table 2 (and later in Fig. 5) as noted, if the taller turrets did not appear to overhang those sampled. Passes labeled “at CT” transected the “dome” of the cloud top, just skimming it at the entrance and exit regions, and were less than 200 m from the topmost center of the cloud as judged by the WCR images.

A cloud pass occurring at cloud top, after it had ascended through the rime-splintering zone, is shown in Fig. 4. The maximum temperature at the aircraft level was $-9^\circ\text{C}$. The aircraft-measured maximum updraft speed was nearly $12\text{ m s}^{-1}$, with a sharp $4\text{ m s}^{-1}$ downdraft occurring on the exiting edge at 1559:27 UTC. The particle motion indicated by the radar above the aircraft level was again substantially less, including a few pixels of downward motion, immediately above a region of weaker ($<4\text{ m s}^{-1}$) updraft sampled by the aircraft $-1559:23\text{ UTC}$. Beneath the aircraft, higher values of LDR are prevalent, although no regions appear completely glaciated (i.e., LDR $=-18\text{ dB}$), and many areas are still dominated by supercooled raindrops (darker pixels). Three broader ribbons of higher LDR indicative of some ice are collocated with positive particle vertical velocities near 1559:20, 1559:22–1559:24, and 1559:25–1559:26 UTC (black boxes). The maximum ice estimated from the manual analysis of probe images was $1\text{ L}^{-1}$. The maximum LDR in the first few range gates beneath the aircraft was approximately $-33\text{ dB}$, implying 3% of the reflectivity was from frozen particles, corresponding to 15% of the particles being frozen, or around $0.3\text{ L}^{-1}$, given that particles less than 1-mm diameter dominate the reflectivity here. A region of particles barely ascending, or even balanced in the updraft, can be found at 1559:22 UTC, while near 1559:18 UTC (white box) some graupel and supercooled raindrops are descending back into the rime-splintering zone at a rate near $5\text{ m s}^{-1}$. Thus, even after the cloud top rose above the rime-splintering temperature zone, graupel (and supercooled drops) were continuing to ascend into it, some were descending through it, and others were temporarily balanced within it. The ribbons

### Table 1. Cloud pass summary for active clouds sampled near 0°C for all three ICE-T flights.

<table>
<thead>
<tr>
<th>Flight date</th>
<th>Time (UTC)</th>
<th>Max vertical velocity (m s$^{-1}$)</th>
<th>Sampling distance from cloud top (m)</th>
<th>Max 2D-C particle concentrations (L$^{-1}$) $\geq$ 100 $\mu$m</th>
<th>Max 2D-P particle concentrations (L$^{-1}$) $\geq$ 300 $\mu$m</th>
<th>Max RWC (g m$^{-3}$)</th>
<th>Max derived rain rate (mm h$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>27 Jul</td>
<td>1509:29–1510:07</td>
<td>7</td>
<td>1500–2000</td>
<td>78</td>
<td>3</td>
<td>1.1</td>
<td>25</td>
</tr>
<tr>
<td></td>
<td>1915:00–1915:57</td>
<td>13</td>
<td>&lt;1000</td>
<td>319</td>
<td>10</td>
<td>4.9</td>
<td>109</td>
</tr>
<tr>
<td></td>
<td>1643:12–1643:34</td>
<td>6</td>
<td>1500–2000</td>
<td>160</td>
<td>6</td>
<td>3.7</td>
<td>87</td>
</tr>
<tr>
<td>28 Jul</td>
<td>1720:25–1720:35</td>
<td>7</td>
<td>At CT</td>
<td>248</td>
<td>5</td>
<td>1.0</td>
<td>10</td>
</tr>
<tr>
<td></td>
<td>1722:05–1722:25</td>
<td>6</td>
<td>&gt;1500</td>
<td>189</td>
<td>3</td>
<td>1.9</td>
<td>42</td>
</tr>
<tr>
<td></td>
<td>1900:44–1900:56</td>
<td>4</td>
<td>&lt;1000</td>
<td>129</td>
<td>5</td>
<td>0.8</td>
<td>11</td>
</tr>
<tr>
<td></td>
<td>1905:30–1906:00</td>
<td>15</td>
<td>&gt;1500$^a$</td>
<td>375</td>
<td>11</td>
<td>5.8</td>
<td>148</td>
</tr>
<tr>
<td>30 Jul</td>
<td>1555:17–1555:36</td>
<td>5</td>
<td>1000</td>
<td>277</td>
<td>4</td>
<td>1.5</td>
<td>29</td>
</tr>
<tr>
<td></td>
<td>1608:23–1609:11</td>
<td>12</td>
<td>&lt;500</td>
<td>314</td>
<td>8</td>
<td>4.1</td>
<td>105</td>
</tr>
<tr>
<td></td>
<td>1609:22–1609:35</td>
<td>8</td>
<td>&lt;1000</td>
<td>245</td>
<td>6</td>
<td>2.1</td>
<td>45</td>
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<tr>
<td></td>
<td>1841:21–1841:38</td>
<td>6</td>
<td>&gt;1500</td>
<td>95</td>
<td>6</td>
<td>3.5</td>
<td>88</td>
</tr>
<tr>
<td></td>
<td>1846:12–1846:22</td>
<td>7</td>
<td>At CT</td>
<td>126</td>
<td>4</td>
<td>1.6</td>
<td>31</td>
</tr>
</tbody>
</table>

$^a$ Cloud top uncertain owing to substantial attenuation of radar signal overhead.
FIG. 3. Data from cloud pass at 1627:25–1627:29 UTC 30 Jul at −6°C. (top) Radar reflectivity and radar-sensed particle velocities as in Fig. 2, with aircraft-measured vertical velocity inset (from −1 to 6 m s⁻¹ with zero marked by thick black line; labeled in 1-s intervals). (bottom left) Corresponding LDR (color scale; dB) below the flight level, with reflectivity superimposed above the flight level for reference, and same inset of aircraft-measured vertical velocity. Also shown are examples of graupel images (mm) from the 2D-C shown where observed at the flight level.
of LDR indicating graupel are only \( \sim 100–300 \) m wide. Similar short bursts of graupel were noted by Rangno and Hobbs (2005) in deeper maritime cumuli in the tropical Pacific.

In Fig. 5, the estimated graupel concentrations for the passes listed in Table 2, derived from manual inspection of the 2D-C images, are shown as a function of sampling temperature and color coded according to the sampling distance from cloud top. For those times when multiple inspectors counted the number of graupel, error bars denote the minimum and maximum estimates from the median value. Estimates from turrets growing alongside other clouds are noted but seem consistent with the more isolated cases. More confidence is placed in the estimates derived from sampling near or within 500 m of cloud top, where new ice nucleation above the sampling altitude would be limited.

Figure 5 shows that increasingly more graupel were observed as the sampling temperature decreased, most being on the order of 0.1 \( \text{L}^{-1} \) from \(-5^\circ\) to \(-7^\circ\)C (within the rime-splintering temperature zone) and on the order of 1 \( \text{L}^{-1} \) directly above. The lowest values are near the sampling volume limits of the 2D-C. All graupel estimates exceed the number of observed primary INP active at these temperatures by two orders of magnitude or more (diagonal black and gray lines). Some graupel is likely attributable to supercooled raindrops capturing newly emitted splinters, within and above the rime-splintering zone. (Splinters that are not immediately collected would soon be advected toward the top and edges of the updraft. HW14 noted large populations of columns occurring in areas of lower vertical velocity and LWC.)

A relationship between the manually estimated graupel and the collocated vertical velocity sampled by the aircraft also emerged. When the aircraft sampling altitude was within a few hundred meters of cloud top, the graupel were found more frequently in the updraft. The SID-2H probe also detected the smaller ice particles at the edges of the updraft or at the updraft–downdraft interface (Johnson et al. 2014). As the sampling distance increased to 500 m or more below the cloud top, the predominant location of the graupel shifted to downdrafts, especially when sampling more than 1000 m below the cloud top. This distribution of particles is likely attributable to their size sorting in the cloud, their location being a function of the particle fall speed and local cloud vertical velocity. HW14 found ice predominantly on the sides of the updraft or in the downdrafts in their statistical analysis of all the ICE-T data but were unable to categorize it by the sampling distance from the cloud tops using the radar data as done here and, thus, were unable to identify this shift.

d. Ice particles exceeding 100 \( \text{L}^{-1} \) for aged, colder cloud tops

Several cloud passes contained ice crystal number concentrations exceeding 100 \( \text{L}^{-1} \) as recorded by the 2D-C probe. These clouds were less isolated (two were embedded within multithermal complexes), were not necessarily sampled while the cloud top was ascending, and were wider (3 km or more), suggesting that high ice number concentrations require colder cloud tops, cloud interaction, more time, or a combination of these factors. Hallett et al. (1978) reported more graupel and columns in older, dying clouds than in fresher turrets; this dependency upon stage of cloud development is more easily documented with the WCR data. One-second segments from three different cloud passes were identified, as discussed in section 3c, as having large numbers of ice particles exceeding 100-\( \mu \text{m} \) diameter (Table 3). Other nearby samples often had higher number concentrations (several hundred per liter), but the Rosemount icing points

<table>
<thead>
<tr>
<th>Flight date</th>
<th>Time of cloud segment (UTC)</th>
<th>Aircraft-measured temperature (°C)</th>
<th>Aircraft-measured vertical velocity (m s(^{-1}))</th>
<th>Distance from cloud top (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>27 Jul</td>
<td>1525:07–1525:42(^a)</td>
<td>(-5)</td>
<td>(-0)</td>
<td>500 to 1000</td>
</tr>
<tr>
<td></td>
<td>1540:15–1540:28(^a)</td>
<td>(-6 to -5)</td>
<td>6 to 10</td>
<td>500 to 1000</td>
</tr>
<tr>
<td></td>
<td>1546:24–1546:35</td>
<td>(-10)</td>
<td>2 to 4</td>
<td>&lt;500</td>
</tr>
<tr>
<td></td>
<td>1559:13–1559:29</td>
<td>(-10 to -9)</td>
<td>(-2 to 12)</td>
<td>At CT</td>
</tr>
<tr>
<td>28 Jul</td>
<td>1515:17–1515:40</td>
<td>(-10)</td>
<td>(-1 to 5)</td>
<td>&lt;500</td>
</tr>
<tr>
<td></td>
<td>1522:48–1523:07</td>
<td>(-9 to -7)</td>
<td>1 to 15</td>
<td>500 to 1000</td>
</tr>
<tr>
<td></td>
<td>1740:28–1740:55</td>
<td>(-7 to -5)</td>
<td>4 to 9</td>
<td>&gt;1000</td>
</tr>
<tr>
<td></td>
<td>1833:01–1833:27</td>
<td>(-11 to -10)</td>
<td>0 to 4</td>
<td>&lt;500</td>
</tr>
<tr>
<td></td>
<td>1839:50–1840:20</td>
<td>(-11)</td>
<td>(-4)</td>
<td>500 to 1000</td>
</tr>
<tr>
<td></td>
<td>1855:04–1855:20(^a)</td>
<td>(-7 to -6)</td>
<td>(-1 to 7)</td>
<td>&gt;1000</td>
</tr>
<tr>
<td>30 Jul</td>
<td>1627:25–1627:29</td>
<td>(-6)</td>
<td>(-4 to 4)</td>
<td>&lt;500</td>
</tr>
<tr>
<td></td>
<td>1725:45–1726:00(^a)</td>
<td>(-12 to -11)</td>
<td>0 to 7</td>
<td>At CT</td>
</tr>
</tbody>
</table>

\(^a\) Possible “seeding” from taller turret off to side.
Fig. 4. As in Fig. 3, but for cloud pass at 1559:13–1559:29 UTC 27 Jul at −9°C. Inset aircraft-measured vertical velocity ranges from −4 to 12 m s⁻¹ with zero marked by thick black line; time axis labeled in 2-s intervals. Black and white boxes on velocity and LDR plots surround features described in the text.
detector showed evidence of supercooled drops being present. Cloud-top temperatures above the sampling altitude were estimated using the radar data and assuming an internal cloud lapse rate of 5°C km⁻¹. Table 3 shows the particle number concentrations before and after the application of the correction algorithm for possible shattering effects. The changes are minor but remove some uncertainty as compared to previous studies where such corrections were not known or attempted. All these cases were near or had achieved their maximum cloud-top heights, as gauged by the low vertical velocities and the general collapse of the reflectivity maxima as seen on the radar.

In the first case listed in Table 3, sampled directly above the rime-splintering zone, the 2D-C images were dominated by large ice and no images of columns were evident (Fig. 6). Presumably any columns grown from splinters at lower levels had collided with supercooled raindrops, or rimed the abundant cloud water up to this level, to comprise the 101-L⁻¹ particles detected. The radar-detected velocities (not shown) indicate that particles immediately above the aircraft at this location are still ascending, while below the aircraft they are descending. The LDR data immediately beneath the aircraft level indicate some ice is always present there, and values near −18 dB occur from 1838:21.5 to 1838:22.8 UTC. Multiple “pockets” less than 100 m wide of ascending particles still exist above the aircraft up to cloud top from 1838:20 to 1838:30 UTC; the splinters/columns might be yet supported by the updraft at those locations while the heavier particles are falling through it.

In contrast, the last two passes listed in Table 3 occurred in the center of the rime-splintering temperature zone (with colder cloud tops farther overhead) and exhibited a plethora of columns as well as graupel in the 2D-C images; one case is shown in Fig. 7. Here in a dying turret (aircraft-measured zero vertical velocity) within a wider cloud, there is evidence of new ice particle formation as gauged by the large number of unrimed columns in the 2D-C images. Broad regions of LDR immediately beneath the aircraft have values around −18 dB, indicative of complete glaciation, and contain descending particles (not shown). Thus as the updraft dies, graupel above the rime-splintering zone fall back through it and can produce a burst of new splinters (provided cloud droplets for riming still exist; the Rosemount probe indicates they were in this case). These splinters, suspended in regions of near-zero vertical velocity, can grow into columns large enough to be detected by the 2D-C, consistent with the locations of abundant needles found by HW14.

4. Summary and discussion

Aircraft observations from three flights of the ICE-T field campaign focused on sampling the microphysical evolution near ascending cloud tops as they cooled from...
Fig. 6. (top) WCR reflectivity and (bottom) LDR (below flight track) for cloud pass on 28 Jul starting at 18:38:17 UTC at −10°C. (left) 2D-C images shown at 18:38:23 UTC, when the maximum measured ice crystal number concentration was 101 L$^{-3}$. 
Fig. 7. As in Fig. 6, but for cloud pass starting at 1816:45 UTC 28 Jul at −6°C. 2D-C images shown at 1817:04 UTC, at location when the maximum ice crystal number concentration was 126 L⁻¹.
In broader, multithermal clouds approaching or having reached their maximum tops at \(-14^\circ\text{C}\) as gauged by decaying updrafts, ice crystal number concentrations (corrected for possible shattering effects on the probe) exceeded \(100 \text{ L}^{-1}\). In a case sampled above the rime-splintering zone, graupel were still observed ascending near cloud top as well as falling back into the rime-splintering zone. In other similar cases sampled within the rime-splintering zone, abundant new (unrimed) columns were found, likely formed by rime splintering of the graupel observed falling back through the zone amidst the remaining supercooled water. These additional ice particles, suspended or ascending in the weakening updrafts, might explain the high ice number concentrations observed.

The complex picture portrayed by this observational analysis argues for detailed, 3D numerical modeling to understand if rime splintering alone can explain the differences in the observed INP, graupel, and total ice number concentrations. The enhanced documentation of cloud-top heights/temperatures, stage of cloud development, updraft speeds, and aircraft sampling heights/temperatures, as well as careful, conservative estimates of early graupel number concentrations supply "quantitative targets" for future numerical modeling studies and potentially for future field and laboratory investigations searching for other possible maritime sources of INPs active at these higher temperatures.

These findings appear applicable to clouds of similar depths and strengths observed during ICE-T, as they do not conflict with the statistical study performed upon the entire C-130 ICE-T dataset by HW14. They will not necessarily be universal to all maritime cumuli. The ICE-T clouds analyzed here had low, warm bases, occurred in low-shear environments, and had sufficiently strong updrafts, producing more raindrops early and transporting them above the \(-6^\circ\text{C}\) level. Maritime clouds in cooler, less unstable, higher-shear, and/or locally polluted environments may be incapable of doing so.

This study raises additional questions and identifies critical areas for future work. The most immediate concern is that the measured INPs active near \(-5^\circ\text{C}\) were three to four orders of magnitude fewer than the graupel observed as the cloud tops first ascended through the \(-5^\circ\text{C}\) level. Hallett et al. (1978) found similar numbers of graupel particles in developing clouds but lacked measurements of INPs active at higher temperatures for comparison. Consideration of ice particle transport (and possible clustering) from the 3D dynamics of a cumulus cloud, while rime splintering is active, is needed to understand the true extent of the discrepancy, if any exists. Additional secondary ice formation processes active at \(-5^\circ\text{C}\), such as that proposed by Knight (2012), might be key. Raindrop shattering upon freezing, if it is more active at higher temperatures than currently thought, might also be contributing to the graupel. Other sources of INPs not detectable with current observational methods are also possible. It was unfortunate that smaller ice could not be quantified with the SID-2H probe as intended; such measurements would have been useful in resolving this issue. Future adaptations to it, or the creation of new instrumentation, must yet meet the challenge of quantifying smaller ice particles in mixed-phase clouds containing abundant supercooled water like those sampled in ICE-T.

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