The liquid–ice mass partitioning in tropical maritime convective clouds is studied using data collected by the National Center for Atmospheric Research C-130 research aircraft during the Ice in Clouds Experiment–Tropical project. The clouds investigated by the C-130 in this study generally contained weak to moderate updrafts. The liquid water content (LWC) is calculated using a combination of hot-wire and imaging probes. The total condensed water content (CWC) is measured by a counterflow virtual impactor. The ice water content (IWC) is calculated as CWC minus LWC. Taking into account potential significant measurement uncertainties, the liquid fraction \[ \frac{\text{LWC}}{\text{IWC}} \] between 0\(^\circ\)C and 215\(^\circ\)C appears to decrease by a factor of about 3 in updrafts near (>500 m) cloud top and a factor of 2 in updrafts far below (>500 m) cloud top. The decrease in liquid fraction as a function of temperature is also correlated with cloud life cycle. In dissipating clouds, ice dominates in all temperature ranges. A comparison between this study and two parameterizations shows that at different geographic locations the liquid fraction in convective clouds differs. Because of the sampling bias and the limitations of instruments, more measurements, especially with more advanced instruments, are needed in the future.

1. Introduction

In nature, 40%–60% of clouds between 0\(^\circ\)C and −30\(^\circ\)C are mixed-phase clouds (Mazin 2006; Shupe et al. 2006). The coexistence of liquid and ice particles, as well as the various properties of ice crystals, has a significant impact on precipitation efficiency, cloud lifetime, and radiative properties (Raymond and Blyth 1992; Sun and Shine 1994; Bower et al. 1996). For those reasons, it is important to better understand the liquid–ice partition in mixed-phase clouds. To that end, many observational studies have been made, including in situ (e.g., Boudala et al. 2004; Noh et al. 2013) and remote sensing measurements (e.g., Wang et al. 2004; Shupe et al. 2006; Zhang et al. 2010). Most of the previous studies, however, have focused on stratiform mixed-phase cloud.

Convective mixed-phase clouds play important roles in hydrology and energy cycles, and their dynamical and microphysical properties are very different from those of stratiform clouds, including, for example, stronger updrafts, the existence of millimeter drops, higher ice concentration, and complex dynamics–microphysics interactions (Arakawa 2004; Lawson et al. 2015). Since 1950, in situ measurement has been the most reliable tool for studying the detailed microphysics of convective clouds (e.g., Warner 1955; Warner and Squires 1958), and many insights into convective cloud microphysical properties have been gained.

One of the most important microphysical properties of convective clouds is the much higher ice concentration than is found in stratiform clouds (e.g., Boudala et al. 2004; Murray et al. 2012; Heymsfield and Willis 2014; Lawson et al. 2015). Primary ice generation in convective clouds is thought to be related to the...
presence of large supercooled drops (e.g., Rangno and Hobbs 2005; Lawson et al. 2015). Koenig (1963) found that high concentrations of ice rapidly form within 5–10 min if millimeter supercooled drops exist in shallow summer cumulus. Hobbs and Rangno (1990) observed high concentrations of ice crystals in maritime cumulus with cloud tops warmer than −12°C in which drizzle-size supercooled drops existed. Recent studies have provided more observational evidence that high concentrations of primary ice crystals in convective clouds are related to the freezing of drizzle-size drops (e.g., Rangno and Hobbs 2005; Taylor et al. 2016) and raindrops (e.g., Zeng et al. 2001; Lawson et al. 2015). Secondary ice production mechanisms [the Hallett–Mossop (H–M) process, ice–ice collision, evaporation of single ice crystals, etc.] can lead to ice multiplication in convective clouds. For example, Mossop (1970) reported that the ice concentration is 10^4 times larger than the primary ice nuclei (IN) concentration in maritime cumulus with cloud-top temperatures of about −8°C. Ono (1971) showed that ice concentrations are three or four orders of magnitude larger than the expected IN concentrations. Recently, Huang et al. (2011) observed high concentrations of pristine and rimed columns at a temperature of about −5°C in convective clouds from airborne measurements in southeastern Germany and eastern France. Heymsfield and Willis (2014) provided more evidence of high concentrations of secondary ice particles generated through H–M process in tropical convective clouds.

Rapid ice production in convective clouds can result in rapid increase in the ice water content (IWC) and significant depletion of the liquid water content (LWC). Bower et al. (1996) showed that the liquid fraction in maritime cumulus near the United Kingdom could be as low as 0.1 at −10°C. Crawford et al. (2012) found that the mean IWC in midlatitude cumulus increases from 0 to −0.25 g m^{-3} between 0° and −6°C, and the maximum IWC could reach 1.3 g m^{-3}. Such rapid IWC growth in convective clouds will result in significant changes of microphysics, dynamics, and radiative properties among different cloud life stages (Mathon and Laurent 2001; Wang et al. 2007; Braga and Vila 2014). However, observations related to our understanding of the liquid–ice mass partition in convective clouds still remain limited because of the limitations of the instruments.

Numerical simulation provides another way to study the liquid–ice partition in convective clouds. Traditionally, the LWC and IWC are simply parameterized as a function of temperature, and LWC is just assumed to exist at certain temperatures (e.g., Smith 1990; Boucher et al. 1995; Bower et al. 1996). However, these parameterizations have large differences. With the development of models, more studies try to prognose the LWC and IWC based on microphysical processes, rather than simply prescribing them as a function of temperature (e.g., Ghan et al. 1997; Rotstyan 1997; Morrison and Grabowski 2008). Studies show that the simulated convective cloud properties are very sensitive to the microphysics schemes. For example, Y. Wang et al. (2009) simulated tropical convective clouds using three sophisticated bulk microphysics schemes and found substantial discrepancies between the simulated and observed ice-phase microphysical properties. Fridlind et al. (2012) compared 10 cloud-resolving models (CRMs) with bulk microphysics schemes and showed that the IWCs from all the simulations were overestimated compared to the observations. Furthermore, an order of magnitude spread was found between the simulations. Other studies have shown that spectral bin microphysics schemes provided more realistic results for modeling convective clouds (e.g., Khain et al. 2004; Grabowski et al. 2015), but these studies lack sufficient evaluation of the detailed microphysics. Large discrepancies between the different modeling schemes again reveal the necessity of more in situ measurements in order to better understand the related microphysical processes and to improve the microphysics schemes for convective clouds.

The objective of this paper is to provide aircraft measurements of the liquid–ice mass partition in tropical maritime convective clouds, using the data collected during the Ice in Clouds Experiment–Tropical (ICE-T) project. Section 2 describes the details of the dataset and the instrumentation. Section 3 presents the analysis method. Section 4 gives the results, and the conclusions are presented in section 5.

2. Dataset

The data used in the present study were collected during ICE-T with airborne in situ and remote sensing instruments. This experiment was conducted during 1–30 July 2011 near St. Croix, U.S. Virgin Islands, with the aim of studying the role of ice generation in tropical maritime convective clouds.

Both the NSF/NCAR C-130 aircraft and the Stratton Park Engineering Company (SPEC), Inc., Learjet were used to penetrate convective clouds over the Caribbean Sea; this study uses only data collected by the C-130. In situ measurements from ICE-T include the liquid water content and total condensed water content (CWC), cloud and precipitating particle concentrations, temperatures, vertical velocities, and the properties of the particles with detailed images, as well as the particle size distributions. Thirteen C-130 research flights (RFs) were
made. The Wyoming Cloud Radar (WCR) (Wang et al. 2012) was operated on the C-130 during seven of those flights (RF07–RF13) to measure two-dimensional reflectivity and the Doppler velocity structures. In this study, we analyze the seven research flights with WCR measurements (RF07–RF13) because the WCR reflectivity images are needed to examine the cloud-top height.

Three main in situ instruments are used here to explore the liquid–ice mass partition: the counterflow virtual impactor (CVI), which measures the CWC; the King probe, which measures the LWC of droplets with diameter smaller than 25 μm; and the Particle Measuring Systems (PMS) two-dimensional cloud (2D-C) and two-dimensional precipitation (2D-P) probes, which record images of “entire in” and “center in” particles (Heymsfield and Parrish 1978) of approximately 25–3200 and 150–19200 μm in diameter, respectively. In this study, the LWC carried by drops larger than 25 μm in diameter is calculated based on the spheres identified from the 2D probes.

The CVI was developed to measure the CWC (Noone et al. 1988, 1993; Twyoh et al. 1997). Cloud particles of a sufficient size (>8 μm) enter the inlet against a counterflow of pressurized, initially dry, carrier air and travel into an evaporation chamber. Drops and ice crystals are evaporated within the inlet at a temperature of about 50°C, and the water vapor and nonvolatile residual nuclei remain after the evaporation of cloud particles is measured. The CVI in general accurately measures the CWC, with an uncertainty of less than 20% (Twyoh et al. 1997). There are some limitations, however. In ICE-T, although it has a detection limit as low as 0.01 g m⁻³, it tends to underestimate the CWC if the CWC exceeds 2.5 g m⁻³, which could be a problem under some conditions, especially in very strong updraft cores with high CWC. In the updrafts near cloud top sampled by the C-130, such high CWC was rarely derived. The Learjet had penetrations in stronger updraft cores with high LWC (Lawson et al. 2015), but the data are not used in this study. Another issue has to do with the hysteresis of this device, since the largest fraction of the water mass is typically carried in droplets with diameter larger than 10 μm. Additionally, the King probe may underestimate the contribution of drops with diameter larger than 30 μm as a result of incomplete evaporation and splashing. In this study, we use the King probe only to cover the LWC carried by droplets smaller than 25 μm in diameter.

The 2D probe is currently one of the main instruments used for particle spectra measurements in airborne studies of clouds. It is an optical array probe that records the two-dimensional images of cloud particles passing through its sample volume by using a focused He–Ne laser beam. The corresponding size distributions, concentrations, and liquid and/or ice water content can be estimated. The detailed optical and electronic principles of this device are fully described by Knollenberg (1970). The “fast” 2D-C probe and the “fast” 2D-P probe (with fast electronics) installed on the C-130 aircraft have resolutions of 25 and 150 μm, respectively. Both probes have 64 diodes in each linear diode array. After increasing the sample volume using the technique developed by Heymsfield and Parrish (1978), the 2D-C and 2D-P probes are able to record the shapes of the entire-in and center-in particles with diameters from 25 to 3200 μm and from 150 to 19200 μm, respectively. The sample volume of 2D probes for entire-in particles depends on true airspeed, depth of field, and width of diode array; for center-in particles the sample volume also depends on particle size. In the present study, we use the 2D probes to differentiate spheres from nonspheres on the basis of the shadow images in order to calculate the LWC carried by the drops larger than 25 μm in diameter. Because of the limitations of 2D probes (e.g., relatively poor resolution and artifacts in particle images), sometimes there could be large uncertainties in the identification of spheres and nonspheres. To better identify drops and ice, more advanced instruments, for example, a cloud particle imager (CPI; Lawson et al. 2006), are needed. But the CPI was not configured properly on the C-130 during the project, and 2D probes are more appropriate for statistical study. On 23 July (RF07), the 2D-P probe was not working well because of a hardware issue; we only use the 2D-C probe on this research flight. Analysis reveals that most of the large drops are smaller than 3 mm. The 2D-C probe covers a size range from 25 to 3200 μm, which includes most of the LWC, but the sample volume of 2D-C is not large.
enough to adequately sample particles larger than 3 mm, so the LWC in the clouds sampled on this research flight may be underestimated.

Several other measurements are also used to characterize the ambient conditions. The temperatures were measured using Rosemount sensors. Vertical air motions were acquired with accurate ($\pm 0.2 \text{ m s}^{-1}$) wind sensing elements. The linear depolarization measurements from the Wyoming Cloud Lidar (WCL) in upward- (RF01–RF13) and nadir- (RF06–RF13) viewing orientations were used to identify ice-dominated and liquid-dominated clouds (Z. Wang et al. 2009).

3. Analysis method

a. Cloud classification

In this study, we mainly focus on updrafts, which are defined as ascending cloud parcels with vertical velocity continuously $\geq 1 \text{ m s}^{-1}$ for at least 500 m (LeMone and Zipser 1980). So both relatively weak and strong updrafts are included based on this definition. In this study, most of the updrafts analyzed were weak to moderate compared to those with stronger updrafts (e.g., Lawson et al. 2015; Yang et al. 2016). We classify the sampled convective clouds into three categories based on the WCR reflectivity structures and in situ vertical velocity in terms of aircraft penetration level: 1) updrafts near (<500 m) cloud top, 2) updrafts far below (>500 m) cloud top, and 3) the dissipating clouds/dissipating part in mature clouds. These classifications reveal the stages of the cloud parcels in terms of fallout of ice: updrafts near cloud top are less affected by fallout of ice particles, while those far below cloud top are more affected by fallout of ice particles and thus stronger liquid–ice interactions (e.g., riming and accretion). The true cloud life stages need to be better examined using more advanced measurements (e.g., video tapes and high temporal and spatial resolution satellite measurements), which are not available in this study. The updrafts near cloud top sampled by the C-130 have scales of from a few hundred meters to a few kilometers. In most of them, the maximum vertical velocities and mean vertical velocities are weaker than 10 m s$^{-1}$, which are relatively weak compared to those sampled by the Learjet (Lawson et al. 2015). For the penetrations very close to the cloud top, the entrainment may reduce the LWC and further affect the ice generations. But in ICE-T, most of the penetrations are a few hundred meters below the cloud top. In dissipating clouds, the vertical velocities are weak (<1 m s$^{-1}$), and in many cases the WCR reflectivity is relatively low. Examples of the classification are presented in Fig. 1. For the updrafts near cloud top (red box), we can see vertical velocity up to 6 m s$^{-1}$ and small IWC near the cloud top, as confirmed by both in situ measurements and WCL signals. For the updraft far below clouds top (orange box), we can see a maximum vertical velocity of 3 m s$^{-1}$, and the cloud top is well above the flight level. The WCL depolarization ratio is much larger compared to that in the updraft near cloud top, which suggests the presence of a large quantity of ice particles. The in situ measurement shows the maximum IWC can reach 2 g m$^{-3}$, while the maximum LWC is only about 0.2 g m$^{-3}$. In the dissipating clouds and the dissipating part of the mature cloud (blue boxes), there is no strong updraft. The high WCL depolarization ratio and large IWC fraction suggest that ice dominates in these clouds. Table 1 shows the number of penetrations, the duration (s), and the flight lengths sampled at different temperature ranges in the three categories. In this study, we only analyze data collected in weak to moderate updrafts from seven research flights of the C-130. The results here cannot be applied to ICE-T convective clouds collected with more advanced cloud particle probes in strong updraft cores, such as those reported by Lawson et al. (2015). More measurements, especially with more advanced instruments, are needed in the future.

b. Drop and ice identification

The 2D probes have been widely used to distinguish liquid drops and ice particles in mixed-phase clouds (e.g., Holroyd 1987; Czys and Petersen 1992; Moss and Johnson 1994; Bower et al. 1996). In this study, we also use the images recorded by 2D probes to differentiate drops from ice crystals by assuming that all spheres are drops and all nonspheres are ice crystals in order to estimate LWC in convective clouds. The detailed procedures are as follows.

In this algorithm, only entire-in and center-in particles (Heymsfield and Parrish 1978) are accepted. Any particle with its center outside of the array is rejected, and the corresponding interarrival time is also reduced. Splashed and shattered particles are removed based on the interarrival time method (Field et al. 2006). Some artifacts in the particle images can be removed through this method, while others cannot. The unremoved artifacts have potential impacts on the drop–ice identification, which will be evaluated below.

When dealing with the entire-in particles, we define the y distance as the distance from the pixel on the edge to the x axis, which horizontally crosses the geometric center of the particle (Fig. 2a). The geometric center is defined as the middle of the x and y coordinates for each particle (i.e., $x_{\text{mid}}$ and $y_{\text{mid}}$). Then we create a y-distance
distribution by arranging the $y$ distances of all the edge pixels counted clockwise, starting from the leftmost one, as illustrated in Figs. 2d and 2g. If the particle is purely spherical (Fig. 2a), we will get a sinusoidal distribution of $y$ distances as a function of $2\pi x' / x'_{\text{max}}$, as presented in Fig. 2b. Here, $x'$ is the index of each counted pixel, and $x'_{\text{max}}$ is the maximum of $x'$. Taking the Fourier transform of this $y$-distance distribution, we get the corresponding frequency spectrum:

$$ F(\omega) = 2\pi \int_{0}^{1} y(x^\prime) e^{-i\omega x^\prime} d(x^\prime), $$

where $y(x^\prime)$ refers to the function of $y$-distance distribution and $x^\prime = x'/x'_{\text{max}}$; $F(\omega)$ refers to the frequency spectrum, and $\omega$ is the angular frequency.

Physically, the maximum value of $F(\omega)$ [i.e., $F(2\pi)$] reveals the radius of a particle, and $F(\omega)$ at higher frequencies indicates the nonsphericity and roughness of

<table>
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<th>TABLE 1. Summary of the number of penetrations, the duration (s), and the flight length sampled in the three defined categories.</th>
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<tr>
<td><strong>Temperature (°C)</strong></td>
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<td><strong>Updrafts near cloud top</strong></td>
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<td><strong>Dissipating clouds</strong></td>
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the 2D-C image boundary. So the ratio between \( F(2\pi) \) and the summation of \( F(\omega) \): \( F(2\pi)/\sum F(\omega) \), can indicate the shape of a particle. Here we use this ratio \( f_1 \) to identify spheres and nonspheres. For the purely spherical particles, \( F(\omega) \) has only one value: \( F(2\pi) \) (Fig. 2c), and thus \( f_1 \) equals 1. However, there are no purely spherical particles in 2D images, and the \( f_1 \) is always smaller than 1. For drops, the 2D images are roughly spherical, with the corresponding \( f_1 \) close to 1. For irregular ice particles, the value of \( f_1 \) is smaller. Two examples are shown in Figs. 2d–i. The left panels show the shapes of the two particles: one is identified as a spherical particle (Fig. 2d), and the other is a graupel particle (Fig. 2g). The y-distance distributions are shown in Figs. 2e and 2h, and the corresponding frequency spectrums are shown in Figs. 2f and 2i, respectively. From this figure, we can see that, for the spherical particle, the y-distance distribution is close to a sinusoidal function, and the value of \( f_1 \) is close to 1 (0.81), while, for the graupel, the shape is more irregular, the y-distance
distribution has more oscillations, the spectrum shows more high frequencies, and the $f_1$ is relatively smaller (0.68). Therefore, the value of $f_1$ can be used as a parameter to approximate the difference between spherical and nonspherical images, given the limitations of the imaging probes.

To better separate spherical from nonspherical particles, we can similarly define $f_2$ from the $x$-distance, which is the distance from the pixels on the edge to the $y$-axis crossing the geometric center.

Other than $f_1$ and $f_2$ determined by the Fourier transformation technique, two area ratios are also defined in this study to investigate the roundness and symmetry of a particle. The first area ratio is defined as the area of the particle divided by the area of the smallest circle that can cover the whole particle. This area ratio has been introduced in several studies (e.g., Baker et al. 2009). For drops, the area ratio is close to 1, while for ice particles such as needles the area ratio is smaller. This area ratio is referred to as $f_3$. The second area ratio is defined by partitioning a particle image into quadrants based on the $x$ and $y$ axes (upper left, upper right, lower left, and lower right; e.g., Fig. 2a), calculating the area of each quadrant separately, and then dividing the smallest area by the largest. If the particle is purely symmetric, this value should equal 1. For less symmetric particles, this value will be less than 1. Here, this area ratio is referred to as $f_4$.

Combining $f_1$, $f_2$, $f_3$, and $f_4$, we now define a parameter $f$:

$$f = f_1 f_2 f_3 f_4$$

(2)

to distinguish spheres and nonspheres.

When dealing with the center-in particles, we first assume that the undetected part is spherical and fit it; then we use the same technique for the entire-in particles to distinguish the sphere and nonspheres.

To find reasonable thresholds of $f$, the occurrence distributions of $f$ are plotted as a function of particle diameter based on the observations from penetrations at temperatures at or above 3°C, where no ice existed (Fig. 3a), and all the dissipating clouds at temperatures colder than $-10^\circ$C, where ice dominated (Fig. 3b). Figure 3 clearly shows that liquid clouds and dissipating clouds are readily separated in the $f$-size space, but there are some overlaps because there are always some artifacts and instrumentation digitizing errors. In order not to significantly underestimate the LWC, the 10th percentile of the accumulated frequency in each diameter bin in Fig. 3a was calculated. Then the 10th-percentile $f$ values is fitted as a function of size, which is used as the threshold to separate drops and ice particles (i.e., 10% particles with $f$ smaller than the threshold are assumed to be ice, 90% with $f$ larger than the threshold are assumed to be drops) and is applied to all the clouds.
sampled in ICE-T. Therefore, this algorithm underestimates at least ~10% LWC in the pure-liquid region; in the mixed-phase region there is no way to evaluate the performance of this algorithm. In addition, for a given data point or short-subperiod measurement, the random error induced by drop and ice identification could be much larger than 10%, especially when there are significant artifacts in the particles’ images. In Fig. 3b, ~60% of the particles of 200-μm diameter are identified as ice; more than 90% of particles larger than 400 μm are identified as ice. For particles smaller than 200 μm (eight pixels for 2D-C) in diameter, this algorithm is not good enough as a result of the insufficient pixels; most of the small particles are identified as drops even in dissipating clouds.

Figure 3 is a composite result of entire-in and center-in particles. A large portion of the uncertainty is due to center-in particles. If center-in particles are excluded, the drop and ice identification will be better; however, statistical analysis of mass partition shows the results are similar regardless of whether the center-in particles are included. In addition, the sample volume of 2D probes will be reduced if center-in particles are excluded, which is not appropriate for statistical analysis. So in this study both entire-in and center-in particles are included.

Now the diameter of each drop can be calculated, and the drop size distributions can be plotted. The diameters for both drops and ice crystals are calculated using

\[ D = \sqrt{4A/\pi}, \]  

where \( D \) is the area equivalent diameter, and \( A \) is the area of the shadow image. When calculating the diameter, the error caused by diffraction is corrected based on Korolev et al. (1998). Additionally, the drops with maximum diameter larger than ~1 mm tend to be oblate; thus, the diameter is calibrated based on the empirical study from Jones (1959). In our study, the size distributions from 25 to 1200 μm are determined from the 2D-C images, while those from 1200 to 19200 μm are from 2D-P images. This 1200-μm threshold is chosen because for 2D-P most of the spheres greater than 1200 μm (eight pixels for 2D-P) in diameter can be well identified (Fig. 3).

With the drop size distribution determined, the LWC contributed by drops larger than 25 μm can be calculated. For droplets smaller than 25 μm, we use the LWC measured by the King probe minus the overlap between the King and 2D probes. The total LWC can be formulated as

\[ \text{LWC} = \text{LWC}_{\text{King}} + \text{LWC}_{\text{2D}} - \sum \left[ \rho_l \frac{\pi}{6} S(D^{2D})^3 E_{\text{King}}(D) \right], \]  

where \( \rho_l \) is the density of liquid water, \( S(D^{2D}) \) is the drop size distribution measured by the 2D probes, and \( E_{\text{King}}(D) \) is the collection efficiency of the King probe. Generally \( E_{\text{King}} \) is equal to 1 for droplets smaller than 30 μm, then starts to decrease as droplets become larger. After the LWC is determined, the IWC can be calculated using

\[ \text{IWC} = \text{CWC}_{\text{CVI}} - \text{LWC}, \]  

where \( \text{CWC}_{\text{CVI}} \) is the CWC measured by CVI. If the calculated LWC is larger than \( \text{CWC}_{\text{CVI}} \), the IWC is set to 0. The uncertainty of the calculated LWC and IWC could be caused by multiple factors: for example the uncertainty induced by drop and ice identification, systematic and random errors of instruments, and the large variability of the microphysics in convective clouds since the instruments were not exactly collocated. Detailed evaluation will be discussed in section 3c.

c. Evaluation

Examples of 2D-C images from the three categories of convective clouds (Fig. 1) are shown in Fig. 4; the images are separated into ice particles and drops based on the classification results. Most of the spheres and non-spheres can be readily identified. The ice particles in the updraft near cloud top are frozen drops, which suggests that the primary large ice particles come from the freezing of drops. This finding is consistent with previous studies (e.g., Zeng et al. 2001; Rangno and Hobbs 2005; Lawson et al. 2015). The concentration and shape of small ice particles cannot be identified using the 2D-C probe. In the updraft far below cloud top, we can see a high concentration of ice crystals; in addition to the frozen drops, there are also a lot of needles and columns, most of which may be secondarily generated (Heymsfield and Willis 2014). The high ice concentrations observed here and by Heymsfield and Willis (2014) are in relatively weak or moderate updrafts, not the vigorous, new updraft cores studied by Lawson et al. (2015). The observation of an occasional strong updraft in these regions is typical of the bubble nature of these clouds. The dissipating clouds are dominated by ice, but the ice concentration here is relatively low when compared to the updraft far below cloud top; we can see various shapes of ice crystals. Spherical frozen drops cannot be identified using this algorithm, so they are regarded as liquid drops. The particle images in this figure were well recorded, but in some other measurements there could be significant artifacts in particle images, which result in larger uncertainty in the drop and ice identification.
LWC evaluation is presented in Fig. 5. Figure 5a shows the scatterplot of the LWC for all particles and spheres measured by 2D-C at temperatures at or above 3°C, which reveals the uncertainty of LWC calculated using the algorithm used in this study. The figure suggests for most of the measurements the uncertainty is within 10%, but sometimes the uncertainty could be larger than 50%, mainly because of the artifacts in 2D images. Figure 5b compares the LWC with CWC measured by the CVI for all penetrations at temperatures at or above 3°C. The red dots represent the LWC measured by the King probe, and the black dots represent the LWC calculated using Eq. (4). From the figure, we can see the 1-s measurements (small dots) show an order-of-magnitude spread; this is mainly caused by random errors of instruments and the spatial inhomogeneity of clouds (e.g., because the King probe and CVI are installed on different areas on the aircraft or as a result of different contributions of large particles). Therefore, a large random error is expected in the calculated IWC. The binned averaged values for given CWC intervals are presented by big dots. It is clear that the King probe (red big dots) significantly underestimates LWC for $LWC > 0.4 \text{ g m}^{-3}$ in these convective clouds. By including large drops, our approach (black big dots) brought the LWC close to the 1:1 line. But for CVI CWC greater than $1 \text{ g m}^{-3}$, our LWC is underestimated because some deformed drops are identified as ice crystals. In ice-dominated clouds, our 2D-probe-based approach has difficulty to identify ice particles smaller than 200 μm, and the King probe might be responding to ice particles (Strapp et al. 1999;
Korolev et al. 2007), resulting in overestimation of LWC.

Figure 6 shows the concentration fraction of drops (i.e., drop concentration divided by total concentration) measured in the updrafts near cloud top. The results are calculated for different size bins and temperatures. As seen from the figure, most of the particles are identified as drops for temperatures warmer than \(258\) C, and the drop concentration fractions are larger than 90% in most of the grid boxes. For temperatures lower than \(258\) C, the drops rapidly freeze; most of the particles larger than 400 \(\mu\)m are regarded as ice below \(-11^\circ\) C. Some images from the 2D-P probe are shown in order to confirm the existence of spheres and quasi spheres larger than 1.5 mm; these particles are assumed to be drops in this study.

Although some particles are identified as ice at temperatures warmer than \(-3^\circ\) C, they are more like deformed drops rather than ice based on 2D-C images. Nonspheres are observed between \(-4^\circ\) and \(-5^\circ\) C in a few cases. Some of them could be artifacts in particle images, and some of them could be ice. Several previous studies suggest ice may be generated at about \(258\) C (e.g., Hobbs and Rangno 1990; Rangno and Hobbs 1991). It is, however, difficult to determine whether they are freshly generated or residuals from adjacent clouds or even just artifacts in particle images. Recently, Lawson et al. (2015) analyzed the data collected in freshly generated convective clouds with strong updrafts over the tropical ocean using CPI images, and they found the “first ice” is not observed at temperatures warmer than \(-8^\circ\) C. Therefore, the nonspheres at temperatures warmer than \(-8^\circ\) C observed in this study may be artifacts in particle images or residuals from adjacent clouds. Using 2D-probes alone is not enough to identify the small amount of fresh ice in updrafts near cloud top at warm temperatures. More data, especially with more advanced instruments, are needed to better explore ice initiation.

4. Results
a. Temperature dependence

Temperature is the primary factor that determines liquid–ice mass partitioning in clouds (Bower et al. 1996; Boudala et al. 2004). Figure 7 shows the temperature dependence of the LWC, the IWC, and the liquid fraction for the tropical maritime convective clouds sampled during ICE-T. The top, middle, and bottom panels represent the updrafts near cloud top, the updrafts far below cloud top, and the dissipating clouds, respectively. The large spread of LWC, IWC, and liquid fraction in each panel reveals the substantial random errors and the large variability of microphysics in convective clouds. In the updrafts near cloud top, the mean LWC (Fig. 7a) tends to increase between 0° and \(-10^\circ\) C, and then decreases from between \(-10^\circ\) and \(-15^\circ\) C. Our calculations suggest that the average values of LWC range between 0.4 and \(~1.1\) g m\(^{-3}\). These values are significantly lower than average values measured by Lawson et al. (2015) in ICE-T clouds and the adiabatic LWC. There are several
possible explanations. 1) Limitations of the instrumentation prevent it from measuring all of the liquid water in these clouds. 2) The sampling strategy of the C-130 was not favorable for measuring LWC in fresh strong updraft cores, so the measurements are made mostly in relatively older and/or weaker updrafts where the LWC has been depleted as a result of entrainment and/or precipitation. 3) Some combination of explanations 1 and 2 above. Using similar instruments, Stith et al. (2002) report high LWC with droplet concentrations within 300–400 cm$^{-3}$, but statistical analyses (Stith et al. 2004) show similar magnitude and temperature trends of LWC to our results. Rangno and Hobbs (2005) showed that LWC peaks at about 2.5 km above the cloud base with mean values less than 1.5 g m$^{-3}$, then decreases with height, and is significantly lower than the adiabatic values. The averaged IWC (Fig. 7b) rapidly increases from 0 to $\sim 1.2$ g m$^{-3}$. This rapid ice generation controls the liquid–ice mass partitioning, resulting in the liquid fraction (Fig. 7c) decreasing from 1 to 0.3 (because of the drop and ice identification uncertainty, the liquid fraction at temperatures warmer than $-4^\circ$C is about 0.9). From $-4^\circ$ to $-5^\circ$C, nonspheres are observed in a few cases. Some of them may be artifacts in particle images and some of them may be ice from adjacent clouds. The fresh ice cannot be identified using 2D probes alone. Lawson et al. (2015) showed no fresh ice was present at temperatures $> -8^\circ$C in the uncontaminated updraft cores that they determined. Some studies of stratiform clouds (e.g., Wang et al. 2014) suggest ice could be generated at warm temperatures because of the biogenic aerosols. Since there are limitations to the instruments and potential sampling biases, more measurements, especially with more advanced instruments, are needed to better study ice initiation in convective clouds in the future.

For the updrafts far below cloud top (Figs. 7d–f), secondary ice production and the fallout of ice particles from lower to higher temperatures can result in high ice concentrations at warm temperatures ($>-10^\circ$C), and the liquid water can be depleted quickly. From Figs. 7d and 7e, we can see the averaged LWC in the updrafts far below cloud top is between 0.4 and 1.1 g m$^{-3}$. This is of a similar magnitude to that in the updrafts near cloud top, maybe because the relatively strong updrafts can still maintain the LWC even if there were fallout of ice particles. But again, notice there are limitations of the instruments and the results may have large uncertainties. The averaged IWC increases from 0.5 to 1.8 g m$^{-3}$ as the temperature decreases from 0$^\circ$ to $-15^\circ$C. Compared to the updrafts near cloud top, the IWC in the updrafts far below cloud top is larger at warmer temperatures ($>-10^\circ$C), suggesting secondary ice production and fallout of ice crystals. Taking into account the potential uncertainties, the resultant liquid fraction (Fig. 7f) decreases from 0.75 to 0.3. Compared to the updrafts near

**FIG. 6.** The size and temperature dependence of the concentration fraction of drops based on all the updrafts near cloud top sampled during ICE-T. The 2D-P image examples are displayed to confirm the existence of the spheres larger than 1.5 mm.
cloud top, the liquid fraction in the updrafts bar below cloud top is obviously smaller at temperatures higher than $-10^\circ C$. For the dissipating clouds (Figs. 7g–i), the averaged LWC and IWC are both very small. The LWC is smaller than 0.1 g m$^{-3}$, and the IWC is smaller than 0.3 g m$^{-3}$ between 0$^\circ$ and $-15^\circ C$. The resultant average liquid fractions are small (between 0.04 and 0.5) in all the temperature ranges, indicating that dissipating clouds are dominated by ice. Since the LWC is overestimated in dissipating clouds, the liquid fraction may be even smaller.

b. Vertical velocity dependence

Vertical velocity is also an important parameter in convective clouds. Different dynamic environments

Fig. 7. Temperature dependence of the LWC, the IWC, and the liquid fraction for (a)–(c) the updrafts near cloud top, (d)–(f) updrafts far below cloud top, and (g)–(i) the dissipating clouds based on C-130 measurements from the seven research flights in ICE-T. The colored backgrounds are occurrence distributions at each temperature bin, and the triangles are the mean values.
may result in different ice generation mechanisms. Figure 8 shows the composite vertical velocity and liquid fraction as functions of the normalized diameter of all the updrafts near cloud top. On the x axis, 0 indicates the upwind edge, and 1 indicates the downwind edge of the updrafts. Generally the vertical velocity is large near the center of updraft cores and small near the edges (1–2 m s$^{-1}$). In many updrafts sampled by the C-130 the maximum vertical velocity exceeds 5 m s$^{-1}$. These updrafts are not as strong as the updraft cores sampled by the Learjet but still can maintain relatively high LWC. The liquid fraction decreases obviously with decreasing temperature; however, it has no obvious correlation with the vertical velocity at the same temperature zone. The liquid fraction near the edge of updraft is in a similar magnitude as that near the center of updraft. The similar liquid fraction suggests that ice generation is strongly linked to water content rather than vertical velocity–related process or properties.

Figure 9 illustrates the drop and ice size distributions on the basis on all the 1-s measurements in the updrafts near cloud top. The diameter is calculated from Eq. (3). The results are given according to different vertical velocity and temperature ranges. At relatively warm temperatures (Fig. 9a), we find a quantity of drizzle-size drops. In stronger updrafts, the drop concentration is higher, especially for millimeter drops. No ice is observed between 0°C and −3°C (Fig. 9f). The drop concentration shows strong dependence on the vertical velocity. In stronger updraft, more drops are found, while relatively fewer drops are found in weak updrafts. The ice size distribution shows continuous ice generation with decreasing temperature. The ice concentration (>125 µm in diameter) increases from less than 1 to 100 L$^{-1}$ as the temperature decreases from −4°C to −15°C. Between −6°C and −15°C (Figs. 10h–j), the size distribution of ice smaller than 1 mm is similar in different vertical velocity ranges, but the concentration of millimeter ice is correlated to the vertical velocity. In stronger updrafts, more millimeter ice particles are found, which suggests greater ice generation in stronger updraft. The consistency between size distributions of drops and ice crystals suggests that primary ice generation in convective clouds is strongly related to the freezing of millimeter drops, consistent with previous findings (Lawson et al. 2015).

Figure 10 presents the LWC, the IWC, and the liquid fraction profiles of the updrafts near cloud top.
The results are calculated based on all the 1-s measurements. From Fig. 10a, we can see relatively small LWC in relatively weak updrafts and relatively large LWC in strong updrafts. The LWC profiles vary similarly for different vertical velocity ranges; all tend to increase between 0°C and -10°C, then decreases between -10°C and -15°C. Taking into account the measurement uncertainties, the magnitude of the LWC in relatively weak updrafts (1–4 m s⁻¹) is about 0.2 g m⁻³ smaller than that in the strong updrafts (≥7 m s⁻¹). The IWC (Fig. 10b) increases as the temperature decreases from 0°C to -15°C. There is no correlation between IWC and vertical velocity at temperatures warmer than -26°C. At temperatures colder than -26°C, higher IWC is observed in stronger updrafts. The IWC is smaller in the weak updraft compared to that in the strong updrafts. Although both LWC and IWC have correlations with vertical velocity, the resultant liquid fraction (Fig. 10c) shows no obvious vertical velocity dependence. All three profiles show that the liquid fraction decreases by a factor of 3 as the temperature decreases from 0°C to -15°C. In the updrafts far below cloud top (Figs. 10d–f), the microphysical processes are more complicated (e.g., riming and secondary ice production). The LWC (Fig. 10d) is roughly correlated to the vertical velocity, similar to what was seen for updrafts near cloud top. The IWC profiles (Fig. 10e), when compared to those in the updrafts near cloud top, show a smaller spread among different updrafts, and there is no obvious vertical velocity dependence. The resultant liquid fraction (Fig. 10f), unlike that in the updrafts near cloud top, shows a slightly larger spread among different updrafts. This reveals the complexity of the microphysics in the updrafts far below cloud top. Between -4°C and -10°C, there is some vertical velocity dependence: a relatively small liquid fraction is found in the relatively weak updrafts, while the liquid fraction becomes larger in stronger updrafts. This finding may be associated with
liquid–ice interaction and secondary ice production; as shown by Heymsfield and Willis (2014), low liquid water content and weak vertical velocities are the cloud conditions that favor secondary ice production. But since there are limitations to the instruments and the results may have large uncertainties, this finding must be better examined in the future.

c. Discussion

Following the analyses above, we can see the liquid–ice mass partition is controlled by the fast ice generation in convective clouds, which is very different from the ice initiation in stratiform clouds. In stratiform clouds, there is no strong updraft to provide large LWC and high concentration of drizzle-size and millimeter drops. Airborne measurements show that the average LWC in stratiform clouds is about 0.1 g m$^{-3}$ at $-3^\circ$C, and it keeps decreasing as temperature decreases (Boudala et al. 2004). However, in convective clouds, the presence of drizzle-size and millimeter drops could be an important source for fast ice generation (Lawson et al. 2015). In addition, the ice generation in stratiform clouds is mainly through deposition, condensation nucleation, and immersion freezing, all of which depend on IN concentration, so the ice concentration is low (Murray et al. 2012). Some studies suggest there is also secondary ice production in stratiform clouds (e.g., Rangno and Hobbs 2001), but the observed ice concentration is still less than 1 L$^{-1}$ at temperatures warmer than $-15^\circ$C, which is higher than the IN concentration but much lower than the ice concentration in convective clouds. The relatively slow ice generation leads to a low IWC in stratiform clouds. Observations show the IWC in stratiform clouds is less than 0.1 g m$^{-3}$ and has no obvious increase with decreasing temperature (Boudala et al. 2004). The resultant liquid fraction is much larger than that in convective clouds. In situ measurements show the liquid fraction in stratiform clouds at $-15^\circ$C is about 0.6–0.7 (Boudala et al. 2004). Ground-based radar measurements show the liquid fraction in Arctic stratiform clouds to be greater than 0.8 for the clouds with top temperatures between 0$^\circ$ and $-15^\circ$C (Zhao 2011). Although there are measurement uncertainties in the previous and present studies, the differences of the liquid–ice partitions found in stratiform and convective clouds are obvious, and therefore it is necessary to treat them differently in models.
The statistical results of this study can be used in a statistically relative sense for comparing with the numerical simulations of the microphysics in tropical maritime convective clouds. Figure 11 compares the temperature-dependent liquid fraction from two early parameterizations and the results from the present study. The first parameterization (solid line) was developed by Smith (1990); it has been used by the Met Office Atmospheric Global Climate Model, and it has been widely referred to in other, related studies. The liquid fraction in the updrafts near cloud top in the present study is roughly consistent with this parameterization at warm temperature, but the liquid fraction of this parameterization decreases more rapidly as temperature decreases. At temperatures colder than $-10^\circ$C, Smith’s parameterization is closer to the results of dissipating clouds. Recall that the algorithm in the present study may overestimate the LWC for ice-dominating clouds, and thus the actual liquid fraction in ICE-T dissipating clouds may be lower than the reported results, which may give a result closer to Smith’s parameterization. Another parameterization (dashed line) was developed by Bower et al. (1996); this is the only parameterization of liquid fraction known to the present authors that was developed for convective clouds. The parameterization was developed based on observations for both midlatitude maritime and continental clouds. The observational data used to develop Bower’s parameterization show a quite large spread (Fig. 9 in Bower et al. 1996) as a result of compositing all the convective clouds in different stages and at different geographic locations, with some clouds almost glaciated while some are still dominated by liquid drops at $-10^\circ$C. Figure 11 shows that the liquid fraction measured in ICE-T is quite different from that of Bower’s parameterization; the present study shows a much faster decreasing liquid fraction. The main reason for the discrepancy could be that this parameterization was developed mainly from measurements sampled in midlatitude continental convective clouds, whereas only a small part of their dataset is from maritime cumulus. Actually, the original data (Bower et al. 1996, their Fig. 9) also show much faster decreasing liquid fraction for the marine cumulus, consistent with the results in the present study. If this is the case, the difference between ICE-T results and Bower’s results indicate potentially systematic differences of liquid–ice mass partition between continental and maritime convective clouds, which need to be further explored.

The liquid–ice mass partitioning is quite different between the three categories. The most obvious differences of the liquid fraction between the three categories are found in the warm temperatures ($T > -10^\circ$C). For temperatures colder than $-10^\circ$C, the three liquid fraction profiles tend to merge together. A schematic diagram is presented in Fig. 12 that gives a visual idea of the liquid–ice mass partitioning in the tropical maritime convective clouds. In Figs. 11 and 12, it is apparent that the liquid–ice mass partition changes significantly as convective clouds evolve, and therefore they cannot be parameterized in the same way. Ice generation in convective clouds controls the liquid-to-ice latent heat release and, thus, affects convective cloud dynamics and detrainment processes. Therefore, high-resolution model simulations need to capture the observed liquid fraction evolution, which requires improved microphysical parameterization. Our results and other ICE-T data offer an important data source to achieve the goal.

5. Conclusions

The liquid–ice mass partitioning in tropical maritime convective clouds is studied using data collected from seven research flights during the ICE-T field campaign. The LWC is calculated by combining the LWC measured by a combination of hot-wire and imaging probes. The CWC is measured by a CVI. The IWC is calculated from the CWC minus the LWC. On the basis of cloud vertical structure provided by the WCR reflectivity and in situ vertical velocity measurements, the sampled clouds are classified into three categories: updrafts near...
In the updrafts near cloud top, fast ice generation controls the liquid–ice mass partition. Taking into account the potential uncertainties, the liquid fraction rapidly decreases by a factor of 3 as the temperature decreases from 0°C to −15°C. In the updrafts far below cloud top, a relatively larger IWC and a smaller liquid fraction are found at warm temperatures (>−10°C), probably associated with liquid–ice interactions and secondary ice production. In the dissipating clouds, a small liquid fraction is found in all the temperature ranges. The largest difference of the liquid–ice mass partition between the three cloud life stages is found at temperatures warmer than −10°C. At temperatures colder than −10°C, the liquid fractions tend to merge together.

3) Compared to the temperature and life stages, the vertical velocity has a minor impact on the liquid–ice mass partition. For the updrafts near cloud top, a larger LWC and IWC are found in the stronger updrafts. However, the liquid fraction shows no statistical vertical velocity dependence. The drop and ice size distributions show that more millimeter drops and ice crystals are found in stronger updrafts, suggesting a greater ice generation in stronger updrafts and that the ice initiation is related to the existence of millimeter drops. For the updrafts far below cloud top, the liquid fractions have a relatively larger spread, suggesting more complicated microphysical processes. The liquid fraction is correlated to the vertical velocity between −4°C and −10°C; this may be linked to liquid–ice interactions and secondary ice productions. Since there are limitations to the instruments and the results may have large uncertainties, the vertical velocity dependence of liquid–ice partitioning in convective clouds needs to be better examined in the future.

Comparisons between the three different categories as well as two temperature-dependent parameterizations suggest that the liquid–ice mass partition changes significantly as the clouds evolve. In addition, the microphysical properties differ between clouds at different geographic locations. For example, Bower et al. (1996) show faster ice generation in maritime convective clouds than that in continental convective clouds. Thus, the liquid–ice mass partition cannot be well described by a single parameterization, and it becomes necessary to treat the liquid–ice
partition differently for different clouds in models through more realistically detailed parameterization of microphysical processes.

Since the dataset used in this study is small and there are uncertainties due to the limitations of the instruments, more measurements, especially with more advanced instruments, are needed to better understand the liquid–ice mass partition in convective clouds and to improve the microphysics schemes in numerical models.

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