1. Introduction

The melting layer (ML), an important feature of stratiform precipitation, is a quasi-isothermal layer resulting from cooling due to the melting of snowflakes and ice particles. Observation of the ML gives useful information about the vertical structure of precipitation falling both from above and into the region where temperatures \( T \) are above \( 0^\circ \text{C} \). In general, one would expect that there is conservation of condensate flux (precipitation rate) into and out of the ML.

The melting layer is shown to have very similar vertical dimensions, independent of the topography and climatic conditions. Wolfensberger et al. (2016), using long-term radar datasets from Davos and Payerne, Switzerland, Ardeche, France, and Iowa, United States, found that the median depth of the ML was 350, 300, 350, and 325 m, respectively. Variations in the depth of the ML during and between precipitation events was shown to be strongly related to the presence of rimed particles, to the fall velocity of hydrometeors, and to the intensity of the radar bright band, a narrow horizontal layer of stronger radar reflectivity in precipitation at the level in the atmosphere where snow melts to form rain. An increased hydrometeor concentration could lead to an increase in the diabatic cooling of the surrounding air during the melting process. This effect can be quite important when the layer is stable and when horizontal temperature advection is small (Kain et al. 2000). The cooling effect increases with the precipitation rate and can significantly lower the freezing level. It has also been shown that a deep ML is usually associated with a strong bright band, a higher vertical extension of precipitation, as well as a larger gradient of radar reflectivity values above the ML (Fabry and Zawadzki 1995; Wolfensberger et al. 2016).

Research into the melting of snow has included surface observations, laboratory experiments, and model studies. Matsuo and Sasyo (1981a) analyzed meteorological surface observations, laboratory experiments, and model studies. Matsuo and Sasyo (1981a) analyzed meteorological surface data from three stations in Japan for the purpose of examining the effect of relative humidity (RH) on melting of snowflakes below the \( 0^\circ \text{C} \) temperature level. They found that the phase of the precipitation (rain or snow) observed at the ground was closely dependent on both the surface RH and \( T \). They developed parameterizations between \( T \) and RH that represented the critical conditions that separated some ice and all rain. Below the critical RH, precipitation was all snow; above it, the phase could be snow, sleet, or rain, according to the situation. The warmest temperature where snow was observed was about \( +6^\circ \text{C} \).

Laboratory experiments have been used to characterize the rate of snowflake melting as a function of the air temperature and relative humidity. Matsuo and Sasyo (1981b) simulated the melting of natural snowflakes while held on a net of thin nylon threads exposed to an airstream of velocity 1 m s\(^{-1}\) and constant temperature of 5.5°C. Such experiments do not properly model the melting of aggregates in clouds because they have increasing fall speed as they melt and fall into increasingly warmer temperatures. Given these limitations, they developed relationships for the melting of snowflakes as a function of the temperature difference between the air and their surface \( (0^\circ \text{C}) \). Mitra et al. (1990) studied the melting of snow in the Mainz Vertical Wind Tunnel. Snowflakes were captured and brought into the tunnel, where they were exposed to steadily
increasing temperatures above 0°C and an RH of 90%. They reported the stages of melting of snowflakes and found that melting commenced when the surface temperature of the particle reached 0°C. Oraltay and Hallett (1989) studied the melting of single crystals and snowflakes suspended on a vertical filament in the laboratory, where they were exposed to varying temperatures and relative humidities. The focus of the study was to examine how the ice particles melted as a function of their crystal type.

From their laboratory experiments and surface-based observations, Matsuo and Sasyo (1981c) developed analytic equations to simulate their observations. Complexities with modeling the process included questions about whether water evaporates from the surface if liquid water is present and/or sublimation of ice from the particle surface. Since their laboratory experiments showed that most of the water penetrates into the interior of the particle, they modeled the process as sublimation. They developed a useful schematic showing the effects of air temperature and water vapor density of the air on melting of snowflakes at temperatures warmer than 0°C.

Sims and Liu (2015), using globally based observations over a long period of time, found that in addition to wet-bulb temperature, vertical temperature lapse rate affects the precipitation phase. Also using surface-based observations from many locations around the world and for a 25-yr period, Behrangi et al. (2018) concluded that wet-bulb temperature yields the highest skill score for determining precipitation phase, rather than the air temperature.

Oraltay and Hallett (1989) pointed out that a complete evaluation of the melting layer structure, leading to diagnostics for the radar bright band and the depth of the overlying isothermal layer, requires measurements of particle size distribution (PSD) together with thermodynamic quantities on the same scale as the size distribution change with location. Such measurements have been made in the past. Willis and Heymsfield (1989) examined the aircraft observations and theoretical evolution of particles above, through, and below the ML in the stratiform region associated with a mesoscale convective system. The aircraft data were obtained from spiral descents whereby the aircraft advected with the wind and the descent rate approximately corresponded to the typical hydrometeor fall speeds (Lo and Passarelli 1982). Willis and Heymsfield (1989) concluded that most of the mass melts, and thus most of the cooling that leads to the 0°C isothermal occurs, in a thin layer above the location of the radar bright band. The production of a few very large aggregates is dramatic after the onset of melting, due in part to a melting-induced increase in the terminal velocity difference between similarly sized hydrometeors and increased sticking efficiency. Below the isothermal layer, the melting process is nearly complete by +3°C, although some large aggregates reach +5°C and warmer as ice. The radar reflectivity maximum (bright band) is due to these relatively few very large aggregates that survive to warmer temperatures. Recent polarimetric radar studies have revealed additional characteristics of the radar bright band (Kumjian et al. 2016; Carlin and Ryzhkov 2019), including inferences about particle habits.

Heymsfield et al. (2002, 2015) used Lagrangian-type spiral descents (Lo and Passarelli 1982), wherein the aircraft drifts with the wind, descending at approximately the mean fall speed of the particle population. Using data from four field programs, they were able to characterize the evolution of the PSD and habits in deep subtropical and tropical stratiform cloud layers. The PSDs broadened from cloud top toward cloud base, with the largest particles increasing in size from several mm at cloud top to 1 cm or larger toward cloud base, and the concentrations of particles smaller than 1 mm in size decreased with decreasing height. Aggregation was found to be responsible for these trends. They also found at relatively high RH, the PSD slope uniformly decreased downward, while the maximum particle size of the largest particle continued to increase. At relatively low RH, there was little melting of the snow particles until a temperature of +2°C was reached. The size of the largest particles increased even at the low RH conditions. For both low and high RH conditions, aggregation was responsible for the increase in the size of the largest particles.

The question of the rate of snowflake melting is a central factor in determining the depth of the melting layer and the associated bright band. It is also important for weather forecast and water hydrology applications, and has important implications for snowfall in a changing climate (see Tamang et al. 2020, and references cited therein). A challenging task for forecasters is to identify the conditions when melting effects might influence the type of precipitation observed at the surface. As pointed out by Kain et al. (2000), because a melting-induced changeover to snow is promoted by high precipitation rates, resulting snowfalls can be heavy, and the consequences of a poor forecast can be exceptionally serious. A snow event was used to demonstrate a day with high precipitation rates when there was a changeover from rain to snow. From simple analytic equations, they showed that the melting of the snow leads to cooling of the layer, and that, given a high enough precipitation rate, the melting layer can be eliminated.

Although several researchers have developed analytical equations to calculate the melting rate of snow (e.g., Matsuo and Sasyo 1981c), it is difficult to model realistically the changes in RH during the fall of the ice particles into and through the melting layer, the properties of the single particles and the PSD, and how the melted water on a particle remains on the surface or within the particle.

In this study, we address the question, “How warm can it snow?” How long can snow survive in the melting layer? A combination of instrumented aircraft Lagrangian spiral descents and ascents, together with along-flight descents through the ML, and observations at the surface from NOAA weather stations are used to characterize the survival of snow under a wide range of temperature and relative humidity conditions. Section 2 provides information on the datasets, section 3 describes the observations, and section 4 presents some interesting aspects of snow gleaned from the Lagrangian spiral descents. In section 5, the results are summarized and conclusions are drawn.
2. Data sources

In this section, details of the airborne measurements and surface observations are presented.

a. Airborne observations

Table 1 summarizes the details of the Lagrangian spiral descents through the ML, with the identification numbers of the flights consistent with the field programs and figures presented below. With a Lagrangian type spiral descent, if conditions are quasi steady and the properties of the atmosphere are fairly uniform over a length scale somewhat larger than the diameter of the loops, the aircraft largely samples the evolution of the PSD properties and the characteristics of the hydrometeors from the particle probes and their imagery as a function of altitude. The average descent rate was 1–2 m s\(^{-1}\) in the ice regions and about 4 m s\(^{-1}\) at temperatures above 0°C. In addition, slow Eulerian descents through the melting layer were conducted.

The data collected here (available on the NASA and DOE/ARM websites) are from five NASA field programs: the Tropical Rainfall Measuring Mission (TRMM) Kwajalein Field Experiment (KW\(A\)JEX), based out of Kwajalein, Marshall Islands, in 1999; the NASA Cirrus Regional Study of Tropical Anvils and Cirrus Layers–Florida-Area Cirrus Experiment (CRY\(S\)T\(A\)L\(F\)ACE) in 2002 over southern Florida and the surrounding oceanic and gulf regions; the Precipitation Measurement Mission (PMM) Midlatitude Continental Convective Cloud Experiment (MC3E) in Oklahoma in 2011 near Lamont, Oklahoma; the Integrated Precipitation and Hydrology Experiment (IP\(H\)E\(X\)), which was conducted from North Carolina in 2014; and the Investigation of Microphysics and Precipitation for Atlantic Coast-Threatening Snowstorms (IMP\(A\)CTS), the first deployment of which was conducted based out of Wallops Island, Virginia, in 2020. An in-depth discussion of the data from the aircraft Lagrangian spiral descents is presented in Heymsfield et al. (2002, 2015). As noted earlier, the aircraft spirals within an ice cloud, drifting with the wind and descending at a rate close to the average ice crystal fall speed. It differs from an Eulerian spiral descent, in which the aircraft spirals over a given geographic location.

We have reanalyzed KWAJEX data collected during three Lagrangian spiral descents from above to below the ML in stratiform ice cloud. Each loop of the spiral had nearly the same diameter during the descent for each case, ranging from about 5–10 km in diameter. The descent on 19 August initiated at a temperature of \(-17^\circ\)C and concluded at +6°C. For the 30 August case, the sampling temperatures were in the range \(-17^\circ < T < +6^\circ\)C.

Three Lagrangian spiral descents by the UND Citation aircraft on 26 July 2002 during CRY\(S\)T\(A\)L\(F\)ACE are examined in detail here. These spirals were in the convective outflow region of a strong thunderstorm, and they are particularly interesting because the RH was highly subsaturated. Correspondingly, downdrafts from \(-1\) to \(-3\) m s\(^{-1}\) were observed at \(T > 0^\circ\)C for the spirals, likely the result of the low RH in this layer. The spirals commenced at temperatures below \(-10^\circ\)C and ended above +6°C. Direct measurements of the condensed water content (CWC) were made, as discussed below.

For MC3E, a descent was made through the ML, with temperatures in the range \(-3^\circ < T < +3^\circ\)C. Updrafts of \(\geq 3\) m s\(^{-1}\) and liquid water were encountered at temperatures just above +3°C.

Five cases on four days during IP\(H\)E\(X\), are examined here. The range of temperatures examined is shown in Table 1. Areas of subsaturation were associated with descending motion, of order \(-1\) m s\(^{-1}\). Several regions of relatively dry air, RH \(\sim 80\%\), were sampled at \(T > 0^\circ\)C.

One case during IMP\(A\)CTS was chosen for study. This was a slow descent through the ML to a temperature of \(+8^\circ\)C, followed by a climb up through the ML. The RH was close to water saturation and peak updrafts and downdrafts of \(\geq 4\) and \(\sim 4\) m s\(^{-1}\), respectively, were measured. Unique results regarding melting are obtained through use of the direct measurements of the condensed water content.

Measurements from all of the field programs except IMP\(A\)CTS were made using a University of North Dakota Citation aircraft. For IMP\(A\)CTS, the NASA P3 aircraft was.
3. Results

The presentation of the results and the corresponding discussion is based on fundamental thermodynamic considerations which are introduced briefly in the following two paragraphs. If the surface temperature of an ice particle is 0°C or below, the particle will not melt, even if the air temperature is 0°C or above. In such situations, the ice particle will sublime rather than melt. The dividing point between sublimation and melting is the RH at an ice-bulb temperature of 0°C (RH_{IB}) for a given air temperature.

The value of RH_{IB} is found using the basic equations that yield the ice-bulb temperature for a psychrometer (American Meteorological Society 2020). The vapor pressure over the ice particle surface (e) is derived as follows.

For temperatures in degrees Celsius and vapor pressure in hектopascals, this formula is

\[
e = e_{s}(T_{IB}) - 5.82 \times 10^{-3}(1 + 0.0015T_{IB})p(T - T_{IB}),
\]

where \( T \) is the ambient temperature, \( T_{IB} \) is the ice-bulb temperature (0°C), and \( p \) is the pressure (hPa). At 0°C, the vapor pressure with respect to ice and water are equivalent and have the value of about 6.11 hPa. The relative humidity is given by the ratio of \( e \) to \( e_{s}(T) \). Figure 1 shows plots of \( T \) versus RH for two different atmospheric pressures: 1000 and 500 hPa. The RH(T) to the left of the line in the figure are regions of sublimation, and RH(T) to the right are regions of melting. Rasmussen and Pruppacher (1982) show a plot similar to Fig. 1 (their Fig. 4) for a pressure level of 1000 hPa.

The physical description of the sublimation/melting process is as follows. If the ambient ice-bulb temperature is below 0°C, the particle will sublime, whereas if the ambient ice-bulb temperature is above 0°C, the particle will melt. Ice-bulb temperature can still be well below 0°C even when air temperature increases to above 0°C if relative humidity is less than 100%. Thus, particles often do not melt when falling through the 0°C level. Rather they do not begin to melt until the ice-bulb temperature reaches 0°C or warmer. The drier the atmosphere, the further they fall to reach the level where the ice-bulb temperature is 0°C or above. Therefore, a single parameter, the ice-bulb temperature, will combine the three parameters in Fig. 1 (\( p \), \( T \), and RH). Using the ice-bulb temperature of 0°C separates sublimation from melting.

a. Airborne observations

The focus of the airborne observations is on the bulk properties of melting of the ice particle ensembles as they fall to \( T \approx 0°C \) and how the conditions they encounter during their transit affect their melting. Examining the particle images by eye from the small-particle probe and the HVPS allows us to qualitatively determine the extent of melting during their fall. Although we considered using automated particle detection for the habit classifications, we favored the manual characterization instead. The main reason was that, even in case of an automated particle detection method, the particle features and thresholds that would have to be applied to delineate melted from nonmelted would still be somewhat arbitrary, would vary by instrument, and would
probably not accurately handle complicated situations, such as a broad mix of habits. For example when graupel is present in the particle mix, even at temperatures above 0°C. Nonetheless, an automated approach should be considered for use in future analysis of larger datasets, such as those advanced by Korolev and Sussman (2000) and Praz et al. (2018), with the inclusion of the identification of melting. One method would be to use the particle area ratio, which is a potential identifier of particle melting.

Examples of particle images are shown in Fig. 2, a period during CRYSTAL-FACE when \( T \) increased from about +5.0° to +6.4°C. Particles likely to be ice are labeled a–c in the panels for the 2D-C particle images, and there are numerous examples of ice in the HVPS data. Examining the images for all flights (Table 1), we made a qualitative designation of the degree of melting (\( T > 0°C \)) primarily from the 2D-C imagery and when there was uncertainty from the HVPS. Designations of “all snow,” “mostly snow,” “mostly rain,” and “all rain” characterize the degree of melting. These designations are subjective and have the approximate bounds: “all snow,” no evidence of melting; “mostly snow,” 50%–90% of the particles are snow; “mostly rain,” 50%–90% of the particles are spherical, presumably mostly melted snow or rain, and “all rain,” no evidence of ice particles or nonspherical drops. These designations are used in the following figures that show the microphysics and relative humidity as a function of temperature for the various cases.

Two Lagrangian spiral descents from the CRYSTAL-FACE field program are used to illustrate a low RH case and how snow can survive even at warm temperatures. Figure 3a shows the temperatures encountered as a function of time during the spiral descent, with the coding used to identify the four qualitative descriptions of the degree of melting. The ice-bulb temperature as a function of time is also plotted. Note that \( T_{IB} \) begins to fall below the air temperature—indicating where melting should begin, at about the same time as melting is observed to begin. Figure 3b shows that RH decreases considerably during the descent and is lower than the associated RH_{IB}. Identifiable melting does not commence until \( T \) approaches +3°C when the RH approaches RH_{IB}. Also plotted in the figure is the relative humidity with respect to ice, RH_{ICE}. During the early part of the spiral, there is the potential for ice particle growth. However, sublimation of the ice begins about midway through the spiral, prior to the air temperature reaching 0°C. The median volume diameter of the PSD (\( D_{MV} \), Fig. 3c) indicates that there is considerable growth through aggregation (deduced from particle images) until the first signs of melting are observed from the images. There is then a rapid decrease in \( D_{MV} \) when melting begins and a decrease in value as melting proceeds. The large fluctuations noted in \( D_{MV} \) in the rain region are due to some artifacts in the HVPS data and to enhancement of the water content in that region.

The condensed water content as measured by the CVI probe (CWC, Fig. 3d) shows fluctuations in the snow region, features that are due to gradients in the RH during each spiral. Note the strong decrease in the CWC due to the loss of condensate in the region of relatively low RH, indicating that there is considerable ice sublimation. Some of the decrease is also attributable to the increased fall velocity (to conserve flux at a higher fall velocity, CWC must decrease). The relationship noted between the air temperature and \( T_{IB} \), and the relative humidity and RH_{IB} and RH_{ICE} are similar in Figs. 4a and 4b to those
found in Figs. 3a and 3b. The enhanced region of CWC near the spiral completion in Fig. 4 shows the measurements for the second CRYSTAL-FACE spiral examined here, illustrating results similar to the first spiral. The data from this case are shown because, even though the RH (Fig. 4b) was a bit lower than spiral 1 and the corresponding CWC (Fig. 4d) decreased much more rapidly, the $D_{MV}$ (Fig. 4c) showed a steady increase even at $T > 0^\circ$C (Fig. 4a). This increase indicates the strong effects of aggregation throughout; the aggregation process allows the snow to fall to warmer temperatures than would otherwise be possible.

An examination of the ratio of the CVI CWC to the particle-estimated ice water content (IWC), which is calculated from the PSD assuming the masses of the particles are ice rather than water, is useful for identifying the extent of melting. Figure 5 shows the ratios for the three spiral descents as a function of $T$ (black symbols) during the spirals. The particle mass is derived using a power mass–dimensional relationship, $m = aD^b$, where $D$ is the maximum particle diameter, $a = 6.1 \times 10^{-3}$ g cm$^{-k}$ is the intercept, and $b = 2.05$ is the power. At $T < 0^\circ$C, where the ratio is close to 1, our mass–dimensional relationship evidently quite accurately represent the masses of the ice particles. What is clearly noted is that the ratio begins to increase above about 1 when $T$ exceeds about +1.5°C, indicating the onset of appreciable melting; the ratio progressively increases with increasing $T$. Also shown in Fig. 5 is the difference $RH_{IB} - RH$ for $T > 0^\circ$C. The RH is considerably below $RH_{IB}$ (positive values, spiral 1) and much less so (spirals 2 and 3). Consistent with the latter, the ratio of CVI CWC/IWC increases slightly and then remains close to 1 (spiral 1), and increases rapidly (spirals 2 and 3). Clearly, $RH_{IB}$ is a good indicator of the extent of melting.

The ratio of the CVI CWC as measured through the spirals to the IWC calculated from the PSD with overlays of the qualitative depiction of the degree of melting provides additional insight into how the mass–dimensional relationship of the particles might change during the melting process (Fig. 6)—this is discussed further in section 4. Our interpretation of where the melting commences and is completed from

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**Fig. 2.** Examples of ice crystal images obtained during CRYSTAL-FACE 26 Jul spiral 2. (top) Images from the 2D-C probe, with the scale shown to the right of the panel. Arrows and numbers indicate discernable, partially melted ice particles. (bottom) From the HVPS probe, with the scale shown to the right of the second panel. The horizontally aligned numbers to the left of each panel are the air temperature.
the particle images independently of these data suggest that our interpretation of the degree of melting is quite close to what actually occurred. Also, Figs. 6b and 6c support our interpretation that some ice still existed at temperatures as warm as +6°C to +7°C.

Attention now turns to two Lagrangian spiral descents through a high RH region during the KWAJEX field program on 30 August (Figs. 7 and 8). Melting is first noted at +1°C, and melting is complete by +3°C (Figs. 7a, 8a). The $T_{IB}$ curves fall to lower temperatures than the air temperature right when melting begins.
and the difference between the two increases as melting proceeds. Although the RH distribution is a bit noisy, the RH are considerably above \( \text{RH}_{\text{IB}} \) throughout, with the initial melting beginning when \( \text{RH} > \text{RH}_{\text{IB}} \) (Figs. 7b, 8b). Correspondingly, \( D_{\text{MV}} \) drops appreciably when melting begins, and the small values near the completion of the spiral suggest that there was considerable evaporation of the drops in the subsaturated air toward this end of the spiral (Figs. 7c, 8c). It is noted that the calculated IWC (Fig. 7d) decreases significantly in the rain region, with very little in the second spiral (Fig. 8d). Given that the RH is not far from water saturation, this decrease implies that the mass–dimensional relationship that we use for ice in spiral 1 does not adequately represent the masses of the melting and melted particles. For spiral 2, the IWCs in the ice regions were relatively low, and the use of the mass–dimensional relationship for ice was such that the resulting IWC was below 0.01 g m\(^{-3}\) and is not shown (Fig. 8d).

The relationship of RH to \( T \) for the Lagrangian spiral descents from CRYSTAL-FACE and KWAJEX are shown in...
Fig. 5. The ratio of the total condensed water derived from the CVI probe to the IWC derived from the particle probe as a function of the temperature for each of the spiral descents on 26 Jul CRYSTAL–FACE. This scale is shown on the bottom axis of the figure. The top axis of each panel shows how much the RH exceeds RH_{IB}(500 hPa) at the given temperature, implying sublimation. Vertical lines are identified in (b). At times, the CVI or the IWC were not measured (derived). That accounts for more points not plotted at temperatures above +3°C.
Fig. 9, along with the codes representing the four precipitation types derived qualitatively from the 2D images. Examination of Fig. 9 suggests that snow that falls to temperatures above freezing encounters an RH consistent with RHIB. In general, for example in nonconvectively generated ice cloud, it is likely that the RH at 0°C will be at 100%. However, an RH ∼50% would be unlikely, as it would be rapidly sublimating above the ML; but if that did occur then the ice would sublimate quickly once in the ML because of the low RH. Consider now what the snow encounters when the air temperature warms progressively to +6°C. Because of the dependence of RHIB on T, as the temperature warms, snow would be increasingly more likely with progressively decreasing RH. A point is reached, for instance when the air temperature exceeds +2°C, that an RH of 90% would also become less likely to have snow. Thus, the snow probability curves of RH versus T would bend back with increasing T and decreasing RH, which may be why the data for the spirals (Fig. 9) approximately straddle the RHIB curve.

Figure 10, derived for the 20 February 2020 case from the IMPACTS field program, shows an example of how melting proceeds when the RH is close to water saturation through the cloud layer sampled. During the descent, as the temperature climbed to just above 0°C, melting was quickly noted from the particle imagery (Fig. 10a). The relative humidity was at water saturation (or slightly above, due to some minor calibration uncertainty) (Fig. 10b). Figure 10c compares the CWC from the WISPER instrument to that derived from the PSD assuming that the particles are ice (and with the particle probes measuring both in horizontal and vertical orientations). At a temperature just at and above 0°C, notice the abrupt decrease in the IWC relative to the CWC. This is clearly shown in Fig. 10d, where there is an abrupt increase in the ratio WISPER CWC/IWC when melting is first noted. On the climb back up through the 0°C level, the same basic trends as described for the descent are noted. Because the relative humidity was close to 100% during the penetration at temperatures above...
0°C, the ice-bulb temperature and ice-bulb relative humidity are not plotted in the figure.

A summary of the number of observations at $T \geq 0$°C from all of the projects identified in section 2 and Table 1 is shown as a function of RH in Fig. 11a. The number of counts is highest for the mixed snow and rain and least for the mostly snow and rain categories. Relatively few counts are noted for any category at RH < 70%. An arbitrary threshold of 12 counts per RH interval is chosen for calculations shown later.
Figure 11b shows the individual data points plotted as a function of $T$ versus RH by category, with the curve for RHIB for a pressure level of 500 hPa plotted. The points for snow largely lie in the sublimation zone, with a progression further into the melting zone from the mostly snow to rain categories. The RHIB therefore provides a good means of characterizing the degree of melting.

Median values of the $T$–RH relationship for the composite dataset are derived for each of the precipitation categories in Fig. 12, accounting for the number of data points per 1°C $T$ and 5% RH intervals based on the discussion for Fig. 11a. One additional category, all mixed, encompassing all observations when snow and rain were observed simultaneously, is included in the figure. The error bars are derived by using the data from all field programs for each specific precipitation category. As expected, there is a migration from the snow to rain categories with increasing RH and temperature. The rain category approximately straddles the

Fig. 8. As in Fig. 7, but for spiral 2.
FIG. 9. Relative humidity measurements as a function of temperature for (a)–(c) CRYSTAL-FACE and (d)–(f) KWJEX. Qualitative estimate of the extent of melting is shown with colors, and RH_{500} with black lines.
RH_{IB} category and the “all mixed” category is not much different than the rain category.

b. Surface observations

Using the SAMSON database for the nine cities and time periods identified in section 2b, the hourly instances that the ice phase was identified by the ground observers was recorded, along with the ambient RH and $T$. Here, the resulting data were binned into intervals of 1°C $T$ and 5% RH. Periods when snow pellets were noted were extracted from the combined dataset. Additional analysis was conducted on this dataset, specifically those instances when snow pellets were observed.
because they are more likely to survive to warmer temperatures and lower RH than snow because of their higher terminal velocities and bulk ice density.

Figures 13 and 14 summarize the data for the precipitation events that were classified as having any ice phase observed, along with the ice-bulb relationship for 1000 hPa and one of the Matsuo and Sasyo (1981c) curves. Fig. 13 shows the number of ice phase events per RH and T intervals binned as above, and Fig. 14 shows the number of events observed for ice pellets. The T dependence of RHIB is plotted in the figures. Note that the T–RHIB curve does not apply directly to data from locations at altitudes above sea level pressures lower than 1000 hPa, such as data from Reno, Nevada. For simplicity, only the T–RHIB curve is plotted for 1000 hPa. The ice phase was indeed observed at T between +8°C and +9°C (Fig. 13), although these instances were rare. The largest number of points were, not surprisingly, observed close to the RHIB and at T between +4°C and +6°C, still straddling the RHIB curve. The Matsuo and Sasyo (1981c) curve fits the SAMSON data reasonably well in comparison with the data. Interestingly, ice pellets were observed much more frequently at RHIB in the melting zone below the RHIB curve. Their higher fall speeds and ice densities allowed them to be sustained in regions where the slower and less dense snow particles mostly melted.

Figure 15 shows the fraction of data points for snow reported in 10% RH intervals as a function of T. The ordinate in the figure is truncated at T = +6°C because there were too few points at temperatures above +6°C to draw contours. The numbers on the left-hand side of the figure shows the number of events per 1°C interval. In the 0°C to +1°C interval,
not surprisingly, the fewest number of events and the smallest fraction are found in the lowest RH interval, and the highest fraction of cases is in the 90%–100% RH interval. Clearly, snow is most likely to fall into the melting layer when the RH is high, but it is almost twice as much likely to persist when the RH < 90%. The fraction of events in the 40%–50% and 50%–60% RH ranges increase steadily with increasing T, while the highest RH ranges exhibit the opposite trend (Fig. 15). Note, however, that a measurement of RH and T at the ground may not accurately reflect the

![Graph showing the relationship between temperature and relative humidity.](image)

**Fig. 12.** Drawing on the points shown in Fig. 11b, median values of the qualitative estimate of the extent of melting of the snow as a function of the air temperature and binned into 5% intervals of relative humidity, along with the standard deviation. Also shown are the $T_{\text{inh}}$–$R_{\text{inh}}$ relationships developed for pressure levels of 1000 and 500 hPa.

![Graph showing the number of hourly data points.](image)

**Fig. 13.** Number of hourly data points in the composite SAMSON dataset derived here that contain ice-phase precipitation identified by NOAA surface observations, binned into 0.5°C intervals of temperature and 5% intervals of relative humidity. The ice-bulb RH is for a pressure level of 1000 hPa and one of the Sasyo and Matsuo curves are plotted.
RH distribution that the snow encountered as it fell below the ML.

4. Discussion

Some general comments about the relationship of ice phase precipitation to RHIB can be drawn from the in situ and SAMSON datasets. Our analysis of the SAMSON data indicates that for some component of snow to reach the ground, the greatest likelihood is that it falls from 90% to 100% RH layer (Fig. 15); while an RH of 40%–50% comprised only 0.1% of the data. If such low RH do occur for particles falling into the melting layer, the ice particles would rapidly sublime as they fell to warmer temperatures. Consider now what snow encounters when the air temperature warms progressively to +6°C and the RH is in the 70%–100% range. With an RH of 70%, snow would not melt until a temperature of about +5°C was reached. Conversely, if RH = 90% or above, melting would begin at higher temperatures.
would occur at $T < +1.5^\circ$C. In both cases, snow would persist until the RH$_{IB}$ curve was reached.

Two primary factors would affect the environmental conditions. First, moistening of the air due to the sublimation of ice would increase the RH (see Fig. 16) and the temperature would decrease because of cooling of the air due to the sublimation (Carlin and Ryzhkov 2019). Second, some indications from the spiral descents suggests that weak downdrafts or subsidence would decrease the RH (and somewhat increase $T$).

In general, these factors would increase the persistence of ice and lower the altitude of the melting. Once the RH$_{IB}$ curve was reached, melting would proceed; and at $T +1.8$ or $2.8^\circ$C higher than the RH$_{IB}$ curve, melting would be complete, depending on the particle type (degree of riming) and maximum particle size. To counter the melting effect, more cooling of the air due to the additional latent heat lost due to the melting of the snow would tend decrease the $T$, increasing the RH. In any event, the demarcation between snow and partially melted particles would be close to the RH$_{IB}$ curve, and perhaps complete melting by $1^\circ$ or $2^\circ$C warmer. This is largely borne out by both the in situ (Fig. 11b) and SAMSON (Fig. 13) data.

Matsuo et al. (1981) collected surface meteorological data at three weather stations in Japan in order to obtain the relationship between the occurrence of the types of precipitation of snow, sleet, and rain on the ground and surface meteorological elements. Their results indicated that the types of precipitation were dependent on the (surface) RH and $T$. They found that for the same $T$, the precipitation types varied depending on RH. They identified two “critical” RH categories. For the lower RH category, precipitation was all snow. Within some range above the critical humidity, precipitation could be snow, sleet, or rain. Above the higher RH category, only rain was observed. They developed relationships for each location to express the critical RH for snow and rain as a function of $T$.

Their results for snow only for one of the locations (the relationships differed by a median of RH $56\%$ over the full range of $T$) is shown in Fig. 16a, along with the RH$_{IB}$–$T$ relationship. Comparison with our SAMSON dataset, comprising ice phase events (Fig. 13), shows a general consistency with the Matsuo et al. (1981) data, specifically the large number of points in red text.

Of note in the Matsuo et al. (1981) study is the difference in the critical RH between the rain only and snow only thresholds. At an RH of 75%, the difference between rain and snow only for their three criteria is $\pm 0.2^\circ$C, and at 80% is about $+2^\circ$C. Our results (Fig. 12) suggest differences closer to about $+2^\circ$C. The difference is possibly due to the difference in the pressure levels of our observations, mostly at 500 hPa, and
their observations at the surface (see Fig. 1). For that reason, we have plotted ice-bulb relationships both for 500 and 1000 hPa. 

To make our results easier to use, linear fits to the $\text{RH}_{\text{IB}}-T$ curves (see Figs. 16a,b) are given for 1000 and 500 hPa by

$$\text{RH}_{\text{IB}}(1000 \text{ hPa}) = 100 - 12.67T, \quad (2a)$$

$$\text{RH}_{\text{IB}}(500 \text{ hPa}) = 100 - 9.37T. \quad (2b)$$

These relationships can be inverted to give $T_{\text{IB}} = f(\text{RH})$, where RH is in percent and $T$ is in degrees Celsius. However, we have tweaked them to give slightly better fits. These are given by

$$T_{\text{IB}}(1000 \text{ hPa}) = 7.90 - 0.079\text{RH}, \quad (3a)$$

$$T_{\text{IB}}(500 \text{ hPa}) = 10.70 - 0.107\text{RH}. \quad (3b)$$

The relationships given by (2) and (3) are plotted in Figs. 16a–d. Also plotted in Fig. 16a are the relationships developed by Matsuo and Sasyo (1981c).

Drawing on the airborne and surface observations, the following guidelines can be used to characterize the conditions “mostly snow” and “mostly rain.” The demarcation between sublimation and melting can be represented by the $\text{RH}_{\text{IB}}-T$ relationship. The category “mostly snow” $T + \Delta$, where $0^\circ < \Delta < +1^\circ$, “mostly rain” $+1^\circ < \Delta < +2^\circ$, and “rain” when $\Delta \geq +2^\circ$.

With knowledge of $T$ and RH, the degree of melting at a given RH can be found for surface conditions from

$$\text{RH}(1000 \text{ hPa}) = 100 - 12.67(T - \Delta T). \quad (4)$$

If $\Delta T < 1^\circ$C, then the precipitation is mostly snow, and if $\Delta T > 2^\circ$C, the snow has likely melted to rain.

The uniqueness of the CRYSTAL-FACE dataset, with its Lagrangian spiral descents and the availability of direct measurements of CVI CWC measurements, makes it amenable to the calculation of some interesting aspects of the melting of particles in highly subsaturated environments. The CVI data summarized in Fig. 6 were collected during the three spirals with average descent rates of $-1.4$, $-1.9$, and $-1.5 \text{ m s}^{-1}$, respectively, for $T < 0^\circ \text{C}$; and $-3.9$, $-3.5$, and $-1.8 \text{ m s}^{-1}$, respectively, for $T > 0^\circ \text{C}$. For two of the three spirals, the descent velocity reasonably approximated the terminal velocity of the snow ($T < 0^\circ \text{C}$), and for the melting snow ($T > 0^\circ \text{C}$). The time rate of change of the CVI CWC can therefore be used, as a first approximation, together with the air temperature, to estimate the rate of change of the RH due to the sublimating ice, in percent per second (Fig. 17). The lower abscissa shows the time rate of change of the CVI CWC as a function of $T$ (ordinate). During each loop of each spiral, there are clear peaks and minima in the relative humidity (e.g., Fig. 3b). Correspondingly, the rate of change of the CVI CWC was derived for the relative minima in the RH, the maxima, and the mean per loop (different colored lines in Fig. 17). Using the means value per loop, the median change in CVI CWC is $-0.00035 \text{ g m}^{-2} \text{ s}^{-1}$. The corresponding potential increase in the relative humidity is shown in Fig. 17 (upper abscissa), with a mean value of 0.0123% s$^{-1}$. Thus, it would take of order 80s for the sublimating ice to increase the RH by 1%. Subsidence would slow this rate of increase. Indeed, whereas there was no obvious downward moving air at temperatures below 0°C, downdrafts of $-1$ to $-2 \text{ m s}^{-1}$ were consistently observed at temperatures warmer than 0°C.

The CVI data from CRYSTAL-FACE also provides a unique opportunity to accurately determine the mass–dimensional power-law relationship for melting snow, $m = aD^b$. Schmitt and Heymsfield (2010) reported on the fractal geometry of simulated aggregates and how that can be used to derive the $a$ and $b$ coefficients. By creating theoretical aggregates of hexagonal crystals, it was found that the two- and three-dimensional fractal dimension values could be simply related. This relationship enabled the development of mass–dimensional relationships analytically from cloud particle images. The exponent in the mass–dimensional relationship, the fractal dimension, was found to be between 2.0 and 2.3 with a dependence on temperature noted for both datasets studied. Application of this method requires the relationship of particle area to diameter. To convert the power in the area–dimensional to the power in the mass–dimensional relationship, a value of 1.275 was found to work well when theoretical aggregates were combined with other aggregates.

In our study, a new approach is used to find $a$. The value of $b$ is derived both from a fractal analysis of the particle images at 5-s intervals and from the approximation noted above. The value of $a$ is determined by using $b$ from fractal geometry together with the particle size distributions to find the value of $a$ that matches the CVI-measured CWC.

The ratio of the coefficient $b$ as derived from the approximation noted above to that derived from the Schmitt and Heymsfield (2010) fractal geometry approach is very close to 1. From the fractal analysis, the coefficient $b$ increases slightly with $T$ from $-12^\circ$ to $0^\circ \text{C}$ (Fig. 18b). The mean value of $b$ in this temperature range is 2.33; it is 2.35 from the approximation. More variability is noted at temperatures above about $+3^\circ \text{C}$.

From the fractal analysis, the coefficient $b$ decreases slightly with increasing $T$ in the range $-12^\circ$ to $0^\circ \text{C}$ (Fig. 18c), with a mean value of 0.010 $\text{ g cm}^{-b}$. A progressively increasing value of $a$ is noted at temperatures above $+2^\circ \text{C}$, indicating that melting is increasing the mass of the snow particles, a result consistent with those shown Figs. 3c and 4c where the median volume diameter of the particle population decreases markedly at temperatures above about $+2^\circ \text{C}$.

5. Summary and conclusions

The basic physics of the melting process can be understood through consideration of the ice-bulb temperature. According to the American Meteorological Society Glossary of Meteorology, the definition of the ice-bulb temperature includes the wording, “In dry conditions the wet bulb of a psychrometer may freeze when the air temperature is appreciably above the freezing point.” That principle can be applied to the study here: in dry conditions at temperatures appreciably
Estimate of the Time Rate of Change of Ambient Relative Humidity

Fig. 17. Rate of change of the CVI TWC (lower abscissa) and estimate of the corresponding increase in RH I (upper abscissa) based on the CVI data from three CRYSTAL-FACE spiral descents for the 26 Jul case study.
FIG. 18. Estimate of the $a$ and $b$ coefficients in the mass–dimensional relationship for the CRYSTAL-FACE spirals. (a) Ratio of estimated $b$ coefficient to that derived from fractal geometry. (b) The $b$ coefficient derived from fractal geometry. (c) The $a$ coefficient derived from a combination of the particle size distributions, CVI TWC measurements, and the $b$ coefficient from fractal geometry.
above 0°C, ice hydrometers will not begin to melt until the RH exceeds the ice-bulb RH at the given temperature.

A combination of Lagrangian spiral descents, beginning in the ice regions at temperatures of −10°C or below and completed at temperatures of +6°C to +8°C, together with slow descents and ascents through the melting layer, are used to characterize the properties of the particle size distributions during the melting process. From the spirals, the melting process appears to follow the T–RH_P curve (Fig. 9). Aggregation continues to proceed during the melting process, possibly accounting for the observations the presence of ice at low humidities and temperatures as warm as +6°C. Taken as a whole, a good predictor of the presence of some component of ice in a melting particle follows the T–RH_P curve (Fig. 12). Although laboratory measurements have intrinsic value for capturing the melting process of individual ice particles, our measurements are more useful for characterizing the properties of ice particles in the melting layer because the particle populations comprise a wide range of sizes, masses, and habits, and natural environmental conditions.

Particles falling through the 0°C level on the way to the ground when the surface temperatures are as warm as +6°C can encounter a wide range of RH that determines whether the precipitation falls as snow, melting snow, or rain. Therefore, the RH measured at the surface may not be an indicator of whether or not there will be some or all ice phase precipitation. For that reason, we have also examined the SAMSON dataset to see how closely it relates to our findings from the in situ observations.

In general, the T–RH dependence noted for some component of ice phase precipitation from the SAMSON dataset examined here straddles the RH_P curve. At a given temperature, most instances we examined fall within ± 10% of the RH_P curve. Ice phase precipitation was reported at temperatures as warm as +6°C to +7°C, although rare and is observed primarily at temperatures above 5°C at substrate saturation. Ice pellets can survive to temperatures of +8°C to +9°C.

Future studies could examine the relationship between reports of ice precipitation at the surface, the associated RH, and from nearly coincident soundings the vertical distribution of the temperature and RH from the melting layer to the surface. This information would be useful for nowcasting applications.

Acknowledgments. We wish to thank Michael Poellot, University of North Dakota, for collecting and providing the UND Citation particle probe and state parameter data. We thank David Noone and Darin Toohey for access to the WISPER data. Meg Miller helped considerably with the editing of the manuscript and Roy Rasmussen and Morris Weisman reviewed this article prior to submission. Special thanks go to Robin Sydney for helping come up with the idea for this study. Dr. Schmitt was partially supported by DOE Grant DE-SC0016354 through a subaward from SUNY Albany. This work was supported by NASA through Grants 80NSSC002K0897 and 80NSSC19K0397. Alexander Theis acknowledges the support of the Deutsche Forschungsgemeinschaft under Grant SZ 260/6-1. The CRYSTAL-FACE data are available at https://espoarchive.nasa.gov/crycryst and the IPHEx, MC3E, and IMPACTS data at https://ghrc.nasa.gov/ (use the drop-down menu to find the appropriate field program). KWAJEX data are available at https://www.csc.uiuc.edu/field_projects/kwajex. Data for IMPACTS are available at https://ghrc.nasa.gov/uso/ds_details/collections/impactsCh.html. The SAMSON dataset is available at https://www7.ncdc.noaa.gov/CDO/cdoselect.cmd. We wish to thank Scott Stevens for his help in locating and downloading portions of the massive SAMSON dataset.

REFERENCES


