Relative Humidity in Liquid, Mixed-Phase, and Ice Clouds

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ABSTRACT

The results of in situ observations of the relative humidity in liquid, mixed, and ice clouds typically stratiform in nature and associated with mesoscale frontal systems at temperatures $-45^\circ C < T < -5^\circ C$ are presented. The data were collected with the help of instrumentation deployed on the National Research Council (NRC) Convair-580. The length of sampled in-cloud space is approximately $23 \times 10^3$ km. The liquid sensor was calibrated in liquid clouds with the assumption that the air in liquid clouds is saturated with respect to water. It was found that the relative humidity in mixed-phase clouds is close to saturation over water in the temperature range from $-5^\circ$ to $-35^\circ$C for an averaging scale of 100 m. In ice clouds the relative humidity over ice is not necessarily equal to 100%, and it may be either lower or higher than saturation over ice, but it is always lower than saturation over water. On average the relative humidity in ice clouds increases with a decrease of temperature. At $-40^\circ$C the relative humidity over ice is midway between saturation over ice and liquid. A parameterization for the relative humidity in ice clouds is suggested. A large fraction of ice clouds was found to be undersaturated with respect to ice. The fraction of ice clouds undersaturated with respect to ice increases toward warmer temperatures.

1. Introduction

Water vapor inside clouds plays a crucial role in their life cycle and precipitation formation. The understanding of the relationships between the vapor pressure and microphysical characteristics of clouds is one of the key questions of theoretical and applied physics of clouds. Knowledge of the humidity inside clouds is important for mesoscale and climate models. The in-cloud water vapor pressure is commonly assumed to be saturated with respect to liquid water in liquid clouds, and saturated with respect to ice in ice clouds. The humidity in mixed-phase clouds has been debated in the cloud community for years. The water vapor pressure ($E_w$) in clouds is generally approximated as a weighted average of the respective saturation values over liquid water ($E_{ws}$) and ice ($E_{is}$)

$$E_w = fE_{ws} + (1-f)E_{is}.$$  

where $f$ is the weighting factor, such that $f = 1$ for liquid clouds, $f = 0$ for ice clouds, and $0 < f < 1$ in mixed-phase clouds. The value of $f$ in mesoscale and global circulation models (GCMs) is usually specified as a function of temperature (e.g., Fowler et al. 1996; Jakob 2002) or cloud liquid (LWC) and ice (IWC) water content (e.g., Lord et al. 1984; Wood and Field 2000). In some numerical schemes (Rotstyan et al. 2000; Tremblay and Glazer 2000) the water vapor in mixed clouds is assumed to be saturated with respect to liquid water; that is, $f = 1$.

Korolev and Mazin (2003) developed a theoretical framework showing that the humidity in mixed-phase clouds is close to the saturation over water, and in ice clouds it can be either lower or higher than saturation over ice but always lower than saturation over water.

There are only a few works related to in situ observations of the humidity in clouds (e.g., Heymsfield and Miloshevich 1995; Gierens et al. 1999; Ovarlez et al. 2002; Strom et al. 2003). Most of these measurements are related to upper-tropospheric cirrus clouds. Thus, observations conducted by Ovarlez et al. (2002) and Strom et al. (2003) showed that the relative humidity
with respect to ice (RH) in cirrus clouds below −40°C may vary from 50% to 150%. Fu and Hollars (2004) presented results of measurements of humidity in Arctic clouds during the First International Satellite Cloud Climatology Project (ISCCP) Regional Experiment Arctic Cloud Experiment (FIRE-ACE) project. They found that in mixed-phase clouds the water vapor pressure can be well parameterized by using the weighting coefficient equal to liquid fraction $f = LWC/(LWC + IWC)$.

The present work studies the relative humidity in liquid, ice, and mixed-phase clouds from in situ measurements. Sections 2 and 3 describe instrumentation used for the measurements and dataset. Section 4 describes a technique for in situ calibration of the humidity sensor. The results of the humidity measurements in liquid, mixed-phase, and ice clouds are discussed in sections 5, 6, and 7.

2. Instrumentation

The cloud microphysical and thermodynamical instrumentation was deployed on the National Research Council (NRC) Convair-580. A description of the probes used in this study is given below.

a. Temperature measurements

The air temperature ($T_a$) was measured by three different probes: two Rosemount total-air temperature probes (model 102DJ1CG; Lawson and Cooper 1990; Frihe and Khelif 1992) and one reverse-flow probe (Rodi and Spyers-Duran 1972; Lawson and Cooper 1990). One of the Rosemount temperature probes was mounted on a boom under the wing, positioned approximately 1 m ahead of the wing’s leading edge. The second Rosemount probe and the reverse-flow sensor were installed underneath the central section of the wing. The reverse-flow and the “wing” Rosemount sensors were calibrated in a methanol stirring bath at temperatures $+50°C > T > −50°C$. During flights the constant biases between these two sensors did not exceed 0.5°C. The “boom” Rosemount sensor was biased by approximately 1.8°C with respect to the other two sensors. This bias was removed during the data processing. The processing and corrections of the temperature measurements were done as described by Williams and Marcotte (2000). In clear sky the standard deviation of the temperature differences between each pair of sensors usually did not exceed 0.08°C. In supercooled liquid and mixed clouds, ice built up on the housing of the reverse-flow probe, which resulted in changing the recovery factor, and eventually degraded the measurements. Both Rosemount sensors were deiced, and they were not subjected to icing in supercooled clouds. No noticeable effect of wetting (e.g., Heymsfield et al. 1979; Lawson and Cooper 1990) was found in supercooled liquid clouds for either of the Rosemount temperature sensors. This observation is in agreement with earlier studies by Lawson and Rodi (1992), who established that in supercooled clouds wetting has a negligible effect on the Rosemount temperature. Figures 1a,b present the distribution and time series of the temperature difference between two Rosemount temperature sensors mounted on the boom and under the wing. The temperature was measured during a descent though a sequence of liquid and ice clouds (Figs. 1c,d). Under a variety of different conditions the amplitude of the temperature differences does not exceed $±0.2°C$ with a standard deviation of 0.06°C (Fig. 1). The temperature difference distribution (Fig. 1a) suggests that the average random error in the temperature measurements can be estimated as approximately $±0.1°C$. The absolute error of the air temperature measured by the Rosemount sensor based on the previous studies is estimated as $±0.5°C$ (Lawson and Cooper 1990; Lawson and Rodi 1992; Frihe and Khelif 1992). The Rosemount sensor mounted on the boom was chosen as a primary for the data analysis, since it was less affected by radio transmissions and other sources of electromagnetic noise causing sporadic spikes in the measurements, as compared to the wing Rosemount sensor.

b. Water vapor measurements

The water vapor concentration was measured by a Licor HO$_2$ analyzer (model LI-6262, LI-COR, Inc.). The principle of operation of the Licor is based on measurements of the absorption of infrared radiation by moist air consecutively passing through a sampling section and calibrating cells (Licor 1996). The Licor probe was mounted inside the fuselage and the air was pumped into the sample cell through a 2-m-long pipe. From outside the fuselage, air was sucked in through a reverse flow inlet, to stop cloud droplets from entering the pipe. If cloud droplets entered the Licor pipeline and evaporated, the absolute humidity would increase. The relative humidity, recalculated back to outside temperature $T_a$, would result in an artificially high supersaturation of the order of tens and hundreds of percent. The absence of such observations gives solid evidence that cloud droplets did not affect the Licor measurements. The declared Licor bench accuracy for measurements of absolute humidity is 1%, which corresponds to a dewpoint accuracy of approximately $ΔT_d = 0.1°C$ at the air temperature $−30°C < T_a < −5°C$. For water vapor pressure 20 mb the peak-to-
peak noise of the Licor signal in terms of dewpoint temperature corresponds to $\Delta T_d = 0.02^\circ$C for 1-s averaging. The time response of the Licor probe, including the transition through the pipe, is estimated as 0.2 s (J. I. MacPherson 2005, personal communication). The dew $T_d$ and frost point $T_f$ temperatures were deduced from the absolute humidity using the Buck (1981) equations. The accuracy of the relative humidity calculations derived from the Licor and air temperature measurements is discussed in sections 4 and 5.

c. Cloud water content measurements

The LWC, the IWC, and the ice water fraction ($\mu = IWC/TWC$) were deduced from the measurements of the Nevzorov probe (Korolev et al. 1998); here $TWC = LWC + IWC$ is the total water content. Calculations of the LWC and IWC were made following the procedure described in Korolev et al. (2003). It was shown in a number of studies that ice particles may cause a response of the LWC sensor (e.g., Korolev et al. 1998; Korolev and Strapp 2002; Field et al. 2004). The residual effect of ice on the LWC sensor is due to the small amount of heat removed from the LWC sensor during impact with ice particles. The residual effect can be characterized by the residual coefficient $\beta = W_{\text{liq}}^{\text{res}}/W_{\text{ice}}$, where $W_{\text{liq}}^{\text{res}}$ is the response of the LWC sensor from ice particles with corresponding ice water content $W_{\text{ice}}$. In practice, the residual coefficient is usually estimated as $\beta = W_{LWC}/W_{TWC}$; here $W_{LWC}$ and $W_{TWC}$ are the water contents measured in ice clouds by the LWC probe.
and TWC sensors, respectively, where LWC = 0 a priori. The value of the residual coefficient $\beta$ depends on the size, shape, and bulk density of ice particles, airspeed, air, and sensor temperatures, and it may somewhat vary from 0.01 to 0.5. Large $\beta$ are typical for a high concentration of small ice particles at the tops of thunderstorms (Strapp et al. 1999). In midlatitude frontal clouds, the residual coefficient is usually close to $\beta = 0.18$, which was the value used for this study.

### d. Cloud particle measurements

Concentration and sizes of cloud droplets were measured by two Particle Measuring Systems (PMSs) forward scattering spectrometer probes (FSSP-100s; Knollenberg 1981), operated in the size ranges 3–47 and 5–95 $\mu$m. Regular maintenance and calibration of the FSSP probes were carried out in order to minimize the potential errors in droplet sizing. The dead time losses and coincidence errors were taken into account during data processing (Baumgardner et al. 1985). The FSSP data were used for calculations of the integral radius of cloud droplets ($N\tau_w$) in liquid clouds; here $N$ and $\tau_w$ are the concentration and average radius of cloud droplets, respectively, over the FSSP size range. The accuracy for the measurement of $N\tau_w$ is estimated at approximately 20%.

Large cloud particles were measured by 2D-imaging optical array probes (OAPs): PMS OAP-2DC (25–800 $\mu$m); a PMS OAP-2DP (200–6400 $\mu$m; Knollenberg 1981); and the SPEC, Inc., High Volume Spectrometer Precipitation Spectrometer (HVPS; 200 $\mu$m–4 cm; Lawson et al. 1998). All three instruments provided shadow binary images and concentrations of hydrometeors within their respective size ranges. The data of the 2D-imaging probes were averaged over 4-s time intervals. Such an averaging time is usually enough to provide a statistically significant number of images for estimations of the particle concentration and habit recognition analysis.

The OAP data were used for identification of ice clouds (section 2f) based on the particle image recognition and concentration calculations. The 2D data were analyzed with the help of a 2D software package developed by Sky Tech Research (www.skytechresearch.com).

### e. Detecting of liquid containing clouds

The Rosemount Icing Detector (RICE) was used for identifying the presence of liquid phase in clouds. The rate of change of the RICE signal ($dV_R/dt$) characterizes the rate of ice accretion on the rod surface ($d\ell/dt$), which can be recalculated to LWC (e.g., Baumgardner and Rodi 1989; Cober et al. 2001; Mazin et al. 2001). The accuracy of the RICE LWC measurements is relatively low, and it is no better than 30%. However, the RICE is rather sensitive to the presence of liquid droplets in clouds and is relatively insensitive to the presence of ice particles. The sensitivity threshold of the RICE is defined by the rate of sublimation of the accreted ice and adiabatic heating at the rod surface. At $100\text{ m s}^{-1}$ the threshold sensitivity is estimated as 0.005 and 0.002 g m$^{-2}$ at $T = -10^\circ\text{C}$ and $T = -20^\circ\text{C}$, respectively (Mazin et al. 2001). Practically, the RICE sensitivity threshold is limited by the noise, related to electronics and interaction of ice particles (Heymsfield and Miloshevich 1989; Strapp et al. 1999). Cober et al. (2001) showed that in clear sky the noise of the RICE is $|dV_R/dt| < 5\text{ mV s}^{-1}$. In ice clouds the noise level is higher than in clear sky because of the interaction of the RICE rod with ice particles. Figures 2b,c,d show the time series of the RICE signal, LWC, and IWC measured during a traverse through precipitating cells with IWC exceeding 1 g m$^{-3}$. As seen from Figs. 2a,b, in ice clouds the fluctuations of $dV_R/dt$ reach 10 mV s$^{-1}$. This value was set as a threshold for the presence of liquid in clouds; that is, clouds with $dV_R/dt > 10\text{ mV s}^{-1}$ were considered as containing liquid droplets, whereas when $dV_R/dt < 10\text{ mV s}^{-1}$ clouds were assumed as glaciated. The RICE signal 10 mV s$^{-1}$ is equivalent to approximately LWC $\sim 0.01\text{ g m}^{-3}$ (Mazin et al. 2001).

The dead time of the RICE consists of deicing heating ($\Delta t_h = 4\text{s}$) with the following period of cooling ($\Delta t_c$) (Heymsfield and Miloshevich 1989; Cober et al. 2001; Mazin et al. 2001). During the cooling period the RICE does not adequately respond to the presence of supercooled liquid and $dV_R/dt$ may exceed 10 mV s$^{-1}$ even in cloud-free air. (Examples of $dV_R/dt > 10\text{ mV s}^{-1}$ in clear sky can be seen in Fig. 7 around 2141 and 2143 UTC, as indicated by the gray bands.) The cooling time depends on LWC, temperature, pressure, and aircraft speed. For the data collected in the present study, the average cooling time was approximately $\Delta t_c = 7\text{s}$. During the data processing the periods of the dead time $\Delta t_h + \Delta t_c$ were automatically excluded from the analysis. Depending on the conditions, the dead time may make up to 90% of the RICE measurement cycle. Under these circumstances a large fraction of data would be excluded from the analysis. To avoid these losses, some extended liquid cloud zones were manually included in the analysis based on the Nevzorov LWC measurements.

At temperatures $T > -4^\circ\text{C}$ the RICE has a reduced response to LWC due to the Ludlam limit (Cober et al. 2001; Mazin et al. 2001). Besides that, from our observation at $T > -2^\circ\text{C}$, ice may build up on the RICE
cylinder in ice clouds with no liquid because of refreezing of the melted water on the rear side of the cylinder. For these reasons the RICE was used for detecting liquid-containing clouds only for the cases with $T < -5^\circ C$.

f. Identifying liquid, mixed-phase, and ice cloud zones

Because of the residual effect of ice on the LWC sensor, ice particles may mask the presence of a small amount of liquid water in mixed clouds in the Nevzorov probe measurements. The ambiguity in segregating ice and low liquid mixed clouds was solved with the help of the RICE, which was used as a detector of clouds with $LWC > 0.01 \text{ g m}^{-3}$ (i.e., $dV_R/dt > 10 \text{ mV s}^{-1}$). Based on this, liquid and mixed clouds were defined from the Nevzorov and RICE measurement using the following conditions. If the ice water fraction $\mu < 0.1$ and $dV_R/dt > 10 \text{ mV s}^{-1}$, then the cloud was considered liquid; if $0.9 \geq \mu \geq 0.1$ and $dV_R/dt > 10 \text{ mV s}^{-1}$, then the cloud was considered as mixed phase. Clouds with the con-
Table 1. Length of measurements in liquid, mixed, and ice clouds for different temperature intervals.

<table>
<thead>
<tr>
<th>Temperature</th>
<th>Liquid clouds (LWC &gt; 0.01 g m⁻³) (km)</th>
<th>Mixed clouds (LWC &gt; 0.01 g m⁻³) (km)</th>
<th>Ice clouds (N &gt; 10 m⁻³) (km)</th>
<th>Total (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>−5° ≤ T &lt; −10°C</td>
<td>798</td>
<td>900</td>
<td>4046</td>
<td>5744</td>
</tr>
<tr>
<td>−10° ≤ T &lt; −15°C</td>
<td>690</td>
<td>473</td>
<td>4893</td>
<td>6056</td>
</tr>
<tr>
<td>−15° ≤ T &lt; −20°C</td>
<td>918</td>
<td>181</td>
<td>3601</td>
<td>4700</td>
</tr>
<tr>
<td>−20° ≤ T &lt; −25°C</td>
<td>172</td>
<td>94</td>
<td>2121</td>
<td>2387</td>
</tr>
<tr>
<td>−25° ≤ T &lt; −30°C</td>
<td>137</td>
<td>19</td>
<td>1348</td>
<td>1504</td>
</tr>
<tr>
<td>−30° ≤ T &lt; −35°C</td>
<td>14</td>
<td>2</td>
<td>1063</td>
<td>1079</td>
</tr>
<tr>
<td>−35° ≤ T &lt; −40°C</td>
<td>0</td>
<td>0</td>
<td>450</td>
<td>450</td>
</tr>
<tr>
<td>−40° ≤ T &lt; −45°C</td>
<td>0</td>
<td>0</td>
<td>920</td>
<td>920</td>
</tr>
<tr>
<td>−5° ≤ T &lt; −45°C</td>
<td>2729</td>
<td>1669</td>
<td>18 442</td>
<td>22 840</td>
</tr>
</tbody>
</table>

The concentration of ice particles $N_{ic} > 10$ m⁻³ and $dV/dt < 10$ mV s⁻¹ were determined as ice clouds. The concentration of ice particles $N_{ic}$ larger than 50, 200, and 400 μm were calculated from the OAP-2DC, OAP-2DP, and HVPS, respectively. Because of large statistical errors in measuring low particle concentration, ice clouds with $N_{ic} < 10$ m⁻³ were not considered here. It worth mentioning that ice clouds with $dV/dt < 10$ mV s⁻¹ may still contain some liquid water.

Because of the reduced response of the LWC sensor to large drops ($D > 100$ μm), the Nevzorov TWC sensor may measure a higher value of water content than that measured by the Nevzorov LWC sensor in drizzling clouds (Korolev et al. 1998; Strapp et al. 2003). As a result, the measured ice water fraction may become $\mu > 0.1$, and the liquid drizzling clouds may be classified as mixed or ice. The drizzling clouds were identified with the help of habit recognition analysis of 2D images (Korolev and Sussman 2000). If the fraction of circular images in the drizzling clouds exceeded 70%, then such clouds were identified as liquid.

4. Relative humidity calculations

4.1. Basic assumptions

The equation describing changes of the water vapor supersaturation in liquid clouds was originally deduced by Squires (1952). Korolev and Mazin (2003) generalized this equation for clouds with any phase composition. The solution of the Squires equation yields that supersaturation approaches its steady-state value

$$S_{qs} = a_0 \mu_x - b^* N_f r_i / b_f N_{ic} r_w + b_f N_f r_i,$$

with the phase relaxation characteristic time

$$\tau_p = 1 / a_0 \mu_x + b_f N_{ic} r_w + (b_f + b^*) N_f r_i;$$

4.2. Field campaign results

The data were collected during three field campaigns: FIRE-ACE in April 1998 over Canadian North and Arctic Ocean, and the Alliance Icing Research Study projects (phases 1.5 and 2) over southern Ontario and Quebec during two winter seasons, 2002/03 and 2003/04 (Isaac et al. 2001, 2005). The bulk of the data was sampled in stratiform clouds (St, Sc, Ns, As, Ac, and Ci), associated with frontal systems. During sampling the airspeed of the airplane was approximately 100 m s⁻¹. The measurements were averaged over 1-s time intervals, which correspond to the spatial resolution of approximately 100 m at the Convair-580 airspeed. The total number of flights included in the analysis is 36, with the total in-cloud portion analyzed here being approximately 22 840 km. The temperature was limited to the range −5° to −45°C. The altitude of measurements ranged from 0 to 7 km. Table 1 shows the length of the in-cloud legs for liquid, mixed, and ice clouds at different temperature intervals. The definitions of liquid, mixed, and ice clouds used in this study are provided in sections 2f and 8a. Table 1 enables an estimate of the statistical significance of measurements for different phase and temperature categories. Approximately 50% of the data were collected at temperatures in the range of −15° < T < −5°C.
ation in liquid clouds is mainly defined by the integral radius $N_w r_w$. In stratiform clouds the integral radius typically varies in the range $10^2 < N_w r_w < 2 \times 10^3 \mu$m cm$^{-3}$ (see section 5), which corresponds to the time of phase relaxation $1 \text{s} < \tau_p < 40 \text{s}$, at $-30^\circ < T < 0^\circ \text{C}$, $350 \text{mb} < P < 1000 \text{mb}$. The characteristic lifetime of stratiform clouds $\tau_{cl}$ is of the order of tens of minutes and hours, and therefore $\tau_{cl} \gg \tau_p$. This inequality suggests that the supersaturation $S$ has enough time to relax to the quasi-steady value $S_{qs}$ after being disturbed by vertical turbulent motions or entrainment of dry air. In stratiform clouds the average component of the vertical velocity is $\bar{w} \sim 0$, whereas the velocity related to vertical turbulent fluctuations is usually $|u'| < 1 \text{m s}^{-1}$ (Mazin et al. 1984; Mazin and Smeter 1989). Substituting $\bar{w}$ and $|u'|$ in Eq. (2) yields a quasi-steady supersaturation related to the regular and turbulent vertical motions of $\bar{S}_{qs} = 0$ and $|S'_{qs}| < 3\%$ for all liquid clouds with $10^2 < N_w r_w < 10^3 \mu$m cm$^{-3}$, $-30^\circ < T < 0^\circ \text{C}$, $350 \text{mb} < P < 1000 \text{mb}$.

Thus, the turbulent fluctuations in liquid stratiform clouds would have no significant effect on $\text{RH}_w$, and, with a high degree of accuracy, it can be assumed that the average relative humidity is $\text{RH}_w = 100(1 + \bar{S}_{qs}) \equiv 100\%$. This is a basic assumption behind the humidity processing in the following sections. It should be noted that this assumption may not be true near the cloud boundaries, or in the zones of liquid precipitations underneath clouds, where the inequality $\tau_{cl} > \tau_p$ is invalid.

### b. Calibration of the relative humidity

Because of the errors of the absolute calibration of the air temperature and humidity sensors, the systematic error of $T_a$, $T_d$, and $T_f$ is of the order of $1\%$. The fine tuning of the humidity measurements was conducted in liquid clouds using the assumption made in section 4a that $\text{RH}_w = 100\%$, and therefore $T_d = T_a$. Figure 3 shows the scatterplot of $T_a$ and $T_d$ measured in liquid clouds during one of the flights. The liquid clouds were identified as described in section 2f. As seen in Fig. 3, the measured $T_a$ and $T_d$ are well aligned along the same line $T_d = aT_a + b$. This suggests that the scaling coefficients $a$ and $b$ remain constant for different $T_a$. It was found that the coefficient $a$ changes from flight to flight within the range $0.98 < a < 1.14$, whereas it stays constant during the same flight. The calibration procedure of the Licor measurements consisted of two steps: 1) adjusting the scaling coefficient $a$ in such a way that the scatterplot $T_a$ versus $T_d$ becomes aligned parallel to the 1:1 line; 2) removing the bias offset $b$ by forcing $T_d$ to become equal to $T_a$ at one of the points in a liquid cloud. The bias offset $b$ usually does not exceed $2^\circ \text{C}$.

Since the random errors of humidity measurements ($\Delta T_a = \pm 0.02^\circ \text{C}$; section 2b) are lower than that of the air temperature ($\Delta T_a = \pm 0.1^\circ \text{C}$; section 2a), the random errors of the relative humidity $\Delta \text{RH}_w$ will be mainly defined by the errors $\Delta T_a$. At $T_a = 35^\circ \text{C}$ and $T_a = -5^\circ \text{C}$ the errors $\Delta T_a = \pm 0.1^\circ \text{C}$ correspond to the errors $\Delta \text{RH}_w \equiv \pm 0.3\%$ and $\Delta \text{RH}_w \equiv \pm 0.75\%$, respectively.

Thermodynamic equations show that the difference $T_f - T_d$ decreases, and the relative humidity at the saturation over ice $\text{RH}_{w, s}$ approaches 100% when the temperature approaches to 0°C. For example, $T_f - T_d \equiv 0.6^\circ \text{C}$ and $\text{RH}_{w, s} = 95\%$ at $T_a = -5^\circ \text{C}$. The relative error in segregation between saturation over liquid and ice becomes too large in the vicinity of 0°C. For the above reason and taking into account the arguments in section 2c regarding accuracy of the RICE at subzero temperatures, our analysis of the relative humidity in clouds only included temperatures $T_a < -5^\circ \text{C}$.

### 5. Relative humidity in liquid clouds

Figure 4 shows the frequency distribution of the relative humidity in liquid clouds with the time of the phase relaxation $\tau_p < 5 \text{s}$, air temperature $-20^\circ < T_a < -5^\circ \text{C}$, ice fraction $\mu < 0.1$, and water content $\text{LWC} > 0.01 \text{g m}^{-3}$. The maximum of the distribution is centered at $\text{RH}_w \sim 99.8\%$, with the average $\text{RH}_w = 99.18\%$ and
standard deviation (STD) of 1.22%. This value of the STD gives an upper estimate of the error of the relative humidity measurements. It is also consistent with the estimate of ΔRHₚ in section 2b.

The selection of clouds with small τₚ is stipulated by the choice of clouds with RHₚ a priori close to 100%. With a small time of phase relaxation, cloud droplets will quickly accommodate the fluctuation in relative humidity and approach RHₚ = 100%.

As seen from Fig. 4, the distribution of the relative humidity is skewed toward low RHₚ. The asymmetry of the RHₚ distribution can be explained by the presence of cloud holes generated by the entrainment of dry air (Korolev and Mazin 1993; Gerber et al. 2005). Based on the analysis of the FSSP data at 10 m (0.1 s) spatial resolution, Korolev and Mazin (1993) found that the cloud holes on average occupy 3%–7% of frontal stratiform clouds. The actual fraction of undersaturated zones in liquid clouds may be higher than that, since some cloud holes smaller than 10 m were missed because of the coarse FSSP spatial resolution (10 m). The dimensions of the undersaturated zones inside clouds may go down to Kolmogorov viscous microscale (10⁻³ m). The presence of pockets of dry air in liquid clouds will result in the reduction of measured RHₚ. In the present study, the data were averaged over 100-m (1 s) intervals. At that spatial resolution cloud holes with horizontal dimensions up to 200 m may be identified as in-cloud zones. Low values of relative humidity 94% < RHₚ < 98% (Fig. 4) can also be explained by a sampling of dry air during transit though the cloud interfaces, when some fraction of the air may be sampled outside the cloud.

Assuming that the average RHₚ should be close to 100%, the distribution of RHₚ > 100% can be considered as a result of the instrumental errors. Extrapolating this distribution toward a lower humidity RHₚ < 100%, symmetrically with respect to RHₚ = 100%, yields an RHₚ distribution distorted as a result of instrumental errors (Fig. 4, area 1). Measurements in area 2 are a result of the reduction of the RHₚ due to averaging over clouds containing cloud holes and in the vicinity of cloud boundaries.

Figure 5 shows a scatterplot of RHₚ versus NₚTₚ and τₚ. The solid lines indicate the average RHₚ. Most values of the integral radius in stratiform supercooled liquid clouds are in the range of 200 < NₚTₚ < 1500 μm cm⁻³ (Fig. 5a), which corresponds to the range of the time of phase relaxation 15 > τₚ > 2 s (Fig. 5b). The maximum value of NₚTₚ reaches 2500 μm cm⁻³ (Fig. 5a). The dispersion of RHₚ values increases toward large NₚTₚ, and the averageRHₚ increases with an increase of NₚTₚ (Fig. 5a) or decrease of τₚ (Fig. 5a). Such behavior is in agreement with the theoretical predictions by Korolev and Mazin (2003), since the probability of deviation of RHₚ from 100% increases with an increase of τₚ (decrease NₚTₚ).

Vertical and horizontal dashed lines in Fig. 5 separate the scatter field into conditional areas of “liquid clouds,” “cloud boundaries,” and “supercooled large drops (SLD) and drizzle.” In the liquid cloud area the relative humidity stays close to 100% since τₚ is relatively small. In cloud boundaries τₚ is still small (τₚ < 40 s, NₚTₚ > 100 μm cm⁻³), whereas the humidity RHₚ < 97%, which exceeds the range of errors in the measurements. The relative humidity in this area is affected by entrainment of dry air, the presence of cloud holes, and the closeness to cloud boundaries.

The analysis of the data from SLD and drizzle showed that most of the cases with NₚTₚ < 100 μm cm⁻³ and τₚ > 40 s (Fig. 5) are related to clouds consisting of drizzle measured underneath the cloud base, when the drops fall down through undersaturated air. As seen in Fig. 5 in the SLD and drizzle area, the average humidity rapidly decreases to RHₚ = 96%.

Figure 6 shows the dependence of RHₚ versus τₚ for different temperature ranges in liquid clouds. As in Fig. 5b there is a clear tendency for RHₚ to decrease as τₚ increases. The dependence of RHₚ versus τₚ is approximately the same for all temperature intervals within the errors of measurements.
6. Relative humidity in mixed-phase clouds

Figure 7 shows the time series of the RICE signal, LWC, IWC, T_a, T_d, and T_f during a transit through a sequence of liquid, ice, and mixed-phase clouds. As seen in Fig. 7, in mixed-phase zones with different ice water fractions, T_a is equal or close to T_d. This suggests that in mixed-phase clouds RH_w ≅ 100%.

Figure 8 shows the dependence of RH_w in mixed-phase clouds versus the ice water fraction f/H9262 for different temperature intervals averaged over clouds with a total sampling length of 2700 km. There is a slight trend for RH_w to decrease toward ice clouds in all temperature intervals; that is, RH_w decreases approximately 2% when the ice fraction f changes from 0.05 to 0.95 for the temperatures -25°C < T_a < -5°C. These changes are close to the accuracy of the measurements. Practically, the relative humidity in mixed clouds can be considered as RH_w = 100% and independent of the ice fraction. This experimental finding is in agreement with the theoretical calculations by Korolev and Mazin (2003).

If RH_w is assumed to be weighted by IWC and LWC [Eq. (1)] in mixed clouds, then substituting f/IWC/LWC in Eq. (1) yields

\[ \text{RH}_{w} = 1 - \mu + \mu \text{RH}_{w_{\text{vis}}}. \]  

(5)

Here, RH_{w_{\text{vis}}} = E_{\text{w}}/E_{\text{vis}} is the relative humidity with respect to liquid water at saturation over ice. The dashed lines in Fig. 8 show RH_w(μ) calculated from Eq. (5) for different temperatures. Our airborne measurements show that the parameterization in Eq. (1) or Eq. (5) do not adequately describe RH_w in mixed-phase clouds for the averaging scale of 100 m.

7. Relative humidity in ice clouds

Figure 9 shows a time series of the RICE signal, LWC, IWC, T_a, T_d, and T_f during a flight in liquid and ice clouds. The agreement between T_a and T_d in liquid clouds on the left of Fig. 9 points out the correctness of the in situ humidity calibration and suitability of the following humidity measurements. Korolev and Mazin (2003) proposed that, since the time of phase relaxation is usually of the order of minutes and hours in ice...
clouds, the relative humidity may be both higher and lower than saturation over ice, but always lower than the saturation over water. This hypothesis found good confirmation from measurements in this study. As seen on the right of Fig. 9 in the ice cloud, the air temperature may be both $T_a > T_f$ and $T_a < T_f$; that is, $\text{RH}_i > 100\%$ and $\text{RH}_i < 100\%$, respectively.

Figure 10 shows the dependence of the average relative humidity with respect to ice $\text{RH}_i$ versus the cloud temperature $T_a$. The most interesting finding is that $\text{RH}_i$ has a distinct tendency to increase with a decrease of air temperature. At $-40^\circ\text{C}$ the average relative humidity is approximately midway between saturation over water and saturation over ice. At the moment there is no clear explanation for such behavior. In section 8c it will be discussed that the increase of $\text{RH}_i$ with the decrease of $T$ may be related to type of clouds, and it is typical for deep frontal cloud systems. The average relative humidity in ice clouds can be parameterized as

\[
\text{RH}_i = \frac{100(1 - \alpha + \mu \text{RH}_i)}{\text{Eq. (5)}}
\]

**Fig. 7.** Time series: (a) RICE signal; (b) IWC measured by the Nevzorov probe, OAP-2DC, and HVPS; (c) LWC measured by the Nevzorov probe; and (d) air, dewpoint, and frost point temperatures measured during a traverse through ice, liquid, and mixed-phase clouds. The measurements were obtained using the NRC Convair-580 during AIRS2 on 22 Jan 2004 in southern Ontario in convective cells.

**Fig. 8.** Dependence of the average humidity $\text{RH}_l$ vs ice water fraction $\text{IWC}/(\text{LWC}+\text{IWC})$ for different temperature intervals measured in mixed-phase clouds. Dashed lines correspond to the parameterization $\text{RH}_l = 100(1 - \alpha + \mu \text{RH}_i)$ [Eq. (5)]. Vertical line on the left side represents an error bar.
a function of the air temperature, where $T_a$ is in degrees Celsius:

$$\bar{RH}_i = 0.0195T_a^2 + 0.266T_a + 100.5, \quad (6)$$

which is shown as the dashed line in Fig. 10. It is worth noting that the behavior of the relative humidity in ice clouds is different in comparison to liquid and mixed clouds, where $\bar{RH}_w$ is independent of $T_a$ and it is always close to 100%.

Figure 11 presents a distribution of $RH_i$ in ice clouds in different 5°C temperature intervals. Vertical dashed lines indicate the saturation humidity over water. The obtained $RH_i$ distributions show that $RH_i$ mostly varies between saturation over ice and water. However, at warm temperatures a large fraction of ice clouds (43%) are undersaturated with respect to ice. Figure 12 shows fractions of ice clouds undersaturated with respect to ice with $RH_i < 100\%$, $RH_{in} < 90\%$, and $RH_i < 80\%$.

![Figure 10](image-url) Average relative humidity with respect to ice $RH_i$ vs air temperature measured in ice clouds. Vertical lines indicate standard deviation of $RH_i$ measurements in different clouds. Dashed line is the parameterized humidity in ice clouds [Eq. (6)].
versus air temperature. Over 90% of ice clouds are supersaturated with respect to ice at $-45^\circ < T_a < -40^\circ$C. The fraction of undersaturated ice clouds increases with an increase in temperature. At $-5^\circ < T_a < -10^\circ$C only 57% of ice clouds are supersaturated with respect to ice. The lowest measured humidity in ice clouds was RH$_{i}$ = 20%.

The tendency toward increasing the fraction of undersaturated ice clouds with warmer temperatures may have a rather simple explanation. In stratiform clouds ice particles usually travel from higher levels with a lower $T_a$, to lower levels with warmer temperatures. It would be reasonable to assume that at some point they fall out of a cloud and enter an undersaturated environment with a higher $T_a$. Therefore, the probability of finding ice particles in an undersaturated environment will increase with an increase in temperature.

The sublimation of ice particles in undersaturated ice clouds may significantly affect their habits. It was found that the fraction of the ice particles with modified shapes, due to sublimation, is approximately 40% (Korolev and Isaac 2004). Sublimating results in the removal of facets and the disappearance of corners. The radiation properties of sublimating ice particles would
be different as compared to growing-facetted ice, due to the disappearance of facets and corners. A gallery of images of sublimated particles was presented in Korolev and Isaac (2004).

Figure 13 shows the dependence of RH\textsubscript{i} versus the time of phase relaxation calculated from Eq. (3) for $u_z=0$. Since the shape factor of ice particles ($c$)\textsuperscript{1} is mostly unknown, the calculations of $\tau_p$ ice particles were assumed to be spheres with $c = 1$. In reality, the shape factor is always smaller than unity and varies over the range $0.01 < c < 1$ (Bailey and Hallett 2004). Therefore, the calculated time of phase relaxation should be considered as an upper estimate of $\tau_p$. Since $\tau_p \propto c^{-1}$, the actual $\tau_p$ can be several times higher than that shown in Fig. 13. As follows from Fig. 13, there is no distinct dependence of RH\textsubscript{i} versus $\tau_p$ and RH\textsubscript{i} is rather a function of $T_a$.

8. Discussion

a. Definition of clouds and its relationship to the results

The results of the statistical analysis of the relative humidity in clouds may significantly depend on the definition of a cloud. There are a variety of different definitions of clouds used by different research groups and can depend on visibility, visible or IR satellite radiances, radar or lidar backscatter characteristics, or cloud microphysical parameters (see Mazin et al. 2000). In general for cloud microphysical parameters, the definition of a cloud consists of two parts. The first part of the definition sets one or several thresholds (i.e., the particle concentration, size, LWC, IWC, etc.) to segregate a cloud from a noncloudy environment. The second part of the definition specifies the spatial averaging scale. Spatial averaging is an important part of the definition, since it characterizes the minimum size of the cloudy objects and the scale of the cloud-free zones, which may be considered as a cloud.

In this study the clouds were defined as follows:
1) liquid clouds: $\mu < 0.1$ and LWC $>0.01$ g m$^{-3}$; 
2) mixed clouds: $0.1 < \mu < 0.9$ and LWC $> 0.01$ g m$^{-3}$; 
3) ice clouds: $N_{\text{ice}} > 10$ m$^{-3}$ and LWC $< 0.01$ g m$^{-3}$ (i.e., $dV_{\text{g}}/dt < 10$ mV s$^{-1}$); and 
4) the spatial averaging scale: 100 m.

The definition used in our study in many ways reflects the instrumental limitations of the probes used in this work. As seen from the above definition the thresholds for the condensed water content in liquid and ice clouds are different. The threshold $N_{\text{ice}} = 10$ m$^{-3}$ may correspond to IWC $\sim 0.001$ g m$^{-3}$ or lower. Such ice clouds usually have small $N\tau_p$ and correspondingly large $\tau_p$. Therefore, one may expect that the ice clouds with a low IWC and a large $\tau_p$ would have a low RH\textsubscript{i}. However, Fig. 13 does not show any significant dependence of RH\textsubscript{i} on $\tau_p$ and, consequently, on $N\tau_p$ and IWC. Even clouds with $\tau_p > 5$ h are, on average, supersaturated with respect to ice $\text{RH}_{\text{ice}} > 100\%$ at $T_a < -10^\circ\text{C}$. On the other hand, as shown in Fig. 9, some parts of ice

\textsuperscript{1} The shape factor of an ice particle or the dimensionless capacitance is defined here as $c = C/\pi$, where $C$ is the electrostatic capacitance of the ice particle in the equation for diffusional growth, and $\tau$ is some linear dimension of the ice particle.
clouds with IWC > 0.3 g m^{-3} may be undersaturated with respect to ice (1703:00–1703:30 UTC).

The larger averaging scale would result in inclusion of more cloud holes and out-of-cloud unsaturated air. This would reduce the average relative humidity in ice and mixed clouds. The reduction of the LWC threshold in the definition of mixed clouds is also expected and would lead to the reduction of the relative humidity.

b. Phase inhomogeneity in mixed clouds

Understanding the spatial phase homogeneity in mixed-phase clouds is an important question for radiation transfer calculations (Sun and Shine 1995) and precipitation formation (Rotstayn et al. 2000). There are two extreme situations: 1) liquid droplets and ice particles are uniformly mixed (Fig. 14a) and 2) liquid droplets and ice particles are separated in space and they form single-phase clusters (Fig. 14b). If the liquid and ice phases in mixed clouds are spatially separated, then the second question arises: What is the characteristic scale of the single-phase zones? Is it 1 cm, 1 m, or 1 km?

Recent in situ observations suggest that in stratiform clouds the horizontal characteristic scale of phase inhomogeneity is of the order of 10^3–10^5 m (Korolev et al. 2003; Field et al. 2004). The measurements for that work were made at 100-m spatial resolution, which does not allow direct judgment about inhomogeneity at smaller scales. However, the presence of such zones can be estimated indirectly from RH_{w} measurements. If we assume the existence of single-phase zones at a scale smaller than 100 m, then the measured humidity (RH_{wm}) will be weighted by the lengths of liquid and ice zones

\[
\overline{\text{RH}_{wm}} = (1 - \lambda_{f})\overline{\text{RH}_{wL}} + \lambda_{i}\overline{\text{RH}_{wI}}. \tag{7}
\]

Here \(\lambda_{f} = L_{w}/(L_{w} + L_{i})\) is a spatial fraction of ice single-phase zones; \(L_{w}, L_{i}\) are the measured lengths of single-phase liquid and ice zones, respectively; and \(\overline{\text{RH}_{wL}}, \overline{\text{RH}_{wI}}\) are the average relative humidity with respect to water in liquid and ice zones, respectively. In liquid zones \(\overline{\text{RH}_{wm}} = 100\%\) (section 4). Figure 10 indicates that in ice clouds at \(-20^\circ < T_{a} < -5^\circ\)C the relative humidity with respect to ice \(\overline{\text{RH}_{i}}\) is close to 100\%.

Then, using \(\mu\) as a surrogate for \(\lambda_{i}\), Eq. (7) can be approximated by Eq. (5). In this case, we would expect that in mixed-phase clouds with spatially separated liquid and ice, the relative humidity would follow one of the dashed lines in Fig. 8, depending on the temperature. For example Fig. 8 indicates that at \(-25^\circ\)C and IWC/TWC = 0.9 for the spatially separated liquid, ice \(\overline{\text{RH}_{wm}} \approx 80\%\). The measurements obtained in this study showed that in mixed clouds \(\overline{\text{RH}_{w}} \approx 100\%\), and it does not depend on \(\mu\) and, therefore, does not depend on \(\lambda_{i}\).

This finding suggests that at scales smaller than 100 m, ice and liquid particles in mixed-phase clouds are well mixed, and single-phase zones typically do not exist.

c. Relation to cloud type

Analysis of quasi-steady supersaturation in Korolev and Mazin (2003) suggests that for a variety of integral radii \(N_{p}\), typical for liquid clouds the supersaturation would not exceed a few percent even for a vigorous updraft in cumulonimbus. This implies that in liquid clouds \(\overline{\text{RH}_{w}} \approx 100\%\), and it does not depend on cloud type. In most mixed-phase clouds \(T_{p} < \tau_{pi}\), where \(T_{p}\) and \(\tau_{pi}\) are the times of phase relaxation associated with liquid droplets and ice particles, respectively. This implies that \(\overline{\text{RH}_{w}}\) in mixed-phase clouds will be defined mainly by liquid droplets, whereas ice particles will play a relatively small role (Korolev and Mazin 2003). Therefore, it is expected that, similar to liquid clouds, \(\overline{\text{RH}_{w}}\) in mixed clouds will be close to 100\% and independent of cloud type. In ice clouds \(T_{pi}\) typically varies from minutes to hours, which suggests that the relative humidity in ice clouds will be sensitive to cloud dynamics, radiative processes, rate of ice nucleation, etc. Since these processes are different in different clouds, it would be reasonable to expect that the statistical characteristics of the relative humidity will be related to cloud types. The frequency distribution of \(\overline{\text{RH}_{i}}\) measured by Ovarlez et al. (2002) in cirrus clouds at \(T_{a} < -40^\circ\)C during the Interhemispheric Differences in Cirrus Properties from Anthropogenic Emissions (INCA) project is centered around 100\%. Approximately 30\%–70\% of the studied INCA clouds were undersaturated with respect to ice. These results are quite different from \(\overline{\text{RH}_{w}}\) in glaciated deep frontal clouds in this study (section 6, Figs. 10 and 11). This difference can be explained by (a) mesoscale slow updrafts in frontal clouds

![Fig. 14. Conceptual diagrams of the phase inhomogeneity in mixed clouds: (a) droplets and ice particles are mixed homogeneously; (b) droplets and ice particles are inhomogeneously mixed, and they are forming single-phase clusters.](image-url)
9. Conclusions

Relative humidity was studied from in situ measurements in liquid, mixed, and ice stratiform clouds associated with frontal systems in the temperature range of $-45^\circ C < T_a < -5^\circ C$. The spatial averaging during measurements was approximately 100 m, and the length of sampled in-cloud space was approximately $23 \times 10^3$ km. Liquid clouds were used for calibration of the humidity sensor, with the assumption that liquid clouds are saturated with respect to water and $\text{RH}_{w} = 100\%$. The following conclusions were obtained.

1) In liquid stratiform clouds the average $\text{RH}_{w}$ slowly decreases with an increase of $\tau_a$ (decrease $N_a$) for all temperatures in the range of $-30^\circ C < T_a < -5^\circ C$. The distribution of $\text{RH}_{w}$ in liquid clouds is skewed toward a low $\text{RH}_{w}$, which can be interpreted as a result of the presence of cloud holes with undersaturated air.

2) In frontal stratiform mixed-phase clouds with LWC > 0.01 g m$^{-3}$ $\text{RH}_{w}$ was found to be close to 100% in the temperature range $-35^\circ C < T_a < -5^\circ C$.

3) In ice clouds the relative humidity may be higher or lower than the saturation over ice, but it is always lower than saturation over water. On average, in ice clouds, the relative humidity $\text{RH}_{i}$ increases with decreasing temperature. At $-40^\circ C$ the average $\text{RH}_{i}$ is midway between saturation over ice and liquid.

4) The fraction of ice clouds undersaturated with respect to ice increases with an increase in temperature.

In general the obtained results serve as an experimental closure of theoretical consideration by Korolev and Mazin (2003). The results here differ from the conclusions reached by Fu and Hollars (2004) as a result of a more detailed scheme identifying liquid and ice clouds and a more extensive handling of the corrections in the air temperature and humidity measurements.

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