Ground-based remote sensing of precipitation in the Arctic

C. Zhao\textsuperscript{1} and T. J. Garrett\textsuperscript{1}

Received 26 July 2007; revised 12 February 2008; accepted 11 March 2008; published 17 July 2008.

In this paper, a simple technique is described for application of radar to retrievals of vertical profiles of precipitation in the Arctic. The method is applied specifically to data from the DOE Atmospheric Radiation Measurement (ARM) North Slope of Alaska–Adjacent Arctic Ocean (NSA-AAO) site near Barrow, Alaska. Profiles of precipitation characteristic size are obtained by developing a relationship between particle size and particle fall speed derived from 35 GHz Doppler radar. The relationship depends on an assumed size- and temperature-dependent particle morphology tailored to the Arctic. Error analysis indicates that typical uncertainties in the precipitation retrieval rate are, to 95\% confidence, about 40\% for ice and 15\% for liquid, although they can be as high as a factor of 2. Monthly averaged precipitation retrievals show strong seasonal variation and compare to within respective uncertainties with local ground-based National Weather Service and ARM NSA-AAO surface measurements. Measurements of precipitation profiles show substantial evaporation of precipitation to the atmosphere below cloud base. Associated atmospheric cooling can be several K d\textsuperscript{-1}, suggesting it may be a substantial component of the boundary layer energy budget in the Arctic.


1. Introduction

Arctic climate is especially sensitive to precipitation variability because of its potential to modify the complex feedbacks between surface albedo and temperature, atmospheric temperature and humidity, clouds, and radiation \cite{Walsh,1983,Curry and Ebert,1992,Curry et al.,1996,Moritz et al.,2002}. Precipitation contributes to the surface snow cover and albedo and moderates conductive, sensible, and latent energy transfer across the cryosphere \cite{Beesley,2000}.

Unfortunately, highly accurate ground-based measurements of Arctic precipitation are difficult to obtain because of uncertainties introduced by winds near the instrument, differences in instrument type, difficulties in distinguishing falling snow from blowing snow, and the frequency of trace amounts \cite{Legates and Willmott,1990,Goodison et al.,1994,Yang et al.,1998,Yang et al.,2000}. For example, the annual mean precipitation measured by the National Weather Service (NWS) at the surface is estimated to be about 120 mm but with an uncertainty of up to 50\% \cite{Curtis et al.,1998}.

A second limitation of ground-based instruments is that they provide only precipitation properties near the surface. However to estimate the effects of precipitation evaporation on the atmospheric temperature structure, profiles of precipitation rates are required. While evaporation is generally recognized as being an important modifier of atmospheric dynamics in lower-latitude systems \cite{Srivastava,1987}, it has not been a focus of observational studies in the Arctic.

In this study, we develop a technique for retrieval of precipitation profiles in the Arctic using a ground-based 35 GHz millimeter wavelength cloud radar (MMCR). The method is a simplified extension of techniques used previously for radar retrievals of cloud \cite{Frisch et al.,1995,Wilson et al.,1997,Mace et al.,2002,Matrosov et al.,2002,Deng and Mace,2006}. The precipitation retrieval algorithm description is presented in section 2. The error of the retrieval algorithm is analyzed in section 3. In section 4, multiyear climatologies of precipitation in the Alaskan Arctic are obtained using data near Barrow, Alaska, which are compared to ground-based measurements. A summary of the method and results is presented in section 5.

2. Precipitation Retrieval Algorithm

The precipitation retrieval process described here is based on an assumed temperature-dependent relationship between particle terminal velocity and cross-sectional area equivalent spherical radius. Figure 1 shows the structure of the precipitation retrieval process. First, the precipitation phase and ice crystal habit are determined on the basis of the precipitation temperature, evaluated at the level of the retrieval. Second, an ice crystal shape factor is derived as a function of precipitation particle size. Third, a relationship between precipitation particle size and 5-min averaged Doppler velocity is derived. Fourth, a precipitation “characteristic radius” is obtained from measured Doppler velocity. Fifth, precipitation rate and concentration are calculated.
2.1. Particle Habit

[7] Ice crystal habit depends on the temperature and supersaturation during formation and growth. In this study, we assume that precipitation is in solid form when the temperature at the level of the precipitation (T) is below freezing, and that precipitation is liquid when the temperature is above freezing. In the Arctic, snow often falls from stratus clouds that are composed at least partly of supercooled liquid [Curry et al., 1996; McFarquhar and Cober, 2004]. Therefore, ice precipitation particles in the Arctic normally form under conditions of water saturation.

[8] Table 1 lists four studies describing ice crystal precipitation habits that are likely to be observed in the Arctic. All studies show that plates are the preferred growth habit of ice crystals for the temperature ranges $-4^\circ$C $< T < 0$ and $-22^\circ$C $< T \leq -8^\circ$C. However, while a majority of studies find that columns predominate between $-40^\circ$C $< T < -22^\circ$C [Kumai, 1982; Takahashi et al., 1991; Chen and Lamb, 1994; Nelson and Knight, 1998; Fukuta and Takahashi, 1999; Wood et al., 2001], and this is the phase assumption employed in the retrieval technique described here, Bailey and Hallett [2004] argue instead that plates dominate in this regime. Therefore, as will be described in section 4, the sensitivity of precipitation retrievals to crystal habit assumptions is tested by switching the habit assumption across all temperature ranges.

2.2. Ice Crystal Shape Factor

[9] For an individual precipitation particle, the terminal velocity $V_T$ can be related to a cross-section area-equivalent spherical radius $r$ through [Heymsfield, 1972; Gelbard et al., 1980; Girard and Blanchet, 2001b]

$$V_T = K_1 r^2 / f, \quad K_1 = 1.19 \times 10^6 \text{ cm}^2 \text{s}^{-1}, \quad r < 40 \mu\text{m}$$  \hspace{1cm} (1)

$$V_T = K_2 r / f, \quad K_2 = 8 \times 10^3 \text{ s}^{-1}, \quad 40 \mu\text{m} < r < 0.6 \text{ mm}$$  \hspace{1cm} (2)

where $K_1$ and $K_2$ are empirical parameters for the Stokes and Non-Stokes regimes, respectively. For ice particles, $f$ is a shape factor, defined as the ratio of the drag force ($F_d$) associated with the crystal to the drag force ($F_{ds}$) of a sphere of equivalent cross-section area going at the same speed [Girard and Blanchet, 2001a], such that $f = F_d / F_{ds}$. A derivation of the ice crystal shape factor for columns and plates is described in Appendix A. For liquid particles, the shape factor $f_l$ is equal to unity. Cross-sectional area-equivalent spherical radius is considered here, rather than some alternative such as a volume-equivalent radius, because it most generally encapsulates the physics that distinguishes the fall speeds of rain versus snow.

[10] Figure 2 shows the relationship between terminal velocity and particle size for spherical liquid particles.

### Table 1. Ice Crystal Habits for Different Temperature Ranges

<table>
<thead>
<tr>
<th>Study</th>
<th>Condition</th>
<th>Observation of Crystal Shape</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kajikawa et al. [1980]</td>
<td>Arctic precipitation</td>
<td>$T &gt; -35^\circ$C, plates</td>
</tr>
<tr>
<td>Ohtake et al. [1982]</td>
<td>Arctic precipitation</td>
<td>$T &gt; -22^\circ$C, plates; $T &lt; -22^\circ$C, columns</td>
</tr>
<tr>
<td>Fukuta and Takahashi [1999]</td>
<td>laboratory</td>
<td>$-4^\circ$C $&lt; T &lt; 0$, plates; $-8^\circ$C $&lt; T \leq -4^\circ$C, columns; $-22^\circ$C $&lt; T \leq -8^\circ$C, plates; $T \leq -22^\circ$C, columns</td>
</tr>
<tr>
<td>Bailey and Hallett [2004]</td>
<td>laboratory</td>
<td>$-4^\circ$C $&lt; T &lt; 0$, plates; $-8^\circ$C $&lt; T &lt; -4^\circ$C, columns; $-22^\circ$C $&lt; T \leq -8^\circ$C, plates; $T \leq -22^\circ$C, columns</td>
</tr>
<tr>
<td>This study</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Table 2. Equations for Calculation of Precipitation Characteristic Radius \( r \) From Measurements of Mean Doppler Velocity \( V_{Dm} \)

<table>
<thead>
<tr>
<th>Phase and Habit</th>
<th>( V_{Dm} (\text{m/s}) )</th>
<th>( r (\mu\text{m}) )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Liquid</td>
<td>( V_{Dm} \geq 0.1 )</td>
<td>( 180V_{Dm,0.05} )</td>
</tr>
<tr>
<td>Column ice</td>
<td>( V_{Dm} \geq 0.1 ) and ( V_{Dm} &lt; 0.3 )</td>
<td>( 460V_{Dm,1.08} )</td>
</tr>
<tr>
<td>Column ice</td>
<td>( V_{Dm} \geq 0.3 )</td>
<td>( 360V_{Dm,1.25} )</td>
</tr>
<tr>
<td>Plate ice</td>
<td>( V_{Dm} \geq 0.1 )</td>
<td>( 480V_{Dm} )</td>
</tr>
</tbody>
</table>

Relationships in the Stokes and Non-Stokes regimes can be combined for \( r > 25 \ \mu\text{m} \),

\[
V_T = 6.1r^{1.05}f_f \quad r > 25 \ \mu\text{m}
\]  

where \( r \) and \( V_T \) are in unit of mm and m s\(^{-1} \), respectively.

On the basis of analysis described in Appendix A, the assumed shape factors for column \( f_c \), plate \( f_p \), and liquid \( f_l \) particles are given by

\[
f_c = 3.3r^{0.28} \quad r < 100 \ \mu\text{m}
\]

\[
f_c = 2.3r^{0.12} \quad r \geq 100 \ \mu\text{m}
\]

\[
f_l = 3.4r^{0.25}
\]

\[
f_l = 1
\]

2.3. Terminal Velocity

The mean fall speed of precipitation particles can be inferred from vertically pointing Doppler radar reflectivity-weighted velocity measurements. Doppler velocity \( V_D \) is the sum of the vertical air motion \( V_a \) and the precipitation particle mean fall velocity \( V_T \):

\[
V_D = V_a + V_T
\]

Two methods can be used to account for the contribution from \( V_a \) [Matrosov et al., 2002]. First, the vertical air motion contribution can be estimated from Doppler wind profile data. The second approach is to minimize the contribution of vertical air motions by time-averaging Doppler velocity measurements. Frisch and Shupe [2002] have argued that, in the Arctic, the mean Doppler radar precipitation velocity averaged over several minutes approaches the mean precipitation particle vertical velocity. This is because, over this timescale, updrafts and downdrafts average to near zero. While this approach might not be appropriate when significant convective activity or large wavelength gravity waves are present, it is well suited for the Arctic because the lower atmosphere tends to have high stability. In this study, a 5-min average is used.

2.4. Precipitation Properties

By substituting equations (4), (5), and (6) into equation (3), equations can be derived that relate the particle radius \( r \) to vertical profiles of time-averaged measurements of Doppler velocity \( V_{Dm} \) (Table 2). There are two important limitations to this technique. First, the value of \( r \) derived using this method can only be considered to be some characteristic value that is representative of an ensemble of precipitation particles of arbitrary particle shape. Second, because the precipitation retrieval technique is based on equation (3), which is suitable only for particles with radii greater than 25 \( \mu\text{m} \), the equations are only suitable when \( V_{Dm} > 0.1 \text{ m/s} \).

Retrieval of precipitation particle concentration \( N \) and precipitation rate \( P \) requires derived values for \( r \) and a measured radar reflectivity \( Z \). Provided particles are liquid and satisfy the Rayleigh approximation for 35 GHz cloud radar reflectivity \( (\frac{Z}{N}) < 1 \) [Schneider and Stephens, 1995],

\[
Z = 2^6 \int_0^{\infty} n(r')r'^6dr' \approx 2^6r^6N
\]

where \( r \) is the derived characteristic size representative of an ensemble of hydrometeors with number size distribution, \( n(r') \). Thus, profiles of precipitation \( N \) and \( P \) are given by

\[
N = Z/(2^6r^6)
\]

\[
P = \frac{4}{3}\pi \rho_p V_{Dm} r^3 N/\rho_l
\]

where \( \rho_p \) is the bulk density of the precipitation (solid ice or water) and \( \rho_l \) is the bulk density specific to liquid water. For ice particles, a 6.5 dBZ correction is applied to \( Z \) in equation (8) to account for its different dielectric constant [Lhermitte, 2002]. Corrections for deviations from Rayleigh behavior are not considered because, as will be shown, precipitation particles in the Arctic generally have characteristic size parameters less than unity.

3. Data

The data sets used in this study were collected from the ARM North Slope of Alaska–Adjacent Arctic Ocean (NSA-AAO) site [Stamnes et al., 1999], and the National Weather Service (NWS) at Barrow, Alaska. Table 3 lists the source, instrument, and uncertainty of the data sets. Measurements of Millimeter-wave Cloud Radar (MMCR) reflectivity, Doppler radar velocity, Vaisala Laser Ceilometer reflectivity, and radiosonde temperature profiles are used for the precipitation retrievals. NWS precipitation amount and ARM precipitation rate are used for retrieval evaluation.

The MMCR system provides continuous profiles of radar reflectivity and Doppler velocity, with 95% confi-

Table 3. Source, Instrument, and Uncertainties of Data Sets Used in This Study

<table>
<thead>
<tr>
<th>Data</th>
<th>Source</th>
<th>Instrument</th>
<th>Uncertainty</th>
</tr>
</thead>
<tbody>
<tr>
<td>Radar reflectivity</td>
<td>ARM</td>
<td>MMCR</td>
<td>0.5 dB</td>
</tr>
<tr>
<td>Doppler velocity</td>
<td>ARM</td>
<td>MMCR</td>
<td>0.1 m/s</td>
</tr>
<tr>
<td>Cloud base height 1</td>
<td>ARM</td>
<td>laser ceilometer</td>
<td>7.6 m</td>
</tr>
<tr>
<td>Cloud base height 2</td>
<td>ARM</td>
<td>MMCR</td>
<td>45 m</td>
</tr>
<tr>
<td>Precipitation amount</td>
<td>NWS</td>
<td>rain gauge</td>
<td>4%</td>
</tr>
<tr>
<td>Precipitation rate</td>
<td>ARM</td>
<td>tower</td>
<td>30%</td>
</tr>
<tr>
<td>Temperature</td>
<td>NWS</td>
<td>radiosonde</td>
<td>1 K</td>
</tr>
</tbody>
</table>

*All data are obtained from NSA-AAO. The uncertainty is defined as the range of probable maximum deviation of a measured value from the true value within a 95% confidence interval.
dence uncertainties of 0.5 dB and 0.1 m/s, respectively. The MMCR system provides measurements of cloud boundaries with an accuracy of 45 m [Dong and Mace, 2003]. The Vaisala Laser ceilometer is a ground-based active remote sensing instrument that detects clouds by transmitting infrared pulses vertically into the atmosphere. In this study, the ceilometer is used to obtain cloud base between 15 and 7350 m with a height accuracy of 7.6 m [Dong, 2005].

Temperature profiles are obtained from NWS balloon-borne measured profiles. The height resolution of the temperature profiles is <60 m (6 s), and the time resolution is 12 h. Temperatures at intermediate heights and times are obtained by linear interpolation. The assumed accuracy of the measured temperature profiles is about ±1 K. Precipitation amount is measured by the NWS at Barrow using a tipping bucket rain gauge, which has an accuracy of about 4%. Since November 2003, ground-based measurements of precipitation rate have been obtained at ARM NSA-AAO, with a resolution of 1 min and an uncertainty of 30%.

4. Error Analysis

Precipitation is assumed to be present when a difference is observed between the hydrometeor base height obtained from the ceilometer ($H_{\text{laser}}$) and the lowest return height sensed by the radar ($H_{\text{radar}}$). The MMCR is more highly sensitive to large precipitation particles that fall below cloud, whereas the laser provides a more reliable estimation of cloud base [Dong, 2005]. Therefore, precipitation is assumed to be present when the lowest radar return is below cloud base, such that $H_{\text{radar}} < H_{\text{laser}} - \Delta H_{\text{radar}}$, where $\Delta H_{\text{radar}}$ (90 m) is the radar range gate.

Error estimates for precipitation characteristic radius $r$ and concentration $N$ are not provided. Precipitation is characterized by a size distribution, and radar return will respond to precipitation accordingly. However, in the absence of time-dependent information for precipitation size distributions below Arctic clouds, we do not assume a size distribution, but rather determine a single-valued size and concentration. While ascribing a single value to $r$ can be satisfactory if the size distribution is indeed approximately single-valued, this is unlikely to be generally true. Therefore, $r$ and $N$ should be considered here as “characteristic” values that have zero error but uncertain definition.

On the other hand, precipitation rates are precisely defined. Errors in the retrieval of precipitation properties depend on uncertainties in the assumed ice crystal habit and aspect ratio, combined with uncertainties in the measurements. Assumed ice crystal habit represents a particularly large potential source of error. This is because it dictates the relative magnitudes of the drag force and crystal mass. Because the true ice crystal habit of precipitation is always unknown, and can only be estimated, we make here a simple estimation of the maximum error in the retrievals due to the ice crystal habit assumptions employed.

Figure 3 shows the difference in precipitation property retrievals associated with switching a habit assumption between plates and columns across all temperatures, as a function of Doppler radar velocity ($V_D$). Approximately 80% of the time, the measured mean Doppler radar velocity $V_{Dm}$ at ARM NSA-AAO lies between 0.3 and 0.9 m/s, with a median value of 0.6 m/s. Over this range, the average difference in retrieved $P$ that is associated with switching the habit assumption from exclusively plates to exclusively columns is a factor 0.60 with a standard deviation of 0.10. For the sake of precipitation retrievals, a 50% habit misclassification is assumed for all temperatures, in which case the respective 95% confidence retrieval error for $P$ is about 35%.
retrieval error in arbitrary precipitation are generally higher and 10/C0ðN can be assumed to be 17% for liquids, and 25% EPh = h x < 0.3 m/s and 20% N2 from Doppler are 215¼ are about 17% for liquid precipitation and 43% for ice precipitation from retrievals at the surface and from the NWS at Vs/C24 Z = Estimated Errors in the Precipitation Retrievals at the Es are.

Liquid – 17% 17%
Liquid – 0.5 dB; 0.1 m/s
Plate 50% 0.5 dB; 0.1 m/s
Plate 35% 0.5 dB; 0.1 m/s
Plate 17% 17%

Table 4. Estimated Errors in the Precipitation Retrievals at the 95% Confidence Level

<table>
<thead>
<tr>
<th>Habit (Aspect Ratio)</th>
<th>Z and V_{Dm}</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Uncertainty</td>
<td></td>
</tr>
<tr>
<td>Column</td>
<td>50%</td>
<td>0.5 dB; 0.1 m/s</td>
</tr>
<tr>
<td>Plate</td>
<td>50%</td>
<td>0.5 dB; 0.1 m/s</td>
</tr>
<tr>
<td>Liquid</td>
<td>–</td>
<td>0.5 dB; 0.1 m/s</td>
</tr>
<tr>
<td>Column</td>
<td>35%</td>
<td>25%</td>
</tr>
<tr>
<td>Plate</td>
<td>35%</td>
<td>25%</td>
</tr>
<tr>
<td>Liquid</td>
<td>–</td>
<td>17%</td>
</tr>
</tbody>
</table>

The aspect ratios for columns and plates assumed in the shape factor derivation described in Appendix A are empirical and therefore also associated with retrieval uncertainty. However, the aforementioned uncertainty associated with ice crystal habit can be seen as an extreme example of aspect ratio variability. Plates are merely columns with an inverted aspect ratio. Therefore, to avoid combining errors with covarying uncertainties, only errors from the habit assumption are included.

Uncertainties in MMCR reflectivity and MMCR Doppler velocity are about 0.5 dB and 0.1 m/s with 95% confidence, respectively. Considering that Doppler velocity is averaged over 5 min, and that the measured Doppler velocity predominantly lies between 0.3 and 0.9 m/s, the relative retrieval uncertainties associated with errors in MMCR reflectivity and mean Doppler velocity are 6% and 8%, respectively. Uncertainties in MMCR reflectivity and Doppler radar velocity covary because both are proportional to particle size. Here, the correlation between Doppler radar reflectivity and velocity is assumed to be unity. Thus, the 1-σ retrieval error in arbitrary precipitation property x due to uncertainties of Doppler radar reflectivity and velocity is

\[ \sigma_x = \left[ \sigma_Z^2 \left( \frac{\partial x}{\partial Z} \right)^2 + \sigma_{I_{\text{in}}}^2 \left( \frac{\partial x}{\partial I_{\text{in}}} \right)^2 + 2\sigma_{Z I_{\text{in}}} \left( \frac{\partial x}{\partial Z} \right) \left( \frac{\partial x}{\partial I_{\text{in}}} \right) \right]^{1/2} \]

(11)

where \( \sigma_Z^2 \), \( \sigma_{I_{\text{in}}}^2 \), and \( \sigma_{Z I_{\text{in}}}^2 \) represent the variance in \( x \), \( Z \) and \( I_{\text{in}} \); and \( \sigma_{Z I_{\text{in}}}^2 \) is the covariance of \( Z \) and \( I_{\text{in}} \) (i.e., \( \sigma_{Z I_{\text{in}}}^2 = (Z - \bar{Z})(I_{\text{in}} - \bar{I}_{\text{in}}) \)). Through equation (11), the 95% (2-σ) confidence retrieval errors in \( P \) from Doppler radar reflectivity and velocity uncertainties are, for liquid precipitation, about 17%, and for column ice crystal precipitation, about 25% when \( V_{Dm} < 0.3 \) m/s and 20% when \( V_{Dm} \geq 0.3 \) m/s. For plate ice crystal precipitation, retrieval errors in \( P \) are about 24%. Therefore, the retrieval errors in \( P \) can be assumed to be 17% for liquids, and 25% for both columns and plates.

Table 4 lists the retrieval errors from uncertainties of measurements and assumptions. Uncertainties associated with ice crystal habit and Doppler velocity dominate the total. The combined errors \( E_{\text{comb}} \) are

\[ E_{\text{comb}}^2 = E_{\text{mthd}}^2 + E_{\text{meas}}^2 \]

(12)

where \( E_{\text{mthd}} \) and \( E_{\text{meas}} \) are the retrieval errors associated with the method and measurements, respectively. As Table 4 shows, the respective combined 95% confidence errors in \( P \) are about 17% for liquid precipitation and 43% for ice crystal precipitation.

Additional uncertainty may be associated with the representation of precipitation size by a characteristic radius rather than a size distribution in equation (8). As will be shown below, despite this simplification, the aforementioned quantified uncertainties appear to reflect well any discrepancy between retrievals and observations.

5. Results and Evaluation

5.1. Frequency and Seasonal Cycle

Figure 4 shows retrieved precipitation properties between 2000 and 2004. Retrieved precipitation properties are restricted to a layer between the surface and cloud base, up to 4 km altitude. Precipitation properties at the surface have a wide range of values in both concentration (\( N \)) and characteristic diameter (\( D \), twice the characteristic radius). On average, \( D \) lies between 50 and 1000 \( \mu \)m, and \( N \) between \( 10^{-1} \) and \( 10^3 \) per liter. Occasionally, values of \( N > 100 \text{ cm}^{-2} \) are seen that may be associated with small precipitation particles and upward air motion, in which case the measured fall speeds may be much smaller than true values. Median values of \( D \) and \( N \) are 215 \( \mu \)m and 5 per liter, respectively. Precipitation particle median size at the ground level is about 15 \( \mu \)m larger; and precipitation particle median concentration at ground level is about 20 per liter less. For the 35 GHz radar, systematically large deviations from Rayleigh behavior begin only when sizes exceed about 3000 \( \mu \)m [Kollias et al., 2007], so the functional form of the radar response assumed in equation (8) appears largely justified.

We define a monthly averaged precipitation rate (\( CP \)) and a monthly averaged instantaneous precipitation rate (\( IP \))

\[ CP = \langle P \rangle \]

(13)

\[ IP = \langle P \rangle \quad P > 0 \]

(14)

Figure 5 shows intercomparison of the seasonal cycle in \( CP \) from retrievals at the surface and from the NWS at Barrow, for the period 2000 to 2004. The precipitation rate increases between winter and summer, from a minimum of 0.03 mm/d in March to a maximum of about 1.0 mm/d in July. The difference in mean precipitation rate between the retrievals and the NWS measurements is sometimes as large as 50%. However, this difference lies within the combined uncertainties in the measurements (\( \sim 40\% \)) and retrievals (\( \sim 20\% \) for liquid and \( \sim 40\% \) for ice). The annual mean precipitation from the retrievals and NWS measurements is 147 mm and 143 mm, respectively.

Since October 2003, ARM NSA-AAO has obtained instantaneous precipitation rate measurements. Figure 6 shows an intercomparison of the seasonal cycle in monthly averaged \( IP \) from the retrievals and from ARM measurements in 2004. Both display the same seasonal cycle. However, the retrieved values of \( IP \) are generally higher.
by about 40%, although, again, the discrepancy lies within the respective margins of error.

[31] We note that no account has been made for intermediate shape factors that might be associated with melting precipitation. While this is a simplification, we nonetheless observed good agreement between retrievals and ground-based measurements independent of whether the precipitation temperature was near the melting point: defining a melting layer to be between 272 and 275 K, an origin-regressed least-squares logarithmic fit of retrieved to NWS measured precipitation rates yields a slope of 1.05 ($r^2 = 0.44$). For comparison, the corresponding fit for all data is 1.09 ($r^2 = 0.41$).

5.2. Sample of Precipitation Profiles

[32] Figure 7 shows two examples of profiles of retrieved instantaneous precipitation rates from 6 September and 6 December 2001. Here, two definitions of lapse rate are defined, $\Gamma_1 = \frac{d \ln P}{dz}$ and $\Gamma_2 = \frac{dP}{dz}$. $\Gamma_1$ is about 0.9 km$^{-1}$ on 6 September 2001, and about 0.6 km$^{-1}$ on 6 December 2001; and $\Gamma_2$ is about 1.5 mm d$^{-1}$ km$^{-1}$ on 6 September 2001, and about 1.0 mm d$^{-1}$ km$^{-1}$ on 6 December 2001. As precipitation particles fall, precipitation evaporates and latent heat is absorbed from the atmosphere. In units of K d$^{-1}$, the associated cooling rate in the boundary layer is

$$\frac{\partial T}{\partial t} = -\frac{\rho L}{c_p \rho_a} \Gamma_2 \approx -2.5 \Gamma_2$$

where $L$ is the latent heat of evaporation, and $\rho_a$ and $c_p$ are the density and specific heat of air, respectively. What this implies is that evaporative cooling below cloud base is approximately 4 K d$^{-1}$ in the September case shown and

Figure 4. Probability distribution functions (PDF) for the retrieved precipitation characteristic diameter ($D$) and number concentration ($N$) for the years 2000 through 2004 at NSA-AAO. The average and variance of cloud base height are about 450 m and 480 m, respectively.

Figure 5. Monthly mean precipitation rate ($CP$) at ground level from retrievals (black line) and NWS measurements (gray line) for the years 2000 through 2004. Bars represent the uncertainties (95% confidence or $2\sigma_{error}$) in the retrievals and measurements.

Figure 6. Monthly averaged instantaneous precipitation rates ($IP$) from retrievals (black) and from measurements (gray) at NSA-AAO in 2004. Bars represent the uncertainties (95% confidence or $2\sigma_{error}$) in the retrievals and measurements.
2.5 K d^{-1} in the December case. Obviously, this could play a substantial role in dynamically decoupling the cloud layer from the surface. The results point to a potential climatology for the cooling below Arctic clouds that is associated with latent heat absorption by precipitation.

6. Summary

[33] This paper has described an efficient technique for the retrieval of profiles of precipitation from a ground-based remote sensing site in the Arctic. Measurements from a vertically pointing 35 GHz MMCR radar are used to retrieve profiles and multiyear climatologies of precipitation. The basis of the proposed retrieval technique is a relationship between particle size and Doppler velocity. A shape factor is introduced to represent the effect of precipitation particle habit and aspect ratio on fall speed.

[34] The main uncertainties associated with the retrieval method surround the assumptions made for ice crystal habit and the instrumental accuracy of the Doppler radar velocity measurements. 95% confidence total retrieval errors in precipitation rates are about 40% for ice crystal precipitation and 15% for liquid precipitation.

[35] To within stated uncertainties, the retrievals agree with ground-based precipitation observation over a wide range of conditions. This suggests that, despite various simplifying assumptions made, the retrieval technique is suitable for long-term characterization of precipitation profiles in the Arctic. Future efforts will be directed at applying this technique to the study of atmospheric cooling associated with the evaporation of precipitation.

Appendix A: Derivation of an Ice Crystal Shape Factor

[36] In this appendix, we derive shape factors for ice crystal columns and plates. First, we derive a general solution relating ice crystal shape factor to its drag coefficient. Second, drag coefficients are derived specifically for columns and plates. Third, shape factors for columns and plates are derived as a function of particle characteristic radius.

A1. General Equation

[37] Girard and Blanchet [2001b] introduced a shape factor $f$ to relate the terminal velocity of an ice crystal relative to an ice particle with a mass-equivalent radius. $f$ is defined to be the ratio of the drag force $F_d$ associated with the crystal relative to the drag force $F_{ds}$ of a sphere of equivalent mass falling at the same speed, $V_T$.

$$f = \frac{F_d(V_T, A)}{F_{ds}(V_T, A_s)}$$  \hspace{1cm} (A1)

where $A$ and $A_s$ are the cross-sectional area which the respective particles present to the flow. For simplicity, we define the shape factor as the ratio of $F_d$ associated with the crystal to $F_{ds}$ of a sphere of equivalent cross-section area falling at the same speed, meaning $A = A_r$.

[38] The drag force $F_d$ is related to $V_T$ and $A$ through [Heymsfield, 1972]

$$F_d = 0.5C_d\rho_f V_T^2 A$$  \hspace{1cm} (A2)

where $C_d$ is the drag coefficient, and $\rho_f$ the density of air. Substituting equation (A2) into equation (A1), the shape factor $f$ is thus

$$f = \frac{0.5C_d\rho_f V_T^2 A}{0.5C_{ds}\rho_f V_T^2 A_s} = \frac{C_d}{C_{ds}}$$  \hspace{1cm} (A3)

A2. Shape-Dependent Parameters

A2.1. Derivation of $C_d$ and $C_{ds}$

[39] The drag coefficient of ice crystals is related to their Reynolds number $Re$ through [Podzimek, 1968]

$$C_d = \alpha Re^{-\beta} = \alpha\mu^{-\beta}(V_T H)^{-\beta}$$  \hspace{1cm} (A4)
where $\alpha$ and $\beta$ are shape-specific coefficients, $Re = V_T / H / \mu$, $\mu$ is the kinematic viscosity of air, $H$ is crystal width ($W$) for plate crystals, length ($L$) for column crystals, and diameter ($2r$) for spherical crystals. In this study, it is assumed that $\mu$ is a constant value of $1.5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$.

[40] Letardi et al. [2001] showed that for values of $Re$ between 0.2 and 150, the drag coefficient for spheres is given by

$$C_{ds} = \frac{24}{Re} + 4Re^{-1/3} \quad V_T < 0.4 \text{ m/s}$$

$$C_{ds} = \frac{22}{Re} + 1.1 \quad V_T \geq 0.4 \text{ m/s}$$

However, in order to facilitate calculation of the shape factor for this study, we convert the functional form of $C_{ds}$ into that of $C_d$ in equation (A4). Assuming that the ice crystal particle terminal velocity lies predominantly between 0.3 and 0.9 m/s, and that particle diameters lie between 50 and 1000 $\mu$m (Figure 4), values of $Re$ in the Arctic should be expected to lie primarily in a range between 1 and 60. From Figure A1, for $Re$ between 0.5 and 90, $C_{ds}$ approximately follows the form

$$C_{ds} = 17 Re^{-0.64}$$

(A5)

[41] Wang and Ji [1997] found drag coefficients for columns ($C_{dc}$) and plates ($C_{dp}$) valid for values of $Re$ between 0.2 and 150

$$\log_{10} C_{dc} = 2.44389 - 4.21639a - 0.20098a^2 + 2.32216a^3$$

(A6)

$$C_{dp} = \left( \frac{64}{\pi Re} \right) \left( 1 + 0.078 Re^{0.945} \right)$$

(A7)

where $a = \frac{\log_{10} Re + 1.0}{3.6026}$. Figure A2 shows an approximate but simplified power law form for these equations is given by

$$C_{dc} = 9.4 Re^{-0.48}$$

(A8)

$$C_{dp} = 11.3 Re^{-0.50}$$

(A9)

[42] Substituting equations (A4) and (A5) into equation (A3), an irregular shape factor $f$ is obtained

$$f = \frac{4\alpha V_T^{2-\beta} r_{th}}{\alpha_{th} \mu r_{th}^{\alpha} H^{1-\beta}}$$

(A10)

where $\alpha$ and $\beta$ are parameters for irregularly shaped particles; $\alpha_{th} (17)$ and $\beta_{th} (0.64)$ are respective parameters for spheres; and $\mu$ is the kinematic viscosity of air.
A2.2. Equivalent Radius of Ice Crystals

Here, we derive a relationship between the maximum length in the horizontally oriented ice crystal cross-section \( H \) and the precipitation characteristic radius \( r \). The equivalent cross-sectional area is

\[
A_c = \Gamma_c L W = \pi r^2 \quad (A11)
\]

\[
A_p = \Gamma_p \frac{\pi}{4} W^2 = \pi r^2 \quad (A12)
\]

where \( \Gamma_c \) and \( \Gamma_p \) are area correction factors for columns and plates, and \( L \) and \( W \) are the horizontally oriented ice crystal cross-sectional length and width as shown in Figure A3. The values of \( \Gamma_c \) and \( \Gamma_p \) are

\[
\Gamma_c = 1
\]

\[
\Gamma_p = \frac{3\sqrt{3} / 8}{\pi / 4} = 0.8
\]

The relationship between \( L \) and \( W \) for columns follows a power law relationship

\[
L = c W^d \quad (A13)
\]

For columns, Heymsfield [1972] showed that

\[
L = 2W \quad L \leq 0.3 \text{ mm}
\]

\[
L = 50 W^{2.416} \quad L \geq 0.2 \text{ mm}
\]

which we modify slightly to

\[
L = 2W \quad L < 0.25 \text{ mm}
\]

\[
L = 50 W^{2.416} \quad L \geq 0.25 \text{ mm}
\]

Substituting equation (A13) into equations (A11) and (A12), the cross-sectional maximum length for columns \( H_c \) and plates \( H_p \) follows

\[
H_c = L = 2.5r \quad L < 0.25 \text{ mm}
\]

\[
= 7.1 r^{1.4} \quad L \geq 0.25 \text{ mm}
\]

\[
H_p = W = 2.2r \quad (A17)
\]

On the basis of the relationship between \( L \) and \( r \) for columns (equation (A16)),

\[
H_c = L = 2.5r \quad r < 100 \mu m
\]

\[
= 7.1 r^{1.4} \quad r > 100 \mu m \quad (A18)
\]

Thus, the relationship between \( H_c \) or \( H_p \) and \( r \) satisfies a general form of

\[
H = b r^p \quad (A19)
\]

and equation (A10) for the shape factor is thus

\[
f = \frac{4 \alpha b^{-\beta} V_T^{\frac{1}{\beta}-\beta} r^{\beta-p}}{\alpha_m \mu^{\beta-p}} \quad (A20)
\]

Table A1 lists the parameters for the various components of the shape factor given by equation (A20).

A3. Shape Factors

[45] On the basis of the values of parameters in Table A1, the values of \( f_c \) and \( f_p \) for columns and plates are given by

\[
f_c \approx \begin{cases} 3.01^{0.16} r^{0.16} & r < 100 \mu m \\ 1.81^{0.3} r^{-0.03} & r \geq 100 \mu m \end{cases} \quad (A21)
\]

\[
f_p \approx 3.21^{0.14} r^{0.14} \quad (A22)
\]

where \( r \) and \( V_T \) are in units of mm and m s\(^{-1}\), respectively. Because \( V_T \) also satisfies (Section 2b)

\[
V_T = 6.1 r^{0.05} / f \quad r > 25 \mu m \quad (A23)
\]

\( f_c \) and \( f_p \) are related to \( r \) through

\[
f_c = \begin{cases} 3.3 r^{0.28} & r < 100 \mu m \\ 2.3 r^{0.12} & r \geq 100 \mu m \end{cases} \quad (A24)
\]

\[
f_p = 3.4 r^{0.25} \quad (A25)
\]

where \( r \) is in unit of mm. Figure A4 shows the dependence of \( f_c \) and \( f_p \) on \( r \). Both \( f_c \) and \( f_p \) are greater than 1 for \( r > 20 \mu m \).

<table>
<thead>
<tr>
<th>Habit</th>
<th>( \alpha )</th>
<th>( \beta )</th>
<th>( b )</th>
<th>( p )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Column ( r &lt; 100 \mu m )</td>
<td>9.4</td>
<td>0.48</td>
<td>2.5</td>
<td>1</td>
</tr>
<tr>
<td>Column ( r \geq 100 \mu m )</td>
<td>9.4</td>
<td>0.48</td>
<td>7.1</td>
<td>1.4</td>
</tr>
<tr>
<td>Plate</td>
<td>11.3</td>
<td>0.50</td>
<td>2.2</td>
<td>1</td>
</tr>
<tr>
<td>Sphere</td>
<td>17.0</td>
<td>0.64</td>
<td>2</td>
<td>1</td>
</tr>
</tbody>
</table>
[46] Acknowledgments. This work was supported by the National Science Foundation through grants 0303962 and 0649570 and through NOAA by grant 2308014. The authors appreciate helpful reviewer comments and discussions with Chuntao Liu, Haiyan Jiang, and Jay Mace.

References


Lorentz, R. (2002), Centimeter and Millimeter Wavelengths in Meteorology, 550 pp., Lherrmite, Miami, Fl.


Podzimek, J. (1968), Aerodynamic conditions of ice crystal aggregation, paper presented at International Conference on Cloud Physics, Int. Assoc. of Meteorol. and Atmos. Phys., Toronto, Ont., Canada.


T. J. Garrett and C. Zhao, Meteorology Department, University of Utah, 135 S 1460 E, Room 819, Salt Lake City, UT 84112-0110, USA. (tgarrett@met.utah.edu)