Utilizing Spaceborne Radars to Retrieve Dry Snowfall

MARK S. KULIE AND RALF BENNARTH
Department of Atmospheric and Oceanic Sciences, University of Wisconsin—Madison, Madison, Wisconsin

(Manuscript received 4 February 2009, in final form 15 June 2009)

ABSTRACT

A dataset consisting of one year of CloudSat Cloud Profiling Radar (CPR) near-surface radar reflectivity $Z$ associated with dry snowfall is examined in this study. The CPR observations are converted to snowfall rates $S$ using derived $Z_e-S$ relationships, which were created from backscatter cross sections of various nonspherical and spherical ice particle models. The CPR reflectivity histograms show that the dominant mode of global near-surface dry snowfall has extremely light reflectivity values ($\sim 3-4$ dB$Z_e$), and an estimated 94% of all CPR dry snowfall observations are less than 10 dB$Z_e$. The average conditional global snowfall rate is calculated to be about 0.28 mm h$^{-1}$, but is regionally highly variable as well as strongly sensitive to the ice particle model chosen. Further, ground clutter contamination is found in regions of complex terrain even when a vertical reflectivity continuity threshold is utilized. The potential of future multifrequency spaceborne radars is evaluated using proxy 35–13.6-GHz reflectivities and sensor specifications of the proposed Global Precipitation Measurement dual-frequency precipitation radar (DPR). It is estimated that because of its higher detectability threshold, only about 7%–1% of the near-surface radar reflectivity values and about 17%–4% of the total accumulation associated with global dry snowfall would be detected by a DPR-like instrument, but these results are very sensitive to the chosen ice particle model. These potential detection shortcomings can be partially mitigated by using snowfall-rate distributions derived by the CPR or other similar high-frequency active sensors.

1. Introduction

Snowfall composes a nonnegligible amount of the total precipitation that falls at many mid- and high-latitude locations and obviously has important hydrological and societal impacts. Snow also plays a crucial role in ice sheet dynamics, so knowledge of annual snowfall accumulations is extremely important to areas of the globe that are covered by large expanses of ice (e.g., Greenland, Antarctica, and alpine glacial regions). Additionally, the importance of obtaining robust global snowfall information and monitoring future fluctuations in its spatial distribution, frequency, and intensity is highlighted by recent reports of the accelerated effects of rapid climatic change experienced at higher latitudes (e.g., Krabill et al. 1999; Hinzman et al. 2005; Luckman et al. 2006). Unfortunately, routine surface measurements of snow are scarce in remote regions where snowfall frequently occurs, and these locales are also largely devoid of ground-based remote sensing observations that could provide useful information about frozen precipitation. Therefore, satellite-based microwave remote sensing remains the most viable option to obtain global snowfall information, and increasing attention is being dedicated to retrieve properties of snow via passive, active, and combined microwave observations (e.g., Skofronick-Jackson et al. 2004; Noh et al. 2006; Kim et al. 2007; Liu 2008a; Grecu and Olson 2008). These recent research avenues are especially critical to prepare for the Global Precipitation Measurement (GPM) mission that is scheduled to launch early in the next decade and will include coincident active and passive measurements at higher latitudes.

Relative to passive-only microwave snowfall observations, active spaceborne observations offer the distinct advantage of providing high-resolution information about the vertical structure of precipitation. However, active satellite-based global snowfall observations are extremely limited. The Tropical Rainfall Measuring Mission (TRMM) precipitation radar (PR) was the first active spaceborne microwave instrument and has been
collecting tropical precipitation data since 1997, but its highly inclined orbit, combined with its minimum detectable signal (MDS) of near 18 dB$_\text{z}$$^*$, precludes it from frequently and effectively observing snowfall events. True global active snowfall measurements have been available since the launch of CloudSat (Stephens et al. 2002) and its 94-GHz nadir-looking Cloud Profiling Radar (CPR) in 2006. The CPR's minimum detectable signal is near −30 dB$_\text{z}$$^*$, thus allowing it to observe precipitating structures and most nonprecipitating clouds. Despite its relatively brief existence, CloudSat's CPR has already demonstrated an ability to effectively detect and retrieve snowfall properties, as indicated by the unique census of global snowfall distribution and intensity highlighted by Liu (2008a). Additionally, Matrosov et al. (2008b) and Hudak et al. (2008) both demonstrate the potential utility of CPR data by comparing them to ground-based radar observations of snowfall cases.

Future active snowfall observations will be also available from the GPM dual-frequency precipitation radar (DPR). The dual-frequency capability of the DPR distinguishes it from the single-frequency TRMM PR and should allow GPM to more effectively observe lighter precipitation that commonly occurs at higher latitudes. The dual-frequency radar should also provide additional information about the droplet size distribution of rain and snow.

This study focuses on a few critical aspects of active microwave remote sensing of dry snowfall, and the valuable role of CloudSat data in studying these issues is highlighted throughout the paper. Similar to Liu (2008a), CloudSat CPR data are used to analyze the annual cycle of snowfall from a global perspective. Considerable effort is also devoted to describe interesting regional snowfall differences based on the CloudSat snowfall dataset, as well as some complications of using CPR data that may bias snowfall retrievals in certain locations. Additionally, attempts are made to address future snowfall observations by a dual-frequency radar with similar characteristics as the GPM DPR. Current DPR instrument specifications anticipate that the MDS for the DPR Ku-PR (13.6 GHz) and Ka-PR (35.5 GHz) will be near 17 and 12 dB$_\text{z}$, respectively (Nakamura and Iguchi 2007). While the Ka-PR’s higher frequency and lower MDS will enable it to detect lighter precipitation rates than the Ku-PR and TRMM PR, a thorough assessment of its ability to realistically observe snowfall has yet to be undertaken. In an effort to assess a DPR-like instrument’s near-surface snowfall detection efficacy, CPR data are used to calculate proxy DPR-like radar reflectivities of global snowfall events, and the possible role of CloudSat data in augmenting DPR-like snowfall climatologies is explored. Last, sensitivity tests are performed that highlight uncertainties due to the assumed model employed to represent frozen hydrometeors, and recently published optical property databases of nonspherical ice particle models (Hong 2007; Kim et al. 2007; Liu 2004, 2008b) are utilized to illustrate this potentially significant source of uncertainty.

Section 2 describes the data used in this study, and section 3 provides a methodology overview of the radar reflectivity factor calculation and conversion procedure. Global and regional results, as well as sensitivity studies of assumed frozen particle backscattering properties and data selection criteria, are provided in section 4. Concluding remarks are given in section 5.

2. Data

CloudSat CPR radar reflectivity factor (hereinafter referred to as “radar reflectivity” or simply “reflectivity”) fields from the official CloudSat 2B-Geometric Profile (GEOPROF) product (Mace 2007), combined with temperature data from the CloudSat European Centre for Medium-Range Weather Forecasts (ECMWF)-Auxiliary (AUX) product (Partain 2007) gridded to the CPR’s resolution, were utilized in this study. The CPR's native vertical resolution is near 500 m, but instrument oversampling enables higher-resolution 240-m bins to be included in the 2B-GEOPROF product. For this study, a dataset of potential snowfall events observed by CloudSat between July 2006 and July 2007 was compiled using the following criteria:

1) CPR reflectivity data were restricted to the 30°–75° latitude belt in both hemispheres. While the CPR routinely provides data at locations poleward of 75°, this latitudinal limit was chosen to better represent the planned orbit of the core GPM satellite (the proposed ~65° inclination angle of the GPM core satellite will prohibit the DPR’s swath from reaching the highest latitudes). Liu (2008a) shows that snowfall occurrence is rare equatorward of 30° latitude, thus providing justification for using this value as a lower latitudinal limit of the dataset.

2) Following Liu (2008a), only near-surface CPR data were used in the ensuing analysis, with “near-surface” defined as the sixth data bin above the surface (~1.3 km). CPR reflectivity data from the lowest five bins were automatically rejected to avoid potential clutter contamination from the surface. Section 4 will provide more discussion about whether this constraint is sufficient enough to reject all clutter events, especially in complex terrain.

3) Only CPR data that coincided with 2-m ECMWF-indicated temperatures at or below 0°C were considered. Even though previous studies have indicated
surface snowfall can be readily expected when surface temperatures are as high as 2°C (e.g., Bennartz 2007; Liu 2008a), the more stringent threshold of 0°C was chosen to reduce the occurrence of partially melted near-surface snow in the dataset and ensure that “dry” snowfall was predominantly sampled.

4) The near-surface CPR reflectivity field had to exceed a threshold value of 15 dBZ_e, thus all snowfall-rate statistics derived from this dataset are “conditional” snowfall rates. Additionally, the reflectivity fields had to possess vertical continuity to be considered as potentially precipitating (vertical continuity was defined as reflectivity exceeding the threshold value of 15 dBZ_e in the five data bins immediately above the near-surface data bin). This constraint is added to help eliminate ground clutter contamination. Section 4 includes a short discussion on the sensitivity of the results to varying the degree of vertical continuity required for an observation to be accepted in the snowfall dataset.

5) Any CPR observations with a nonzero 2B-GEOPROF data quality field that indicated a potentially problematic observation were rejected.

6) Similar to Liu (2008a), no attenuation corrections were applied to the reflectivity fields before inclusion into the snowfall dataset. Dry snowfall itself is not highly attenuating at 94 GHz unless the snowfall rates are large and the snowfall layer is thick (e.g., Matrosov 2007a). While the CPR can experience high levels of attenuation under certain meteorological conditions (e.g., heavy liquid precipitation, melting precipitation, high-cloud liquid water contents, and/or large columnar water vapor amounts), it is assumed that the majority of snowfall cases in this study will be largely immune to severe attenuation effects.

The resulting dataset after applying these conditions was populated by over 4.9 million snowfall occurrences, which composed about 3% of the total possible CPR observations within the latitude belts considered and 13% of the total possible CPR observations associated with subzero 2-m ECMWF temperatures in the same latitudinal boundaries.

3. Methodology

Appropriate equivalent radar reflectivity factor (Ze)-snowfall rate (S) relationships were first derived using three different ice particle models. The Ze–S relationships were used for two primary purposes: 1) converting the radar signal (Ze) to a physically useful geophysical parameter (S) and 2) reconstructing the actual CPR snowfall-related reflectivities to proxy 13.6- and 35-GHz reflectivities given a particular snowfall rate.

The Ze–S relationships were derived via the following two-step process:

1) The ice particle size distribution (PSD) for a specified range of snowfall rates was derived via the Field et al. (2005) ice PSD moment conversion scheme.

2) The dependent variable, Ze, was calculated for a given snowfall rate and ice particle model using the PSD derived in the previous step and ice particle backscattering characteristics from various ice habit models. Final Ze–S relationships were derived using a power-law curve-fitting routine.

These steps are described further in the following subsections.

a. Snowfall rate to PSD conversion

The liquid equivalent snowfall rate is given by

\[ S = \rho_w \int_{D_{\min}}^{D_{\max}} m(D) v(D) N(D) dD, \]

where \( \rho_w \) is the liquid water density, \( m(D) \) and \( v(D) \) are the mass and fall speed of a frozen particle (assumed to be unrimed or dry) of maximum dimension \( D \), respectively, and \( N(D)dD \) is the particle concentration within the size bin \( D \) (note that \( D \) is expressed in terms of maximum dimension of the frozen particle, not melted diameter, throughout this study). A liquid equivalent snowfall rate range of 0.01–3.0 mm h\(^{-1}\) is specified a priori for all subsequent calculations. The integral limits \( D_{\min} \) and \( D_{\max} \) in Eq. (1) (and in the following equations that are integrated over the PSD) are set to 0.1 and 5.5 mm, respectively. These values are dictated by the PSD parameterization employed in this study, as well as the database used to obtain backscattering coefficients for calculating radar reflectivities (both of which will be explored in more detail later in this section). The following mass–fall-speed particle-size relations are assumed for calculating snowfall rates:

\[ m(D) = a D^b \]

\[ v(D) = a D^\gamma, \]

where \( a = 0.034 \), \( b = 1.95 \), \( a = 8.83 \), and \( \gamma = 0.36 \) in Eqs. (2) and (3), respectively (SI units of kilograms and meters per second are assumed). The prefactors and exponents in Eqs. (2) and (3) result from best-fit curves derived from previously published studies of unrimed aggregate snowflake properties (Locatelli and Hobbs
Inserting Eqs. (2) and (3) into (1) and then solving for the integrand in Eq. (1) yields the following expression for the $b + \gamma$ moment ($M_{b+\gamma}$) of the PSD:

$$M_{b+\gamma} = \int_{D_{mn}}^{D_{mn}} D^{b+\gamma} N(D) \, dD = \frac{S_{bw}}{a_k},$$  

(4)

Given any arbitrary moment of the PSD, Field et al. (2005) provide a procedure to calculate any other moment of the PSD and to derive the PSD itself. For a given input snowfall rate (expressed in SI units), the “reference” second moment of the PSD ($M_2$) can be derived by inserting the right-hand side of Eq. (4) into the following relationship from the Field et al. (2005) PSD moment conversion scheme and solving for $M_2$:

$$M_n = a_F(n, T_c) M_2^{b_F(n, T_c)},$$  

(5)

where $n$ indicates the moment number of the PSD [e.g., $n = b + \gamma$ from Eq. (4)], $T_c$ (°C) is the temperature, and $a_F$ and $b_F$ are temperature- and moment-dependent parameters outlined in Field et al. (2005) derived from airborne measurements of frozen PDS near the United Kingdom. It should be noted that the Field et al. (2005) results are valid only for particle sizes greater than 0.1 mm, which constrains the lower integral limit in Eqs. (1), (4), and (6) as previously discussed. After $M_2$ has been calculated, the PSD itself can also be derived as outlined by Eqs. (4)–(6) in Field et al. (2005). A similar conversion procedure using the Field et al. (2005) methodology also has been documented by Kim et al. (2007).

b. Calculating $Z_{e-S}$ relationships

The last step to derive frequency-dependent $Z_{e-S}$ relationships is to obtain $Z_e$ for each input snowfall rate using the previously calculated PSDs, where $Z_e$ can be written (mm$^6$ m$^{-5}$) as

$$Z_{e,\lambda} = \frac{\lambda^4}{\pi^2 |K|^2} \int_{D_{mn}}^{D_{mn}} \sigma(D) b_{\lambda} N(D) \, dD,$$  

(6)

where $\lambda$ is the radar wavelength, $|K|^2$ is related to the dielectric constant of water (assumed to be 0.93, 0.88, and 0.75 at 13.6, 35, and 94 GHz, respectively), and $\sigma_{b\lambda}$ is the frequency-dependent backscatter cross section for an individual frozen particle shape and size. Figure 1 depicts $\sigma_{b\lambda}$ at 94 GHz for various shapes based on discrete dipole approximation (DDA) calculations of nonspherical particles (Kim et al. 2007; Hong 2007; Liu 2004, 2008b), as well as frequency-dependent low-density spherical representations of frozen particles following the approach of Surussavadee and Staelin (2006, 2008). The $\sigma_{b\lambda}$ results from these studies were interpolated to a common particle size grid between 0.1 and 5.5 mm. The upper limit of 5.5 mm was chosen to correspond to the Hong (2007) database. It is possible that reflectivity values may be artificially depressed by this upper limit when large snowflakes exceed this size threshold, but calculations reveal this is an issue only at high-liquid equivalent snowfall rates exceeding $\sim 2.5$ mm h$^{-1}$ (not shown). The sometimes large variation of $\sigma_{b\lambda}$ between the different particle shapes is a very obvious feature highlighted in Fig. 1, and the spread is especially accentuated at larger particle sizes where differential non-Rayleigh scattering effects of the particles causes the $\sigma_{b\lambda}$ values to differ by many orders of magnitude. The resonant Mie scattering features of the spherical particles are also noteworthy when compared to the nonspherical, DDA-based results. Note that the single-scattering values from Fig. 1 for three different shapes are inserted directly into Eq. (6) and then averaged over the PSD to obtain $Z_e$.

To assess the sensitivity of the results to the assumed frozen particle model, $Z_{e-S}$ relationships for three different particle shapes at 94.0, 35.0, and 13.6 GHz were derived (Table 1). It should be emphasized that these idealized ice particle models best represent unrimed—or perhaps lightly rimed—frozen hydrometeors, thus providing further motivation for limiting this study to dry snowfall. The Liu (2008b) three-bullet rosette (denoted as LR3), Hong (2007) aggregate (HA), and the Surussavadee and Staelin (2006, 2008) low-density spherical snow particles (SS) were chosen as three representative ice particle models. The LR3 backscattering values generally fall within the middle of the $\sigma_{b\lambda}$ envelope at 94 GHz (Fig. 1), especially at larger particle sizes that contribute most to the calculated reflectivity, and will be used throughout this study as the representative “average” ice particle model. The HA and SS shapes show stronger and weaker backscattering characteristics at 94 GHz, respectively, than LR3 and are utilized to show the sensitivity of the results to other ice particle models. Other particles, such as the Hong (2007) droxtal and Liu (2008b) column and plate shapes display much higher $\sigma_{b\lambda}$ values than the HA shape, but were not chosen since these habits realistically occur only at much smaller particle sizes than what typically contribute most significantly to the total radar reflectivity.

The $Z_{e-S}$ relationships are further illustrated in Fig. 2. Figure 2a shows the 94-GHz $Z_{e-S}$ relationships as both a scatterplot and a best-fit power-law line that corresponds to the results shown in Table 1. The relative error of the best-fit lines is small, with only small differences
observed at most $Z_e$–$S$ pairings. As expected from the $\sigma_b$ results in Fig. 1, the LR3 $Z_e$–$S$ relationship falls between the HA and SS shapes at 94 GHz, but the resulting $Z_e$ value for a given snowfall rate is extremely sensitive to the ice particle model. For instance, an assumed snowfall rate of 0.1 mm h$^{-1}$ produces corresponding reflectivity values of about 0.14, 0.52, and 1.7 mm$^6$ m$^{-3}$ for the SS, LR3, and HA ice models, respectively—a range exceeding 10 dB. At a snowfall rate of 1.0 mm h$^{-1}$, the potential range of reflectivity values increases to about 14 dB. For this study, it is probably more meaningful to investigate the differences in retrieved values of $S$ given a $Z_e$ observation. As shown in Fig. 2b, calculated snowfall rates for the SS, LR3, and HA shapes are about 0.76, 0.22, and 0.10 mm h$^{-1}$ for a $Z_e$ value of 1.6 mm$^6$ m$^{-3}$ ($\approx 2$ dB$Z_e$), and 3.55, 0.82, and 0.32 mm h$^{-1}$ when $Z_e$ is 10.0 mm$^6$ m$^{-3}$ (10 dB$Z_e$). The backscattering properties of the chosen ice model clearly represent a large potential source of uncertainty in retrieving snowfall rates from radar data.

Complicating matters further, note that the $Z_e$–$S$ relationships for each ice particle change relative to one
another as frequency decreases (Figs. 2c,d). At 35 GHz, the LR3 $Z_r-S$ curve converges with the SS results at higher snowfall rates. At 13.6 GHz, the LR3 $Z_r-S$ relationship produces similar results to the HA shape at low snowfall rates and the SS shape at high snowfall rates. [It should be noted that no 13.6-GHz DDA results were available from Hong (2007), so the HA shape’s $\sigma_b$ results at this frequency were scaled directly from the SS 35-GHz $\sigma_b$ results.] Overall, these results highlight the complex shape-dependent transition of PSD-averaged backscattering behavior from higher to lower frequencies resulting from each shape’s distinctive deviation from Rayleigh scattering behavior, especially at higher frequencies.

Table 1 also indicates three other recently published $Z_r-S$ relationships for the frequencies of interest (Noh et al. 2006; Matrosov 2007a; Liu 2008a). There are noticeable differences between the current results with these recent studies at all frequencies. The choice of ice particle model explains some of these differences. The Noh et al. (2006) and Liu (2008a) studies both use DDA results from Liu (2004, 2008b), similar to this study. However, Noh et al. (2006) considers dendrite and sector snowflakes exclusively, while Liu (2008a) uses a best fit of all Liu (2004, 2008b) rosette shapes, plus sector and dendrite snowflakes, to derive the $Z_r-S$ relationship. The Matrosov (2007a) $Z_r-S$ relationships are not derived using the DDA approach, but instead aggregate snowflakes are modeled as spheroids and combined with the T-matrix method to calculate backscattering coefficients. Differences between the current study and these related works can also be partially attributed to the PSD utilized, as Noh et al. (2006), Matrosov (2007a), and Liu (2008a) all employ exponential PSDs that differ from the Field et al. (2005) PSD used in this study. Matrosov (2007a) also notes considerable variability in $Z_r-S$ relationships when the assumed aspect ratio of the model spheroid is changed, or when the mass–fall-speed relationships are altered to mimic the variability in observed aggregate masses and fall speeds. The overall variability in $\sigma_b$ and $Z_r-S$ relations illustrated in Figs. 1 and 2, however, combined with possible PSD effects, appears to dominate these other sources of potential uncertainty.

<table>
<thead>
<tr>
<th>Ice habit (or reference)</th>
<th>$Z_r$ (94 GHz)</th>
<th>$Z_r$ (35 GHz)</th>
<th>$Z_r$ (13.6 GHz)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LR3</td>
<td>13.16$^{+4.40}_{-0.85}$</td>
<td>24.04$^{+5.51}_{-1.20}$</td>
<td>34.63$^{+5.56}_{-1.20}$</td>
</tr>
<tr>
<td>HA</td>
<td>56.43$^{+5.52}_{-4.85}$</td>
<td>313.29$^{+5.85}_{-4.85}$</td>
<td>163.51$^{+5.98}_{-4.85}$</td>
</tr>
<tr>
<td>SS</td>
<td>2.19$^{+0.20}_{-0.20}$</td>
<td>19.66$^{+0.74}_{-0.74}$</td>
<td>36.10$^{+0.97}_{-0.74}$</td>
</tr>
<tr>
<td>Liu (2008a)</td>
<td>11.50$^{+0.25}_{-0.25}$</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>Matrosov (2007a)</td>
<td>10.00$^{+0.80}_{-0.80}$</td>
<td>56.00$^{+0.20}_{-0.20}$</td>
<td>—</td>
</tr>
<tr>
<td>Noh et al. (2006)</td>
<td>—</td>
<td>88.97$^{+0.64}_{-0.64}$</td>
<td>250.00$^{+0.08}_{-0.08}$</td>
</tr>
</tbody>
</table>

The derived 94-GHz $Z_r-S$ relationships for the LR3 shape shown in Table 1 were used to convert CloudSat CPR near-surface radar reflectivity observations to snowfall rates. Once the snowfall rate was known for each CPR observation, corresponding proxy DPR-like reflectivities were calculated using the derived 35.0- and 13.6-GHz $Z_r-S$ relationships. It should be acknowledged that the DPR will be a scanning radar capable of producing a much larger data swath than CloudSat’s CPR, so the proxy DPR-like database of snowfall events presented here assumes all statistics and comparisons are made using only the nadir viewing angle.

4. Results

4.1 Global results

The CPR 94-GHz near-surface reflectivity distribution in 1-dBZ$_r$ data bins for global dry snowfall events located between the 30° and 75° latitudinal belts is shown in Fig. 3a. The CPR distribution peaks near the 3- and 4-dBZ$_r$ bins, while an estimated 94% of all CPR near-surface radar observations in the snowfall dataset were less than 10 dBZ$_r$. It should be noted that this percentage also assumes a lower threshold of $-10$ dBZ$_r$ that corresponds to a snowfall rate of about 0.03 mm h$^{-1}$ (assuming the LR3 $Z_r-S$ relationship), since radar reflectivities lower than about $-10$ dBZ$_r$ do not contribute significantly to the total snowfall accumulation (see Fig. 3b). These results highlight the prevalence of very light radar reflectivities associated with dry snowfall and hint at potential detection difficulties using active remote sensing instruments with a higher MDS than the CPR. To better assess the detection efficacy of global snowfall by a DPR-like instrument, the 35- and 13.6-GHz reflectivity distributions that were calculated using the LR3 $Z_r-S$ relationships in Table 1 are also indicated in Fig. 3a. The histograms for these two lower frequencies peak at the higher-reflectivity bins of near 6 and 7 dBZ$_r$, respectively, because of the differing $Z_r-S$ relationships shown in Table 1. Even though the peak reflectivities are shifted higher, the ability of a 35–13.6-GHz radar to observe snowfall may be limited by their MDS. The portion of the reflectivity distribution that can be captured by a 35-GHz instrument with an MDS of 12 dBZ$_r$ is indicated in Fig. 3a and clearly shows the potential undercount of global snowfall events by a 35-GHz radar, and only about 7.1% of the reflectivity values associated...
with near-surface snowfall would be detected (Table 2). A 13.6-GHz instrument with an MDS of 17 dBZ is even less sensitive to snowfall and would potentially capture only about 1.2% of the near-surface snowfall cases (Table 2). Similar to the CPR 94-GHz results presented earlier, these percentages are calculated assuming measurable snowfall results from a near-surface reflectivity exceeding −9 and −8 dBZ at 35 and 13.6 GHz (which corresponds to a snowfall rate of 0.03 mm h⁻¹ assuming the LR3 Zₑ-S relationship), respectively.

An alternative, and arguably more physically significant, perspective of global snowfall is shown in Fig. 3b, which highlights both the distribution of conditional snowfall rates and the cumulative density function of the average conditional snowfall rate (which serves as a proxy for snowfall accumulation) that reveals how much each snow-rate bin contributes to the conditional global snowfall rate. The snowfall-rate bins are derived directly from the CPR reflectivity observation bins shown in Fig. 3a using the LR3 94-GHz Zₑ-S relationship in Table 1. According to Fig. 3b, the average conditional global snowfall rate is near 0.3 mm h⁻¹. Assuming an MDS of 17 and 12 dBZ, 35- and 13.6-GHz instruments will only be sensitive to snowfall rates greater than about 0.76 and

![Graphs showing radar reflectivity factor (Zₑ) as a function of snowfall rate (S) at 94, 35, and 13.6 GHz for LR3, HA, and SS ice particle models.](http://journals.ametsoc.org/jamc/article-pdf/48/12/2564/3548119/2009jamc2193_1.pdf)
1.27 mm h\(^{-1}\), respectively, which translates to about 80% and 94% of the total global snowfall accumulation being missed by a DPR-like radar based on the cumulative distribution curve shown in Fig. 3b.

b. Sensitivity to particle type

The sensitivity of the global radar reflectivity and snowfall-rate distributions to assumed particle type is shown in Fig. 4. At 35 GHz, noticeable shifts in the radar reflectivity histograms to higher values are evident if the HA and SS \( Z_e \)-\( S \) relationships are used instead of the LR3 shape. The HA histogram peaks near 8 dB\( Z_e \), while the SS histogram peaks near 12 dB\( Z_e \) and substantially broadens. From a radar reflectivity detection standpoint, both of the scattering characteristics of these shapes would allow a 35-GHz active radar to observe more snowfall compared to the LR3 histograms shown in Fig. 3, and the detection rate increases substantially to about 18% and 41% for the HA and SS shapes, respectively. At 13.6 GHz, the SS distribution again shifts to higher reflectivity values, peaking near 16 dB\( Z_e \), and increases the potential detection rate significantly to about 33% (versus only about 1.2% for the LR3 shape). The HA shape, however, peaks lower than the LR3 distribution at 13.6 GHz and would have a reduced detection efficacy of under 1%.

The conditional snowfall-rate histograms and cumulative plots shown in Figs. 4c and 4d enable both an assessment of the physical reality of each particle shape and a crude error estimate from an average global snowfall-rate perspective. The snowfall-rate distributions for each shape are clearly different, as the LR3 snowfall-rate

<table>
<thead>
<tr>
<th>Table 2. The first two columns are percentages of near-surface proxy DPR-like radar reflectivities at 35 (13.6) GHz greater than or equal to the proposed DPR minimum detectable signal of 12 (17) dB( Z_e ), calculated from CPR observations. The last two columns are the percentage of total snowfall accumulation that would be detected at 35 (13.6) GHz for the same snowfall dataset. Boldface numbers indicate percentages after quality-control measures were applied to clutter-contaminated data points.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Percentage Percentage of total</td>
</tr>
<tr>
<td>dB( Z_{35} ) &gt; 12 dB( Z_{13.6} ) &gt; 17</td>
</tr>
<tr>
<td>-----------------------------------------------</td>
</tr>
<tr>
<td>Global</td>
</tr>
<tr>
<td>Greenland (land)</td>
</tr>
<tr>
<td>Greenland (ocean)</td>
</tr>
<tr>
<td>Antarctica</td>
</tr>
<tr>
<td>Russia</td>
</tr>
</tbody>
</table>
distribution is bounded by the HA snowfall rates at the lower end and SS at the higher end (Fig. 4c). According to Fig. 4d, the average global conditional snowfall rate is near 0.28 mm h\(^{-1}\) based on the LR3 shape, but large deviations from this value of 60% (HA) to 300% (SS) are possible depending on what particle shape is chosen. The potential detection rate of accumulated snowfall for a DPR-like radar is also very sensitive to the particle shape and increases markedly for the HA shape at 35 GHz and the SS shape at both frequencies.

A natural question to pose is what particle model produces the most physically realistic results, and also is our assumption of the LR3 shape as a representative average value valid? Liu (2008a) showed reasonable comparisons between calculated yearly accumulation statistics averaged over a large expanse of Canadian and historical ground-based observations. From this comparison, the LR3 shape used in this study probably produces the most plausible results out of the three shapes considered, as the SS average conditional snowfall results seem skewed much too high and the HA too low, but more comparisons with other independent measurements are needed to further address this issue. Overall, these sensitivity tests illustrate the large potential uncertainty in snowfall retrievals based exclusively on the choice of ice particle model.

c. Sensitivity to vertical continuity threshold

Recall from section 2 that a vertical continuity threshold is applied in the screening process to select CPR
snowfall observations since precipitation-producing clouds are likely to possess some degree of vertical development. Also, ground clutter tends to display limited vertical structure, so a secondary motive is to introduce further safeguards to reduce contamination by ground returns.

Figure 5 illustrates the sensitivity of the global CPR histograms to the assumed vertical reflectivity structure threshold. As shown in Fig. 5a, the low end of the distribution is drastically altered if this threshold is reduced to ~0.5 km or completely eliminated, and the resulting histograms are skewed to much lower reflectivity values that peak near the −1- and −10-dBZ<sub>e</sub> reflectivity bins, respectively. These changes to the reflectivity distributions are also reflected in the total number of snowfall counts (Fig. 5c) that increase tremendously as the vertical threshold is relaxed to ~0.5 km (increases ~44%) or removed (increases ~80%). The addition of such a large quantity of lower-reflectivity observations depresses the conditional average snowfall rate shown in Fig. 5d by about 21% (0.5-km threshold) and 36% (no threshold) compared to the default threshold. The upper end of the reflectivity distribution is also sensitive to the vertical continuity threshold, although it
is not readily evident in Fig. 5a. If the ordinate in Fig. 5a is changed to a logarithmic scale, noticeable secondary maxima near 25 dBZ_e appear in each of the distributions (Fig. 5b). As will be discussed in greater detail in the next section, these secondary maxima are likely due to ground clutter. Recent studies have also indicated that 94-GHz precipitation-related reflectivities cannot exceed a threshold of about 25 dBZ_e because of the potential dampening effects of resonant Mie scattering (e.g., Matrosov 2007b; Hudak et al. 2008), thus lending further support that the secondary CPR observational maxima near 25 dBZ_e in Fig. 5b are not due to precipitation. While the frequency of occurrence of these potential clutter-contaminated entries in the snowfall dataset is substantially less than the counts tallied at the lower end of the reflectivity distribution when the vertical continuity threshold is lowered, they can significantly influence the retrieved average conditional snowfall rates, especially on a regional basis (see section 4d for more details). The effectiveness of the vertical reflectivity threshold in reducing clutter contamination is clearly illustrated in Fig. 5b. The secondary frequency of occurrence maxima with no vertical threshold is lowered by over an order of magnitude when the 1-km vertical threshold is utilized. Furthermore, the reduction is not as substantial for the 0.5-km threshold case, thus justifying the more stringent vertical continuity threshold to help mitigate clutter contamination. Note, however, that the 1-km vertical continuity threshold does not completely remove the occurrence of the elevated reflectivities above 20 dBZ_e so further quality-control measures must be taken to further reduce their effect on retrieved snowfall rates (see section 4d).

The detection statistics for a DPR-like instrument shown in Table 1 are also affected by the assumed reflectivity depth constraint, as the percentage of near-surface 35 (13.6)-GHz reflectivities exceeding 12 (17) dBZ_e would decrease from 7.1% (1.2%) to 5.0% (1.0%) due to the increased reflectivity counts at the lowest part of the reflectivity distributions shown in Fig. 5. Interestingly, the percentage of total snowfall accumulation detected would increase by 5%–10% if no vertical threshold is applied, which seems counter-intuitive if there are substantially lighter snowfall rates contributing to the total global snowfall accumulation. This paradox, however, is again related to the secondary maximum in the reflectivity distribution exceeding about 20 dBZ_e associated with ground clutter that artificially inflates the results. If a quality-control step is introduced to partially correct some of these inflated values, the snowfall detection percentages fall substantially when the vertical continuity thresholds are removed.

These sensitivity tests raise a few important points. First, there is a distinct possibility the vertical continuity constraint is too restrictive in this study, and some snowfall counts are being rejected. But as shown in Fig. 5, most of these rejected snowfall occurrences reside at the light end of the reflectivity–snowfall-rate spectrum. However, the sheer number of these rejected cases could possibly contribute a substantial, or at least non-negligible, amount to the total global snowfall accumulation. If this is the case, the global near-surface detection values for a DPR-like instrument shown in Table 2 would decrease, and the results shown in this study would be conservatively located at the high end of the potential DPR-like detection rate. Additionally, since the near-surface reflectivity is defined as ~1.3 km above the surface in this study, and removing any vertical reflectivity structure threshold significantly increases the number of near-surface snowfall counts, there may be a substantial mode of global shallow snowfall. This topic bears further study since no definitive conclusions can be drawn from this analysis. But great care must be taken when analyzing the CPR data without any vertical reflectivity continuity thresholds, as significant ground clutter contamination exists in such a dataset without applying additional safeguards.

d. Regional results

A regional perspective of snowfall detected by the CPR and the corresponding proxy DPR-like reflectivity and snowfall-rate distributions is indicated in Fig. 6. These remote regions were chosen because CloudSat observations—and future missions like GPM—provide extremely valuable information in regions previously devoid of routine active remote sensing observations of snowfall. While passive microwave observations of these regions are comparatively plentiful, snowfall retrieval—especially light snowfall over continental regions—by passive microwave methods is extremely difficult (e.g., Kongoli et al. 2003; Skofronick-Jackson et al. 2004; Kim et al. 2008). The regions illustrated in Fig. 6 also display a relatively high frequency of snowfall occurrence (Liu 2008a), thus providing further motivation for selecting these specific locations. Since the CPR dataset utilized in this study only extends to 75°N–S, “Greenland” is assumed to be all land regions on Greenland located south of 75°N, while the “Greenland ocean” region includes all over-ocean observations near Greenland (bounded by 58°–75°N and 62°–18°W). “Antarctica” describes all CPR observations north of 75°S (i.e., only the northern periphery of continental Antarctica). The north-central Russia region is bounded by 58°–75°N and 75°–100°E and includes exclusively continental observations.
FIG. 6. As in Fig. 2, but histograms are derived on a regional, not global, basis. Also, the average conditional snowfall-rate thin dotted line shown for the Greenland, Greenland ocean, and Antarctica regions represents the quality-controlled (QC) cumulative distribution function that alters the reflectivity pixels potentially associated with ground clutter in topographically complex regions.
Figure 6 highlights the inherent variability in the radar reflectivity distributions between the selected regions. The reflectivity histograms of the land regions are dominated by very light CPR reflectivities over Greenland, north-central Russia, and Antarctica, while snowfall over the ocean environs surrounding Greenland exhibits higher intensities than the adjacent land regions, similar to the findings of Liu (2008a). Table 2 and Fig. 6 combine to show the tremendous variability in the percentage of reflectivity values exceeding the assumed MDS at 35 and 13.6 GHz, as well as the percentage of the average snowfall rate that would be captured by a DPR-like instrument, for each region. The following subsections highlight some interesting details that emerge from analyzing these data, with special attention paid to the Greenland dataset.

1) Greenland

The 94-GHz CPR-derived reflectivity histogram for Greenland peaks between −1 and 1 dBZₑ (Fig. 6a), so very light reflectivities dominate this regional distribution. The converted 35-(13.6-)GHz reflectivities are greater than or equal to 12 (17) dBZₑ at about an 8% (3%) rate, which is slightly higher than the global results (Table 2). The Greenland data display an interesting subtle secondary reflectivity maximum between 20 and 30 dBZₑ in Fig. 6a. This feature is also evident in the conditional snowfall-rate histogram and appears more prominently in the average conditional snowfall rate that almost doubles its value due to snowfall rates exceeding about 10 mm h⁻¹ (Fig. 6b). These extremely intense snowfall rates produce impressive detection rates from a DPR-like instrument of about 48% (39%) of the total snowfall accumulation that are abnormally large relative to the global values in Table 2.

The CPR data were closely inspected to find the cause of these high-reflectivity values. Greenland can receive very large snowfalls, especially on its southeastern edge (Hanna et al. 2006), but the snowfall rates contributing to the large average snowfall rate value in Fig. 6b seem physically improbable. Therefore, the entire CPR snowfall dataset was searched for reflectivity values exceeding 25 dBZₑ. Almost 1400 instances of elevated reflectivities were found that represented only about 0.03% of the entire dataset, and 60% of these aberrant values occurred over Greenland. A handful of other regions like the Andes Mountains, the Canadian Rocky Mountains, the Himalaya Mountains, and some parts of Antarctica—all regions of complex terrain with potential snow- and ice-covered surfaces—preferentially contained these anomalous reflectivity values as well.

Figure 7 illustrates actual CPR data (sixth bin above the surface and higher) for three cases over Greenland. The black line indicates the surface height taken directly from the 2B-GEOPROF product. These cases all contain very light reflectivities aloft (generally under 10 dBZₑ), although Fig. 7b does contain some reflectivities exceeding 15 dBZₑ that are indicative of heavier snowfall (e.g., near 3-km altitude at −60.3°N). However, note the numerous exceedingly high near-surface reflectivities approaching 30 dBZₑ in numerous locations that are obviously unphysical. Also note that Figs. 7a and 7c are almost the exact same overpass location on different dates, and the same signatures appear in both plots in regions of complex terrain. Interestingly, some of the signatures in Fig. 7b do not appear to coincide with highly structured terrain, but these are areas where the elevation database may not be trustworthy in remote, complex topographic regions, and official CloudSat data literature does indeed warn of potential errors in the elevation database (Li et al. 2007). It should be noted that many of these potentially contaminated CPR observations are embedded within legitimate snowfall with significant vertical structure, thus making it very difficult to detect without using more complex vertical reflectivity gradient tests to identify ground clutter signatures. Melting snow could also possibly cause elevated reflectivities, but, as indicated in Fig. 7, this does not appear to systematically occur over Greenland or the other regions mentioned.

In an attempt to mitigate the clutter contamination problem, all near-surface CPR observations exceeding 20 dBZₑ were replaced with data from the eighth data bin above the actual surface instead of the sixth bin. As shown in Fig. 6b, this rudimentary quality-control method completely removed the large increase in average snowfall rate above the 10 mm h⁻¹ snow-rate bin. Table 2 also indicates corrected values in boldface print for the various percentages with most of the clutter eliminated. The total accumulation percentages detected by a DPR-like instrument decrease substantially to about 22% (7%) for the 35-(13.6-)GHz frequencies. Also note that the global total accumulation detection percentage values in Table 2 were affected by the small amount of clutter-contaminated near-surface reflectivity data points, but the reflectivity percentages were not significantly altered.

2) Greenland Ocean

As previously mentioned, the Greenland ocean environment produces consistently higher radar reflectivities associated with snowfall, as the peak CPR 94-GHz reflectivity bin is located near 5 dBZₑ. The Greenland oceanic region contains the highest percentage of 35-GHz radar reflectivities exceeding 12 dBZₑ (11.8%) and total snowfall accumulation detected by 35- and 13.6-GHz
frequency measurements (27.3% and 8.8%, respectively). The conditional average snowfall rate of near 0.34 mm h$^{-1}$ is about 30% higher than the related value over Greenland. Note that this region displays some of the highest snowfall frequency occurrence values and the most intense snowfall on the entire globe (Liu 2008a). The detection statistics of a DPR-like instrument for the Greenland oceanic region can therefore be considered a best-case regional scenario.

Surprisingly, ground clutter affected the Greenland ocean dataset, as indicated by the secondary increase in average conditional snowfall rate in Fig. 6d that is similar to continental Greenland. Two clusters of data points were discovered over near-coastal Greenland regions.

FIG. 7. CloudSat CPR radar reflectivity observations (dBZ) and ECMWF temperature (K) of three snowfall events over Greenland. The sixth reflectivity bin above the surface (“near-surface” reflectivity) and above are shown to correspond with the actual dataset used in this study. The land surface, derived from a digital elevation map database used in the official CloudSat products, is also indicated by the black line: (a) 2 Dec 2006, (b) 2 Sep 2006, and (c) 25 Apr 2007.
that were apparently misclassified as ocean instead of land and were the source of the excessive clutter contamination. These clusters appear to be located within a few elevated conditional mean reflectivity pixels in the Liu (2008a) study and may have artificially inflated their results as well.

3) Antarctica and North-Central Russia

The conditional average snowfall rates retrieved over these continental regions are very similar and much lighter than the Greenland land and ocean regions (Figs. 5f,h). Note, however, that the reflectivity and snowfall-rate histograms are very different between these two locations. North-central Russia’s histogram peaks near the 1-dB$Z_e$ data bin, while Antarctica’s is much lower at about −3 dB$Z_e$ (Figs. 5e,g). Antarctica’s 94-GHz reflectivity distribution is much broader than that of north-central Russia though, which has major implications for how a DPR-like instrument would observe the snowfall over each area. A DPR-like radar would have difficulty retrieving much snowfall over north-central Russia, as its near-surface reflectivity detection efficacy is about 2.4% (0.2%) for 35 (13.6) GHz (Table 2), while Antarctica is substantially improved at 35 GHz (4.8%) and slightly better at 13.6 GHz (0.9%) because of its broad reflectivity distribution shape. The percentage of total accumulation that could be observed by a DPR-like instrument is 16.2% (4.3%) for Antarctica, which is very near the global averages, while north-central Russia is much lower [5.8% (0.5%)]. Note that north-central Russia is also not affected by ground clutter contamination, but Antarctica is susceptible to it. The north-central Russia and Antarctica comparison highlights the importance of knowing regional differences in the snowfall-rate distributions, as one might conclude that these regions would be similarly sampled by an active spaceborne radar based on their comparable conditional average snowfall rates.

5. Conclusions

With the advent of CloudSat, global radar observations of snowfall are for the first time possible. Such observations arrive at an especially crucial time, as pressing scientific issues related to the rapid and dramatic effects of climatic change at higher latitudes make sustained monitoring of global snowfall extremely important in the coming years. The main goals of this study were to highlight the utility of global CloudSat snowfall observations, illustrate interesting regional differences in the reflectivity and retrieved snowfall-rate distributions, provide necessary guidance related to how future spaceborne radars may observe snowfall on a global and regional basis, and address some of the uncertainties associated with active, spaceborne snowfall retrievals.

Properly characterizing the scattering properties of snowfall remains one of the largest sources of uncertainty related to snowfall retrieval. In the last few years, various authors have developed databases of optical properties of nonspherical precipitation-sized ice particles. One goal of the present study was to attempt to summarize these efforts and address the question of how different retrieved snowfall rates and accumulations depend on the chosen ice microphysical model. It is shown that the annually and globally averaged conditional dry snowfall rate varies significantly depending on what ice scattering model is used. Some of the more unlikely ice particle shapes with extreme backscatter behavior, such as precipitation-sized droxtals, might be disregarded using heuristic arguments. However, the remaining spread of backscattering properties from various frozen particle models is significant. Clearly, based on the 60+ years of experience with regular weather radars, a unique, globally valid answer is unlikely to be found. Thus, efforts to estimate and report uncertainties and errors associated with snowfall observations are highly desirable. Further studies should, in particular, perform closure experiments using additional information. Dedicated aircraft campaigns and long-term ground validation sites, as well as combined active and passive observations, might help to establish smaller error margins and reject unrealistic estimates.

Closely related to the accuracy of instantaneous retrievals is the problem of precipitation detection and clutter removal. The high vertical resolution of spaceborne radars offers a distinct advantage over the ground-based scanning system. Nevertheless, this study shows that ground clutter effects are problematic in the CloudSat data and need to be carefully removed, especially in highly structured terrain. This study also illustrates that the use of “near surface” reflectivity bins (i.e., bins located ~1 km above the surface) alone might be insufficient to effectively eliminate all sources of ground clutter. The use of vertical continuity thresholds provides a simple measure to eliminate many false returns, although some embedded clutter remains even if such thresholds are utilized and further quality-control measures are necessary. Vertical continuity thresholds may also reject legitimate low-topped precipitation events, so an enhanced cost–benefit analysis of using these thresholds to help remove clutter that can severely bias the snowfall retrievals must be undertaken. Ground-based, vertically pointing cloud radars might offer a unique perspective to study whether significant low-topped snowfall does occur since it is immune to the clutter contamination that affects the lowest CloudSat
data bins. Ground-based instruments can also be effective tools to develop enhanced relationships between the near-surface reflectivity bins (located above 1 km) used in this study and the actual surface reflectivity for snowfall events.

It is important to acknowledge further potential deficiencies in this study that warrant further attention in subsequent research efforts. The results of this study may suffer from the reliance on near-surface reflectivity observations, as situations may arise that can cause an overestimation (e.g., virga) or underestimation (e.g., very shallow precipitation) of precipitation if no reflectivity data are available below the 1-km level [see Hudak et al. (2008) for examples of these issues]. Additionally, errors in the temperature data used in this study may adversely affect the results, as any systematic ECMWF-derived temperature biases may also cause the number of snowfall events in the dataset to be misrepresented. Furthermore, vertical temperature information is not used in this study, so under certain meteorological conditions like elevated temperature inversions, freezing rain or brightband signatures may be inadvertently included in the snowfall dataset and artificially elevate reflectivity counts and alter the detectability statistics. The main assumption of “dry” snowfall utilized in this study may also be unrealistic in snowfall cases where significant supercooled water exists, as the backscattering characteristics of the ice particle models used in this study may not adequately represent heavily rimed frozen particles. Combined active and passive microwave observations of snowfall events may provide a better assessment of the percentage of snowfall cases that have significant columnar supercooled water content and the potential for excessive riming. Last, attenuation and multiple scattering effects need to be further quantified for snowfall events. Attenuation probably does not play an important role for a large percentage of the light snowfall events represented in this study, but the higher end of the reflectivity distribution may be underestimated by as much as 1–2 dB, especially for events like heavy lake effect snow that contain significant cloud liquid water. Multiple scattering effects have been shown to artificially increase the radar signal and would tend to counteract the weakening of the signal due to attenuation, but preliminary evidence using CloudSat data has indicated that these effects are most pronounced at moderate–heavy precipitation rates (e.g., Battaglia et al. 2008; Matrosov et al. 2008a). Further studies will be necessary to quantify the degree to which they adversely affect the results presented in this study.

Despite these shortcomings and the large uncertainties associated with snowfall retrievals, this study shows how multifrequency, active spaceborne observations—with the inclusion of 94-GHz observations—offer a distinct advantage over lower-frequency single- or dual-frequency methods for retrieving dry snowfall. The results of this study indicate that the near-surface dry snowfall detection efficacy of a dual-frequency radar operating at 35 (13.6) GHz may suffer and might only approach 7% (1%), which translates into about 17% (4%) of the total global snowfall accumulation. These results should be considered preliminary though, and are subject to large potential errors that have been previously discussed. However, this study shows that high-frequency active radar observations can be extremely beneficial and can augment future dual-frequency observations of the GPM DPR at the lowest snowfall rates by building a priori or concurrent snowfall-rate distributions that adequately capture the entire snowfall-rate spectrum. This study reveals strong regional differences in snowfall-rate/reflectivity distributions, however, so these regional effects must be taken into consideration when using the CloudSat data (or future cloud radars) to enhance snowfall retrievals of lower-frequency instruments. Additionally, the 1-yr snowfall dataset presented in this study is admittedly limited, so sustained multiyear CloudSat observations will be essential to build more robust global and regional snowfall climatologies, and a strong argument can be made for an extended CloudSat mission that would truly benefit snowfall research. Ideally, future active remote sensing missions that study precipitation at higher latitudes will include a multifrequency instrument to gain the most benefit from spaceborne observations, or minimally include a parallel cloud radar mission that will assist the lower-frequency instruments at adequately capturing the lowest snowfall rates.

Acknowledgments. The authors thank Dr. Gang Hong for generously supplying DDA results that were used in this study and Mr. Michael Hiley for extensive data processing assistance. This research was partially supported by NASA Grant NNX07AE29G and by the Joint Center for Satellite Data Assimilation (JCSDA). The comments and suggestions from three anonymous reviewers are also gratefully acknowledged.

REFERENCES

