The annual cycle of snowfall at Summit, Greenland

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THE ANNUAL CYCLE OF SNOWFALL AT SUMMIT, GREENLAND

by

BENJAMIN BRIAN CASTELLANI

B.S., The Pennsylvania State University, 2010

A thesis submitted to the
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has been approved for the Department of Atmospheric and Oceanic Science

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The final copy of this thesis has been examined by the signatories, and we
Find that both the content and the form meet acceptable presentation standards
Of scholarly work in the above mentioned discipline.
Abstract

The mass balance of interior Greenland is much less understood than the oft-studied melting coastal regions due to a dearth of observations. The ICECAPS project, which launched in the summer of 2010 at Summit, Greenland, offers a unique, ground-based opportunity to study precipitation in central Greenland where the mass balance is marginally positive. Combining the perspectives from a Precipitation Occurrence Sensor System (POSS), a Millimeter-wavelength Cloud Radar (MMCR), and an accumulation field, the annual cycle of precipitation at Summit is examined. The annual average snowfall measured by the POSS (MMCR) is 83.3 mm (88.8 mm) of liquid equivalent, with the seasonal cycle defined by a large peak in summer and a smaller one in late winter. Accumulation showed a similar seasonal pattern, though with damped variability and a one or two month time lag. Daily snowfall increases by a factor of 3 from June through October compared to the rest of the year, while accumulation only increases 18% during the same timeframe. This reduced variability is explained by the seasonally-varying nature of latent heat flux, compaction, and wind contributions. The ICECAPS remote sensors and the accumulation field measurements do not completely agree as far as total annual liquid equivalent. The deposition of snow by wind, among other factors, is suggested as a possible contributor to the discrepancy. To further examine the seasonal cycle, snowfall measurements by the POSS were linked to other local meteorological parameters, including wind direction, liquid water path (LWP), 2-m
temperature, and precipitable water vapor (PWV). Snowfall rarely occurs and is typically
very light if the wind does not have a southerly component, except in the summertime, for
which moderate snowfall often coincides with all wind directions. Snowfall rate and
occurrence are higher when PWV exceeds the current month’s mean PWV. The wind
direction and moisture dependence are consistent with snowfall being linked to pulses of
moist air that originate over nearby, ice-free ocean, a resource that becomes more readily
available in summertime as the winter sea ice retreats. LWP is shown to have little
relationship to snowfall, indicating that ice-phase precipitation processes are quite important
for snowfall at Summit.
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The annual cycle of snowfall at Summit, Greenland

1. Introduction

The global mean temperature has increased 0.75 °C between 1880 and 2008, a change that has been attributed to both natural and anthropogenic influences [Intergovernmental Panel on Climate Change (IPCC) 2007]. The most significant changes are apparent in the Arctic (Washington and Meehl 1989; Haeberli 1990; Luckman et al. 2006), where temperature increases are proposed to be at least two times greater than in lower and middle latitudes (IPCC 2007; Chylek et al. 2009). The precise strength of this Arctic amplification, as it is called, varies significantly due to the non-linear nature of Earth’s climate system.

Peltier and Tushingham (1989) were some of the first researchers to propose that global mean sea level (GMSL) rise was related to warming from the “greenhouse effect.” Church and White (2006) found that GMSL has risen 195 mm from 1870 through 2004, and that the increase has been accelerating. The two main contributors to GMSL rise are the thermal expansion of the sea water and the addition of fresh water to the ocean from melted ice.

The Greenland Ice Sheet (GIS) is made up of nearly 3 million cubic kilometers of ice, which if melted would raise the GMSL six to eight meters (Ohmura and Reeh 1991; Alley et al. 2005; Gregory and Huybrechts 2006). Furthermore, the injection of fresh melt water from the GIS into the North Atlantic Ocean has been modeled to have a potentially significant and rapid impact on the global thermohaline circulation (Rahmstorf 1995; Rahmstorf 2000; Fichefet et al. 2003, Stouffer et al. 2006). In general, net melting is seen on the edges of the ice sheet, while slight net accumulation occurs in the interior (Krabill et al. 2000). These two regions compete as the major players in the overall GIS mass balance. Determining their
exact magnitudes and trends is essential for understanding how GMSL will be impacted, as it is the addition of mass from melted ice that dominates this change (Ohmura and Wild 2006). The latest estimates indicate the overall negative mass balance of the GIS has contributed around 8mm to sea level rise for the period 1992-2011 and the rate is increasing (van den Broeke et al. 2009; Shepperd et al. 2012). Rignot and Kanagaratnam (2006) found through the use of satellite interferometry that Greenland’s contribution to GMSL rise more than doubled from 1996 to 2005. Many coastal glaciers have shown acceleration over this time period (Howat et al. 2005), further confirming the decreasing mass balance. While much of the current research is focused on the melting coastal areas, a more complete understanding of the positive mass balance in the interior regions of the GIS is essential for diagnosing Greenland’s net contribution.

New measurements can help fill some important knowledge gaps regarding the central GIS mass balance. Recently the Integrated Characterization of Energy, Clouds, Atmospheric state and Precipitation at Summit (ICECAPS) project launched at Summit, Greenland (72.6°N, 38.5°W, 3260m above sea level; Fig. 1), a research site located near the apex of the GIS in a region where the mass balance is thought to be near equilibrium or slightly positive (Thomas et al. 2000). The overarching goal of the project is to examine how clouds and atmospheric state impact the mass and energy balances of the GIS through the use of ground-based remote sensing instruments (Shupe et al. 2013). This project commenced in the summer of 2010 and is a collaborative effort between several universities and agencies. Included in the instrumentation (Fig. 2) are two depolarization lidars, a ceilometer, a millimeter-wave cloud radar, a precipitation occurrence sensor system, two microwave radiometers, a spectral infrared interferometer, a sodar, a sky imager, radiosondes, and periodic ice crystal imaging. ICECAPS offers an unrivaled opportunity to study snowfall in
the center of the GIS because of the vast array of complementary, co-located, ground-based remote-sensing instruments and the project’s prolonged measurement period for understanding the interactions among clouds, atmospheric state, and precipitation.

Figure 1. Map of Greenland with the location of Summit shown.
This paper focuses on ICECAPS measurements that are useful for characterizing precipitation. The first objective is to evaluate the ICECAPS snowfall measurements against each other to gauge the uncertainty associated with the measurements and retrievals. Then, using these measurements, the annual cycle of precipitation at Summit will be presented and compared to independent measurements from an accumulation field. Finally, snowfall will be examined in the context of local meteorological conditions to better understand the sources of precipitation.

2. Instrumentation
Numerous data sets from several instruments will be utilized to understand the precipitation at Summit. A brief description of these particular instruments follows.

2.1. Precipitation Occurrence Sensor System

The Precipitation Occurrence Sensor System (POSS; Sheppard and Joe 2008) is a bistatic, continuous-wave, X-band (wavelength of 2.85cm) Doppler radar that was originally developed by the Meteorological Service of Canada for the observing precipitation occurrence, rate, and type. The transmitter and receiver are separated by 45 cm and are both pointed upward, angled 20 degrees from vertical, with an approximate sampling volume of one cubic meter located around their intersecting sight lines. The POSS measures the Doppler velocity spectrum of any hydrometers in its sampling volume. The POSS has mainly been used since the early 1990s for liquid precipitation applications (Sheppard 1990; Sheppard and Joe 1994; Campos and Zawadzki 2000; Sheppard 2007; Sheppard and Joe 2008), but, to a lesser extent, has been used for solid precipitation as well (Sheppard and Joe 2000; Sheppard and Joe 2008; Shiina et al. 2010).

2.2. Millimeter-wave Cloud Radar

The Millimeter-wave Cloud Radar (MMCR; Moran et al. 1998), which operates at 35 GHz and has Doppler capability, measures the backscattered power and Doppler spectrum of scatterers for a profile of range gates in the vertical and is used to detect cloud presence, boundaries, and phase (Shupe 2007). Radar reflectivity can be derived from the backscattered power and is roughly proportional to the liquid equivalent diameter to the sixth power, summed over the size distribution, assuming the scatterers meet the Rayleigh criteria. The MMCR self-calibrates the entire system, except the antenna, on a monthly basis and has a nominal uncertainty in reflectivity of 1 dB (personal communication, Duane Hazen 2014).
Reflectivity and snowfall measurements from the MMCR will be compared to those from the POSS.

2.3. Snow Accumulation Field

To provide an additional perspective on the mass balance at Summit, the data from a nearby snow accumulation field are incorporated into this analysis. The accumulation field is a 10 x 10 grid of bamboo stakes with 8 m spacing outfitted with measurement markings. The field is located in a remote, undisturbed location within 1 km of the ICECAPS measurements, and is used to measure changes in the height of the snow surface (hereafter referred to as accumulation; Dibb and Fahnestock 2004). Roughly once per week, a measurement is taken by an observer at each stake in the field to a precision of 0.1 cm (lower precision for the earlier part of the data set). The field provides the longest standing semi-real-time snowfall record at Summit stretching back ten years to 2003. Excluding human subjectivity or error, most variability in this data set comes from drifting and blowing snow. Often many of the stakes will see opposing height changes during the same week. To reduce this error, all 100 stakes are averaged to give a single value for each measurement period. These data are not directly associated with the ICECAPS project and are used to provide an additional perspective in Sect. 5.1.

2.4. Microwave Radiometers

Two microwave radiometers are present at Summit that passively measure downwelling atmospheric microwave radiation. The Humidity and Temperature Profiler microwave radiometer (HATPRO; Rose et al. 2005) operates at numerous channels ranging from 22 to 58 GHz, while the High-Frequency Microwave Radiometer (MWRHF) operates at 90 and 150 GHz. Measurements from 20 to 30 GHz are used to derive the column-
integrated precipitable water vapor (PWV; Turner et al. 2007), while measurements from 23, 31, 90, and 150 GHz are used to derive the column-integrated liquid water path (LWP; Liljegren 1999; Turner et al. 2007). These parameters play an important role in relating the snowfall measurements by active sensors to the local atmospheric state.

2.5. Temporary Atmospheric Watch Station

The NOAA Earth System Research Laboratory Global Monitoring Division has had a presence at Summit since the mid-1990s. Currently their Temporary Atmospheric Watch Observatory (TAWO) includes measurements of 2-meter temperature and relative humidity, 10-meter wind speed and direction, 10-meter temperature and station pressure. Several of these measurements will be used in the analysis contained in Sect. 5. TAWO is not directly associated with the ICECAPS project.

2.6. IcePIC Camera

The IcePIC camera is a Nikon D50 DSLR camera mounted on ~5.6x magnifying microscopic body (designed after one developed by Kenneth G. Libbrecht) used to image ice crystals that have fallen to the surface. Snowfall is routinely collected on a microscope slide to be imaged on average once per week, with the frequency being higher in the summer and lower in the winter. While this process is subjective, it does give information about the actual ice crystal habits being observed at the surface. These images are used to make qualitative generalizations about the snow at Summit. Actual images of typical ice crystals observed at Summit are shown in Fig. 3.
Figure 3. Ice crystal images captured by IcePIC including the crystal types (a) capped column, (b) hexagonal plate, (c) bullet rosette, (d) hollow column, (e) stellar plate, (f) needle, (g) stellar dendrite, (h) sectored plate, (i) simple star, (j) radiating dendrite, and (k) fernlike stellar dendrites.

2.7. MicroPulse Lidar

The MicroPulse Lidar (MPL; Campbell et al. 2002) emits polarized 532 nm light upwards and measures the backscattered radiation for various vertical range gates. The backscatter responds most strongly to the total surface area of the scatterers and hence is larger for liquid clouds, which consist of high concentrations of small droplets, than it is for ice clouds, which are typically composed of a lower concentration of larger crystals. The polarization of the MPL allows for cloud phase to be further distinguished by the
depolarization ratio, which helps determine if the scatterers are spherical liquid droplets or non-spherical ice crystals (Intrieri et al. 2002). The MPL works well to detect cloud presence, base, and phase. This data is used exclusively in Sect. 2.8.

2.8. Example Snowfall Cases

The ICECAPS instruments work to complement each other and to give the clearest possible understanding of the clouds, atmospheric state, and precipitation in the skies above. Two typical, but inherently different snowfall cases are presented in order to demonstrate the complimentary perspective of all relevant ICECAPS instruments in regards to snowfall.

2.8.1. Case A

This snowfall event occurred between 0 and 12 Z on 6 August 2012 and represents a typical moderate snowfall event that occurs regularly during the warmer summer months at Summit. The imaged sectored plates from this case taken at 9Z are lightly rimed (Fig. 4a), indicating to have originated within or fallen through a mixed-phase cloud. Figures 4b-h shows the data for the date for several of the ICECAPS remote sensing instruments. The POSS shows a moderate snowfall rate peaking near 0.5 mm hr$^{-1}$ (Fig. 4h). The backscatter from the MPL (Fig. 4c) shows the presence of three separate liquid cloud layers that persist most of the day, located at 3.8 km, 300 m, and 100 m above ground level, indicated by the layers of high backscatter shown in white. The upper layer at 3.8 km occasionally is not seen by the MPL, as its signal quickly gets attenuated by the liquid cloud layers below, as is the case between 6 and 10 Z. Figure 4d
shows the radar reflectivity and it is apparent from the high reflectivity values (dark orange and red colors) and low MPL backscatter that ice crystals are forming within and falling out of the thin liquid cloud layers.

Figure 4. Case A: 6 August 2012 moderate snowfall event. (a) IcePIC images of rimed sectored plates taken at 9Z (b) Radiosonde data from the 12 Z launch (c) Relative backscatter from a lidar (d) Reflectivity from the MMCR (e) Doppler velocity from the MMCR (f) Spectral width from the MMCR (g) LWP measured by the microwave radiometers (h) Snowfall rate derived from the POSS.
The upper two liquid cloud layers are also evident in the radar spectral width data throughout the day (Fig. 4f). Unlike ice cloud crystals, liquid cloud droplets are quite effective at radiatively cooling their cloud tops, a process that drives substantial turbulence in the layer below (Shupe et al. 2008). This turbulence is seen as high values of spectral width. Figure 4e and 4f hints at the presence of riming between 6 and 10 Z as well, indicated by the particularly large Doppler velocities and spectral widths stretching all the way to the ground. The accretion of liquid cloud droplets increases the mean fall speed of the ice crystals, but also increases the spectral width as riming can vary significantly depending on ice crystal habit, size, and orientation. This timing also coincides with 50 to 150 g m$^{-2}$ of LWP measured by the microwave radiometers (Fig. 4g), further confirming the potential for riming. The radiosonde data from 12Z (Fig. 4b) indicates the temperature of the upper cloud deck where the crystals likely formed was around -20°C. Only the upper cloud layer has an accompanying temperature inversion. The 2-meter temperature was about -5°C, and the cloud was not decoupled from the surface (no surface-based inversion present). This coupling with the surface moisture source may play a role in producing the summertime moderate snowfall events.

3.2. Case B

This snowfall event occurred sporadically between 9 and 15 Z on 30 September 2012 and represents a typical non-summer light snowfall scenario. The dominant crystal types imaged by IcePIC at around 12Z were columns and bullets (Fig. 5a). Again, remote sensing data for the day is given in Figures 5b-h. Snowfall rates shown by the POSS were
intermittent and light, peaking at only around 0.15 mm hr\(^{-1}\) (Fig. 5h). A single liquid layer was present from 8 to 11 Z at about 1.5 km above ground level (Fig. 5c). However, at the time the crystals were imaged, no liquid clouds were present, only a sole ice cloud with a top at around 1.5 km above ground level. This is shown by the lack of high spectral widths seen by the cloud radar (Fig. 5f) and no substantial MPL backscatter (Fig. 5c) at this time. The low LWP (less than 30 g/m\(^2\) which is approximately the error bars for the LWP retrieval itself) measured by the microwave radiometers is in agreement as well. The radiosonde from 12 Z (Fig. 5b) shows the temperature of the cloud at 1.5 km where the columns and bullets likely originated was about \(-30^\circ\)C and the 2-meter temperature was around \(-35^\circ\)C. In this case, there was a strong surface-based inversion present and the cloud layer was decoupled from the surface, increasing the likelihood that the moisture for this event was advected in to cloud level from afar. This will be discussed further in Sect. 6.
Figure 5. Case B: 30 September 2012 light snowfall event. (a) IcePIC images of columns and bullets taken around 12 Z. (b) Radiosonde data from the 12 Z launch. (c) Relative backscatter from a lidar. (d) Reflectivity from the MMCR. (e) Doppler velocity from the MMCR. (f) Spectral width from the MMCR. (g) LWP measured by the microwave radiometers. (h) Snowfall rate derived from the POSS.
3. Radars and Snowfall

3.1. Data

Radars have been used to examine precipitation processes for many years. Using radar for solid precipitation is more challenging because snowflakes come in many forms, with widely varying sizes and habits (geometric shapes), which all have their own unique and complex scattering properties (Fujiyoshi et al. 1990; Matrosov 1992; Macke et al. 1996; Liu 2008). This variation serves as a major source of uncertainty for snowfall measurements by radar and has been a topic of research for radar meteorologists for decades, in both identifying which habits are occurring and the associated scattering properties. Fortunately, improvements in computational power have allowed scattering codes, such as the T-matrix method (Mischenko 2000) and the Discrete Dipole Approximation (Purcell and Pennypacker 1973), to model the backscatter properties of intricate non-spherical particles. However, this still does not address the ambiguity of which ice crystal habits are actually occurring during any given event that are needed for the derivation of parameters such as snowfall rate. Habits can vary significantly from location-to-location, storm-to-storm or even hour-to-hour at a single location. The implication of this will be discussed more in Sect. 3.2.

The POSS is a vital component to understanding the snowfall at Summit. One important data set used from this instrument is the equivalent reflectivity factor, $Z_e$, (hereafter, reflectivity), and is derived using two equations. The first being the bi-static radar equation given by

$$P_t = \frac{P_t G_t G_r A^2 \sigma(\theta)}{64 \pi^3 R_t^2 R_r^2 R},$$

(1)
where $P_r$ is the average power received by the radar from a single scatterer, $P_t$ is the average transmitted power, $G_r$ is receiver antenna gain, $G_t$ is the transmitter antenna gain, $\lambda$ is the wavelength of transmitted radiation, $\sigma(\theta)$ is the radar scattering cross section and is a function of $\theta$ (scattering angle), and $R_t$ and $R_r$ are the distances to the scatterer from the transmitter and receiver, respectively. Equation (1) is solved for $\sigma$, which then contributes to

$$Z_e = \frac{\sigma \lambda^4}{|K_w|^2 \pi^5},$$

where $\sigma$ is the integral of backscattering cross section from all particles in the volume and $K_w$ is a function of the dielectric constant and represents the scattering and absorption properties of the scatterers. The units of $Z_e$ are defined as mm$^6$ m$^{-3}$. Due to several factors specific to the POSS, the reflectivity here is not true reflectivity. However, using a simulation model, the true reflectivity can be estimated from the POSS reflectivity by accounting for the non-180 degree angle between the transmitter and receiver of the bistatic system, as well as the fact that the sampling volume of the POSS is different for particles with different scattering cross sections (which is related to particle size). The greater scattered signal produced by larger particles can be detected at further distances from the center of the volume. This produces a central region where the measured scattered signal is maximized, hence the need for a simulation model.

There are several inherent sources of uncertainty that need to be considered for measurements taken by the POSS. The first is attenuation from radome wetting. In heavy rainfall, attenuation of the reflectivity signal is on the order of a few dB (van de Beek et al. 2010). For a location like Summit, it never rains and falling snow is relatively light and rarely melts, making this error almost non-existent. Attenuation of the radar signal by the scatterers themselves is also negligible because of the close proximity of the scatterers and
the small size of sampling volume. Another source of potential uncertainty arises from the horizontal wind speed. Because of the particularly slow vertical fall speed of most solid precipitation, even a slight horizontal wind can cause the hydrometers to fall at angles far from vertical, which will broaden the Doppler velocity spectrum. However, Sheppard and Joe (2008) have shown that wind has a minimal effect on the 0\textsuperscript{th} moment of the Doppler spectra and therefore on the estimate of the reflectivity. Finally, there is the uncertainty associated with large particle non-Rayleigh effects. Snowfall at Summit is somewhat unique, generally consisting of dry, pristine crystals with snowfall rates rarely exceeding 0.5 mm hr\textsuperscript{-1} of liquid equivalent. A distinct lack of aggregates means snowflakes are mostly smaller than a couple of millimeters and thus non-Rayleigh effects are likely not important for the X-band POSS at Summit (Sheppard and Joe 2008; Matrosov et al. 2009).

While the POSS samples about one cubic meter of atmosphere located a few meters above ground level (AGL), the MMCR is a pulsed radar system that measures a vertical profile. Sample volumes are on the order of hundreds of cubic meters, with the lowest reliable range gate located near 200m AGL. In order to choose the measurements that best represent the surface, the reflectivity values in the lowest reliable MMCR range gate are used. This provides the best guess as to what is actually reaching the ground, under the assumption that the reflectivity will be relatively constant down to the surface on average (Matrosov et al. 2008).

During a portion of the measurement period, a saturation issue impacted the data in the lowest MMCR range gates. Saturation occurs when the input signal is too strong for the detector. The dynamic range of the MMCR detector is finite and, for a portion of the experimental record, a component failure effectively shifted this range toward lower powers, causing detector saturation at the highest observed powers. This issue affected the lowest
several range gates and only when the actual reflectivity was above a certain threshold. For the affected gates of the MMCR, this threshold has been found to be ~5 dB. The saturation issue caused the MMCR to underestimate the reflectivity in these lower gates during events of moderate snowfall. A correction was applied to data where saturation was deemed to be occurring. During these times, the height that is selected to best represent the surface is at a higher location in order to be above the gates impacted by saturation, and could be as high as 700m AGL, instead of the usual 200m AGL.

3.2. Snowfall Retrievals

Radar reflectivity is often related to snowfall rate, \( S \), through equations known as \( Z_e - S \) relationships, which have the general power-law form,

\[
Z_e = BS^\beta
\]

where \( B \) and \( \beta \) are coefficients and \( S \) has units of mm hr\(^{-1}\) and is expressed as the liquid equivalent snowfall rate. These coefficients are not fixed and can vary significantly based on crystal habits and the size distribution of the snow. Thus, no one set of coefficients is generally applicable. \( Z_e - S \) relationships are more complicated for snow than for rain due to uncertainties in crystal habit, orientation, fall velocity, density, degree of riming, size, and the impacts all of these have on the scattering properties. Some of the coefficients are shown in Table 1 for such types of snow as plates, columns, dendrites, generic “snowflakes”, and dry snow. As an example, the \( Z_e - S \) relationships for several habits from the POSS are shown in Fig. 6. Notice the large variation in snowfall rate between various habits, especially at larger reflectivities.
Table 1. B and β coefficients for various Z_e-S relationships from the literature (adapted from Rasmussen et al. 2003)

<table>
<thead>
<tr>
<th>Source</th>
<th>Crystal Type</th>
<th>B</th>
<th>β</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ohtake and Henmi (1970)</td>
<td>Plates/Columns</td>
<td>90</td>
<td>1.6</td>
</tr>
<tr>
<td>Ohtake and Henmi (1970)</td>
<td>Needles</td>
<td>208</td>
<td>1.9</td>
</tr>
<tr>
<td>Ohtake and Henmi (1970)</td>
<td>Stellar Crystals</td>
<td>403</td>
<td>1.5</td>
</tr>
<tr>
<td>Ohtake and Henmi (1970)</td>
<td>Spatial Dendrites</td>
<td>739</td>
<td>1.7</td>
</tr>
<tr>
<td>Imai (1960)</td>
<td>Dry Snow</td>
<td>120</td>
<td>2</td>
</tr>
<tr>
<td>Puhakka (1975)</td>
<td>Dry Snow</td>
<td>235</td>
<td>2</td>
</tr>
<tr>
<td>Gunn and Marshall (1958)</td>
<td>Snowflakes</td>
<td>448</td>
<td>2</td>
</tr>
<tr>
<td>Sekon and Srivistava (1970)</td>
<td>Snowflakes</td>
<td>339</td>
<td>2.21</td>
</tr>
</tbody>
</table>

Figure 6. Colored lines show Z_e-S relationships derived specifically for the POSS using the T-matrix method by Sheppard and Joe (2008) for six different ice crystal habits. The solid black line is the actual effective relationship used by the POSS. The dashed black line shows the relationship used for the MMCR in this analysis.
Ze-S relationships for radars are largely derived in one of three ways. The first involves taking simultaneous measurements of both reflectivity and snowfall rate using a snow gauge or balance giving the direct relationship between the two (Muramoto et al. 2003; Super and Holroyd 1998; Hunter and Holroyd 2002; Fujiyoshi et al. 1990; Rasmussen et al. 2003). The second method is similar, but is only based on observations of snow size distributions (Sekhon and Srivastava 1970; Ohtake and Henmi 1970; Yagi et al. 1979). These two methods allow for the relationship to be determined based entirely on real measurements. The third means for determining Ze-S relationships is by using scattering models (Barber and Yeh 1975; Matrosov 2007; Matrosov et al. 2009). This method models the backscatter properties of the snowflakes. By making assumptions about the characteristics of the snowflake size distribution, both a theoretical reflectivity and snowfall rate can be determined, as well as their power-law relationship to one another. The results, even within each method, vary widely. In this paper, the relationships used for the MMCR and POSS measurements were developed using the T-matrix scattering model.

Due to the hydrometer size dependence of the POSS’s sampling volume, the 0th moment (Po) of the measured Doppler spectrum is not linearly related to Ze. This makes determining S more of a challenge than for a pulsed radar system like the MMCR. A second-order regression of log(S) on log(Po) can be used to find S (Sheppard and Joe 2008). This method is unique to the POSS, but is equivalent to Ze-S methods for other radars. As with the Ze-S power-law relationships, the coefficients involved depend on the form of the SSD. The assumed SSD used by the POSS operational software is that found in Sekhon and Srivastava (1970), which has the general form,

\[ N(D_m) = N_0 e^{-\lambda D_m}, \]

\[ (4) \]
where \( N_0 \) is 2500 \( S_{ss}^{-0.94} \text{ m}^{-3} \text{ mm}^{-1} \), \( \lambda \) is 2.29 \( S_{ss}^{-0.45} \text{ mm}^{-1} \), \( S_{ss} \) is the model water equivalent snowfall rate in mm h\(^{-1}\) and \( D_m \) is the “melted” diameter given in the Sekhon-Srivistava (1970) model. More details on this POSS retrieval are given in Sheppard and Joe (2008).

MMCR-based snowfall retrievals follow many of the same principles, however, since the MMCR is a pulsed system, a \( Z_e-S \) relationship for dry snow described and developed by Matrosov (2007) under a basic set of assumptions for millimeter wavelengths is used to derive all mass-related parameters,

\[
Z_e = 56 \, S^{1.20}. \tag{5}
\]

The MMCR’s uptime during the measurement period has been intermittent at times. For this reason, the POSS is the primary instrument considered here, but it will be compared to and supplemented by measurements from the MMCR.

4. Evaluation of Methods

The objective here is to evaluate and compare the POSS and MMCR measurements and retrievals for consistency. The POSS has been collecting data at Summit since September of 2010 with >99% uptime. Unfortunately, no independent precipitation gauges exist at Summit to directly evaluate the POSS for precipitation measurements at the surface. However, the MMCR, which was operational periodically through the time period, can be used. The equivalent reflectivity derived from the POSS \( (Z_{EP}) \) is compared to the equivalent reflectivity measured by the MMCR \( (Z_{EM}) \), an instrument with a relatively well-tracked and characterized calibration when it is operating properly. This intercomparison provides some baseline evaluation of the POSS calibration, as well as the assumptions that go into the scattering code that has been sued to develop the retrievals for both instruments.
The POSS and the MMCR operate in different radar bands, with wavelengths of 2.85 cm and 8.5 mm, respectively. This difference causes dissimilarities in the way the radar beams interact with larger snowflakes. For centimeter wavelengths, the Rayleigh approximation holds well for particle sizes of up to 4 or 5 mm (Matrosov et al. 2009). However, for the millimeter-scale wavelength of the MMCR, scattering can deviate significantly from Rayleigh theory. The derivation of $Z_e$ for the POSS is calculated from Mie theory and makes no assumptions about particle size. For particles smaller than a couple of millimeters, the $K_a$-band reflectivities will be smaller by a few dB or less (Matrosov 1998) and even less for the dry snowflakes at Summit (Matrosov 2007). Based on qualitative IcePIC images, most snowflakes at Summit are relatively small and fall below these thresholds. However, the potential contributions of non-Rayleigh effects to differences in reflectivity between these instruments gets larger for increasing particle sizes (higher $Z_e$).

Other concerns when comparing measurements by the POSS and MMCR are differences in sampling volume location and the MMCR’s saturation issue discussed in Sect. 3. Additionally, due to the measurement height difference of the POSS and MMCR, which may range from 200m to 700m AGL, a time lag of 10 minutes is introduced for the comparison to account for the time it takes ice crystals to fall from those heights to the surface. To further reduce the uncertainty in time and space, the data for both instruments was averaged temporally. Averaging windows ranging from one minute up to one hour were evaluated and 10 minutes was chosen as the best compromise between statistical representation and temporal resolution. While some of these details complicate the comparison, it is the best that can be achieved given the suite of instrumentation available at Summit over the time period of study.
The POSS and MMCR reflectivities are compared by considering instantaneous difference, $\Delta Z$ (Fig. 7a). This comparison is useful for evaluating the model-derived regressions used for calculating the POSS reflectivity. Consequently, since the same formalism and assumptions are used to derive snowfall rate from the POSS, the evaluation may also inform some of the uncertainties in that retrieval. The bolded black box-and-whisker plot on the far right contains all of the data. The POSS-derived reflectivity compares adequately with the MMCR, showing only a small bias of the mean $\Delta Z$ of -0.6 dB, and an even smaller negative median-bias (POSS lower than MMCR) of -0.2 dB. Here we evaluate the total difference between POSS and MMCR, $\sigma_{TOT}$, expressed as a standard deviation, by considering the individual components that might contribute to the differences, using

$$\sigma_{TOT} = \sqrt{\sigma_{MMCR}^2 + \sigma_{POSS}^2 + \sigma_{COMP}^2},$$

where $\sigma_{MMCR}$ is the uncertainty associated with the MMCR reflectivity and has a known nominal value of 1 dB, $\sigma_{POSS}$ is the effective uncertainty of POSS reflectivity that includes both the POSS calibration uncertainty and the uncertainty contributed by the regressions for determining the true POSS reflectivity, and $\sigma_{COMP}$ is the contribution introduced from differences in sampling volumes and time. Rearranging this equation and solving for $\sigma_{POSS}$ gives an estimate of POSS uncertainties:

$$\sigma_{POSS} = \sqrt{(\sigma_{TOT}^2 - \sigma_{COMP}^2 - \sigma_{MMCR}^2)}.$$  

Based on all three years of data (the bolded box and whisker plot in Figure 1a), $\sigma_{TOT}$ is equal to 5.3 dB. Thus, at the maximum possible uncertainty of the POSS (i.e., assuming no differences in sample volume and $\sigma_{COMP} = 0$) is 5.2 dB. However, we have summarized several notable complications with the comparison. One of the most influential is likely the
difference between the two sampling volumes. The MMCR measures a comparatively large volume a few hundred meters off the ground, while the POSS is just a few meters above the surface measuring a small volume about one cubic meter in size. The degree to which the POSS and MMCR measure the same general aspects of the snowfall depends on the horizontal wind speed, the homogeneity of the distribution of snow on the local scale, and even the properties of the snowflakes themselves that impact their fall speed, which determines the time lag before the snow at the height of the MMCR observations makes it to the height of the POSS observations. These differences due to sampling volume likely contribute significantly to the overall variability in observed differences in reflectivity. While these conditions cannot be fully characterized, some of them are examined below.
Figure 7. a) Box-and-whisker plots of the reflectivity difference, $\Delta Z$, for all data (bolded plot on the right) and for various $Z_{EP}$ bin subsets (6 plots on the left). b) Bar graph showing the number of occurrences that fall into each $Z_{EP}$ bin (black) and the relative mass-weight occurrence in each bin (red).
Figure 8 gives insight into potential contributions to observed differences by characterizing the differences between 10-min averaged bins of POSS and MMCR data and 10-min averaged bins of the same data that were offset by 1, 3, 5, and 10 minutes. Essentially, this is a proxy for how much reflectivity changes in time and can shed light on the variability of data expected by potential temporal offsets between POSS and MMCR sample volumes. At the top of each box-and-whisker in the figure, the standard deviation is shown. With a 3 or 5 minute offset, the standard deviation is already near 3 dB. It is important to note that both instruments show basically the same temporal self-variance statistics, suggesting that this variability is more likely related to how snowfall varies and is independent of the instruments. The blue box-and-whisker represents the difference between reflectivity measured by the MMCR at 200m and 700m during times when snowfall was occurring and serves as a proxy for the potential vertical variation in reflectivity that could impact the comparison. The positive bias (i.e., reflectivity at 200m being larger) suggests a net growth of snowflakes as they fall towards the surface. The standard deviation representing vertical variability across this 500m was found to be 4.8 dB.
Figure 8. Box-and-whisker plots showing the result of choosing 10-minute averaging bins for the POSS (red) and MMCR (black) that are offset from one another by 1, 3, 5, and 10 minutes. The plots show the difference between the offset bins with the one that was used in the analysis. These gauge the variability associated with choosing slightly different time windows for comparing the POSS and MMCR reflectivities. The blue box-and-whisker plot on the right has no bin offset and simply represents the difference between the MMCR reflectivities at 200m and 700m AGL.

Even though the overall measured uncertainty of the POSS was found to have an upper limit of 5.2 dB, based on the various proxies, the uncertainty associated with the time-space comparison alone can easily range between 3 and 5 dB. This gives more confidence to the POSS’s calibration and regressions. As a result of this information, the POSS is likely to be accurate to within 3 dB (i.e., a factor 2), if not better. This estimate is close to the 2.9 dB standard deviation found by Huang et al. (2014), who analyzed $Z_e$ differences between the POSS and a 2D-Video Disdrometer for snowfall.
Returning to Fig. 7a, the six leftmost black box-and-whisker plots show reflectivity differences categorized by reflectivity (subset by ranges of $Z_{EP}$). At lower reflectivities, the POSS tends to measure a lower reflectivity than the MMCR. Moving towards higher reflectivities, the MMCR begins to measure increasingly lower reflectivity than the POSS. This downward trend towards larger reflectivities has a few potential explanations. The likely culprit is the introduction of non-Rayleigh effects to the MMCR at higher reflectivities due to the larger particle sizes. Also, the MMCR saturation issue may not be fully accounted for in all situations. As mentioned, saturation only affects the MMCR at times of higher reflectivity, and it does so by lowering the measured MMCR reflectivity. Both of these effects would contribute in the manner observed in Fig. 7a. The apparent negative POSS bias at low $Z_{EP}$ is likely related to the small sampling volume and minimum detectable signal of the POSS. Most of the data fall within $Z_{EP}$ bins that have mean and median differences less than 3 dB. In total, 93% of the snowfall events and 78% of snowfall mass are within these bins (Fig. 7b).

A point-to-point comparison of instantaneous derived snowfall rates shows that the bulk of the data closely follows the one-to-one line (Fig. 9) with a correlation of 0.75 and a RMSE of 0.08 mm/hr. Despite the POSS appearing to have lower measurements of reflectivity than the MMCR at the smallest reflectivities, the opposite is true for snowfall rate (i.e., at the smallest snowfall rates, the POSS rates are higher). This is likely due to the differences in the way the $Z_e$-S relationships are developed for these two radar systems. Overall, the POSS has been shown to compare adequately with the MMCR for their overlapping measurement periods, giving us a better understanding and more confidence in the POSS-focused results presented in Sect. 5.
Figure 9. Scatter plot of POSS and MMCR snowfall rates for all times that both instruments were actively measuring snowfall. The dashed red line is the one-to-one line and the dotted red lines show a factor of two difference.

5. Results and Discussion

5.1. Annual Cycle of Snowfall

One basic, but extremely valuable piece of information is the annual cycle of snowfall in the central GIS. The POSS has been operating 24 hours a day since September of 2010 (with greater than 99% up-time), until October of 2013 for the analysis presented here. Figure 10a shows the annual cycle of snowfall through monthly means for the POSS during the full measurement period (red), and the MMCR for intermittent times during the period (green), as well as accumulation from the nearby accumulation field over the same period (dark blue). Additionally, a longer, ten-year period of accumulation from the same field stretching back into 2004 is included (light blue). Unlike the instantaneous measurements
taken by the POSS and MMCR, the accumulation field only provides three to five measurements per month.
Figure 10. a) The annual cycle of monthly snowfall at Summit measured by the POSS derived for Sep 2010 to Oct 2013 (red) and the MMCR for a subset of the period (green). The annual cycle of accumulation is given for the same period as the POSS (dark blue) and for Jan 2004 to Dec 2013 (light blue). Asterisks indicate monthly mean values for all data sets. For the short-term data, the extent of the vertical lines indicates the maximum and minimum monthly values during the period. For the longer term accumulation data, the extent of the vertical lines indicates plus and minus one standard deviation from the monthly mean. b) The annual cycle of deposition due to latent heat flux for Summit from Apr 1996 to Aug 1999, constructed from Table 5 of Box and Steffen (2001). c) The monthly density of the accumulation implied by POSS snowfall and latent heat contribution measurements, assuming no other contributing factors. The asterisks indicate monthly mean values, while the extent of the vertical lines indicates maximum and minimum monthly values during the three-year period.

The annual snowfall cycle indicated by the POSS and MMCR is defined by two peaks and two troughs. The summertime peak in JJAS is largest and tails off into the fall. There is another increase moving through winter, with a smaller peak seen in February. The annual
minimum occurs in springtime. July sees the most snowfall, averaging 17.0 mm of liquid equivalent from the POSS. MAM receives the lowest amount relative to the rest of the year, only averaging 3.0 mm per month. The mean annual snowfall over the time period from the POSS (MMCR; differing time period) is 83.3 (88.8) mm of liquid, with 61% (49%) occurring in JJAS. This annual value is comparable to those found in previous studies whose average value for precipitation was 90 mm of liquid per year for the area around Summit when derived from reanalysis and model products including the ERA (1979-1993), ECMWF Moisture Budget (1985-1995), NCEP/NCAR (1958-1996), the Keen Model (1964-1988), and the Chen-Bromwich model (1985-1995), that were given by Bromwich et al. (1998). In general, the reanalyses captured a similar annual pattern as observed by the ICECAPS sensors. The snowfall measured for July and January, 17 mm and 4 mm, match quite well with the values for Summit derived using the omega equation by Bromwich et al. (1999). The largest monthly variability occurs in May, June and September, where the difference between the maximum and minimum monthly snowfall totals was more than 8 mm, while the rest of the months see a range of variability between 3 and 7 mm. Despite the varying temporal coverage and differing snowfall retrievals, the POSS and MMCR annual snowfall cycles compare well, generally falling within 30% of each other for monthly means year-round. There are monthly differences, with MMCR larger during lower snowfall months and POSS larger during higher snowfall months, likely related to the differences discussed in Sect. 4 and Fig. 7.

The measured accumulation over the same time period also shows seasonal variability, though the maxima and minima are not in sync and the amplitude is smaller. Accumulation peaks in August and September, measuring 7.7 cm and 8.3 cm of solid accumulation, respectively. The late winter maximum seen in snowfall is not evident.
However, a spring minimum is likewise observed, with the lowest monthly mean accumulation occurring in June (2.2 cm). Each monthly mean of accumulation over the measurement period falls within one standard deviation of the 10-year monthly mean accumulation, suggesting that the 3-year data set is approximately representative of accumulation over the last decade. Over the entire 10-year period, the mean annual solid accumulation was 71 ± 11 cm. This value compares well with Bales et al. (2001) who found an average annual solid accumulation of 69 cm for the period of 1904 to 1974 and Dibb and Fahnestock (2004), who measured an annual mean solid accumulation of 65 cm for the periods of Aug 2000 to Aug 2002 and 1991 to 1995 based on accumulation field data.

To further understand the relationship between snowfall and accumulation, cumulative series of both data are plotted in Fig. 11. Snowfall and accumulation qualitatively track each other quite well throughout the time period. The rate of accumulation throughout the measurement period is relatively constant, which agrees with the findings of Steffen and Box (2001) and Dibb and Fahnestock (2004) for Summit. The snowfall rate, on the other hand, appears more akin to a step function, with minimal snowfall for a little more than half of the year in the winter months (Periods W1, W2, and W3), and increased snowfall in the summer periods (Periods S1, S2, and S3). For consistency purposes, the MMCR cumulative snowfall is shown in Fig. 11 for the time period when the data were reliable and continuous. Over that period, the MMCR measured a nearly identical amount of cumulative snowfall as the POSS, though their paths were not always in complete agreement.
Mean annual statistics are useful to quantify the major difference in amplitude between snowfall and accumulation (Table 2). In general, the months of June through October can be defined as the summer snowy season. The snowfall rate in the snowy season is 0.36 mm of liquid per day, three times larger than the rest of the year. The accumulation rate during the snowy season is 2.0 mm of solid accumulation per day, only about 15% larger than during the rest of the year.
Table 2. Snowfall and accumulation data for 6 consecutive time periods during the record of POSS snowfall measurements. The beginning and end of each period was subjectively chosen based on distinct regimes of snowfall rate visible in Figure 11. The mean values for the summer (S) and winter (W) periods are also shown.

<table>
<thead>
<tr>
<th>Time Period</th>
<th>Period Label</th>
<th>Total Days</th>
<th>Total Snow (liquid mm) POSS</th>
<th>Snow Rate (liquid mm/day) POSS</th>
<th>Total Snow (liquid mm) MMCR</th>
<th>Snow Rate (liquid mm/day) MMCR</th>
<th>Total Accumulation (solid mm)</th>
<th>Accumulation Rate (solid mm/day)</th>
</tr>
</thead>
<tbody>
<tr>
<td>11/1/10 – 5/31/11</td>
<td>W1</td>
<td>212</td>
<td>25.0</td>
<td>0.12</td>
<td>349</td>
<td>1.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td>6/1/11 – 10/31/11</td>
<td>S1</td>
<td>153</td>
<td>41.0</td>
<td>0.27</td>
<td>257</td>
<td>1.7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>11/1/11 – 5/31/12</td>
<td>W2</td>
<td>212</td>
<td>27.2</td>
<td>0.13</td>
<td>407</td>
<td>1.9</td>
<td></td>
<td></td>
</tr>
<tr>
<td>6/1/12 – 10/31/12</td>
<td>S2</td>
<td>153</td>
<td>58.5</td>
<td>0.38</td>
<td>331</td>
<td>2.1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>11/1/12 – 5/31/13</td>
<td>W3</td>
<td>212</td>
<td>27.9</td>
<td>0.13</td>
<td>333</td>
<td>1.6</td>
<td></td>
<td></td>
</tr>
<tr>
<td>6/1/13 – 10/31/13</td>
<td>S3</td>
<td>153</td>
<td>64.6</td>
<td>0.42</td>
<td>321</td>
<td>2.1</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mean (Winter)</td>
<td>W</td>
<td>212</td>
<td>26.7</td>
<td>0.13</td>
<td>363</td>
<td>1.7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mean (Summer)</td>
<td>S</td>
<td>153</td>
<td>54.7</td>
<td>0.36</td>
<td>303</td>
<td>2.0</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

In order to examine these seasonal differences in snowfall and accumulation, we first define accumulation, A, as the measured change in the height of the snow surface, according to:

$$A = P + L - C + W,$$  \hspace{1cm} (8)

where P is snowfall, L is the net contribution from latent heat flux (can be positive [deposition] or negative [sublimation]), C is compaction, and W is the influence of snow redistribution by the wind and can be positive or negative. The overall correlation between POSS snowfall and accumulation from Fig. 11 is 0.39. This relatively low correlation with snowfall suggests that other factors in (8) may be important. Either the latent heat flux contribution, wind redistribution, compaction, or a combination thereof is playing an influential role as well, and that their roles change seasonally.

In the summer months, when snowfall is at its peak, the net latent heat flux at Summit is negative or very small (i.e., net sublimation to the atmosphere, see Fig. 10b; Box and
Steffen 2001). Compaction, including the influences from destructive snow metamorphism and firn deflation, increases with increasing surface temperature (Spencer et al. 2001; Steffen and Box 2001; Arthern et al. 2010). Both of these factors would counteract the summertime snowfall maximum to lessen its impact on accumulation (Dibb and Fahnestock 2004).

Another result of these counteracting factors is that the annual peak in accumulation is notably delayed by two months, into September, relative to the snowfall peak which occurs in July. In June and July, when snowfall starts to increase, latent heat flux is negative and significant compaction of the wintertime accumulation occurs due to the warming temperatures. By the end of summer, in August and September, snowfall is still large. However, latent heat flux is no longer negative and the initial large rate of compaction has diminished. This allows for accumulation to increase and peak, even though snowfall is already trending downward. In the winter (October through April), the net latent heat flux is often positive (i.e., net deposition onto the surface; Box and Steffen 2001) and compaction is reduced as a result of the colder temperatures. Both of these would explain accumulation being large relative to snowfall, as seen in Fig. 10a.

Figure 10c shows the monthly mean implied snow pack density, $\rho_I$, that is implied by combining the snowfall and accumulation measurements with the estimates of latent heat flux. $\rho_I$ is generally less than 300 kg m$^{-3}$, except in January, which has seen widely varying $\rho_I$ up to 430 kg m$^{-3}$. The annual average implied density is 130 kg m$^{-3}$, which is far too low considering the density profile at Summit in the upper meter of snow typically ranges from 300 to 450 kg m$^{-3}$ (Albert and Shultz 2002; Jacobi et al. 2004). Stated in other terms, the mean snowfall of 83.3 mm (liquid equivalent) observed by the POSS is significantly smaller than the mean liquid equivalent accumulation of 210 mm that would result from a modestly low assumed density of 300 kg m$^{-3}$, a value that is similar to the 200 mm found by Bales et
al. (2009). The annual mean latent heat contribution (6 mm, Box and Steffen 2001) can only account for a small fraction of the difference.

There are several plausible explanations for this discrepancy. The first is uncertainty in both the snowfall and accumulation measurements. The mean accumulation of 71 cm tends to agree with values found in other studies for areas near Summit, determined from methods including accumulation fields, snow pits, ice and firn cores, and acoustic ranging (Steffen and Box 2001; Dethloff et al. 2002; Dibb and Fahnestock 2004). It is conceivable that accumulation near Summit is slightly enhanced compared to nearby undisturbed areas due to spatial drifting effects caused by the camp itself. This has not been quantified.

Uncertainty in the remote sensors can be a factor, yet the agreement between two independent retrievals is quite good. Based on the estimated uncertainty of the POSS detailed in Sect. 4 (a factor of 2), the mean annual snowfall could be as high as 167 mm. The POSS has also been shown in past studies to underestimate precipitation, with median “catch ratios” ranging from 0.63 to 0.79 and inter-quartile ranges of 0.36 to 0.50 when compared to Nipher gauges and Double Fenced Intercomparison Reference snow gauges for solid precipitation (Sheppard and Joe 2008; Wong 2012). “Under catch” could be due to sensitivity issues or other causes. Even including this expected “under catch,” the difference is still too large to be explained by the uncertainty in the POSS and accumulation data alone. The uncertainty in the coefficients of the MMCR Z_e-S relationship may be as large as a factor of 2, so the uncertainty in derived mass parameters could be substantial (Matrosov 2007). Finally, since the snowfall retrievals used by the instruments were designed for snowfall in a mid-latitude setting, they may not necessarily be completely applicable in a high altitude, extremely cold Arctic environment like Summit, where individual crystal densities may be smaller.
Another possible contributor to this disparity is the only additional source of mass contained in (8), the redistribution of snow by the wind below the height of the POSS (~6m). At Summit, the predominant wind direction is from the south-southwest (Steffen and Box 2001; Cohen et al. 2007; Shupe et al. 2013). A strong north-south gradient in precipitation and accumulation across Greenland is ubiquitous throughout the literature, as well as enhanced accumulation towards the coastlines to the southeast and southwest of Summit (Ohmure and Reeh 1991; Bromwich et al. 1998; Bromwich et al. 1999; Bales et al. 2001; Drinkwater et al. 2001; Bales et al. 2009). The combination of wind direction and snowfall gradient could make wind transport a source of added accumulation at Summit. Unfortunately, direct observations and modeling studies of drifting and blowing snow over Greenland are very sparse (Lenaerts et al. 2012; Budd et al. 2013), so pinpointing the magnitude of this potential contribution would be challenging.

5.2. Influence of Local Meteorology

The strong seasonal variability in snowfall at Summit suggests seasonally varying influences from meteorological conditions. Figure 12 shows the seasonal cycles of four parameters that may play roles in snowfall variability: 2m temperature, 10m wind speed, liquid water path (LWP), and precipitable water vapor (PWV). Temperature, LWP, and PWV show maxima in the summer and a general minimum throughout the rest of the year, while wind speed has a summertime minimum.
Figure 12. Monthly box-and-whisker plots for (a) 2m temperature, (b) 10m wind speed, (c) liquid water path, and (d) precipitable water vapor observed at Summit (Jun 2010 - Oct 2012).

The direction from which air masses impinge on Summit can influence snowfall. Figure 13 shows four panels that relate snowfall to seasonal variations in local 10m wind direction. The dominant wind direction throughout the year is southeasterly to southwesterly, with a more dominant westerly component in June (Fig. 13a). Figure 13b shows the percentage of time it was snowing when the wind was blowing from a given direction for each month. A mid-summer maximum is seen in all wind directions. Outside of June, July and August, there is less than 5% snowfall occurrence for all wind directions that have a northerly component. Figure 13c considers the mean snowfall rate when snowing. Again, we see the directional dependence of snowfall, with very low rates when the wind has any
northerly component for November through May. For all other wind directions and seasons, rates are approximately within a factor of 2, with the highest rates for southerly winds in May through September. The large maximum in May is likely not representative, but instead the result of one or two intense snowfall events during the measurement period occurring in a month which sees very little snowfall in general.

Figure 13. a) The occurrence of 10-m wind directions normalized by month of year, b) The percent of time it was snowing as a function of 10-m wind directions and month of year, c) The mean snowfall rate when it was snowing as a function of 10-m wind directions and month of year, d) The total liquid equivalent snowfall as a function of 10-m wind direction and month of year. Wind direction bin sizes are 15 degrees.
Combining Figs. 13a-c gives Fig. 13d, which highlights the most important parameter, mass. This plot confirms that nearly all of the mass at Summit between September and May occurs when the local wind direction has a southerly component. This is the result of three factors: the low occurrence of northerly wind directions during these months, a low occurrence of snowfall when the wind does have a northerly component and minimal snowfall rates during snowfall occurring under northerly winds. In summer, some mass results from conditions with northerly winds. The seasonal cycle of snowfall mass from northerly winds is likely a result of the presence of winter sea ice surrounding all of Greenland except the North Atlantic Ocean to the south. The diminished directional dependence in summer may be related to the springtime retreat of the sea ice surrounding Greenland, exposing open-ocean in most directions.

Another local variable that may be part of the linkage to open water is the moisture content of the air above Summit. The atmospheric boundary layer at Summit (and presumably over much of the GIS) is quite shallow, often less than 100 m in depth and frequently contains a very cold, surfaced-based inversion (Miller et al. 2013; Shupe et al. 2013). Because of this stratification, the regional ice sheet itself is likely not an adequate source of moisture for cloud and precipitation formation. As a result of this decoupling, most of the snowfall at Summit must be generated by clouds that are formed and sustained by moisture that was advected in aloft from afar. To examine moisture content, PWV derived from the microwave radiometers is used. Figure 14a shows the monthly occurrence of PWV peaking in the summer above 3 mm and generally less than 1.5 mm throughout the remainder of the year. Not surprisingly, the odds of snowfall occurring during any given month increases with increasing PWV, particularly when PWV is above the monthly mean value (Fig. 14b). With the exception of July and August, the occurrence of snowfall when PWV is
below the monthly mean is less than 5% of the time. Additionally, most appreciable snowfall rates occur under the same constraints (Fig. 14c). However, large PWV values are relatively infrequent. As a result, most snow mass occurs when PWV is near or above its monthly mean, and appreciable snowfall mass can even occur with PWV less than 1 mm in winter (Fig. 14d).

Figure 14. a) The occurrence of PWV normalized by month of year, b) The percent of time it was snowing as a function of PWV and month of year, c) The mean snowfall rate when it was snowing as a function of PWV and month of year, d) The total liquid equivalent snowfall as a function of PWV and month of year. PWV bin sizes are 0.5 mm. The black lines in each panel denote the mean PWV for all times, when it was snowing, and when it was not snowing.
Figure 14c shows the summertime maximum in LWP corresponding with warmer temperatures and more moisture availability. An intriguing unknown is whether or not the amount of cloud water has any influence on snowfall. As 2-m temperature increases, larger values of LWP are more frequently observed (Fig. 15a). For a given LWP, snowfall occurrence increases with increasing T until -15°C, where snowfall occurrence is maximized at all LWPs (Fig 15b). This T coincides with the maximum saturation vapor pressure difference between water and ice. While there is some indication of more frequent snowfall at higher LWP, the snowfall rate when it is snowing does not appear strongly dependent on LWP (Fig. 15c) and actually appears to have a weak maximum for LWP ~15-30 g/m² at all T. On the other hand, the snowfall rate when it is snowing clearly increases with T at all LWP. Figure 15d shows that, on average, LWP is lower when it is snowing compared to times when it is not snowing, and most mass occurs when LWP is below the mean for all T ranges. The snowfall mass distribution actually decreases at higher LWP, in large part due to the infrequent occurrence of high LWP. Together, these results indicated that (1) LWPs are quite small at Summit; (2) more efficient precipitation appears to coincide with periods where the Bergeron process has acted to enhance ice growth while depleting liquid water; and/or (3) ice-phase precipitation systems and processes can be quite important for snowfall at Summit.
Figure 15. a) The occurrence of LWP normalized by temperature, b) The percent of time it was snowing as a function of LWP and temperature, c) The mean snowfall rate when it was snowing as a function of LWP and temperature, d) The total liquid equivalent snowfall as a function of LWP and temperature. Bins for LWP are 5 g/m² and for temperature are 10 degrees. The black lines in each panel denote the mean LWP for all times, when it was snowing, and when it was not snowing.

6. Conclusions

To further the understanding of the central GIS mass budget, measurements of snowfall rate at Summit since September of 2010 are examined. POSS reflectivity and snowfall rate measurements where compared to those of the co-located MMCR, an instrument with a generally well-characterized uncertainty. The POSS was found to agree with the MMCR, both in terms of reflectivity and retrieved snowfall rate. Overall, the POSS
was found to have an uncertainty that is likely less than 3 dB for reflectivity. Such an error implies an uncertainty of a factor of 2 for snowfall retrievals, which is consistent with past evaluations of POSS-based retrievals and similar to MMCR-based retrieval uncertainties.

Using the POSS and MMCR measurements, the annual cycle of snowfall was presented showing a clear summertime maximum and a smaller, late-winter maximum. The annual mean liquid equivalent snowfall measured by the POSS (MMCR) was 83.3 mm (88.8 mm), which is within 10% of the estimates provided by reanalysis products. A comparison to the seasonal cycle of surface accumulation showed similar seasonal patterns, though the variability is dampened and a one or two month time lag exists. A cumulative time series of accumulation and snowfall shows that while accumulation increases at a relatively steady rate throughout the year, snowfall occurs in two distinct regimes. Snowfall rate increases by nearly a factor of three between June and October, compared to the rest of the year. Accumulation rate sees just an 15% increase during this same timeframe in relation the rest of the year. A relatively low correlation between snowfall and accumulation indicates that other factors like compaction, latent heat flux contribution, and the redistribution of snow by wind are also important contributors to the seasonal variability of solid accumulation. These factors combine to give a smaller amplitude seasonal cycle for accumulation relative to snowfall.

The observed liquid equivalent snowfall, when combined with estimated latent heat processes (i.e., deposition) and characteristic snow pack density, is substantially smaller than the observed surface accumulation. Some portion of this difference may be due to underestimates of the total snowfall by the POSS/MMCR retrievals, particularly for light snowfall. However, this uncertainty alone may not be enough to explain the disagreement in mass. It is possible that redistribution of snow by wind near the surface also contributes a
significant amount of mass. While studies on drifting snow across the GIS are few, the background snowfall and accumulation gradient across Greenland and the prevailing wind direction at Summit make this explanation plausible. Additionally, the broader drifting effects of the Summit camp may contribute to enhanced local effects on the net accumulation. Further investigation is needed to explore this potential wind and drifting effect. Despite the discrepancy between the ground-based remote sensors and in-situ accumulation measurements, the POSS and MMCR provide snowfall data at a much higher temporal resolution, something that is required for linking snowfall with system processes and meteorological events. Moreover, after further studies, the instantaneous measurements could be “calibrated” in time using the accumulation data to provide a more robust data set. To help address these issues, the suite of ICECAPS precipitation instruments will be expanded in the summer of 2014 to include important new perspectives from a hot plate precipitation gauge (Rasmussen et al. 2011) and a Multi-Angle Snowflake Camera (MASC; Garrett et al. 2012).

Snowfall at Summit, including occurrence, total, and rate, was found to be closely linked with local wind direction. Outside of the summer months, little snowfall accumulates when wind directions do not contain a southerly component. This relationship is believed to be linked to the seasonally-changing sea ice coverage of the area immediately surrounding Greenland. Likewise, a dependence of snowfall on PWV is seen. Snowfall occurrence and rates are drastically higher when PWV exceeds the monthly mean value, consistent with snowfall being linked to moist air masses that are influenced by nearby, ice-free ocean. LWP is shown to have little relation to snowfall occurrence, with most snowfall occurring under conditions with little to no LWP. However, snowfall generally increases with temperature, perhaps related to the increase in moisture availability. Interestingly, at any temperature, the
LWP is generally smaller during snowfall than when snow is not occurring. Clearly, the ice-phase precipitation processes are quite important for snowfall at Summit.

While the meteorological connections and snowfall analysis discussed here are local to Summit, they may offer broader insights into Greenlandic mass balance processes. Future work can utilize this snowfall data set, within the context of other ICECAPS instruments and collaborative projects at Summit, to develop a process-level understanding of precipitation as it relates to clouds, isotopic signatures, atmospheric stability, and others. These measurements can also be linked with satellite observations and/or model reanalysis products to examine and quantify spatial precipitation patterns impacting the central Greenland Ice Sheet.

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