THE EFFECTS OF SUMMER SNOWFALL ON ARCTIC SEA ICE RADIATIVE FORCING

By
Hannah Chapman-Dutton, B.A.

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APPROVED:

Melinda Webster, Committee Chair
Matthew Sturm, Committee Member
Thomas Ballinger, Committee Member
Simon Zweiback, Committee Member
Bernard Coakley, Chair
Department of Geosciences
Karsten Hueffer, Dean
College of Natural Sciences and Mathematics
Richard Collins, Director
Graduate School
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Abstract

The decline in Arctic sea ice has had major impacts on the climate system, particularly relating to the ice-albedo feedback. Since fresh snow on top of bare or melting sea ice increases the surface albedo on local scales, the impact of summer snow events can have a negative radiative forcing effect, which could inhibit sea ice surface melt.

In this study, we compared snow depth and meteorological data from buoys and satellite retrievals of surface and atmospheric conditions to identify and characterize summer snow accumulation case studies across the Arctic from 2003 to 2017. Clouds and Earth’s Radiant Energy System (CERES) retrievals were used to quantify the changes in surface albedo before and after the snow accumulation events. Information from these case studies was then scaled up to find similar events on a pan-Arctic scale using a Lagrangian sea ice parcel database. In this way, we characterized the frequency, magnitude, and duration of summer snow accumulation events similar to those observed by buoys. Finally, a simple radiative transfer model was used to quantify the impact of summer snowfall events on the surface and top-of-atmosphere radiative forcing over the entire Arctic region.

The following work provides new information on observed snow accumulation events over Arctic sea ice in summer by combining multiple sources of in situ, satellite, and modeled data. Such results will be particularly useful in understanding the impacts of ephemeral summer weather on surface albedo and their propagating effects on the radiative forcing over Arctic sea ice.
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General Introduction

Sea ice is simply frozen seawater, but it plays a major role in moderating earth’s climate. It forms a physical boundary between the ocean and atmosphere which, in the sunlit season, reflects light away from the earth’s surface and prevents that light from heating the upper ocean while also cooling the local atmosphere. The decline in Arctic sea ice extent, thickness, and age throughout the last century is well documented, but the greatest losses have occurred in the last two decades (Cavaliere & Parkinson, 2012; Stroeve & Notz, 2018). As increasing temperatures across the Arctic extend the summer melt season, sea ice now has a shorter window in which to form, which has caused a change from a multiyear ice-dominated regime to one that is increasingly seasonal (Markus et al., 2009; Stroeve et al., 2014).

Significant changes to Arctic weather patterns and precipitation have also been noted, with some uncertainty around how meteorological conditions may change in the future. In general, warmer temperatures allow for more water vapor retention within the atmosphere. This leads to an increase in total precipitation and overall likelihood of heavy precipitation events, while also amplifying the greenhouse effect which causes surface warming (Boisvert & Stroeve, 2015). Under these conditions, Bintanja et al. (2020) predict that interannual precipitation variability will increase in some areas of the Arctic by up to 50% throughout the 21st century. As warming continues, Arctic precipitation rates will continue to increase, with a predicted shift away from snow and towards rain-dominated summers (Boisvert et al., 2023; Bintanja & Andry, 2017). The combination of warmer temperatures and greater rainfall is very likely to decrease the summer sea ice thickness and extent. Therefore, it becomes increasingly important to understand past Arctic precipitation patterns, both rain and snow, so that we can make better predictions for how the Arctic will function in the future.

For most of the year, below-freezing temperatures limit precipitation to snowfall. However, freezing conditions can still occur during the summer months and allow for snow accumulation on the sea ice and on refrozen melt ponds. In general, spring and summer snow depth in the western Arctic is currently decreasing, particularly in the Chukchi and Beaufort regions (Webster et al., 2014) and climate models predict that this trend in snowpack thinning will continue (Webster et al., 2021). As Arctic warming continues, most models also project a
decrease in summer snowfall across the Arctic (Lique et al., 2016; Bintanja & Andry, 2017) which could have major implications for the thickness and survivability of the sea ice.
Chapter 1 Research Results

1.1 Background and Motivation

Fresh snow is one of the most reflective natural materials on earth (Warren, 2019). In the Arctic, bare sea ice has a fairly high albedo ($\alpha$) of 0.6-0.7, but the albedo of fresh snow can be as high as 0.9 (Perovich, Nghiem, et al., 2007; Perovich & Polashenski, 2012; Light et al., 2022). A typical ice-dominated region in the summer months may include many combinations of ice and melt ponds ($\alpha = 0.12-0.32$) on which the presence of snow can significantly increase the overall albedo of the given system (Figures 1.1 & 1.2). During the peak of summer, undeformed seasonal ice can have over 50% of its surface area covered by melt ponds (Scharien & Yackel, 2005; Polashenski et al., 2012), creating a complex icescape for snow to potentially accumulate on. As multiyear ice cover decreases and smoother seasonal ice becomes dominant across the Arctic, we could see a 38% increase in solar heat input to the ice layer and subsequently, to the upper ocean (Perovich & Polashenski, 2012). This extra heat would have major implications for sea ice mass balance as well as for ocean circulation and primary productivity in the Arctic Ocean. However, if fresh snow accumulates on top of sea ice in summer, the solar absorption within sea ice and the ocean could be strongly mitigated.

![Figure 1.1](image-url)
Relatively little is known about summer snowfall on Arctic sea ice. Most research regarding snow on sea ice relates to the winter months, but a few studies have documented individual summer snowfall events and the significant effect that they can have on the progression of ice melt (Perovich et al., 2002; Perovich et al., 2017, Light et al., 2015; Lim et al., 2022). The 19 June 1998 snowfall during the Surface Heat Budget of the Arctic Ocean (SHEBA) expedition is a well-documented example of such an event during which 4 cm of fresh snow caused an increase in albedo from 0.2 to 0.8 over areas of bare sea ice and refrozen ponds before returning to pre-snow values two days later (Perovich et al., 2002). In a different event described in Perovich et al. (2017), an early June snowfall likely delayed the ice surface melt onset by several weeks. While the albedo impact of this snowfall was relatively small and short-lived (increasing from 0.68 to 0.77), the latent heat associated with melting the new snow caused the
snow layer to act as a negative heat flux and therefore protect the underlying ice from surface melt.

A 2022 study by Lim et al. used satellite and reanalysis products to investigate summer snowstorms in the North Atlantic and Eurasian-Pacific regions between 1980 and 2019. In their definition, a snowstorm occurs “when the daily snowfall averaged over the Eurasian-Pacific sector of the Arctic (80–240°E; 69–90°N) exceeds 0.75 standard deviations for 3 or more consecutive days”. The results showed that the 95-96 storms identified increased the areal albedo by 0.01-0.02, which was associated with a reduction in shortwave absorption of around 2 W/m². Lim et al. (2022) found that snowstorms are also associated with about a week of below-freezing surface air temperatures, which allowed snow to persist. These cold temperatures were the main driver behind a positive sea ice extent anomaly of $0.12 \times 10^6$ km² associated with snowstorms. Even so, the snowfall itself explained about 25% of the ice extent anomaly, with the remaining effect relating to cold air temperatures.

As a result of the sparsity of in situ observations, little is known about the frequency of summer snow-on-ice events—how much it snows, how long the snow cover lasts, and the impact of these events on the energy balance. This study looks to address these knowledge gaps by investigating in situ, model, and remotely sensed data between 2003 and 2017. After characterizing summer snow accumulation events at a local scale and using remote sensing data to scale up across the Arctic, we determined the radiative forcing impact of such events on the Arctic as a whole. These results may provide a baseline understanding from which to gauge changes in a future climate.

1.2 Objectives

We approached the study of summer snow on Arctic sea ice with the objective of answering the question: how do summer snowfall events impact the surface albedo of Arctic sea ice? Within this broader question, we also investigated the following:

- How much does albedo change during snow accumulation in the summer?
- What proportion of albedo change can be attributed to snowfall as opposed to increased ice concentration or melt ponds refreezing?
• What effect does the thickness of the snow layer have on the magnitude and duration of the albedo change?
• How frequent and impactful are snow accumulation events in different regions of the Arctic?
• What is the radiative forcing associated with summer snow accumulation events?

1.3 Data

This analysis focuses on sea ice, snow, and atmospheric data from 2003 through 2017 to investigate the impacts of snow accumulation on Arctic summer sea ice. To do this, a variety of complementary data sets are used, ranging from direct surface measurements to satellite and model data. Each data type used in the analysis is described below and listed in Table 1.1.

<table>
<thead>
<tr>
<th>Data Product</th>
<th>Key Variables</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sea Ice Mass-Balance Buoys</td>
<td>Snow depth (m)</td>
<td><a href="http://imb-crrel-dartmouth.org">http://imb-crrel-dartmouth.org</a> (Perovich et al., 2023)</td>
</tr>
<tr>
<td></td>
<td>Air temperature (C)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Air pressure (hPa)</td>
<td></td>
</tr>
<tr>
<td>Ice-Parcel Database</td>
<td>SnowModel-LG (ERA5) Snow Depth (m)</td>
<td>Liston et al., 2020</td>
</tr>
<tr>
<td>(Horvath et al., 2023)</td>
<td>NSIDC Sea Ice Concentration (%)</td>
<td>Fetterer et al., 2017</td>
</tr>
<tr>
<td></td>
<td>ERA5 Surface Air Temperature (C)</td>
<td>Hersbach et al., 2023</td>
</tr>
<tr>
<td></td>
<td>ERA5 850hPa Air Temperature (C)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>CERES Surface Albedo</td>
<td>Wielicki et al., 1996</td>
</tr>
<tr>
<td>MODIS (MOD09GA)</td>
<td>500m Daily Surface Reflectance (unitless)</td>
<td>Vermote et al., 2015</td>
</tr>
<tr>
<td>Melt Pond Distribution on Sea Ice</td>
<td>Melt Pond Fraction (%)</td>
<td>Rösel et al., 2015</td>
</tr>
</tbody>
</table>

1.3.1 Local Scale: Ice Mass Balance Buoys

We began our study of summer snow accumulation at the local scale with in situ observations of snow depth, air temperature, and surface air pressure. The Cold Regions Research and Engineering Laboratory (CRREL)-Dartmouth Ice Mass Balance Buoys (IMBs) (Perovich et al., 2023) provide the in situ data for this project, with a time-series beginning in 1997 and extending through 2018. While the specific instrumentation can vary, the basic buoy sensors used include an air temperature probe, thermistor string, electric barometer for sea level
pressure, and two acoustic sounders facing upwards and downwards to measure snow depth and ice draft. The sounders can measure snow depth and ice draft changes down to 5 mm, but have an inherent uncertainty of about 1 cm (Perovich et al., 2023). Buoys are most often deployed in the Central Arctic and the Beaufort Sea and are left to drift with the sea ice. Data are recorded every one or four hours, but have been averaged to daily values for the purposes of this study.

1.3.2 Pan-Arctic Scale: Ice Parcel Database

We then expanded the analysis to the pan-Arctic scale to understand the larger scale effects of snow accumulation events on sea ice using remote sensing, reanalysis, and model data. The Ice-Parcel Database (Horvath et al., 2023) takes a Lagrangian approach to tracking sea ice as it drifts throughout the Arctic. Each 25-km by 25-km parcel represents a grid cell with an ice concentration of at least 15%. The parcels are tracked from October through to the end of September each year from 2002 to 2020. The database was created by combining multiple sources of satellite, model, and reanalysis data and includes a wide range of information about each ice parcel, including radiative forcing and energy balance terms. The primary variables used in this analysis are the Clouds and Earth’s Radiant Energy System (CERES) albedo and energy balance products (Wielicki et al., 1996), National Snow and Ice Data Center (NSIDC) sea ice concentration (Fetterer et al., 2017), surface and 850 hPA air temperature from the ERA5 reanalysis product (Hersbach et al., 2023), and snow depth values modeled by SnowModel-LG (Liston et al., 2020) forced with ERA-5 reanalysis. All incorporated data have been re-gridded to a 25-km resolution.

1.4 Methods

Using local to pan-Arctic scale data sets, we investigated the effects of snowfall on surface albedo and radiative forcing for initially snow-free conditions over Arctic sea ice in the summer months (May-August). Using the buoy snow depth and SnowModel-LG (Liston et al., 2020), we identified distinct snow accumulation events across the Arctic based on the following criteria: (1) initial nearly snow-free conditions (one centimeter or less), (2) at least one centimeter of new snow accumulation, and (3) a subsequent melt period that reduces the snow depth to one centimeter or less. This means that our “events” occur on the ice surface at a single point or ice parcel and do not represent a larger storm or weather system.
1.4.1 Local Scale

Case studies from the IMB buoy records serve as an in-situ baseline for this study. Out of the sixty-four buoys evaluated, sixteen summer snow accumulation events were identified. Seven of the buoys recorded the same storm as another buoy. Thus, there is a total of twelve distinct weather events resulting in snow accumulation over nearly snow-free conditions at the buoys. Specifics of each case study can be found in Table 1.2. For each selection, the total snow accumulation was determined, as well as the number of days that passed before the snow melted completely or returned to the initial depth of 1 cm or less. In some cases when a snow accumulation event occurred near the end of the summer season, it formed the first layer of the winter snowpack and remained through the end of the year. These cases were not included in the calculation of the average duration of snow accumulation events.

Table 1.2: IMB Buoy snow accumulation events used in the case study analysis. Lines grouped together represent buoys which experienced the same large-scale weather system. Durations marked with * did not melt below 1 cm before the end of August and became the base of the winter snowpack for that year.

<table>
<thead>
<tr>
<th>Buoy Name</th>
<th>Region</th>
<th>Snowfall Start</th>
<th>Snowfall End</th>
<th>Accumulation (cm)</th>
<th>Duration</th>
</tr>
</thead>
<tbody>
<tr>
<td>2013B</td>
<td>Central</td>
<td>2 Jun 2013</td>
<td>7 Jun 2013</td>
<td>10.7</td>
<td>9 Jul 2013</td>
</tr>
</tbody>
</table>
Since these buoys only report snow depth, ice thickness, air temperature, and surface air pressure, data from the ice parcel database can be used to comprehensively characterize each snow event. While the buoys do not line up exactly with the parcel drift track, the nearest parcel was collocated with each buoy. With this combination of data, each case study was analyzed to determine the albedo response to the new snow, changes in total precipitation and atmospheric conditions (pressure, temperature, wind speeds), and to compare the in-situ snow accumulation amount and duration with that from the model.

1.4.2 Pan-Arctic Scale

We used the case study results to expand the analysis to the pan-Arctic scale to learn how often such events occur, how long the new snow cover lasts, what areal coverage the events have, and what impacts the new snow has on surface albedo. A snow accumulation event is defined as an increase in snow depth of any amount from an initial depth of one centimeter or less and extending until the snow melts below one centimeter of depth. Events that showed an accumulation of three centimeters or more were classified as being optically thick based on the results in Figure 1.3 and data from Perovich (2007). For all events, we analyzed the atmospheric characteristics and ice properties as well as the duration of the event. We divided the parcels into regions (Figure 1.3, Meier et al., 2007) and did further analysis to determine the geographic differences of snow accumulation events across the regions of the Arctic. Southerly regions such as the Bering and Okhotsk seas, Baffin Bay, Hudson Bay, and the Gulf of St. Lawrence were not included.

1.4.3 Radiative Forcing Impacts

We used a simple two-layer reflectance model (Thomas & Stamnes, 1999, Section 6.11) to compute the change in the top-of-the-atmosphere (TOA) radiative forcing resulting from the addition of a fresh snow layer on top of melting, bare sea ice. In this way, we determined how much the TOA albedo changed after summer snowfall events, and by how much the new albedo increased the amount of radiation reflected back into space during the summer melt season.
The methodology presented here largely follows those presented in Webster & Warren (2022); specifically, we use the same two-layer multiple reflection model (Thomas & Stamnes, 1999, Section 6.11) and the following assumptions: (1) the Arctic climatological cloud cover is 80% in space and time (Environmental Working Group, 2000), (2) the monthly climatological TOA solar irradiance follows those in Hartmann (2016) and is weighted by 70-90°N latitude, (3) the ice concentration values have a ±15% accuracy for the summer months (Meier et al., 2021), (4) albedo values for different surface types (cold thin snow; bare melting sea ice) under clear and cloudy sky conditions remain static (Light et al. (2022), Perovich, Nghiem et al. (2007), Perovich & Polashenski (2012)), and (5) the area of the Arctic Ocean can be defined by the Meier et al. (2007) region mask. Values for these conditions and variables, as well as other details on the methodology, can be found in Webster and Warren (2022) and are additionally listed in Table 1.3. Further explanation of the optical values can be found in the Appendix.
Table 1.3: Variables and values used in calculating the radiative forcing impact of summer snow accumulation events. The area of the Arctic Ocean is calculated from the Meier et al. (2007) region mask.

<table>
<thead>
<tr>
<th></th>
<th>Clear Sky</th>
<th>Cloudy Sky</th>
</tr>
</thead>
<tbody>
<tr>
<td>(a_{\text{snow}})</td>
<td>Fresh Snow Albedo</td>
<td>0.8</td>
</tr>
<tr>
<td>(t_{\text{snow}})</td>
<td>Fresh Snow Transmittance</td>
<td>0.027</td>
</tr>
<tr>
<td>(a_{\text{ice}})</td>
<td>Bare Ice Albedo</td>
<td>0.58</td>
</tr>
<tr>
<td>(a_{\text{atm}})</td>
<td>Atmospheric Albedo</td>
<td>0.14</td>
</tr>
<tr>
<td>(t_{\text{atm}})</td>
<td>Atmospheric Transmittance</td>
<td>0.76</td>
</tr>
<tr>
<td>(A_{AO})</td>
<td>Area of the Arctic Ocean</td>
<td>(1.71 \times 10^7) km(^2)</td>
</tr>
</tbody>
</table>

To determine the TOA albedo change, broadband albedo values for two surface conditions (thin, cold snow vs. bare, melting ice) are first considered in clear vs. cloudy conditions. Equation 1a (Figure 1.4) is the two-layer multiple reflection model for the system (combined) albedo of fresh, thin snow (with an albedo, \(a_{\text{snow}}\), and transmittance, \(t_{\text{snow}}\)) on top of bare ice (\(a_{\text{ice}}\)). This surface system albedo (\(a_{\text{surf}}\)) is calculated using the corresponding albedo and transmittance values of snow and ice (Table 1.3) for both clear and cloudy conditions separately. We then repeat the calculation using the surface system albedo (\(a_{\text{surf}}\)) as the lower layer and the atmospheric albedo (\(a_{\text{atm}}\)) and transmittance (\(t_{\text{atm}}\)) as the upper layer (Figure 1.4) (Eq. 1b) to determine the TOA system albedo (\(a_{\text{sys}}\)).

\[
\begin{align*}
\text{Eq. (1a)} & \quad a_{\text{surf}} = a_{\text{snow}} + \frac{a_{\text{ice}} t_{\text{snow}}^2}{1-a_{\text{ice}} a_{\text{snow}}} \\
\text{Eq. (1b)} & \quad a_{\text{sys}} = a_{\text{atm}} + \frac{a_{\text{surf}} t_{\text{atm}}^2}{1-a_{\text{surf}} a_{\text{atm}}}
\end{align*}
\]

Using the TOA system albedo of fresh snow on bare ice (\(a_{\text{sys\_clear}}\), \(a_{\text{sys\_cloudy}}\)), we calculated the change in albedo when going from bare ice to full snow cover under both sky conditions.

\[
\begin{align*}
\Delta a_{\text{sys\_clear}} & = a_{\text{sys\_clear}} - a_{\text{ice\_clear}} \quad \text{Eq. (2a)} \\
\Delta a_{\text{sys\_cloudy}} & = a_{\text{sys\_cloudy}} - a_{\text{ice\_cloudy}} \quad \text{Eq. (2b)}
\end{align*}
\]
Since the Arctic is frequently cloudy during the summer, we weighted the system albedo to appropriately incorporate clear and cloudy sky conditions. To do this, we assume that the sky is cloud-covered 80% of the time and clear the remaining 20% (Webster & Warren, 2022). We then calculate a weighted average of the TOA albedo change ($\Delta a_T$) from bare ice to snow-covered ice under both sky conditions (Eq. 3).

$$\Delta a_T = (0.2 \cdot \Delta a_{sys,clear}) + (0.8 \cdot \Delta a_{sys,cloudy}) \quad \text{Eq. (3)}$$

To determine the radiative forcing related to snow on sea ice across the Arctic Ocean region, we used the results from the Lagrangian ice parcel database to account for the average number of parcels experiencing snow accumulation events, the duration of the events, and the areal coverage of events. We used the:

- TOA albedo change associated with a snowfall event ($\Delta a$) determined above,
- the monthly average downward TOA solar flux ($F_m$) determined from climatological solar irradiance data from Hartmann (2016) weighted by 70-90°N latitude,
- the average number of events per month ($N_m$) for 2003-2017,
- the average number of parcels that exist during each month ($P_m$) for 2003-2017, and
- the average monthly duration of snow accumulation events ($D_m$) for 2003-2017.
To determine the monthly areal coverage of events, we used the average number of parcels experiencing events \( A_m \) and multiply that number by the sea ice concentration \( SIC_m \) of those ice parcels. We then divided the result by the total area of the Arctic Ocean \( A_{AO} \) to get an Arctic domain average.

We computed the monthly radiative forcing, summed the values across all months and divided by 12 to get the annual average change in TOA radiative forcing associated with summer snow events \( \Delta RF_{ArcticOcean} \). This result reveals the impact that summer snow events have on an Arctic-wide scale.

\[
\Delta RF_{ArcticOcean} = \frac{1}{12} \sum_{m=1}^{12} \Delta a \cdot F_m \cdot \frac{D_m}{30.5} \cdot N_m \cdot \left( \frac{A_m \cdot SIC_m}{A_{AO}} \right)
\]

Eq. (4)

1.5 Results

The approach for this study is broken down into three sections. First, Section 1.5.1 looks at case examples of summer snow accumulation recorded by IMB buoys on a local scale. This provides a source of direct measurements to compare against when looking at other environmental variables from the ice parcel database. Section 1.5.2 expands these events to the pan-Arctic scale using satellite, model, and reanalysis data from the Lagrangian ice parcel database to characterize summer snowfall patterns and the impacts of such events on surface albedo. Finally, Section 1.5.3 uses the areal coverage and duration results from the previous sections to quantify the potential radiative forcing effects of summer snowfall on sea ice by combining reanalysis and satellite data with in-situ reflectance measurements of snow and sea ice.

1.5.1 Local Scale

First, we present the details of one case study from the 2008E buoy and its corresponding ice parcel as an example of the analysis applied to all case studies (Section 1.5.1.1). Following are the collective results from all case studies and corresponding ice parcels (Section 1.5.1.2).

1.5.1.1 Case Study: IMB 2008E

Between 8 July and 14 July 2008, 6.4 cm of snow accumulated at the 2008E buoy north of Greenland. Based on the IMB data, this snow cover persisted until 23 July. The greatest snowfall occurred on 12 July according to the simulated snow depths, which aligns with the buoy
While the snow depth increased, there were also distinct increases in the local albedo and ice concentration values. A further look into the event conditions show buoy temperatures below freezing, high winds, and a drop in sea level pressure (Figure 1.6), all of which point to stormy weather.

In this case, the snow accumulation event was the result of a nearby cyclone, as identified in the ice parcel database. Cyclones are often associated with high winds, which can compress the sea ice and close leads through convergence. Accordingly, an albedo increase during a snowfall event may not be solely due to fresh snow accumulation but could also result from changes in sea-ice coverage. To explore this topic further, we examined the corresponding 500-m resolution MODIS reflectance data (Vermote et al., 2015) over the same spatial domain before and after the cyclone event which shows an overall increase in the surface reflectance distribution after the snow accumulation event (Figure 1.7). Furthermore, freezing temperatures persisted throughout and after the snowfall event, which likely contributed to the observed decrease in melt pond coverage. Combined, these factors indicate a widespread coverage of new snow accumulation, which likely led to the positive shift in the reflectance distribution of the ice parcel domain.
Figure 1.6: Surface air temperature, wind speed, and surface air pressure from the Ice Parcel database and IMB 2008E from June through August with the snow accumulation event (see Table 1.2) shaded in gray.

Figure 1.7: MODIS Daily surface reflectance values before the 2008E event (top left) and after the event (top right). Bottom panel shows the pixel reflectance distribution for both scenarios.
1.5.1.2 Background Buoy Data and Case Studies

While the in-depth case study of Buoy 2008E provides a detailed look at a specific snow accumulation event, the full set of buoy data used in this analysis (1993-2017) contains valuable information about environmental conditions and sea-ice state over longer time scales and at different points across the Arctic. Based on all buoy data, the boxplots in Figure 1.8 show the following characteristics: (1) snow depth remains fairly stable throughout the year with only a noticeable drop in the summer months, particularly July and August; (2) sea ice thickness steadily increases throughout the winter and declines in late summer, with the greatest variability in thickness occurring in September; (3) surface air temperatures remain consistently below 0°C all year except during the summer when temperatures stay close to 0°C or slightly above; and (4) surface air pressure is highest in February and lowest during the summer months. Each season shown here contains a relatively equal number of buoys with 50 active buoys from December to February, 57 from March to May, 57 from June to August, and 55 from September to November. During the summer, there were 26 active buoys in June, 16 in July and 30 in August.

The complete IMB data set is also useful for quantifying how much snow accumulates per day on average. Table 1.4 shows the average amount of positive snow depth change per day across the full buoy record as well as the standard deviation. Interestingly, while the overall snow depth shows a major decline in June, the variability in average depth change per day is much higher than May, June, and August (3.9 cm in July vs 2.2-2.6 cm in other months). When only considering the case study events, the average daily accumulation is much higher, at 6.5 cm per day (standard deviation = 3.5) while the monthly average snow depths are also considerably higher than the overall average (May: N/A (no events identified), June: 7.3 cm, July: 5.5 cm, August: 6.8 cm).
The buoy case studies were combined with the corresponding ice parcels in space and time to get a sense of each snow accumulation event’s environmental conditions and effects on albedo. Of the events that occurred, the buoy data show that the average observed snow accumulation occurred over 5.4 days and the snow lasted for an average of 13.7 days after the first snowfall. The mean snow accumulation amount for the events was 6.5 cm (standard deviation: 3.5 cm). Each buoy value was correlated with a selection of ice parcel parameters using a Pearson correlation coefficient, shown in Figure 1.9.

In an ideal world, the buoy and parcel data would be strongly correlated during snow accumulation events; however, the results show a poor relationship between the two data sets. The main reason for this is likely the difference in scale. While the buoys represent a single point
measurement, each corresponding ice parcel covers a 25-km-by-25-km grid cell. Calculating a single average value over such a large area will inherently erase much of the local-scale heterogeneity that dominates the buoy measurements. Despite this, some relationships are worth noting. The albedo values are highly correlated with both ice concentration and SnowModel snow depth, which indicates the importance of accounting for changes in ice concentration when considering the albedo response to changes in the snowpack.

Table 1.4: Mean snow depth and daily snow depth increase across all IMB buoys from May to August (1993-2017). Standard Deviations included in parentheses.

<table>
<thead>
<tr>
<th>Month</th>
<th>Mean Snow Depth [cm]</th>
<th>Mean Daily Snow Depth Increase [cm]</th>
</tr>
</thead>
<tbody>
<tr>
<td>May</td>
<td>24.7 cm (13.6)</td>
<td>2.2 cm (1.8)</td>
</tr>
<tr>
<td>June</td>
<td>20.4 cm (13.9)</td>
<td>2.4 cm (2.5)</td>
</tr>
<tr>
<td>July</td>
<td>4.5 cm (13.4)</td>
<td>3.9 cm (6.4)</td>
</tr>
<tr>
<td>August</td>
<td>2.7 cm (8.4)</td>
<td>2.6 cm (3.6)</td>
</tr>
<tr>
<td>Total</td>
<td>13.1 cm</td>
<td>2.8 cm</td>
</tr>
</tbody>
</table>

Figure 1.10 shows the same relationships but only includes data from the days in which the buoy snow depth increased rather than the full snow cover duration. These are important to consider since precipitation and snowfall should increase at the same time as snow depth, but the snow will remain on the ground after the precipitation ends, therefore leading to a break-down in correlation when all dates are included. This change allows for a few strong anti-correlations to emerge within the buoy data. For example, as the date progresses through the summer months, the albedo and ice concentrations decline. Interestingly, reanalysis surface temperatures also appear to decrease as the date increases. Since this matrix only shows dates during which snow accumulated, there is the potential that the events later in the summer experienced more extreme cooling than those earlier in the season, despite overall temperatures increasing during the early-to-mid melt season.
Figure 1.9: Pearson correlation values across all days during the case study snow accumulation events for the IMBs and their corresponding ice parcels. Black dots represent p-values significant at the 0.05 level. These results include cases where snow did not melt before the end of August.

In short, these results uphold the current understanding of the Arctic climate system in which summer precipitation is associated with low pressure systems and cool atmospheric temperatures. Increases in snow depth often lead to higher albedo while warming temperatures through the summer lead to decreases in albedo, snow depth, and ice concentration. This local-scale analysis provides a foundation for using the ice parcel data independent from in-situ buoys in order to learn about such snow events on larger spatial scales.
Figure 1.10: Pearson correlation values for case study IMBs and corresponding ice parcels using only days when snow depth increased at the buoy level. Black dots represent p-values significant at the 0.05 level. The results include cases where snow did not melt before the end of August.

1.5.2 Pan-Arctic Scale Results

Using the ice parcel database, we characterized the frequency, spatial extent, albedo change, and snow conditions associated with each snow accumulation event over the 2003-2017 period. In general, events occur most frequently in the East Siberian, Chukchi, and Beaufort Seas with upwards of 6 events per year on average compared to 1 or fewer across most of the Greenland and Barents Seas (Figure 1.11). In the Pacific and Central Arctic regions, most events occur in July and early August, while the Atlantic regions (Barents, Greenland, Kara, Labrador, and Laptev) experience events at a consistent rate throughout the summer. The relative lack of events in the Atlantic regions is likely a result of generally higher temperatures which lead to precipitation in the form of rain and earlier ice loss. Note that, while snow could be falling at the
same rate in all regions, the lower ice concentration in one region would prevent high fractional coverage of fresh snow accumulation.

![Event density map showing where snow accumulation events occurred between 2003 and 2017 based on SnowModel-LG results from the ice parcel database.]

Figure 1.11: Event density map showing where snow accumulation events occurred between 2003 and 2017 based on SnowModel-LG results from the ice parcel database.

We often assume that snow can only fall when temperatures are below freezing, despite snow accumulation in above freezing temperatures being possible in the Arctic. To address this discrepancy, we examined the temperatures at two meters and at 850 hPa, as well as the albedo change, during the snow accumulation events. Figure 1.12 compares the distribution of temperatures at the surface and at the 850 hPa level, which demonstrates the high number of above-freezing temperature events that occur across the Arctic. At the 850 hPa level, 60% of the snow accumulation events that had temperatures above 0°C showed an increase of 0.06 in albedo (Figure 1.12A). In comparison with the surface level, the albedo increased 98% of the time, with an average increase of 0.07 when surface temperatures were above freezing (Figure 1.12B). Through July and August, surface temperatures during events remained fairly similar to temperatures outside of events, and rarely warmed above 0°C.
Next, we investigated the pan-Arctic changes in albedo resulting from the snow accumulation events. Throughout the study period, the average albedo at Day 0 of an event (snow depth of 1 cm or less) was 0.47 and the day with the deepest snow during the event had an average albedo of 0.49, an increase of 0.02 (Table 1.5). When considering the lowest and highest albedo during the event, not tied to the thinnest or thickest snow depth, the average albedo ranged from 0.44 to 0.52, an increase of 0.08.

When comparing the average event albedo by region (Figure 1.13A), some interesting geographic differences appear. The Greenland and Barents Seas show much lower albedos than any other region, while those in the Central Arctic are highest. Other regions have comparable albedos, being close to the 0.50 average of all events. Interestingly, the same regions with the lowest average albedo also experience the greatest increase in albedo from snow accumulation events (Figure 1.13B). Since relatively few events occur in the Greenland and Barents regions (Figure 1.11) despite lying within the North Atlantic storm track region, we determine that their low average albedo is a result of low ice concentrations and high rates of surface melt. More simply, even if many storm systems occur, there is less ice surface for the snow to accumulate on than other regions. When conditions are favorable for accumulation, however, the fresh snow can lead to larger temporary increases in albedo than are seen in regions where the underlying ice surface is consistently more continuous.
Table 1.5: Average albedo during events. Left two columns show albedo on the date with the minimum amount of snow and the maximum snow. Right columns show minimum and maximum albedo during the event, regardless of snow depth.

<table>
<thead>
<tr>
<th>Year</th>
<th>Day of Min Snow</th>
<th>Day of Max Snow</th>
<th>Min During Event</th>
<th>Max During Event</th>
</tr>
</thead>
<tbody>
<tr>
<td>2003</td>
<td>0.49</td>
<td>0.51</td>
<td>0.46</td>
<td>0.54</td>
</tr>
<tr>
<td>2004</td>
<td>0.50</td>
<td>0.52</td>
<td>0.47</td>
<td>0.55</td>
</tr>
<tr>
<td>2005</td>
<td>0.47</td>
<td>0.49</td>
<td>0.44</td>
<td>0.51</td>
</tr>
<tr>
<td>2006</td>
<td>0.49</td>
<td>0.51</td>
<td>0.47</td>
<td>0.53</td>
</tr>
<tr>
<td>2007</td>
<td>0.48</td>
<td>0.51</td>
<td>0.46</td>
<td>0.53</td>
</tr>
<tr>
<td>2008</td>
<td>0.48</td>
<td>0.49</td>
<td>0.45</td>
<td>0.52</td>
</tr>
<tr>
<td>2009</td>
<td>0.47</td>
<td>0.49</td>
<td>0.44</td>
<td>0.51</td>
</tr>
<tr>
<td>2010</td>
<td>0.45</td>
<td>0.46</td>
<td>0.42</td>
<td>0.49</td>
</tr>
<tr>
<td>2011</td>
<td>0.39</td>
<td>0.39</td>
<td>0.36</td>
<td>0.42</td>
</tr>
<tr>
<td>2012</td>
<td>0.45</td>
<td>0.46</td>
<td>0.41</td>
<td>0.49</td>
</tr>
<tr>
<td>2013</td>
<td>0.48</td>
<td>0.50</td>
<td>0.45</td>
<td>0.53</td>
</tr>
<tr>
<td>2014</td>
<td>0.48</td>
<td>0.49</td>
<td>0.45</td>
<td>0.52</td>
</tr>
<tr>
<td>2015</td>
<td>0.49</td>
<td>0.50</td>
<td>0.46</td>
<td>0.53</td>
</tr>
<tr>
<td>2016</td>
<td>0.49</td>
<td>0.51</td>
<td>0.46</td>
<td>0.53</td>
</tr>
<tr>
<td>2017</td>
<td>0.49</td>
<td>0.51</td>
<td>0.46</td>
<td>0.53</td>
</tr>
<tr>
<td>Average</td>
<td>0.47</td>
<td>0.49</td>
<td>0.44</td>
<td>0.52</td>
</tr>
</tbody>
</table>

Figure 1.13: (A) The albedo for all snow accumulation events in each region over 2003-2017. (B) The average change in albedo during snow accumulation events in each region over 2003-2017.
To characterize the duration of events, we created two categories: thin snow accumulation events (maximum snow depth less than 3 cm) and thick snow accumulation events (maximum depth at least 3 cm). Table 1.6 shows the average duration of events each year. Overall, optically thick snow events lasted an average of 3.4 days longer than optically thin events. This is likely a result of the thicker snow layer taking longer to melt away, but it could also be related to other atmospheric conditions associated with enhanced snowfall, such as colder temperatures (i.e., Lim et al., 2022) or increased cloudiness (Perovich, 2018).

Table 1.6: Average duration (days) of events with thick snow covers (3 cm or greater) and thin snow covers (<3 cm).

<table>
<thead>
<tr>
<th>Year</th>
<th>Duration: Thick Snow Cover</th>
<th>Duration: Thin Snow Cover</th>
</tr>
</thead>
<tbody>
<tr>
<td>2003</td>
<td>5.3</td>
<td>2.8</td>
</tr>
<tr>
<td>2004</td>
<td>5.4</td>
<td>2.6</td>
</tr>
<tr>
<td>2005</td>
<td>5.7</td>
<td>2.6</td>
</tr>
<tr>
<td>2006</td>
<td>6.5</td>
<td>2.7</td>
</tr>
<tr>
<td>2007</td>
<td>7.6</td>
<td>2.7</td>
</tr>
<tr>
<td>2008</td>
<td>5.9</td>
<td>3.0</td>
</tr>
<tr>
<td>2009</td>
<td>7.2</td>
<td>3.1</td>
</tr>
<tr>
<td>2010</td>
<td>6.0</td>
<td>2.7</td>
</tr>
<tr>
<td>2011</td>
<td>5.9</td>
<td>3.1</td>
</tr>
<tr>
<td>2012</td>
<td>6.5</td>
<td>3.1</td>
</tr>
<tr>
<td>2013</td>
<td>7.1</td>
<td>3.1</td>
</tr>
<tr>
<td>2014</td>
<td>7.2</td>
<td>2.7</td>
</tr>
<tr>
<td>2015</td>
<td>6.7</td>
<td>2.6</td>
</tr>
<tr>
<td>2016</td>
<td>5.3</td>
<td>2.8</td>
</tr>
<tr>
<td>2017</td>
<td>5.9</td>
<td>3.0</td>
</tr>
<tr>
<td>Average</td>
<td>6.3</td>
<td>2.8</td>
</tr>
</tbody>
</table>

To investigate the relationships between the event duration and other variables, we computed the correlations between duration and other potentially important factors. Of albedo, snow depth, ice concentration, surface temperature, 850 hPa temperature, surface air pressure, and 10m wind speed, only snow depth showed a significant positive correlation, shown in detail in Figure 1.14. This suggests that the main driver in the relationship between snow depth and duration is the time required to melt a deeper snow layer. Similar findings hold when the parcels are grouped by region and averaged over the entire period.

Since snow depth plays an important role in the duration of the events, we looked further into the factors that may impact variations in snow depth. By comparing average event snow depth across regions, we expect to find deeper snow in the N. Atlantic regions since this area is known to have higher precipitation and snowfall rates than other regions of the Arctic. Figure 1.15 confirms this prediction. It is important to note that while the Atlantic and Eastern regions have a considerably deeper average snow depth, the Central and Pacific regions had far more events by
area. It is likely that the storms in the Atlantic region, which are generally larger and contain more moisture than those in the Central Arctic and Pacific regions, are more likely to precipitate rain rather than snow. While N. Atlantic snow events are less frequent, their greater size and intensity can produce considerably more snow than those in other regions.

Figure 1.14: Random sample of data representing the relationship between snow accumulation event duration and snow depth. Red line represents a polynomial trend line. Histogram (gray) shows the distribution of event durations.

Figure 1.15: Regional average snow depth during snow accumulation events.
1.5.3 Forcing Impacts

Using the two-layer albedo model of Thomas & Stamnes (1999, Section 6.11), we calculated the surface and TOA radiative effect of fresh snow on bare, melting ice. We first determined the system albedo of the surface with a fresh snow cover over bare ice under clear skies and cloudy skies, yielding a combined ice with snow surface albedo of 0.80 under clear skies and 0.86 under clouds. We repeated this calculation using the surface albedo as the lower layer and the atmosphere as the upper layer and found that the TOA albedo for the combined system is 0.57 for clear skies and 0.73 for cloudy skies. Assuming that the Arctic is cloudy for 80% of the summer as in Webster and Warren (2022), which is based on an Arctic cloud climatology, we calculated a weighted average albedo of 0.85 at the surface and 0.70 from TOA.

Next, based on the results in Section 1.5.2, we used the average spatial extent of the Arctic that experienced a snow accumulation event over the 15-year study period to determine the Arctic-wide radiative forcing effect. With these values and the climatological surface and TOA incident solar flux (Table 1.7), we compared the radiative forcing associated with snow accumulation events with that of bare ice for each month, which is shown in Table 1.7. Although peak insolation occurs in June, the greatest impact of summer snow accumulation events occurs in July, when the largest fraction of the Arctic sea ice is snow-free and dominated by bare, melting ice. The dampening effect of the cloud cover at TOA throughout the summer is highlighted in Figure 1.16, with the greatest values occurring in July and under clear skies in June. We also calculated an annual Arctic radiative forcing of \(-0.08\, \text{W m}^{-2}\).

Table 1.7: Average monthly values from snow accumulation events in 2003-2017 used in Equation 4 to calculate a total yearly Arctic radiative forcing.

<table>
<thead>
<tr>
<th></th>
<th>May</th>
<th>June</th>
<th>July</th>
<th>August</th>
</tr>
</thead>
<tbody>
<tr>
<td>(F)</td>
<td>TOA Solar Irradiance (W m(^{-2}))</td>
<td>412</td>
<td>498</td>
<td>460</td>
</tr>
<tr>
<td>(SIC)</td>
<td>Sea Ice Concentration (%)</td>
<td>69.6</td>
<td>78.5</td>
<td>72.3</td>
</tr>
<tr>
<td>(D)</td>
<td>Event Duration (days)</td>
<td>8.0</td>
<td>2.7</td>
<td>2.8</td>
</tr>
<tr>
<td>(A)</td>
<td>Event Area (km(^2))</td>
<td>(2.7\times10^5)</td>
<td>(10.2\times10^5)</td>
<td>(15.9\times10^5)</td>
</tr>
<tr>
<td>(N)</td>
<td>Number of Events</td>
<td>1,553.7</td>
<td>6,374.9</td>
<td>12,644.4</td>
</tr>
<tr>
<td>(RF_{\text{Arctic Ocean}})</td>
<td>Arctic Radiative Forcing (W m(^{-2}))</td>
<td>-0.02</td>
<td>-0.14</td>
<td>-0.47</td>
</tr>
</tbody>
</table>
Figure 1.16: The change in TOA radiative forcing in clear and cloudy sky conditions from snow accumulation events in 2003-2017.

1.6 Discussion

The results presented here can tell us much about the frequency, duration, and magnitude of summer snow accumulation events and the role they play in the Arctic energy balance. However, these analyses only account for surfaces that are completely covered by fresh snow on continuous sea ice. In reality, a situation in which there is no change in ice concentration or surface conditions during the events is highly unlikely. In Section 1.6.1, we investigate the impact of ice concentration and melt pond coverage on surface albedo. Section 1.6.2 compares this work with other studies of Arctic summer snow.

1.6.1 Alternate Sources of Albedo Change

From the case studies and ice parcel database, we found evidence that ice concentration could play a role in albedo changes during snow events. This also opened discussion for other factors besides snow which could change the measured albedo at an ice parcel. We decided to further investigate the role of melt ponds as well as that of ice concentration in order to better distinguish the different sources of albedo change.

In order to estimate the effects of sea ice convergence on albedo, we performed a back-of-the-envelope calculation of the aggregate-scale albedo based on measured albedos of four different surface conditions (bare sea ice, fresh snow, bare ice with melt ponds, and snow with
melt ponds), sea ice concentration, and melt pond fraction following Perovich, Light, et al. (2007). We began with the IMB case studies by calculating the expected albedo of the given parcel assuming only bare sea ice and open water were present using Equation 5a:

\[ a = (a_{ice} \times SIC) + (a_o \times (1 - SIC)) \]  \hspace{1cm} \text{Eq. (5a)}

where \( a \) represents the aggregate albedo of a given grid cell, \( a_{ice} \) is the bare sea ice albedo (0.65), \( a_o \) is the open water albedo (0.07), and SIC is the NSIDC ice concentration. The albedo values are based on measurements in Light et al. (2022). By replacing the bare sea ice albedo with that of fresh snow (0.85), we compared the observed albedo with the two calculated scenarios (bare ice with open water and snow with open water) shown in Figure 1.17A for the 2008E case study. In this case, we found that the expected albedo for bare ice was higher than the ice parcel albedo, implying that the ice surface had developed melt ponds, which are darker than bare sea ice. To investigate this further, we modified Equation 5a to include three surface types: bare sea ice, melt ponds, and open water, as shown in Equation 5b:

\[ a = a_{ice} \times (SIC - MPF) + (a_{mp} \times MPF) + (a_o \times (1 - SIC)) \]  \hspace{1cm} \text{Eq. (5b)}

where \( a_{mp} \) represents the albedo of melt ponds (0.22) based on Light et al. (2022), and MPF is the melt pond fraction of the ice area, integrated to daily values based on the 8-day composite product from Rösel et al. (2015). With this process, we calculated the expected albedo for a scenario containing bare ice with melt ponds and repeated again for fresh snow with melt ponds (Figure 1.17B).

From these results we conclude that during the 2008E case study, the ice was bare and melting throughout the month of June and had likely formed melt ponds throughout the second half of the month. During this particular snowfall event, the albedo increased to that of fresh snow with melt ponds before settling in the range of bare ice or ice with some surface melt after the event. While values did not remain as high after the new snow melted, the local albedo did remain distinctly higher than the period directly before the snow. Since this equation accounts for the changes in sea ice concentration that also occurred, the resulting albedo changes can be attributed to the actual change of the snow and ice albedos rather than solely due to ice convergence during this snow accumulation event.
Figure 1.17: Recorded CERES albedo from the ice parcel database for the 2008E event compared with (A) fresh snow and bare ice albedos calculated using sea ice concentration and (B) snow, bare ice, and melt pond albedos using the melt pond fraction and sea ice concentration.

We then applied these equations to all the case study events, including those which did not melt out at the end of the season, and found that across all cases, the average ice parcel albedo was very close to that of bare ice with ponds at the beginning and end of the event, and slightly higher during the middle of the event (Figure 1.18a). Since ponds were present before the snowfall events and peak albedos did not reach those of bare ice during the event, we speculate that the ponds partially refroze before melting again by the end. Alternatively, the ponds may not have refrozen and the increased albedo could have been due to fresh snow covering a portion of the ice surface, but not all of it.

Next, we applied the same equations to the pan-Arctic scale events. Across all the snow accumulation events, the results are similar to those of the case studies: the average ice parcel albedo is similar to that of bare ice, with the melted or ponded ice albedo being much lower (Figure 1.18b). While the temporal evolution of the albedo within the case study events (Figure
1.18a) is muted when including all events, we can still partially attribute the albedo increase during events to fresh snow accumulation rather than increases in ice concentration.

![Figure 1.18: Measured and expected albedos for (a) all buoy case studies and (b) all snow accumulation events from the ice parcel database showing the average albedos at different stages of the event duration (e.g., 0% = Day 0 (day before the snowfall event), 50% = snowfall event midpoint, 100% = snowfall event end date)](image)

In Section 1.5.3, we calculated radiative forcing of snow accumulation by assuming that all ice (at the given average concentration for each month) was completely covered by fresh snow and no melt ponds were present. While the observed albedos are distinctly higher than those of bare ice with ponds, they are more similar to those of bare ice or snow with ponds than to fully snow-covered ice surface. Because of this, our forcing estimates are likely an overestimate; however, more detailed analysis incorporating melt pond coverage is needed to determine the extent of the over-estimate. With that said, it is important to note that the products used here also include their own inherent biases. For example, the CERES albedo product tends to overestimate downwelling shortwave radiation and underestimate upwelling shortwave (Huang, et al., 2022), which leads to an underestimate in the observed surface albedo values in the ice parcel database.

1.6.2 Comparative Work

This work complements the results presented by Lim et al. (2022), specifically those related to albedo, despite the different criteria used to define summer snowfall events. The Lim et al. (2022) analysis defines a snowstorm as occurring “when the daily snowfall averaged over the
Eurasian-Pacific sector of the Arctic (80–240°E; 69–90°N) exceeds 0.75 standard deviations for 3 or more consecutive days”; however, the results of both methods are complementary. For example, they estimated that a 0.01 increase in albedo may increase upwelling shortwave radiation at the surface by 2.0-3.5 W m⁻² based on the CERES climatological mean (2000-2019). For June-August 2003-2017, our results show an increase of -12.3 W m⁻² when considering an 80% cloud cover and a 203 W m⁻² mean irradiance (Hartmann, 2016). While including cloud cover in our estimate decreases the amount of energy reflected at the surface, the main difference between our findings is based on the different definitions of a snow event. Namely, since our events start with no snow on the ice surface, the increase in surface albedo during the event will generally be greater than if a snow layer was already present. Despite the different methodologies, we also found an average of ~2 snow accumulation events in the Arctic each year, which agrees with the Lim et al. (2022) study’s 2-3 storm events per year.

Until recently, there has been a gap in the scientific literature regarding summer snow on sea ice. An investigation on individual observed snowfalls show the importance of these events to the local energy balance (Perovich, et al., 2017), and the role of storm systems on surface albedo during the summer (Lim et al., 2022); however, little has been done to combine the two topics and expand local-scale surface processes to the broader Arctic. By analyzing observations, satellite, and reanalysis data across the Arctic, this study bridges the gap by quantifying the frequency and extent of summer snow accumulation on sea ice and calculating snowfall’s role in reflecting incoming shortwave radiation from the Arctic sea-ice surface using observationally based methods. These results can provide a valuable basis for climate models to be assessed by (e.g., Light et al., 2015), as they represent a more complete picture of summer snowfall and the Arctic energy balance over sea ice.

1.6.3 Conclusions

The results of this analysis advance our understanding of the frequency, extent, and radiation impacts of summer snowfall over Arctic sea ice. Beyond the results shown here, there are several other ways in which snow can impact the sea ice environment. When sea ice is actively melting, fresh snow accumulation can cool the ice surface as a result of latent heat loss from the ice as the snow melts. The magnitude of this effect varies with snow density and temperature but can be enough to postpone the onset of surface melt and the formation of melt
ponds for a period of time in the beginning of the melt season (Perovich et al., 2017). If all the Arctic sea ice melts during the sunlit season, there would be an increase in the Arctic’s energy absorption of 21 W m$^{-2}$ compared to 1979 values (Pistone, et al., 2019). While our results show that summer snow events contribute only -0.08 W m$^{-2}$ to the Arctic energy balance, every small amount of cooling is important when considering the massive changes that may come with the loss of summer sea ice.

The greatest radiative forcing impact from snow accumulation occurs in July with the combination of very high downwelling shortwave radiation and melt conditions persisting across the Arctic, but more work is needed to investigate the complex processes occurring in the early spring and autumn. Although not included in this study, it is common for snow to accumulate on top of an existing snowpack during these periods. If the existing snowpack is old or melting, it will have a slightly lower albedo than that of the fresh snow and would therefore would also be expected to have a negative radiative forcing effect (Perovich and Polashenski, 2012). As the Arctic climate continues to warm, with melt onset beginning earlier and freeze-up occurring later in the year, the effects of snow accumulation during the shoulder seasons will likely play an increasingly important role in the seasonal evolution of Arctic sea ice mass balance.
General Conclusions

Summer snow is important not only for its role in the Arctic energy balance, but also for how it can impact wildlife, particularly near coastlines. Snow’s high albedo on sea ice has a large influence on primary productivity of the ocean, particularly for the algae and biota living in the sea ice and upper ocean, which form the base for the Arctic food chain. On the surface of the ice, snow provides denning material for polar bears and protection for seal cubs. Near inhabited areas, sea ice provides transportation and hunting grounds for local communities, while further from shore, a snow-covered sea ice environment can make ship travel through the Arctic more difficult and costly than in open water or bare ice conditions.

In short, this study shows the quantitative impact that summer snow accumulation can have on the Arctic sea ice environment. In a changing climate, the radiation response to the high albedo of fresh snow will cool the region and, more broadly, the planet. With the Arctic expected to be effectively ice-free in the summer by the 2030s-2050s under multiple carbon emission scenarios (Kim et al., 2023), the base for snow to accumulate on will also disappear, leading to a complete loss of the negative radiative forcing effect gained from summer snowfall events.
References


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Appendix

Arctic Atmosphere: Reflectance and Transmittance
By Steve Warren, 28 July 2023

At solar zenith angle $\theta_o=66^\circ$, a typical summertime zenith angle for the Arctic Ocean, the clear (unpolluted) atmosphere has reflectance $a_a=0.14$ (Table 5.1 of Paltridge and Platt, 1976). For the broadband transmittance of the clear Arctic atmosphere we use $t_a=0.76$ (Warren and Wiscombe, unpublished, cited by Charlson et al., 1991). Together, these values imply an absorptance $a_a=0.10$ for the clear Arctic atmosphere.

In summer about 80% of the Arctic Ocean is covered by low stratus clouds. For $\theta_o=60^\circ$, Wiscombe (1975, Table 2) calculated cloud reflectance $r_c=0.44$, using some early measurements. Subsequent data from flights above and below Arctic stratus clouds gave absorptance $a_c=0.07$ (Table 3 of Herman, 1977). From more such flights, Herman and Curry (1984, their Table 6 and Figures 2 and 3) plotted transmittance $t_c$ and reflectance $r_c$ versus optical depth $\tau$. For their median optical depth $\tau=10$, they obtained $t_c=0.44$ and $r_c=0.54$. Adjusting these values slightly to give Herman’s $a_c=0.07$, we choose $t_c=0.42$; $r_c=0.51$.

These values we have chosen are for Herman and Curry’s median cloud, $\tau=10$. Measurements of downward spectral irradiance at the surface of the Arctic Ocean by Grenfell and Perovich (2008) are best fitted by a similar optical depth, $\tau=11$ (Section 2 of Dang et al, 2015). Tsay and Jayaweera (1984) also found similar optical depths for Arctic stratus clouds. These results are all consistent with the general observation that Arctic stratus clouds are relatively thin in comparison to stratiform clouds at lower latitude. For example, Twohy et al. (1989) measured clouds in the eastern Pacific between California and Hawaii, obtaining an average optical depth of 22.

Clear-sky broadband albedos at the top of the atmosphere (TOA) from CERES, for $\theta_o=60^\circ$, are about 0.58 over “bright” (snow-covered) ice, and 0.47 over “dark” (snow-free) ice (Figure 2 of Hudson, 2011). These values are somewhat lower than our calculation gives, 0.66 over snow and 0.505 over ice. These discrepancies are consistent with the finding of Hudson et al. (2010) that CERES’s TOA albedos are low by ~8% over snow and ice.
Appendix References


