Northwest Territories and Nunavut Snow Characteristics from a Subarctic Traverse: Implications for Passive Microwave Remote Sensing

CHRIS DERKSEN,* MATTHEW STURM, † GLEN E. LISTON, # JON HOLMGREN, † HENRY HUNTINGTON,@ ARVIDS SILIS,* AND DANIEL SOLIE&

* Climate Research Division, Environment Canada, Toronto, Ontario, Canada
† Cold Regions Research and Engineering Laboratory, U.S. Army Engineer Research and Development Center, Fort Wainwright, Alaska
# Cooperative Institute for Research in the Atmosphere, Colorado State University, Fort Collins, Colorado
@ Huntington Consulting, Anchorage, Alaska
& University of Alaska Fairbanks, Fairbanks, Alaska

(Manuscript received 24 July 2008, in final form 23 September 2008)

ABSTRACT

During April 2007, a coordinated series of snow measurements was made across the Northwest Territories and Nunavut, Canada, during a snowmobile traverse from Fairbanks, Alaska, to Baker Lake, Nunavut. The purpose of the measurements was to document the general nature of the snowpack across this region for the evaluation of satellite- and model-derived estimates of snow water equivalent (SWE). Although detailed, local snow measurements have been made as part of ongoing studies at tundra field sites (e.g., Daring Lake and Trail Valley Creek in the Northwest Territories; Toolik Lake and the Kuparak River basin in Alaska), systematic measurements at the regional scale have not been previously collected across this region of northern Canada. The snow cover consisted of depth hoar and wind slab with small and ephemeral fractions of new, recent, and icy snow. The snow was shallow (<40 cm deep), usually with fewer than six layers. Where snow was deposited on lake and river ice, it was shallower, denser, and more metamorphosed than where it was deposited on tundra. Although highly variable locally, no longitudinal gradients in snow distribution, magnitude, or structure were detected. This regional homogeneity allowed us to identify that the observed spatial variability in passive microwave brightness temperatures was related to subgrid fractional lake cover. Correlation analysis between lake fraction and Advanced Microwave Scanning Radiometer for Earth Observing System (AMSR-E) brightness temperature showed frequency dependent, seasonally evolving relationships consistent with lake ice drivers. Simulations of lake ice thickness and snow depth on lake ice produced from the Canadian Lake Ice Model (CLIMo) indicated that at low frequencies (6.9, 10.7 GHz), correlations with lake fraction were consistent through the winter season, whereas at higher frequencies (18.7, 36.5 GHz), the strength and direction of the correlations evolved consistently with the penetration depth as the influence of the subice water was replaced by emissions from the ice and snowpack. A regional rain-on-snow event created a surface ice lens that was detectable using the AMSR-E 36.5-GHz polarization gradient due to a strong response at the horizontal polarization. The appropriate polarization for remote sensing of the tundra snowpack depends on the application: horizontal measurements are suitable for ice lens detection; vertically polarized measurements are appropriate for deriving SWE estimates.

1. Introduction

Snow cover is a defining characteristic of arctic and subarctic environments, covering the land surface for up to nine months of the year. Because of this temporal persistence, the importance of snow cover to the ecology, climatology, and hydrology of the tundra cannot be overstated. Snow plays a synergistic role in linking processes that span these different systems. For instance, snow–shrub interactions driven by wind transport can control local-scale snow depth distributions and affect the surface energy balance, soil temperature, nutrient release, and the water cycle via spring runoff (Sturm et al. 2001). Snow depth also controls winter season biogeochemical cycling through soil temperature influences on respiration rate (Grogan and Jonasson 2006). The vertical properties of the snowpack also
influence the accessibility of scarce food resources to grazing ungulates—unusually deep snow or the formation of ice layers in the pack can have dire consequences on wildlife populations (Grenfell and Putkonen 2008). The snow cover is also important for human use as burgeoning northern development, such as diamond mining in the Canadian Northwest Territories (NWT), creates increased demand for water resources, both to support industrial activities and for the construction and maintenance of winter ice roads.

Despite this importance, systematic measurements of tundra snow cover across Canada are limited by a sparse conventional observing network that has a strong coastal bias. The measurements themselves are also often flawed—snowfall measurements require correction because of systematic undercatch due to wind and the cumulative effect of numerous trace snowfall events (Benson 1982; Yang et al. 1998, 1999)—whereas individual station snow depth measurements are influenced by local-scale drifting and can poorly represent the surrounding area. Satellite measurements provide systematic observations but with varying levels of effectiveness and uncertainty. Microwave scatterometer observations are well-suited for snowmelt detection (i.e., Kimball et al. 2004) but snow extent datasets derived largely from optical data have problems capturing rapid changes in the snow line during spring (Wang et al. 2005; Brown et al. 2007), and algorithm development for snow water equivalent (SWE) estimates from passive microwave data lag behind other landscape regions (Derksen et al. 2005) as a result of difficulties in signal interpretation.

Given predicted and observed increases in high-latitude precipitation (Kattsov and Walsh 2000; Min et al. 2008; Peterson et al. 2006), we are faced with a scenario where information about the tundra snow cover is increasingly important, but the conventional observation network is inadequate to supply that information. At the same time, solutions derived from satellite data are still unreliable. Unique observational datasets are required to address this situation, and during April of 2007, a coordinated series of snow measurements was made across the Northwest Territories and Nunavut, Canada, during a 4200-km snowmobile traverse from Fairbanks, Alaska, to Baker Lake, Nunavut. Data from this snow survey were used in this study to

1) provide information on the characteristics and spatial variability of the tundra snow cover across this region, an area for which systematic ground-based snow measurements have never been previously acquired; and

2) relate these regional snow cover characteristics to Advanced Microwave Scanning Radiometer for Earth Observing System (AMSR-E) passive microwave brightness temperatures to identify fundamental issues in the development of a tundra-specific SWE retrieval scheme.

This snowmobile-based traverse (Snow Science Traverse–Arctic Region or SnowSTAR-2007) crossed an expansive, largely unpopulated region for which there had been no previous systematic snow measurements. SnowSTAR-2007 followed from previous extensive regional snow surveys conducted in Alaska (such as SnowSTAR-2002 and SnowSTAR-2004; see Sturm and Liston 2003) and was a joint effort between American and Canadian scientists, a contribution to International Polar Year activities in both nations.

2. Study area and measurements

The study area and route of SnowSTAR-2007 are shown in Fig. 1. Measurements were made at 45 locations. The west-to-east route included long stretches on frozen lakes (including Great Bear Lake, the seventh largest lake in the world), areas of rolling tundra, and areas of rocky, moraine-covered tundra. A more detailed description of the exploration history and physiography of the region can be found in Tyrrell (1897) and Douglas (1914).

The most intensive period of sampling occurred during the Daring Lake, Northwest Territories, to Baker Lake, Nunavut portion of the traverse (16–26 April 2007). Sample sites were located at least every 1° of longitude. At these sites, we measured snow depth, density, snow water equivalent (SWE), stratigraphy, and grain size. The goal was to make measurements at paired sites: one on tundra (land) and one on ice (lake or river ice, depending on what was available). It was not always possible to achieve this goal. At selected sites, we collected snow samples for analysis of soot content, ionic loading, and mercury on behalf of other research groups.

Standard methods were used to measure snow stratigraphy, density, and SWE (Colbeck et al. 1992). More extensive depth transects were measured than is customary by using a self-recording snow depth probe (U.S. Patent No. 5864059; cf. Sturm and Liston 2003) linked to a GPS. At each site, 201 depth measurements were made at 0.5-m spacing over a 100-m line. These closely spaced measurements were followed by several hundred more measurements made at random spacing, their location recorded by the GPS. Bulk SWE and density were determined from snow cores (measured in pairs) taken every 25 m along the sampling line. An ESC-30 snow corer (cross sectional area of 30 cm²) was used for these measurements. Grain size was measured using a
stereo microscope and comparator card. Near-infrared (NIR) images of snow pit faces were made using a Sony DSC-P200 Cybershot 7.2M pixel digital camera equipped with an NIR filter (MaxMax Xnite 850). Sensitivity of NIR reflectance to snow grain specific surface area (SSA) allows for clear delineation of snowpack stratigraphy. These images were processed in accordance with the methods of Matzl and Schneebeli (2007).

One of the primary objectives of the traverse was to acquire the necessary measurements to evaluate the capability of passive microwave satellite data to estimate SWE across the tundra. While passive microwave data are capable of retrieving SWE information across prairie and forest environments (Derksen et al. 2003; Derksen 2008), no algorithm specific to tundra areas has yet been developed. The traverse measurements complement the spatially intensive tundra snowpack measurements made at Daring Lake, NWT since 2004, which have illustrated the high degree of error in contemporary passive microwave SWE algorithms in the tundra (Rees et al. 2006). Daily AMSR-E measurements in level 2 swath and 25-km EASE-Grid formats were acquired for the complete 2006/07 snow cover season from the National Snow and Ice Data Center (NSIDC) for the traverse region (Knowles et al. 2006). AMSR-E is on board the EOS–Aqua platform and has provided multifrequency and multipolarization measurements since 2002 at an enhanced spatial resolution compared to previous microwave sensors [the Scanning Multichannel Microwave Radiometer (SMMR); the Special Sensor Microwave Imager (SSM/I)].

3. Results: General nature of the snowpack

a. Depth, density, and water equivalent

Typical snow depth transect profiles (100 m long) from the traverse are shown in Figs. 2a,b for lake and tundra sites, respectively. The more spatially expansive roving snow depth measurements for these same sites are shown in Figs. 2c,d. These transects yielded similar profiles at all sampling locations. Probability distribution functions (PDFs) of snow depth, density, and SWE measurements based on all the measurements from all the sites are shown in Fig. 3. Statistics describing these PDFs are given in Table 1.

Separate PDFs were plotted for the measurements from 1) all sites, 2) terrestrial sites, and 3) lake/river sites. For snow depth, all distributions have a peak at approximately 30–40 cm, indicating this was the modal range for snow depth over lakes and tundra. Differences in the depth distributions, however, are clear between terrestrial (Fig. 3b) and lake sites (Fig. 3c). Snow depth range and variability were higher at terrestrial sites: the more extensive right skew toward higher depth values is driven by variability in snow catchment and retention related to slope and vegetation. These two parameters are absent on lakes, where accumulation patterns are produced solely by wind drifting (Sturm and Liston 2003).
Unlike snow depth, the snow density measured on lakes (Fig. 3f) had a more uniform distribution (i.e., the probability of a bulk density of 0.3 g cm\(^{-3}\)) was not significantly different than the probability of density of 0.4 g cm\(^{-3}\)) compared to a normal distribution centered on 0.35 g cm\(^{-3}\) derived for terrestrial sites (Fig. 3e). The PDFs illustrate a clear lower threshold in snow density across the entire region of 0.3 g cm\(^{-3}\). Measurements below this value were very rare (~5% of all density measurements), regardless of land cover and substrate. Measured SWE values exhibited greater variability at terrestrial (Fig. 3h) versus lake (Fig. 3i) sites. The trapping influence of vegetation creates the strong right skew toward higher SWE values in the terrestrial measurements, while a SWE maximum on lakes of approximately 130 mm was evident.

Because SWE is a product of depth and density, the underlying driver of SWE variability can be determined from the PDFs in Fig. 3. On land, a lognormal distribution for depth was coupled with a sharply peaked density PDF centered on about 0.35 g cm\(^{-3}\). Land snow depths varied with the distribution of tussocks, rocks, and shrubs—possibly even with the microtopography—but the bulk density was consistent because wind action was the same across all of the vegetation and terrain types. This produced a broad, skewed SWE PDF (Fig. 3h). In contrast, without any shrubs or tussocks, the snow on the lakes and rivers tended to have a uniform depth as a result of much more consistent, uninterrupted wind action, but the same processes produced a wide range in density and are the result of the successive pattern of hard-packed drifts interspersed with softer snow. For lakes, this density variation broadened the SWE distribution in comparison to the strongly peaked depth distribution.

In short, tundra SWE is variable because tundra snow depth is variable (density is relatively constant), while lake/river SWE is variable because lake/river ice snow density is variable (though depth is relatively consistent).

b. Stratigraphy

Snowpits were excavated to determine the snowpack stratigraphy at each site. Layer and grain types were noted following the classification of Colbeck et al. (1992). Because depth hoar and wind slab are ubiquitous in the tundra snow, the classification system was broadened for those types of snow to allow for a more nuanced approach. Specifically, we recognized four types of slabs based on hardness (soft, moderate, hard, and very hard), and we noted depth hoar layers whose texture indicated they had origins as a wind slab. This latter type of layer is generally not seen outside of the Arctic and arises because the temperature gradients are so strong (Sturm et al. 1995) that dense, fine-grained layers of wind slab, and even ice layers, eventually metamorphose into large, faceted, and striated depth hoar grains. A key characteristic of these slab-to-hoar layers is that they are tough and not fragile like most depth hoar. If they are composed of mainly depth hoar crystals, yet still cohesive, we called the layers “indurated” (meaning hardened). If a significant number of small wind grains remained, we called them “slablike.”

NIR photographs of snowpit faces were taken at each site. Techniques remain to be refined for the quantitative retrieval of parameters, such as grain size (see Matzl and Schneebeli 2007), so quantitative analyses of these images are not included in this study. We found, however, a simple setup and camera configuration procedure allowed photographs to be acquired even under extremely...
cold and often windy conditions. The NIR photographs are usable for documenting stratigraphy rapidly and efficiently (see Fig. 4 for an example), so we believe the method should become a standard snow pit practice.

As noted previously for Alaskan tundra snow (Benson 1969; Benson and Sturm 1993), the snowpack of NWT and Nunavut was made up of a limited number of layers, almost always fewer than seven (Table 2). There were

Table 1. Summary of measured depth, density, and SWE PDFs (see Fig. 3). Automatic depth probes could only measure up to 122-cm depth but not deeper, hence the uncertainty in max and range values.

<table>
<thead>
<tr>
<th>Depth (cm)</th>
<th>Density (g cm$^{-3}$)</th>
<th>SWE (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>All</td>
<td>Tundra</td>
</tr>
<tr>
<td>Mean</td>
<td>33.1</td>
<td>38.9</td>
</tr>
<tr>
<td>Standard deviation</td>
<td>16.2</td>
<td>19.5</td>
</tr>
<tr>
<td>Min</td>
<td>0.2</td>
<td>0.2</td>
</tr>
<tr>
<td>Max</td>
<td>&gt;122</td>
<td>&gt;122</td>
</tr>
<tr>
<td>Range</td>
<td>&gt;122</td>
<td>&gt;122</td>
</tr>
<tr>
<td>CV</td>
<td>0.49</td>
<td>0.50</td>
</tr>
<tr>
<td>Skew</td>
<td>1.21</td>
<td>0.78</td>
</tr>
<tr>
<td>Kurtosis</td>
<td>2.99</td>
<td>1.31</td>
</tr>
</tbody>
</table>
generally more layers on the tundra (terrestrial) than on ice substrates (lake/river), probably because early season snowfalls fell into unfrozen lakes and rivers. At the time of measurement (16–26 April 2007) in all locations, more than two-thirds of the pack by thickness was depth hoar (including slab to hoar), while a surprisingly low percentage (11%–14%) was wind slab. In northwestern Alaska, we have typically seen about twice this percentage of wind slab (Sturm and Liston 2003). We do not know whether 2007 was a low wind year in Nunavut and the NWT, or whether depth hoar metamorphism driven by low temperatures converted more slabs into hoar, or some combination of both. As we discuss later, there is reason to think metamorphism was more extreme over the traverse route than in northwestern Alaska.

The contrast in bulk density between tundra and lake sites (Table 2) was expected (windier lakes produce more dense snow layers, hence overall a denser snowpack), but the low density for river sites is more difficult to explain. Despite these minor differences, overall, the snow across all terrain units sampled during the traverse tended to be similar. All of it was thin, with relatively few layers, and almost all of it was fully converted into coarse-grained depth hoar by the end of winter.

Two of the snow types—new/recent and icy—tended to change rapidly in response to weather events and, by thickness, were not major components of the snowpack. What little new/recent snow was present (Table 2; average of 11% of total pack by thickness), most often was observed to be an ephemeral layer at the top of the pack. As soon as the wind started to blow, the thickness of this layer would double or fall to zero, a process we watched in action many times. If the last six sites of the traverse are excluded (each of which was about 30% new/recent snow because of a recent snow storm), the average percentage of new/recent snow across the traverse falls to 7.5% of the total. Although new snow is important in many ways, it is a transient feature of the snowpack and therefore hard to predict in time and space. Icy snow was even a smaller percentage of the pack by thickness, on average 3%. In general, we found overflow and snow–ice to be rare. Most of the other icy layers in the snowpack were the product of a single warm period with rain-on-snow. These layers were never more than 5 mm thick, though as we discuss later, they may have radiometric importance.

Excluding new/recent and icy snow, we examined the relative distribution of depth hoar layers and wind slabs. These account for almost 80% by thickness of the pack. The key to understanding the stratigraphy observed on the traverse, and elsewhere in the Arctic as observed on other traverses, is how these two basic types of snow textures play out. They are linked. As noted before, wind slabs can be converted into depth hoar, while depth hoar can be eroded and reconstituted as a wind slab. Strong winds and vigorous wind transport tend to form dense, tough slabs that resist conversion to depth hoar. Less strong winds form slabs that are less cohesive and fine-grained, which tend to metamorphose more easily. Early snow and low winter temperatures produce strong temperature gradients across the snowpack that can accelerate slab-to-hoar conversion; late snow and mild temperatures tend to preserve wind slabs.

In Fig. 5a, a pie chart shows the relative abundance (by count rather than thickness or SWE) of 11 types of

---

**Table 2. Stratigraphy statistics for different types of sites.**

<table>
<thead>
<tr>
<th></th>
<th>All sites</th>
<th>Tundras</th>
<th>Lakes</th>
<th>Rivers</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number of layers</td>
<td>5.8</td>
<td>6.6</td>
<td>4.9</td>
<td>5.5</td>
</tr>
<tr>
<td>Avg layer thickness (cm)</td>
<td>6.6</td>
<td>6.7</td>
<td>6.4</td>
<td>6.3</td>
</tr>
<tr>
<td>Total thickness (cm)</td>
<td>37.1</td>
<td>43.4</td>
<td>30.1</td>
<td>33.6</td>
</tr>
<tr>
<td>SWE (cm)</td>
<td>11.4</td>
<td>14.0</td>
<td>10.1</td>
<td>8.1</td>
</tr>
<tr>
<td>Bulk density (g cm(^{-3}))</td>
<td>0.311</td>
<td>0.319</td>
<td>0.340</td>
<td>0.236</td>
</tr>
<tr>
<td>Hoar fraction</td>
<td>0.71</td>
<td>0.65</td>
<td>0.76</td>
<td>0.71</td>
</tr>
<tr>
<td>Slab fraction</td>
<td>0.12</td>
<td>0.14</td>
<td>0.11</td>
<td>0.12</td>
</tr>
<tr>
<td>Icy fraction</td>
<td>0.03</td>
<td>0.02</td>
<td>0.04</td>
<td>0.02</td>
</tr>
<tr>
<td>New/recent fraction</td>
<td>0.11</td>
<td>0.14</td>
<td>0.07</td>
<td>0.12</td>
</tr>
<tr>
<td>Other fraction</td>
<td>0.03</td>
<td>0.03</td>
<td>0.02</td>
<td>0.03</td>
</tr>
</tbody>
</table>

---

Fig. 4. NIR photograph of a snowpit face on Yamba Lake. Layers have been manually delineated following contrast enhancement.
snow seen in the 259 total snow layers that were identified in the field. Thirty-three percent of all layers were neither depth hoar nor wind slab. If we remove these layers from the count (remove new, recent, medium, and fine-grained snow, melt clusters, and icy snow) and look at the remaining layers (Fig. 5b), a more interesting story appears. Forty-six percent of this remaining subset is classified as pure depth hoar. These are layers that probably did not start out as wind slabs. They were initially new, recent, or fine-grained snow layers. Seventeen percent of the layers were still wind slab at the time of the traverse. The remaining 37% was depth hoar with textural indicators, suggesting it had been transformed into depth hoar from wind slab. Adding this 37% to the 18% wind slab that remained suggests 55% of the snowpack was initially wind slab. In essence, about half the snow started as nonwind slab snow and became depth hoar, while the other half started as wind slab; however, most of the snow metamorphosed into depth hoar. This distribution is consistent with tundra snow results from northwestern Alaska (Sturm and Liston 2003; Fig. 2), where about half of the snow was wind slab and the other half depth hoar. The implication of this finding is that in the traverse region conditions may favor even more extreme metamorphism than has been typically measured in arctic tundra snow of other regions.

Average grain size by snow layer type is shown in Table 3. The values in the table are averages computed for all layers identified as that type of snow. There were 259 layers for which grain size was measured. The percentages listed in Fig. 5a, multiplied by 259, give the approximate \( n \) value for each grain size average. Ice layers do not have distinct grains, so they have been excluded from the table. The results show that the largest grains were found in indurated depth hoar layers and depth hoar chain assemblages, and that these grains (i) were 5 times larger than the grains in wind slabs, and (ii) the hoar grains tended to be plate-like with short-to-long axial ratios near 1:4. We tested whether the grain sizes differed if the snow was on tundra versus ice substrate but found no significant differences.

c. Regional snowpack variability

The conventional wisdom is that snow varies along altitudinal and latitudinal gradients but not necessarily along longitudinal gradients. Since the traverse route was essentially west to east, we examined if there were any longitudinal gradients in the snow properties. Storm tracks and synoptic patterns for Northwest Territories and Nunavut are complex, so both latitudinal
Table 3. Grain sizes (determined by field microscope and comparator card) by snow texture type. The hardness code is a relative measure of layer hardness: the higher the number, the harder the snow; “fg” is fine grained and “mg” is medium grained.

<table>
<thead>
<tr>
<th>Snow texture type</th>
<th>Grain size</th>
<th>Hardness Code</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Long axis (mm)</td>
<td>Short axis (mm)</td>
</tr>
<tr>
<td>New/recent</td>
<td>0.9</td>
<td>0.4</td>
</tr>
<tr>
<td>fg and mg</td>
<td>0.9</td>
<td>0.4</td>
</tr>
<tr>
<td>Soft slab</td>
<td>0.8</td>
<td>0.3</td>
</tr>
<tr>
<td>Moderate slab</td>
<td>0.8</td>
<td>0.6</td>
</tr>
<tr>
<td>Hard slab</td>
<td>2.0</td>
<td>0.3</td>
</tr>
<tr>
<td>All slabs</td>
<td>0.9</td>
<td>0.5</td>
</tr>
<tr>
<td>Slab to hoar</td>
<td>2.3</td>
<td>0.8</td>
</tr>
<tr>
<td>Indurated hoar</td>
<td>5.7</td>
<td>1.1</td>
</tr>
<tr>
<td>Slab to hoar</td>
<td>3.8</td>
<td>0.9</td>
</tr>
<tr>
<td>Indurated hoar</td>
<td>4.0</td>
<td>1.1</td>
</tr>
<tr>
<td>Depth hoar</td>
<td>6.4</td>
<td>1.6</td>
</tr>
<tr>
<td>Hoar chains</td>
<td>1.3</td>
<td>0.5</td>
</tr>
<tr>
<td>Melt clusters</td>
<td>1.3</td>
<td>0.5</td>
</tr>
</tbody>
</table>

Table 4. Correlation (r value) results for snow parameters vs longitude.

<table>
<thead>
<tr>
<th>Snow parameter</th>
<th>Tundra</th>
<th>Lakes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Depth</td>
<td>-0.20</td>
<td>0.73</td>
</tr>
<tr>
<td>Density</td>
<td>0.18</td>
<td>-0.52</td>
</tr>
<tr>
<td>SWE</td>
<td>-0.10</td>
<td>0.52</td>
</tr>
<tr>
<td>Layers</td>
<td>-0.39</td>
<td>0.79</td>
</tr>
<tr>
<td>Grain long</td>
<td>0.25</td>
<td>0.61</td>
</tr>
</tbody>
</table>

and east–west snow gradients across NWT and Nunavut are possible. The most intensive period of sampling occurred between Daring Lake and Baker Lake (16–26 April; see Fig. 1). We used a correlation analysis of mean site values versus longitude to determine if there are gradients reflected in the data. Two sites along the Thelon River were omitted from this analysis because the Thelon “Oasis” is known to be an anomalous area, a large outlier of taiga forest (and taiga snow) hundreds of kilometers north of the normal tree line. The results show only weak relationships with longitude for the tundra sites (Table 4), though stronger relationships were evident for the lake sites. The snow cover on the lakes became shallower and more dense and was composed of fewer layers with smaller grains as we moved eastward. A possible explanation is a gradient in lake freeze-up dates from west to east, but there are no observations to confirm this.

All measurements were also segmented into 1° longitudinal bins. Box plots of depth, density, and water equivalent are shown in Fig. 6. No longitudinal trends in the variability of snow properties were evident for tundra and lake sites. The depth, density, and SWE gradients for snow on lakes suggested by the correlation analysis are evident in the box plots but as before, the gradients are very subtle.

4. Results: Passive microwave observations

a. Spatial brightness temperature variability

A primary objective of the traverse was to assist in the efforts to develop and validate satellite passive microwave SWE retrieval algorithms specific to open tundra environments (for a general overview, see Rees et al. 2006). Regional brightness temperature patterns from AMSR-E measurements are shown in Fig. 7 for the Daring Lake to Baker Lake segment of the traverse. Aside from the oasis of trees along the Thelon River, the domain in Fig. 7 is almost exclusively north of the treeline. As discussed in the previous section, there were no strong gradients in snow cover properties across this region. Examination of HYDRO1K topographic datasets showed little variability in elevation or topographical complexity along this portion of the traverse. Despite this consistency in snow cover, land cover, and topography, spatial patterns in the AMSR-E measurements are evident (Figs. 7a–d) and qualitatively appear to be related to lake fraction (Fig. 7f).

The traverse route shown in Fig. 7 can be broken down into three basic units: (1) a lake-rich area in the west composed of a connected sequence of large lakes (i.e., Lac de Gras, Thonokied Lake, Afridi Lake, Aylmer Lake, and Clinton-Colden Lake); (2) a relatively lake-sparse region that includes the Thelon Oasis; and (3) the interconnected large lake system that flows into Chesterfield Inlet at Baker Lake (Beverly Lake, Aberdeen Lake, Schultz Lake to the north, and Lake Dubawnt to the south). The low-frequency AMSR-E measurements (6.9 and 10.7 GHz) are approximately 10–15 K lower across the lake-rich areas (Figs. 7a,b) than the lake-poor area in the center. The 18.7-GHz measurements exhibit only subtle variations over the region, with a total range of only 5–10 K (Fig. 7c). The largest brightness temperature range was observed at 36.5 GHz (Fig. 7d). Unlike the lower frequencies, 36.5-GHz brightness temperatures were up to 30 K higher over the two lake-rich portions of the study area.

Sensitivity to lake fraction is problematic when conventional approaches to retrieving SWE from microwave brightness temperatures are used (Green 2007; Duguay et al. 2005). Most SWE retrieval algorithms exploit the difference between a measurement frequency sensitive to snow grain volume scattering (~37 GHz) and a measurement frequency considered insensitive to snow (~19 GHz; Chang et al. 1990; Goodison and Walker 1995; Kelly et al. 2003; Mognard and Josberger 2002;
Pulliainen 2006): the larger the difference between these two measurements, the higher the estimates of SWE. Nineteen and 37 GHz are commonly used because these frequencies extend continuously through the passive microwave satellite record from 1978 to the present. Our snow surveys, however, indicate no strong longitudinal gradient in flat tundra and lake SWE across the Daring Lake to Baker Lake portion of the traverse (see Table 1). If the 36.5-GHz measurements were predominantly influenced by only the volume scatter of the snowpack, the pattern in Fig. 7d should be nearly spatially homogeneous. Instead, there is a strong spatial pattern that appears to be driven by the lake fraction.

The frequency-dependent response of the AMSR-E measurements to lake fraction compounds the problem. If a high lake fraction increases the 36.5-GHz brightness temperatures by 30 K but has little influence on the 18.7 GHz measurements, then higher lake fractions will be
systematically associated with a smaller 36.5–18.7 brightness temperature difference, and hence lower SWE estimates. This is illustrated in Fig. 7e, which shows the AMSR-E 36.5–18.7 brightness temperature difference. Across lake-rich areas the difference is closer to zero than in lake-poor regions, so retrieved SWE with a simple brightness temperature difference algorithm uncorrected for lake fraction produces regional biases unrelated to the snowpack.

b. The influence of lake ice fraction

To understand and potentially exploit microwave brightness temperature response across lake-rich tundra

FIG. 7. Vertically polarized AMSR-E brightness temperatures at (a) 6.9, (b) 10.7, (c) 18.7, and (d) 36.5 GHz. (e) Difference of 36.5 and 18.7 GHz and (f) subgrid lake fraction. Symbols denote surface measurement sites.
environments, the seasonally evolving influence of the snowpack must be separated from the seasonally evolving influence of lake ice across the range of satellite measurement frequencies. Because of wavelength specific variations in penetration depth, low-frequency measurements are controlled by emission below the ice layer, as illustrated by Hall et al. (1981) for lake ice in Colorado. Higher-frequency measurements are influenced by the water when the snow and ice are thin, but the combined snow and ice thickness will eventually exceed the penetration depth. A schematic illustrates this concept in Fig. 8.

To investigate these competing influences, correlation analysis was performed on weekly averaged EASE-Grid AMSR-E brightness temperatures and lake fraction. The lake fraction estimates (Fig. 7f), produced by the U.S. Geological Survey (USGS) and regridded to the EASE-Grid, are static but five winter seasons (November through April) of AMSR-E data were analyzed, producing five seasonal correlation values for each weekly interval. Boxplots of these correlation values for the first week of each month are shown in Fig. 9 and provide an overview of the seasonally evolving relationships between brightness temperature and lake fraction at microwave frequencies ranging from 6.9 to 36.5 GHz.

To assist in the interpretation of the brightness temperature versus lake fraction correlation results, simulations of lake ice thickness and snow depth on lake ice were produced from the Canadian Lake Ice Model (CLIMo; see Duguay et al. 2003) driven by observations from the Environment Canada surface observing station at Baker Lake. CLIMo is a one-dimensional thermodynamic model that simulates lake ice phenology, including freeze-up and break-up dates, temperature profiles, ice thickness (including clear ice versus snow ice), and snow depth. It has been evaluated extensively across northern Canada and Alaska (see Menard et al. 2002; Duguay et al. 2003). The Baker Lake simulation should be representative of general lake ice conditions along the traverse route because of the consistency in the observed snow and ice conditions. The simulation results for the same period investigated in the correlation analysis (2002/03–2006/07) are also plotted in Fig. 9 and show relatively consistent ice thickness and snow depth evolution through the five seasons considered. Maximum ice thickness ranged from approximately 150 to 175 cm, with snow depth peaking near 50 cm.

Interseasonally consistent relationships were identified between AMSR brightness temperatures and lake fraction. At low frequencies (6.9; 10.7 GHz), these are constant through the winter season at r values around 0.6 (Figs. 9a,b). High lake fraction is associated with lower brightness temperature as a result of the influence of subice liquid water that consistently influences microwave emission at these frequencies through the complete ice growth season. At higher frequencies (18.7; 36.5 GHz), the strength and direction of the correlations evolve consistently with the penetration depth as the influence of the subice water is replaced by emissions from the ice and snowpack (Figs. 9c,d).

Mätzler (2001; Fig. 2) illustrated the microwave penetration depth for frequencies between 1 and 100 GHz (prescribed snow density of 0.300 g cm$^{-3}$; snow temperature of 270 K) simulated using the Microwave Emission Model of Layered Snowpacks (MEMLS; Wiesmann and Mätzler 1999; Mätzler and Wiesmann 1999). Very similar values for freshwater ice were calculated by Surdyk (2002). From these simulations, the penetration depths at AMSR-E frequencies were compared to the seasonal ice thicknesses simulated by CLIMo (illustrated in an idealized fashion in Fig. 8) and the correlations between brightness temperature and lake fraction shown in Fig. 9. A summary of this assessment is provided in Table 5. Of note, the penetration depth at 18.7 GHz is between 1 and 2 m. This combined ice/snow thickness is reached some time in February or March, when the correlation between 18.7 GHz brightness temperature and lake fraction changed from weakly negative to positive (Fig. 9c). Similarly, the penetration depth at 36.5 GHz is less than 1 m; this combined snow and ice thickness was typically reached in November or December, when the correlation with lake fraction jumped from significantly negative to significantly positive (Fig. 9d). The changing state of ice thickness clearly influences the seasonal brightness temperature evolution at the satellite scale, and this influence is dependent on
frequency. The correlation results show that multifrequency approaches must consider the temporally changing influence of subgrid lake fraction on satellite brightness temperature measurements.

c. The influence of a surface ice layer

The choice of polarization is not consistent among microwave SWE algorithms—some use horizontal measurements (Chang et al. 1990 and the modified versions of Foster et al. 1997; Kelly et al. 2003), while others are based on vertical measurements (Goodison and Walker 1995; Goita et al. 2003; Mognard and Josberger 2002; Pulliainen 2006). One advantage of vertical polarization measurements is the reduced sensitivity to ice lenses within the snowpack. An opportunity to assess this sensitivity arose because of a rain-on-snow event in the vicinity of Daring Lake on 7 April 2007. The 36.5-GHz polarization gradient [vertical (V) and horizontal (H)] was calculated for 3 April (pre-event and cold conditions) and 9 April 2007 (postevent and cold conditions). The difference between these two dates is illustrated in Fig. 10. No difference is evident over the majority of the study area, except in the region of Daring Lake (and west), where the difference in polarization gradient between the two dates reaches 30 K. This difference is driven by a strong response in the horizontal polarization measurements (not shown). The extent of the polarization gradient difference agrees well with observations of a surface ice layer (1–5 mm thick) at sites extending from north of Yamba Lake, south to Lake Providence, and east to Thonokied Lake (as noted in Fig. 10).

5. Discussion and conclusions

The SnowSTAR-2007 traverse spatially extends more than a decade of intense snow cover measurements in arctic North America. In Alaska, a series of twice-yearly

<table>
<thead>
<tr>
<th>Frequency (GHz)</th>
<th>Penetration depth (m)</th>
<th>Ice thickness = penetration depth</th>
<th>Change in correlation</th>
</tr>
</thead>
<tbody>
<tr>
<td>6.9</td>
<td>&gt;15</td>
<td></td>
<td>From negative to positive</td>
</tr>
<tr>
<td>10.7</td>
<td>~10</td>
<td></td>
<td>From positive to positive</td>
</tr>
<tr>
<td>18.7</td>
<td>~2</td>
<td>February/March</td>
<td>From positive to positive</td>
</tr>
<tr>
<td>36.5</td>
<td>~0.75</td>
<td>November/December</td>
<td>From negative to positive</td>
</tr>
</tbody>
</table>
oversnow traverses were implemented between 1994 and 1997 in the Kuparuk basin (König and Sturm 1998; Liston and Sturm 2002; Taras et al. 2002). This transect was then moved west in 2000, forming a line between the Brooks Range and Barrow (Sturm and Liston 2003). These traverses were about 200 km in length. The snow measurement expeditions were then extended to northwest Alaska in 2002 and 2004, where snow along a 1000-km route between Nome and Barrow was measured. In Canada, five years of intensive snow sampling to determine landscape controls on snow (re)distribution and to provide ground measurements for comparison with satellite and airborne remote sensing data started at Daring Lake in 2003 (Rees et al. 2006). Other campaigns were conducted near Churchill, Manitoba (DerkSEN et al. 2005). At the broadest scale, all these field datasets validate an idea put forward in 1995 (Sturm et al. 1995) that there was a distinct class of snow called tundra snow. In general, the snow properties we observed on the traverse were clustered within narrow limits of depth, density, and grain/layer characteristics, and these were similar to the range of properties we had observed farther west in Alaska and farther south in Manitoba. The new measurements, combined with the older, indicate the tundra snow class stretches more than 6000 km across arctic and subarctic North America. It is likely that within that band, there are regional differences in the amount of wind slab and depth hoar, but overall, these are second-order effects.

Collecting snow data at this large regional scale can be sobering, because it is easy to see how much of the complexity of the snow cover on the landscape is being ignored. By necessity, we have to drive past many kilometers of snow before stopping at a measurement site. Is that site typical of the snow we just passed by? Does the necessity of traveling where there is a trail or a lake bias the selection of sample points? There is a constant tension between traveling (there are always time constraints on such a sample campaign) and stopping to make measurements. Despite these issues, the data collected in 2007 show surprisingly consistent patterns, suggesting that sampling bias was not too severe and that while local variations did exist, regional variations (or the lack thereof) were captured. In addition, working at regional-to-continental scales forced us to think broadly about the snow cover, and this is a valuable perspective for climate studies and remote sensing applications.

With the exception of a single site that was in the taiga, all of the snow examined during the SnowSTAR-2007 traverse could be classified as tundra snow. The tundra snow cover is a thin snowpack that consists mainly of wind slab and depth hoar (the latter usually making up layers at the bottom of the pack), with small but variable and transient amounts of new and recent snow at the top of the pack. Features associated with wind deposition and erosion (dunes and sastrugi) are ubiquitous in this tundra-snow environment. The mature tundra snowpack rarely has more than 5–8 layers, a product of a limited number of winter precipitation events. Minus a thin veneer of new or recent snow, the top layers are wind slabs due to the free play of the wind during the long winter, and the basal layers are depth hoar due to the strong metamorphic temperature gradients imposed by generally low air temperatures. Wind slabs formed in early winter often completely metamorphose into depth hoar by the end of the season as a result of these temperature gradients (Sturm and Benson 1997). The grains in these layers will be morphologically similar to regular depth hoar, but the layers will be stronger and more cohesive than normal for depth hoar layers, a relict feature of the original wind slab. We call these layers indurated.

Because of wind action, the depth and thickness of individual layers vary markedly over short distances, as do the density, hardness, and grain size. Across longer distances, and across the entire region, there is a surprising amount of convergence in the snow properties. The biggest variations in snow properties are not regional—they are local and are associated with topographic breaks and local variations in vegetation.

One within-region source of variability in snow distribution arises from the presence of lakes. Results from previous SnowSTAR campaigns (see Sturm and Liston 2003) and measurements from Daring Lake (A. Rees 2008, unpublished manuscript) are shown in Table 6. A consistent difference between tundra and lake snow properties is evident from multiple years of data and
from sites distributed across North America. SWE on lakes tends to be 63%–81% of that measured at tundra sites, while the snow density on lakes ranges from 107% to 126% of that measured on land. As discussed in Sturm and Liston (2003), these differences can be readily explained by processes that operate differently on the tundra than on the lakes. The range of values (Table 6) is surprisingly narrow given the generally heterogeneous nature of snow cover: 18% for SWE, and 19% for density.

The traverse data also confirm that if we want to use remote sensing to monitor tundra snow, we must deal with the influence of lakes. The problem is that the passive microwave frequencies traditionally used to retrieve SWE (~19 and 37 GHz) have penetration depths that are strongly influenced by water beneath the ice for part of the season but are also influenced by the ice and overlying snowpack by the end of the season (see Fig. 8). The timing of this shift in the primary emission source is different at 19 versus 37 GHz, so the traditional brightness temperature difference approach is not appropriate for lake-rich tundra environments. Instead, we need to consider other options. One potential pathway is to drive a model that produces estimates of lake ice thickness, such as CLIMo, with atmospheric reanalysis datasets to provide near-real time ice thickness datasets useful for interpreting brightness temperatures from lake-rich areas.

A complicating factor in this approach is that if we correctly simulate the lake ice thickness, we need information on how much snow is on the lake ice. Incorporating modeled ice thickness would remove one unknown parameter, leaving only the SWE (on both lakes and land) to resolve and allow for the use of multiple frequencies despite the variable lake-ice influence. Multiple frequencies mitigate problematic physical temperature and atmospheric effects that occur when single frequencies are used (see Wang and Tedesco 2007; Markus et al. 2006). A key remaining challenge for tundra SWE retrieval is the slope reversal in the SWE versus 37-GHz brightness temperature relationship that occurs beyond a theoretical limit of approximately 120-mm water equivalent (Ulaby et al. 1986; Schanda et al. 1983; Derksen 2008). The approximate point at which this slope reversal occurs in high-density, fine-grained tundra snowpacks has not been documented.

Regardless of the approach, the snow measurements from the traverse provide a unique dataset for the assessment of satellite measurements, including the evaluation of new retrieval techniques and radiometric models. The sensitivity of passive microwave measurements to a thin surface ice layer was also illustrated. The strong influence on horizontally polarized 36.5-GHz data shows both the ability to detect these layers with microwave measurements and the need to account for these layers if horizontally polarized data are used to retrieve snow depth or water equivalent.

The area covered by the SnowSTAR-2007 traverse is a part of the Arctic that is warming rapidly (Hassol 2004). One ramification of the warming is likely to be a change in the winter snow-cover magnitude, distribution, and characteristics (Raisanen 2007). Unfortunately, we have not found any descriptions of the snow cover against which we can compare our 2007 measurements. Most travelers across the region were struggling to stay alive, looking for gold or diamonds, or moving too fast to make measurements. Our field observations may be useful in the future as a baseline against which present conditions can be compared. Rain-on-snow and winter-thaw events may become more common. This would lead to more ice layers, as we observed near Daring Lake. These layers have implications for wildlife (see Grenfell and Putkonen 2008) and would clearly complicate remote sensing. Depth hoar percentages may decrease, while wind slab percentages may increase. Warmer winters would produce less metamorphism, hence fewer slabs would be converted into depth hoar. Paradoxically, because slabs have higher thermal conductivity than depth hoar (Sturm and Benson 1997), this may actually lead to lower winter soil temperatures, which would also have ecosystem effects.

Acknowledgments. SnowSTAR-2007 was supported by the National Science Foundation, the Cold Regions Research and Engineering Laboratory, and Environment...
REFERENCES


