**The influence of snow microstructure on dual-frequency radar measurements in a tundra environment**

**Joshua King1\*, Chris Derksen1, Peter Toose1, Alexandre Langlois2, Chris Larsen3, Juha Lemmetyinen4,5, Phil Marsh6, Benoit Montpetit7, Alexandre Roy8, Nick Rutter9, and Matthew Sturm3**

1 Environment and Climate Change Canada, Climate Research Division, Toronto, Canada

2 Université de Sherbrooke, Centre d'Applications et de Recherches en Télédétection, Québec, Canada

3 University of Alaska Fairbanks, Geophysical Institute, Fairbanks, Alaska, USA

4 Finnish Meteorological Institute, Arctic Research, Helsinki, Finland

5 Chinese Academy of Sciences, Institute of Remote Sensing and Digital Earth, Beijing, China

6 Wilfrid Laurier University, Cold Regions Centre, Waterloo, Canada

7 Environment and Climate Change Canada, Wildlife and Landscape Science Division, Ottawa, Canada

8 Université de Montréal, Département de Géographie, Montréal, Canada

9 Northumbria University, Department of Geography and Environmental Sciences, Newcastle upon Tyne, UK

**\*** Correspondance: Joshua.King@Canada.ca

**Abstract:**

Recent advancement in the understanding of snow-microwave interactions has helped to isolate the considerable potential for radar-based retrieval of snow water equivalent (SWE). There are however, few datasets available to address spatial uncertainties, such as the influence of snow microstructure, at scales relevant to space-borne application. In this study we introduce measurements from SnowSAR, an airborne, dual-frequency (9.6 and 17.2 GHz) synthetic aperture radar (SAR), to evaluate high resolution (10 m) backscatter within a snow-covered tundra basin. Coincident in situ surveys at two sites characterize a generally thin snowpack (50 cm) interspersed with deeper drift features. Structure of the snowpack is found to be predominantly wind slab (65%) with smaller proportions of depth hoar underlain (35%). Objective estimates of snow microstructure (exponential correlation length; ), show the slab layers to be 2.8 times smaller than the basal depth hoar. In situ measurements are used to parametrize the Microwave Emission Model of Layered Snowpacks (MEMLS3&a) and compare against collocated SnowSAR backscatter. The evaluation shows a scaling factor () between 1.37 and 1.08, when applied to input of , minimizes MEMLS root mean root mean squared error to less than 1.1 dB. Model sensitivity experiments demonstrate contrasting contributions from wind slab and depth hoar components, where wind rounded microstructures are identified as a strong control on observed backscatter. Weak sensitivity of SnowSAR to spatial variations in SWE is explained by the smaller contributing microstructures of the wind slab.

**Keywords:** Snow, SWE, radar, SAR, tundra, Arctic

# 1. Introduction

Across the Northern Hemisphere, snow on the ground plays a critical role in climatological, hydrological and ecological processes, represents an essential freshwater resource, and influences natural hazards. Improved understanding of where, and in what ways, snow mass is changing is an important consideration for advancement of associated studies, including numerical weather prediction and hydrological forecasting, where spatially continuous observations with high temporal availability are desirable (Carrera et al., 2010; Bernier et al., 2012). Sparse in situmeasurement networks have made satellite remote sensing an attractive option to satisfy observational requirements for initialization and verification of land surface models within these forecast schemes, and in a more general sense, appear key to monitoring of global snow resources (Takala et al., 2011).

While progress has been made in the retrieval of snow cover extent from missions such as MODIS (e.g. Hall et al., 2010), retrieval of volumetric properties, including snow water equivalent (SWE), remain a challenge. Current space-borne methods rely on passive microwave radiometry, an approach inherently limited in spatial resolution and negatively impacted by spatiotemporal variations in snow microstructure (Kelly et al., 2003; Tedesco and Jeyaratnam, 2016). As a potential complement to existing methods, radar remote sensing is also sensitive to volumetric changes with the added advantage of higher spatial resolution to address uncertainty at scales relevant to snow physical processes (<100 m). Recent advancement in radar-based observation of terrestrial snow was motivated in large part by the Cold Regions High Resolution Hydrological Observatory (CoReH20) mission concept. While not selected as the European Space Agency (ESA) Earth Explorer 7, CoReH2O provided the impetus for both technical study on instrument design (Rott et al., 2010), and applied algorithm development (Chang et al., 2014; King et al., 2015; Lin et al., 2016).

Although the theoretical response of radar is well documented (Ulaby and Stiles, 1980), the observed relationship between backscatter and SWE, much like in the passive case, is indirect due to the strong influence of snow microstructure and basal surface interactions. The response of two snowpacks with the same SWE, but differing stratigraphic composition (for example, predominantly rounded grains typical of taiga snow; faceted grains typical of tundra snow), can produce contrasting volumetric interactions, governed in part by the nonlinear relationship amongst observing wavelength, microstructure, and backscatter. The use of traditional grain size (i.e. Fierz et al., 2009) to address this uncertainty has clear limitations, both in the level of confidence that can be placed in the subjective values, and their applicability in microwave modeling (Durand et al., 2008; Löwe and Picard, 2015). Instead, specific surface area (SSA; Domine et al., 2007) and correlation length (Mätzler, 2002) represent metrics that can be robustly estimated in the field (Gallet et al., 2009; Montpetit et. al., 2012; Proksch et al., 2015a), and applied in models to study snow-microwave interactions (Roy et al., 2013; Kontu et al., 2017; Sandells et al., 2017).

The use of airborne radar to determine first order sensitivity to SWE across basin domains remains relatively unexplored (see Yueh et al., 2009), with much of the recent focus given to plot-scale experiments (see King et al., 2015; Lemmetyinen et al, 2016; Lin et al., 2016). In this study, we introduce airborne dual-frequency (9.6 and 17.2 GHz) synthetic aperture radar (SAR) acquired with the ESA SnowSAR instrument within the Trail Valley Creek (TVC) research basin. The SnowSAR measurements are used to demonstrate backscatter diversity across a snow-covered tundra environment during a period corresponding closely with peak SWE. Concurrent in situ snow measurements, including objective estimates of microstructure, are used to identify and decompose backscatter contributions. We apply the Microwave Emission Model of Layered Snowpacks adapted to include backscattering (MEMLS3&a; referred to as MEMLS; Proksch et al., 2015b) to simulate perceived influences using field-based parametrizations. Extensive airborne lidar measurements facilitate spatial extrapolation of the snowpack and model analysis. The specific objectives of the study were to:

1. Introduce co-located snow property and airborne SAR (9.6 and 17.2 GHz) measurements to characterize spatial variability and discuss geophysical attribution.
2. Parametrize MEMLS using in situ measured snow properties to evaluate forward skill against SnowSAR backscatter and optimize inputs to improve modeled results.
3. Characterize SnowSAR and MEMLS sensitivity to observed variations in microstructure and discuss implications for future radar retrieval development.

# 2. Data and methods

2.1. Measurement area

Located 50 km northeast of Inuvik, Northwest Territories, Canada, the TVC research basin sits at the southern edge of the Arctic tundra (68°45'N, 133°39'W). Airborne and in situ measurements discussed as part of this study were collected within the primary 57 km2 basin, as well as in an adjacent subcatchment known as SikSik Creek between April 5 and April 13, 2013. Land cover in the study area is predominantly graminoid tundra (62% of total coverage) with patches of willow or alder shrub (25%) and isolated black spruce stands (12%). Figure 1A identifies regions of the latter two features, where shrub tundra contains standing vegetation 50 to 125 cm in height and forests are in excess of 300 cm (Marsh et al., 2010). The remaining open areas are characterized by short or ground covering vegetation (<50 cm in height), mineral earth hummocks, and rolling topography.

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Figure 1: Overview of the Trail Valley Creek (TVC) research basin with vegetation (A) and SnowSAR 17.2 GHz VV backscatter overlaid (B). Boundaries of the Upland tundra (UT) and SikSik (SS) study sites are outlined in red. The higher elevation UT site was generally free of shrub or forest vegetation while SS contains discontinuous open coverage and a range of vegetated features.

In situ sampling was focused in two areas, identified as SikSik (SS; 9.0 km2) and Upland Tundra (UT; 2.7 km2) in Figure 1. The larger SS site contained a diverse set of sub-basin elements including rolling tundra to the north, a transecting valley, and isolated forest stands. As in the larger TVC domain, substantial portions were free of tall vegetation (65.2%, 5.9 km2) with small clusters of shrub tundra (26.8%, 2.4 km2), and isolated black spruce (8.0%, 0.7 km2). In contrast, the smaller UT site was homogeneous, sitting atop a windswept plateau (146±6.0 mASL) composed of flat (1.3±1.0°) and open terrain (97%, 2.7 km2). Although no vegetation was visible above the snow surface, shrubs were known to be present in isolated quantities based on previous surveys.

Conditions throughout the study were consistent, with cold air temperatures (-20.9±5.3°C) and trace precipitation recorded at local meteorological stations. Typical of a wind-exposed environment, accumulated snow is shallow across much of the basin (<50 cm), with deeper areas associated with topography or vegetation where redistributed snow is deposited as substantial drifts (Essery et al., 2004). Observed wind speeds within the basin were on average 6.6 ms-1, sufficient to support sustained transport and subsequent drift development. In previous years, snow drifts have been estimated to occupy as much as 17% of the basin, representing a considerable storage of freshwater (Marsh et al., 1994, 2010).

2.2. In situ snow property measurements

To establish an understanding of spatial variability in the tundra snow properties, a distributed set of transect and pit-based measurements were completed. Bulk (i.e. vertically integrated) snow properties including depth, density, and water equivalent were sampled at both sites to produce a reference dataset (Figure 2). Geo-located measurements of snow depth were made with Snow-Hydro GPS Snow Depth Probes (Sturm and Holmgren, 1999) along multi-kilometer linear transects. Individual depth measurements were separated by 2 operator paces, roughly equating to a sampling interval of 3 m. In total, 15,436 measurements were collected, covering an approximate distance of 25 km. Along the same transects, 347 snow cores were extracted using a 30-cm2 corer, commonly known as the ESC-30, to estimate bulk density and SWE. The slower-to-complete snow cores were spaced at a broad interval of 200 m in the interest of creating comparable coverage with the reference snow depths.

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Figure 2: Transect, pit, and airborne snow measurements at the Upland Tundra (UT) and SikSik (SS) sites. Green shaded areas indicate areas of shrub or forest vegetation greater than 50 cm in height. The background shows lidar derived snow depths detailed in section 2.3.2.

Vertical variations in density, temperature, stratigraphy, and microstructure were characterized with a spatially distributed set of 11 snow pits (Figure 2). At each location, continuous vertical profiles of snow density were sampled with a 100-cm3 box-style cutter and hanging scale. Uncertainty of box-style density measurements have been approximated at 8% (Proksch et al., 2016) but may be higher due to the use of a hanging scale. Vertical temperature profiles were measured using a handheld digital thermometer with accuracy of ±0.5°C. Stratigraphy was manually interpreted from the pit face via finger hardness and visual inspection. Snow samples were extracted from each of the identified layers to classify by grain type with a 2-mm comparator card and field microscope (Fierz et al., 2009).

Although traditional grain size estimates were made, the values are considered subjective, and represent a substantial uncertainty in parameterization of microwave models (Durand et al., 2008). To objectively characterize microstructure, measurements of SSA were made with the shortwave InfraRed Integrating Sphere system (IRIS; Montpetit et al., 2012). The IRIS is a laser-based device adapted from the original infrared reflectometry work of Gallet et al. (2009), after Domine et al., (2006). Measurements of snow reflectance () at 1300 nm are physically linked to SSA following Kokhanovsky and Zege (2004):

(1)

where is an escape function set at 1.26 based on geometry of the system, is a shape factor set at 4.53 corresponding to spheres (Picard, et al., 2009), and isthe absorption coefficient of ice (Wiscombe and Warren, 1980). IRIS outputs a photodiode measured voltage which can be converted to reflectance when calibrated against reference targets (Montpetit et al., 2012). Reference target calibrations were performed at each snow pit prior to vertical sample profiles at a 3-cm resolution.

2.3 Airborne measurements

2.3.1. Airborne SAR measurements

SnowSAR, an airborne radar system, was deployed to collect measurements coincident with the in situ campaign. This dual-frequency (9.2 and 17.2 GHz), dual-polarization (VV and VH) synthetic aperture radar (SAR), was developed to emulate the proposed configuration of CoReH2O (Rott et al., 2010; Di Leo et al., 2015). Mounted aboard a Cessna 208 Grand Caravan, SnowSAR was flown at a nominal altitude of 1220 mAGL, resulting in a ground-projected swath of 400 m, and nominal spatial resolution of 2 m. Flight missions were executed over two days (April 8 and 9, 2013) with 4 flight lines intersecting each of the TVC sites (Figure 3).

Prior to analysis, SnowSAR radar cross-sections () were normalized to the ground projection area, taking into account local variations in slope:

(2)

where A is the nominal pixel area (4 m2) and is incidence angle. Radar incidence angles were evaluated normal to the observed surface (i.e. local angle) using aircraft attitude data and a snow-free digital elevation model (DEM) introduced in Marsh et al. (2010). SnowSAR measurements with outside of the anticipated range of 35 to 45° were removed from further analysis to minimize terrain associated uncertainty. Given the nominal mid-swath angle of 40° and low variability in the observed terrain, a vast majority of the data was retained in this process, leaving no appreciable impact on total coverage. To minimize the influence of radar speckle and geolocation errors, SnowSAR measurements were aggregated to a spatial resolution of 10 m by taking the mean of an equivalent moving window. An extended description of the SnowSAR system and details on the radiometric calibration procedure can be found in Di Leo et al. (2015).

2.3.2. Airborne lidar measurements

The University of Alaska Fairbanks (UAF) airborne lidar system was flown on April 6, 2013 to produce a spatially continuous surface of snow height within the TVC domain. The UAF system is a Riegl LMS-Q240in with a 905 nm wavelength and 10 kHz sampling rate (Johnson et al., 2013). Flown at 500 mAGL, typical vertical errors are on the order of ±10 cm, however, extreme cold during the acquisition period led to greater than normal uncertainty. While this uncertainty is associated with the absolute magnitude of the snow surface elevations, spatial patterns and characterization of pixel-scale variability remain accurate, particularly in open areas. For estimates of snow depth, lidar elevations of the snow surface were gridded and differenced from the snow-free DEM (Marsh et al., 2010) and then aggregated to the SnowSAR resolution (10 m). The use of lidar derived snow depth is differentiated from the reference dataset () by subscript here forward ().

Measurements of at each site were compared against to determine a root mean squared error (RMSE; Table 1). The comparisons agreed closely in their mean and standard deviation, but had slightly larger RMSE within the larger SS site. As a bias correction, the SDlidar data was offset in translation by an amount that minimized the difference with specific to each site.

Table 1: Comparison of the lidar () and in situ()snow depths in open tundra coverage. Analysis was completed separately for the SikSik (SS) and Upland tundra (UT) sites. Errors described as root mean squared error (RMSE).

|  |  |  |  |
| --- | --- | --- | --- |
|  | (cm) | (cm) | RMSE (cm) |
| SS | 52.6±15.3 | 55.4±18.2 | 11.2 |
| UT | 58.4±18.0 | 55.6±17.5 | 8.5 |

2.3 Model configuration

In an effort to link in situ and airborne datasets, as well as diagnose causal mechanisms of interaction, we employ a microwave radiative transfer model. Originally developed for microwave emission modelling (Wiesmann et al., 1999), MEMLS was recently extended to include backscatter for active applications (Proksch et al., 2015). Self-described as a model of intermediate complexity, MEMLS uses exponential correlation length () as a microstructural quantity in computation of the scattering coefficient (Matzler, 1998). As a result, objective lab-based measurements or field-based approximations of can be used to create meaningful input. In practice, while many of the required inputs were satisfied with snow pit measurements (Table 2), additional steps were necessary to derive from the IRIS measurements, as well as constrain unknowns including basal reflectivity.

Table 2: MEMLS model inputs as modeled or derived from the available TVC datasets. Salinity and liquid water were set as constant 0 values in all simulations to reflect the dry terrestrial nature of the snowpack.

|  |  |  |  |
| --- | --- | --- | --- |
| **MEMLS parameter** | **Value [unit]** | **Method** | **Reference** |
| **Snow**: |  |  |  |
| Layer thickness | [cm] | Traditional/Pit | Fierz et al., 2009 |
| Density ( | [kg m-3] | Traditional/Pit | Fierz et al., 2009 |
| Temperature | [K] | Traditional/Pit | Fierz et al., 2009 |
| Liquid water content | 0 [%] | Traditional/Pit | Fierz et al., 2009 |
| Exp. Corr. Length. ( | [mm] | IRIS, Eqn. 3 | Montpetit et al., 2012 |
| **Basal surface**: |  |  |  |
| Incidence angle (θ) | 35-45 [°] | DEM, Eqn. 2 | Di Leo et al., 2015 |
| Reflectivity () | 0-1 | Modeled | Wegmüller and Mätzler, 1999 |
| Specular reflect. () | 0.75 | Constant | Proksch et al., 2015b |

2.3.1 MEMLS snow property inputs

Configured as a stack of horizontal layers, MEMLS requires characterization of temperature, liquid water content, density, thickness, salinity, and to compute forward estimates of (Table 2). Among the snow property inputs, is the most sensitive in dry snow, particularly at high frequencies where volume scattering dominates (Durand et al., 2008; Sandells et al., 2017). Although was not measured directly (i.e. X-ray tomography), it was possible to approximate input from the IRIS measurements following Debye et al. (1957) and Matzler (2002):

(3)

where is optically equivalent diameter () and is ice volume fraction (. Snow pit measurements of density () were used to estimate along with a constant ice density () of 917 kg m-3.

Using IRIS-based estimates of and direct measurements of the remaining snow inputs, three configurations of increasing complexity were generated to evaluate forward modelling skill at the point scale. At the lowest level of complexity, a thickness weighted mean of each snow parameter was used to construct a 1-layer representation. Increasing in complexity, the snowpack was abstracted as two layers by dividing the volume into contrasting wind slab and depth hoar components based on snow pit descriptions of stratigraphy. Finally, an n-layer configuration was created whereby all available IRIS measurements were treated as separate layers. In all configurations, salinity and volumetric water content were set to 0, reflecting the cold (-20°C on average) and dry terrestrial condition of the observed snowpack.

2.3.2 MEMLS snow-ground reflectivity

MEMLS requires input of reflectivity at the snow-ground interface to account for background contributions. As snow-free measurements were unavailable to provide a reference, first order approximations of basal surface reflectivity were made with the empirical model of Wëgmuller and Mätzler (1999):

(4)

(5)

where is the observing wave number, is the standard deviation of surface height, and is incidence angle. Given that the incident wave originates in snow, was revised to account for variations in refractive index using in situ estimates of bulk snow density (). Similarly, was adjusted to account for refraction within the snowpack prior to interaction at the snow-ground interface (Snell’s Law).

Fresnel coefficients ( were calculated with a constant relative permittivity of 4 + 0.5i based on soil sensors buried 5 cm below the surface at each study site. Although spatial variations in permittivity were probable, the influence was assumed to be conservative given consistent soil temperatures below -15°C, adequate for complete freezing of the organic material (Mironov et al., 2010). Collecting distributed estimates of was not possible during the winter field campaign, requiring treatment as a free parameter. The evaluated range of was limited to reflect previous tundra-like parametrizations (see Montpetit et al., 2013) and optimized at locations were all other inputs could be constrained (e.g. at snow pits). Given that MEMLS requires input of specular reflectivity () to estimate backscatter, modeled total reflectivity () was scaled by a constant factor to produce a proportional estimate (; Proksch et al., 2015).

# 3. Results

3.1. Observed snow conditions

3.1.1. Snow depth, density and water equivalent

Regardless of differences in landscape, mean at the two sites differed by only 3.2 cm when all in situ measurements were considered (Table 3). Agreement was improved when depths were limited to those in open tundra, with means of 51.8±16.5 and 50.9±17.4 cm (±1σ), at the UT and SS sites, respectively. Deeper snow and increased variability were associated with shrub tundra and forest, with means of 61.8±22.7 cm and 78.8±19.4 cm, at the SS site. Although a smaller number of measurements were collected within areas of standing vegetation (28.6% of total measurements), the sampled proportions were comparable to the SS site fractional coverage (34.8% of total coverage). Relative contributions to total snow mass were evaluated using the reference snow core measurements (). Despite having the lowest mean , open tundra was identified as the principle storage at both sites due to its spatial predominance (Table 3). As a result, a narrowed focus on open tundra where biophysical complexity is minimized can be shown to cover both the largest land cover fraction (>71%), as well as a high proportion of the basin-wide snow mass (>60%).

Table 3: In situ measured snow depth (), density (), and water equivalent () at the SikSik (SS) and Upland tundra (UT) sites separated by land cover type.

|  |  |  |  |  |  |  |  |  |  |
| --- | --- | --- | --- | --- | --- | --- | --- | --- | --- |
|  | **Upland Tundra** | | | |  | **SikSik** | | | |
|  | **All** | **Tundra** | **Shrub** | **Forest** |  | **All** | **Tundra** | **Shrub** | **Forest** |
| **Coverage (%)** | N/A | 96.3 | 3.7 | 0.0 |  | N/A | 65.2 | 26.8 | 8.0 |
| **Measurements (%)** | N/A | 99.6 | 0.4 | 0.0 |  | N/A | 71.4 | 22.6 | 6.0 |
| **Mean (cm)** | 51.8 | 51.8 | 65.1 | N/A |  | 55.0 | 50.9 | 61.8 | 78.8 |
| **Mean (kg m-3)** | 266 | 266 | 268 | N/A |  | 242 | 241 | 245 | 238 |
| **Mean (mm)** | 135 | 135 | N/A | N/A |  | 118 | 106 | 141 | 180 |
| **Total mass (107 kg)** | 3.8 | 3.6 | 0.2 | 0.0 |  | 12.1 | 7.3 | 3.7 | 1.2 |

Focusing on open tundra, probability density functions (PDFs) of SWE were generated to evaluate inter-site variability (Figure 3). Here, individual estimates of SWE were calculated as the product of and the nearest neighbor snow core density (), leveraging strong covariance with depth to increase the number of available estimates (Sturm et al., 2010). At the UT and SS sites, similar unimodal, positively skewed distributions characterized a broad range of SWE (>320 mm) with means of 138±50 and 122±45 mm, respectively. The high degree of observed variability is typical of a tundra environment where depressions between tussocks and hummocks can produce decimeter variations in SWE at sub-meter scales (Sturm and Holgren, 1994). Much larger increases in the PDF tails however, are not accounted for by microtopography, suggesting additional influence at larger length scales.

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Figure 3: Probability density functions (PDFs; A) and length scales of variability (B) for open tundra SWE at the Upland Tundra (UT) and SikSik (SS) sites.

To quantify spatial scales of variability, estimates of autocorrelation (Moran’s I; Moran, 1950) were calculated for in situ pairs of SWE at lag distances of up to 200 m (Figure 3B). From this, it can be shown that bulk SWE varied at length scales greater than the aggregate SnowSAR resolution across the basin. The rate of decay however, was unequal between sites, as noted by the shorter UT length scale (correlation of 0.5 at 41 m lag distance). Unexpectedly, this implies that larger units of coherence were present within the diverse SS landscape (correlation of 0.5 at 77 m lag distance).

Given the limited areal support of the in situ measurements, it was advantageous to evaluate the length scale inequality using the lidar-derived estimates of snow depth (). Nearest neighbor measurements of were again used to estimate SWE, this time as a product of the spatially continuous . Areas contributing to the positive distribution tails () were found to be characteristic of drifted snow, oriented perpendicular to the prevailing northwestern winds and strongly associated with leeward slopes (Figure 4). Drifts with area greater than the computed length scales of variability (4 pixels or 1600 m2) occupied 11.5% of the UT domain, representing a substantial component of total storage. In comparison, similar drifts were found to represent only 6.8% of the larger SS domain. Increased retention within drifts accounts for the higher mean SWE and shorter length scale at the UT site where these features punctuated the otherwise thin snowpack with increased frequency.

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Figure 4: Lidar derived snow water equivalent (Left) and digital elevation model derived slope (right) and at the UT site. The locations of large snow drifts are outlined in black.

3.1.1. Snow stratigraphy and microstructure

Comprehensive evaluation of the observed radar interactions required characterization of the internal structure of the open and drifted snowpack. Common amongst the 11 snow pits used to describe vertical heterogeneity were contrasting slab and hoar layers, a function of sustained wind action and temperature gradient metamorphism, respectively (Colbeck, 1983). Despite generalized structure, relative quantities of each component varied across the TVC domain leading to a wide range of snow pit configuration (Table 4).

Table 4: Thickness weighed bulk snow pit properties measured within the swath of the SnowSAR measurements. Values of were calculated using Equation 3.

|  |  |  |  |  |  |  |
| --- | --- | --- | --- | --- | --- | --- |
| **Site** | **Pit** | **Depth** | **Density** | **SWE** | **Depth Hoar** |  |
| (SS/UT) | (#) | (cm) | (kg m-3) | (mm) | (%) | (mm) |
| SS | 1 | 33 | 208 | 67 | 29 | 0.345 |
| SS | 2 | 44 | 295 | 127 | 40 | 0.231 |
| SS | 3 | 36 | 276 | 97 | 35 | 0.185 |
| SS | 4 | 43 | 290 | 122 | 52 | 0.238 |
| SS | 5 | 38 | 266 | 99 | 61 | 0.272 |
| SS | 6 | 79 | 272 | 213 | 32 | 0.193 |
| SS | 7 | 35 | 340 | 116 | 23 | 0.177 |
| UT | 8 | 37 | 224 | 81 | 20 | 0.198 |
| UT | 9 | 39 | 213 | 81 | 40 | 0.289 |
| UT | 10 | 41 | 240 | 96 | 28 | 0.212 |
| UT | 11 | 57 | 275 | 152 | 20 | 0.193 |
|  | **Average** | 44 | 264 | 114 | 35 | 0.230 |

Originating at the air-snow interface, slab layers were characterized by their higher density (280±43 kg m-3) and wind-pulverized grains (0.25-0.5 mm diameter). On average, these layers composed 67% of the snowpack volume, with as many as 6 independent layers representative of sequential wind or snowfall events. Once buried, slab layers exhibited incipient kinetic faceting. Density of the faceting slabs remained comparable to the adjacent surface features, reflecting a hardened slab-to-hoar conversion process referred to as indurated (Derksen et al., 2009). At the base of the pack, depth hoar composed the remaining 35%, easily distinguished by its unconsolidated structure, lower density (222±36 kg m-3), and large cup shaped grains (4-6 mm). Widespread aggregation of the depth hoar into large chained units and the presence of prismatic hoar were consistent indicators of the advanced metamorphic state of the basal layer (Trabant and Benson, 1972).

Persistent winds acted as a strong control on depth, scouting open areas and redistributing accumulation downwind as drifts (Figure 4). As a result, increases in snow pits thickness were commonly associated with additional or expanded slab layers rather than strong variations in depth hoar thickness. An example of this slab dominated change is illustrated in Figure 5 where on a leeward slope, total thickness increased by 30 cm over the local mean, but depth hoar remained a comparatively small component (23% of total thickness). Amongst the snow pits, the maximum depth hoar thickness was 25 cm, similar to the height of variations in microtopography.

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Figure 5: Near infrared photograph of snow pit 6 within the SS domain (Left) where basal depth hoar is easily distinguished by texture due to its unconsolidated structure. Microstructure estimates from IRIS, shown as , were largest near the basal surface.

Availability of IRIS measurements provided a means to objectively evaluate stratigraphic differences in microstructure as related to the identified snowpack features. Estimates of exponential correlation length () derived from IRIS (see Section 2.2) showed a wide range of variation with both layer type and measurement height (Figure 6). Composing a bi-modal distribution, the 198 estimates in open tundra ranged from 0.044 to 0.482 mm. Measurements taken near the snow surface contained little variability (0.131±0.045 mm), consistent with qualitative snow pit descriptions of homogeneous, fine grained structure. Increases in departing from the air-snow interface corresponded with observed faceting of the buried slab layers. These transitionary forms were limited to less than 0.241 mm despite clear evidence of sustained kinetic growth. In strong contrast, depth hoar was on average 2.8 times greater in than the overlying slab with a mean of 0.360±0.066 mm.

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Figure 6: Exponential correlation length () within the 11 snow pits. Probability density functions (PDFs) show difference between slab or depth hoar layer classified measurements. Estimates of were derived from IRIS measurements as detailed in Section 2.2.

As a whole, the measurements in Figure 6 describe an inverse relationship between and height above the basal surface. Thermodynamically, slab layers driving this relationship are increasingly likely to remain close to their deposited wind rounded state as thermal gradients are depressed with height above the basal surface. Consequently, increasing proportions of slab found in deeper or drifted snow have the effect of driving mean downwards when depth hoar remains comparatively thin. From a scattering perspective, small changes in shallow snowpack (<30 cm) may produce strong variations in backscatter where predominantly large microstructures of the depth hoar are involved. Inversely, where changes are slab dominated (i.e. drifted snow) sensitivity to SWE may be diminished because of the reduction in contributing microstructure.

3.2. MEMLS validation and optimization

In this section we generate a representative suite of MEMLS parametrizations and simulations to evaluate model skill and sensitivity when applied to the observed tundra environment. MEMLS is used in forward simulation to compare against co-located SnowSAR measurements at the location of each snow pit. This serves to gauge model skill when detailed snow microstructure inputs are available, and to address the influence of unknowns, such as basal surface reflectivity.

3.2.1. Forward model parametrization of

As has been established in passive microwave radiometry, scaling of field-derived (Montpetit et al., 2013; Royer el al., 2017) and modeled (Sandells et al., 2017) microstructural quantities is often necessary to minimize forward model bias. Scaling factors applied in these studies vary by model, measurement method, and climate class of snow, but in general offer a practical approach to establish representational equivalence between model inputs and real-world observables. To assess the need for scaling in the active case, a factor between 0.1 and 2.0 was applied to snow pit estimates of . An optimized factor ()was determined by minimizing the mean absolute error (MAE) between SnowSAR () and MEMLS () for all available snow pits:

(6)

(7)

where snow property inputs ( other than were held constant and are in linear units. Only the higher frequency 17.2 GHz SnowSAR measurements were considered in this evaluation to limit influence of the snow-ground interface. Background reflectivity was approximated with *h* of 1.0 cm, and additional input parameters detailed in section 2.3.2. Given that the cross-polarized outputs of MEMLS are scaled quantities of the co-polarized (Proksch et al., 2015), they add no independent information and were therefore not considered in the optimization.

Figure 7 introduces modeled MAE as a function of in 1-, 2-, and n-layer configurations of MEMLS. In each case, errors were minimized with positive scaling of (i.e. larger than the IRIS estimated values), although the retrieved was dependent on layer configuration. As a single layer, global MAE was minimized with scaling of 1.37. While inherently describes a single point of convergence, optimal scaling of individual 1-layer snow pits ranged from 1.13 to 1.73 with a standard deviation of 0.17. Divided into separate slab and hoar components as a 2-layer configuration, was reduced to 1.08. The change in was accompanied by a smaller standard deviation of 0.09, signifying stronger agreement amongst the individual snow pit optimizations. The n-Layer results were comparable, with of 1.09, and a standard deviation of 0.11. Given the power relationship between and scattering, the difference between configurations when depth hoar is considered as a separate layer is important, as reflected in the 21% reduction of . The close agreement between the 2- and n-layer models is viewed positively, as it suggests that a 2-layer simplification is sufficient for tundra snow.

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Figure 7: Optimization of exponential correlation length () at 17.2 GHz (A) by way of scaling () in 1-, 2-, and n-layer configurations of MEMLS. Soil roughness () sensitivity analysis at 9.6 GHz (B) using the reflectivity model of Wegmüller and Mätzler (1999).

Microstructure optimized estimates of MEMLS backscatter at 17.2 GHzare compared against the co-located SnowSAR measurements in Figure 8A, using the derived of 1.37, 1.08 and 1.09 for 1-,2-, and n-layer cases, respectively. In all three configurations correlations are strong (0.71<R<0.73), demonstrating forward skill in the snow pit abstractions. Root mean square error (RMSE) was largest in the 1-layer configuration (1.15 dB), corresponding with the largest adjustment. The 2- and n-layer configurations were of reduced RMSE at 0.80 dB and 0.98 dB, respectively.

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Figure 8: Comparison of optimized MEMLS simulations and SnowSAR measured at each snow pit in 1-, 2- and n-layer configurations. Adjusted () input of listed for each configuration in addition to the optimized soil input (*h*).

3.2.2. Forward model evaluation of *h*

To evaluate background sensitivity and determine optimized input, standard deviation of surface height (*h*) in the soil reflectivity model was assessed using values between 0.01 and 4.00 cm. As in the microstructure evaluation, simulations were completed in 1-, 2-, and n-layer configurations for each of the available snow pits and compared with SnowSAR measurements to calculate MAE. Background sensitivity was initially evaluated at 9.6 GHz to maximize penetration of the tundra snowpack.

Close agreement was noted between the different snow layer configurations in the evaluation of *h* (Figure 7B). This was consistent with higher penetration at 9.6 GHz, indicating reduced dependence on the properties of the snowpack. Optimized minimums were identified at 1.37, 1.28, and 1.21 cm in the 1-, 2-, and n-Layer configurations, although each lacked clear convergence. The reason for this was evident in the individual snow pit simulations, where optimal values of *h* spanned a large range of 0.30 to 3.82 cm, with a standard deviation of 1.09 cm. Using the retrieved standard deviation to bound potential influence, we find that *h* contributes on average 2.9 dB of variability amongst the 1-layer snow pit simulations, or in the worst case, 3.9 dB if the full input range of *h* is considered (i.e. 0.01 to 4.00 cm). Evaluating the same standard deviation range of *h* at 17.2 GHz, a much lower sensitivity of 1.3 dB is found, or 2.5 dB if the full range of input is considered.

Comparison of the *h* optimized MEMLS simulations and 9.6 GHz SnowSAR measurements shows little difference between the three snow layer configurations (RMSEs of 0.86, 0.90, and 0.90 dB), along with lower correlations (0.72, 0.69, 0.67; Figure 8B). However, the moderate correlations at 9.6 GHz were largely driven by the two strongest SnowSAR backscatter values which were both located in the north of the SS site. As a whole, the sensitivity to *h* demonstrates considerable, and spatially variable, influence from the underlying surface at 9.6 GHz, with a much narrower influence at 17.2 GHz.

3.3. Comparison of observed and MEMLS simulated backscatter

3.3.1. SnowSAR backscatter response

Although previous airborne studies have identified direct relationships between backscatter and SWE (i.e. Yueh et al., 2009; Chang et al., 2014), we anticipate it necessary to account for microstructure because of the strong contrast between slab and depth hoar. The TVC backscatter response, irrespective of microstructure, is introduced in Figure 9 where SnowSAR is delineated by site and frequency in open tundra. Within the smaller SS domain, 17.2 GHz measurements were normally distributed with a mean of -9.1±0.5 dB. Higher 17.2 GHz backscatter and increased variability were found within the larger SS site with a mean of -8.5±0.8 dB. The 9.6 GHz response was expectedly lower at both sites due to its reduced dependence on the snowpack, with means of -14.8±0.6 and -14.3±0.9 dB. Much like in the higher frequency case, a wider range at 9.6 GHz was captured within the UT domain.

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Figure 9: Distributions of SnowSAR at 9.6 (Black) and 17.2 GHz (Grey) within the SikSik (SS) and Upland Tundra (UT) sites. Measurements have been limited to those located in open tundra.

The SnowSAR response to snow mass is introduced in Figure 10 where dual-frequency is shown with co-located estimates of SWE. Overlaid boxplots summarize sensitivity across the first standard deviation of SWE in 10 mm increments. At the UT site the 17.2 GHz median values describe a weak change in backscatter with increasing SWE (+0.13 dB over 108 to 206 mm). The interquartile range (IQR) of 17.2 GHz UT boxplots were on average 0.61 dB, larger than the observed trend with SWE. Similar behaviour was identified at the SS site, where a small increase in 17.2 GHz (+0.3 dB) was characterized across the standard deviation range of SWE (between 95 and 165 mm). A larger IQR (0.90 dB on average) was associated with the SS backscatter, particularly at lower SWE. At 9.6 GHz, decreases in median were noted at both sites (-0.24 dB at UT and -0.36 dB at SS), although as in the higher frequency case, the change was small in comparison to the IQR (0.77 dB at UT and 1.13 dB at SS).

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Figure 10: Relationship between SnowSAR and SWE at the SS and UT sites (Black dots and boxplots). Grey shaded area shows the MEMLS 1-layer simulated range of variability based on snow pit measurements. The solid and dashed lines are for of 0.383 and 0.247 mm, representing the first standard deviation of measurements.

3.3.2. MEMLS sensitivity to microstructure and SWE

Given the flat response of SnowSAR, it was clear that factors other than snow mass were strong contributors, requiring additional information to decompose. Two MEMLS sensitivity studies were generated to address this ambiguity, contrasting the response of a microstructure invariant snowpack against stratigraphic variability similar to the observed slab-hoar dichotomy. These simulations attempted to characterize the effect of wind rounding and temperature gradient metamorphosis on backscatter, as controlling processes within the drifted and thin snowpack elements. In all simulations, MEMLS input was derived from the layer weighted properties of the snow pits, was adjusted with , and a radar incidence angle of 40° was applied.

As a first step, field-parametrized 1-layer simulations were completed to establish bounds on expected variability as related to microstructure and SWE. The grey shaded areas in Figure 10 show the range of simulated relationship between SWE and when is parametrized according to the first standard deviation of observed microstructure (0.247 to 0.383 mm; Table 4). Snow density in the simulations was set at 264 kg m-3 (mean from Table 4) and background contributions were optimized according to the process in Section 3.2.2. Variability of SnowSAR was well contained by the 1-layer simulations, but in each case, the simulations show a strong positive relationship absent in the observations. Between 50 and 250 mm SWE, MEMLS increased by 5.2 dB at 17.2 GHz and 2.9 dB at 9.6 GHz, using the observed mean of 0.315 mm. Sensitivity to increasing was comparable, as shown by average differences of 4.6 dB at 17.2 GHz and 3.3 dB at 9.6 GHz between the 0.247 and 0.383 mm simulations.

While the 1-layer simulations indicated strong microstructural sensitivity, the invariant nature of with respect to SWE was uncharacteristic of the observed environment. In particular, the simulations failed to address the impact of accumulating slab, where the contributing microstructures were small relative to the mean (Figure 4). A set of 2-layer simulations were completed to evaluate backscatter response to increasing slab thickness between 5 and 75 cm (14 to 213 mm SWE) when underlain with the observed mean depth hoar of 15 cm (34 mm SWE). Density of the slab and hoar layers were set at 288 and 225 kg m-3, according to their respective means. Slab microstructure was parametrized across a range of equivalent to the observed standard deviation (0.102-0.198 mm; Figure 6) and depth hoar was assigned the mean of 0.388 mm. Additional control simulations were completed where both layers were assigned equal according to the depth hoar mean (0.388 mm).

Figure 11 shows the result of the 2-layer simulations where large increases in SWE produced small increases in when slab was the driver of change. Fractional slab composition increased from 25 to 84%, representative of the full range of snow pit observations (Table 4). The addition of 70 cm of slab with of 0.150 mm increased by 0.25 dB at 17.2 GHz and 0.23 dB at 9.6 GHz, over the constant depth hoar contributions. Sensitivities of the 2-layer simulations were weak with maximums of 0.06 dB per cm SWE at 17.2 GHz and 0.04 dB per cm SWE at 9.6 GHz with the largest slab parametrization of 0.198 mm. The smallest slab of 0.102 mm produced no sensitivity across the 200 mm range of evaluated SWE at either frequency. In contrast, sensitivities of 0.27 and 0.18 dB per cm SWE at 17.2 and 9.6 GHz were found when upper layer was replaced with depth hoar, similar to the response of the 1-layer simulations.

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Figure 11: MEMLS 2-layer sensitivity to increasing wind slab composition in a snowpack with total SWE between 50 and 250 mm. A constant depth hoar layer of 15 cm (34 mm SWE) was maintained while increasing thickness of the wind slab between 5 and 75 cm. The grey shaded area shows the response within the standard deviation range of snow pit observed slab . The top panel shows the thickness weighted mean for each of the slab parametrizations as a function of total SWE. Blue lines show control simulations where microstructures of both layers were assigned of 0.388 mm.

The weighted mean of the 2-layer input decreased by approximately 42% (0.317 to 0.183 mm) with the mean slab parametrization ( 0.150 mm; Figure 11), showing the potential for drastic change in bulk microstructure where drifted snow is involved. At the mean fractional composition of 65% (or 114 mm total SWE), the difference between the slab (0.150 mm ) and control simulations (0.388 mm ) were 3.5 dB at 17.2 GHz and 2.3 dB at 9.6 GHz. In contrast, variations in slab (0.102 to 0.198 mm) at the same fractional composition produced much smaller differences of 0.57 and 0.29 dB at 17.2 and 9.6 GHz, respectfully. Thus, if abstracted as a single layer, backscatter and sensitivity will be overestimated unless covariance of and snow mass is considered in wind controlled environments.

# 4. Discussion and conclusions

As generalized from snow pits, the TVC volume was principally composed of wind slab and indurated layers (65%), with smaller proportions of underlying depth hoar (35%). Their fractional composition however, diverged with increasing total depth where changes were slab dominated, consistent with a wind exposed tundra environment (Domine et al. 2012; Rutter et al., 2016). The restricted variation in depth hoar thickness is attributed to the small range of tussock micro-topography and its finite ability to trap early season accumulation (Benson and Sturm, 1993, King et. al. 2015). Lacking tall standing vegetation, larger variations in the open tundra were realized as substantial drift features where topographic relief was sufficient to accommodate deposition of additional wind transported snow (Essery and Pomeroy, 2004). Microstructures of the principal slab layers, estimated as , were on average 2.8 times smaller than the depth hoar, despite evidence of sub-surface faceting and slab-to-hoar conversion (Figure 6). Thus, spatial increases in depth were generally not of equivalent bulk microstructure, but rather were characterized increasingly by larger proportions of small, wind-pulverized or transitional grain types. Additionally, depth of the drifted snow was sufficient to decouple the basal interface from air temperature (Taras et. al. 2002, Slater et. al, 2016), thermodynamically differentiating these features from the predominantly thin, heavily faceted, open tundra snowpack.

The observed snowpack conditions had complex implications for scattering where contributions from increasing depth (increasing scattering) and slab microstructure (decreasing scattering) were in apparent compensation, inferring nonlinearity in the expected radar response. Forward MEMLS simulations, proposed to decompose these complexities, were able to replicate co-polarized SnowSAR measurements (9.6 GHz and 17.2 GHz) with RMSE of less than 1.1 dB (Figure 8). To do so, optimization of field-measured microstructural quantities and soil unknowns was required. An optimized scaling factor, , for IRIS-based inputs of exponential correlation length (), was shown to be configuration dependent, where separate representation of the slab and hoar layers greatly reduced the need for adjustment at 17.2 GHz (from 1.37 to 1.08). Chang et al., (2014) reported similar findings with multilayer QCA/DMRT and SnowSAR, also attributing the reduced simulation bias to improved representation of the basal snow layer. Positive scaling of at the snow pits is in agreement with previous active and passive studies of MEMLS (Brucker et al. 2011; Montpetit, et. al. 2013; Lemmetyinen et. al, 2015). In particular, the 1-layer optimization agrees closely with Montpetit, et. al. (2013) who reported of 1.3 with IRIS derived microstructure. In general, scaling and sensitivity testing of highlighted the contrasting importance of the two layer types, where depth hoar in the observed tundra environment scatters disproportionally to its limited SWE or thickness.

Evaluation of background sensitivity at the same snow pits, via standard deviation of surface height (*h*), showed sensitivity at 9.6 GHz (2.2-4.1 dB) to be nearly twice that of 17.2 GHz (1.3-2.5 dB). Although snow volume contributions were expected at both frequencies (Rott et. al. 2010), sensitivity to *h* confirmed that the evaluated snow pits were of insufficient depth to fully decouple spatial variations in the background signal. A lack of convergence in Figure 7B supports this finding at 9.6 GHz, showing high diversity in the optimized background reflectivity, and general insensitivity to changes in snow configuration. Conversely at 17.2 GHz, the range of SnowSAR measurements (4.2 dB) was much larger than the anticipated background sensitivity, allowing conclusion that the snow volume was the substantive contributor to spatial backscatter variability. In future study, the acquisition of early season measurements may provide a means to constrain background contributions including soil and vegetation.

Apparent insensitivity of SnowSAR backscatter to increasing SWE confirmed that factors other than snow mass were robust contributors across the basin (Figure 10). Using 1-layer MEMLS simulations, it was demonstrated that the observed range of was capable of driving backscatter variations comparable to the full range of SnowSAR (Figure 10). The simulated positive relationship between SWE and backscatter was predicated on an assumption of spatially invariant microstructure (i.e. static input), an undoubtedly invalid condition for both sites. Extended 2-layer simulations used to conceptualize microstructural variability showed that slab dominated changes in snow volume contributed little to total backscatter over the already strong depth hoar and background signals. The modeled behavior indicates a shift from thermodynamic to wind process control on backscatter, where areas of thin and heavily faceted snowpack scatter comparatively to substantially thickener slab dominated snowpack. This shift may explain the flat response of the SnowSAR backscatter where depth dependent variations in microstructure counter the expected increase in backscatter.

Passive microwave studies have demonstrated that physically based snow models, when carefully treated, can provide microstructural information needed to improve model simulations and enable retrieval of SWE (Kelly et al., 2003; Langlois et al., 2012; Sandells et. al. 2017; Kontu et al., 2017). In the case of TVC, and other tundra environments, a key consideration to decoupling the microstructural signals will be the ability of physically based snow models to characterize the relevant snow processes of wind redistribution and temperature gradient metamorphosis. Given that satellite observations are likely to be of much lower resolution (>50 m), the influence of smaller footprint features resolved by SnowSAR may be reduced, thus minimizing the need to strictly resolve spatial variations in microstructure. Much like background contributions (see Rott et al., 2010), spatial complexities related to microstructure may also be easier to constrain when viewed as a seasonal evolution where baselines (snow-free or sequential acquisitions) can be used to decompose change from a known state.

At present, space-borne radar is one of few concepts which can combine a wide swath with moderate to high resolution measurements suitable for operational monitoring of volumetric snow properties. It is clear that advancement of radar-based retrievals of snow mass must occur alongside improved understanding, measurement, and modeling of microstructure and background signals. The expansive and coincident nature of the TVC datasets provided an opportunity to evaluate forward model skill and to diagnose geophysical contributors in a characteristic tundra environment. Similar analysis for other climate classes of snow cover will be vital because of the inherent link between snow microstructure and radar signatures. This is a fundamental gap which must be addressed at multiple frequencies in order to further the justification for future space borne radar mission concepts.

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# List of figure captions

Figure 1: Overview of the Trail Valley Creek (TVC) research basin with vegetation (A) and SnowSAR 17.2 GHz VV backscatter overlaid (B). Boundaries of the Upland tundra (UT) and SikSik (SS) study sites are outlined in red. The higher elevation UT site was generally free of shrub or forest vegetation while SS contains discontinuous open coverage and a range of vegetated features.

Figure 2: Transect, pit, and airborne snow measurements at the Upland Tundra (UT) and SikSik (SS) sites. Green shaded areas indicate areas of shrub or forest vegetation greater than 50 cm in height. The background shows lidar derived snow depths detailed in section 2.3.2.

Figure 3: Probability density functions (PDFs; A) and length scales of variability (B) for open tundra SWE at the Upland Tundra (UT) and SikSik (SS) sites.

Figure 4: Lidar derived snow water equivalent (SWE; Left) and digital elevation model derived slope (right) and at the UT site. The locations of large snow drifts are outlined in black.

Figure 5: Near infrared photograph of snow pit 6 within the SS domain (Left) where basal depth hoar is easily distinguished by texture due to its unconsolidated structure. Microstructure estimates from IRIS, shown as , were largest near the basal surface.

Figure 6: Exponential correlation length () within the 11 snow pits. Probability density functions (PDFs) show difference between slab and depth hoar layer classified measurements. Estimates of were derived from IRIS measurements as detailed in Section 2.2.

Figure 7: Optimization of exponential correlation length () at 17.2 GHz (A) by way of scaling () in 1-, 2-, and n-layer configurations of MEMLS. Soil roughness () sensitivity analysis at 9.6 GHz (B) using the reflectivity model of Wegmüller and Mätzler (1999).

Figure 8: Comparison of optimized MEMLS simulations and SnowSAR measured at each snow pit in 1-, 2- and n-layer configurations. Adjusted () input of listed for each configuration in addition to the optimized soil input (*h*).

Figure 9: Distribution of SnowSAR at 9.6 (Black) and 17.2 GHz (Grey) within the SikSik (SS) and Upland Tundra (UT) sites. Measurements have been restricted to those located in open tundra.

Figure 10: Relationship between SnowSAR and SWE at the SS and UT sites (Black dots and boxplots). Grey shaded area shows the MEMLS 1-layer simulated range of variability based on snow pit properties. The solid and dashed lines are for of 0.383 and 0.247 mm, representing the first standard deviation of the in situ measurements.

Figure 11: MEMLS 2-layer sensitivity to increasing wind slab composition in a snowpack with total SWE between 50 and 250 mm. A constant depth hoar layer of 15 cm (34 mm SWE) was maintained while increasing thickness of the wind slab between 5 and 75 cm. The grey shaded area shows the response within the standard deviation range of snow pit observed slab . The top panel shows the thickness weighted mean for each of the slab parametrizations as a function of total SWE. Blue lines show control simulations where microstructures of both layers were assigned of 0.388 mm.