Implications of representing snowpack stratigraphy for large-scale passive microwave remote sensing

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Abstract. The layered character of snowpacks increases the complexity of algorithms intended to retrieve snow properties, such as water equivalent, from the snowpack microwave return signal. However, it also offers the opportunity to infer snowpack microphysical properties by combining different frequencies and polarizations, and therefore has the potential for fully self-consistent retrieval algorithms applicable to multi-frequency sensors, which avoids the necessity for additional information sources which are otherwise required. Implementation of a multifrequency snow property retrieval strategy requires knowledge of the stratigraphy of snowpack microphysical properties, which as a practical matter can only be produced by predictive (forward) models that include representation of spatial variability of vegetation and snow characteristics. We describe a multi-layer snow model designed for such applications. The model is computationally efficient to the extent that it can be implemented at the scale of large watersheds, or even over continents. The model’s ability to replicate large-scale snowpack layer features is evaluated using observations from the Cold Land Processes Experiment (CLPX) and a 2002 Nome-Barrow snowpit transect (SnowSTAR 2002). The multi-layer model coupled with a radiative transfer scheme improved the estimation of brightness temperatures both in terms of absolute values and frequency/polarization differences (error reductions ranging from 47 to 72%). The effects of certain stratigraphic features on large-scale microwave emissions are demonstrated, while potential impacts of using a multi- versus a single-layer model when assimilating radiances are explored.
1. Introduction

Snow is a key component of the global hydrologic cycle, especially in mid to high latitudes. From a hydrological standpoint, snowpacks act as storage reservoirs which modulate the seasonal cycle of runoff. The space-time distribution of snow can also affect atmospheric processes as a result of the strong contrast in albedo and surface temperature between snow covered and snow free surfaces. These contrasts can alter atmospheric circulation patterns \cite{Heim and Dewey, 1984; Cohen and Entekhabi, 1999}. The spatial and temporal sparseness of in-situ observation networks globally have led to reliance on remotely sensed observations of snow properties, especially snow cover extent, at large scales \cite{Schmugge et al., 2002}. However, the production of consistent, accurate satellite-based estimates of snow water equivalent, the key hydrological variable associated with snow, remains elusive.

Satellite observations of snow properties have been available for more than 40 years, mostly based on visible and passive microwave sensors. Although visible wavelength sensors provide a basis for estimating snow cover extent, they do not provide any information about snow water storage, which is hydrologically more important. Passive microwave remote sensing has been used to operationally map snow cover and depth since 1978 \cite{Chang et al., 1987; Kelly et al., 2003}, however its use is generally limited to relatively thin, cold snowpacks. Previous studies have shown correlations between passive microwave brightness temperatures and snow depth \cite[e.g.][]{Künzi et al., 1982; Hallikainen and Jolma, 1992; Josberger and Mognard, 2002; Derksen et al., 2003}, but many limitations nevertheless make direct retrieval of snow properties from passive microwave sensors problematic.
These problems include the coarse spatial resolution of current sensors, signal “saturation” (related to the penetration depth for different frequencies), presence of liquid water in the snowpack which diminishes volume scattering and hence limits retrieval algorithms, dense forest cover (which dominates the microwave signal), as well as snow metamorphism which can strongly alter microwave emissivity and thus complicates retrieval algorithms.

The experience to date is that satellite observations alone cannot provide a means for estimating snow water storage at large scales, aside from a few highly specific cases. An alternative approach is to merge remotely sensed observations with physically-based model predictions to constrain retrieval algorithms and potentially account for the uncertainties (e.g. through data assimilation [Durand and Margulis, 2006; Pulliainen, 2006]). Such an approach would require coupling a large-scale snow hydrology and a microwave emission model. Although there have been a few studies [Andreadis et al., 2008a; Durand et al., 2008] that have evaluated microwave brightness temperatures predicted by such coupled models, a key issue is that a discrepancy exists between the assumption of homogeneous snowpacks inherent in most land surface models and the fact that snow is a naturally layered medium [Andreadis et al., 2008a]. Moreover, anomalous behavior of microwave emissivity, such as increasing brightness temperatures with increasing snow depth, is sometimes observed from satellite and ground measurements [Rosenfeld and Grody, 2000a], and can be attributed to snow metamorphic processes and the snowpack layering structure [Hofer and Mätzler, 1980; Rosenfeld and Grody, 2000b]. In cases when the wavelength is comparable to either layer thicknesses or particle sizes, and when there are sufficient differences in the dielectric properties between layers, microwave signals can
be very different for otherwise comparable snowpacks [Colbeck, 1991]. For example, the
presence of an ice layer within a snowpack can change its brightness temperature by up
to 50 K [Edgerton et al., 1971], while depth hoar can significantly decrease brightness
temperature by increasing scattering due to its prevailing larger snow crystals [Hall et al.,
1986].

The layered character of snowpacks increases the difficulties in deconvolving the return
microwave signal (passive or active), but it also offers the opportunity to infer the meta-
morphic signature of the observed snowpack [Rosenfeld and Grody, 2000b] and to extract
snowpack microphysical information by combining different frequencies and polarizations.

In order to exploit this potential, snowpack stratigraphic information is required. Given
the spatial scales supported by existing satellite technology, this prior information can
only be provided by snow models applicable to the spatial scale of passive microwave
sensors (typically tens of km) that can represent layered snowpacks. Key questions then
become how to capture the spatial variability of snow layering in a forward model, and
how that variability is reflected in the satellite microwave signal.

Snowpack layering is affected by meteorological (wind, air temperature, precipitation,
solar radiation) as well as topographic (slope, aspect) and physiographic (presence or ab-
sence of vegetation) controls. Sturm and Benson [2004] examined the spatial heterogeneity
of snowpack stratigraphy over distances ranging from 10 m to more than 200 km using
snowpit measurements collected on the Arctic coastal plain of Alaska. They found that
at scales greater than 100 m, topographic variations can result in substantial variations
in precipitation, wind, and solar radiation which in turn are reflected in heterogeneity in
snowpack stratigraphy, while such variations can also be caused by snow-vegetation in-
teractions [Sturm, 1992]. However, notwithstanding these small scale variations, at scales larger than about 10 km, there was a general coherence in the snowpack layers which had relatively strong spatial correlations across the Kuparuk basin [Sturm and Benson, 2004] for distances ranging from 10-25 km. Weather appeared to be the primary driver for this synoptic-scale variability that led to snow layers remaining recognizable over distances order of 160 km, notwithstanding that the landscape and its interactions with weather introduced relatively smaller-scale heterogeneity that distorted the larger-scale trends. The combined effects of controls at these two scales was noted by Schweizer and Kronholm [2007], who observed a layer of surface hoar over an approximately 250 km$^2$ area, the spatial coherence of which was eventually reduced by local effects of wind and aspect variability.

The objective of this study is to develop a macroscale multilayer snow model that can reflect spatial variability in the controlling processes at multiple scales, and can replicate large-scale snowpack layer features and their effect on passive microwave emissivity. Such a model, which would account for sub-grid variability in topography and land cover and their interactions with precipitation, air temperature and wind would potentially be able to capture the large-scale layering features that would affect satellite retrievals, such as surface and depth hoar, by simulating snowpack physics within these relatively homogeneous areas. The model we have developed with this objective in mind is described in the following section. Section 3 presents the model validation results. Validation is performed both in terms of snow stratigraphy and microwave emissivity, while this study also examines the effects of certain stratigraphic features on the satellite-scale observations, as well
as present a comparison of the potential use of a multi-layer versus a single-layer forward model to assimilate satellite passive microwave observations.

2. Model description

The multi-layer snowpack model is based on a mass and energy balance. We describe the foundations for the model below along with a brief description of the overall model structure in a macroscale setting.

2.1. General model structure

Many snowpack models have been developed for different applications including hydrological prediction, global circulation modeling, and avalanche forecasting [Etchevers et al., 2004]. These models vary in complexity, ranging from simple force-restore schemes [Yang et al., 1997], to approaches assuming snowpacks as homogeneous media [Verseghy, 1991; Wigmosta et al., 1994; Koren et al., 1999], to multi-layered representations with detailed snowpack physics [Brun et al., 1989; Jordan, 1991; Bartelt and Lehning, 2002]. Given our objectives, a physically-based representation of snowpack internal processes is required that would be of intermediate complexity and efficient enough to simulate stratigraphy over large scales, [e.g. Loth et al., 1993; Sun et al., 1999]. Our approach here is similar to previous studies, adapting existing models [Jin et al., 1999] and coupling them to a large-scale model [Boone and Etchevers, 2001].

The macroscale model, which provides a framework for our development, is the Variable Infiltration Capacity (VIC) model [Liang et al., 1994], which solves the water and energy balance of a soil column over a gridded domain accounting for sub-grid variability in land cover by partitioning each model grid cell into tiles based on topography and land cover.
Spatial heterogeneity in runoff generating processes is represented by a parameterization of the combined effects of topographic and small scale spatial variability in soil properties. Water and energy states and fluxes are simulated for each “homogeneous” tile and averaged over each model grid cell. The multi-layer snowpack model we develop here is intended to be coupled to VIC in such a way as to retain the interactivity between the current essentially single-layer snowpack model and other components including the canopy snow interception and energetics [Andreadis et al., 2008b], frozen soils [Cherkauer and Lettenmaier, 2003], and blowing snow [Bowling et al., 2004] sub-models.

2.2. Energy balance

The snowpack is represented as a layered medium, with energy exchange between snow and the atmosphere limited to the surface layer taking the form

\[ Q^* = (1 - \alpha)Q_s \downarrow + Q_l \downarrow - Q_l \uparrow + Q_h + Q_e + Q_a - Q_c - Q_{lw} - \Delta H \]  

where \( Q_s \downarrow \) is the downward shortwave radiation, \( \alpha \) is the snow albedo, \( Q_l \downarrow \) is the downward longwave radiation, \( Q_l \uparrow \) is the emitted longwave radiation (= \( \epsilon \sigma T^4_s \)) where \( \epsilon \) is the emissivity, \( \sigma \) is the Stefan-Boltzmann constant and \( T_s \) is the snow surface temperature, \( Q_h \) is the sensible heat flux (including advected sensible heat flux), \( Q_e \) is the latent heat flux (including heat flux from sublimation), \( Q_a \) is the advected heat from rainfall, \( Q_c \) is the heat conducted to/from the deeper snowpack, \( Q_{lw} \) is the heat associated with liquid water percolation, \( \Delta H \) is the change in heat content \( \left( = c_i h_m \frac{T_t - T_{t-1}}{\Delta t} \right) \) where \( c_i \) is the heat capacity of ice, \( h_m \) is the snow water equivalent, and \( T_t, T_{t-1} \) are the temperatures at the current and previous time steps respectively, and \( Q^* \) is the energy available for melting or refreezing. Albedo is calculated as a function of snow age using an
empirical relationship [Andreadis et al., 2008b], while incoming shortwave and longwave radiations are provided either from measurements or can be estimated [Thornton and Running, 1999]. Turbulent heat fluxes and advected energy from rainfall are calculated using the formulations in Wigmosta et al. [1994] depending on atmospheric pressure, air temperature, rainfall amount, and saturation vapor pressure and temperature at the snow surface.

Heat conduction through the snowpack is modeled as

\[ Q_{c,j} = k_j \frac{\partial T}{\partial z} \] (2)

where \( k_j \) is the heat conductivity, and \( \frac{\partial T}{\partial z} \) is the temperature gradient across layer \( j \) (if this layer is the bottom layer, \( Q_c \) becomes the ground heat flux). Heat conductivity can be estimated in a number of ways; here, it is calculated from a quadratic relationship between conductivity and density derived from a measurement data set that encompassed most types of seasonal snow cover [Sturm et al., 1997]. The energy balance for the internal layers can be formulated as

\[ Q^*_{j} = Q_{s,j} - Q_{lw,j-1} + Q_{lw,j} - Q_{c,j-1} + Q_{c,j} - Q_{h,j-1} + Q_{h,j} - \Delta H_j \] (3)

where \( Q_{s,j} \) is the shortwave radiation penetrating to this layer, \( Q_{lw,j} \) is the energy associated with water movement from layer \( j \) to \( j+1 \) \( = c_w \frac{U_{l,j} T_j}{\Delta t} \) and \( c_w \) is the heat capacity of water and \( U_{l,j} \) the liquid water flux from layer \( j \), \( Q_{c,j} \) is heat conducted from layer \( j \) to \( j+1 \), and \( Q_{h,j} \) is the latent heat released from vapor diffusion from layer \( j \) to \( j+1 \) \( = L_s \frac{U_{v,j} \rho_w}{\Delta t} \) with \( L_s \) being the latent heat of sublimation, \( U_{v,j} \) the vapor flux from layer \( j \), and \( \rho_w \) the water density. The energy balance is solved as a system of equations.
using the iterative discrete Newton algorithm [Dennis and Schnable, 1983], or as an Euler backward tridiagonal system if the first method does not converge [Loth et al., 1993].

2.3. Mass balance

After solving for the snowpack temperature profile, the amounts of water that have melted or refrozen are calculated and liquid water is allowed to percolate from one layer to another using a 35% retention capacity. Along with the liquid water flux, vapor fluxes are calculated from the Clausius-Clapeyron equation

\[ U_v = -D_s \frac{p_s}{RT^2} \left( \frac{L_s}{RT} - 1 \right) \frac{\partial T}{\partial z} \]  

(4)

where \( p_s \) is the vapor saturation pressure, \( R \) is the water vapor gas constant, and \( D_s \) is the vapor diffusion coefficient for snow estimated from Colbeck [1993].

Two important processes related to snow metamorphism are densification and grain growth. The former is caused by compaction of snow layers, destructive (equi-temperature), constructive (temperature gradient), and melt metamorphism. Anderson [1976] developed a model for snow densification taking these processes into account, which has been adapted to the VIC model with the essentially single-layer formulation [Andreadis et al., 2008b], and has been extended to the multi-layer version by calculating snow density changes for each layer. The snow grain growth model is adapted from Lehning et al. [2002], and uses empirical relationships for equilibrium growth metamorphism (small temperature gradients, less than 5 K), and wet snow metamorphism (presence of liquid water in the snowpack) [Brun, 1989]. When temperature gradients are larger than 5 K, higher growth rates occur (kinetic growth metamorphism) and both layer-to-layer and intra-layer vapor transport contribute to the water vapor supply [Sturm and Benson, 1997].
The snow model is formulated in a way such as to be computationally feasible for simulating snowpacks across large scales. Therefore, some constraints are required in the combination of layers to keep the number of layers below a prescribed maximum. A new layer is created after every snowfall event as long as it has a snow water equivalent of 5 mm, while the surface layer snow mass is not allowed to exceed 20 mm. At the end of each time step, two neighboring layers are combined if both are cold and their temperature gradient is less than 5 K/m. When the maximum number of layers is exceeded by invoking these criteria, the two layers with the minimum temperature gradient and density difference are combined.

3. Results

3.1. Cold Land Processes Experiment, Colorado

The Cold Land Processes Experiment (CLPX) was a multi-sensor and multi-scale field campaign conducted during the winters of 2002 and 2003 over a set of nested areas in Colorado and Wyoming. A set of snowpit detailed measurements were taken in a 100 × 100 m clearing, designated as the Local Scale Observation Site (LSOS), in addition to radiometric measurements from a Ground-Based Microwave Radiometer (GBMR-7). Snowpit measurements included snow depth, water equivalent, density, temperature and grain size profiles which were collected between November 2002 and March 2003, while the GBMR-7 measurements included brightness temperatures at 18.7, 23.8, 36.5 and 89 GHz during selected days in January and December 2002 and February and March 2003 [Graf et al., 2003].

This set of measurements offered an opportunity to simultaneously test the ability of VIC to reproduce stratigraphic profiles, and to simulate the corresponding passive
microwave response after being coupled with a radiative transfer model. The model was allowed to have a maximum number of five layers, and was forced with daily precipitation, maximum, minimum air temperature and wind speed in a manner similar to [Maurer et al., 2002]. This was done to emulate the general data availability and computational feasibility when applying such macroscale hydrological models in off-line settings. The simulation period was from 1 October 2002 (no snow) to 25 March 2003. The model used internal algorithms to disaggregate the daily meteorological data (by partitioning precipitation into equal-magnitude increments with diurnal variability introduced to air temperature) and/or to calculate required inputs at the model time step (in the case of relative humidity, incoming shortwave and longwave radiation) as outlined in Maurer et al. [2002].

3.1.1. Snowpack stratigraphy

Before examining how well snowpack stratigraphy is simulated by the multilayer model, it is important to evaluate the accuracy of the model in simulating snow water equivalent. Fig. 1 shows the simulated snow water equivalent (SWE) for both the 5-layer and standard (surface and pack layer) VIC snow models compared with snow pit measurements. Both model simulations are quite close to the observed SWE, although they seem to slightly underestimate snow accumulation in March 2003. The differences in simulated snow mass between the two model versions are minimal, with the single- and multi-layer relative having mean squared errors of 9.4 mm (6.1%) and 11.7 mm (7.3%) respectively. The differences are mostly attributable to small differences in sublimation rates.

Fig. 2 shows a representative set of simulated and observed snow temperature, density and grain size profiles for selected dates (22 January, 2 and 20 February, and 11 March
The model appears to be able to simulate snowpack temperature quite well, with small discrepancies occurring either because of underestimated incoming shortwave radiation (Fig. 2b) or underestimated snow depth (Fig. 2c). The model predicts a colder temperature than observed at the middle of the snowpack for the March 11 pit (Fig. 2d), although the surface, top 20 cm and ground temperatures are predicted correctly. Results are similar for snow density, the model reproduces the density profile quite accurately for the January snowpit (Fig. 2a) but underestimates the densification rate for the upper portion of the pack during February (Fig. 2b,c). When partial melting occurs during March, the model predicts a larger density gradient while the observations show a more gradual change in density with depth (Fig. 2d). However, the model does capture the qualitative features of the increase in density near the surface and the decrease in the deeper layers. Evaluation of the predicted grain size is difficult because of the range in the measurements, and the single model value for each layer. Fig. 2a shows that the model underestimates grain size for the deeper layers, but predicts it quite well for the upper layers. During February (Fig. 2b,c) the model is able to predict the large near-surface grain size, as well as the bottom depth hoar evident in the measurements. Simulated grain size is relatively close to the observed range in the March snowpit (Fig. 2d) as well, with a small underestimation near the snowpack mid-depth.

3.1.2. Ground-based microwave brightness temperatures

Potential improvements in microwave emission model predictions when using a multiple layer coupled (hydrologic and radiative transfer) model over a single-layer one were assessed using the GBMR-7 measurements. The latter were taken on selected dates in February 2003 within the LSOS site and included measurements of brightness tempera-
ture at 18.7, 23.0, 36.5 and 89.0 GHz. The microwave emission model used is based on the Dense Media Radiative Transfer (DMRT) theory and quasi-crystalline approximation \cite{Tsang et al., 1985, 2000}. The model does not make the assumption of independent scattering, and takes into account inter-particle forces that lead to adhesion and collective scattering calculating microwave emissivity as a function of snow depth, density, grain size and temperature \cite{Tsang et al., 2007}. The degree of particle clustering is modeled through a stickiness parameter, which also affects the frequency dependence of the extinction coefficient. Recently, the DMRT was extended to a multi-layer formulation, and was shown to agree very well with the GBMR-7 ground measurements when forced with observed snow stratigraphy \cite{Liang et al., 2008}. Here, we used the VIC multilayer snow model stratigraphy in place of observations for DMRT, with the stickiness parameter was set to the default value of 0.1 for all snowpack layers.

Fig. 3 compares the single and multiple layer simulations of brightness temperatures with the GBMR-7 measurements at 18.7 GHz (Fig. 3a) and 36.5 GHz (Fig. 3b) at both horizontal and vertical polarizations. The improvement in predicting 18.7 GHz TB using 5 layers is evident for all measurement dates except 2/19 for the vertical and 2/20 for both polarizations. Similarly, for 36.5 GHz the 5-layer model better predicted TB with the exception of 2/21 for the horizontal and 2/22 for the vertical polarization. The corresponding prediction root mean squared errors (RMSE) for the four frequency channels (18.7 GHz horizontal and vertical, and 36.5 GHz horizontal and vertical) were 14.0, 5.5, 13.3, 9.4 K for the single-layer coupled model, and 5.5, 2.9, 5.2, 3.4 K for the 5-layer VIC/DMRT model. These results appear to confirm the hypothesis in previous work using the single-layer VIC/DMRT \cite{Andreadis et al., 2008a}, that predictions of TB for
horizontal polarizations were problematic because of the reflection at layer boundaries, which was not captured by the single-layer model and mostly affects horizontal polarization [Wiesmann and Mätzler, 1999].

The 5-layer coupled model is also able to reproduce the observed frequency and polarization differences in the LSOS site better than the single layer model (Fig. 4). The errors in predicting polarization differences (in terms of RMSE) for the 1-layer model were 9.8 and 8.7 K for 18.7 and 36.5 GHz, which were reduced by the 5-layer model to 3.9 and 3.3 K respectively. The error reduction is also evident in the scatter plot in Fig. 4a, which shows simulated versus observed polarization differences. Fig. 4b shows simulated versus observed frequency differences for both 1-layer and 5-layer models and both polarizations. The 1-layer model has similar frequency difference prediction errors for both polarizations, 11.9 and 11.4 K for the horizontal and vertical, while the 5-layer model decreases those errors to 9.1 and 5.6 K respectively.

3.2. SnowSTAR 2002 Transect, Alaska

A series of snowpack stratigraphy measurements (designated SnowSTAR 2002) were taken in arctic Alaska along a transect extending 750 km from Nome to Barrow, AK during March and April 2002 [Sturm and Liston, 2003; Sturm and Benson, 2004]. These measurements span distances that range from 10 m to 200 km between snowpits, and include snow depth, density, grain size, temperature, water equivalent, hardness, type, and fractions of wet, recent snow, slabs, and hoar. The prevalent land cover along the traverse is tundra, while the area is fairly flat with relatively low relief and most elevations being below 500 m. These data the opportunity to evaluate the relationship between large-
scale variability in stratigraphy and satellite-observed passive microwave emissions, given
the relative homogeneity and absence of dense forest cover.

Satellite observations of microwave brightness temperature, taken from the Special Sen-
sor Microwave Imager (SSM/I) [Grody and Basist, 1996], were used to examine the change
in microwave emissivity with snow depth. Fig. 5 shows snow depth measurements from
different dates for each co-registered SSM/I 25×25 km pixel and the corresponding bright-
ness temperatures for 37 GHz, which is probably the most appropriate frequency given
snowpack depths and the typical penetration depths for passive microwave frequencies.
Satellite pixels which contained lakes were excluded to isolate the effects of snowpack
variability on brightness temperatures [Duguay et al., 2005], while both polarizations
(horizontal and vertical) and orbits (ascending and descending) are shown for the SSM/I
observations. Although there is a general pattern of increasing (decreasing) brightness
temperature with decreasing (increasing) snow depth across the transect, there are clearly
cases where the opposite is true. Possible reasons for this behavior include the effects of
different snowpack stratigraphy as well as the potential spatial variability within each
satellite pixel. We examine below two particular cases to demonstrate the effects of cer-
tain stratigraphic features on the large-scale microwave emission.

3.2.1. Depth hoar

Depth hoar is mostly formed in snowpacks with strong temperature gradients. Its
characteristics include large snow crystals and relatively lower density (mechanically weak
layers), with depth hoar crystal growth depending directly on increased vapor diffusion.
Microwave brightness temperature should be lower for snowpacks with increased snow
grain sizes [Chang et al., 1982], since this allows for greater radiative scattering, especially
when crystals are of equivalent size to the wavelength. The larger crystals of depth hoar layers cause a decrease in brightness temperatures relative to snowpacks where such layers don’t exist. In order to demonstrate this effect, we examine the snowpack stratigraphy at two sites with similar snow accumulation, and the respective satellite observations. Figs 6a-c show snow density, grain size and temperature profiles for the IC12 site, with the measurement taken on 4 April 2002 and snow depth of 56 cm. Depth hoar was prevalent in this snowpack, as in other measurement sites across a distance of 15 km shown in Fig. 6e with hoar fraction ranging from 60 to 80%. Measurements at those sites were taken between 15-17 April 2002, with snow similar depths being similar (coefficient of variation 0.18). The satellite observations for 18.7 and 36.5 GHz (both polarizations, Fig. 6d) appear very consistent (with only 2 of the sites in the same SSM/I pixel) suggesting that the effective snowpack properties observed by SSM/I across the 15 km were quite similar.

The stratigraphic profile at the CB04 site (Figs 6f-h) shows that snow crystals are much smaller than at the IC12 site, while the density profile is similar except for two thin ice layers near the bottom. Snow depth at the site was 56 cm, with the spatial variability from two sites within a 15 km distance relatively small (coefficient of variation: 0.14). All measurements were taken on 28-29 March 2002 (Fig. 6j). SSM/I brightness temperatures are relatively consistent across these sites (Fig. 6i), and they are higher than the ones for the depth hoar sites. The differences were about 40 K for both polarizations at 36.5 GHz (42.7 and 43.2 K for horizontal and vertical respectively), while the respective differences for 18.7 GHz were smaller (24.0 and 14.5 K). An interesting feature becomes evident when we examine the microwave emissivity of the site at ~5 km from site CB04, which is lower than the other sites at both frequencies. This discrepancy could be attributed to the
slightly higher depth (61 versus 56 cm), but it can also be argued that the larger depth
hoar fraction for that site caused that decrease in brightness temperature. Nonetheless,
the effect of depth hoar on the microwave emission, even at large scales, has been well
known [Hall et al., 1986; Foster et al., 2005] and is also verified by the SnowSTAR 2002-
SSM/I measurements.

3.2.2. Inter-layer variability

Two sites with similar snow depths were selected to examine the effects of the density
vertical variability on the large scale microwave emission observed by the satellite. Snow
depth at site ARM was 40 cm, and the observed stratigraphy displays sharp changes in
density with depth (Fig. 7a), while grain size (Fig. 7b) and temperature (Fig. 7c) show
a more typical profile for an Arctic Alaska snowpack. Measurements were taken at three
more sites within a distance of 25 km from ARM on 25, 27, 28 April and 2 May 2002. The
spatial variability in snow depth was relatively high (coefficient of variation 0.46), but the
sites shown are in different SSM/I pixels from ARM. The higher spatial variability is also
evident in the layer fractions (Fig. 7e), where ARM has 50% depth hoar, while the site at
Chukchi (3 km distance) has a much shallower snowpack (11 cm) and a large fraction of
slab (81%, probably caused by local wind effects), and the other two sites have comparable
snow depths (25 and 29 cm) with somewhat smaller depth hoar fractions. The differences
in the stratigraphy between these sites is arguably evident when looking at the SSM/I
observations (Fig. 7d, higher $T_B$ for lower depths and smaller hoar fractions). Figs 7f,g,h
show density, grain size and depth for the SA03 site, where snow depth was 44 cm and
the density profile is “smoother” than ARM with the exception of a thin slab layer near
the top and a basal ice layer. The sites at a distance of 25 km from SA03 (measurements
taken 4-6 April 2002) have similar snow depths (35 and 32 cm) as well as layer fractions
with depth hoar prevalent (Fig. 7j). The brightness temperatures along this transect
(Fig. 7i) range within 7 K for 36.5 GHz, and 3 and 6 K for 18.7 GHz, which can be
attributed to the difference in snow depth between SA03 and the other two sites. When
compared to the ARM site’s microwave signature, the feature that stands out is the much
smaller polarization difference for both frequencies (12.9 and 26.1 K at 36.5 GHz and 18.0
and 32.9 K at 18.7 GHz for SA03 and ARM respectively). The layering differences in
the ARM snowpack increases reflection (at layer interfaces), reducing horizontal $T_B$ and
increasing the polarization difference [Mätzler, 1987; Mätzler and Hüppi, 1989].

3.2.3. Refrozen snow

The microwave signature of melting snow changes dramatically [Tedesco et al., 2006]
raising the emissivity because of increased absorption. When snow refreezes scattering is
increased because of larger effective grain sizes compared to dry snow, while polarization
differences (vertical minus horizontal) become smaller [Choudhury, 1995]. Although re-
refrozen snow layers were found in the SnowSTAR 2002 transect measurements and sites
with consistent microwave signatures were found, there were sites with refrozen layers that
did not exhibit the expected behavior. This might be related to horizontal discontinuities
in the refrozen layers within each 25 km pixel affecting the aggregate microwave emission.

3.2.4. Hydrologic simulations

The accuracy of modeling predictions of snow mass and properties depends to a large
degree on the meteorological forcings (e.g. precipitation and air temperature). In-situ
measurement networks of meteorological variables are quite sparse, both spatially and
temporally, while satellite-observed or model-derived meteorological data sets can have
large uncertainties [Adam and Lettenmaier, 2003] and even large differences between data
sets [Adler et al., 2001]. A first assessment of the information content of passive microwave
satellite observations in terms of errors in meteorological forcings that propagate to the
model predictions of snow mass, can be made by comparing the model-predicted $T_B$ with
the actual satellite observations. Snow hydrologic simulations were performed along the
SnowSTAR 2002 transect with both the 1 and 5-layer VIC models. The required daily
meteorological forcings (precipitation, air temperature and wind speed) were provided
by the ERA-40 re-analysis of meteorological observations globally, including conventional
and satellite measurements and atmospheric model predictions [Uppala et al., 2005]. The
SYMAP algorithm [Shepard, 1984] was used to interpolate the data from the ERA-40
spatial resolution of $\sim1.125$ degrees to the measurement locations.

Simulated SWE was overestimated across the transect, shown in Fig. 8 with an estima-
tion error of 89.3 mm. This discrepancy can be attributed to the ERA-40 precipitation
being overestimated and in fact comparisons with in-situ observations from the SNOTEL
network showed large differences in cumulative precipitation [Shi et al., 2009]. On the
other hand, daily maximum and minimum air temperatures from ERA-40 showed rela-
tively good agreement with SNOTEL measurements. SWE simulations between the 1 and
5-layer models were almost identical, but the snow depth estimates were different (due
to differences in predicted snow density) although both were higher than the observed
depths (RMSE of 43.7 cm for the 5-layer model).

The corresponding simulated $T_B$ from both the single and multi-layer models along with
the SSM/I observations at the SnowSTAR 2002 transect measurement locations are shown
in Fig. 9. The 1-layer $T_B$ predictions are consistently higher than the satellite observa-
tions for 18.7 GHz (Fig. 9a), despite the fact that simulated snow depths were also higher than the ones observed. Generally, it would be expected that $T_B$ would decrease with increasing depth, but the single-layer model cannot predict the larger depth hoar crystals, underestimating grain size. The smaller grain size increases $T_B$, offsetting the decrease of $T_B$ because of the higher snow depth. If the SSM/I radiance observations were assimilated into the single-layer model, predicted SWE (or snow depth) would theoretically be increased to accommodate the negative difference between the observed and the model-predicted $T_B$. On the other hand, the 5-layer VIC is able to reproduce the larger grain sizes and predict the lower $T_B$ relative to the SSM/I observations. Assimilating the latter into the 5-layer model would theoretically lead to a decrease in snow depth and SWE. Clearly an actual data assimilation experiment would be needed to correctly evaluate the analysis SWE and depth estimates, but the innovations (actual minus model-predicted observations) do provide insight into the potential impact of the assimilation.

The 36.5 GHz $T_B$ simulations (Fig. 9b) exhibited the same qualitative behavior as the 18.7 GHz. The 1-layer model generally over-predicted $T_B$, while the 5-layer model predicted lower microwave emissivities consistent with the simulated larger snow mass. $T_B$ predicted by the 5-layer model at the measurement sites at distances from Nome between 600 and 700 km are higher than the observed ones, despite SWE being higher as well. This is mostly attributable to the underestimation of grain sizes at the mid to upper parts of the snowpack, leading to an increase in simulated $T_B$ for 36.5 GHz and not 18.7 GHz suggesting that depth hoar is reproduced by the model. This highlights the difficulties in consistently predicting $T_B$ despite the improvement in the forward models, as well as the
relatively large sensitivity to snow grain size, which needs to be taken into account when assimilating microwave radiances by defining the appropriate uncertainties for it.

4. Conclusions

A multilayer snow model was incorporated into a macroscale hydrology model that is able to account for the effects of topography and vegetation on snow accumulation and ablation. Snow temperature, density and grain size profiles were reproduced reasonably well by the model when evaluated with point measurements and using meteorological forcings that would generally be available for large-scale basin simulations. Improvements were also shown in predicting microwave brightness temperatures in terms of both absolute value and frequency/polarization differences when compared with in-situ measurements from a ground radiometer. Measurements from a snowpit transect across Alaska were used to examine the effects of stratigraphy on SSM/I satellite observations. Sites with depth hoar and inter-layer variability exhibited very different passive microwave emission signatures when compared with sites of comparable snow depth. These effects were evident across 15-25 km, in agreement with a previous study [Sturm and Benson, 2004] that showed snow layers remaining recognizable over long distances. Because of unavailability of meteorological forcings along the transect, the ERA-40 re-analysis data set was used to simulate snow properties and microwave radiances thereafter. The multi-layer coupled model provided simulations with the expected behavior of lower brightness temperatures for overestimated snow depth, in contrast with the single-layer model simulations, offering insight into the potential use as the forward model in a data assimilation system.

The lower relief at the satellite scales and the sparse forest cover in Alaska are ideal for evaluating the passive microwave observations. However in other areas where snow
is of key significance to water resources (such as the western U.S.), complex topography, dense forest cover, and higher spatial variability in snow properties will impede estimation from data assimilation. Future work should perform data assimilation experiments using single- and multi-layer models to evaluate the information content of passive microwave satellite observations, and to identify the optimal model configurations for such a system (e.g. number of model layers, errors in precipitation, grain size). Although improvements in snow mass estimates may be incremental and dependent on a number of factors, it is important to explore and develop approaches to incorporate and digest the long-term satellite data set, especially for regions where no only in-situ measurements of snow properties are essentially non-existent, but meteorological forcings for hydrological models are uncertain.

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Figure 1: Simulated snow water equivalent for the multiple and single-layer VIC models compared with snowpit measurements from the CLPX Local Scale Observation Site.
Figure 2: Simulated (black) and observed (gray) snow temperature (left), density (middle), grain size (right) profiles for selected dates, 22 January (a), 3 February (b), 20 February (c) and 11 March 2003 (d), at the CLPX LSOS site, Colorado.
Figure 3: Observed (vertical bars), 1-layer (circles) and 5-layer (triangles) simulated brightness temperatures at horizontal (black) and vertical (grey) polarizations at 18.7 (top panel) and 36.5 (bottom panel) GHz from the CLPX LSOS site, Colorado.
Figure 4: (a) Simulated versus observed brightness temperature polarization differences for the 5-layer (triangles) and 1-layer (circles) models, 18.7 (black) and 36.5 (grey) GHz. (b) Simulated versus observed brightness temperature frequency differences for the 5-layer (triangles) and 1-layer (circles) models, horizontal (black) and vertical (grey) polarizations.
Figure 5: SnowSTAR 2002 transect snow depth (circles) at each co-registered SSM/I 25×25 km pixel and corresponding observed brightness temperature at 37 GHz, vertical (triangles) and horizontal (squares) polarization.
Figure 6: Snowpack stratigraphy profiles (density, grain size, temperature) for two SnowSTAR 2002 sites: IC12 (prevalent depth hoar, (a), (b), and (c)) and CB04 (fine/middle grain, (f), (g), and (h)); SSM/I observed brightness temperatures at measurement sites over 15 km distances from both sites ((d) and (i) respectively) at 18.7 horizontal (triangles), vertical (inverted triangles) and 36.5 GHz horizontal (circles) and vertical (squares); and layer fractions at measurement sites over 15 km distances from both sites ((e) and (j) respectively) with new snow (grey line), wet/icy snow (black dashed line), depth hoar (black line), and slab (grey dashed line) fractions.
Figure 7: Snowpack stratigraphy profiles (density, grain size, temperature) for two SnowSTAR 2002 sites: ARM (snowpack with large inter-layer density differences, (a), (b), and (c)) and SA03 (mostly depth hoar and slab layers, (f), (g), and (h)); SSM/I observed brightness temperatures at measurement sites over 25 km distances from both sites ((d) and (i) respectively) at 18.7 horizontal (triangles), vertical (inverted triangles) and 36.5 GHz horizontal (circles) and vertical (squares); and layer fractions at measurement sites over 25 km distances from both sites ((e) and (j) respectively) with new snow (grey line), wet/icy snow (black dashed line), depth hoar (black line), and slab (grey dashed line) fractions.
Figure 8: Simulated (grey squares) and observed (black circles) snow water equivalent along the SnowSTAR 2002 transect.
Figure 9: Observed $T_B$ from SSM/I, and simulated $T_B$ from the 1 and 5-layer VIC/DMRT along the SnowSTAR 2002 transect for both horizontal and vertical polarizations at 18.7 (a) and 36.5 (b) GHz.