Greenland Snow Ablation and Accumulation Observed Using ERS and SSM/I Data

Jed D. Pack , Jon W. Wallace, Michael A. Jensen
Brigham Young University, MERS Laboratory, 459 CB, Provo, UT 84602
Tel: 801-378-5736, Fax: 801-378-6586
packj@et.byu.edu, wallacej@et.byu.edu, jensen@ee.byu.edu

INTRODUCTION

Long term change of the Greenland ice sheet serves as a sensitive indicator of global environmental trends. As a result, scientists have focused on delineating the various facies and determining the surface characteristics in each region. This paper highlights research aimed at identifying the extent of summer snow melt and winter accumulation. First, we explain the trends seen in both active ERS1/2 (5.3 GHz, VV pol.) and passive SSM/I (37 GHz, V pol.) data and discuss identification of snow melt and accumulation. Next, we use active and passive Radiative Transfer (RT) models to justify our interpretation of the data.

Our research serves as an extension of work by Wismann, et. al [1] which focused on the active record only. The use of passive data greatly extends the observation history (back to the earliest space-borne microwave sensors), and provides opportunities for sensor fusion on more recent data sets.

DATA OBSERVATIONS

It has been indicated by Wismann, et. al [1], that sudden increases in the Normalized Radar Cross Section (NRCS) in the percolation zone represent the formation of reflective ice structures as a result of melt/refreeze events. These events are manifested as sudden *decreases* in brightness temperature in the SSM/I data record. We have processed both ERS1/2 and SSM/I datasets to create maps of the magnitude of these discontinuities and of the melt duration for each summer from 1992 to 1998.

Measuring Melt Discontinuity

Discontinuities are measured by convolving time series data for each location during each summer (JD 152-272) using an averaging derivative filter with impulse response $h[n] = u[n] - 2u[n\text{-}10] + u[n\text{-}20]$. The magnitude of the melt discontinuity is defined as the maximum value of the convolution output for ERS data, and the magnitude of the minimum value for SSM/I data. Results are shown in Fig. 1, where darker areas indicate larger discontinuities.

As snow accumulates on top of these reflective ice layers, the NRCS decreases steadily while the brightness temperature increases. If snow continues to accumulate for two or three years without another melt/refreeze event, the brightness temperature reaches a stable value. The NRCS, however, continues its decline for many more years before reaching a stable level. We intend to use data fusion techniques to take advantage of this and other distinguishing characteristics of the two instruments. This observation also suggests the possibility of using ERS data to approximate the number of years passed since the last melt at a given location in the percolation zone by comparing the slope of the NRCS to the snow accumulation rate. Fig. 2 shows time-series ERS data for two locations: 70° 54'N, 45° 36'W and 70° 21'N, 45° 35'W. It is clear that the first location did not experience a melt/refreeze cycle for many years until 1997. On the other hand, the second location (south about half a degree latitude) experienced this in 1995 and 1997 and also some time prior to 1992.

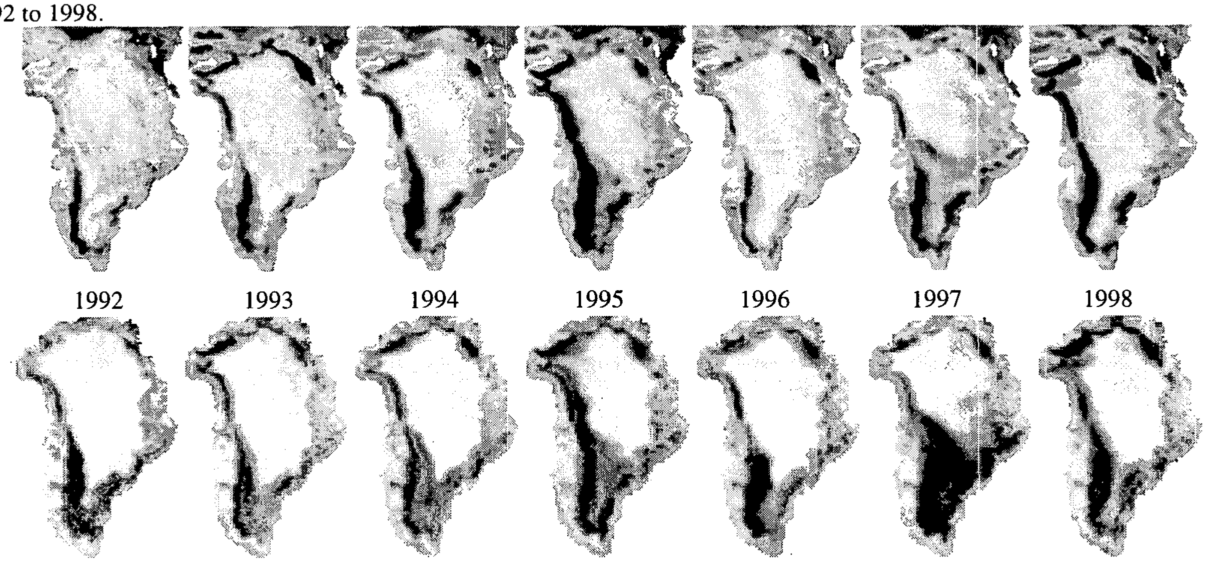

Fig. 1: Magnitude of summer melt discontinuity in ERS (top) and SSM/I (bottom) datasets.

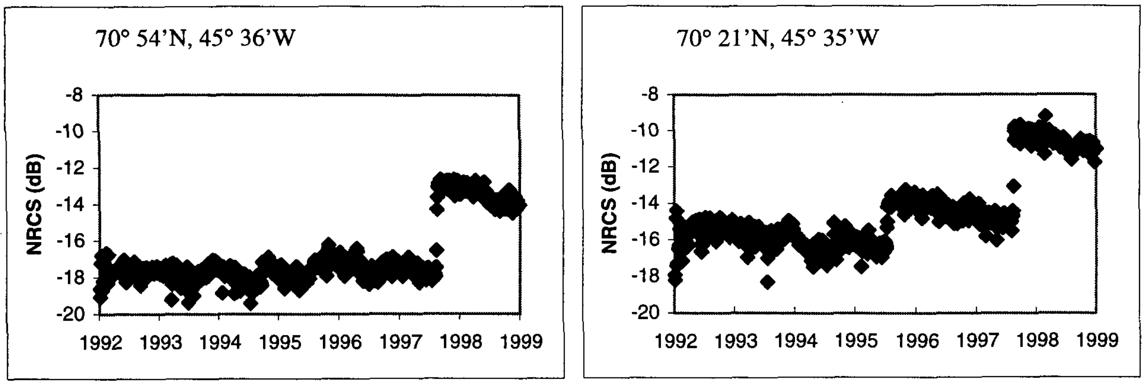

Fig. 2: NRCS time-series plots for two nearby locations.

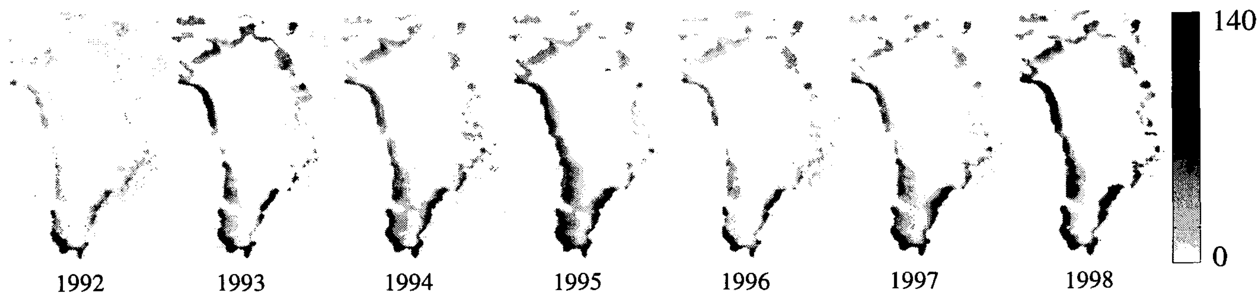

Fig. 3: Number of melt days

Counting Melt Days

Melt discontinuities occur in the percolation zone as a result of brief melt events and in the ablation zone as a result of more prolonged melting. In an effort to distinguish between these events, maps were generated showing the number of melt days at each location during each year. A melt day is defined as one in which the measured NRCS is significantly (at least eight standard deviations) less than the mean value for the given location during the winter months of 1992-1998 [2]. These maps are given in Fig. 3. It should be noted that some of the ERS data from 1992 appears to be erratic. Therefore, less confidence should be placed in the first map in Fig. 1(a) and 3.

RADIATIVE TRANSFER MODELS

We developed an advanced RT model to interpret ERS and SSM/I data. The following results use data from 67° N 42° W, identified as point **A** here as well as in [1].

Active Model

First, we use a simple RT model to account for decrease in NRCS with snow accumulation. Fig. 4a shows the scenario after melt and refreeze with subsequent snow accumulation. Assuming independent Rayleigh scattering and absorption, snow extinction is 0.242-dB/m. The RT model quantifies the loss as $L = 2\,k_e\,d / \cos\theta$, where k_e is the extinction coefficient of the snow, d is layer depth, and θ is the propagation angle inside the medium. Accumulation in water equivalent (*WE*) is then $WE = L\,(\rho_i / \rho_w)\cos\theta / (2\,k_e)$, where ρ_i is density of pure ice, and ρ_w is density of water.

Applying a linear fit to ERS NRCS data before the melt event (1992 to 1995) provides a slope of -0.727-dB/year. Using this value for L gives a rate change of *WE* of 484-mm/year which agrees with the annual accumulation of 500-mm *WE* [3] at this point. The slope after the melt event is -0.610-dB/year, yielding a rate change in *WE* of 407-mm/year.

Next, we model the jump in NRCS using the scenario in Figs. 4b and 4c. Fig. 4b shows a layer of dry snow overlying strata of snow and ice. σ_0' is the backscatter value that would be measured if the top dry snow layer were not present, and σ_0 is the actual measured value with the snow layer present. Fig. 4c shows the situation after melt and refreeze. If the measured value of backscatter now is assumed to be the same as σ_0' (the values should be similar when averaged over many melt and refreeze cycles), then it appears that the jump is due to extinction in the snow layer. The depth required to explain the 1.5-dB jump in the ERS data is 2.7-m, representing a few years of accumulation.

Passive Model

A more sophisticated model is required to interpret SSM/I data at 37 GHz due to increased scattering. Here, we apply a passive numerical RT code [4] to account for this effect.

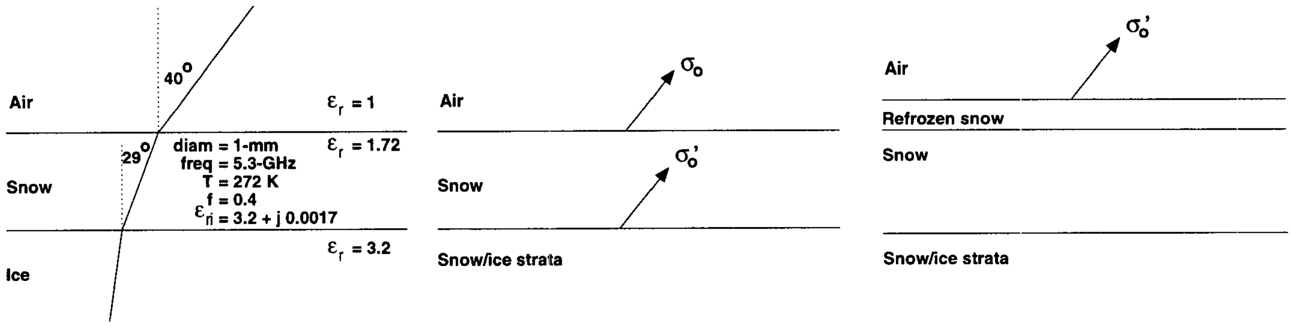

Fig. 4: Models used to account for active NRCS data. From left to right: (a) Simple one-layer model to account for NRCS decrease with accumulation, (b) and (c) model of media before and after melt and refreeze.

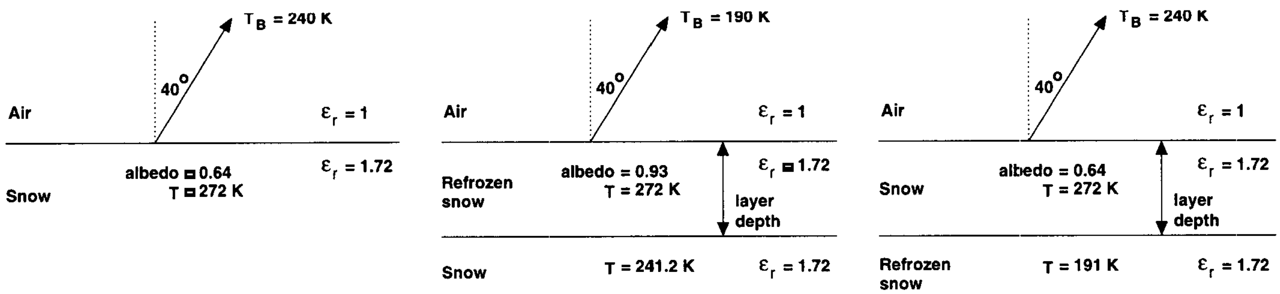

Fig. 5: Models used to account for passive SSM/I data. From left to right: (a) half-space snow model present before melt event, (b) single layer scattering model after melt and refreeze, (c) model used to explain increase in T_B with accumulation.

A 50K drop in T_B is observed at **A** in the SSM/I data coincident with the melt event. Fig. 5a shows a half-space of dry snow assumed to be present before the melt event. The brightness temperature (T_B) at point **A** was approximately 240K before the melt event. The passive RT code requires a scattering albedo of 0.64 for the snow medium to match the 240K value. For simplicity, we replace the half space of snow with a homogeneous medium giving the same T_B value. At 40°, transmissivity at the snow/air interface is 0.995, requiring a homogenous medium with a physical temperature of 241.2K.

Fig. 5b shows the situation present after the melt event. The refrozen surface is assumed to consist of coarse-grained snow having the same effective permittivity as the dry snow. Scattering albedo for the refrozen snow is estimated using a numerical QCA-CP code [4]. The ice grains in the dry and refrozen snow were assumed to have $\varepsilon_r = 3.2 + j\,0.0017$ and a fractional volume of 0.4. The QCA-CP code gives a scattering albedo of 0.64 and $k_e = 0.0243/\lambda$ for ice grains of radius 0.44-mm. If the ice grains double in size as a result of melt and refreeze, the QCA-CP code predicts that albedo increases to 0.93. In order to match the drop to 190K after melt and refreeze, the radiative transfer code requires a layer depth of 50-cm.

Correlating increase in T_B after refreeze with accumulation is difficult due to seasonal temperature variation. A simple approach is to compare T_B at the same month in consecutive years. Fig. 5c shows a model for a layer of dry snow overlying the refrozen snow. At the same point in the year following the melt event (1996), T_B is measured at 240K. The RT model requires an optical thickness of 4, or 1.33-m (492-mm *WE*) for $k_e = 0.0243/\lambda$. Recall, however, that a half space of snow also gave T_B = 240K, indicating that accumulation is sufficient to saturate the value of T_B after a single year.

CONCLUSION

This paper discusses use of active and passive data to explore snow ablation and accumulation on the Greenland ice sheet over a 7-year period (1992 to 1998). Data observations allow characterization of melting extent. Coupling of the data with analysis from a RT model aids in quantification of melting and accumulation.

REFERENCES

[1] V. Wismann, D. P. Winebrenner, K. Boehnke, R. J. Arthern, "Snow Accumulation on Greenland Estimated from ERS Scatteromer Data," Proc. of the Intl. Geosci. and Remote Sensing Symp., Singapore, 3-8 August, pp. 1823-1825, 1997.

[2] I. Ashcraft and D. Long, "SeaWinds views Greenland," Proc. of the Intl. Geosci. and Remote Sensing Symp., Honolulu, HI, 24-28 July, 2000.

[3] A. Ohurma and N. Reeh, "New precipitation and accumulation maps for Greenland," Journal of Glaciology, vol. 37, no. 125, pp. 140-148, 1991.

[4] L. Tsang, J. A. Kong, R. T. Shin, Theory of Microwave Remote Sensing, New York: Wiley, pp. 258-272, pp. 498-505, 1985.