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TOWARDS THE DEVELOPMENT OF A LAND DATA ASSIMILATION SYSTEM FOR SNOW

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The present paper introduces the first steps done towards the development of a land data assimilation system for snow, based on the assimilation of passive microwave brightness temperature (TB) observation. It introduces the coupling of a land surface model (JMA-SiB) used by the Japanese Meteorological Agency (JMA) and a radiative transfer model for snow.

In a first step JMA-SiB was extended to consider the change of the snow grain during the winter season, which have a significant impact on the TB observation and the radiative transfer in snow. The model results indicate, that the new version is capable to predict the snow grain size. Furthermore a first version of the coupled system was applied to data observed during the Cold Land Processes Field Experiment (CLPX) in Boulder Colorado (2003), to assimilate the TB change at 89 GHz after snowfall events and the results are in agreement with the observed change of the snow water equivalent at the site.

Key Words **:** *Land Data Assimilation, Snow Radiative Transfer, Snow Model, Snow Grain Metamorphism, Cold Land Processes Field Experiment, Shuffeled Complex Evolution*

1. INTRODUCTION

Due to the high albedo and thermal insulation of snow, it an important role in the global energy and water balance, e.g. snow changes the runoff characteristics of a catchment and influences the soil moisture and evaporation[1)].

Up to 53% of the northern hemisphere and up to 44% of the world land mass can be covered with snow at any given time[2)] and world wide one third of the water used for irrigation is temporarily stored as snow[3)].

The Climate and Cryosphere (CliC) Project stated, that "Knowledge of the amount, distribution, and type of precipitation and its temporal and spatial variability on a wide range of scales, is essential for the study of cold climate and related hydrological processes"[4)].

To increase the temporal and spatial resolution of snow data, several approaches are possible. Passive microwave remote sensing observations are sensitive to the size and number of the snow particle and therefore provide information about the current snow pack conditions. This is used in current satellite algorithm to estimate the snow depth or the snow water equivalent[5)].

If meteorological forcing data is available, e.g. from observations or in the form of model output, it is also possible to use a snow model or a land-surface scheme, which try to predict the evolution of the snow cover, based on the available forcing data.

Both approaches have their own drawbacks. Current satellite snow products, do not consider the effect of the snow metamorphism and in the case of snow models, the instability of long-term model runs due to the accumulation of model errors or inaccurate initial conditions. Furthermore, in case of a forward modeling approach, accurate forcing data is necessary, especially the amount of precipitation is a very important input parameter.

In this paper the first steps towards the development of a land data assimilation system are introduced, which tries to combine the merits of satellite observations and forward models. The paper focuses on the extension of JMA-SiB to consider the snow grain metamorphism, an important part for calculating the radiative transfer model in snow. The land surface scheme is then coupled with a radiative transfer model. The coupled system is applied to data collected during the Cold Land Processes Experiment for the assimilation of new snow on the ground[6)].

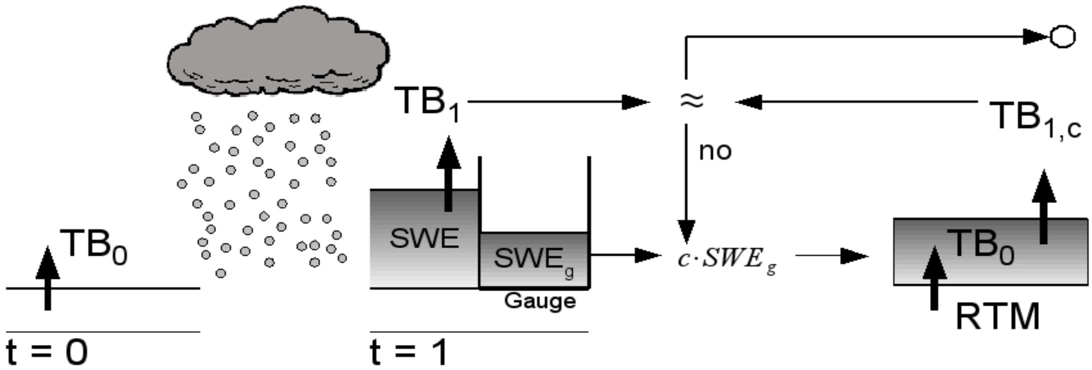

Fig. 1 New Snow Assimilation

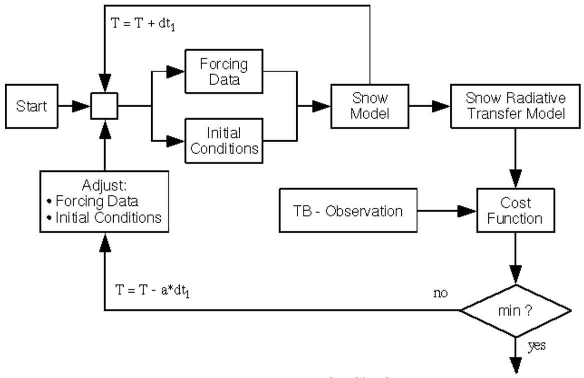

Fig. 2 Data Assimilation

2. METHODOLOGY

(1) Data Assimilation

a) Overview

Brightness temperature observation in the microwave region, are sensitive to snow on the ground. This is mainly due to the scattering effect of the snow particles, which is expressed by the number and size of the particles. Therefore the TB observations are providing information about the state of the snow pack. The basic idea of data assimilation is, to repeatedly introduce these informations into a land-surface or snow model, to get improved initial conditions for the model run.

Fig. 2 provides a general overview of the assimilation process. The snow model, which is called model operator, is used with observed forcing data and initial conditions as input to predict the evolution of a snow cover until new radiometer observations are available. The predicted snow pack state is then used as input parameter for the radiative transfer model, which is called the observation operator.

The observation operator calculates the brightness temperature for the predicted snow pack structure, which would be observed by a radiometer. By changing the initial conditions of the model operator in a predefined range, this step will be repeated until the estimated and observed brightness temperatures are in close agreement. To minimize the cost function of this algorithm the shuffled complex evolution method (SCE) will be introduced. The parameter range of the initial condition can for example be decided based on the variance of the parameter. Which can be either known from experience and/or historical data.

Similar approaches have been for example successfully used to assimilate passive microwave satellite observation to determine the spatial and temporal variation of soil moisture and soil temperature[7].

b) New Snow Assimilation

In a first step we implemented an approach to assimilate the change of brightness temperature observation at 89 GHz to estimate the amount of new snow on the ground[6]. This is done by correcting observed data from a precipitation gauge, which often only observes a fraction of the actual precipitation, due to wind-induced under-catch or evaporation

The brightness temperature at 89 GHz is showing a high sensitivity to the amount of fresh snow on the ground. Therefore it is possible to relate the change in the brightness temperatures at this frequency between two different times (at $t=0$ and $t=1$) to the accumulated new snow. The advantage of this algorithm is, that it is not necessary to know the physical properties of the old snow. In addition the fresh snow on the ground can be assumed to have a homogeneous vertical profile, which further simplifies the problem.

Fig. 1 provides an overview of the algorithm. The gauge data (SWE_g) and the first guess of the correction factor (c) are input parameters for a snow model. The result of the snow model and the brightness temperature for the old snow (TB_0) are then used to calculate the radiative transfer in the fresh snow. The results of the radiative transfer model ($TB_{1,c}$) are then compared with (TB_1). During the assimilation process, the correction factor for the observed solid precipitation will be updated, until $TB_{1,c}$ is in good agreement with TB_1.

An important assumption in this algorithm is, that TB_0 does not change between the two observations.

Snow grain size, density and temperature are parameters, which can influence the brightness temperature. The change of the grain size and the density are small within the selected time period between $t=0$ and $t=1$, e.g. 1 day, and therefore the effect of the change on TB_0 can be neglected.

c) Assimilation of Lower Frequencies

Current satellite sensors (e.g. AMSR-E) are providing TB observations at lower frequencies like 18.7 GHz and 36.5 GHz (e.g. AMSR-E), those frequencies are also sensitive to the snow on the ground, but show only a low sensitivity to fresh snow. Therefore the next step in the development of the land data assimilation scheme will be the introduction of TB observations at lower frequencies. This will make it necessary to consider longer assimilation steps and therefore the effect of the snow grain growth becomes more important and needs to be considered.

(2) Model Operator – JMA-SiB

a) Overview

We selected the new land-surface model developed by JMA as observation operator, called JMA-SiB[8] in this paper. The model is an extension of the Simple Biosphere (SiB) model developed by Sellers et. al.[9]. SiB was developed to express the transfer of energy, mass and momentum between the land-surface and the atmosphere, a part where snow plays an important role. The introduced changes are based on experience at JMA with using SiB as part of the global NWP model (JMA-GSM). The changes include the introduction of:

- Multiple Snow Layers and
- Sophisticated Snow Processes such as aging of albedo, temporal change of density and so on.

b) Snow Grain Metamorphism

The original JMA-SiB does not consider the effect of the snow grain growth, which is an important parameter for the radiative transfer in snow.

We implemented an grain growth model proposed by Rachel Jordan[10]. For dry snow, the grain growth is estimated based on the temperature gradient:

$$\frac{dr}{dt}=\frac{1}{4}\frac{g_1}{r}D_{eos}\left(\frac{1000}{P_a}\right)\left(\frac{T}{273.15}\right)^6 C_{iT}\left|\frac{dT}{dz}\right| \qquad (1)$$

where r represents the mean grain radius, D_{eos} the effective diffusion coefficient of water vapor in snow, C_{iT} the variation of saturation vapor pressure with temperature relative to ice, dT/dz the temperature gradient in the snow pack and g_1 $(=5.0\cdot 10^{-7} m^4/kg)$ is an empirical parameter.

For wet snow the following equations are used:

$$\frac{dr}{dt}=\frac{1}{4}\frac{g_2}{r}\cdot\left(\theta_l+0.05\right) \qquad \theta<0.09 \qquad (2)$$

$$\frac{dr}{dt}=\frac{1}{4}\frac{g_2}{r}\cdot 0.14 \qquad \theta\geq 0.09 \qquad (3)$$

where g_2 $(=4.0\cdot 10^{-12} m^2/kg)$ is an empirical coefficient and θ_l the snow wetness.

(3) Observation Operator - MEMLS

The Microwave Emission Model of Layered Snowpack was used as observation operator. The model was explicitly developed for radiative transfer modeling in snow and successfully applied[11,12].

(4) Minimization Scheme

a) Cost Function

The assimilation scheme is used to minimize the cost function J by adjusting the state vector x[13]. In general J can be separated into two different costs, one represents the background error J_B and the other one the observation error J_0:

$$J=J_B+J_0 \qquad (4)$$

For this application, we currently neglect the background error, and only use the observation error J_0, which usually expresses the difference between the observed and modeled values by considering the error covariance matrix R:

$$J(x_{0,}f)=\frac{1}{2}\sum_{i=1}^{N}\left(H[M(x_{0,}f)]-y_i^0\right)^T R^{-1}\left(H[M(x_{0,}f)]-y_i^0\right) \qquad (5)$$

y_i^0 is a vector representing the satellite observation at time i, which will be assimilated. H is the radiative transfer model (observation operator) and R is the error covariance matrix of the observation. M is the model operator, which calculates the model state at i from initial conditions and forcing data f.

b) Shuffled-Complex Evolution

During the assimilation the cost function J (Eq. 5) is minimized by adjusting the amount of solid precipitation within one assimilation window. This is done using the Shuffled-Complex Evolution method (SCE)[14].

The method involves the evaluation of the function usually at a random sample of points in the feasible parameter space, followed by a subsequent manipulation of the sample using a combination of deterministic and probabilistic rules.

3. APPLICATION

(1) Dataset from CLPX 3

The data[15),16] used in this study was obtained during the IOP 3 of the Cold Land Processes Field Experiment (CLPX) at the Local-Scale Observation Site (LSOS) in Fraser, Colorado, USA. The objective of the CLPX was to improve our

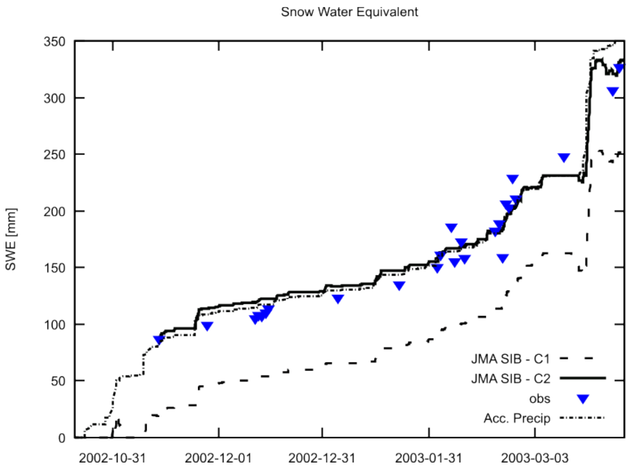

Fig. 3 SWE for Case 1 and Case 2.

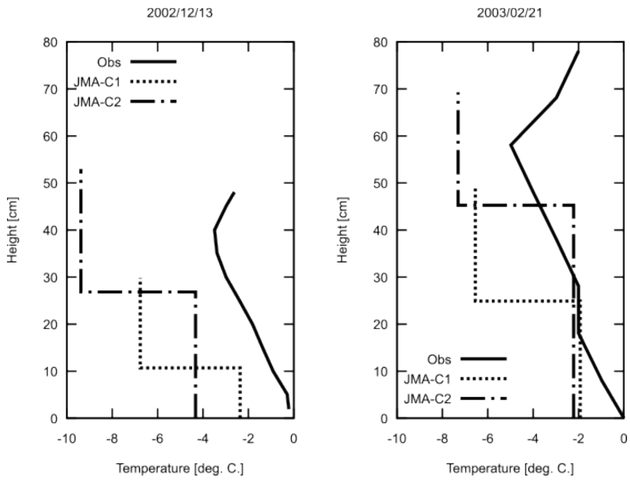

Fig. 5 Temperature Profile for Case 1 and Case 2.

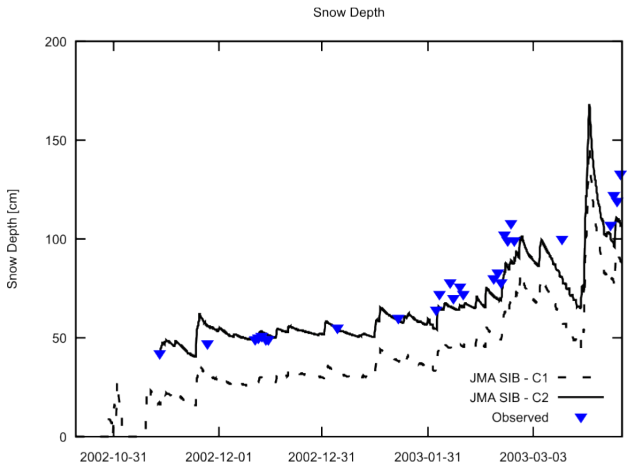

Fig. 4 Snow Depth for Case 1 and Case 2.

understanding of the terrestrial cryosphere, using a multi-sensor, multi-scale approach[17].

Within the CLPX the LSOS was used to implement very detailed observation of the local snow conditions, soil properties, vegetation and energy balance characteristics, which allow the investigation of scaling issues between ground based observations and airborne-/satellite-based sensors. Hardy et. al.[18] provides a complete overview of all data collected at the LSOS.

(2) Modeling Results

In this section we introduce results of the extended version of JMA-SiB.

The used forcing data set provides observation between 1 October 2002 and 29 March 2003. Two different cases are presented for the first case (C1) the model was initialized before the snow accumulation started (Oct. 15) and the second case (C2) was initialized after the first snow pack data was collected at Nov. 13, 2002.

For all application the raw precipitation data was used, not considering possible losses due to e.g. wind induced under-catch.

a) Snow Depth & SWE

Fig. 3 and **Fig. 4** show the modeled snow water equivalent (SWE) and snow depth for two different cases and compare the results to observed data. **Fig. 3** furthermore includes the total accumulated observed precipitation.

For C1 the snow, which accumulated around Oct. 31, melted at the beginning of November. This error propagates through the results for the whole winter season.

As it can be seen for C2, using the observation data as new (improved) initial conditions significantly improves the model results. The snow depth is well represented by JMA-SiB for the whole winter season, but especially in 2003, the snow depth is a little bit underestimated, which is caused by an overestimation of the snow density.

(b) Snow Pack Temperature

Fig. 5 shows a comparison of the snow pack temperature using data from Dec. 13 and Feb. 21. In both cases JMA-SiB is underestimating the snow pack temperature, the gap is especially large for Dec. 13.

The model currently does not consider the effect of the snow micro-structure on the heat transfer in the snow pack. This might be a possible reason for the difference between the modeled and observed snow temperature. In Fraser rather large depth hoar snow crystals are expected, which will reduce the thermal conductivity of the snow cover[19].

(c) Grain Size

As it can be seen from **Fig. 6** the observed and modeled grain size is in a good agreement, but the model seems to overestimate the snow grain size.

Eq. (1) shows the importance of the snow pack temperature for the calculation of the grain growth and one reason could be the underestimation of the snow pack temperature, which can lead to a larger temperature gradient in the snow cover and therefore to a faster snow grain growth.

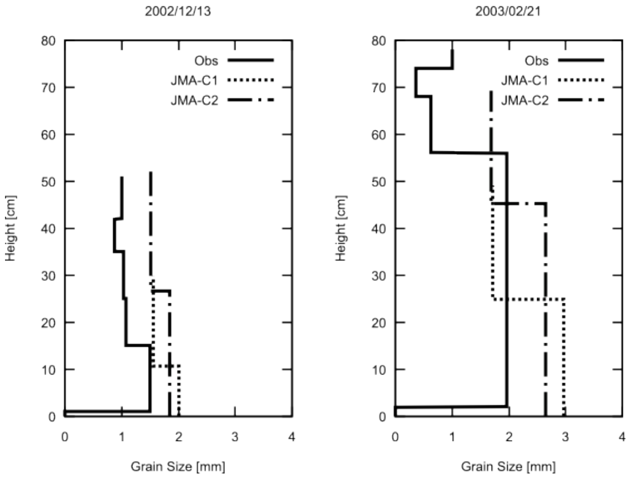

Fig. 6 Grain Size for Case 1 and Case 2.

(3) Assimilation Results

In this section the results of the new snow assimilation scheme using radiometer data observed during IOP3 of the CLPX[16] are introduced.

The assimilation system runs between Feb. 21, 12:00 to Feb. 25, 14:45. During this period a total of 23 mm snowfall was recorded at the precipitation gauge. The LSOS snow pit data was collected close to the radiometer at two different locations, each location was sampled every second day.

The snowpit sites where located in a large forest clearing of the LSOS. The distance between L1 and L2 was about 15 m, but 'L1' was closer to the forest edge, whereas 'L2' more in the open field[15].

Table 1 provides an overview of the observed SWE for each location, the change between two samples (ΔSWE). The column 'Precip.', indicates the amount of snowfall since the previous day. The data which was used as initial conditions is marked gray. During the days before, no snowfall was observed.

Comparing the SWE at Feb. 21 with the one at Feb. 19, there seems to be a loss of 23 mm, the only possible explanations for this are a re-allocation of the snow due to strong winds, or an error in sampling the snow pack. If the observation at Feb. 19 is considered as a reference, ΔSWE is 21 mm, rather then 44 mm.

For the assimilation two different cases have been considered. For the first case, the assimilation window was selected to cover one day ('A1'), which includes one set of TB and for the second case the assimilation window was two days ('A2'), including two sets of TB observations.

The results are summarized in **Table 2**. Because the SWE change between the snow pits can only be compared for the same location, the assimilation results have been accumulated to cover the same periods. In **Table 2** 'A1' and 'A2' correspond to the accumulated new snow on the ground, 'Δ' shows the change of the snow pack SWE and 'P' the accumulated precipitation.

Table 1 Observed SWE [mm] and Raw Precipitation [mm]

Date	***SWE***		***ΔSWE***		***Precip.***
	L1	L2	L1	L2	
Feb. 19, 12:10		182.0			
Feb. 20, 15:30	189.0				0.0
Feb. 21, 11:30		159.0		-23.0	0.5
Feb. 22, 11:00	206.0		17.0		10.4
Feb. 23, 11:00		203.0		44.0/21.0	4.8
Feb. 24, 13:30	229.0		23.0		4.8
Feb. 25, 11:00		211.0		8.0	3.0

Table 2 Assimilation results [mm]

Period	***L2***				***L1***		
	Δ	***A1***	***A2***	***P***	***Δ***	***A1***	***P***
Feb. 21 – 22	44.0/				17.0	22.6	10.4
Feb. 22 – 23	21.0	23.0	23.3	15.2			
Feb. 23 – 24					23.0	14.3	9.6
Feb. 24 – 25	8.0	46.1	37.4	7.8			

For the period of Feb. 21 to 23 (at L2), both the observed precipitation and also assimilation results are much lower than the observed change in the SWE. On the other hand, if the snow pit data from Feb. 19 is considered, the SWE change is only 21 mm, which is in good agreement with the assimilation results.

For the period of Feb. 21 to 22 the assimilation system overestimates the ΔSWE by 5.6 mm and for Feb. 22 to 24, the assimilation underestimates the change by 8.7 mm, still the results appear to be better than the observed precipitation, for which the observed snowfall was lower by 6.6 and 13.4 mm.

Only for the period between Feb. 23 and Feb. 25, the assimilation results are much worse than the observed precipitation, the system strongly overestimates the snowfall. Comparing 'A1' and 'A2' for 'L2' shows, that 'A2' seems to provide a little bit better results for the period between Feb. 23 and Feb. 25. In this case the observed the brightness temperature change does not correspond to the results from the radiative transfer model. This could indicate a problem with the radiative transfer model and further investigation needs to be done to analyze the results.

4. DISCUSSION & OUTLOOK

The application of JMA-SiB to the CLPX data showed good results of the land-surface model for the snow conditions in Fraser. Except for the temperature a good agreement was achieved. For the two selected cases, the observed snow pack temperature is underestimated, one possible reason is that JMA-SiB does not consider the effect of the snow micro-structure on the thermal conductivity of the snow. The error in the snow temperature can be the reason for the slight overestimation of the snow grain size. By improving the estimation of the thermal conductivity, the representation of the snow temperature might be improved, which might lead to an improved representation of the snow micro-structure. More efforts need to be done, to analyze the heat transfer in the snow pack.

The estimated amount of new snow from the assimilation is, with the exception of one case, closer to the SWE change in the snow pack, than the observed precipitation data. One problem can be the change of the new snow, from previous assimilation steps, during the assimilation period.

The current system only uses the TB data at 89 GHz but current satellite systems do also provide observation at lower frequencies, e.g. for example 18.7 and 36.5 GHz (AMSR-E). The lower frequencies are not sensitive to new snow and therefore can be used independent of the new snow assimilation system, but in this case the consideration of the snow metamorphism is necessary and an important step was done by introducing the snow grain growth.

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