SIMULATION OF INTERANNUAL VARIABILITY OF SNOW COVER AT VALDAI (RUSSIA) USING A DISTRIBUTED BIOSPHERE HYDROLOGICAL MODEL WITH IMPROVED SNOW PHYSICS

Maheswor SHRESTHA¹, Lei WANG² and Toshio KOIKE³

¹Student Member of JSCE, Ph. D. Student, Dept. of Civil Eng., University of Tokyo (Bunkyo-ku, Tokyo 113-8656, Japan)

²Member of JSCE, Dr. Eng., Research Associate, Dept. of Civil Eng., University of Tokyo (Bunkyo-ku, Tokyo 113-8656, Japan)

³Member of JSCE, Dr. Eng., Professor, Dept. of Civil Eng., University of Tokyo (Bunkyo-ku, Tokyo 113-8656, Japan)

The distributed biosphere hydrological model (WEB-DHM) was physically improved in snow physics, by incorporating the 3-layer energy balance based snow module of the Simplified Simple Biosphere 3 (SSiB3) model and albedo module of the Biosphere Atmosphere Transfer Scheme (BATS). The new version of WEB-DHM is referred to as WEB-DHM-S. In this study, the WEB-DHM-S is evaluated at Valdai site (Russia) from 1966 to 1983, to check its performance in simulating the interannual variability of snow cover. Results show that the WEB-DHM-S is capable of reproducing the point-scale interannual variations of snow water equivalent, soil and snow surface temperature well. In addition, the good performance of WEB-DHM-S was further confirmed in simulating soil moisture (top 1 m) and evapotranspiration.

Key Words : WEB-DHM-S, energy balance, Valdai, interannual variability, snow water equivalent

1. INTRODUCTION

Seasonal snow cover is of great importance in understanding the energy and water balance in cold regions. With its inherent properties such as high albedo and low thermal conductivity, it acts as an insulating body for the exchange of energy fluxes between the atmosphere and the land surface. It has ability to store or release water within the hydrological cycle and hence influences the temporal variability of spring snowmelt runoff.

Many studies have been conducted for better representation of snow physics in several climate and hydrological models^{1,2,3,4,5,6,7,8,9}. Some snow model intercomparison studies have been carried out to address the issues related to the current state of snow modeling process^{10,11,12}. The complexity of existing snowmelt models in atmospheric and hydrologic research community ranges from simple degree day models² to advanced physically based energy balance based models^{1,3}. Degree day models are the

black box models while energy balance based models are physically based. One layered energy balance based snow melt models are extensively used for simplicity but these models can not simulate the complete snow physics and many studies^{7,10,13,14} concluded that at least three layered energy balance based snowmelt model is needed to simulate the diurnal cycle of snow accumulation and snow melting. Furthermore, study towards simulation of an interannual cycle of snow accumulation and melting process is very important to gain an improved understanding of snow processes and to verify the model physics in snow/soil vegetation atmosphere transfer (SVAT) schemes.

This study attempts to evaluate the performance of three layered energy balance based snow melt model in interannual scale. The model here used is Water and Energy Budget – based Distributed Hydrological Model with improved snow physics (WEB-DHM-S)¹³, developed by coupling three layered snow physics of Simplified Simple Biosphere 3 (SSiB3)^{14,15} and albedo scheme of

Biosphere Atmosphere Transfer Scheme (BATS)¹⁶ into Water and Energy Budget – based Distributed Hydrological Model (WEB-DHM)^{17,18}. The originality of the newly developed model is that it is capable of simulating the snow processes more accurately through the physically based multi layer energy balance modeling approach in the basin scale. However, this study is focused on evaluating the model performance in point scale for the long-term simulation of seasonal and interannual variability of snow cover as long-term high-quality datasets for snow model inputs and evaluations are currently not available over large regions.

2. STUDY AREA AND DATA

The study area is the grassland field of Usadievskiy catchment (0.36 km²) in Valdai, Russia (57°58' N, 33°14' E). 18 years of continuous meteorological forcing data (1966 – 1983) are used to drive the model. The complete dataset were downloaded from the global moisture data bank¹⁹. The atmospheric forcing data includes temperature, pressure, and humidity measured at 2m and wind speed at 10 m. These data along with precipitation were originally sampled at 3-h intervals and were interpolated to 30-min intervals using a cubic-spline interpolation method, but this research used all forcing data at 1-h interval. Downward shortwave radiation and longwave radiation were estimated based on observed cloudiness and temperature²⁰. The threshold temperature for snow/rain transition was taken as 0°C²⁰.

A continuous snow cover exists from early November to late April. The average annual temperature range is around 35°C and temperature falls below -10°C in winter. The soil water is frozen in winter. Average annual precipitation is 730 mm with the majority falling in summer and autumn months. The annual maximum snow water equivalent (SWE) value ranges from 120 to 170 mm. Evaluation dataset, used in this study, includes SWE, snow and soil surface temperature, total soil moisture in top 1 m and evapotranspiration (ET). Details of the dataset were given in Schlosser et al.²⁰. This site was also used in the Project for Intercomparison of Land Surface Schemes (PILPS) - Phase 2(d)^{10,20} to investigate the snow processes in 21 land surface models.

3. MODEL

WEB-DHM-S is the improved version of Water and Energy Budget – based Distributed Hydrological Model (WEB-DHM), with the improvements in the snow physics by coupling three

layered energy balance based snowmelt module of SSiB3 and albedo scheme of BATS. WEB-DHM-S can simulate the variability of snow density, snow depth and snow water equivalent, liquid water and ice content, snow albedo, snow layer temperature and thermal heat due to conduction.

In WEB-DHM-S, the snow parameterizations for the canopy are kept the same as in WEB-DHM, but the three layer energy balance based snow scheme is activated when the snow depth is greater than 5 cm. Initially, the snowpack is divided into three layers that start with the same initial snow temperatures. The top layer thickness is kept at a fixed depth of 2 cm regardless of the total snow depth to provide reasonable simulation of the diurnal changes in the snow surface temperature. The maximum thickness of the middle layer is kept at 20 cm, and the bottom layer represents the remaining body of the snowpack. A surface energy balance equation is formulated only for the top layer, which is influenced by the surface radiation budget and sensible and latent heat fluxes. The heat budget of the second and third layers is controlled by the heat conduction and the penetrating shortwave radiation. And the mass budget for each layer is calculated accordingly by taking account of the precipitation, evaporation, condensation, compaction, liquid water retention, snowmelt runoff and infiltration into the underlying layers. The soil model coupled with a three layer snow model in WEB-DHM-S is shown in **Fig. 1**.

The equation for enthalpy of each snow layer is

$$\frac{\partial H(Z_j)}{\partial t} = - \frac{\partial G_{sn}(Z_j)}{\partial Z} \quad (1)$$

where H (Jm⁻³) is the volumetric enthalpy of water, Z_j is the snow depth of layer j and G_{sn} (Wm⁻²) is the heat flux through the snow layer. H is defined as

Fig. 1 Coupled soil/snow model in WEB-DHM-S.

$$H(Z_j) = C_v(Z_j) \times \{T_{sn}(Z_j) - 273.16\} - f_{ice}(Z_j) \times h_v \times \rho_s(Z_j) \quad (2)$$

where C_v ($\text{Jm}^{-3}\text{K}^{-1}$) and T_{sn} (K) are mean snow volumetric specific heat capacity and snow temperature respectively. C_v is parameterized as a function of the bulk density of snow and intrinsic density of ice following Versegny⁴. f_{ice} is the dry-snow mass fraction of the total mass in the snow layer, h_v (Jkg^{-1}) is the latent heat of fusion for ice and ρ_s (kgm^{-3}) is the bulk density of snow. G_{sn} is defined as

$$G_{sn}(Z_j) = \begin{cases} R_{nsn} - H_{sn} - \lambda E_{sn} + G_{pr}, & j = 3 \\ K(Z_j) \frac{\partial T_{sn}(Z_j)}{\partial Z} + SW_{sn}(Z_j), & j = 2, 1 \end{cases} \quad (3)$$

where R_{nsn} (Wm^{-2}), H_{sn} (Wm^{-2}), λE_{sn} (Wm^{-2}), G_{pr} (Wm^{-2}), K ($\text{Wm}^{-1}\text{K}^{-1}$) and SW_{sn} (Wm^{-2}) are net radiation, sensible heat, latent heat flux, thermal energy from rain at the snow surface, thermal conductivity of snow and shortwave radiation flux absorbed by the snow layer respectively. The shortwave radiation SW_{sn} at the snow layer is defined following Jordan³ accounting the penetration of shortwave radiation flux into the snow layers. The process of the calculations of R_{nsn} , H_{sn} and λE_{sn} is the same as Sellers *et al.*²¹ except that the snow surface temperature is used instead of the average bulk snow temperature for the surface energy balance. The thermal conductivity of snow K ($\text{Wm}^{-1}\text{K}^{-1}$) is calculated from bulk density of ice using the formulations adopted by Jordan³. Thermal energy from rain (G_{pr}) can be calculated as

$$G_{pr} = \rho_w \times C_w \times (T_{rain} - 273.16) \times IF_0 \quad (4)$$

where IF_0 (ms^{-1}) is the infiltrated flux rate of rain at the snow surface, T_{rain} (K) is the temperature of rainfall, ρ_w (kgm^{-3}) and C_w ($\text{Jkg}^{-1}\text{K}^{-1}$) are the density and specific heat capacity of water. For simplicity, T_{rain} is considered as air temperature (T_{air}). Ground surface temperature (T_g) and deep soil temperature (T_d) are obtained by considering conductive heat flux at the snow/soil interface and the force-restore model²² of the heat balance in the soil surface.

$$C_g \frac{\partial T_g}{\partial t} = -K(Z_1) \frac{\partial T_{sn}(Z_1)}{\partial Z} - \frac{2\pi C_g (T_g - T_d)}{\tau_d} \quad (5)$$

$$C_d \frac{\partial T_d}{\partial t} = \frac{2\pi C_g (T_g - T_d)}{\tau_d \sqrt{365\pi}} \quad (6)$$

where C_g and C_d are the effective heat capacity ($\text{Jm}^{-2}\text{K}^{-1}$) for the soil surface and deep soil, τ_d is the day length (s) and $K(Z_1)$ is the effective thermal conductivity at the snow/soil interface. T_d is the deep soil temperature.

The mass balance for snow is represented by the change in liquid water and ice content in the snowpack. The relative change in snow mass is

controlled by snowfall/rainfall, compaction, snow melting, runoff, infiltration into the underlying snow layer/soil and evaporation/sublimation at the snow surface. The mass balance equations for the snow layer are

$$\frac{\partial M_{snow,j}}{\partial t} = \begin{cases} P_s + IF_0 - IF_j - R_j - E_{sn}, & j = 3 \\ IF_{j+1} - IF_j - R_j, & j = 2, 1 \end{cases} \quad (7)$$

where $M_{snow,j}$ (m) corresponds to the SWE at snow layer j , P_s (ms^{-1}) is the rate of snowfall, IF_j (ms^{-1}) = $\min(O_j, P_{avs})$, is the actual liquid water infiltration flux at the interfaces, R_j (ms^{-1}) is runoff from the lower interface and E_{sn} (ms^{-1}) is the combined evaporation and sublimation rate. O_j is the outflow flux rate which is the liquid water drained to the underlying layer as the total liquid water in layer exceeds its liquid water holding capacity (C_r). Liquid snow mass fraction, $f_{liq} = (1 - f_{ice})$ is used to calculate the total amount of liquid water. P_{avs} is the pores available in the layer. R_j is calculated as the difference between IF_j and O_j . Three snow compaction processes, namely destructive metamorphism, densification due to snow overburden and compaction due to snow melting, are parameterized following Anderson¹. Compaction processes are accounted during the calculation of IF_j . C_r is taken as a function of the snow layer density following Anderson¹. The bulk density of ice for new snowfall is calculated following the formulation used in the CROCUS snow model⁵:

$$\gamma_i = 109 + 6 \times (T_{air} - 273.16) + 26 \times \sqrt{u_m} \quad (8)$$

where u_m is the wind speed (ms^{-1}). Lower value of density of new snowfall is limited to 50 kgm^{-3} .

The snow albedo is parameterized using a physically based prognostic snow albedo scheme of the BATS¹⁶ model. The albedo is computed for VIS and NIR spectral bands with adjustments for illumination angle and snow age. The total snow albedo (α_s) is the weighted average of VIS and NIR albedo, which depends on the spectral ratio of the incident shortwave radiation. VIS and NIR albedo (α_{vis} , α_{nir}) are defined as

$$\begin{aligned} \alpha_{vis} &= \alpha_{vd} + 0.4 \times f_{zen} \times (1 - \alpha_{vd}) \\ \alpha_{nir} &= \alpha_{nird} + 0.4 \times f_{zen} \times (1 - \alpha_{nird}) \\ \alpha_{vd} &= \alpha_{vis0} \times (1 - 0.2 \times f_{age}) \\ \alpha_{nird} &= \alpha_{nir0} \times (1 - 0.5 \times f_{age}) \end{aligned} \quad (9)$$

where α_{vd} and α_{nird} are the albedo of the diffused shortwave radiation in the VIS and NIR bands respectively, α_{vis0} (0.85) and α_{nir0} (0.65) represent fresh-snow albedo for the VIS and NIR bands, f_{zen} is the correction term for a solar zenith angle larger than 60° and f_{age} is the snow aging factor accounting for the effect of grain growth due to vapor diffusion and the effect of dirt and soot. Details of f_{zen} and f_{age}

Table 1 The parameters used in WEB-DHM-S at Valdai.

VIS(a_{vis0}) & NIR(a_{nir0}) albedo of fresh snow	0.85;0.65
VIS & NIR albedo of soil	0.10;0.20
Top soil, root soil and deep soil zone (m)	0.1;0.9;1.0
Vegetation coverage (%)	90.0
Bare soil coverage (%)	10.0
Canopy top height, z_2 (m)	0.60
Canopy base height, z_1 (m)	0.10
Roughness length of snow (m)	0.0024
Roughness length of soil (m)	0.035
Porosity of soil	0.401
Sat. hydraulic conductivity for soil (ms^{-1})	0.00002
Sat. hydraulic conductivity for snow (ms^{-1})	0.01
Zero plane displacement height (m)	0.25
Threshold temperature to separate rain/snow ($^{\circ}\text{C}$)	0.0

can be found in Yang *et al.*¹⁶⁾. The soil moisture and ET are calculated as SiB2²¹⁾. Model parameters specific to Valdai site are given in **Table 1**. The model is run from January 1966 to December 1983 with initial conditions of observed soil temperature, soil moisture and SWE. The model performance is evaluated by Mean Bias Error (MBE) and Root Mean Square Error (RMSE), defined as followings;

$$MBE = \sum_{i=1}^n (S_i - O_i) / n \quad (10)$$

$$RMSE = \sqrt{\sum_{i=1}^n (S_i - O_i)^2 / n} \quad (11)$$

where n is the total number of time series, S_i is the simulation result and O_i is the observed data.

4. RESULTS AND DISCUSSIONS

(1) Snow Water Equivalent (SWE)

SWE is one of the very important variables in snow hydrology as it is a direct measure of the total liquid water available in the snowpack. Proper simulation of SWE is critical in predicting seasonal discharge. The comparison of SWE simulated by WEB-DHM-S and the observed monthly SWE values are shown in **Fig. 2**. The observed value represent the catchment averaged SWE based on snow course measurements from 44 individual sites throughout the catchment.

Fig. 2 shows that model is able to simulate long term interannual variability of snow cover well. SWE in 1967/68 and 1968/69 were less simulated following excessive simulation in some other years, similar to the study by Slater *et al.*¹⁰⁾. Over estimation of SWE in some years may be due to specific blowing snow conditions. Model BIAS and RMSE are given in **Table 2** and its value averaged over 18 year simulation, are given in **Table 3**. The model is able to simulate the observed low accumulation year (1971/72) and high accumulation year (1975/76)

quite well but it is unable to simulate mid-season ablation in the winters of 1970/71 and 1973/74 (**Fig. 2**), consistent to the results in PILPS-2d¹⁰⁾. We simulated the SWE by changing the threshold temperature to 1 and 2 $^{\circ}\text{C}$, but the bias in mid-season ablation is not improved (not shown here). The biases for many years may be basically due to the problem with the forcing, such as overcatch.

2) Skin Surface Temperature (SST)

SST is another important parameter of the land surface energy balance. It directly affects the sensible and latent heat fluxes. It is the key variable for the estimation of upward longwave radiation when the in-situ data are not available. The comparison between observed and simulated daily SST is shown in **Fig. 3**.

Fig. 2 Observed and simulated snow water equivalent (SWE) from hydrological year 1966/67 to 1981/82

Fig. 3 Observed and simulated skin surface temperature (SST) from hydrological year 1966/67 to 1981/82

Table 2 MBE and RMSE for SWE & SST (1966-1983)

Year	SWE (mm)		SST (°C)	
	MBE	RMSE	MBE	RMSE
66/67	2.31	10.24	-0.35	2.29
67/68	-9.18	16.81	-0.52	2.42
68/69	-20.03	26.5	-0.20	2.14
69/70	22.38	25.00	-0.62	2.41
70/71	15.73	28.69	-0.29	2.00
71/72	-8.57	14.37	-0.39	2.51
72/73	11.08	20.39	-0.29	2.07
73/74	12.19	29.58	-0.43	2.17
74/75	44.49	52.84	-0.54	2.20
75/76	5.42	23.99	-0.47	2.19
76/77	37.98	40.01	-0.66	2.06
77/78	2.74	14.25	-0.81	2.22
78/79	7.45	28.06	-0.56	2.41
79/80	10.21	16.10	-0.72	2.67
80/81	63.44	71.59	-0.80	2.43
81/82	-7.55	18.20	-0.27	2.30
82/83	43.57	49.48	0.10	2.12

Table 3 Statistic analysis of model simulation (1966-1983)

	MBE	RMSE
Snow water equivalent (mm)	12.09	31.62
Skin surface temperature (°C/daily)	-0.48	2.29
Soil moisture (mm/month)	12.11	30.17
Evapotranspiration up to 1973(mm/month)	-6.09	10.86

SST is calculated as the soil surface temperature in the absence of snow cover and as the snow surface temperature in the presence of the snow cover. The observed SST shows strong interannual variation in snow cover season. The overall seasonal and annual cycle of SST are well captured for the almost all months of the simulation years (1966/67-1981/82). This implies that the model is capable of simulating the surface energy budget over the soil and snow surface. MBE and RMSE for SST are given in **Table 2** and its value averaged for 18 years are found to be -0.48°C and 2.29°C respectively (**Table 3**).

3) Albedo

Albedo is a measure of fraction of solar radiation that is absorbed by the earth surface. It has significant contribution to the radiation budget during the snow melting period. Hence, the accurate simulation of snow cover period is highly governed by the snow albedo. **Fig. 4** shows the interannual variation of albedo values simulated by the model. The simulated daily albedo has a range of 0.2 to 0.8. As albedo is a function of grain size, age and snow surface temperature, the seasonal and interannual variability of albedo highly depend on the snow depth. The simulated albedo is relatively small in 1971/72 (with the lowest peak of SWE) and relatively big in 1975/76 (with the highest peak of SWE).

4) Soil moisture and ET

Fig. 5 compares the simulated and observed mean monthly soil moisture in top 1 m. The results

show that the model can capture the observed seasonal and interannual variability of soil moisture well in the pattern but accompanied with large bias. However, an increase in soil moisture due to snowmelt infiltration is reasonably simulated. The model is able to simulate relatively dry conditions in 1972 and 1975, although the model overestimated them. The remarkable positive bias may be due to the effect of frozen soil. Overall, MBE (12.11 mm) and RMSE (30.17 mm) values for monthly soil moisture are obtained for the 18-year simulation (**Table 3**). The monthly ET simulated by the model is shown in **Fig. 6** and compared with the measurements by Fedorov²³⁾ up to 1973. The results show that the capability of the model in simulating the seasonal cycle of ET is found quite well. The observed monthly-mean ET ranges from 4.2 mm in November to 102 mm in June. The ET during snow cover days is underestimated by the model whereas that in summer is comparable to the observations.

Fig. 4 Model simulated daily averaged albedo from 1966/67 to 1982/83 **Fig. 5** Observed and simulated mean monthly soil moisture in top one meter soil from 1966/67 to 1982/83 **Fig. 6** Observed (up to 1973) and simulated monthly ET from 1966/67 to 1982/83. **Fig. 7** Annual water balance from 1966/67 to 1982/83.

The annual water balance is shown in **Fig. 7**. The runoff is computed as the sum of surface runoff and outflow from vertical soil column. **Fig. 7** shows that the ET and runoff usually have equal weights to

the total precipitation. The 18 years averaged ratio of ET to the total precipitation is found to be 0.52. Remarkable positive storage was observed in the years 68/69, 70/71, 72/73, 75/76, 76/77 whereas negative storage was predominant in the years 69/70, 71/72, 74/75, 77/78, which mostly follows the soil moisture evolution as shown in **Fig. 5**.

5. CONCLUSIONS

In this study, the newly developed energy balance based snow melt model, WEB-DHM-S was evaluated at Valdai site for long term simulation of snow cover from 1966 to 1983. The model is found to be able to simulate reasonably well the seasonal and interannual variability of observed snow water equivalent (SWE), skin surface temperature (SST), soil moisture in top 1 m and ET. Model biases in some years may be attributed to the error in forcing data and/or blowing condition of snow. The WEB-DHM-S provides physical details of snow processes, which make it possible to continuously simulate long-term snow hydrology in the basin scale with a distributed manner. Furthermore, blowing snow module and advanced schemes for representation of frozen soil processes will be added to the current model for better representation of snow processes in future.

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