New Snow-Physics to Complement SSiB

Part I: Design and Evaluation with ISLSCP Initiative I Datasets

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Abstract

In evaluation(s) as a part of the Global Soil Wetness Project using ISLSCP Initiative-I data, the snow-melt in SSiB in the Russian Wheat Belt region (RWB) was found to be substantially delayed, with very deficient meltwater infiltration as compared to observations. Furthermore, most of the meltwater emerged as runoff, as opposed to soil moisture recharge. The deficiency emanated from the crudeness of snow-physics of the combined snow and ground layer of SSiB. In the current work, a new snow model employing a separate snow-layer was included. The snow-pack absorbs and transmits the incoming solar flux, thereby affecting the snow and ground temperatures through the winter and snow-melt periods. In the ISLSCP Initiative-I data evaluations, the snow-melt over the RWB region occurs 2–3 weeks sooner in the new model, and its soil thaws quite early in the snow-melt duration, which helps to infiltrate more meltwater into the soil. The new-model produces a more realistic simulation of soil-moisture, as well as Volga river runoff in RWB evaluations. Some residual delay in the snow-melt (varying from 1–4 weeks) seems to be related to the following: (1) inaccuracies in the satellite retrievals of snow under dense forest canopies; (2) the modeling assumptions, *e.g.*, neglecting the influence of snow aging on its thermal diffusivity, and simplifications in absorption of solar-flux in the snow cover, leading to an inadequate distinction between snow-pack surface and mean temperatures; and, (3) possible cold bias of the ISLSCP surface air temperature data.

1. Introduction

The ISLSCP Initiative-I project has provided a unique blend of global surface-datasets (Meeson *et al.*, 1995; and Sellers *et al.*, 1995) gathered from *in situ* observations and 4-dimensional data assimilation products produced by a 3-D VAR analysis. The datasets cover a two-year period, January 1, 1987 through December 31, 1988, and were released for the Global Soil Wetness Project (hereafter GSWP) to evaluate Land Surface Models (LSMs) (Chen and Mitchell, 1999). SSiB or Simple SiB (Simple Biosphere Model of Sellers *et al.*, 1986), as improved by Xue *et al.* (1991, and 1997), is currently used in several GCMs including our version of GEOS II GCM (Takacs *et al.*, 1994). SSiB was previously used in the GLA GCM (Sud and Walker, 1993). It is similar to the COLA (Center for Land Ocean Atmosphere Interactions) SSiB except for minor implementation differences. Mocko and Sud (1998) evaluated SSiB in GSWP and noted several satisfactory

features of its simulation, such as: (i) generally realistic annual cycle of soil moisture and runoff, (ii) a wilt-free (wilt-prone) time series of soil moisture in the time-transacts in forested (semiarid) regions such as Amazon rainforest (Sahel in north Africa), and (iii) an observationally well-verified annual cycle of soil moisture in the Illinois region. However, they also discovered a few deficiencies, particularly in regions where snow-melt infiltration is huge and occurs after a hard winter freeze, *e.g.*, the Russian Wheat Belt Region (55N–65N and 20E–60E, hereafter RWB), a subregion of the Russian observations in Vinnikov and Yeserkepova (1991). In RWB, SSiB produced virtually no soil-water recharge in the early spring (Fig. 1a). Snowmelt was accompanied by an excessive, and rather delayed spring-season runoff (also noted by R. Koster, personal communications, 1998). The peak runoff of SSiB and Volga River (largely fed by RWB) observations (Fig. 1b) shows huge errors, even though the time delay between grid-scale lateral flow and the Volga River runoff was ignored in this comparison. Moreover, SSiB simulates no runoff from October through April, whereas the Volga River yields persistent

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Fig. 1. Time series for Russian Wheat Belt region (RWB) of (a) available water in the top 1-meter of the soil (mm); and, (b) runoff (mm/day) of the Volga-river basin simulated by old SSiB vis-a-vis observations.

runoff for the period. Similar deficiencies were also noted by Oki *et al.* (1997) in other regions with large snow-melt. Through the analysis of SSiB-GCM simulations, similar soil moisture and snow-melt related deficiencies were isolated by Robock *et al.* (1995), while consistent snow-cover related deficiencies were reaffirmed by Foster *et al.* (1996).

In GSWP, the participating modellers opted for comparing Soil Moisture Indices (SMIs) — defined as a ratio of the actual to the maximum available soil moisture. However, these SMIs revealed large variations among the participating models. See, for example, the Lena river basin simulations with different land-surface models (Sato and Nishimura, 1999). Here again, the SSiB runoff is not only a month too late, but it is so huge that it virtually eliminates the proverbial early spring meltwater soil-moisture recharge — a vital necessity for realistic soil moisture and evapotranspiration in summer when the region receives the highest sunshine. Satellite observations also show very little snow cover after mid-May in RWB, while SSiB maintains snow cover till the end of June. Based on these findings, we concluded

that the snow-melt and snow hydrology of SSiB were deficient.

First, we ascertained that large spring-season runoff in SSiB occurred because its meltwater could not enter the frozen soil. We suspected this to be the consequences of: (i) ignoring the insulating effect of snow on soil temperature and hydrology; (ii) disallowing the physical influence of solar radiation that transmits through the snowpack and prepares the ground for infiltrating meltwater; and, (iii) neglecting refreeze of snow-melt reaching the frozen ground. These aspects of snowphysics have been addressed in the new snow-model.

Several studies have stressed the importance of an accurate soil moisture initialization on GCM simulations. Dirmeyer (1999) found that if the soil moisture is better initialized with GSWP data produced with COLA-SSiB, the COLA GCM has the ability to produce a more skillful climate forecast. Dirmeyer had reaffirmed the importance of prescribing a more accurate initial soil moisture field for the seasonal climate and rainfall prediction. Snow-model improvements can be expected to yield two benefits: i)

Table 1. Differences in Old and New Treatment of Ground Snow/Ice in SSiB.

ITEM	OLD	NEW	COMMENTS
1. Land Temp. Model	Force-restore with all surface fluxes influencing the ground temperature.	Force-restore using solar influx transmitted in and heat conduction from soil.	Ground gets insulated in winter while snow cools.
2. Snow Temp. Model	Snow and top layer of the ground have one temperature. All surface fluxes affect the temperature.	Snowpack is separated from ground. Ground interacts with the snowpack by heat conduction and LW radiation.	Snow can get much cooler while the ground remains insulated by the snowpack.
3. Snow Evaporation or Meltwater	Snow temperature is allowed to rise above freezing and the snow-melt occurs to restore the temperature to freezing.	Snow temperature is not allowed to exceed freezing. Reduces surface fluxes during melting.	It holds the surface fluxes during the snow-melt events in the new model.
4. Energy exchange between the ground and snowpack	None. Both have one temperature.	Occurs through heat conduction and longwave interaction through the air gap between the snow and the ground.	Snow has lower heat capacity as opposed to that of ground and snow as one.
5. Refreezing snow-melt	Not allowed. Leads to runoff of all snow-melt over frozen ground.	Snowmelt above the frozen ground refreezes and does not runoff.	Eliminates runoff over the frozen ground helping soil-moisture recharge.
6. Transition from snow-covered to bare ground	Heat capacity of snow which is an add-on to the ground; becomes first very small and then zero.	When the snow cover reduces to 0.5 kg m ² , the model transforms into a single layer model as in old SSiB.	Transition reduces the land-model to the old SSiB.

2.1 Solar transmission:

Observations show that approximately 10 % of the incoming solar-flux transmits through a 10-cm thick snow/ice layer (J. Foster, personal communications, 1997). The commonly used equation with an exponential transmittance function (Verseghy, 1991), typical of radiative transfer through a linear absorber, was written. The transmission to the ground, SW_g , is:

$$SW_g = SW_{sn} \exp(-23.0Z_{sn}/\rho_{sn}), \tag{1}$$

where Z_{sn} is the thickness of the snowpack in m , and ρ_{sn} is the specific gravity of snow/ice and SW_{sn} is the net solar radiation available for absorption at the snow surface.

2.2 Heat conduction

The heat conduction through the snowpack has three path-segments. The first segment is from the middle of the first ground-layer to its upper surface; the second segment is longwave transfer from the upper ground-surface to the bottom of the snowpack (air-gap transmission); and, the third segment

is from the bottom to the top of the snowpack. Assuming a constant heat flux, the conductance from the bottom of snow to the center of the ground-layer, κ_g , and the combined ground-to-snow conductance, κ_c , would be:

$$\kappa_g = (0.5Z_{sl}/\kappa_{sl} + (4\sigma T_m^3)^{-1})^{-1}, \tag{2a}$$

$$\kappa_c = (0.5Z_{sl}/\kappa_{sl} + (4\sigma T_m^3)^{-1} + Z_{sn}/\kappa_{sn})^{-1}, \tag{2b}$$

where conductivities, κ , and conducting medium thicknesses, Z , are for soil (sl) and snow (sn); σ is the Stefan-Boltzmann constant; T_m is the average of the ground and snow temperatures; and T_g , T_{sn} and T_f are ground, snow, and freezing temperatures, respectively. The heat conduction out of the ground, F_{gd} , is the larger of the following two fluxes:

$$F_{gd} = \kappa_c(T_g - T_{sn}), \tag{2c}$$

$$\text{or} \qquad \qquad \qquad = \kappa_g(T_g - T_f). \tag{2d}$$

Eqs. (2c) and (2d) give different answers only when the temperature at the bottom of the snowpack exceeds freezing (an unattainable temperature for

snow/ice) in the combined calculation. This happens when the temperature drop across the air gap is smaller than the temperature drop in the snowpack and the ground temperature is above freezing.

2.3 Snowmelt by ground-heat flux

The difference in the fluxes (one from the ground and other into the snow) causes snow-melt at the bottom of the snowpack. The bottom melt flux, F_{mb} , is the difference of the ground-to-snow flux: $F_{mb} = F_{gd} - F_{sm_{max}}$, and the maximum possible snow flux given by:

$$F_{sm_{max}} = \frac{\kappa_{sn}}{Z_{sn}}(T_f - T_{sn}), \quad (2e)$$

where T_{sn} is the snow temperature averaged for the whole snowpack.

2.4 Snowmelt refreezing

Total snow-melt flux is the sum of the component fluxes at the bottom and top of the snowpack:

$$F_{melt} = F_{mt} + F_{mb} \quad (4a)$$

where F_{mt} is the net available surface energy for melting snow/ice; it is equal to the sum of all the surface fluxes and rainfall. If the ground has freezing capacity, F_R , as determined by:

$$F_R = \frac{\kappa_g}{L}(T_f - T_g), \quad (4b)$$

a fraction of the snow-melt reaching the soil is assumed to refreeze into ice:

$$F_{ice} = \min(F_{melt}, F_R). \quad (4c)$$

There are two other important modifications. First, the snow fluxes are held constant for melting conditions; this eliminates fictitious warming of snow in the implicit backward snow temperature prediction equation (See Appendix). For conditions when $T_{sn} \cong T_f$, when the net energy is positive into the snowpack, we set $\partial H / \partial T_{sn} = \partial E / \partial T_{sn} = \partial LW / \partial T_{sn} \equiv 0$, where H , E , and LW represent sensible, latent, and emitted longwave at the surface. The second modification is that vapor pressure for sublimation of snow is corrected for saturation conditions over snow (and not water). All the other surface energy interactions remain identical to old SSiB (Sato *et al.*, 1989; and Xue *et al.*, 1991), and are described in the Appendix.

3. GSWP-GEWEX Simulations

The old version of SSiB, that was evaluated in the GSWP comparison, was integrated for several years by forcing the land repeatedly with 1987 ISLSCP Initiative I data until the soil moisture at the end of the integration cycle was within 5 % or 5.0 kg-m⁻² of the soil moisture at the beginning of the year. The spin-up integration continued up to 10

years or until equilibration, whichever was achieved first. The integration used an hourly timestep. The 6-hourly near-surface meteorology data was linearly interpolated to an hourly forcing data. The only exception to this is the incoming shortwave data, which was interpolated using the cosine of the solar zenith angle to produce a more realistic diurnal cycle. After the spin-up, the resultant soil moistures were used as the initial conditions for the following two-year integration, 1987–1988. This criteria was used by all the participating LSMs including SSiB in which snowphysics deficiencies have been identified. In fact, SSiB took a long time to reach hydrologic equilibrium. The new snow physics in SSiB was also integrated following the above criteria, and then for 1987 and 1988. Mocko and Sud (1998) and Mocko *et al.* (1999) provide more details of the integration procedure and the temporal resolution of the datasets. One of the problems with a hydrology evaluation project is the lack of reliable validation datasets. We decided to concentrate on the RWB region, which has valuable *in situ* soil-moisture and runoff datasets.

4. Results

We performed a comparison of the snow cover, soil-moisture, surface fluxes, and ground hydrology simulations with the new and old snowphysics in SSiB. We used the Russian Wheat Belt region (hereafter RWB) for verification, because it has useful validation datasets (Robock *et al.*, 1995). For snow-free regions, the two models are virtually identical and produce exactly analogous simulations of land-processes and hydrology. However, regions that have snowfall produce systematic differences even during the snow-free periods.

4.1 Simulated ground/snowpack temperatures

Figure 3a shows that in winter the new model produces a relatively small, but gradual cooling of land (a low temperature of about -3.0°C) beneath the snowpack. This is much warmer than the old model's ground and snow (a low temperature of -25°C). The snow in the new model can adjust to the ambient conditions much more quickly and follows the forcing air temperature closely. During the latter part of the spring melting, and the initial snow-accumulation period of early fall, the snow temperatures are near freezing, but air temperatures are above freezing (Fig. 3a). The warmer ground (see new vis-a-vis old physics simulations) in the beginning of spring requires less energy to reach the thawing temperature. The solar radiation absorbed by the ground beneath the snowpack helps to thaw it roughly 2–3 weeks earlier. This helps snow-melt infiltration. Accordingly, we see large differences in the apportionment of net radiation into sensible and latent energies in the following summer. Figure 3a

Fig. 3. Comparison of area-averaged fields for new and old snowphysics simulations vis-a-vis observations (if available) for RWB. The six panels are: (a) temperatures ($^{\circ}\text{C}$) for new model ground (green), old model snow covered ground (light blue), old model all ground (dark blue), new model snow (red), and forcing air temperatures (yellow); (b) fractional-area of snow cover for observations and model simulations; (c) sensible heat flux (W m^{-2}); (d) latent heat flux (W m^{-2}); (e) water equivalent of snow/ice depth (mm) and runoff (mm day^{-1}); and, (f) available water in the top 1-meter of soil — observations vis-a-vis simulations. {Note: In panel (a) snow temperatures are averaged only over snow-covered land}

also shows remarkably similar ground temperatures for the new and old summer (JAS) period models, suggesting that whatever differences in temperature may be at individual locations, they largely average out while the ground follows the air temperature closely.

4.2 Regional average snow cover and snow-melt

One of the major evaluation datasets is the satellite derived fractional area of snow-cover, which is drawn from satellite datasets generated by Matson *et al.* (1986). If this is true, snow models simulate snow accumulation much too soon (not shown). However, this cannot be true unless the observations of precipitation and/or 4DDA scaled snowfall were systematically erroneous. A better match with observations is achieved by recognizing that satellite is

unable to detect snow/ice below 2.0 cm (Fig. 3b). Using such an assumption, the snow accumulation of both models compares well with the satellite data provided 10 kg m^{-2} of snow (roughly 20 mm physical snow depth, and 500 kg m^{-3} of snow density) is taken as the threshold snowdepth. However, varying the threshold for the spring-season snow cover does not make much difference in the inferred snow cover. The influence of important details such as satellite data retrievals, which counts snowy-pixels only if more than 50 % of the grid has snow cover, can not be ignored, because snow drifting can pile snow unevenly and cause in homogeneous snow-melt. Indeed, snow-melt proceeds in a nonhomogeneous way and counting no-snow for less than 50 % snow cover will under-represent fractional snow cover. This is

Fig. 3. (Continued)

likely to yield an early snow melt in the observations as compared to a model in which snow is constrained to melt uniformly, and disappears suddenly. The same snow retrieval algorithm had no effect on snow accumulation. Therefore, we cannot attribute the entire error to the retrieving algorithm. The remaining discrepancy in snow melt timing raises some doubts about the accuracy of surface temperatures in the ISLSCP Initiative-I dataset. In other words, if the forcing data provided air temperatures that are too cold, the land model would interpret rainfall as snowfall, as well as require more energy to melt it. A similar delay in the snow-melt timing was noted by Douville (1998). However, the LSM-model used in data assimilation producing ISLSCP data also employs another LSM; therefore, the problem ultimately reduces to an LSM deficiency. Above all, there are some crude assumptions in the snow model too, particularly a lack of subgrid scale snow cover variability. Therefore, additional snowphysics improvements are envisioned to resolve the remaining errors.

4.3 Sensible and latent heat fluxes

Figures 3c and 3d shows the sensible and latent heat fluxes for RWB. In winter, the sensible heat flux is mostly into the snow. The new snow model shows much less evaporation as compared to the old model, a deficiency of SSiB also noted by Schlosser *et al.* (1997). Indeed, its timeseries is much smoother in the early part of winter when the heat capacity of the snow cover is comparatively smaller than the snow-plus-ground heat capacity of the old model. In the spring season, the new snowmodel yields much larger surface fluxes of heat and moisture, because its snow cover is warmer and its snow-melt occurs sooner. However, its snow temperature is not allowed to exceed freezing (a persistent feature of the old model, as a consequence of the implicit backward scheme for time-integration of ground temperatures). This correction can be expected to yield lower outgoing sensible heat flux, evaporation, and surface-OLR during the melting phase (as was noted in the single grid point diagnostics, but is not evident in the RWB-area averages). A warm episode

Fig. 3. (Continued)

in the spring of 1987 produced large downward sensible fluxes, but the new model with a snowpack of smaller heat capacity adjusts to the anomalous weather much more quickly as compared to the old model. However, the warm spell was also a moist spell; hence, the evaporation reduction. In the snow-free periods of July, August, and September (JAS) for which the LSM-physics of the new and old models is identical, the sensible (latent) heat fluxes are smaller (larger) for the new model as compared to the old, even though the ground temperatures were very similar as can be seen in Fig. 3a. This is the result of larger soil moisture of the new snow-model.

4.4 Snowmelt, soilmoisture, and runoff

There are huge differences seen between snow melt timing and soil-water recharge. The corresponding snow depth and lateral runoff time-series also show large differences (Fig. 3e). In the new model, snow accumulation in the fall is somewhat larger, but it melts much earlier and is in better agreement with Douville *et al.* (1995b). The new model snow-melt infiltrates into the soil far more efficiently. If the soil

is frozen, the model causes refreezing of snow-melt as opposed to maintaining runoff at freezing temperatures over the sub-freezing ground as in the old scheme. Timing estimates of these differences are available in Table 2. From the river discharge diagnostic analysis of Oki *et al.* (1997), we know that the old model produced excessive runoff for the spring snow-melt period. The new model corrects this deficiency. The soil moisture increases after snow-melt (about July 1) too small in the old model, while the soil water recharge is much more realistic in the new model (Fig. 3f). Another major difference appears in the mid-winter runoffs of the two models. The old model produced none, while the new model produces a more realistic Volga River runoff. Similar beneficial effects were reaffirmed (not shown) in the river-routing model of Oki and Sud (1998). In the new model, the snowfall events of the early fall season are followed by a succession of quick snow-melt events. Alternatively, the old model which assumed the ground and snow to be a unified layer did not warm and cool so quickly, and showed a smooth

Table 2. Comparison of Old and New Models. Data is listed as 1987 value/1988 value.

Data	Old	New	Observations
Snowmelt Date (days)	May 23/May 29 143/149	May 10/May 17 130/137	May 3/April 20 123/110
Available Water (mm)			
May 1	109/108	157/144	236/218
June 1	95/87	187/186	206/151
July 1	92/69	158/148	165/140
August 1	96/56	149/117	167/114
End of Melt	97/91	230/231	
Average Ground Temp °C			
Jan 1 – Snowmelt Date	–10.3/–9.3	–11.0/–9.2	
Summer Evaporation W/m ²	49.2/53.9	52.7/60.3	
Summer Sensible Heat W/m ²	22.0/23.2	18.4/17.7	
Ground Thaw Date (days)	May 22/May 28 142/148	April 30/May 3 120/123	

growth of its snowmass. In this way, the RWB-averaged snow depth timeseries represents the integrated influence of more realistic snow processes of the new model, versus comparatively simplistic snowphysics of the old model.

4.5 Spring snow-melt, and thawing of soil

Even though separate maps were examined and analyzed for 1987 and 1988, we use averages for both years. A computer animation of snow cover progression revealed a clear south-west progression of the increase of snow cover during the fall season and north-east progression of decrease in snow cover during spring season snow-melt (not shown). At any given time, the snow cover in the spring season was much reduced for the simulation with the new model. We show the days advanced to snow-melt (old minus new) in Fig. 4a. The gain is anywhere from 10–20 days. At no place does the snow-melt in the new model occur later than the old model. The time gained is larger for forested regions (inferred by higher NDVI in January), and least in the middle of the RWB or fields, where wheat is grown. Higher heat capacity of soil in forested regions delayed snow-melt as discussed earlier. The local topography, or nearness of oceans, did not matter (not shown) for the gain in days-gained-to-snow-melt. The number of days gained in achieving earlier thawing of soil, rough by increased by 15–20 days (Fig. 4b). This represents an advance of five days within the snow-melt period, and is primarily responsible for higher snow-melt infiltration.

4.6 New snow-model vis-a-vis observations

The number of weeks lapsed to completion of snow-melt starting from April 1 is shown in Fig. 5a. It represents an average for both years, 1987 and

1988. Clearly simulated snow-melt is delayed by 2–4 weeks everywhere, as compared to the satellite data (Fig. 5b). A four-week delay may appear truly huge, but an examination of its distribution reveals it may not entirely be a model deficiency. In the southern regions, where the land is cleared for wheat cultivation (as can be inferred from high leaf-area indices for January shown in the background), a two-week delay includes model errors plus inability of microwave-sensing to label less than 50 % snow cover as snow-free regions. Since snow melting is a process with positive feedback, it causes patchiness. The digitized data may infer no-snow for the region somewhat early. However, the largest errors occur in the forested regions. These regions contain evergreen forests and maintain healthy LAI even in January (background LAI maps). It is well known that the satellite algorithms have problems in these regions, because snow often remains on the ground while the canopies are cleared within a few days after snowfall. The snow below the forest canopies is naturally patchy, as a result of heat generated by the metabolizing or thermally radiating biomass (neglected in SSiB). Thus, a large difference in simulated and observed snow-melt may largely be caused by the above effects in forests.

Figure 5c shows that the available soil moisture has errors of 20–50-mm mon⁻¹ in a total of roughly 160–200-mm mon⁻¹. This is a 10–20 % error in a field which can easily reduce by 30 % merely by gravitational drainage in a matter of 3–5 rain-free days. Indeed, the *in situ* soil moisture data at the observing stations is very reliable, but it is somewhat sparse because the entire region is not covered at all times. We note a large variability in the observations, from one place to another, which must

Fig. 4. Two year averaged Old minus New snow-model differences for : (a) "days-to-complete snow-melt"; (b) "days-to-thawing of the first soil-layer." The background shading shows NDVI for January (Los *et al.*, 1994; also ISLSCP Initiative-I data). NDVI's larger than 0.1 (shaded regions) are assumed to be evergreen, with NDVI-definition on the bar-scale below the figures. Black boxes represent water points. Horizontal resolution is 1 deg. $\times$ 1 deg.

be real. But, the model is too simple as well as too coarse to resolve subgrid scale variability. For these reasons, we submit that the new snow-melt and its runoff-infiltration is performing satisfactorily despite several deficiencies. Indeed, for the region as a whole, we again note the same kind of inaccuracies (Fig. 3f), but it is much better than the old model simulation, (Fig. 1a) which is also typical of other versions of SSiB.

5. Summary and conclusion

Modeling snowpack essentially requires synthesis of existing knowledge of the behavior of snow physics, and soil-hydrology. As variously suggested, presumably the biggest inaccuracy emerges from sub-grid scale inhomogeneity and its influence on the snowpack, snow-melt, and runoff. Since in situ observations are sparse, it is not clear what level of sophistication and accuracy would be sufficient for a GCM. The authors believe that if an internally consistent physical model, but without ad hoc assumptions, can produce a realistic simulation, it may be adequate for GCMs.

There is a long treatment history of snowpack, and ground as a unified layer with one temperature.

However, the GSWP exercise has clearly shown that such an assumption causes major deficiencies in snow-melt timing and infiltration. Thus, it became abundantly clear that such a model is inadequate. A few models employ separate snow-cover physics, and produce better simulations of snowfluxes (see, Douville *et al.*, 1995b). We have significantly improved the snowphysics of SSiB by including a separate snow-layer. The new model simulates a far better snow-melt timing and infiltration. This, in turn, has led to a much higher soil moisture in the summer season causing lesser (larger) sensible (latent) heat fluxes. The new model also produces some runoff throughout the year, thereby providing water flow, while its early spring runoff is much more realistic. There are consistent changes in the winter and spring temperatures and surface fluxes, which largely mitigate the deficiencies of the old model. Based on the above findings, we conclude that the new snow model is more realistic than the old. Dirmeyer's (1999) results suggest that such an improvement would help to simulate more realistic climate with GCMs; we evaluate this in Part II. We also infer that a better prescription of the soil properties, as proposed by Xue *et al.* (1997), may not be

Fig. 5. Two year averaged plot of (a) number of weeks lapsed to completion of snow-melt after April 1 for the new snow-model; and, (b) new model minus satellite retrieval error in weeks for "completion of snow-melt." (c) Available soil moisture differences: simulated minus in situ data for the top one-meter of the soil. The background shading are the same as in Figs. 4a and 4b.

adequate to mitigate snow-model deficiencies of old SSiB.

Despite the need for avoiding ad hoc assumptions, a few such assumptions were necessary in the new snow model. For example, we do not calculate the age of the snowpack nor it is clear as what density should be used for approximating the snow/ice mixture. Since the conductivity of snowpack can be 100 to 1000 times less than that of ice (Gray and Male, 1981), we have used the average of the snow and ground temperature as surrogates for the snow/ice mixture. The need to improve the current snow model is suggested by our diagnostic analysis. There is also a long history of the force-restore calculation of ground temperature (Bhumralkar, 1975). One needs to investigate the advantage of multilayer soil models in which frost-depth is predicted so that layers below the frost-depth can produce runoff.

We also submit that the remaining timedelay of

snow-melt could be caused by one or more of the following: 1) an assumption of grid-scale homogeneity of the snow pack; 2) deficiencies in the simulated diurnal cycle of snow-surface temperatures emanating from lack of force-restore calculation of surface temperature of the snowpack; 3) use of a simple solar extinction function within the snowpack; 4) neglect of interactive surface albedo and conductivity of the snowpack as a function of snow density, which is affected by the age of snow, and 5) low bulk transfer coefficients for heat and moisture for stable conditions as shown in studies by Jones *et al.* (1994). Further studies are needed to address these issues. Another important consideration would be the influence of in-flight sublimation of blowing snow particles. Such a sublimation will draw the latent heat from the air, without affecting the surface energy budget of the snowpack.

Appendix

Equations for land-atmosphere interaction with snow-cover

The following equations of the biosphere-atmosphere scheme of SSiB are solved simultaneously for tendencies of ground temperature, T_g , canopy temperature, T_c , PBL temperature, T_{PBL} , and specific humidity, q_{PBL} . Some additional symbols invoked are: R_n for net radiation, H and E for sensible and latent fluxes; C for heat capacity, t for time, SW for shortwave and LW for longwave fluxes, V is for the vegetation cover fraction. Subscripts g , c , and a are used for ground, canopy, and near-surface ambient air; flux out of the BLT is for the boundary-layer top. The four temperature tendency equations in the original model were:

$$C_g \frac{\partial T_g}{\partial t} = Rn_g - H_g - E_g - \frac{2\pi C_g}{\tau} (T_g - T_d), \quad (A.1)$$

$$C_c \frac{\partial T_c}{\partial t} = Rn_c - H_c - E_c, \quad (A.2)$$

$$\Delta m_{SL} c_p \frac{\partial \theta_{PBL}}{\partial t} = H_g + H_c - H_{BLT}, \quad (A.3)$$

$$\Delta m_{SL} L \frac{\partial q_{PBL}}{\partial t} = E_g + E_c - E_{BLT}. \quad (A.4)$$

Due to partial vegetation cover, net radiation is a weighted average of absorption between the canopied and clear areas; shortwave albedo is α , longwave emissivity of ε , and $\tau = 1$ day. The equations determining net radiation absorbed by the canopy and ground are:

$$Rn_c = SW_c \downarrow (1 - \alpha) + LW \downarrow [V_c(1 - \varepsilon_c)] - 2\sigma T_c^4 V_c(1 - \varepsilon_c) + \sigma T_g^4 V_c(1 - \varepsilon_c), \quad (A.5)$$

$$Rn_g = SW_g \downarrow (1 - \alpha) + LW \downarrow [1 - V_c(1 - \varepsilon_c)] + \sigma T_c^4 V_c(1 - \varepsilon_c) - \sigma T_g^4. \quad (A.6)$$

The above equations for snow-free land remain the same, but for the new snowphysics they are revised to yield the following modified equations. The new tendency equations for snow temperature, T_s (instead of T_g) and net surface radiation are as follows:

$$C_s \frac{\partial T_s}{\partial t} = Rn_s - H_s - E_s + \kappa_c(T_g - T_s). \quad (A.7)$$

while the net radiation in (A.7) is modified to:

$$Rn_s = SW_{sn} \downarrow (1 - \alpha) [1 - \exp(-\kappa_t Z_{sn})] + LW_{sn} \downarrow [1 - V_c(1 - \varepsilon_c)] + \sigma T_c^4 V_c(1 - \varepsilon_c) - \sigma T_s^4. \quad (A.8)$$

The new ground temperature tendency equation under the snow-cover becomes:

$$C_g \frac{\partial T_g}{\partial t} = SW_g - \kappa_c(T_g - T_s) - \frac{2\pi C_g}{\tau} (T_g - T_d), \quad (A.9)$$

where

$$SW_g = SW_{sn} \downarrow (1 - \alpha) \exp(-\kappa_t Z_{sn}) \quad (A.10a)$$

$$\kappa_c = \left[\frac{0.5 Z_{sl}}{\kappa_{sl}} + \frac{1}{h_{air}} + \frac{Z_{sn}}{\kappa_{sn}} \right]^{-1}, \quad (A.10b)$$

and

$$h_{air} = 4\sigma \left(\frac{T_s + T_g}{2} \right)^3. \quad (A.10c)$$

The additional symbols are Z for depth, h_{air} for heat transport coefficient of the air gap, the solar extinction coefficient, κ_t , and snow and ground conductivities, κ_{sn} and κ_{sl} . (which give a composite conductivity, κ_c). Canopy interception of snowfall or rainfall is handled consistently in the previous model and remains the same.

The temperature increments for the physics time-step are calculated by integrating the equations simultaneously, using implicit backward integration. The algebra is tedious, but the approach is straightforward. We show the numerical form of one of the equations, (A.2), to explain the procedure. Its implicit backward form is:

$$\begin{aligned} C_c \left(\frac{T_c^{n+1} - T_c^n}{\Delta t} \right) &= Rn_c + \frac{\partial Rn_c}{\partial T_c} (T_c^{n+1} - T_c^n) \\ &+ \frac{\partial Rn_c}{\partial T_g} (T_g^{n+1} - T_g^n) + \frac{\partial Rn_c}{\partial T_a} (T_{PBL}^{n+1} - T_{PBL}^n) \\ &+ \frac{\partial Rn_c}{\partial q_a} (q_{PBL}^{n+1} - q_{PBL}^n) \\ &- \left[H_c + \frac{\partial H_c}{\partial T_c} (T_c^{n+1} - T_c^n) + \frac{\partial H_c}{\partial T_g} (T_g^{n+1} - T_g^n) \right. \\ &+ \left. \frac{\partial H_c}{\partial T_a} (T_{PBL}^{n+1} - T_{PBL}^n) + \frac{\partial H_c}{\partial q_a} (q_{PBL}^{n+1} - q_{PBL}^n) \right] \\ &- \left[E_c + \frac{\partial E_c}{\partial T_c} (T_c^{n+1} - T_c^n) + \frac{\partial E_c}{\partial T_g} (T_g^{n+1} - T_g^n) \right. \\ &+ \left. \frac{\partial E_c}{\partial T_a} (T_{PBL}^{n+1} - T_{PBL}^n) + \frac{\partial E_c}{\partial q_a} (q_{PBL}^{n+1} - q_{PBL}^n) \right]. \end{aligned} \quad (A.11)$$

Equation (A.11) can be recast to yield an equation of the type: $f(T_g^{n+1}, T_c^{n+1}, T_{PBL}^{n+1}, q_{PBL}^{n+1}) = \text{constant}$. When each of the equations (A.1) to (A.4) have been recast in the implicit backward form, we can express them in the form of a 4×4 matrix, plus a column matrix for the RHS. The equations are solved simultaneously by inverting the matrix of coefficients of $T_g^{n+1}, T_c^{n+1}, T_{PBL}^{n+1}, q_{PBL}^{n+1}$. In fact, even the simple looking partial derivatives can get quite involved, e.g., let us determine:

$$\frac{\partial H_c}{\partial T_g} = \frac{\partial}{\partial T_g} \left[\frac{\rho c_p}{r_c} (T_c - T_a) \right], \quad (A.12)$$

It requires $\frac{\partial T_a}{\partial T_g}$, which is produced by first equating the sum of fluxes at "a" to zero:

$$H_g + H_c + H_a = \frac{\rho c_p}{r_g} (T_g - T_a) + \frac{\rho c_p}{r_c} (T_c - T_a)$$

$$+\frac{\rho c_p}{r_a}(T_{PBL} - T_a) \equiv 0. \quad (\text{A.13})$$

Equation (A.13) is used to get all the partial derivatives $\frac{\partial T_a}{\partial T_g}$, $\frac{\partial T_a}{\partial T_c}$, $\frac{\partial T_a}{\partial T_{PBL}}$. Putting $\frac{\partial T_a}{\partial T_g}$ into (A.12), we get:

$$\frac{\partial H_c}{\partial T_g} = \rho c_p [r_g^{-1} + r_c^{-1} + r_a^{-1}] [r_c r_g]^{-1} \quad (\text{A.14})$$

The above serves as an example of the complexity of partial derivatives required for all the terms of all the equations that are cast like (A.11). For the snow model, we use the same methodology, but we invoke the constraint that if T_{sn} is 0°C , and the energy fluxes at t^n are positive representing warming of the snow cover, we set all $\partial(\)/\partial T_s$ to zero which disallows the influence of snow temperature rising above 0°C , a physical unrealizable state, on the surface fluxes. This correction reduces all the other fluxes, while invigorating snow-melt particularly in the spring season. Moreover, T_s and T_{sn} are distinguished only for heat conduction through the snow pack. As described in the body of the paper, there is a possibility of snow melt from the bottom, which again follows the constraint that $T_{sn} \leq 0^\circ\text{C}$.

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SSiB を改良した新雪モデル

I: 設計と ISLSCP Initiative I データによる評価

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ISLSCP Initiative-I データを用いた全球土壌水分プロジェクトでの評価で、SSiB モデルによるロシア小麦地帯 (RWB) での融雪が観測と比べて非常に遅れ融雪水の浸透が極端に少ないことが示された。さらに、融雪水の多くが土壌水分増加ではなく流出となった。この欠点は SSiB の雪モデルと土壌層のモデル化の不十分のためであった。本研究では独立の雪層を考慮した新雪モデルを採用している。雪は入射太陽フラックスを吸収・射出し冬と融雪期を通じて雪温・地温に影響する。ISLSCP Initiative-I データによる評価で、新雪モデルは RWB 域での融雪が 2~3 週間早くなり、融雪期の初期に土壌が融け、より多くの融雪水が土壌に浸透する。このように新モデルは土壌水分やボルガ河流出をより現実的に再現する。融雪の遅れ (1~4 週間) の理由として、(1) 密な森林での衛星による雪観測の不正確さ、(2) モデリングの仮定、例えば雪の年齢の影響を無視していることや雪による太陽放射吸収の単純化のために雪面温度と平均気温の区別が不適切になること、(3) ISLSCP 気温データの低温バイアスの可能性、が考えられる。