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VALIDATION OF A SNOW-FROZEN GROUND PARAMETERIZATION OF THE ETA MODEL

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1. INTRODUCTION

Cold season processes significantly influence water and heat fluxes between the atmospheric boundary layer and the land surface [Barnett et al, 1989; Namias, 1985]. Frozen ground and snow cover also influence rainfall-runoff partitioning and, therefore, the amount of soil moisture that subsequently is available for evapotranspiration in spring and summer. Nearly impermeable soil layers can be developed under some critical heat-moisture conditions during winter and spring seasons. Because of the high sensitivity of the climate system to winter as well as summer land surface forcing, there has been strong recent interest in upgrading the cold as well as warm season land surface parameterizations of atmospheric models [Verseghy, 1991; Marshall et al, 1994; Lynch-Stieglitz, 1995].

The Eta model employs the NWS National Center for Environmental Prediction (NCEP) and Office of Hydrology (OH) extensions to the multilayer OSU (Oregon State University) soil/vegetation scheme [Ek and Mahrt, 1991]. Although the model includes a crude snow pack treatment, it does not account for the effects of frozen ground, patchy snow cover, or temporal/spatial variability in snow properties. This study presents extensions of the Eta model Land Surface Subsystem (Eta LSS) that include the latter effects. These extensions were developed so that the added physical complexity and soil profile treatment is compatible with general complexity and configuration of the present model. Accordingly, a physically based parameterization of the frozen ground, and a more realistic snow accumulation-ablation scheme were introduced. Off-line tests of the parameterization were performed using experimental data from the Rosemount site in Minnesota, and PILPS2d forcing data (Valdai, Russia).

2. LAND SURFACE PARAMETERIZATION

The existing Eta model land-surface parameterization couples the Penman potential evaporation approach, the layer-integrated soil model, the canopy resistance model, and the surface runoff component from the SWB model [Schaake et al., 1996]. The surface energy and water budgets are computed for a single unified ground-vegetation surface. Soil moisture and heat fluxes are simulated separately at each time step assuming no significant heat transfer during redistribution of liquid water.

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Ground heat and water fluxes are controlled by the diffusion equations for soil temperature, T, and volumetric soil moisture content, θ, [Chen et al., 1996]. Soil thermal and hydraulic parameters are formulated as functions of total soil moisture content. The original parameterization does not account for the latent heat of soil moisture phase transitions, assuming instead that water is liquid at any temperature.

The implicit Crank-Nicholson scheme is applied to the layer-integrated form of diffusion equations to simulate a layer-average soil temperature and soil moisture assuming to be located in the middle of each soil layer. Although 2-4 layers that extent at least over the root zone were used in different applications, more layers could be easily accommodated. Skin temperature, evaporation, and infiltration are used as the upper boundary conditions. The skin temperature is determined from a linearized ground-vegetation/snow surface energy balance equation. Evaporation from the ground-vegetation surface is the sum of the direct evaporation from the top shallow soil layer, evaporation of precipitation intercepted by the canopy, and transpiration via the canopy and roots. The direct evaporation from the ground surface is driven by the water flux to the soil surface. The transpiration is defined as a ratio of the potential evaporation depending on the canopy resistance. The green vegetation fraction acts as a weighting factor between the three components. The total evaporation and its components are bounded by the potential evaporation from a Penman-based energy balance approach. It assumed that the total transpiration via roots is partitioned between root zone layers according to the layer weights. The SWB formulation is used to partition rainfall/snowmelt into infiltration and surface runoff. The mean annual air temperature and gravitational percolation are the bottom boundary conditions. More detailed description of the Eta hydrology can be found in Chen et al. [1996].

2.1 *Snow accumulation-ablation scheme extension*

Snow accumulation and ablation scheme of the original Eta model is based on the energy and mass balance of the snowpack. The parameterization neglects heat transferred by water movement and assumes that all liquid water immediately reaches the soil surface. It assumes also that snow surface temperature is at the freezing point during snowmelt. Fractional snow covered area is accounted for by a linear function depending on snow water equivalent. Snowpack physical characteristics, thermal conductivity and density, are assumed constant during the cold season. This can lead to significant overestimation of snow depth and as a result can cause biases in snow-soil surface heat exchange.

Snow compaction and a more flexible distribution function of snow cover are introduced to overcome the original model restrictions. An approximate solution of the Anderson-Kojima's model of snow compaction [Anderson, 1976] was derived to calculate an average snow density of snowpack, ρ_s, during cold season:

$$\rho_{s,t+\Delta t} = \rho_{s,t} \frac{e^{BW_{s,t}} - 1}{BW_{s,t}} \quad (1)$$

where

$$B = \Delta t C_1 e^{0.08 T_s - C_2 \rho_{s,t}} \quad (2)$$

where T_s is a snow temperature, Δt is the time step, and C_1 and C_2 are parameters. Values of 0.01 $cm^{-1}\ hr^{-1}$ and 21 $cm^{-3}g^{-1}$ are assigned based on Anderson [1976] results. Snow density is also adjusted during new snowfall and snowmelt. Anderson's [1976] formulation is used in estimating snow liquid-water-holding capacity depending on snow density. The holding capacity is bounded by 0.03 and 0.10. The minimum bound value is used when snow density exceeds 0.4 g cm^{-3}. Freezing of liquid snow water, W_{frz}, is calculated based on snow surface temperature assuming that it occurs at the snow surface if snow density is above 0.2 g cm^{-3}, W_{frz}=0.215$\sqrt{(\Delta t\ T_s)}$ mm where Δt in hours, and T_s in degree Celsius.

A linear distribution function of snow is replaced by an exponential function with an upper bound limit. To account for bare soil patches during snowmelt, skin temperature is weighted between the melting point temperature and an estimated skin temperature for free surface depending on the snow cover fraction.

2.2 *Frozen ground parameterization*

To account for the latent heat of soil moisture phase transitions, the heat flux equation in the original Eta parameterization was replaced by a diffusion equation with the source/sink term. The volumetric heat capacity, the thermal conductivity, and the hydraulic conductivity of the soil are now functions of total volumetric soil moisture content θ and volumetric ice content θ_{ice}.

The same layer-integrated form of the diffusion equation is also used to apply the implicit Crank-Nicholson scheme. An explicit approximation is applied to the source/sink term. To reduce numerical error during fast soil freezing/thawing, two iterations are used. This is similar to how water flux is simulated during rainfall events [Chen et al., 1996]. The ice content at each soil layer is estimated as a function of soil temperature and total soil moisture content. It is assumed that when ice is present, soil water potential remains in equilibrium with the vapor pressure over pure ice. A simple relationship can be drawn between the freezing point of soil water and soil water potential, ψ, after neglecting of soil water osmotic potential:

$$\psi = \frac{LT}{g(T + 273.16)} \quad (3)$$

where soil temperature T is in degrees Celsius, L is the latent heat of fusion, and g is the acceleration of gravity. Combination of Campbell's relationship between water potential and water content, modified on frozen soil effects, and (3) leads to

$$\frac{g\psi_s}{L}(1 + c_k\theta_{ice})^2 \left(\frac{\theta - \theta_{ice}}{\theta_s}\right)^{-b} - \frac{T}{T + 273.16} = 0 \quad (4)$$

where ψ_s is the air entry potential, and parameter c_k accounts for the effect of increasing of specific surface between soil minerals and ice-liquid water during freezing.

Equation (4) indicates that the ice content (or unfrozen soil water) is a function of both the soil temperature and the soil moisture content. This agrees with laboratory/field experiments. Because actual amount of water converted into ice or vise versa depends on an available incoming heat flux, the potential increase/decrease of ice estimated from (4) is bounded by an amount that can be converted using available heat flux. This approach does not account explicitly for the freezing front propagation because the integrated form of the diffusion equation is used. Phase transitions can occur at each layer at the same time depending on distribution of soil temperature and heat fluxes.

Surface runoff is adjusted depending on frozen ground conditions. Practically impermeable layer can be formed under some critical conditions. Field experiments suggest that the spatial extent of an impermeable layer depends on the areal average ice content of the frozen soil. Koren [1991] expression is used in this study to estimate a fraction of an area on which an impermeable layer is formed. It is assumed that this area produces direct surface runoff, and the rest of an area produces surface runoff depending on the rainfall-runoff partitioning mechanism of the original Eta parameterization.

3. VALIDATION RESULTS

Two experimental data sets were used in this study: (a) From the Rosemount site of the University of Minnesota Agricultural Experiment Station where field measurements were performed during a cold season; (b) Long-term measurements of water balance components from the water balance station Valdai (Russia). The first data set is a profile type measurements of soil temperature and unfrozen soil moisture content at 8 depths from 2.5 cm to 1 m. The time interval of measured raw data varies from 10 min to 1 hour. All forcing data ((incoming and reflected solar radiation, incoming and outgoing longwave radiation, wind speed, air temperature, precipitation, air pressure, and specific humidity at the surface) were also available. The second data set represents spatial averages over a small river basin Usadievskiy (watershed 0.36 km^2). Measurements of total soil moisture were taken at the end of each month using gravimetric technique for every 10-cm layer to a depth of 1 m. Snow water equivalent measurements were also available in non-regular time intervals during snow accumulation and ablation periods. 18 years forcing data from the PILPS2d project were used in the study. A four-layer version and a half-hourly time step was used in simulations for both data sets.

3.1 *Soil temperature and soil moisture results at the Rosemount site*

10 cm, 20 cm, and 60 cm layers were used to represent a root zone of Waukegan silt loam profile at the Rosemount site, and 70 cm layer was assigned for the bottom soil storage. Model parameters that are responsible for the water redistribution were estimated based on measured soil properties at the site. Vegetation-related parameter values were adopted from Chen et al. [1996]. Monthly values of the green leaf area index (LAI) and the green vegetation fraction were derived from Eta databases for the Rosemount site location. The soil temperature at 3 m, the bottom boundary condition, is defined as the long-term annually averaged air temperature, which is 7.5°C. Initial conditions at four soil layers were estimated from soil moisture contents and temperatures measured at the beginning of October 1995. Simulations were performed continuously to the end of the cold season.

Simulated soil temperatures and unfrozen soil moisture contents at the three top layers were compared to measured data. Overall, the frozen ground version simulations are reasonably close to measured data for both temperature and soil moisture. Simulated temperatures from the original Eta model are generally much too cold. Frozen ground and snow component adjustments could affect on these differences. To estimate a frozen ground contribution only, a version of the new parameterization that includes only the snow component without frozen ground was run on the same data set. Results from this version were closer to the snow-frozen ground version results, but differences between them could be as much as 5 degrees in the two deeper layers. The results confirm the hypothesis that the latent heat of the ice fusion significantly affects soil temperature variability. This can be seen clearly in Figure 1 where hourly soil temperatures during multiple transitions from freezing to thawing of the top layer are shown. The frozen ground version produces reasonably stable soil temperature during this period as are actually observed. Diurnal variation of soil temperatures from the original Eta parameterization follows very closely to the air temperature variation, even when the soil is freezing. This is not consistent with the observed behavior.

Soil moisture simulation results are shown in Figure 2. The original Eta parameterization and the new version without frozen ground have not frozen water component, and only total (unfrozen) soil moisture contents are plotted in Figure 2. As expected, these two versions produced very close results at all layers. The frozen ground version results reasonably agree with measured liquid water contents. The hydrological importance of the frozen ground can be seen from

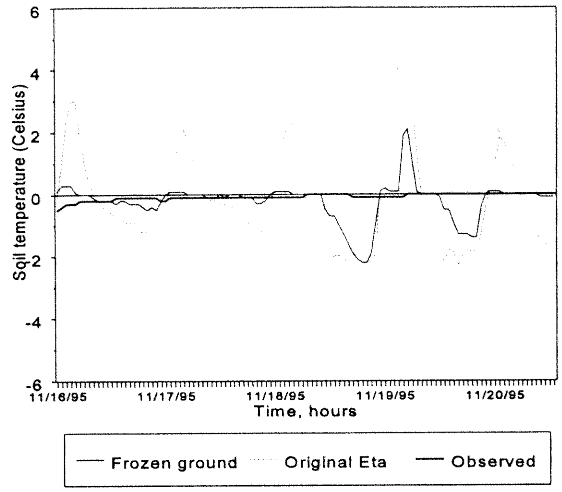

Figure 1. Observed and simulated hourly soil temperatures at the 10 cm layer, Rosemount site.

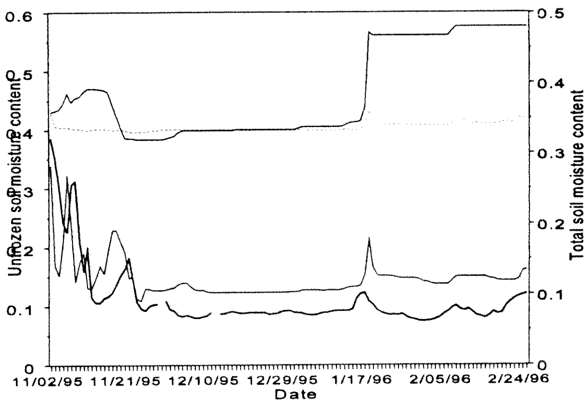

Figure 2. Observed (thick line) and simulated (thin line) unfrozen water, and total soil moisture (top panel) simulated by the frozen (solid line) and Eta (dashed line) versions at the top 10 cm layer; Rosemount site.

comparison of the total soil moisture contents before the snowmelt simulated by the frozen ground version and the original Eta model version. Soil water storages from the original Eta version are lower then the frozen ground version results for all soil layers. Differences for the two top layers can be as much as 20-25 percent. It means there will be less available water for evaporation and less runoff during spring time.

3.2 *Snow component results for the Valdai basin*

The Usadievskiy basin (81% grassland with shallow root zone) soil profile was modeled by 10 cm, 30 cm and 60 cm root zone layers, and 1 m layer of the bottom soil storage. 70% roots were allocated at the first layer. Most of model parameters were adopted from the PILPS2d definition. Snow albedo was 0.75, and it was reduced during snowmelt depending on snow coverage to its minimum 0.23 for the bare soil. To initialize states, the model was run first in spin-up mode, and then it was run continuously for the entire 18 years period.

Simulated and observed snow water equivalents are close for the most of winter-spring periods (see e.g. Figure 3). Simulation results suggest that it is very important to account for the retention and freezing of liquid water during snowmelt. The original Eta version (without retention and freezing of snow liquid water) simulates significantly shorter ablation period with much faster reduction in snow water equivalent at the beginning of snowmelt. Figure 4 illustrates the diurnal cycle of liquid water retention and freezing, and its effect on snow water equivalent.

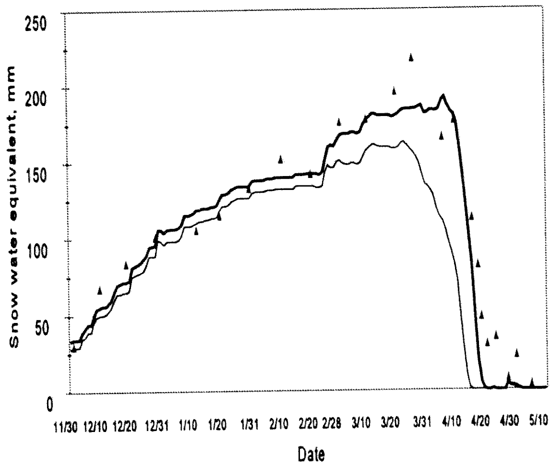

Figure 3. Observed (triangles) and simulated from the new (thick line) and Eta (thin line) versions; Valdai, 1975-1976 cold season.

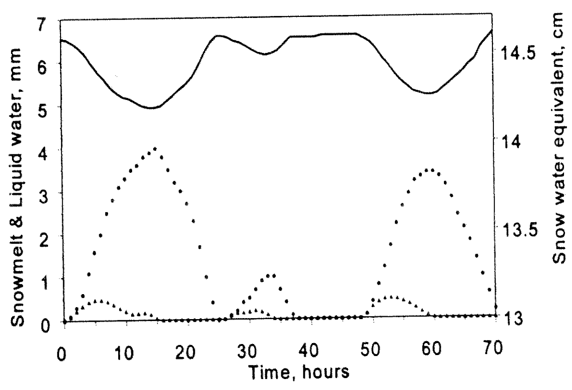

Figure 4. Snowmelt rate (triangles), liquid water content in snow (circles), and snow water equivalent (top panel) simulated during few retention-freezing cycles, Valdai.

4. CONCLUSIONS

The frozen ground version over performed the original Eta parameterization as compared to measured data on both soil temperature and soil moisture. Neglecting of soil freezing/thawing leads to under prediction of soil temperatures and reduction of total soil moisture content at the end of cold season. Processes of retention and freezing of liquid water in snowpack can affect significantly on snow ablation. Snow ablation period is shorter and snow water equivalent reduction is faster if these processes are neglected.

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