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Variable infiltration capacity cold land process model updates

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Abstract

The Variable Infiltration Capacity (VIC) macroscale hydrologic model is distinguished from other Soil–Vegetation–Atmosphere Transfer schemes (SVATS) by its focus on runoff processes. These are represented via the variable infiltration curve, a parameterization of the effects of subgrid variability in soil moisture holding capacity, from which the model takes its name, and a representation of nonlinear baseflow. Recent upgrades to the model have improved its representation of cold land processes, and the effects of surface storage in lakes and wetlands. Specific improvements described in this paper include the following: (1) explicit representation of the canopy energy balance separate from the land surface when snow is intercepted in the canopy; (2) parameterization of the effects of spatial variability in soil freeze–thaw state and snow distribution on moisture and energy fluxes; and (3) effects of advection on snowmelt under conditions of partial snow cover. The effects of these model updates are demonstrated using data from the PILPS Phase 2(e) validation catchments within the Torne-Kalix River basin, Sweden.

Keywords: frozen soil; snow; surface energy balance; spatial variability; land surface scheme

1. Introduction

The variable infiltration capacity (VIC) model is a surface water and energy balance model designed for large-scale applications. It is applied to grid cells, typically with spatial dimensions from 1/8° to 2° latitude by longitude, connected so as to represent large continental river basins. In common with other soil–vegetation–atmosphere transfer schemes (SVATS), it balances water and energy at the land surface at sub-daily time steps. It is distinguished from other SVATS by its parameterization of the subgrid variability of soil moisture, and its effect on runoff generation, and by its nonlinear baseflow representation (see Fig. 1). Land cover within each grid cell is represented using a mosaic scheme. Within each vegetation class, spatial variability of infiltration and runoff generation is simulated using the variable infiltration curve (Liang et al., 1994) and baseflow is represented using the empirically based Arno baseflow curve (Liang et al., 1994).

Early simulations with the model were conducted using resolutions of one or more degrees latitude by longitude per grid cell, with two soil layers. Liang et al. (1996) determined that a thin top layer (5–15 cm) significantly improved evapotranspiration predictions in arid climates, and a third soil layer was added. The current version of the model is capable of using an arbitrary number of soil layers, although most applications (Bowling et al., 2003a; Maurer et al., 2001; Wood et al., 2002) use three layers.

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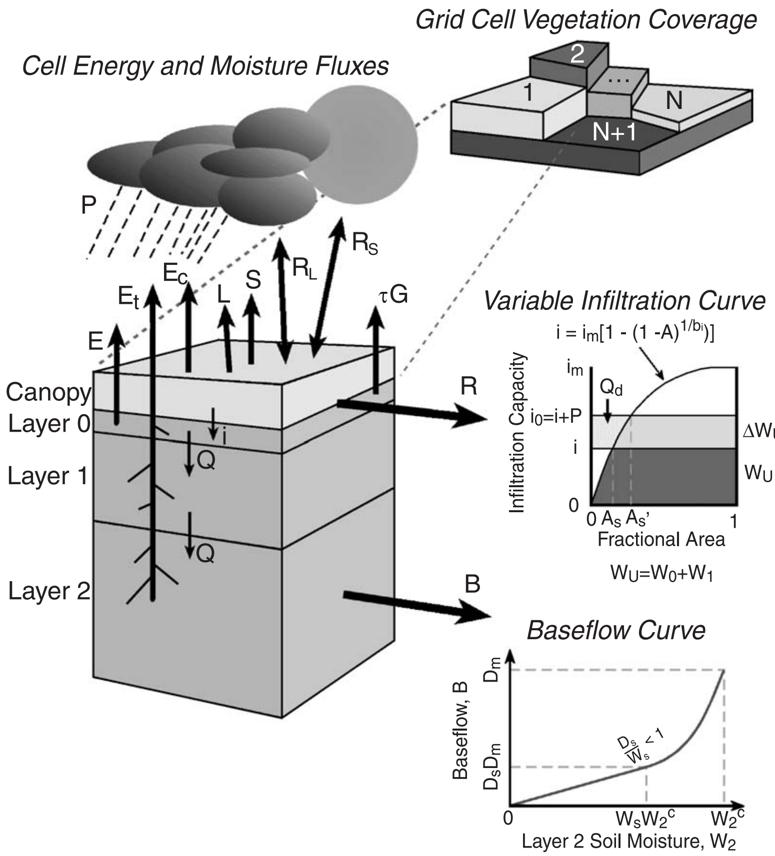

Fig. 1. Schematic of the variable infiltration capacity (VIC) macroscale hydrologic model with mosaic representation of vegetation coverage and three soil moisture layers.

Over the last 3 years, a number of changes have been made to improve the VIC model's representation of cold season processes, in conjunction with the Global Energy and Water Balance Experiment (GEWEX) Continental-scale International Project (GCIP) activities in the upper Mississippi River basin (Cherkauer and Lettenmaier, 1999). These involved the development of an algorithm to represent the effects of frozen soil, the addition of an algorithm to simulate the interception of snow by forest canopies, and general improvements to the snow accumulation and ablation algorithm. Detailed descriptions of these changes and a demonstration application to the upper Mississippi River basin are included in Cherkauer and Lettenmaier (1999).

This paper presents the most recent modifications to the VIC model that are not included in previous archival literature. In particular, all changes to the model as used in the PILPS Phase 2(e) experiment in the Torne-Kalix River basin, Sweden (Bowling et al., 2003a) not previously documented are described here. The changes include an improved representation of the canopy energy balance, parameterization of spatial variability in snow and frozen soil characteristics, modification of the frozen soil algorithm of Cherkauer and Lettenmaier (1999) to represent permafrost (in contrast to ephemerally frozen soil, as in the upper Mississippi river basin), and an algorithm that represents the effects of lakes and wetlands on surface moisture and energy fluxes. These updates, as well as sample applications to the Kaalasjärvi River, are detailed below. The Kaalasjärvi River is a headwater of the Torne-Kalix River basin and is within the PILPS Phase 2(e) domain.

2. Cold processes model updates

2.1. Canopy energy balance

Interception of snow by the vegetation canopy was added to improve the VIC model's representation of snow accumulation and ablation in forested regions (Cherkauer and Lettenmaier, 1999). The interception algorithm was developed by Storck and Lettenmaier (1999) for the high-resolution Distributed Hydrology–Soil–Vegetation Model. By design, this algorithm does not iterate for canopy temperature and therefore does not balance energy. Closing the energy balance was motivated by two concerns: (1) improving the simulation of surface energy fluxes by including the interactions between the under and overstories; and (2) ensuring that reported surface energy fluxes were accurate.

The VIC model surface energy balance, as developed by Liang et al. (1994, 1999, 2001), included the entire vegetation column in the control volume. That method is still used when there is no snow present or falling, and when there is no vegetation canopy. Otherwise, three control volumes are used (see Fig. 2): (1) the overstory; (2) the understory; and (3) the canopy air. Several other large scale land surface schemes (e.g. SiB (Sellers et al., 1996) and BATS (Dickinson et al., 1993) use similar schemes. To close the energy balance for the entire soil–vegetation–atmosphere column, VIC solves separate energy balances for each of the three control volumes, and then solves an overall balance:

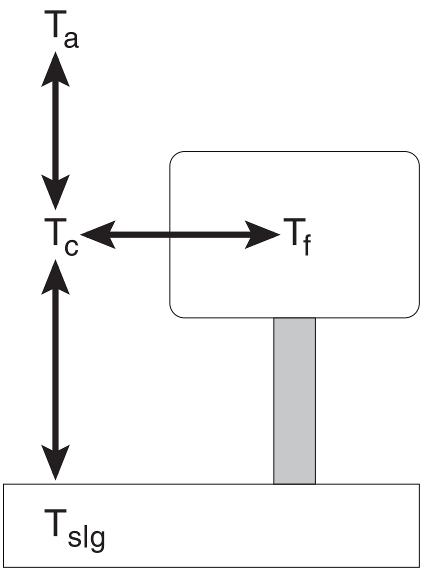

Fig. 2. Schematic of energy flux interactions. T_a is the observed air temperature. T_c, T_f, T_s and T_g are simulated temperatures for the canopy air, canopy foliage, ground snow surface and the soil surface, respectively. Arrows represent the interaction connections of latent and sensible heat fluxes.

- If snow is falling or is present in the canopy, the model iterates to find a foliage temperature (T_f).
- Downwelling shortwave radiation is reduced due to foliage interception before reaching the understory. Long-wave radiation from the sky is assumed to be completely absorbed by the foliage, so long-wave input to the understory is based on the foliage temperature. The ground snowpack surface (T_s) and ground surface (T_g) temperatures are solved iteratively (Cherkauer and Lettenmaier, 1999).
- The energy fluxes from the over and understory are used to compute the canopy air temperature (T_c), which controls sensible heat exchange between the under and overstory and the atmosphere (T_a).
- Energy fluxes between the canopy and the atmosphere are compared with those between the canopy air and the overstory and understory. If the fluxes do not balance, the previous steps are repeated after updating the exchanged fluxes for the new temperature estimates.

Energy fluxes predicted from this process balance for the canopy, the ground surface and for the entire soil–vegetation–atmosphere column. Because the iteration associated with closing the full canopy energy balance increases computational requirements significantly, the model can optionally be run without the final overall iteration process. Thus, fluxes, which should really interact, are estimated using the previous temperature solution. Simulations with and without the final iteration show little difference in the predicted water and energy fluxes, but for some specific cases the energy balance error can be substantial. The "shortcut" method (without the final iteration) can be useful during parameter estimation studies, where computational load is especially critical.

Fig. 3 compares annual average monthly sensible heat fluxes for the VIC model with and without the closed canopy energy balance. Without the canopy

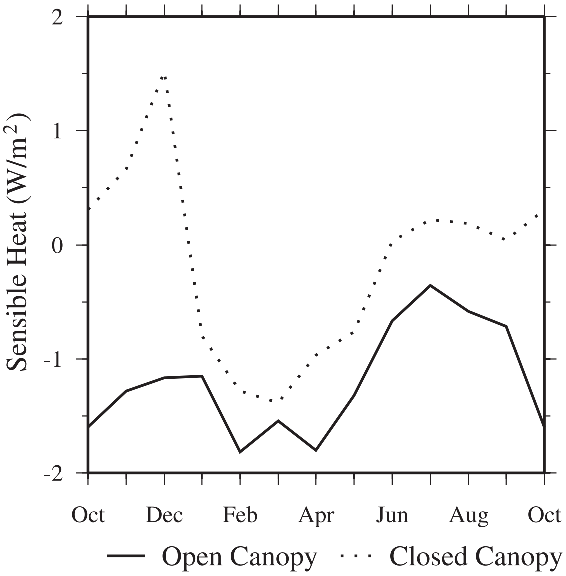

Fig. 3. Average monthly simulated sensible heat flux with and without the new canopy energy balance algorithm for the Kaalasjärvi River, Sweden.

energy balance (method of Cherkauer and Lettenmaier, 1999), the reported sensible heat comes from the understory and therefore does not show the range of the sensible heat flux exchange as seen above the canopy. This clearly demonstrates the importance of closing the canopy energy balance on the simulated surface energy fluxes.

2.2. Spatially variable snow cover

Snow-free areas play an important role in the exchange of energy at the land surface and in controlling the rate of snowmelt. Because of their lower albedo, snow-free areas absorb more shortwave radiation than neighboring snow-covered areas. Increased absorption means that the snow-free areas warm up faster, increasing the regional exchange of sensible heat as compared with a fully snow-covered area. The increased sensible heat can also be advected from the snow-free areas to neighboring snow-covered areas, increasing the rate of melt (Shook et al., 1993). The snow accumulation and ablation algorithm described by Cherkauer and Lettenmaier (1999) does not allow for spatial variability in snow coverage, and therefore does not represent these important processes. The snow cover spatial variability algorithm consists of two parts: (1) prediction of snow spatial extent; and (2) advection of sensible heat from snow-free areas.

2.2.1. Partial snow cover

During snow accumulation, the snowpack is assumed to cover the ground surface completely (Fig. 4a). Once melt begins the coverage fraction,

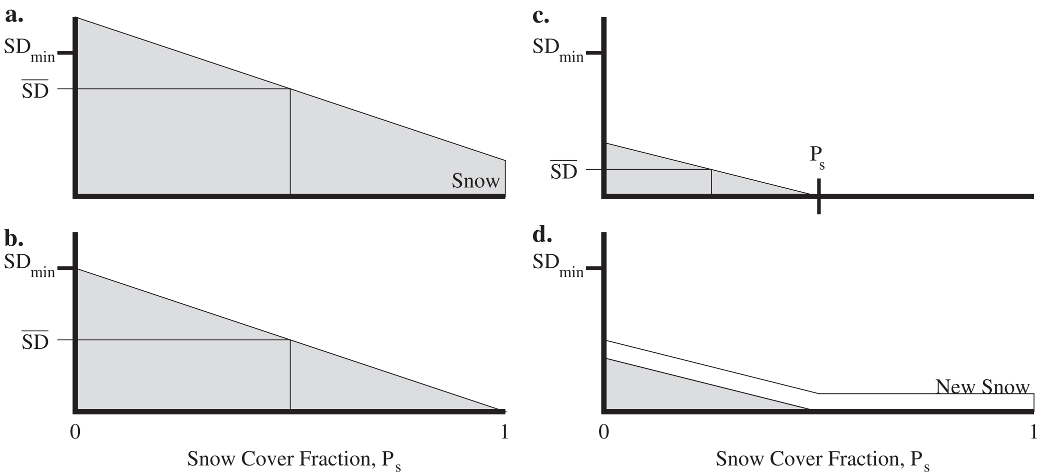

Fig. 4. Spatial snow algorithm schematic where P_s is the fractional snow coverage, $\overline{SD}$ is the average snow depth and SD_{min} is the minimum snow depth for complete snow cover: (a) deep snowpack that completely covers the area, (b) snowpack depth equal to minimum depth for full coverage, (c) snowpack covers less than the full cell area, and (d) new snow accumulation over an established snowpack distribution (Cherkauer and Lettenmaier, 2003).

P_s, is adjusted according to the depth of the snowpack at the current time step. As long as the pack depth is greater than the minimum depth required for full coverage, SD_{min}, snow cover is continuous (Fig. 4b). Once the pack depth falls below SD_{min}, the coverage fraction is calculated based on a statistical distribution of snow depth (Cherkauer and Lettenmaier, in preparation) (Fig. 4c). A uniform distribution derived from field observations of snow depth in central Minnesota (Cherkauer et al., submitted for publication) is currently used; however, this method is generalizable to other probability distributions.

There are two special cases: (1) melting of thin accumulating snowpacks and (2) fresh snow on a partially snow-free grid cell. If snowpack depth does not exceed SD_{min} prior to a melt event, the algorithm computes a new distribution by assuming that SD_{min} is equal to the current depth. This reduces the initial change in P_s, which could be quite large for a thin snowpack. The second special case is the accumulation of new snow when a grid cell has partial snow coverage. In this case, the depth distribution has already been established by a previous melt event, but new snow accumulates uniformly over the cell (Fig. 4d). If another melt event occurs shortly after this accumulation, it is assumed that the new snow has not been redistributed so the previous spatial distribution of snow cover is preserved. If accumulation continues until the mean pack depth exceeds SD_{min}, the old distribution is discarded and the snow cover distribution is reset for the next melt event. Otherwise, if melt occurs before the pack depth exceeds SD_{min}, the algorithm melts the new stored snow before returning to the original distribution.

When snow cover is not complete, iterative solutions determine the surface temperatures and energy fluxes for the snow and snow-free surfaces. A weighted average of the fluxes for snow and snow-free areas is used to balance incoming radiative fluxes. Therefore, as the snow coverage decreases the solution approaches that of snow-free ground.

2.2.2. Advection of sensible heat

The advection of sensible heat from neighboring snow-free areas can have a large influence on the melt rate of the remaining snowpack (Marsh and Pomeroy, 1996; Marsh et al., 1997). The snow-free portion of a partially snow-covered region will be warmer than the snowpack during the day as it absorbs more of the incoming radiation. Turbulent or advected sensible heat results when warm air above the snow-free surface is transported over the neighboring snow-covered areas by local winds. Observations by Marsh and Pomeroy (1996) and Marsh et al. (1997) showed that the snowmelt rate peaks when snow coverage is about 60%, at which time nearly 100% of the sensible heat flux from the bare patches is advected to the snow-covered patches. As the snow cover fraction falls below 60%, less of the surrounding snow-free areas contribute sensible heat to the melt process. Semi-empirical equations based on the work of Marsh et al. (1997) were implemented in the VIC model by Cherkauer and Lettenmaier (in preparation). Fig. 5 compares VIC model simulations using the fractional snow cover and sensible heat advection parameterizations. The figure shows that the revised model predicts higher sensible heat fluxes during spring and early summer, as compared with the model with uniform snow cover.

2.3. Spatially distributed frozen soil

The spatial variability of frozen soil is an important consideration for prediction of the impact of frozen soil on runoff generation by spring snowmelt and

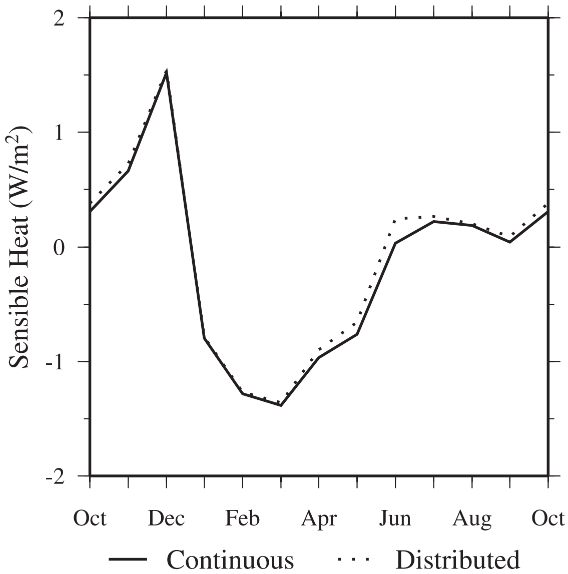

Fig. 5. Average monthly simulated sensible heat flux with uniform and spatially distributed snow cover algorithm for the Kaalasjärvi River, Sweden.

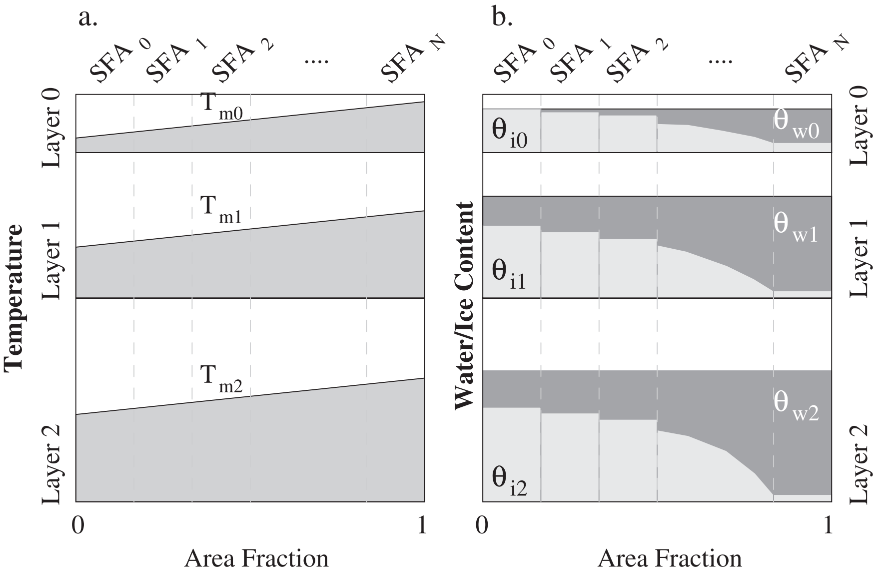

Fig. 6. Spatial frost algorithm schematic: (a) uniform soil temperature distribution and (b) nonuniform ice content distribution. Layers indicate the three VIC soil moisture layer, T_{mN} is the mean temperature for layer N, θ_{iN} is the ice content for layer N, θ_{wN} is the total moisture content for layer N and SFA_M is the fractional area of bin M (Cherkauer and Lettenmaier, 2003).

rainfall on frozen soil. Soil ice reduces infiltration, but meltwater will pond or flow across the surface until it finds an area with less soil ice that has a higher infiltration capacity (Cherkauer et al., submitted for publication). The VIC model frozen soil algorithm described by Cherkauer and Lettenmaier (1999) assumed that the soil was uniformly frozen, and as a result tends to over-predict spring floods (Cherkauer and Lettenmaier, 1999). To improve spring peak flow predictions, a parameterization of the spatial distribution of soil frost was added to the model.

As with the spatially variable snow cover algorithm, the frozen soil spatial distribution algorithm is based on a uniform probability distribution, in this case derived from soil temperatures in central Minnesota (Fig. 6a). Soil temperatures estimated from the surface energy balance calculations are assumed to be the spatially averaged values across the grid cell. The uniform distribution is only applied to soil temperatures for the calculation of unfrozen water content, estimated using the method described by Cherkauer and Lettenmaier (1999). To account for the spatial variability of unfrozen water content, each layer is divided into a series of separate bins for which ice content is computed (Fig. 6b). Liquid water is distributed evenly through all of the bins and allowed to drain proportionally to the ice content in that bin. Bins with higher ice contents will allow less water to drain, so the total soil moisture

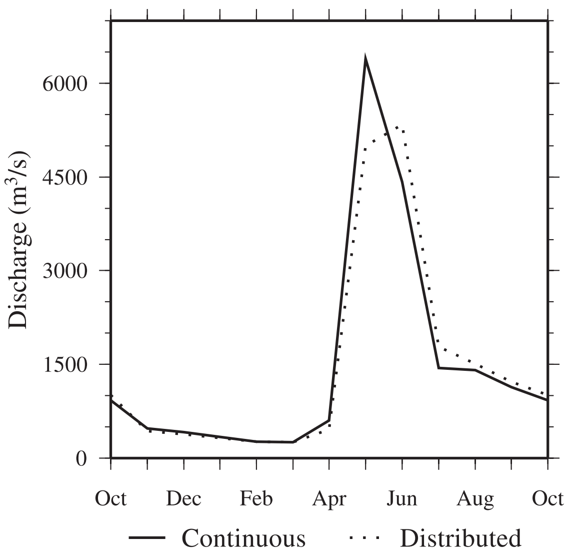

Fig. 7. Average monthly simulated discharge with uniform and spatially distributed frozen soil algorithm for the Kaalasjärvi River, Sweden.

drainage and surface infiltration for the layer is reduced. The inclusion of the spatial distribution of ice content results in some bins restricting almost all drainage, while others drain virtually unrestricted.

Fig. 7 shows the difference in predicted streamflow between the spatially uniform and spatially distributed frozen soil algorithm. Because more meltwater is able to infiltrate during spring melt, the simulated maximum flood is reduced and the spring response lasts longer relative to spatially uniform frozen soils. This is consistent with observed runoff response observed by Cherkauer and Lettenmaier (in preparation) in the upper Mississippi River basin.

2.4. Frozen soil algorithm improvements

With the operational release of the frozen soil algorithm described by Cherkauer and Lettenmaier (1999), several limitations were discovered. These included its inability to simulate permafrost, and the computational requirements for its inclusion in the full soil–vegetation–atmosphere column energy balance.

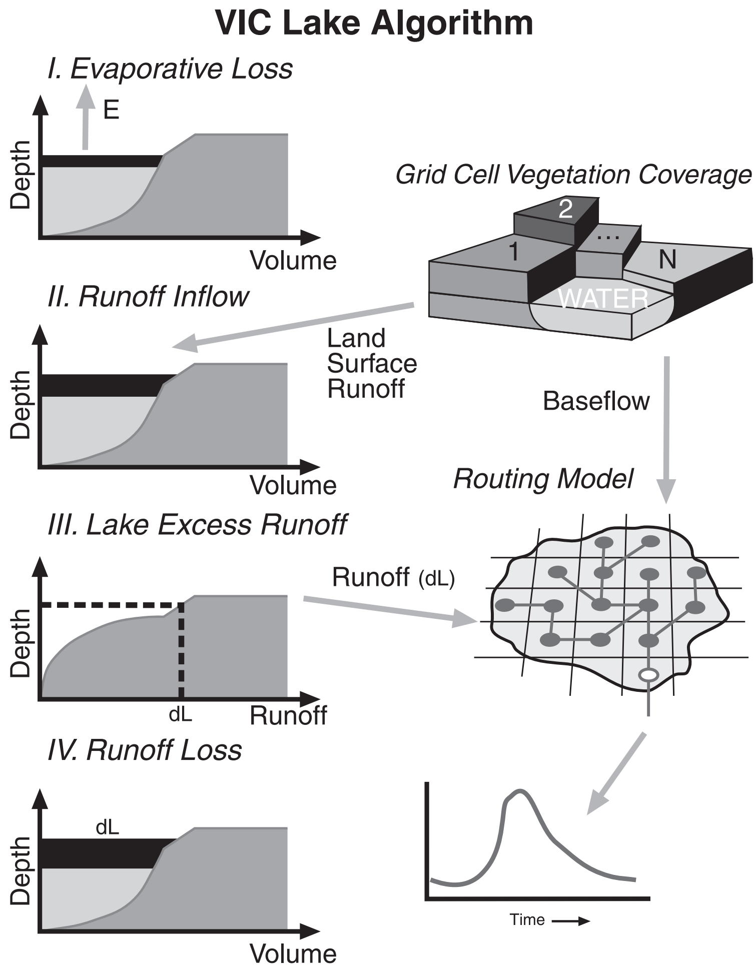

Fig. 8. Schematic of the VIC lake algorithm. Lakes are included in the mosaic representation of grid cell vegetation classes. During each time step, the lake level is adjusted by the following: (I) evaporative losses driven by the surface energy balance, (II) inflow as runoff from the rest of the grid cell and as precipitation, (III) inflows that exceed the maximum holding capacity of the lake are not affected by the presence of the lake and (IV) outflow from the lake when it is not full is controlled by a rule curve that relates lake depth to runoff rate.

Permafrost is especially important to simulations of arctic hydrology, while improvements in simulation speed will greatly increase the usefulness of the routine. The following improvements have been made to the frozen soil algorithm:

- The accounting of frozen layers within the soil column was changed to allow the algorithm to track multiple frozen and unfrozen layers. This also allows simulation of permafrost.
- An optional no-flux bottom boundary condition was added to improve solutions over multi-decade time periods.
- The soil heat flux solution was made more efficient by dynamically reducing the number of soil nodes required in the iteration for ground heat flux. When the soil is unfrozen only the top two nodes are used; as the freezing front penetrates the soil column, more nodes are added. This method reduces the computation time considerably. Since total heat flux through the soil column is not included in the iteration for the surface energy balance, this simplification does not close the energy balance completely. However, as with the closing of the canopy energy balance, the energy balance can be closed completely by changing the run-time flag for the solution method.

2.5. Lakes and wetlands

An algorithm to represent the effects of lakes and wetlands on moisture storage and evaporation has been incorporated into the VIC model by Bowling et al. (in preparation). These processes are particularly important for representing runoff in high latitude, low relief watersheds where much of the surface storage is in lakes and wetlands, rather then in the soil column. A surface water land class can be indicated for any grid cell with surface water coverage exceeding a given threshold (typically 1% or 2%). This class is treated in a manner similar to the vegetation and bare soil land classes currently employed by the VIC model (Bowling et al. in preparation). The algorithm can be summarized as follows (see Fig. 8):

- Evaporation from the water surface is calculated in each time step by solving a surface energy balance in the manner of Hostetler (1991) and Hostetler et al. (1993).

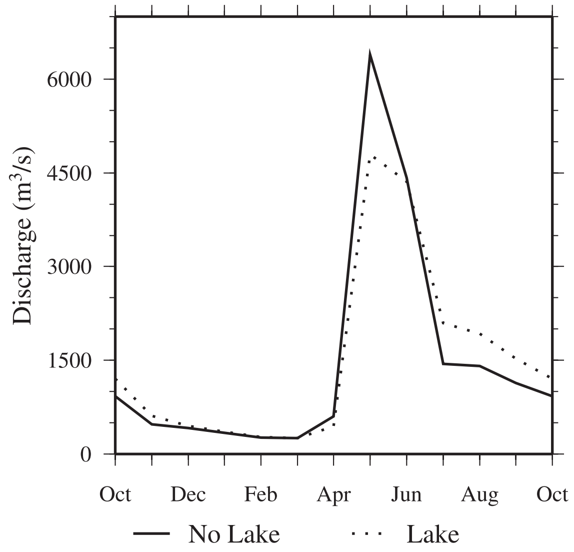

Fig. 9. Average monthly simulated discharge with and without the lake algorithm for the Kaalasjärvi River, Sweden.

- Lake freeze–thaw and snow accumulation over the frozen surface are also included (Patterson and Hamblin, 1988).
- A user-defined fraction of grid cell runoff from vegetated areas is routed through the lake, to represent the storage retardation effect of lakes on spring melt.
- Once the new stage is calculated, runoff is released from the simulated grid-cell lake according to a stage-based rule curve.

Fig. 9 demonstrates the effect of lakes on the average monthly hydrograph. The maximum monthly discharge rate is reduced, but timing changes vary little. The lake algorithm also increases baseflow during the summer and early fall.

3. Discussion

Cold season processes are important to the hydrologic cycle over much of the northern hemisphere. Properly representing these processes within a large scale hydrologic model should lead to improved understanding of the role of these processes in land surface–atmosphere interaction. Closing the canopy energy balance when snow is present means that the VIC model now returns balanced energy fluxes at the top of the canopy. Representation of the spatial distribution of snow cover improves simulation of energy fluxes and snowmelt rates during the spring melt period. On a monthly or annual average basis, the snow spatial distribution algorithm increases monthly average sensible heat fluxes by a small amount—a maximum of 0.25 W/m^2 during spring melt (see Fig. 5). Daily differences, however, can be quite significant, reaching up to 40 W/m^2 during snowmelt (Cherkauer and Lettenmaier, in preparation). Representation of the spatial distribution of soil frost reduces the overestimation of spring flood peaks, while increasing soil moisture recharge. This can have a substantial effect on monthly discharge as well as daily energy and water fluxes. The parameterization of lakes and wetlands made a large impact on VIC model simulations during PILPS Phase 2(e) where their storage effects attenuated the observed discharge significantly (Bowling et al., 2003a). The algorithms discussed in this paper facilitate use of the VIC model to explore land surface–atmosphere interactions in the arctic and the implications of climate change on high latitude hydrologic processes.

Acknowledgements

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