Modeling Hydraulic Responses to Meteorological Forcing: From Canopy to Aquifer

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An understanding of the hydrologic interactions among atmosphere, land surface, and subsurface is one of the keys to understanding the water cycling system that supports our life system on earth. Properly modeling such interactions is a difficult task because of the inherent coupled processes and complex feedback structures among subsystems. In this paper, we present a model that simulates the land-surface and subsurface hydrologic response to meteorological forcing. This model combines a state-of-the-art land-surface model, the National Center for Atmospheric Research (NCAR) Community Land Model version 3 (CLM3), with a variably saturated groundwater model, TOUGH2, through an internal interface that includes flux and state variables shared by the two submodels. Specifically, TOUGH2 in its simulation uses infiltration, evaporation, and root-uptake rates, calculated by CLM3, as source–sink terms; CLM3 in its simulation uses saturation and capillary pressure profiles, calculated by TOUGH2, as state variables. This new model, CLMT2, preserves the best aspects of both submodels: the state-of-the-art modeling capability of surface energy and hydrologic processes from CLM3 and the more realistic physical process–based modeling capability of subsurface hydrologic processes from TOUGH2. The preliminary simulation results show that the coupled model greatly improves the predictions of the water table, evapotranspiration, surface temperature, and moisture in the top 20 cm of soil at a real watershed, as evaluated from 18 yr of observed data. The new model is also ready to be coupled with an atmospheric simulation model, representing one of the first models capable of simulating hydraulic processes from the top of the atmosphere to deep ground.

Abbreviations: E, model efficiency; ET, evapotranspiration; NCAR, National Center for Atmospheric Research; WT, water table.

Land hydrologic responses to meteorological forcing involve complicated exchanges of moisture and energy between soil, vegetation, snowpack, groundwater, and the overlying atmospheric boundary layer. These exchanges occur in the form of many interactive natural hydrologic processes, including precipitation, snow and soil water melting and freezing, infiltration, storage and movement of soil moisture, surface and subsurface runoff, recharge of groundwater, and evapotranspiration. Through these processes, soil, vegetation, snowpack, groundwater, and the overlying atmospheric boundary layer often become an integrated hydrologic system at various scales. Quantitatively understanding or modeling the behavior of this integrated system is critical not only in modeling regional climate or predicting global energy and water balances but also in assessing the impact of climate change and human modifications of the natural hydrologic system on the water resources that sustain our civilizations. However, the integrated system is often modeled separately for each subsystem because the land surface is traditionally the boundary between different disciplines in the scientific and engineering community. For example, many climate models, surface-water models, and vegetation and ecology models often take the land surface as the lower boundary, parameterizing the subsurface processes in various simplified ways (e.g., runoff coefficient, evaporation coefficient). On the other hand, many physically based subsurface or groundwater models often take the land surface as the upper boundary by lumping the complex processes above the surface as known boundary conditions (e.g., net infiltration or hydraulic head). However, in nature, the hydraulic processes from canopy to aquifer often form an integrated surface–subsurface system through complicated interactions. As a result, such simplified models cannot properly describe how the real system behaves, in many cases resulting in unacceptable errors. During the last few decades, much progress has been made in developing more realistic models to simulate hydraulic interactions through the land surface. Instead of simply taking the land surface as the boundary of the modeling domain, many recent models simulate with various approaches the lower portion of the atmosphere and upper portion of the subsurface as an integrated system, by which the atmosphere–land interactions become internal processes (Abromopoulos et al., 1988; Famiglieti and Wood, 1991; Wood et al., 1992; Liang et al., 1994; Bonan, 1998; Dai and Zeng, 1997; Walko et al., 2000; Gutowski et al., 2002; York et al., 2002; Liang et al., 2003; Oleson et al., 2004; Niu and Yang, 2006). CLM3 is one such model, primarily developed to meet the needs of regional climate modeling. In CLM3, radiation, sensible and latent heat transfer, zonal and meridional surface stresses, and ecological

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and hydrological processes are simulated as interrelated subprocesses, using hybrid approaches (i.e., combinations of physically based dynamic modeling and experientially based parameterization models). However, the model of subsurface moisture flow in CLM3 is still overly simplified. In this regard, TOUGH2 offers a more realistic physical process-based modeling capability for subsurface hydrologic processes (including heterogeneity, three-dimensional flow, seamless combining unsaturated and saturated zones, and water table). Therefore, coupling these two models is an attractive way to build a useful model of a surface–subsurface hydraulic system.

The objectives of this study are (i) to develop a new model of atmosphere–land–subsurface hydraulic interactions at watershed or regional scales by combining the best aspects of both CLM3 and TOUGH2, and (ii) to show the importance of realistically modeling both surface and subsurface processes, as well as their interactions in predicting the hydrologic responses to meteorological forces, by applying the new model to a watershed in Russia over an 18-yr period.

Modeling Approaches

The new model, CLMT2, can be seen as combining CLM3 and TOUGH2 (Module EOS9 only, called "TOUGH2" below for simplicity) in a sequential coupling. It inherits most of the modeling capabilities of both CLM3 and TOUGH2. A detailed technical description of CLM3 can be found in the NCAR technical note (Oleson et al., 2004); a summary of EOS9, an unsaturated–saturated water flow simulation module within the TOUGH2 package, is found in Wu et al. (1996).

From the perspective of CLM3, the new model no longer simulates the subsurface moisture movement as a one-dimensional process by explicit scheme. Instead, the three-dimensional Richards equation is solved implicitly by TOUGH2. In particular, the assumption that the permeability decreases exponentially from top to bottom of the soil is no longer used, and the groundwater depth is no longer a parameter calculated as saturation-weighted depth. Therefore, CLMT2 can be more flexible in addressing complex subsurface environments. From the perspective of TOUGH2, the new model no longer takes the net infiltration or root uptake as a prescribed boundary condition or source–sink term. Instead, the net infiltration and root uptake result from simulations of coupled energy, wind, vegetation, and hydraulic processes by CLM3. As a result, CLMT2 expands the scope of TOUGH2 such that more realistic modeling of land-surface conditions is possible.

Table 1 lists the major differences in simulating subsurface flow between CLM3 and the coupled model, CLMT2.

Spatial Discretization and Grid Structure of CLMT2

The modeling domain below land surface is discretized into connected grid cells similar to a TOUGH2 grid. Different from a regular TOUGH2 grid, however, the grid cells in the upper portion (the root zone) of a CLMT2 grid must be geometrically "regular," so that they can form grid columns. The aerial extent of each grid column corresponds to the grid cell of a regional climate model. Above each grid column, nested hierarchical grid structures are created to capture land-surface heterogeneity within the area. An area can contain multiple, noninteractive land units (e.g., glacier, wetland, vegetated, lake, and/or urban). Each land unit (except lake) can contain multiple, noninteractive "snow–soil" subcolumns. Similarly, each snow–soil type can contain multiple, noninteractive plant functional type patches (Bonan et al., 2002). The term *noninteractive* indicates that there is no communication among substructures at the same level. In other words, they are logically isolated subareas splitting the entire area. Besides the snow–soil subcolumns, which can have multiple layers, all other substructures are one-layer or single-node structures. Note that the soil subcolumns spatially overlap the root zone of the subsurface grid column where the communication between TOUGH2 and CLM3 takes place. In addition, the snow–soil subcolumns are also used for calculations of thermal transfer and freezing–melting processes in snow cover and soil, because the EOS9 of TOUGH2 does not account for those processes.

Modeling of Processes in CLMT2

Models of water flow in subsurface are based on numerical solutions of the Richards equation:

$$\frac{\partial\theta}{\partial t} = \nabla\cdot\left[k_s k_r \nabla\psi_h\right] + q_s - q_{root} \qquad [1]$$

with a flux continuation condition at land surface:

$$-k_s k_r \left.\frac{\partial\psi_h}{\partial z}\right|_{\text{at landsurface}} = q_{net} \qquad [2]$$

where θ, ψ_h, and k_s, k_r are the volumetric water content, the hydraulic potential, the saturated hydraulic conductivity, and the relative permeability, respectively. The term q_{root} is root uptake rate, while q_s indicates other source–sink terms that may exist in the subsurface (e.g., wells). The root uptake rate varies spatially and depends on the root distribution in the root zone and the transpiration from dry leaf surfaces (E_v^t):

TABLE 1. Major differences between the models CLM3 and CLMT2 in simulation of subsurface flow.

CLM3	CLMT2
Assumes that permeability decreases with depth exponentially.	Spatially variable permeability is user specified.
Richards equation is solved explicitly (no iteration in each time step).	Richards equation is solved fully implicitly.
Clapp and Hornberger (1978) relationships are used for hydraulic functions of soil.	van Genuchten relationships are used for hydraulic functions of soil.
Hydraulic properties are assigned generally based on the soil texture classification.	Hydraulic properties are provided as input by the user for the specific site.
Soil moisture stress for root uptake is either 0 or 1 (dead or live).	A piecewise linear function is used to simulate the soil moisture stress for root uptake.
Soil columns are isolated from one another, and subsurface drainage (base flow) is calculated as a value proportional to the saturation weighted average saturated hydraulic conductivity in lower soil layers and exp(− water table), which is then deducted from the soil each time step.	Lateral subsurface flow if any is included naturally in three-dimensional flow simulation. No artificial subsurface drainage is included.
Soil depth is limited to 3.5 m.	Soil depth, usually larger than 3.5 m, is specified by the user so that the domain bottom is deeper than the water table.

$$q_{\text{root}} = E_v^t r(z) = \left[-\frac{\rho_{\text{atm}}\left(h_{\text{can}} - h_{\text{can}}^{\text{sat}}\right)}{r_b} r_{\text{dry}} \beta_t \right] r(z) \quad [3]$$

where $r(z)$, varying with the depth z, is the effective root fraction, a product of the root fraction and the soil stress. The terms ρ_{atm}, h_{can}, $h_{\text{can}}^{\text{sat}}$, r_b, and β_t are the density of atmospheric air, specific humidity of canopy air, saturated water-vapor specific humidity at the vegetation temperature, leaf boundary stomatal resistance, and total soil moisture stress to the root uptake, respectively. The shade factor (r_{dry}) is calculated as a function of the sunlit (L^{sun}) and shaded (L^{sha}) leaf area indices:

$$r_{\text{dry}} = \frac{f_{\text{dry}} r_b}{L} \left[\frac{L^{\text{sun}}}{r_b + r_s^{\text{sun}}} + \frac{L^{\text{sha}}}{r_b + r_s^{\text{sha}}} \right] \quad [4]$$

The term f_{dry} is the fraction of leaves that are dry, and r_s^{sun} and r_s^{sha} are the sunlit and shaded stomatal resistances, respectively.

The net infiltration rate (q_{net}) in Eq. [2] is calculated from the surface water-balance equation (the run-on process is not simulated in the model and all runoff water will be removed immediately):

$$q_{\text{net}} = q_0^{\text{liq}} - q_{\text{runoff}} - E_g \quad [5]$$

where q_0^{liq} is the rate of liquid water reaching the soil surface and E_g is the water vapor flux at soil surface. The rate q_0^{liq} could be the summation of throughfall rate ($q_{\text{thru}}^{\text{liq}}$) and canopy drip rate ($q_{\text{drip}}^{\text{liq}}$) if no snow cover exists or the flow rate of liquid water reaching the soil surface from the snow layers (including melting water). The throughfall rate is the liquid precipitation (q_{rain}) that directly falls through the canopy and is calculated as

$$q_{\text{thru}}^{\text{liq}} = q_{\text{rain}} \exp\left[-0.5\left(L+S\right)\right] \quad [6]$$

where L and S are the exposed leaf and stem area index, respectively. The canopy drip rate is calculated from the canopy interception model, while the flow rate of liquid water reaching soil surface from the snow layers is an output of the snow processes model. Both models are described in detail in Oleson et al. (2004) and are not repeated here.

The other two terms in Eq. [5], the surface runoff (q_{runoff}) and the water vapor flux at soil surface (E_g), along with the transpiration (E_v^t) and the net infiltration rate (q_{net}) mentioned above, are four important fluxes that connect the surface and subsurface processes in CLMT2.

If the top soil layer is not impermeable, the surface runoff is the sum of runoff from saturated and unsaturated areas:

$$q_{\text{runoff}} = \left[f_{\text{sat}} + \left(1 - f_{\text{sat}}\right) w_m^4 \right] q_0^{\text{liq}} \quad [7]$$

where f_{sat} and w_m are the fraction of saturated area and the mean wetness in the top three layers, respectively. In particular, the fraction of saturated area is a function of water table depth (z_w):

$$f_{\text{sat}} = w_{\text{fact}} \min\left[1, \exp\left(-f_z z_w\right)\right] \quad [8]$$

where w_{fact} and f_z are the fraction of wet land area and a constant scaling factor (f_z = 1 m^{-1}), respectively.

The water vapor flux at soil surface (E_g) reflects the net result of soil surface evaporation and dew. It is driven by the gradient of specific humidity between the ground surface and the atmosphere (nonvegetated surface) or the canopy (vegetated surface) as follows:

$$E_g = -\frac{\rho_{\text{atm}}\left(h_{\text{atm}} - h_g\right)}{r_{\text{aw}}} \quad [9a]$$

for nonvegetated surface, and

$$E_g = -\frac{\rho_{\text{atm}}\left(h_{\text{can}} - h_g\right)}{r_{\text{agc}}} \quad [9b]$$

for a vegetated surface, where ρ_{atm}, h_{atm}, h_g, and h_{can} are the density of atmospheric air, the atmospheric specific humidity, the specific humidity of the soil surface, and the canopy air specific humidity, respectively. The other two terms, r_{aw} and r_{agc}, are the aerodynamic resistance to water vapor transfer between the ground and the atmospheric air at the reference height, and that between the ground and the canopy air, respectively. The aerodynamic resistances are calculated using a surface-layer model based on Monin–Obukhov similarity theory. The water-vapor flux is simulated as a part of the coupled surface energy, momentum, and moisture model, described in detail in Oleson et al. (2004) and not repeated here.

Figure 1 shows a brief flowchart of CLMT2 for one time step. For a given meteorological forcing at each time step, CLM3 modules simulate canopy and surface processes sequentially and column by column, using the water table (WT), water content [$W(i)$], and capillary pressure [$Pc(i)$] calculated by the TOUGH2 module at the previous time step. The resulting net infiltration

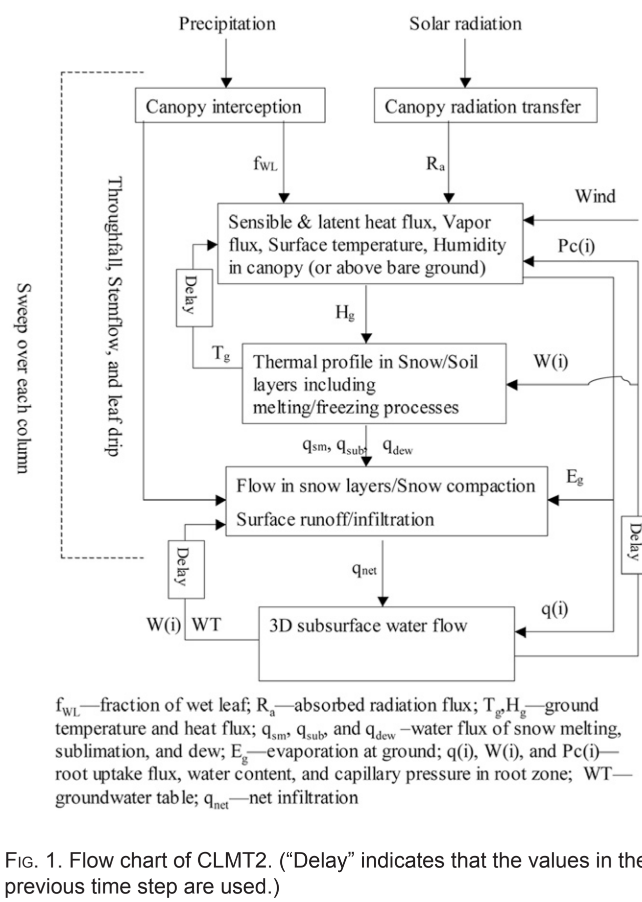

Fig. 1. Flow chart of CLMT2. ("Delay" indicates that the values in the previous time step are used.)

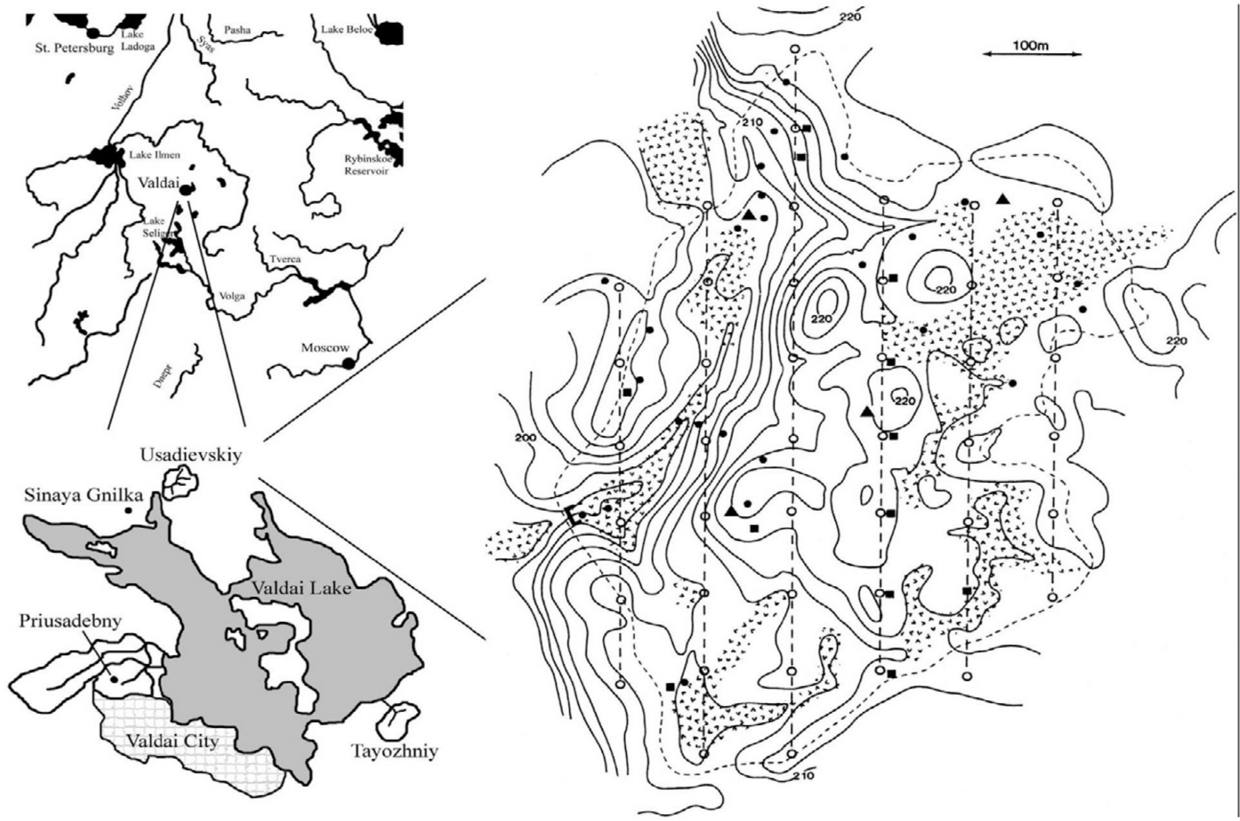

Fig. 2. Map of the Usadievskiy Catchment at Valdai, Russia, and its location (adapted from Fig. 1 in Luo et al., 2003). Filled circles are water-table measurement sites. Open circles with dashed lines indicated the snow measurement sites and routes, respectively. Discharge is measured at the stream outflow point of the catchment (see bold bracket) at the lower left-hand corner of the catchment map. Filled triangles indicate the measurement sites of soil freezing and thawing depths. The short dash line denotes the catchment boundary. Hatched areas denote regions of swampy conditions. Elevation contours are in increments of 2 m.

rate (q_{net}) and root uptake flux [$q(i)$] are then used as source–sink terms in subsurface flow simulation by the TOUGH2 module. This sequential coupling approach is mass conserved but could be inaccurate if the size of the time step was too large. However, the time step required for simulating the surface processes by CLM3 is usually so small (e.g., in order of hours) with respect to the subsurface processes that the sequential coupling approach will not add any new time-stepping limit.

Results and Discussion

Usadievsky Catchment, Valdai, Russia, is a midlatitude grassland catchment, with deep snow cover in the winter and significant precipitation in the summer (Fig. 2). Eighteen years of observation data related to this catchment were used extensively within the Project for Intercomparison of Land-Surface Parameterization Scheme (PILPS) and provided a very robust validation for surface–subsurface models (Maxwell and Miller, 2005). All of the observations were made available by Robock et al. (2000) and Luo et al. (2003) as part of the Global Soil Moisture Data Bank. The precipitation data within the original meteorological forcing data at 3-h intervals were scaled by the observed monthly precipitation, so that the precipitation as model input was consistent with the observed ones at the temporal scale of 1 mo.

For subsurface simulation in CLMT2, the hydraulic parameters used in this study are the same as those in Maxwell and Miller (2005). The entire catchment (0.36 km^2) is simulated as a one-dimensional column down to the depth of 6 m. The vertical discretization of subsurface is 0.1 m except the top two cells, for which 0.01 and 0.09 m are used, respectively. Table 2 lists the major model parameters used in the simulation.

For subsurface simulation in CLM3, the entire catchment (0.36 km^2) is also simulated as a one-dimensional column, but the depth of the domain is 3.5 m as hard coded in CLM3. Furthermore, the vertical spatial discretization varies from 0.0175 to 1.137 m from top to bottom, which is also hard coded in CLM3. All the parameters for the land surface and the meteorological forcing data are the same for both models.

The simulated and observed daily snow depths are presented in Fig. 3. Both CLM3 (blue dash line) and CLMT2 (red

Table 2. Model parameters used in Vaidai simulation.

Parameter	Value	Unit
van Genuchten alpha	1.95	m^{-1}
van Genuchten exponent	1.74	unitless
Saturated hydraulic conductivity	1.21	m d^{-1}
Effective soil porosity	0.401	m^3 m^{-3}
Residual saturation	0.136	unitless
Lower critical point at which root uptake stops	–5270.81	mm H_2O
Upper critical point at which root uptake stops	0.1	mm H_2O
Fraction of model area with high water table	0.15	unitless
Latitude	57.6N	degree
Longitude	33.1E	degree
Vegetation type index	7 (grassland)	unitless
Soil type index	6 (loam)	unitless

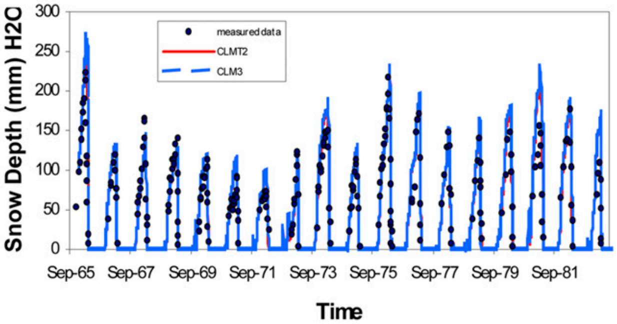

FIG. 3. Simulated and observed snow depth (in mm of equivalent water).

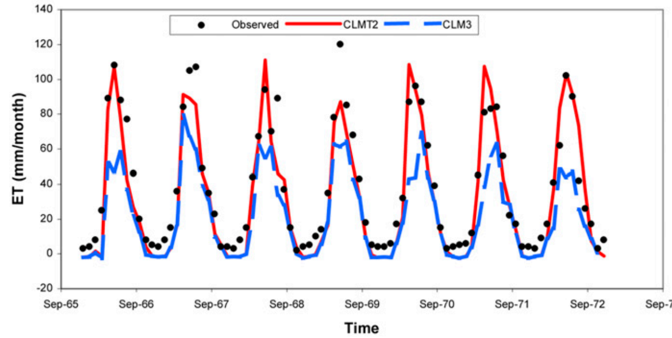

FIG. 4. Simulated and observed monthly evapotranspiration (ET).

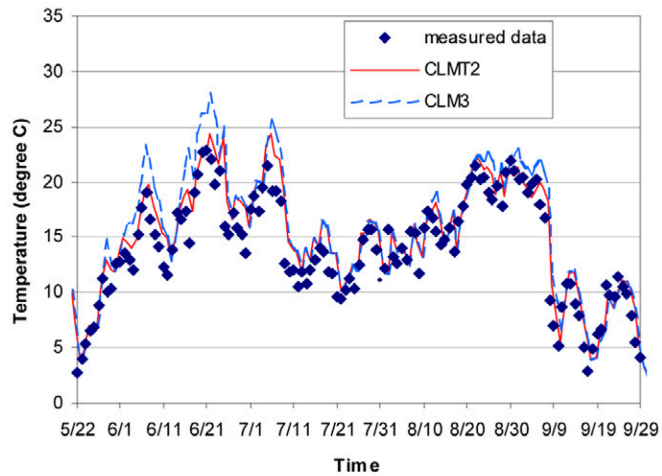

FIG. 5. Simulated and observed ground surface temperature during summer (22 May 1968–29 Sept. 1968 shown; others are similar).

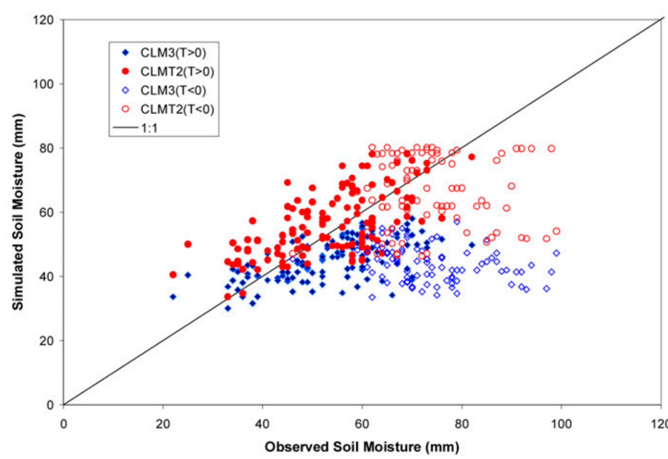

FIG. 6. Comparison between the simulated monthly averaged moisture in the top 20 cm of soil by CLM3 (blue diamond) and CLMT2 (red circle) vs. measured data over 18 yr. Here T is the monthly average ground surface temperature (°C). Therefore, T < 0 indicates freezing month whereas T > 0 indicates warm month.

solid line) predict almost-identical results that agree well with the observed snow depth (black dots). This convergence between the two models is expected because of the halt in surface–subsurface hydraulic interactions during the frozen winter season. As a result, the accuracy of the subsurface simulation does not matter in simulating the snow accumulation process on the land surface.

However, CLMT2 does significantly improve the predictions of monthly evapotranspiration (ET) (Fig. 4). As shown in Fig. 4, CLM3 underestimates the ET compared with the observed data, whereas CLMT2 is in close agreement with the observed data. This model improvement can be evaluated quantitatively by the model efficiency (E) proposed by Nash and Sutcliffe (1970) that is defined as one minus the sum of the squared differences between the predicted and observed values normalized by the variance of the observed values during the period under investigation. The range of E lies between 1.0 (perfect predict) and $-\infty$. For the 84 mo of monthly ET data shown in Fig. 4, E is 0.635 for CLM3 and 0.865 for CLMT2, respectively. The improvement is significant.

Consistent with underestimating ET, CLM3 often overestimates the surface temperature during the summer season (Fig. 5, only 4 mo of 1968 shown for a clear presentation; others are similar). The coupled model, CLMT2, is more accurate in this case as well. These results indicate that the impact of subsurface flow on surface processes during nonfrozen seasons is significant and that correctly simulating the subsurface flow is very important.

Evapotranspiration is one of the important moisture and energy exchanges between land and atmosphere and is the most distinct process that tightly connects the near-surface atmosphere, vegetation, soil, and groundwater together. Temperature, humidity, and wind speed are three major meteorological factors that drive ET processes. The canopy structure and wetness regulate how effective these factors will be in driving the ET processes, whereas the type of plant root controls how deeply the subsurface moisture movement will be affected. Soil moisture status in the root zone directly controls the availability of the soil moisture for ET, while the groundwater serves as a major buffer that tends to reduce soil moisture variations. Logically, the major reason for CLM3 to underestimate the ET would be its underestimation of the soil wetness in the root zone, because the modeling approaches for the surface processes are identical in the two models. Comparisons between observed moisture amounts in the top 20 cm of soil and the values predicted by CLM3 and CLMT2 confirm that this is the case (Fig. 6). The observed moisture data were computed using data from 9 to 11 observational points distributed over the basin area. Note that the observed soil moisture data contain significant noise; especially in winter when the soil is frozen, the measured moisture often exceeds the holding capacity of 802 mm for the 20-cm soil (as defined by the porosity of 0.401). Except for these outliers,

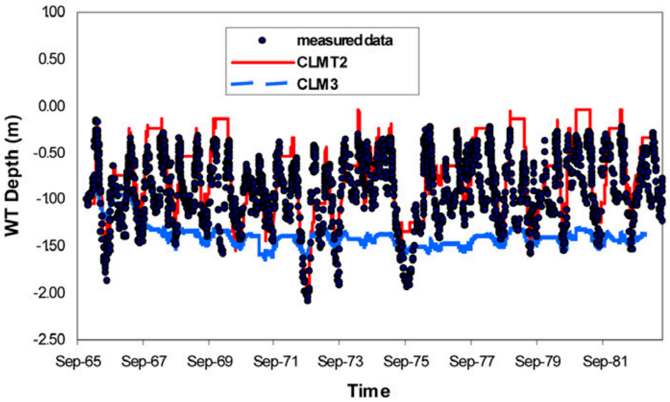

FIG. 7. Simulated and observed daily water table (WT).

the values predicted by CLMT2 are much closer to the 1:1 line than those by CLM3. The underestimation of the available soil moisture in the root zone, especially in the top 20 cm, where most roots are located, causes CLM3 to underestimate ET. Correct simulation of subsurface processes is thus important not only in catching the dynamic responses in the subsurface itself but also in estimating surface moisture and energy fluxes.

Figure 7 shows a comparison of the observed daily WTs with those simulated by CLM3 (blue line) and CLMT2 (red line), respectively. The observed WT data are derived from a site average of 19 observation wells at a subweek scale. CLM3 uses a special parameterization scheme to calculate the WT as a soil saturation-weighted depth, while the WT is automatically determined as the interface between the unsaturated and saturated soil layers simulated by CLMT2. As shown in Fig. 7, CLMT2 replicated most groundwater seasonal responses to meteorological forcing. CLM3, however, poorly estimated such responses, especially in the magnitude of WT variations. The Nash–Sutcliffe efficiency (E) is 0.216 for CLMT2 and –3.921 for CLM3, respectively. The negative E obtained by CLM3 indicates that the mean value of the observed time series of WT would have been a better predictor than CLM3. Although CLM3 was not designed to be an accurate predictor of the WT variations at first place, the result shown in Fig. 7 implies that poor estimation of water table could be an important reason for the reduced accuracy of ET prediction. In CLM3, the assumption that the permeability of soil decreases exponentially with the increase of the depth is used, which unrealistically limits the moisture movement in the soil, including both unsaturated and saturated zones. As a result, the simulated WT is much less responsive to meteorological forcing than the reality. One example of such weak responsiveness is the underestimated capability of the groundwater to supply the moisture to the root zone that feeds to the ET requirement. Furthermore, the overall low predicted WT by CLM3 can be attributed to its parameterization scheme of subsurface drainage, which tends to overestimate the subsurface drainage rate.

Note that neither models caught the lowering of the WT during winter (Fig. 7). This is most likely a result of subsurface discharge flow below the frozen zone, which could not be accounted for by CLMT2 with this single column model, whereas CLM3 accounts for the subsurface drainage improperly, as discussed above. As a sensitivity study, a "constant-head" cell is added to the TOUGH2 grid to mimic the subsurface discharge flow in the CLMT2 model. The parameters of the "constant-head" cell that regulate the subsurface discharge flow are estimated based on the stream/catchment ratio and the average slopping of the catchment. As shown in Fig. 8, the WT depth simulated by CLMT2 only partly catches the winter lowering of WT. The efficiency E improved to 0.439. Because the Usadievsky Catchment is a small part of the Valdai watershed (Fig. 2), the subsurface discharge problem is further complicated by the unknown regional groundwater flow. Consequently, a distributed model would be required to investigate this problem (which should be a good topic for further studies). Unlike CLM3, the new model, CLMT2, has the capability to simulate three-dimensional regional groundwater flow, provided that adequate field information is available.

Conclusions

A model that combines the ability to simulate the land-surface and subsurface hydrologic responses with meteorological forcing, CLMT2, has been developed, by combining a state-of-the-art land-surface model, the NCAR Community Land Model version 3 (CLM3), and a variably saturated groundwater model, TOUGH2, through an internal interface that includes flux and state variables shared by the two submodels. CLM3

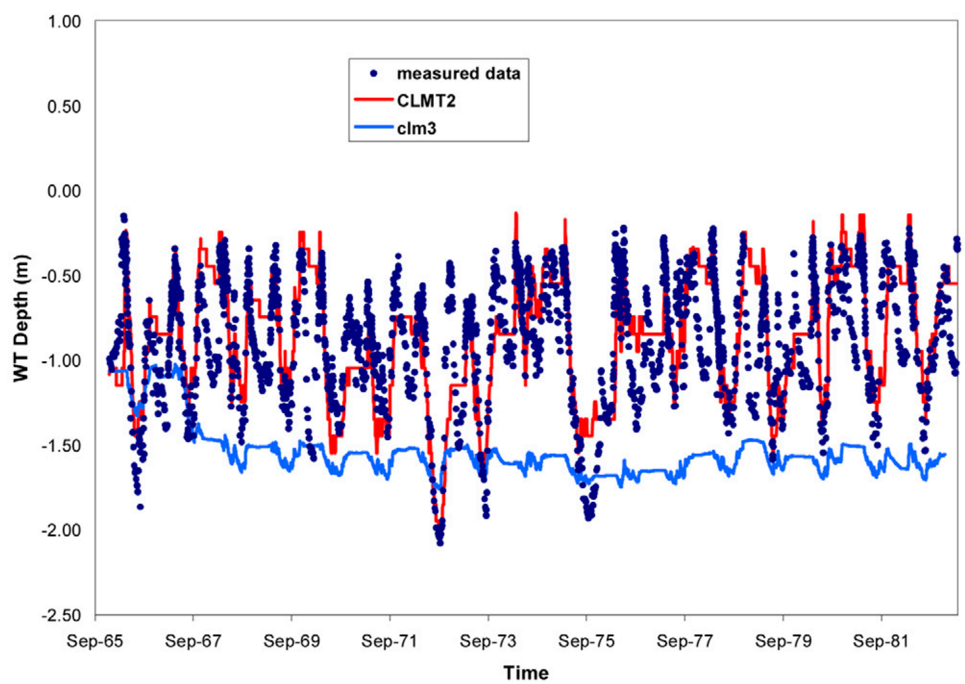

FIG. 8. Simulated and observed water tables (WT) (CLMT2 has a "constant head" cell).

provides the state-of-the-art modeling capability for surface energy and hydrologic processes, including snow, runoff, freezing and melting, evapotranspiration, radiation, and biophysiological processes. TOUGH2 offers the more realistic physical process–based modeling capability of subsurface hydrologic processes, including heterogeneity, three-dimensional flow, seamless combining of unsaturated and saturated zone, and water table. This new model, CLMT2, preserves the best aspects of both submodels. It is also ready to be coupled with an atmospheric simulation model, representing one of the first models that is capable of simulating hydraulic processes from top of the atmosphere to deep ground.

Eighteen years of observed data from Usadievsky Watershed, Valdai, Russia, were used to evaluate the performance of the new model. Compared with the old model, CLM3, the new model, CLMT2, greatly improves the predictions of the water table, evapotranspiration, surface temperature, and the moisture in the top 20 cm of soil at the real watershed. This is particularly true in the nonfrozen season, when the interactions between surface and subsurface are significant. These results also indicate that correct simulation of subsurface flow (including the water table) is very important not only in assessing subsurface water resource itself but also in simulating surface processes such as evapotranspiration or land-surface temperature, the two most important feedback factors for regional climate.

Although the results obtained in this study are promising, the comparisons so far have been limited to a one-dimensional column simulation. The three-dimensional simulation capability of CLMT2 needs to be tested in the future. In particular, the modeling capability of surface lateral flow (e.g., runoff/run-on processes) needs to be improved in future development of CLMT2 to simulate the spatial details of the hydrological responses within a watershed. In addition, modeling the hydrologic processes in the frozen soil (i.e., nonrigid media) is another challenge for improving CLMT2, as revealed in comparison with the observed soil moisture data.

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