Implications of a regional scale soil temperature and freezing model in the Upper Danube Basin for climate change scenarios

Markus Muerth & Wolfram Mauser

Abstract

Frozen soils influence Alpine runoff generation especially during snow melt events. Though exact quantification of lateral runoff due to soil ice is difficult, it is shown that a functional model of soil freezing and frozen soil runoff can reproduce snow melt runoff peaks more closely than the same mesoscale hydrological model without the freezing component. The physically based soil heat transfer module (SHTM) of the Global Change Decision Support System DANUBIA reproduces soil temperatures well with the surface energy balance being computed and is therefore ready for scenario applications. First validation results of simulated high flows in winter and a preliminary scenario analysis are presented in the following.

Keywords: climate change, energy balance, hydrology, soil ice, soil temperature

1 Introduction

The project GLOWA-Danube aims at the development and application of DANUBIA, an integrated Global Change Decision Support System to simulate water and related matter and energy fluxes in the Upper Danube Basin. A more detailed description of DANUBIA and its landsurface component can be found in this issue (Mauser & Muerth 2008). Part of the landsurface model in DANUBIA is the soil module that was upgraded to predict soil temperature and soil freezing in conjunction with the surface energy balance in the context of climate change scenarios. In case of snow cover, we also implemented a coupling algorithm between the upper soil layer and the base of the snow pack. Outputs of the Soil Heat Transfer Module (SHTM) are used by the DANUBIA components for the biogeochemical cycle, the farming practices, runoff generation and others. SHTM combines simplified physical algorithms for the computation of the actual temperature in the upper soil layers and an analytical lower boundary condition to represent climate change conditions. Changes in soil moisture and soil freezing are explicitly taken into account. The ground heat flux as the driving force of the model is provided by DANUBIA components for the radiation balance and surface fluxes. Soil temperature of the upper layer feeds back into the energy balance of the soil surface or snow pack, while soil temperature in the root zone is used as input by biological and agro-economical model components of DANUBIA.

Regarding Alpine regions, frozen soil water plays an important role in evaluating climate change impacts, because the relative soil frost duration increases with elevation up to the permafrost regions in high Alpine environments. . The validation of SHTM against time series of soil temperature is a result of this work. Then we show the modeled influence of frozen soil water on runoff generation especially during snow melt events in comparison to gauge measurements. A first climate change scenario run gives an estimate of potential future changes.

2 Model Description

2.1 The Landsurface Component

The DANUBIA landsurface component consists of five interdependent modules that simulate the surface water fluxes of each proxel (process pixel, resolution: 1 km²). Recent developments in DANUBIA account for the processes needed to close the energy and matter cycles of the landsurface and the coupling of the landsurface component to regional climate models (Marke & Mauser 2008). The modules of the landsurface component are:

a) RADIATION BALANCE: The radiation module calculates the radiation balance according to its geographical location, sun angle and cloud cover for both vegetation layers and the ground surface (Mauser & Bach 2008).
b) SNOW: The energy and water balance of snow covers and glaciers are computed by physically based algorithms and can be simulated at the subpixel scale (Prasch et al. 2008).
c) BIOLOGICAL: A plant physiological module based on the work of Farquar et al (1980) calculates the transpiration, biomass production and energy balance of the canopy.
d) SOIL: The soil hydraulic module calculating the soil water content with an extended, multi-layer Eagleson approach (Eagleson 1978) can produce lateral runoff as well as percolation. The soil heat transfer module calculates soil energy fluxes and storage (including soil freezing) for the soil layer stack.
e) SURFACE: the aerodynamic module consists of algorithms for the removal of transpired water vapor into the atmosphere (Monteith & Unsworth 1990) and an energy balance algorithm for non-vegetated surfaces and the soil surface below canopies (Muerth 2008).

2.2 The SOIL Module

Recent improvements and additions to the soil model required the adjustment of the soil layer stack to 4 layers (0.05/0.15/0.45/1.35 m). This led to a better representation of lateral runoff processes and was essential for a realistic simulation of soil energy fluxes (figure 1). Simulation of fluxes in the upper layers was tested with the *Newman* criterion and sufficient overclocking of these layers was implemented.

To simulate water fluxes in the soil column, SOIL uses a modified version of the *Eagleson* model (Eagleson 1978) that predicts infiltration and exfiltration of the soil column. Originally used in PROMET (Mauser & Schädlich 1997), in recent years

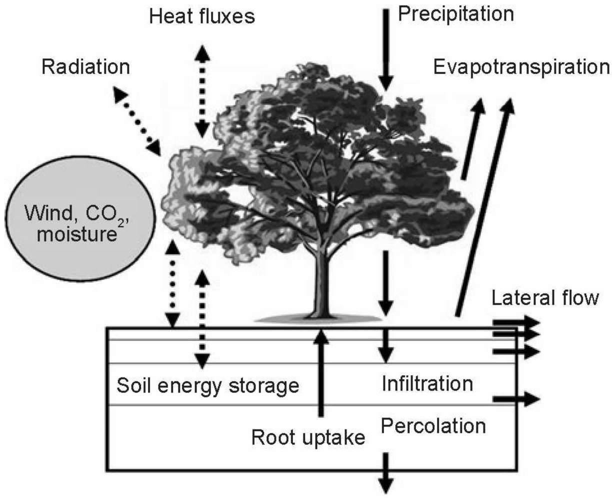

Figure 1: Energy and mass transfer at the DANUBIA land surface.

the model was extended for a soil layer stack with up to 4 layers (Mauser & Bach 2008) and implemented in the decision support system DANUBIA. The algorithm basically distinguishes between "wet" and "dry" time steps. Water sources for a soil layer can be infiltration from above (effective precipitation or percolation from upper soil layer) and capillary rise from the groundwater table or the lower soil layer. Water sinks can be evaporation (top layer), root water uptake (all layers with roots) and gravitational drain (which is summed up with capillary rise for the net percolation of a soil layer). The *Philips* equation handles actual infiltration. Excess water is added to the overland flow. If the net percolation of a soil layer exceeds the infiltration capacity of the soil layer below, the remaining water is added to the model output "interflow". All computations in the soil layer stack are run "top down", which means that the most active upper layer is run first and the lowest, i.e. the least dynamic layer is run last.

2.3 Basic equations of the soil temperature model

Basis for the computation of soil temperatures for each soil layer are the one-dimensional, conductive heat transfer equations

$$G(z) = -\lambda \frac{\partial T}{\partial z} \tag{1}$$

$$C_s \frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(\lambda \frac{\partial T}{\partial z}\right) \rightarrow \frac{\partial T}{\partial t} = \frac{\lambda}{C_s}\left(\frac{\partial^2 T}{\partial z^2}\right) \tag{2}$$

where $G(z)$ is the heat flux [W/m²] at depth z, C_s is the heat capacity [J/m³K] and λ is the heat conductivity [W/mK].

For homogenous layers and a fixed time step, one can reduce the heat flow $G_{1,2}$ from mean depth of layer 1 (z_1) to the mean depth of layer 2 (z_2) by writing

$$G_{1,2} = -\lambda \frac{T_1 - T_2}{z_2 - z_1} \tag{3}$$

The driving variable of any soil temperature model is the surface ground heat flux G_0, which can significantly change the temperature of the upper soil layer during a time step of one hour. If the upper boundary is forced by a soil surface temperature T_0, the resulting surface ground heat flux is computed by

$$G_0 = \lambda \cdot \frac{T_0(t) - T_1(t_0)}{0.5 \cdot d_1} \tag{4}$$

Like in the modified *Eagleson* model, the forcing at the upper boundary is transfered from top down to the next soil layer. After computing the heat fluxes, the new mean temperature of any layer of the soil layer stack with thickness d at time step $t = t_0 + \Delta t$ is calculated by

$$T(t) = \frac{(G_{upper} - G_{lower}) \cdot \Delta t}{d \cdot C_v} + T(t_0) \tag{5}$$

The heat flux commences downwards with the downward heat flux being the upper one of the next lower layer. For the lowest layer n of the soil layer stack (here: n=4) the lower heat flux $G_{n,n+1}$ is influenced by lower boundary condition T_{n+1} computed by an analytical solution (e.g. Hillel 1998):

$$T(z,t) = T_{av} + A_y \cdot e^{-z/D} \cdot \cos\left(\omega(t - t_{max}) - \frac{z}{D}\right) \tag{6}$$

with T_{av}: mean annual air temperature, A_y: annual amplitude of air temperature, z: mean depth of virtual layer, t_{max}: time of maximum air temperature and $\omega = 2\pi/\tau$: angular velocity of cosine function, while damping depth D is a function of thermal conductivity λ and volumetric heat capacity C_v

$$d_D = \sqrt{2\lambda/C_v\omega} \tag{7}$$

The required air temperature dependent parameters have to be initialised and are dynamically updated at the end of each hydrological year from input weather data to account for annual differences in air temperature.

To include a realistic simulation of winter temperatures and soil freezing, the potentially storable latent energy $LE_{pot,j}$ [J] of the soil water in each layer is computed as soon as the layer temperature drops below the freezing point. $LE_{pot,j}$ of a given layer j acts as a buffer before the soil layer temperature T_j further diminishes. Because some of the water in a soil matrix is influenced by freezing point depression, we included an empirical relationship between liquid and frozen water against the difference between soil and freezing point temperature $\check{T}_j$ derived from the laboratory findings of Watanabe & Mizoguchi (2002). Because the logarithmic curve is extremely steep for very small $\check{T}_j$ we assume soil freezing starts at $\check{T}_j$ = 0.1 K. Using this empirical approximation, about 70% of the water is frozen at $\check{T}_j$ = 1 K and about 90% at $\check{T}_j$ = 6 K. If $\check{T}_j = (273.15\text{ K} - T_j) > 0.1$ K, the volumetric frozen soil water content $\theta_{ice,j}$ is related to total soil water content θ_j by

$$\Theta_{ice,j}(\check{T}_j) = (0.7 + 0.11 \cdot \ln(\check{T}_j)) \cdot \Theta_j \tag{8}$$

So the potential latent energy $LE_{pot,j}$ is dependent on the actual temperature T_j

$$LE_{pot,j}(T_j) = \Theta_{ice,j}(\check{T}_j) \cdot d_j \cdot 1\text{m}^2 \cdot 334{,}000 \frac{\text{J}}{\text{kg}} \cdot 1{,}000 \frac{\text{kg}}{\text{m}^3} \tag{9}$$

For a given layer j: if $0 \leq LE_{act,j}(t_0) < LE_{pot,j}$ then

$$LE_{act,j}(t_0 + \Delta t) = LE_{act,j}(t_0) - G_{j,j+1} \cdot 3{,}600\text{s} \tag{10}$$

If the soil layer temperature later rises above 273.05 K the procedure is repeated until the stored latent energy is consumed ($LE_{act,j} = 0$). Vertical soil water movement in an frozen soil is impeded by a reduction of hydraulic conductivity like in Lundin (1990) and for soil layers with minimal liquid water or air content, infiltrating water is split up between lateral runoff and macropore percolation by an empirical parametrisation based on literature (as in Bayard et al. 2005, Stähli et al. 1996 and others).

2.4 The energy balance algorithm

For the computation of the energy balance terms on the ground surface an iteration of the surface temperature ($T_{surface}$) is done until incoming radiation (R_{in}) equals the outgoing terms latent (LE) and sensible heat flux (H), outgoing radiation (R_{out}) and ground heat flux (G), plus an error term of ± 5 W/m² to reduce computational workload (figure 2).

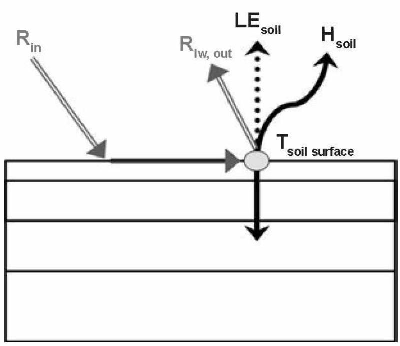

Figure 2: Representation of the energy fluxes at the soil surface.

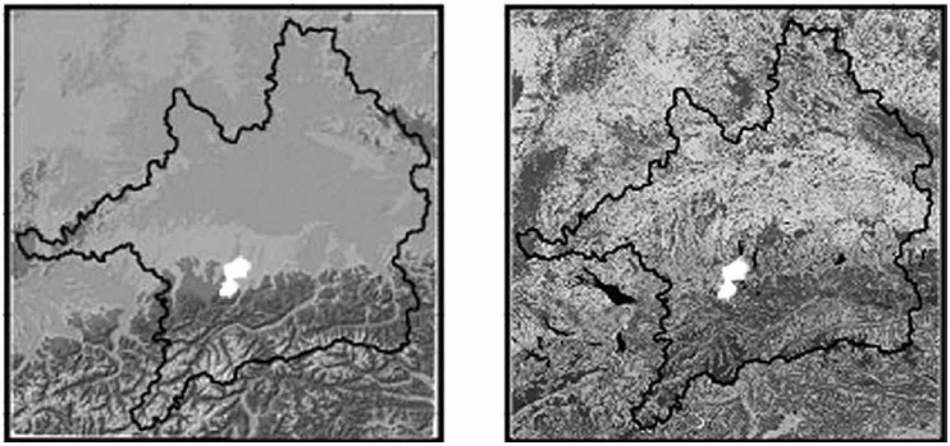

Figure 3: The Ammer subcatchment within the Upper Danube basin.

3 The test site

The Upper Danube catchment has a heterogeneous physiogeographic characteristic with the Alpine terrain covering about one third of its total area (~77,000 km^2). One of its largely unregulated tributaries is the Ammer (figure 3).

At the gauge "Weiheim" the Ammer (A ~ 709 km^2) has a mean discharge of 15.5 m^3/s. It is a typical river crossing from the Alps to the Alpine foreland and one of the few rivers not regulated by technical means. Winter high waters occur mainly in times of snow melt events, sometimes accompanied by heavy rain.

3.1 First results

All year temperature time series have been validated at 27 agrometeorological stations of the Bavarian State Research Center for Agriculture and at 15 measurement sites of the German Weather Services DWD. The Root Mean Square Error (RMSE) between simulated and measured daily soil temperature values was found to be lower than 2°C (Muerth 2008). Periods of soil temperatures below 0°C were reproduced well, despite unknown measurement conditions and model simplifications. This led to the implementation of a simple empirical algorithm to represent changes in runoff generation due to soil ice. Figure 4 shows for a ten year time series, that the model with soil ice simulates the mean monthly maximum better then the model without it.

Similarily, the observed frequencies of certain daily runoff values during times of high flow are better reproduced by the soil model with lateral runoff due to soil ice blocking (figure 5). Extreme high flows (> 50 m^3/s) are still underestimated by the combined soil water and energy model, but some winterly high flow events are apparently reproduced by the reduced infiltration capacity caused by soil ice in near-surface soil layers.

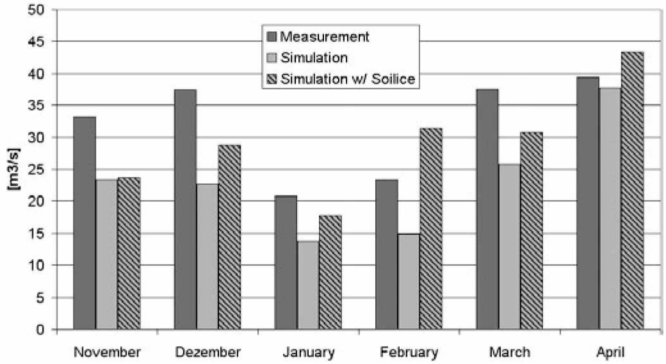

Figure 4: Simulated vs. measured monthly mean maxima of daily runoff at gauge Weilheim (1991–2000).

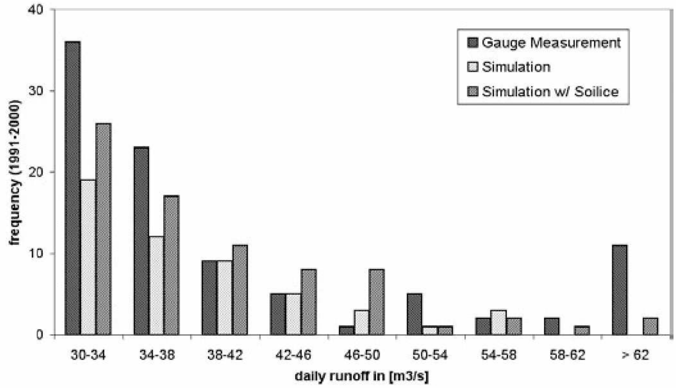

Figure 5: Simulated vs. measured frequencies of daily runoff at gauge Weilheim during the winters of 1991–2000.

For a scenario run the stochastic climate generator (Mauser & Muerth 2008) generated one possible weather data set for the years 2041 to 2050 based on the A1B emission scenario (IPCC 2001) with a fit on the recent regional temperature trend. Measurements show a temperature increase for the Upper Danube basin that was nearly two-fold the global increase from 1970 to 2006. As a consequence, the mean annual temperature rise in the Upper Danube basin was assumed to be ~2.5 °C until 2050 in this first scenario run. The majority of events of runoff on frozen ground shifted from March to February, with some extreme snow melt/precipitation events still occurring in April. As figure 6 shows, the mean monthly maximum of daily runoff had decreased for all months except for March in the scenario run. This run produced only slightly increased precipitation during the winter months, but increasingly dry summer months, resulting in a lower mean annual runoff. But as depicted in figure 7, the reduced water storage in form of snow and ice led to considerable changes in runoff during the winter months. Mean runoff in February was notably higher in the scenario decade than in the validation period, but the frequency of high flows in April due to snow melt were greatly reduced.

Regardless of the input data, the regional warming trend will lead to shorter retention times of water as snow and ice. The combination of less summer precipita-

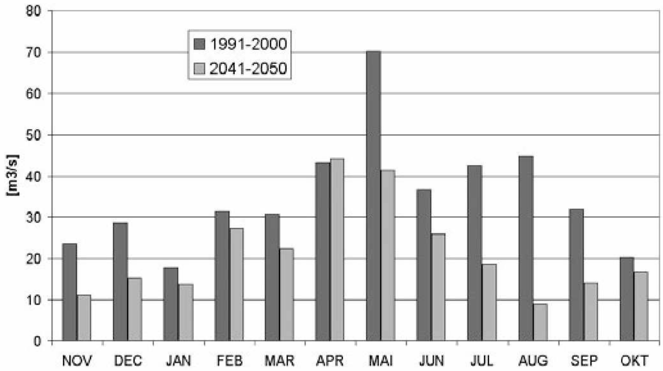

Figure 6: Comparison of mean monthly maximum runoff at gauge Weilheim computed for the validation and the scenario decade.

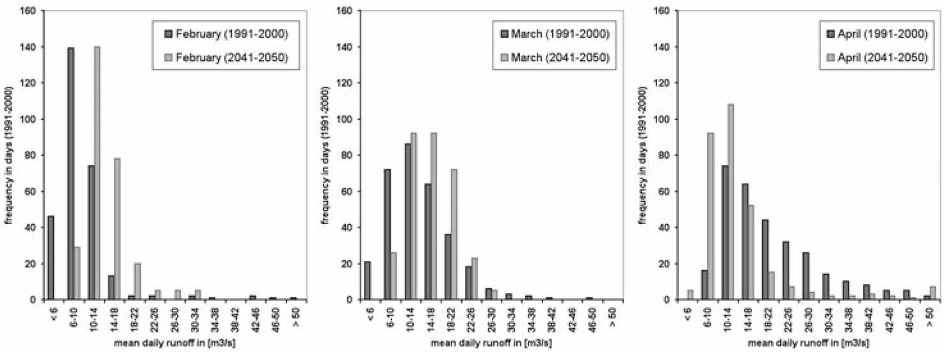

Figure 7: Simulated daily runoff frequencies Weilheim in late winter at gauge Weilheim for the validation and the scenario decades.

tion and early snow melt could lead to lower water levels in summer. In this case an estimation of the effects of soil ice on runoff generation could be helpful to better quantify future amounts of water recharging the aquifers or running off laterally in winter.

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