Runoff parameterization in a SVAT scheme: sensitivity tests with the model SEWAB

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Abstract A Soil-Vegetation-Atmosphere-Transfer (SVAT) scheme, which solves the Surface Energy and Water Balance (SEWAB) equations considering partly vegetated surfaces, has been developed. SEWAB is based on the one-layer-concept for vegetation cover. Within the soil column the vertical diffusion equations for soil temperature and soil moisture are solved. The impact of the subsurface runoff parameterization and the inclusion of a variable infiltration capacity are investigated. Modifications after the ARNO model conceptualization, in combination with a variable infiltration capacity, proved to be most suitable. In SEWAB precipitation is partitioned into runoff, evapotranspiration and soil moisture and is therefore appropriate when linking atmospheric and hydrological models.

NOTATION

A	fraction of the grid cell/basin where the infiltration capacity is less than i_0 (-)
c	volumetric heat capacity (J m^{-3} K^{-1})
D, D_s	hydraulic diffusivity, saturated hydraulic diffusivity (m^2 s^{-1})
E	total evapotranspiration (mm s^{-1})
E_g	evaporation from bare soil (mm s^{-1})
i, i_{max}, i_0	infiltration capacity, maximum infiltration capacity, infiltration capacity at beginning of time step (mm)
J	layer-index (-)
K, K_s	hydraulic conductivity, saturated hydraulic conductivity (m s^{-1})
P	precipitation (mm s^{-1})
Q_{lat}, Q_{sens}, Q_{soil}	latent heat flux, sensible heat flux, soil heat flux (W m^{-2})
Q_{rad}	radiation flux density (W m^{-2})
R, R_j	total runoff, runoff from j-th soil layer (mm s^{-1})
ΔS	soil moisture storage change (mm s^{-1})
S_η	sink term for soil moisture, e.g. due to transpiration from the root zone (s^{-1})
T	soil temperature (K)
Δt	time step (s)
Veg	vegetated fraction of the grid cell (-)
W_s	fraction of saturated soil moisture (-)
z, Δz_j	depth, thickness of jth soil layer (m)

β	shape parameter for the area under consideration (-)
λ	thermal conductivity (W m^{-1} K^{-1})
η, η_s	soil moisture content, saturated soil moisture content (m^3 m^{-3})
ρ_w	density of water (kg m^{-2})

THE SVAT SCHEME SEWAB

Global Circulation Models (GCM) and mesoscale atmospheric models usually have a resolution of 10–110 km. The interaction of the atmosphere with the land surface is not resolved by those scales. This implies a parameterization of the land surface within the atmospheric models. Such parameterizations are commonly one-dimensional (vertical) SVAT schemes, e.g. SEWAB which solves the coupled system of the surface energy and water balance equations. Applying such a one-dimensional SVAT scheme on a grid cell of an atmospheric model requires some kind of regionalization. Therefore interpolated (meteorological forcing) and aggregated (vegetation and soil) data were used to run the model.

SEWAB

SEWAB solves the surface energy balance equation:

$$Q_{rad} + Q_{lat} + Q_{sens} - Q_{soil} = 0 \tag{1}$$

as well as the water balance equation:

$$P - E - R = \Delta S \tag{2}$$

Following Noilhan & Planton (1989) the evapotranspiration is calculated separately for bare soil and vegetated surfaces. The evapotranspiration from the vegetated part is calculated after Deardorff (1978).

Within the soil column of a variable number of layers, the vertical diffusion equations of soil temperature and soil moisture content:

$$c(\eta)\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(\lambda\frac{\partial T}{\partial z}\right) \tag{3}$$

$$\frac{\partial \eta}{\partial t} = \frac{\partial}{\partial z}\left(D(\eta)\frac{\partial \eta}{\partial z} + K(\eta)\right) - S_{\eta} \tag{4}$$

are solved semi-implicitly. $D(\eta)$ and $K(\eta)$ are parameterized after Clapp & Hornberger (1978). For the soil temperature, equation (3), the upper boundary condition is given by the surface energy balance, equation (1). At the lower boundary soil temperature is prescribed by a time series representing the seasonal cycle. The boundary conditions for the soil moisture are taken following Abramopoulos *et al.* (1988), i.e. the upper boundary condition for equation (4) is determined by throughfall, evaporation and soil moisture. With the maximum infiltration capacity:

$$i_{max} = \left[D_s\left(\frac{\eta_1 - \eta_s}{\Delta z_1}\right) + K_s\right]\rho_w \Delta t \tag{5}$$

the upper boundary condition is given by:

$$D(\eta_1)\left.\frac{\partial\eta}{\partial z}\right|_{\eta_1} + K(\eta_1) = \min\left((1-veg)P + RRLD - E_g, i_{\max}\right) \tag{6}$$

and:

$$R_1 = \max\left((1-veg)P + RRLD - E_g - i_{\max}, 0.\right) \tag{7}$$

If N is the number of layers, the lower boundary condition is specified assuming no diffusion between layer N and N + 1 and saturated soil moisture in layer N + 1.
In this basic version runoff from layer $j > 1$ (interflow) is:

$$R_j = \max\left(\frac{\eta_j \Delta z_j \rho_w}{\Delta t} - \frac{\eta_s \Delta z_j \rho_w}{\Delta t}, 0.\right) \tag{8}$$

with $\eta_j = \min(\eta_j, \eta_s)$, i.e. runoff is only produced from saturated soil. Water draining into the saturated layer N + 1 is assumed to be baseflow and added to the total runoff.

EXPERIMENT

Data

SEWAB participated in the Project for Intercomparison of Land-surface Parameterization Schemes (PILPS) (2c) where runoff modelled with SVAT schemes was compared with measured streamflow (e.g. Wood *et al.*, 1998). The only component of the hydrological cycle that can be measured with high accuracy, at the scales under consideration, is streamflow. Comparison of modelled with measured streamflow data shows the ability of the schemes to partition precipitation into runoff on the one hand and the sum of evapotranspiration and soil moisture storage on the other.

For experiment PILPS(2c), land-surface characteristics and meteorological and hydrological data were provided for the Arkansas Red River basin (USA). Vegetation data were taken from ISLSCP CD-ROM. For the ISLSCP data set a 1° × 1° grid was laid over the source data field (satellite data) and area-weighted averages of values falling within each new grid square were calculated. A detailed description is provided by Sellers *et al.* (1995). Monthly leaf area index, roughness length, albedo and zero-plane displacement height were obtained by averaging corresponding values of 1987 and 1988. The US Department of Agriculture Soil Conservation Service (SCS) created the STATSGO (State Soil Geographic) data from soil surveys with the US Geological Survey 1:250 000 scale topographic quadrangles as base maps. Using GRASS, a public domain GIS developed by the US Army/CERL, the soil data were aggregated to 1° × 1° resolution for PILPS(2c). From January 1979 to December 1987, meteorological forcing data were provided on a 30 min timestep. They were interpolated from hourly data from NCDC Surface Airways stations. Daily precipitation data were obtained from NCDC stations and interpolated uniformly to 30 min intervals. Among those were the data for the semiarid Black Bear Creek basin of 1491 km^2 (36.34°N, –96.80°E). The Black Bear Creek is a stream of 38 km length

Table 1 Land-surface characteristics of the Black Bear River.

Parameter	Black Bear River
Vegetation	Mixed forest and woodland
Soil depth (m)	1
Root depth (90% of roots) (m)	0.3
Saturated soil moisture η_s	0.48
Saturated hydraulic conductivity K_s (m s^{-1})	2.64E–6
Saturated suction head Ψ_s (m)	–0.479
Sand (%)	23
Clay (%)	19
Clapp and Hornberger parameter b	4.55

contributing to the Arkansas River at Pawnee. Table 1 shows the land-surface characteristics of this basin. During the period from 1979–1987 most rain fell for about 6 weeks in spring while the rest of the year was rather dry. Figure 1 shows the daily precipitation exemplary for 4 May–13 June 1982. Additionally, measured daily streamflow data were given for comparison with runoff data produced by the SVAT scheme (Fig. 2).

Saturation runoff excess parameterization (run 1)

For PILPS(2c) SEWAB was run in its basic version as described above with five soil layers and a 30 min timestep (run 1). Here, the water drains fully to the lowest layer where all runoff is produced. There is no surface runoff production.

Figure 2 shows the daily sum of the runoff calculated with SEWAB and the measured daily streamflow example for 4 May–13 June 1982. During this period the runoff parameterizations (run 1 – run 3) showed their differences most clearly. No runoff was calculated before the 15 May, though there were precipitation events and streamflow was measured from the 28 April onwards. The precipitation of 50 mm on the 12 May and 20 mm the following day caused a peak in the measured streamflow. The strongest precipitation event occurred on the 17 May. The calculated runoff reached its peak on 19 May, i.e. the runoff was calculated with a delay of up to several days. Therefore the following decay of the calculated runoff is too smooth, although it roughly follows the qualitative pattern of the measured streamflow until the 13 June albeit it is too high.

For the Black Bear Creek basin the mean absolute difference between monthly simulated and observed runoff, normalized by the observed monthly mean flow, was compared in PILPS(2c). With a value of 0.62 SEWAB is well within the range of 0.3 and 1.2 of the 16 participating SVAT schemes.

The annually calculated and measured streamflow (Table 2) compare fairly well. In 1980, 1982 and 1985 the calculated annual runoff fits the measured values best; in these years the precipitation occurred quite steadily for two months. In 1986 strong precipitation occurred from the end of September to December starting with 170 mm on the first day and causing strong overestimation in the model. Although similar amounts of rainfall occurred in 1981 and 1982, streamflow in 1982 was seven times higher than in 1981. This is reproduced by SEWAB.

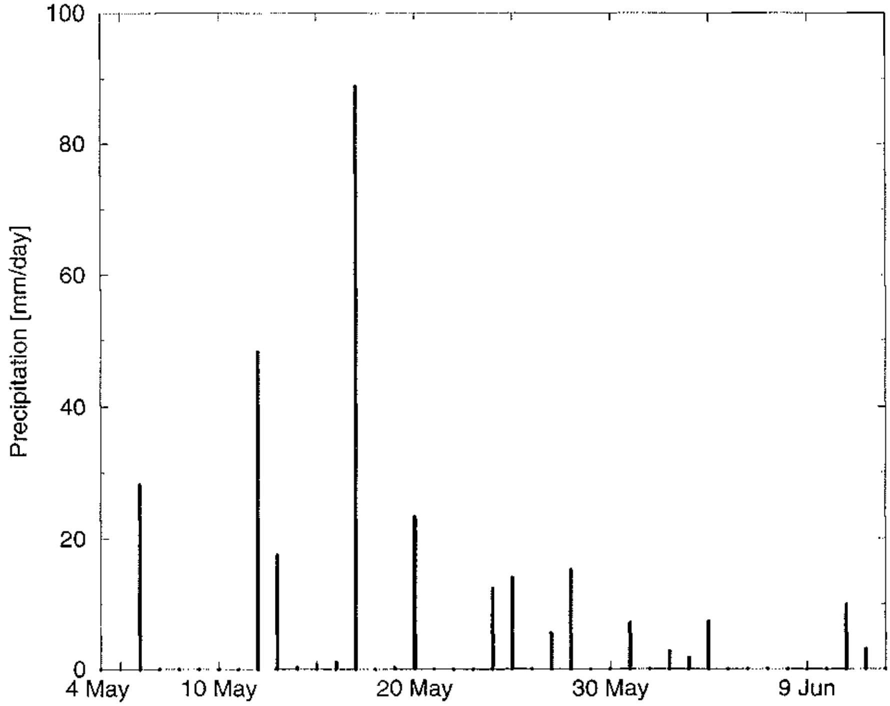

Fig. 1 Daily precipitation in the Black Bear River basin in 1982 (bar chart).

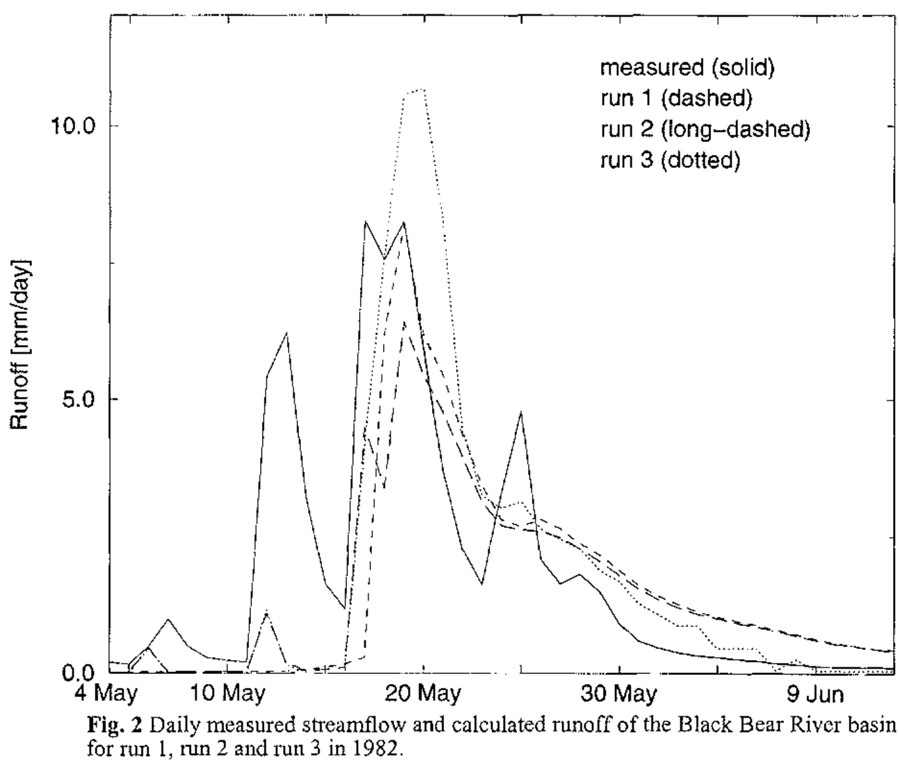

Fig. 2 Daily measured streamflow and calculated runoff of the Black Bear River basin for run 1, run 2 and run 3 in 1982.

Effect of a variable infiltration capacity (run 2)

Experience from PILPS(2c) resulted in a modification of SEWAB by implementing a variable infiltration capacity after the Nanjing model described by Wood *et al.* (1992). The assumption is a spatial variation of the infiltration capacity with:

$$i = i_{max}\,(1 - (1 - A)^{1/\beta}) \tag{9}$$

over a grid cell or basin due to heterogeneity in topography, soil, vegetation and precipitation within the area. Depending on precipitation and initial soil moisture a saturated fraction of the area is assumed, that allows surface runoff production as a result of a precipitation event even though the total area is not saturated, i.e. in case of $(i_0 + P) < i_{max}$ surface runoff is:

$$R_1 = P + \frac{\eta_1 \Delta z_1 \rho_w}{\Delta t} - \frac{\eta_s \Delta z_1 \rho_w}{\Delta t}\left[1 - \left[\frac{i_{\max} - (i_0 + P)}{i_{\max}}\right]^{(1+\beta)}\right] \tag{10}$$

Otherwise equation (7) applies.

For the Black Bear Creek basin β is 0.02. The variable infiltration capacity has an effect on the timing of the runoff events. Figure 2 shows that there is little runoff production on 6 and 12 May, i.e. when measured streamflow occurs, though the amount is underestimated. The two peaks on 16 and 18 May and the low on the 19 May, are met qualitatively, but the runoff is underestimated. Afterwards the calculated runoff is dominated by subsurface runoff and follows the calculated curve from run 1.

Table 2 shows that there is hardly any difference between the annual runoff of this run and run 1. The inclusion had a positive effect on the annual runoff in 1981, where the precipitation events caused hardly any runoff.

Table 2 Annual runoff of the Black Bear River.

Year	Precipitation (mm year^{-1})	Measured runoff (mm year^{-1})	run 1* runoff (mm year^{-1})	run 2* runoff (mm year^{-1})	run 3* runoff (mm year^{-1})
1980	914	144	149	161	161
1981	781	15	6	14	14
1982	753	101	93	93	93
1983	890	111	171	174	171
1984	653	87	46	49	53
1985	1016	188	196	196	188
1986	1314	348	448	451	448
1987	1000	234	155	162	163

*run 1: saturation runoff excess.
run 2: run 1 with variable infiltration capacity.
run 3: baseflow parameterization and variable infiltration capacity.

Effect of a subsurface runoff modification (run 3)

Since there was no improvement in explanation over the previous run, run 2 was repeated with the calculation of the subsurface runoff being modified after the ARNO model conceptualization (e.g. Francini & Pacciani, 1991). This scheme (run 3) assumes that below a limiting factor, runoff is produced linearly depending on the soil moisture content, and above this value increases nonlinearly.

$$R_N = k_1 \eta_N \Delta z_N \rho_w + \max\left(k_2\left((\eta_N - W_s \eta_s)\Delta z_N \rho_w\right)^{k_3}, 0.\right) \tag{11}$$

For the Black Bear Creek, calibration resulted in k_1 2.6×10^{-6} s^{-1}, k_2 0 s^{-1}, k_3 0 and W_s 1. Run 3 results in a higher runoff production from 18–27 May than in the previous runs, but less runoff from then onwards. The runoff is usually delayed by two days compared to the measured streamflow. The mean annual runoff difference to measured streamflow is slightly reduced by this parameterization.

CONCLUSION

At present SEWAB provides the surface fluxes as the lower boundary condition for an atmospheric model. In such models the soil for an area of about 100 km^2 to 11 000 km^2 is considered as homogeneous and therefore treated as one soil column. Here SEWAB was applied to the Black Bear River basin treating the whole area of 1491 km^2 as one soil column. SEWAB is able to calculate the annual runoff fairly well under changing meteorological conditions (1981 and 1982). With a correlation coefficient of 0.72 between calculated and measured daily runoff over a period of 8 years (1980–1987) in a semi-arid basin, its performance is reasonable for application in a mesoscale atmospheric model.

The inclusion of a variable infiltration capacity allows surface runoff production and results in a slight improvement regarding the timing of the runoff calculation. Allowing additional subsurface runoff before the soil is saturated reduced the difference between the calculated runoff and the measured streamflow not only annually but also on a daily basis.

The SVAT scheme SEWAB has proved to be able to partition the precipitation into runoff, evapotranspiration and soil moisture storage change. It is therefore a suitable land surface parameterization to be used as a link between atmospheric and hydrological models as is planned in the Baltic Sea Experiment BALTEX, a continental scale experiment under GEWEX (BALTEX, 1995). Usually a river basin is covered by several grid boxes of an atmospheric model. Coupled to an atmospheric model SEWAB calculates the local runoff for each grid cell of the atmospheric model. Using these values, applying a horizontal routing model for channel delivery, and channel routing (e.g. Lohmann *et al.*, 1996) to the area under consideration, results in runoff data that can be compared with measured streamflow data at gauging stations as done e.g. by Lohmann (1996) for the Weser basin (Germany) and in PILPS(2c) for the Arkansas Red River basin (USA).

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