SEWAB – a parameterization of the Surface Energy and Water Balance for atmospheric and hydrologic models

Heinz-Theo Mengelkamp *, Kirsten Warrach, Ehrhard Raschke

GKSS Research Center, Institute for Atmospheric Physics, Max-Planck-Str., D-21502 Geesthacht, Germany

Received 29 September 1997; received in revised form 10 August 1998; accepted 6 April 1999

Abstract

A soil-vegetation-atmosphere-transfer scheme, which solves the coupled system of the Surface Energy and Water Balance (SEWAB) equations considering partly vegetated surfaces, is presented. In terms of complexity, SEWAB is similar to many other land–surface schemes with respect to the calculation of the turbulent fluxes of sensible and latent heat the latter being based on the one-layer concept for vegetation. Emphasis is put on the description of the soil processes as the link between the atmospheric and hydrologic system. The diffusion equations for heat and moisture are solved semi-implicitly on a multi-layer grid. Surface runoff and baseflow may be calculated as saturation excess runoff as usually done in land–surface schemes for atmospheric models. In addition to this, the variable infiltration capacity (VIC) approach is included alternatively which takes into account the influence of topographic heterogeneities inside a grid cell on surface runoff prediction. Subsurface runoff may also be described by the ARNO conceptualization allowing a gradual increase with soil moisture content. The saturation hydraulic conductivity is a function of depth. SEWAB has been validated with field data from the FIFE experiment and has participated in the PILPS project for intercomparison of land–surface parameterization schemes. SEWAB partitions reasonably well the incoming solar radiation and the precipitation into sensible and latent heat fluxes as well as into runoff and soil moisture storage. The inclusion of a variable infiltration capacity description slightly improves the surface runoff estimation and the timing of the total runoff. Changes in the parameterization of the subsurface runoff production and the drainage show only minor effects. © 1999 Elsevier Science Ltd. All rights reserved.

Keywords: Land–surface scheme; Evapotranspiration; Runoff

1. Introduction

Atmospheric processes on all spatial and temporal scales are sensitive to variations in land–surface properties and soil characteristics. The exchange of momentum, heat and moisture between the earth’s surface and the atmosphere is controlled by the vegetation and soil type, among other factors. An appropriate description of these transfer processes is therefore essential for atmospheric simulation models. A Soil-Vegetation-Atmosphere-Transfer (SVAT) scheme solves the coupled system of the surface energy and water balance equations. The latter partitions the precipitation into evapotranspiration, soil moisture storage and horizontal runoff which may be further separated in fast overland flow and slow subsurface runoff. The amount of local runoff is a key input parameter for models determining the runoff in rivers, as in use to forecast floods.

Various parameterizations for the Soil-Vegetation-Atmosphere-Transfer have been developed during the last decade and were implemented in weather forecast and climate models. They basically follow principles outlined by Deardorff [4] but treat the particular components of the soil-vegetation system with different complexity. This concerns the description of physical, biophysical and physiological processes as well as the numerical approaches and the number of layers in the soil and the vegetation. The Simple Biosphere Model (SiB) of Sellers et al. [29] includes three soil layers and two vegetation layers. In many mesoscale models, the vegetation is commonly represented by a single layer, which may partly cover the model grid box (e.g. [24]). The treatment of heat and water in the soil ranges from the simple bucket model by Manabe [20] to multi-layer soil models ([30]). The appropriate level of complexity of the different components of SVATs for particular
applications on different time and spatial scales is still under discussion.

In order to compare land–surface schemes with one another and with observations, the Project for the Intercomparison of Land–Surface Parameterization Schemes (PILPS) was initiated as a research activity in the framework of the World Climate Research Program ([11]). Participation in two phases of this intercomparison study proved the reliability of the parameterization scheme SEWAB and led to substantial improvements particularly concerning the soil water treatment.

In this paper, the model SEWAB is described and its performance within the PILPS project is discussed. The parameterization of the surface energy and water balance is described in Section 2, the soil model in Section 3, and Section 4 describes the program structure. We show some validation examples with data from the FIFE experiment and the PILPS intercomparison project in Section 5. The sensitivity of the runoff production to formulations of the infiltration, the baseflow runoff and the drainage is discussed in Section 6.

2. The surface energy and water balance

2.1. Balance equations

The radiation fluxes are the most important of the various kinds of energy the earth’s surface receives. The shortwave solar radiation that reaches the surface after some atmospheric alterations heats the ground or is reflected. Thermal infrared radiation is emitted from both the surface and the atmosphere, with a net loss to the surface since the ground is usually warmer than the atmosphere above. The net radiative energy is used to evaporate water (latent heat), to heat directly the atmosphere and to heat the deeper layers of the soil. Below the surface heat diffusion takes place. Fig. 1 illustrates the main processes of the energy and water exchange at the surface.

The latent heat flux couples the surface energy with the water balance. If rain reaches the ground the water may be stored on the surface, infiltrate into the soil or be transported horizontally as surface runoff. These processes mainly depend on soil type and the total amount of water already stored in the soil column. Below the surface the water may reach greater depths through gravitational drainage and diffusion processes and may form interflow or groundwater runoff.

The presence of vegetation influences the transfer processes at the surface in different ways. Increased surface roughness enhances turbulent mixing and consequently the transfer of heat and water. Absorption and reflection of solar radiation depend on the physiological characteristics of the vegetation cover and the ability of plants to transfer water from the root zone to the leaves controls the latent heat flux. Over forested areas the interception of rainfall and subsequent evaporation from wet leaves may even prevent water reaching the ground, depending on rainfall intensity and interception storage capacity.

SEWAB does not currently treat snowfall and frost explicitly both of which significantly influence the soil–atmosphere exchange in Northern Latitudes. A slight modification of the roughness and optical properties of the ground surface accounts for snow cover.

Assuming that no heat and water are stored at the surface, the energy balance equation may be written as

$$Q_{\text{rad}} + Q_{\text{sens}} + Q_{\text{lat}} - Q_{\text{soil}} = 0,$$  \hspace{1cm} (1)

and the corresponding surface water balance during rainfall events

$$P - \frac{Q_{\text{lat}}}{L_v} - I = -R.$$  \hspace{1cm} (2)

Between rainfall events $P$, $I$ and $R$ in Eq. (2) vanish because we do not allow water storage at the surface (e.g. ponding is not included). $Q_{\text{rad}}$ is the radiation flux density, $Q_{\text{sens}}$ and $Q_{\text{lat}}$ are the sensible and latent heat fluxes, respectively, and $Q_{\text{soil}}$ is the soil heat flux. $L_v$ is the latent heat of vaporization, $P$ precipitation and $R$ the horizontal surface water runoff. The infiltration $I$ represents that amount of water which directly infiltrates the soil. Upward fluxes are positive, downward fluxes negative throughout this paper. Symbols and units are listed in Appendix A.

Eq. (1) is solved for the surface temperature $T_g$ iteratively by Brent’s method ([26]). From our experience
this method is superior in terms of convergence to other methods like the false position/regular falsi approach or a Newton-Raphson iteration scheme. This holds in particular for extreme situations like very dry soils and the development of an evaporation barrier and for rapidly changing situations e.g. a change of solar radiation due to cloud development or the beginning of a rainfall event which are associated with rapid changes in surface temperature.

With \( Q_{\text{lat}} \) from Eq. (1), surface runoff \( R \) is calculated with \( P \) given by the forcing data and the infiltration \( I \) as described in Section 2.4.

### 2.2. Radiation, sensible heat and soil heat flux densities

The radiation flux density \( Q_{\text{rad}} \) is written as

\[
Q_{\text{rad}} = -R_t(1 - a) - \varepsilon_a R_t + \varepsilon a T_g^4,
\]

where \( a \) is the albedo and \( \varepsilon \) the surface emissivity. The shortwave radiation \( R_t \) reaching the surface depends on the radiation for a cloudless sky \( R_{d0} \) and a factor for cloudiness \( N \) which was derived by Kasten and Czeplak [14] for a weather station at the North Sea coast \( R_t = R_{d0}(1 - 0.75N^{3.4}) \). This formulation is used when SEWAB is coupled to an atmospheric model. For the validation exercises reported here, the measured downward shortwave radiation was prescribed as part of the forcing data. The downward longwave radiation flux emitted from the atmosphere, \( R_l \), is parameterized after Deardorff [4]. \( T_g \) is the surface temperature and \( \varepsilon_a \) an effective atmospheric emissivity.

The turbulent flux of sensible heat is calculated by a bulk-aerodynamic formula

\[
Q_{\text{sens}} = -\rho c_p C_{H1} u_a (\Theta_a - \Theta_g)
\]

with air density \( \rho \), specific heat of air \( c_p \), drag coefficient \( C_{H1} \) after Louis [18] and windspeed \( u_a \). \( \Theta_a \) denotes air potential temperature, \( \Theta_g \) surface potential temperature.

The soil heat flux between the surface and the very thin first soil layer (of order 2 cm) underneath is given by

\[
Q_{\text{soil}} = -\lambda \left( \frac{\partial T_g}{\partial z} \right) \approx \lambda \frac{T_g - T_1}{\Delta z_1}
\]

with the thermal conductivity \( \lambda \) being either dependent on soil moisture (e.g. \([21,25]\) for soils with a high organic content) or assigned a constant value. \( T_1 \) is the temperature of the first soil layer, \( T_g \) the surface temperature. \( \Delta z_1 \) denotes the thickness of the first soil layer which typically is 2 cm.

No change in temperature is assumed for water bodies. In this case, \( Q_{\text{soil}} \) is the residual of the surface energy balance \( Q_{\text{soil}} = Q_{\text{rad}} + Q_{\text{sens}} + Q_{\text{lat}} \). This assumption seems to be justified if relatively short time periods were considered due to the high heat capacity of water and its mixing ability.

### 2.3. Evapotranspiration

The surface water balance represents the partitioning of precipitation into evapotranspiration, soil moisture storage, infiltration and surface runoff. Evapotranspiration and interception are described in this section, runoff and infiltration in Section 2.4 and the soil moisture storage is treated in Section 3.

The description of the evapotranspiration follows in large parts the Interaction Soil Biosphere Atmosphere model (ISBA) of the French Weather Service ([24]). The flux of latent heat \( Q_{\text{lat}} = L \dot{E} \) is described by bulk formulæ for the surface moisture flux \( \dot{E} \) which is the sum of the moisture flux from bare surfaces \( \dot{E}_g \) and of evapotranspiration from vegetated surfaces \( \dot{E}_v \)

\[
\dot{E} = (1 - \text{veg}) \dot{E}_g + \text{veg} \dot{E}_v,
\]

veg is the fraction of a grid square covered by vegetation.

The evaporation from bare soils is

\[
\dot{E}_g = C_Q \rho u_a \omega_q(T_g - q_a).
\]

The drag coefficient \( C_Q \) equals that for sensible heat \( C_{H1} \). \( q_a(T_g) \) is the saturation specific humidity at surface temperature \( T_g \). \( q_a \) is the air specific humidity.

The relative humidity \( z \) at the ground surface has been the subject of different investigations. Refs. [23,19] list various formulations for \( z \) which either depend on the surface suction head or are based on the concept of the field capacity. We have implemented a formulation according to Ref. [24].

\[
z = \left\{ \begin{array}{ll}
\frac{1}{2} & \text{if } \eta_1 < \eta_{lc} \\
1 & \text{if } \eta_1 \geq \eta_{lc}
\end{array} \right.
\]

This expression requires information about the surface volumetric water content \( \eta_1 \) which is deduced from the soil model. The field capacity is often related to the saturation water content \( \eta_{lc} \approx 0.75\eta_g \) (e.g. [24]). We followed a suggestion by Lee and Pielke [15] who related the field capacity to a hydraulic conductivity of 0.1 mm per day although there are some indications that this value is exceptionally small and may well be orders of magnitude larger for particular soil types.

The evapotranspiration from the vegetated part consists of the evaporation from the wet foliage \( \dot{E}_v \) and the transpiration from the remaining dry part \( \dot{E}_{\text{tr}} \):

\[
\dot{E}_v = \delta_w \dot{E}_v + (1 - \delta_w)\dot{E}_{\text{tr}}.
\]

The wet fraction \( \delta_w \) of the foliage is a power function of the liquid water content of the interception reservoir \([4] \):

\[
\delta_w = \left( \frac{w_d}{w_{\text{max}}} \right)^{2/3}
\]

with the maximum interception reservoir \( w_{\text{max}} = 0.2 \text{ veg LAI} \). LAI is the leaf area index which is defined as the total one-sided leaf area of the foliage relative to
the ground area of the same region. The rate of the interception water content $w_i$ is given by

\[
\frac{\partial w_i}{\partial t} = \text{veg} P - E_t \text{veg} \delta_w
\]  

(11)

for $0 \leq w_i \leq w_{\text{max}}$. If $w_i$ exceeds $w_{\text{max}}$, leaf drip $R_{\text{ld}}$ reaches the ground. Over vegetated areas (veg $> 0$) we have:

\[
R_{\text{ld}} = \frac{w_i - w_{\text{max}}}{\Delta t}.
\]

(12)

The evaporation from the wet foliage is

\[
E_t = C_Q \rho u_a [q_s (T_g) - q_a]
\]

(13)

and the transpiration over the dry part

\[
E_{tr} = \frac{1}{R_a + R_{\text{ST}}} \rho [q_s (T_g) - q_a]
\]

(14)

with the aerodynamic resistance

\[
R_a = \frac{1}{C_Q u_a}.
\]

(15)

The stomatal resistance is formulated after Pinty et al. [27]:

\[
R_{\text{ST}} = R_{\text{ST} \text{min}} \left[ 1 + \left( \frac{\psi_f}{\psi_{\text{fc}}} \right)^{5.5} \right] \left[ 1 + 0.0055 R_s \frac{R_{\text{ST} \text{min}}}{R_{\text{ST} \text{max}}} + 0.0055 R_s \right],
\]

(16)

where $R_{\text{ST} \text{min}}$ and $R_{\text{ST} \text{max}}$ are minimum and maximum stomata resistances which can be determined for particular vegetation types. The bracket term represents a moisture factor for the stomata resistance which limits transpiration when water stress occurs. It is a function of the soil moisture potential at field capacity $\psi_{\text{fc}}$ and a leaf moisture potential $\psi_f$, which depends on the average soil moisture potential in the root zone $\psi_{sr}$, the mean canopy height $h_t$ and a transpiration resistance term which merely accounts for internal resistances to water flow within plants $R_t$ and the resistance to water flow through the soil to the root area $R_s$. The leaf moisture potential is written after Federer [8]

\[
\psi_f = \psi_{sr} - h_t \frac{E_t}{\rho_w} (R_t + R_s).
\]

(17)

The last term of Eq. (17) represents the solar radiation factor of the stomatal resistance ([5]).

2.4. Runoff and infiltration

Two approaches are implemented in SEWAB to estimate infiltration and surface runoff.

The first version corresponds to a saturation excess approach ([7]). Here it is assumed that all the water which reaches the ground surface immediately infiltrates into the uppermost soil layer until saturation. Then if the sum of precipitation $P$ and soil moisture storage in the first layer $W_i = \eta_s \Delta z_i \rho_w / \Delta t$ exceeds the saturation soil moisture storage $W_{\text{max}} = \eta_s \Delta z_i \rho_w / \Delta t$ surface runoff $R$ is produced.

\[
R = P + W_i - W_{\text{max}},
\]

(18)

where the subscript $s$ refers to saturation and $\Delta z_i$ is the thickness of the uppermost soil layer.

As an alternative, the variable infiltration capacity (VIC) method ([32]) is implemented (infiltration excess or Hortonian runoff, [13]). The assumption is a spatial variation of the infiltration capacity with

\[
i = i_m [1 - (1 - A)^{1/b}]
\]

(19)

over a grid cell due to heterogeneity in topography, soil, vegetation and precipitation within the area. $i_m$ is the maximum infiltration capacity within the grid cell or catchment. It represents the maximum depth of water which can be stored in the soil column and depends on the maximum soil moisture content of the uppermost layer. $i_m = \eta_s \Delta z_i \rho_w (1 + \beta)$ with $\Delta z_i$ being the thickness of the uppermost layer and $\beta$ a shape parameter. Dependent on precipitation and initial soil moisture, a saturated fraction $A$ of the area is assumed that allows surface runoff production as a result of a precipitation event although the total area is not saturated. In this case, surface runoff is calculated from

\[
R = P + W_0 - W_{\text{max}} \left[ 1 - \left[ \frac{i_m - (i_0 + P)}{i_m} \right]^{1/b} \right]
\]

(20)

$W_0$ and $W_{\text{max}}$ represent the actual and the maximum soil moisture storage, $i_0$ the initial infiltration capacity and the shape parameter $\beta$ is a calibration coefficient which is a characteristic parameter of the catchment. Ref. [32] varied $\beta$ in the range of 0.01–5.0 while Dümenil and Todini [6] found values within the range of 0.01 and 0.5 for GCM land–surface parameterizations. We have used a value of $\beta = 0.02$ as found from calibration exercises for the Arkansa-Red-River basin.

The number of layers in the soil and their thickness is generally variable. Subsurface runoff in the lowest layer usually between 2 and 3 m depth is calculated after the ARNO model conceptualization ([10]). It is assumed that the baseflow out of the lowest layer increases logarithmically as a function of the soil moisture if this exceeds 90 percent of the saturation value. Runoff for the intermediate layers is assumed as saturation excess runoff.

3. The soil model

When defining a temperature diffusivity $B_t$ in the soil as the relationship of the thermal conductivity and the volumetric heat capacity the temperature diffusion equation is written as

\[
\frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left( B_t \frac{\partial T}{\partial z} \right)
\]

(21)
and the vertical diffusion equation for soil moisture content (liquid plus vapor phase) is the Richards equation ([27])
\[
\frac{\partial \eta}{\partial t} = - \frac{\partial}{\partial z} \left( D_{\eta} \frac{\partial \eta}{\partial z} + K_{\eta} \right) .
\] (22)

Relationships between the soil moisture characteristics \(D_{\eta}, K_{\eta}, \Psi_{\eta}\) and \(\eta\) are given by Clapp and Hornberger [3]:
\[
\Psi_{\eta} = \Psi_s \left( \frac{\eta}{\eta_s} \right)^b ,
\] (23)
\[
K_{\eta} = K_s \left( \frac{\eta}{\eta_s} \right)^{2b+3} ,
\] (24)
\[
D_{\eta} = - \frac{bK_s \Psi_s}{\eta} \left( \frac{\eta}{\eta_s} \right)^{b+3} .
\] (25)

Subscript \(s\) refers to the saturation value and \(b, \Psi_s, \eta_s, K_s\) are listed for 11 soil classes in the USDA (US Department of Agriculture) textural classification (see e.g. Table 1 in [31]). Following a suggestion by Clapp and Hornberger [3] Eq. (23) is modified for conditions \(\eta > \eta_{inf}\) where \(\eta_{inf}\) is a so-called inflection point above which a gradual air entry (as shown by most soils near saturation) is accounted for. The inflection point is assumed at \(\eta_{inf} = 0.92\eta_s\). For \(\eta > \eta_{inf}\) we formulate
\[
\Psi_{\eta} = -m \left( \frac{\eta}{\eta_s} - n \right) \left( \frac{\eta}{\eta_s} - 1 \right) ,
\] (26)
with
\[
m = \frac{\Psi_{inf}}{(1 - \eta_{inf}/\eta_s)} - \frac{\Psi_{inf} b}{\eta_{inf}/(1 - \eta_{inf}/\eta_s)}
\] (27)
and
\[
n = 2 \frac{\eta_{inf}}{\eta_s} - \frac{\Psi_{inf} b}{m(\eta_{inf}/\eta_s)} - 1 .
\] (28)

The functional relationships \(\Psi_{\eta}\) and \(D_{\eta}\) represent the soil moisture characteristic curves. The hysteresis effect, stating that the equilibrium soil wetness at a given suction is greater in the drying than in the wetting phase ([12]), is generally not accounted for in soil water transport models.

Since the soil moisture characteristics (hydraulic diffusivity, conductivity and soil moisture potential) change exponentially with soil moisture and in order to avoid excessively small time steps due to the very thin uppermost layer, Eq. (21) and the diffusion term of Eq. (22) are solved with a semi-implicit method described as the Thomas-algorithm by Richtmyer and Morton [28]. At the upper boundary, the temperature and soil moisture of the uppermost layer are updated every time step according to the soil heat flux and the infiltration rate, respectively. At the lower boundary, the soil temperature is prescribed either as a fixed value for short-time integrations or as time series. Concerning soil moisture, we do not allow diffusion through the lowest layer but gravitational drainage out of it. The gravitational drainage term of Eq. (22)
\[
\frac{\partial \eta}{\partial t} = - \frac{\partial K_{\eta}}{\partial z} .
\] (29)
is calculated explicitly.

4. The program structure

If SEWAB is used in an off-line mode, the time step corresponds to the averaging period of the measured forcing variables (wind speed, temperature, specific humidity, surface pressure, radiation and precipitation) which usually ranges from 10 min to 3 h. In a coupled mode, SEWAB uses the same time step as the atmospheric host model. At any instant of time, both soil and atmospheric fluxes are algebraic functions of the externally imposed meteorological forcing variables and the soil heat and moisture stores. Irrespective of prior history, all fluxes are calculated from the state of the system at any time.

The soil heat and moisture contents are time dependent and their evaluation is given by differential equations (Eqs. (21) and (22)). The change in heat or moisture storage depends on current fluxes and the soil state at the former time step. This distinction between fluxes and storages implicates the overall program structure:

Step 1: Definition of soil and vegetation type. Initialize soil heat and moisture content.

Step 2: Specification of atmospheric forcing variables for time \(t\).

Step 3: Calculate fluxes from iteration of Eq. (1).

Step 4: Calculate infiltration and runoff from Eq. (2).

Step 5: Update soil moisture distribution by removing transpired water from the root zone soil layer, evaporated water from the first layer and the runoff from the respective layers and adding infiltrated water to the first soil layer.

Step 6: Integrate Eqs. (21) and (22).

Step 7: Update soil characteristic curves (Eqs. (23)–(25)) and proceed with step 2 for \(t + \Delta t\).

5. Validation with FIFE and PILPS data

We evaluated SEWAB using data from three field experiments namely FIFE, PILPS2a and PILPS2c.

5.1. FIFE

A data set from the FIFE experiment (First ISLSCP Field Experiment, FIFE special issue [9]) was used to
validate SEWAB with respect to surface fluxes. The data set consists of site averages from 10 meteorological stations having measured forcing data (air temperature, wet bulb temperature, pressure, wind speed, downward shortwave radiation, net longwave radiation, precipitation and cloudiness) as 30-min averages. The flux data were averaged from 17 stations, 6 of them with eddy correlation measurements, the remaining 11 were Bowen ratio stations. The instrumentation was deployed within a $15 \times 15$ km tallgrass prairie area in Kansas. The soil type is silty loam. The soil temperatures were initialized identical to the air temperature whereas the initial soil moisture content ($\eta = 0.5\eta_s$) and the fraction of vegetation ($\text{veg} = 0.95$) were adapted.

Calculated and measured turbulent fluxes were compared for a period of 12 days (30 June–11 July 1987) which was mainly dry with some small showers during June 30 and July 7 and heavy rainfall around noon on July 5. Because the net longwave radiation was measured the last two terms of Eq. (3) were prescribed and one possible major source of uncertainty was eliminated.

Rapidly changing synoptic conditions caused larger differences between observations and simulations on the 8th and 9th day and during the night from the 11th to the 12th day (Fig. 2). Latent and sensible heat fluxes are fairly well predicted with some overestimation of the latent heat flux on July 10 and 11. During these days, relatively high wind speeds and high soil moisture contents prevailed.

The fact that the forcing data and the flux data were not measured at the same locations has to be considered in particular during small scale and rapid variations in the forcing data.

5.2. PILPS2a

The five phases of the Project for Intercomparison of Land–surface Parameterization Schemes (PILPS) are described in Ref. [11]. In phase 2a forcing data from a meteorological tower at Cabauw in the Netherlands were provided on a 30-min interval for 1987 together with a description of the soil and vegetation type. The site is characterized by grassland with narrow ditches and the soil was supposed to be saturated during the whole year. The observed flux data were not provided before the experiment in order to prevent any tuning of the participating SVAT schemes. Ref. [2] summarizes the results of the Cabauw experiment and shows the intercomparison of all participating schemes; here only the performance of SEWAB is discussed. Fig. 3 shows the monthly mean values of the sensible and latent heat flux. There seems to be a slight overestimation of the sensible heat flux during summer while the latent heat flux is underestimated at the same time. Because this underestimation is balanced to some extent during winter the latent and sensible heat fluxes came close to the observations in the yearly mean. Although runoff was not measured at the Cabauw site a rough estimate indicates a slight overestimation of runoff by SEWAB on the average.

5.3. PILPS2c

In PILPS phase 2c (Arkansas river) meteorological forcing data, hydrological data and land–surface characteristics were provided for 61 grid boxes of $1^\circ$ by $1^\circ$ size covering the Arkansas-Red-River basin (USA). The climate ranges from arid in the southwest to humid in the eastern parts. Snow has a relatively small climatological and hydrological effect. The vegetation varies from grassland in the west to deciduous forests in the east. From January 1979 to December 1987 the forcing data were given on a 30 min timestep. Ref. [16] describes this experiment and gives preliminary results.
SEWAB was run with 6 soil layers. Fig. 4 shows the observed mean annual basin runoff and the spatial standard deviation of annual runoff with respect to the 61 grid cells and the corresponding SEWAB results. While the spatial variability seems to be reasonably well reproduced, the total amount of runoff is slightly higher than the observations indicate. A more detailed investigation into the runoff production revealed that in SEWAB no surface runoff was produced and all the runoff was due to the slow subsurface component. The reason for this behaviour is twofold. The basic SEWAB version only produces saturation excess runoff. Also water is transported from the surface to the lower layers quite fast by the diffusion and the drainage process. This caused an underestimation of evapotranspiration and an accumulation of water in the lowest layer. Consequently, compared to observations, too much runoff was produced with a certain time delay after rainfall events.

6. Sensitivity of runoff production to soil moisture processes

The tests described in Section 4 were performed with the basic version of SEWAB in which we assumed saturation excess runoff at all levels and prescribed the same saturation conductivity at all soil layers. In order to investigate the sensitivity of the runoff production the semi-arid Black Bear River basin of 1491 km$^2$ (36.34°N, −96.80°E) was chosen as a test catchment. Its land surface and soil characteristics are given in Table 1.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Characteristic</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vegetation</td>
<td>Mixed forest and woodland</td>
</tr>
<tr>
<td>Soil depth (m)</td>
<td>1</td>
</tr>
<tr>
<td>Root depth (90% of roots) (m)</td>
<td>0.3</td>
</tr>
<tr>
<td>Saturated soil moisture $\eta$</td>
<td>0.48</td>
</tr>
<tr>
<td>Saturated hydraulic conductivity $K_s$ (m/s)</td>
<td>$2.64 \times 10^{-6}$</td>
</tr>
<tr>
<td>Saturated suction head $\Psi_s$ (m)</td>
<td>−0.479</td>
</tr>
<tr>
<td>Sand (%)</td>
<td>23</td>
</tr>
<tr>
<td>Clay (%)</td>
<td>19</td>
</tr>
<tr>
<td>Clapp and Hornberger parameter $b$</td>
<td>4.55</td>
</tr>
</tbody>
</table>
Leaf area index, roughness length, albedo and displacement height were provided monthly. The atmospheric forcing from January 1979 to December 1987 was given on a 30-min timestep. Runoff production was calculated with the basic SEWAB version (run 1) and with inclusion of the variable infiltration capacity (run 2). Additionally, run 2 was repeated with the subsurface runoff parameterization after the ARNO model conceptualization (run 3) and run 3 was repeated with a depth-dependent saturation hydraulic conductivity (run 4). Because a constant saturation hydraulic conductivity might not be justified due to a varying soil density, in run 4 we included a saturation hydraulic conductivity varying exponentially with depth \( z \) after Beven [1]:

\[
K_S = K_{S0} \exp(-fz)
\]

with \( f \) varying between 1 and 13 m\(^{-1}\) according to land characteristics. Here \( f \) is 2 m\(^{-1}\). \( K_{S0} \) is the saturation hydraulic conductivity at the surface.

The runs were performed for the whole 9 yr period. The rainfall-runoff relation varies largely for particular rainfall events. This holds for the observed and the simulated runoff and their interrelationship. Reasons for a non-uniform relationship between simulated and measured runoff are numerous. The size of the catchment and the heterogeneity of the landscape as well as the distribution of soil moisture and its history may limit the accuracy of runoff calculations and measurements as being representative for the whole area. In addition to this, the limited number of rain gages and the variability of rainfall patterns in space and time sometimes do not allow an exact estimate of the amount of rainfall representative for the grid box.

A 40 day period in May and June 1982 is shown in Fig. 5. The daily sums of the measured runoff are shown as a solid line, the SEWAB simulations for run 1 dotted and for run 2 dashed. Although there were precipitation events and streamflow observations before this period, no runoff was calculated before day 11 in run 1. The strongest streamflow was observed on day 13 while the calculated runoff reaches its peak on day 15. The following decay is much slower for the calculated runoff though it roughly follows the measured streamflow pattern. With the inclusion of the variable infiltration capacity (run 2) little runoff is produced on days 2 and 8 at the same time when streamflow was measured but the amount is too small particularly on day 8. On day 12 the calculated and observed runoff start at the same time but the high peak period on days 13 to 15 is underestimated. Afterwards the runoff is dominated by subsurface runoff and follows the curve from run 1. Runs 3 and 4 do not show any substantial difference to run 2 for this particular time period and are not presented.

The mean annual measured runoff of the Black Bear River basin for each of the 8 yr is listed in Table 2 together with the observed precipitation and the calculated runoff. There is hardly any difference between the basic SEWAB version (run 1) and runs 2–4. Overall the measured and calculated streamflow agree quite well concerning the interannual variation and also the absolute amount. In particular, during the years 1981 and 1982 with roughly the same amount of precipitation the large difference in the measured runoff is also reflected in the calculations.

7. Summary and conclusion

We have described the land–surface scheme SEWAB which is intended to serve as the link between atmospheric circulation models and hydrologic models. Basically a one-layer concept accounts for the vegetation while for the soil model a multi-layer approach has been implemented. For the vegetation and the soil surface the same temperature is used to describe the transfer of energy between the surface and the atmosphere. The evapotranspiration is separated in evaporation from bare soil and the wet foliage and the transpiration from the dry part of the vegetation. To describe the transport of heat and water in the soil the heat conduction and the Richards equation are solved semi-implicitly on a multi-layer vertical grid. Surface runoff is calculated either as saturation excess runoff or through the variable infiltration capacity approach. The slow runoff component at the lowest soil layer is described by saturation excess runoff or the ARNO scheme.

In addition to the atmospheric forcing data, either from an atmospheric model or from observations, the scheme requires information about the dominant soil and vegetation type within each grid cell. The secondary parameters describing thermal and hydraulic properties...
of the soil and the physical and physiological properties
of the vegetation and the bare soil are estimated from
field experiments and listed in tables.

Validation of SEWAB with field data from the FIFE
experiment proved that the turbulent sensible and latent
heat flux can quite well be reproduced by adapting the
initial soil moisture content and the fraction of vegeta-
tion in reasonable ranges. Participation in PILPS phase
2a (Cabauw experiment) showed good agreement of
observed and calculated sensible heat flux over a time
period of one year while the latent heat flux is under-
estimated during summer. Experiment PILPS 2c (Ar-
 kansas-Red-River catchment) focused on the runo/C128
production of SVAT schemes. Simulations with the
saturation excess approach showed that SEWAB pro-
duces slightly too much runoff although the fast surface
component did not contribute. Inclusion of a variable
infiltration capacity allows surface runoff production
and improves the timing of the runoff calculation. Al-
lowing in addition subsurface runoff before the soil is
saturated (ARNO scheme) and the inclusion of a depth-
dependent saturation hydraulic conductivity influenced
the runoff production only little. Overall, the simulated
runoff coincides quite well with the average runoff ob-
servations for the whole Arkansas-Red-River basin as
well as for one particular test catchment, namely the
Black Bear River.

SEWAB will now be implemented into a non-
hydrostatic mesoscale model and replace a simple
bucket-type approach ([22]) for process studies in the
framework of the Baltic Sea Experiment (BALTEX), the
European continental scale experiment for GEWEX. It
is also used to calculate runoff data from rainfall ob-
servations as input to a conceptual large-scale hydro-
logical model [17]. Extensions of SEWAB will include
the parameterization of snow and frozen soil as well as
the consideration of land–surface heterogeneity inside
grid cells.

Acknowledgements

The FIFE data set was prepared by A. Betts and
gratefully provided by Y. Xue and I. Takayabu. The
authors thank the organizers of PILPS 2a, T.H. Chen,
A. Henderson-Sellers and W. Qu, and PILPS 2c, D.P.
Lettenmaier, X. Liang, E.F. Wood and D. Lohmann,
for providing the data and allowing adaption of figures
of the workshop reports.

Appendix A

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Description</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>$a$</td>
<td>reflectivity of the surface</td>
<td></td>
</tr>
<tr>
<td>$A$</td>
<td>saturated fraction of grid area</td>
<td></td>
</tr>
<tr>
<td>$B_g$</td>
<td>soil temperature diffusivity</td>
<td>m$^2$ s$^{-1}$</td>
</tr>
<tr>
<td>$C_H$</td>
<td>drag coefficient for heat flux</td>
<td></td>
</tr>
<tr>
<td>$C_Q$</td>
<td>drag coefficient for moisture flux</td>
<td></td>
</tr>
<tr>
<td>$c_p$</td>
<td>specific heat of air</td>
<td>J m$^{-3}$ K$^{-1}$</td>
</tr>
<tr>
<td>$D_h$</td>
<td>hydraulic diffusivity</td>
<td>m$^2$ s$^{-1}$</td>
</tr>
<tr>
<td>$E$</td>
<td>moisture flux</td>
<td>kg m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$g$</td>
<td>gravitational constant</td>
<td>m s$^{-2}$</td>
</tr>
<tr>
<td>$h_f$</td>
<td>canopy height</td>
<td>m</td>
</tr>
<tr>
<td>$i$</td>
<td>infiltration</td>
<td>kg m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$i$</td>
<td>infiltration capacity</td>
<td>kg m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$K_g$</td>
<td>hydraulic conductivity</td>
<td>m s$^{-1}$</td>
</tr>
<tr>
<td>$K_s$</td>
<td>saturation hydraulic conductivity</td>
<td>m s$^{-1}$</td>
</tr>
<tr>
<td>$LAI$</td>
<td>leaf area index</td>
<td></td>
</tr>
<tr>
<td>$L_v$</td>
<td>latent heat of vaporization</td>
<td>J kg$^{-1}$</td>
</tr>
<tr>
<td>$N$</td>
<td>cloud cover</td>
<td></td>
</tr>
<tr>
<td>$P$</td>
<td>precipitation</td>
<td>kg m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$q_a$</td>
<td>specific humidity of air</td>
<td>kg m$^{-3}$</td>
</tr>
<tr>
<td>$q_s$</td>
<td>saturation specific humidity</td>
<td>kg m$^{-3}$</td>
</tr>
<tr>
<td>$Q_{rad}$</td>
<td>radiation flux density</td>
<td>W m$^{-2}$</td>
</tr>
<tr>
<td>$Q_{sens}$</td>
<td>sensible heat flux</td>
<td>W m$^{-2}$</td>
</tr>
<tr>
<td>$Q_{lat}$</td>
<td>latent heat flux</td>
<td>W m$^{-2}$</td>
</tr>
<tr>
<td>$Q_{soil}$</td>
<td>soil heat flux</td>
<td>W m$^{-2}$</td>
</tr>
<tr>
<td>$R$</td>
<td>surface runoff</td>
<td>kg m$^{-2}$ s$^{-1}$</td>
</tr>
<tr>
<td>$R_a$</td>
<td>aerodynamic resistance</td>
<td>s m$^{-1}$</td>
</tr>
<tr>
<td>$R_f$</td>
<td>plant resistance to water flow</td>
<td>s</td>
</tr>
<tr>
<td>$R_{ld}$</td>
<td>downward longwave radiation</td>
<td>W m$^{-2}$</td>
</tr>
<tr>
<td>$R_{ld}$</td>
<td>leaf drip</td>
<td>kg m$^{-2}$ s$^{-1}$</td>
</tr>
</tbody>
</table>

Table 2
Annual runoff of the Black Bear River

<table>
<thead>
<tr>
<th>Year</th>
<th>Precipitation (mm/y)</th>
<th>Measured runoff (mm/y)</th>
<th>run 1 runoff (mm/y)</th>
<th>run 2 runoff (mm/y)</th>
<th>run 3 runoff (mm/y)</th>
<th>run 4 runoff (mm/y)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1980</td>
<td>914</td>
<td>144</td>
<td>149</td>
<td>161</td>
<td>161</td>
<td>150</td>
</tr>
<tr>
<td>1981</td>
<td>781</td>
<td>15</td>
<td>6</td>
<td>14</td>
<td>14</td>
<td>13</td>
</tr>
<tr>
<td>1982</td>
<td>753</td>
<td>101</td>
<td>93</td>
<td>93</td>
<td>93</td>
<td>92</td>
</tr>
<tr>
<td>1983</td>
<td>890</td>
<td>111</td>
<td>171</td>
<td>174</td>
<td>171</td>
<td>167</td>
</tr>
<tr>
<td>1984</td>
<td>653</td>
<td>87</td>
<td>46</td>
<td>49</td>
<td>53</td>
<td>49</td>
</tr>
<tr>
<td>1985</td>
<td>1016</td>
<td>188</td>
<td>196</td>
<td>196</td>
<td>188</td>
<td>186</td>
</tr>
<tr>
<td>1986</td>
<td>1314</td>
<td>348</td>
<td>448</td>
<td>451</td>
<td>448</td>
<td>448</td>
</tr>
<tr>
<td>1987</td>
<td>1000</td>
<td>234</td>
<td>155</td>
<td>162</td>
<td>163</td>
<td>163</td>
</tr>
</tbody>
</table>


References


[28] Richtmyer RD, Morton KW. Difference methods for initial value

[29] Sellers PJ, Mintz Y, Sud YC, Dalcher A. A simple biosphere
model (SiB) for use within general circulation models. J Atmos Sci

heat and moisture in the soil and their application to boundary

[31] Tjernström M. Some tests with a surface energy balance scheme,
including a bulk parameterization for vegetation, in a mesoscale

hydrology parameterization with subgrid variability for general