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Estimation of Evapotranspiration Using Soil Water Balance Modelling

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1. Introduction

Assessing evapotranspiration is a key issue for natural vegetation and crop survey. It is a very important step to achieve the soil water budget and for deriving drought awareness indices. It is also a basis for calculating soil-atmosphere Carbon flux. Hence, models of evapotranspiration, as part of land surface models, are assumed as key parts of hydrological and atmospheric general circulation models (Johnson et al., 1993). Under particular climate (represented by energy limiting evapotranspiration rate corresponding to potential evapotranspiration) and soil vegetation complex, evapotranspiration is controlled by soil moisture dynamics. Although radiative balance approaches are worth noting for evapotranspiration evaluation, according to Hofius (2008), the soil water balance seems the best method for determining evapotranspiration from land over limited periods of time. This chapter aims to discuss methods of computing and updating evapotranspiration rates using soil water balance representations.
At large scale, Budyko (1974) proposed calculating annual evapotranspiration from data of meteorological stations using one single parameter w_0 representing a critical soil water storage. Using a statistical description of the sequences of wet and dry days, Eagleson (1978 a) developed an average annual water balance equation in terms of 23 variables including soil, climate and vegetation parameters with the assumption of a homogeneous soil-atmosphere column using Richards (1931) equation. On the other hand, the daily bucket with bottom hole model (BBH) proposed by Kobayashi et al. (2001) was introduced based on Manabe model (1969) involving one single layer bucket but including gravity drainage (leakage) as well as capillary rise. Vrugt et al. (2004) concluded that the daily Bucket model and the 3-D model (MODHMS) based on Richards equation have similar results. Also, Kalma & Boulet (1998) compared simulation results of the rainfall runoff hydrological model VIC which assumes a bucket representation including spatial variability of soil parameters to the one dimensional physically based model SiSPAT (Braud et al. , 1995). Using soil moisture profile data for calibration, they conclude that catchment's scale wetness index for very dry and very wet periods are misrepresented by SiSPAT while captured by VIC. Analyzing VIC parameter identifiability using streamflow data, DeMaria et al. (2007) concluded that soil parameters sensitivity was more strongly dictated by climatic gradients than by changes in soil properties especially for dry environments. Also, studying the measurements of soil moisture of sandy soils under semi-arid conditions, Ceballos et al. (2002) outlined the dependence of soil moisture time series on intra annual rainfall

variability. Kobayachi et al. (2001) adjusted soil humidity profiles measurements for model calibration while Vrugt et al. (2004) suggested that effective soil hydraulic properties are poorly identifiable using drainage discharge data.
The aim of the chapter is to provide a review of evapotranspiration soil water balance models. A large variety of models is available. It is worth noting that they do differ with respect to their structure involving empirical as well as conceptual and physically based models. Also, they generally refer to soil properties as important drivers. Thus, the chapter will first focus on the description of the water balance equation for a column of soil-atmosphere (one dimensional vertical equation) (section 2). Also, the unsaturated hydrodynamic properties of soils as well as some analytical solutions of the water balance equation are reviewed in section 2. In section 3, key parameterizations generally adopted to compute actual evapotranspiration will be reported. Hence, several soil water balance models developed for large spatial and time scales assuming the piecewise linear form are outlined. In section 4, it is focused on rainfall-runoff models running on smaller space scales with emphasizing on their evapotranspiration components and on calibration methods. Three case studies are also presented and discussed in section 4. Finally, the conclusions are drawn in section 5.

2. The one dimensional vertical soil water balance equation

As pointed out by Rodriguez-Iturbe (2000) the soil moisture balance equation (mass conservation equation) is "likely to be the fundamental equation in hydrology". Considering large spatial scales, Sutcliffe (2004) might agree with this assumption. In section 2.1 we first focus on the presentation of the equation relating relative soil moisture content to the water balance components: infiltration into the soil, evapotranspiration and leakage. Then water loss through vegetation is addressed. Finally, infiltration models are discussed in section 2.2.

2.1 Water balance

For a control volume composed by a vertical soil column, the land surface, and the corresponding atmospheric column, and under solar radiation and precipitation as forcing variables, this equation relates relative soil moisture content *s* to infiltration into the soil I(s,t), evapotranspiration E(s,t) and leakage L(s,t).

$$nZ_a\ \delta s/\delta t = I(s,t) - E(s,t) - L(s,t) \quad (1a)$$

Where t is time, n is soil effective porosity (the ratio of volume of voids to the total soil matrix volume); and Z_a is the active depth of soil.
Soil moisture exchanges as well as surface heat exchanges depend on physical soil properties and vegetation (through albedo , soil emissivity, canopy conductance) as well as atmosphere properties (turbulent temperature and water vapour transfer coefficients, aerodynamic conductance in presence of vegetation) and weather conditions (solar radiation, air temperature, air humidity, cloud cover, wind speed). Soil moisture measurements require sampling soil moisture content by digging or soil augering and determining soil moisture by drying samples in ovens and measuring weight losses; also, in situ use of tensiometry, neutron scattering, gamma ray attenuation, soil electrical conductivity analysis, are of common practice (Gardner et al. (2001) ; Sutcliffe, 2004; Jeffrey et al. (2004)).

The basis of soil water movement has been experimentally proposed by Darcy in 1856 and expresses the average flow velocity in a porous media in steady-state flow conditions of groundwater. Darcy introduced the notion of hydraulic conductivity. Boussinesq in 1904 introduced the notion of specific yield so as to represent the drainage from the unsaturated zone to the flow in the water table. The specific yield is the flux per unit area draining for a unit fall in water table height. Richards (1931) proposed a theory of water movement in the unsaturated homogeneous bare soil represented by a semi infinite homogeneous column:

$$\delta\theta/\delta t = \delta/\delta z\ [\ K\ \delta\psi/\delta z - K(\theta)] \tag{1b}$$

Where t is time; θ is volumetric water content (which is the ratio between soil moisture volume and the total soil matrix volume cm^3cm^{-3}); z is the vertical coordinate (z>0 downward from surface); K is hydraulic conductivity (cms^{-1}); ψ is the soil water matrix potential. Both K and ψ are function of the volumetric water content. Richards equation assumes that the effect of air on water flow is negligible. If accounting for the slope surface, it comes:

$$\delta\theta/\delta t = \delta/\delta z\ [K\ \delta\psi/\delta\theta\ \delta\theta/\delta z] - \delta\ K/\delta\theta\ \ \delta\theta/\delta z]\ \cos\beta \tag{2}$$

Where β is surface slope angle and cos is the cosinus function. We notice that the term [K $\delta\psi/\delta z$ – K(θ)] represents the vertical moisture flux. In particular, as reported by Youngs (1988) the soil-water diffusivity parameter D has been proposed by Childs and Collis-George (1950) as key soil-water property controlling the water movement.

$$D(\theta) = K(\theta)\ \delta\psi/\delta\theta \tag{3}$$

Thus, the Richards equation is often written as following:

$$\delta\theta/\delta t = \delta/\delta z\ [\ D(\theta)\ \delta\theta/\delta z\ ] - \delta\ K(\theta)/\delta z] \tag{4}$$

Eq. (4) is generally completed by source and sink terms to take into account the occurrence of precipitation infiltrating into the soil $I_{nf}(\theta,z_0)$ where z_0 is the vertical coordinate at the surface and vegetation uptake of soil moisture $g_r(\theta,z)$,. Vegetation uptake (transpiration) depends on vegetation characteristics (species, roots, leaf area, and transfer coefficients) and on the potential rate of evapotranspiration E_0 which characterizes the climate. Consequently, Eq. (4) becomes:

$$\delta\theta/\delta t = \delta/\delta z\ [\ D(\theta)\delta\theta/\delta z - K(\theta)] - g_r(\theta,z) + I_{nf}(\theta,z_0) \tag{5}$$

Youngs (1988) noticed that near the soil surface where temperature gradients are important Richards equation may be inadequate. We find in Raats (2001) an important review of evapotranspiration models and analytical and numerical solutions of Richards equation. However, it should be noticed that after Feddes et al. (2001) "in case of catchments with complex sloping terrain and groundwater tables, a vertical domain model has to be coupled with either a process or a statistically based scheme that incorporates lateral water transfer". So, a key task in the soil water balance model evaluation is the estimation of $I_{nf}(\theta,z_0)$ and $g_r(\theta,z)$. Both depend on the distribution of soil moisture. We focus here on vegetation uptake (or transpiration) $g_r(\theta,z)$ which is regulated by stomata and is driven by atmospheric demand. Based on an Ohm's law analogy which was primary proposed by Honert in 1948 as outlined by Eagleson (1978 b), the conceptual model of local transpiration uptake u(z,t)= $g_r(\theta,z)$ as volume of water per area per time is expressed as (Guswa, 2005)

$$u(z,t)=\Delta z\ (\psi(z,t) - \psi_p)\ /[\ R_1(\theta\ (z,t))+R_2] \tag{6}$$

ψ soil moisture potential (bars), ψ_p leaf moisture potential (bars); R_1 (s cm^{-1}) a resistance to moisture flow in soil; it depends on soil and root characteristics and is function of the volumetric water content; R_2 (s cm^{-1}) is vegetation resistance to moisture flow; Δz is soil depth. It is worth noting that $\psi_p > \psi*$ where $\psi*$ is the wilting point potential; In Ceballos et al. (2002) the wilting point is taken as the soil-moisture content at a soil-water potential of -1500 kPa.
Estimations of air and canopy resistances R_1 and R_2 often use semi-empirical models based on meteorological data such as wind speed as explanatory variables (Monteith (1965); Villalobos et al., 2000). Jackson et al. (2000) pointed out the role of the Hydraulic Lift process which is the movement of water through roots from wetter, deeper soil layers into drier, shallower layers along a gradient in ψ. On the basis of such redistribution at depth, Guswa (2005) introduced a parameter to represent the minimum fraction of roots that must be wetted to the field capacity in order to meet the potential rate of transpiration. The field capacity is defined as the saturation for which gravity drainage becomes negligible relative to potential transpiration (Guswa, 2005). The potential matrix at field capacity is assumed equal to 330 hPa (330 cm) (Nachabe, 1998). The resulting u(z,t) function is strongly non linear versus the average root moisture with a relative insensitivity to changes in moisture when moisture is high and sensitivity to changes in moisture when the moisture is near the wilting point conditions. We also emphasize the Perrochet model (Perrochet, 1987) which links transpiration to potential evapotranspiration E_0 through:

$$g_r(\theta,z,t) = \alpha(\theta)r(z)\ E_0(t) \tag{7}$$

Where r(z) (cm^{-1}) is a root density function which depends both on vegetation type and climatic conditions, $\alpha(\theta)$ is the root efficiency function. Both r(z) and $\alpha(\theta)$ represent macroscopic properties of the root soil system; they depend on layer thickness and root distribution . Lai and Katul (2000) and Laio (2006) reported some models assigned to r(z) which are linear or non linear. As out pointed by Laio (2006), models generally assume that vegetation uptake at a certain depth depends only on the local soil moisture. It is noticeable that in Feddes et al. (2001), a decrease of uptake is assumed when the soil moisture exceeds a certain limit and transpiration ceases for soil moisture values above a limit related to oxygen deficiency.

2.2 Review of models for hydrodynamic properties of soils

Many functional forms are proposed to describe soil properties evolution as a function of the volumetric water content (Clapp et al. , 1978). They are called retention curves or pedo transfer functions. We first present the main functional forms adopted to describe hydraulic parameters (section 2.2.1). Then, we report some solutions of Richards equation (section 2.2.2).

2.2.1 Functional forms of soil properties

According to Raats (2001), four classes of models are distinguishable for representing soil hydraulic parameters. Among them the linear form with D as constant and K linear with θ and the function Delta type as proposed by Green Ampt D= $\frac{1}{2} s^2 (\theta_1 - \theta_0)^{-1} \delta(\theta_1 - \theta_0)$ where s is the degree of saturation (which is the ratio between soil moisture volume and voids volume; s=1 in case of saturation) and θ_1; θ_0 parameters. Also power law functions for $\psi(\theta)$ and $K(\theta)$ are proposed by Brooks and Corey (1964) on the basis of experimental observations while Gardner (1958) assumes exponential functions. The power type model proposed by Brooks & Corey (1964) are the most often adopted forms in rainfall-runoff transformation models. The Brooks and Corey model for K and ψ is written as:

$$K(s) = K(1)\, s^{c'} \; ; \; \psi(s) = \psi(1)\, s^{-1/m} \tag{8}$$

where m is a pore size index and c′ a pore disconnectedness index (Eagleson 1978 a,b); After Eagleson (1978a, b), c′ is linked to m with c′=(2+3m)/m. In Eq. (8), K(1) is hydraulic conductivity at saturation (for s=1); $\psi(1)$ is the bubbling pressure head which represents matrix potential at saturation. During dewatering of a sample, it corresponds to the suction at which gas is first drawn from the sample; As a result, Brooks and Corey (*BC*) model for diffusivity is derived as:

$$D(\theta) = s^{d}\, \psi(1)\; K(1)\, /(nm) \tag{9}$$

where n is effective soil porosity; and d=(c′-1- (1/m)). Let's consider the intrinsic permeability k which is a soil property. (K and k are related by $K= k\, \rho_w /\mu$ where μ dynamic viscosity of water; ρ_w specific weight of pore water). After Eagleson (1978 a, b), three parameters involved in pedo transfer functions may be considered as independent parameters: n, c′ and k(1) where k(1) is intrinsic permeability at saturation.
On the other hand, Gardner (1958) model assumed the exponential form for the hydraulic conductivity parameter (Eq. 10):

$$K(\psi) = K_S\, e^{-a'\psi} \tag{10}$$

Where K_S saturated hydraulic conductivity at soil surface; a' pore size distribution parameter. Also, in Gardner (1958) model, the degree of saturation and the soil moisture potential are linked according to Eq. (11). The power function introduces a parameter l which is a factor linked to soil matrix tortuosity (l= 0.5 is recommended for different types of soils);.

$$s(\psi) = [e^{-0.5\, a'\psi}\, (1+ 0.5\, a'\, \psi\,)]^{2/(l+2)} \tag{11}$$

Van Genutchen model (1980) is another kind of power law model but it is highly non linear

$$K(\psi) = K_S\, s^{\,\lambda+1}\, [\, 1- (1- s^{\,(\lambda+1)/\lambda})^{\lambda/(\lambda+1)}]^2 \tag{12}$$

$$\begin{aligned} s(\psi) &= [1+ (\psi(1)/\psi))^{-(\lambda+1)}\,]^{-\,\lambda/(\lambda+1)} && \text{for } \psi\leq\psi(1); \\ s&=1 && \text{for } \psi<\psi(1) \end{aligned} \tag{13}$$

In Eq. (12) and (13) λ is a parameter to be calibrated. Calibration is generally performed on the basis of the comparison of computed and observed retention curves.
In order to determine K_S one way is to adopt Cosby et al. (1984) model (Eq. 14).

$$\mathrm{Log}(K_S)= -0.\,6 +(\,0.0126\, S_{\%} - 0.0064\, C_{\%}) \tag{14}$$

Where $S_{\%}$ and $C_{\%}$ stand for soil percents of sand and clay. Also, we may find tabulated values of K_S (in m/day) according to soil texture and structure properties in FAO (1980). On the other hand, soil field capacity S_{FC} plays a key role in many soil water budget models. In Ceballos et al. (2002) the field capacity was considered as "the content in humidity corresponding to the inflection point of the retention curve before it reached a trend parallel to the soil water potential axis". In Guswa (2005), it is defined as the saturation for which gravity drainage becomes negligible relative to potential transpiration. As pointed out by Liao (2006) who agreed with Nachabe (1998), there is an "intrinsic subjectivity in the definition of field capacity". Nevertheless, many semi-empirical models are offered in the literature for S_{FC} estimation as a function of soil properties (Nachabe, 1988). In Cosby (1984), S_{FC} expressed as a degree of saturation is assumed s:

$$S_{FC} = 50.1 + (-0.142\, S_{\%} - 0.037\, C_{\%}) \quad (15)$$

On the other hand, according to Cosby (1984) and Saxton et al. (1986) S_{FC} may be derived as:

$$S_{FC} = (20/A')^{1/B'} \quad (16)$$

where
$A'=100*\exp(a_1+a_2C_{\%}+a_3S_{\%}^2+a_4S_{\%}^2C_{\%})$; $B'=a_5+a_6C_{\%}^2+a_7S_{\%}^2+a_8S_{\%}^2C_{\%}$; $a_1 = -4{,}396$; $a_2 = -0{,}0715$; $a_3 = -0{,}000488$; $a_4 = -0{,}00004285$; $a_5 = -0{,}00222$; $a_6 = -0{,}00222$; $a_7 = -0{,}00003484$; $a_8 = -0{,}00003484$
Recently, this model was adopted by Zhan et al. (2008) to estimate actual evapotranspiration in eastern China using soil texture information. Also, soil characteristics such as S_{FC} may be obtained from Rawls & Brakensiek (1989) according to soil classification (Soil Survey Division Staff, 1998). Nasta et al. (2009) proposed a method taking advantage of the similarity between shapes of the particle-size distribution and the soil water retention function and adopted a log-Normal Probability Density Function to represent the matrix pressure head function retention curve.

2.2.2 Review of analytical solutions of the movement equation

Two well-known solutions of Richards equation are reported here (Green &Ampt model (1911), Philip model (1957)) as well as a more recent solution proposed by Zhao and Liu (1995). These solutions are widely adopted in rainfall-runoff models to derive infiltration.
In the Green &Ampt method (1911), it is assumed that infiltration capacity f from a ponded surface is:

$$f = K_{av}\,(1 + \Delta\psi\, \Delta\theta\, F^{-1}) \quad (17)$$

K_{av} average saturated hydraulic conductivity ; $\Delta\psi$ difference in average matrix potential before and after wetting; $\Delta\theta$ difference in average soil water content before and after wetting; F the cumulative infiltration for a rainfall event (with $f = dF/dt$).
In the Philip (1957) solution, it is assumed that the gravity term is negligible so that $\delta K(\theta)/\delta z] \approx 0$. A time series development considers the soil water profile of the form:

$$z(\theta,t) = f_1(\theta)\, t^{1/2} + f_2(\theta)\, t + f_3(\theta)\, t^{3/2} + \ldots \quad (18)$$

Where f_1, f_2, … are functions of θ. Hence, the cumulative infiltration $\Omega_f(t)$ is:

$$\Omega_f(t) = S\, t^{1/2} + (A_2 + K_S)\, t + A_3\, t^{3/2} + \ldots \quad (19)$$

Where S soil sorptivity, K_S is saturated hydraulic conductivity of the soil and A_1, A_2, … are parameters. Philip suggested adopting a truncation that results in:

$$\Omega_f(t) = S\, t^{1/2} + K_S / n'\, t \tag{20}$$

Where n' is a factor $0.3 < n' < 0.7$. It is worth noting that the soil sorptivity S depends on initial water content. So it has to be adjusted for each rainfall event. This is usually performed by comparing observed and simulated cumulative infiltration. For further discussion of Philip model, the reader may profitably refer to Youngs (1988).
Another model of infiltration is worth noting. It is the model of Zhao and Liu (1995) which introduced the fraction of area under the infiltration capacity:

$$i(t) = i_{max}\,[1 - (1 - A(t))^{1/b''}] \tag{21}$$

Where i(t) is infiltration capacity at time t. Its maximum value is i_{max}. A(t) is the fraction of area for which the infiltration capacity is less than i(t) and b′′ is the infiltration shape parameter. As out pointed by DeMaria et al. (2007), the parameter b′′ plays a key role. Effectively, an increase in b′′ results in a decrease in infiltration.

3. Review of various parameterizations of actual evapotranspiration

Many early works on radiative balance combination methods for estimating latent heat using Penman – Monteith method (Monteith, 1965) were coupled with empirical models for representing the conductance of the soil-plant system (the conductance is the inverse function of the resistance). Based on observational evidence, these works have assumed a linear piecewise relation between volumetric soil moisture and actual evapotranspiration. Thus, several water balance models have been developed for large spatial and time scales assuming this piecewise linear form beginning from the work of Budyko in 1956 as pointed out by Manabe (1969)), Budyko (1974), Eagleson (1978 a, b), Entekhabi & Eagleson (1989) and Milly (1993). In fact, soil water models for computing actual evapotranspiration differ according to the time and space scales and the number of soil layers adopted as well as the degree of schematization of the water and energy balances. Moreover, specific canopy interception schemes, pedo transfer sub-models and runoff sub-models often distinguish between actual evapotranspiration schemes. Also, models differ by the consideration of mixed bare soil and vegetation surface conditions or by differencing between vegetation and soil cover. In the former, there is a separation between bare soil evapotranspiration and vegetation transpiration as distinct terms in the computation of evapotranspiration. In the following, we first present a brief review of land surface models which fully couple energy and mass transfers (section 3.1). Then, we make a general presentation of soil water balance models based on the actualisation of soil water storage in the upper soil zone assuming homogeneous soil (section 3.2).Further, it is focused on the estimation of long term actual evapotranspiration using approximation of the solution of the water balance model (section 3.3). In section 3.4, large scale soil water balance models (bucket schematization) are outlined with much more details. Finally a discussion is performed in section 3.5.

3.1 Review of land surface models

In Soil-Vegetation-Atmosphere-Transfer (SVAT) models or land surface models, energy and mass transfers are fully coupled solving both the energy balance (net radiation equation, soil heat fluxes, sensible heat fluxes, and latent heat fluxes) in addition to water movement equations. Usually this is achieved using small time scales (as for example one hour time increment). The specificity of SVAT models is to describe properly the role of vegetation in the evolution of water and energy budgets. This is achieved by assigning land type and soil information to each model grid square and by considering the physiology of plant uptake. Many SVAT models have been developed in the last 25 years. We may find in Dickinson and al. (1986) perhaps one of the first comprehensive SVAT models which was addressed to be used for General circulation modelling and climate modelling. It was called BATS (Biosphere-Atmosphere Transfer Scheme). It was able to compute surface temperature in response to solar radiation, water budget terms (soil moisture, evapotranspiration and, runoff), plant water budget (interception and transpiration) and foliage temperature. ISBA model (Noilhan et Mahfouf, 1996) was further developed in France and belongs to "simple models with mono layer energy balance combined with a bulk soil description" (after Olioso et al. (2002)). An example of using ISBA scheme is presented in Olioso et al. (2002). The following variables are considered: surface temperature, mean surface temperature, soil volumetric moisture at the ground surface, total soil moisture, canopy interception reservoir. The soil volumetric moisture at the ground surface is adopted to compute the soil evaporation while the total soil moisture is used to compute transpiration. The total latent heat is assumed as a weighted average between soil evaporation and transpiration using a weight coefficient depending on the degree of canopy cover. Canopy albedo and emissivity, vegetation Leaf area index LAI, stomatal resistance, turbulent heat and transfer coefficients are parameters of the energy balance equations. It is worth noting that soil parameters in temperature and moisture are computed using soil classification databases. Without loss of generality we briefly present the two layers water movement model adopted by Montaldo et al. (2001)

$$\delta\theta_g / \delta t = C_1 / (\rho_w d_1) [P_g - E_g] - C_2 / \tau [\theta_g - \theta_{geq}] \qquad 0 \le \theta_g \le \theta_s \tag{22}$$

$$\delta\theta_2 / \delta t = C_1 / (\rho_w d_2) [P_g - E_g - E_{tr} - q_2] \qquad 0 \le \theta_2 \le \theta_s \tag{23}$$

d_1 and d_2 depth of near surface and root zone soil layers; ρ_w density of the water; θ_g and θ_2 volumetric water contents of near surface and root zone soil layers; θ_{geq} equilibrium surface volumetric soil moisture content ideally describing a reference soil moisture for which gravity balances capillary forces such that no flow crosses the bottom of the near surface zone of depth d_1; P_g precipitation infiltrating into the soil; E_g bare soil evaporation rate at the surface; E_{tr} transpiration rate from the root zone of depth d_2; q_2 rate of drainage out of the bottom of the root zone; It is assumed to be equal to the hydraulic conductivity of the root zone at $\theta=\theta_2$. τ ; C_1 and C_2 are parameters. In this model, the rescaling of the root zone soil moisture θ_2 seems to be highly recommended in order to achieve adequate prediction of θ_g in comparison to observations (Montaldo et al. (2001)). Using an assimilation procedure, Montaldo et al. (2001) achieved overcoming misspecification of K_S of two orders magnitude in the simulation of θ_2.

According to Franks et al. (1997), the calibration of SVAT schemes requires a large number of parameters. Also, field experimentations needed to calibrate these parameters are rather important. Moreover up scaling procedures are to be implemented. Boulet and al. (2000) argued that "detailed SVAT models especially when they exhibit small time and space steps are difficult to use for the investigation of the spatial and temporal variability of land surface fluxes".

3.2 Review of average long term evapotranspiration or "regional" evapotranspiration models

Considering the soil water balance at monthly time scale, Budyko (1974) introduced one single parameter which is a critical soil water storage w_0 corresponding to 1 m homogeneous soil depth. According to Budyko (1974), w_0 is a regional parameter seasonally constant and essentially depending on the climate-vegetation complex. The main assumption is that monthly actual evapotranspiration starts from zero and is a piecewise linear function of the degree of saturation expressed as the ratio w/w_0 where w is the actual soil water storage. Either, for $w \geq w_0$ actual evapotranspiration is assumed at potential value E_0.

Average annual water balance equation is also developed in Eagleson (1978 a) in terms of 23 variables (six for soil, six for climate and one for vegetation) with the assumption of a homogeneous soil-atmosphere column using Richards equation. Further, the behaviour of soil moisture in the upper soil zone (1 m deep or root zone) is expressed in terms of the following three independent soil parameters: effective porosity n, pore disconnectedness index c' and saturated hydraulic conductivity at soil surface K_S while storm and inter storm net soil moisture flux are coupled to storm and inter storm Probability Density Functions. The average annual evapotranspiration E_m is finally expressed as :

$$E_m = J(E_e, M_v, k_v)\,(E_{pa} - E_{ra}) \tag{24}$$

J(.) evapotranspiration function; E_{pa} average annual potential evapotranspiration; E_{ra} average annual surface retention; E_e exfiltration parameter as function of initial degree of saturation s_0; k_v plant coefficient. It is approximately equal to effective transpiring leaf surface per unit of vegetated land surface; M_v vegetation fraction of surface.

Further, Milly (1993) developed similar probabilistic approach for soil water storage dynamics based on Manabe model (Manabe, 1969). A key assumption is that the soil is of high infiltration capacity. The model adopts the so-called water holding capacity $W_{0,}$ which is a storage capacity parameter allowing the definition of the state "reservoir is full". For well developed vegetation, W_0 is interpreted as the difference between the volumetric moisture contents θ_f of the soil at field capacity and the wilting point θ_w ($W_0=\theta_f-\theta_w$). Furthermore, Milly (1994) adopted seasonally Poisson and exponential Probability Density Functions, together with seasonality of evapotranspiration forcing. To take into account horizontal large length scales, the spatial variability of water holding capacity W_0 was introduced, adopting a Gamma Probability Density Function with mean W_{m0}. In total, the model involved only seven parameters: a dryness index EDI = P / ETP, the mean holding capacity of soil W_{m0} and a shape parameter of the Gamma distribution,, mean storm arrival rate, and one measure of seasonality for respectively annual precipitation, potential evapotranspiration and storm arrival rate. Performing a comparison with observed annual runoff in US, it was found that the geographical distribution of calculated runoff shares at least qualitatively the large scale features of observed maps. In effect, 88% of the variance of grid runoff and 85% of the variance of grid evapotranspiration is reproduced by this model. However, it is outlined that the model presents failures within areas with elevation. Average annual precipitation and runoff over 73 large basins worldwide were also studied by (Milly and Dunne, 2002). Using precipitation and net radiation as independent variables, they compared observed mean runoff amounts to those computed by Turc-Pike and Budyko models. In northern Europe, they found a tendency for underestimation of observed evapotranspiration.

3.3 Empirical model for estimating regional evapotranspiration

Combining the water balance to the radiative balance at monthly scale, Budyko proposed an asymptotic solution in which R_n stands for average annual net radiation (which is the net energy exchange with the atmosphere equal to net radiation – sensible heat flux – latent heat flux), P average annual precipitation, E_m average (long term) annual evapotranspiration, ϕ a function expressed in Eq. (26).

$$E_m / P = \phi (R_n / P) \tag{25}$$

$$\phi (x) = [x (\tanh(x^{-1})) (1 - \cosh(x) + \sinh(x))]^{1/2} \tag{26}$$

Where tanh(.) stands for hyperbolic tangent, cosh(.) hyperbolic cosines, sinh(.) hyperbolic sinus

According to Shiklomavov (1989) and Budyko (1974), Ol'dekop was the first to propose in 1911 an empirical formulation of the relationship between climate characteristics and water balance terms (rainfall and runoff) assuming the concept of « maximum probable evaporation» E_{max} and using the ratio P / E_{max}. According to Milly (1994), works of Budyko in 1948 resulted, on the basis of dimensional analysis, to propose the ratio R_n/P as radiative index of aridity. Conversely, the function ϕ (Eq. 26) was empirical and was derived assuming that in arid climate E_m approaches P while it approaches R_n under humid climate.Budyko model was validated using 1200 watersheds world wild computing E_m as the difference between average long term annual observed rainfall and annual observed runoff. Model accuracy is reflected by the fact that the ratio E_m /P is simulated within a relative error of 10% (Budyko, 1974). However, larger discrepancy values are found for basins with important orography. Choudhury (1999) proposed to adopt Eq. (27) to derive ϕ :

$$\phi (x) = (1+x^{-\nu})^{-1/\nu} \tag{27}$$

where ν is a parameter depending of the basin characteristics. Milly et Dunne (2002) reported that ν=2.1 closely approximates Budyko model, while ν=2 corresponds to Turc-Pike model. According to Choudhury (1999), the more the basin area is large, the more ν is small and smaller is E_m. ν=2.6 is recommended for micro-basins while ν=1.8 for large basins. According to Milly et Dunne (2002), it was found that for a large interval of watershed areas, ν=1.5 to 2.6.

Another approximation of Budyko model is the Hsuen Chun (1988) model (H.C.) introducing the ratio $ID_{etp} = E_0/P$ and an empirical parameter k'.

$$E_m = E_0 [ID_{etp}^{\,k'} / (1+ ID_{etp}^{\,k'})]^{1/k'} \tag{28}$$

After Hsuen Chun (1988) the value *k'*=2.2 reproduces Budyko model results. According to Pinol et al. (1991), the adjusted values of k' are in the interval 1.03 <k'< 2.40. Also, they noticed that k' depends on the type of vegetation cover. After Donohue et al. (2007), Eq. (28) may be adopted for basins with area < 1000 Km^2 and series of at least 5 year length.

3.4 Modeling of actual evapotranspiration for long time series and large scale applications

Simple soil water balance models based on bucket schematization have been developed to fulfil the need to simulate long time series of water balance outputs allowing the calculation of actual evapotranspiration. We focus the review on the Manabe model (1969), the

Rodriguez-Iturbe et al. (1999) model and the Bottom hole bucket model of Kobayachi et al. (2001).

3.4.1 Manabe bucket model

In fact, the single layer single bucket model of Manabe (1969) takes a central place in large scale water budget modelling. It was proposed as part of the climate and ocean circulation model. This conceptual model runs at the monthly scale and adopts the field capacity S_{FC} as key parameter. Also, it assumes an effective parameter W_k representing a fraction of the field capacity ($W_k = 0.75^* S_{FC}$). Here we notice that the field capacity S_{FC} is now expressed as a water content. The climatic forcing is represented by the potential evapotranspiration E_0. Let w be the actual soil water content. The actual evapotranspiration E_a is expressed as a linear piecewise function:

For $w \geq W_k$ $E_a= E_0$

For $w<W_k$ $E_a= E_0^*(w/W_k)$

On the other hand, the surface runoff R_s component in Manabe model depends on the actual soil moisture content in comparison to the field capacity as well as on the precipitation forcing compared to the potential evapotranspiration uptake. Let Δw the change in soil water content. Thus, surface runoff is assumed as following:

For $w= S_{FC}$ and $P> E_0$; $\Delta w=0$ and $R_s= P- E_0$

For $w< S_{FC}$; $\Delta w=P-E_a$; $R_s=0$

Another well-known model is FAO-56 model (Allen et al. (1998)). In fact, it is based on Manabe soil water budget. However, it takes into account the water stress through an empirical coefficient K'_s. First of all, in FAO-56 model, it is important to outline that the potential evapotranspiration is replaced by a reference evapotranspiration E_r computed using Penman-Montheith model with respect to a reference grass corresponding to an albedo value equal 0.23. Then, a seasonal crop coefficient K_c is introduced. The parameter K_c depends on both the crop type and the vegetative stage. Default K_c values are reported in (Allen et al. (1998)) for various crop types. This crop coefficient corresponds to ideal soil moisture conditions related to no water stress conditions and to good biological conditions. In real conditions, K_c is corrected by a correction coefficient K'_s ($0<K'_s<1$) such that the product $K_c K'_s$ includes the vegetation type as well as the water stress conditions. So actual evapotranspiration is written as:

$$E_a = K_c K'_s E_r \tag{29}$$

According to Biggs et al. (2008) mild stress conditions would correspond to K'_s of 0.8 and moderate stress conditions to K'_s of 0.6. Based on the findings that default K_c values underestimate lysimeter experiments K_c values, Biggs et al. (2008) built a non linear regression relationships between the product ($K_c K'_s$) and the ratio of seasonal precipitation to potential evapotranspiration for various crop types. To that purpose they fitted a Beta Probability Density Function to the correction factor K'_s. They adopted lysimeter observations to fit this modified FAO-56 model.. The model explained (49–90%) of the variance in actual evapotranspiration, depending on the crop type.

3.4.2 Rodriguez-Iturbe model

In Rodriguez-Iturbe et al. (1999), the point of departure is infiltration into the soil which is expressed as function of the existing soil moisture which is reported in terms of saturation (corresponding to s= w/nZ_a where Za is effective depth of soil and n soil effective porosity). Soil drainage varies according to a power law although it is approximated by two linear segments. Consequently, it is assumed that soil drainage occurs for s exceeding a threshold value s_1, going from zero for $s=s_1$ to K_S for saturated condition (s=1) where K_S is the saturated hydraulic conductivity of the soil. Moreover, a saturation threshold s* is assumed to reduce evapotranspiration in case of water stress. Its value depends on the type of vegetation. Thus, for s≤s*, the evapotranspiration is computed as the potential rate scaled by the ratio s/s* while the evapotranspiration is at potential value for s> s*.

$$E_a(s)=E_0\, s/s^* \qquad \text{For } s\leq s^* \tag{30}$$

$$E_a(s)=E_0 \qquad \text{For } s>s^* \tag{31}$$

Milly (2001) model corresponds to the case s* → 0 and K_S → infinity. According to Milly (2001), the introduction of the threshold parameter s* is much recommended especially under arid conditions. In the case where no distinction is made between forested and bare soil areas, Rodriguez-Iturbe et al. (1999) pointed out that s* is considerably lower than the field capacity S_{FC} conversely to Manabe model which corresponds to s* = 0.75 S_{FC}. Laio (2006) adopted a generalized form of Rodriguez-Iturbe et al. (1999) model by accounting for the reduction of evapotranspiration in case of water stress by introducing the soil moisture at wilting point s_w. He represented s* as a soil moisture level above which plant stomata are completely opened (Eq. 32 and Eq. 33).

$$E_a(s)=E_0\, (s-s_w)/(s^*-s_w) \qquad \text{For } s\leq s^* \tag{32}$$

$$E_a\,(s)=E_0 \qquad \text{For } S_{FC}>s>s^* \tag{33}$$

On the other hand, Rodriguez-Iturbe et al. (1999) model the leakage component is represented by the exponential decay Gardner model. This model was also adopted by Guswa et al. (2002). Leakage component is assumed as exponential decay function of the effective degree of soil saturation, as well as soil characteristics (saturated hydraulic conductivity, drainage curve parameter and field capacity).

3.4.3 Bottom hole bucket model

The daily bucket with bottom hole model (BBH) proposed by Kobayashi et al. (2001) is also based on Manabe model involving one layer bucket but including gravity drainage (leakage) as well as capillary rise. Kobayashi et al. (2001) outlined that the soil moisture dynamics is better simulated by BBH than by Bucket (Manabe) model. Kobayashi et al. (2007) developed a new version of BBH named BBH-B including a second soil layer in order to take into account for the variability of the soil profile when the root zone is rather deep (1 m or more).

In the following, we focus on BBH model where forcing variables are precipitation P and potential evapotranspiration E_0. The actual evapotranspiration is assumed as:

$$\begin{aligned} E_a &= M'\, E_0 \qquad \text{For } s\leq s^* \\ E_a &= E_0 \qquad \text{For } s>s^* \end{aligned} \tag{34}$$

Where M′ is a water stress factor updated at each time step and expressed as:

$$M' = Min\,(1, w/(\sigma W_{max})) \qquad For\ s \leq s^* \tag{35}$$

σ: parameter representing the resistance of vegetation to evapotranspiration; $W_{max}=nZ_a$ where W_{max}: total water-holding capacity (mm); Z_a: thickness of active soil layer (mm); n: effective soil porosity.
Percolation and capillary rise term Gd(t) is assumed according to exponential function.

$$Gd(t) = exp\,((w(t)-a)/b)-c \tag{36}$$

Where a: parameter related to the field capacity (mm); b: parameter representing the decay of soil moisture (mm); c: parameter representing the daily maximal capillary rise (mm). On the other hand, daily surface runoff Rs(t) is expressed as:

$$Rs(t) = Max\,[P(t)-(W_{BC}-W(t))-E_a(t)-Gd(t), 0] \tag{37}$$

Where $W_{BC}= \eta\ W_{max}$; η : parameter representing the moisture retaining capacity ($0< \eta <1$).
According to Kobayachi and al. (2001) the parameter a (which corresponds here to a/Wmax) is “nearly equal to or somewhat smaller than the field capacity”. After Teshima et al. (2006), parameter b is a measure of soil moisture recession that depends on hydraulic conductivity and thickness of active soil layer Z_a. In Iwanaga et al. (2005), a sensitivity analysis of BBH model applied to an irrigated area in semi-arid region suggests that error soil moisture is most sensitive to σ, η and c.

3.5 Discussion

According to the previous presentation and model comparison, bucket type models involves one parameter in Manabe model (W_k) up to six parameters in BBH (W_{max},a,b,c,σ, η). The minimum level of model complexity for bucket type models is discussed using a daily time step by Atkinson et al. (2002). These authors introduced the permanent wilting point θ_{pwp} to refine the bucket capacity $S_{bc} = (n-\theta_{pwp})Z_a$. Also, complexity is raised by the inclusion of a separation between transpiration and evaporation from bare soil. Hence a parameter which represents the fraction of basin area covered by forests is incorporated. A linear piecewise function is assumed similarly to Rodriguez-Iturbe et al. (1999) in both cases (bare soil areas and forest areas). They suppose that storage at field capacity S_{fc} is the bucket capacity S_{bc} scaled by a threshold storage parameter fc with $S_{fc} = fc\ S_{bc}$ and $fc = (\theta_{fc}- \theta_{pwp})/\ (n-\theta_{pwp})$ where θ_{fc} is volumetric water content corresponding to field capacity. In addition, they assume that saturation excess runoff occurs when the storage exceeds S_{bc} and that subsurface runoff occurs when the storage exceeds S_{fc} with a piecewise non linear drainage function involving two recession parameters. These parameters are further calibrated using observed discharge recession curves while the other parameters are adapted from soil properties (via field data interpretation). Under wet, energy limited catchments authors conclude that the threshold storage parameter fc has a little control on runoff. Conversely, under drier catchments they conclude that the threshold storage parameter fc controls runoff volumes. Either, Kalma & Boulet (1998) compared simulation results of the hydrological model VIC which assumes a bucket representation including spatial variability of soil parameters to the one dimensional physically based model SiSPAT. Using soil moisture profile data for calibration, they conclude that catchment scale wetness index for very dry and very wet periods are misrepresented by SiSPAT while VIC model may better capture the water flux near and by the land surface. However, they outlined that

the difficulty of physical interpretation of the bucket VIC model parameters (maximum and minimum storage capacity) constitutes a major drawbacks of the bucket approach.
Guswa et al. (2002) also compared simulations of Richards (1D) and daily bucket model for African Savanna. They outlined that the differences between models outputs are mainly in the relationship between evapotranspiration and average root zone saturation, timing and intensity of transpiration as well as uptake separation between transpiration and evaporation. Vrugt et al. (2004) as well compared the daily Bucket model to a 3-D model (MODHMS) based on Richards equation while taking into account drainage observations. They concluded that Bucket model results are similar to MODHMS results. They also noticed that physical interpretation of MODHMS parameters is difficult since they represent effective properties. Moreover it is noticed that soil control on evapotranspiration is important in dry conditions. Besides, the introduction of a threshold parameter for evapotranspiration uptake is much recommended under arid conditions. Else, according to Rodriguez-Iturbe et al. (1999) under dry conditions, the spatial variation in soil properties has very little impact on the mean soil moisture. DeMaria et al. (2007) analyzed VIC parameter identifiability using stream flows data. Classifying four basins according to their climatic conditions (driest, dry, wet, wettest) they concluded that parameter sensitivity was more strongly dictated by climatic gradients than by changes in soil properties.

4. Rainfall runoff hydrological models

Soil water balance represents a key component of the structure of many Rainfall-runoff (R-R) models. Rainfall-runoff models are primarily tools for runoff prediction for water infrastructure sizing, water management and water quality management. On the basis of rainfall and temperature information, they aim to simulate the water balance at local and regional scales often adopting daily time step. In the majority of cases, model structure is a conceptual representation of the water balance, model parameters having to be adjusted using climatic and soil information as well as hydrological data, in order to match model outputs to observed outputs (Wagener et al., 2003). R-R models have two main components: a soil moisture-accounting module (also named production function) and a routine module (also named transfer function). In the former, the soil moisture status is up-dated while in the latter the runoff hydrograph is simulated. Models differ by the sub-models which are used for each hydrological process in both modules. The way of computing infiltration, evapotranspiration and leakage is of amount importance in the moisture-accounting module which simulates the soil moisture dynamics. It is worth noting that the Rainfall-Runoff Modelling Toolkit (RRMT), developed at Imperial College offers a generic modeling covering to the user to help him (her) to implement different lumped model structures to built his (her) own model (http://www3.imperial.ac.uk/ewre/research/software/toolkit). The system architecture of RRMT is composed by the production and transfer functions modules, and either an off-line data processing module, a visual analysis module and optimization tools module for calibration purposes (Wagener et al. 2001). In this section, we focus on evapotranspiration sub-models of two well-used R-R models (section 4.1). Then, we review the main steps of the calibration process required to estimate the model parameters (section 4.2). Finally three case studies are reported (section 4.3).

4.1 Evapotranspiration sub models

Despite the focus on runoff results in R-R modeling, evapotranspiration computation is a key part of R-R models. As an example, we emphasize the evapotranspiration sub-model of

GR4 model which is a parsimonious lumped model proposed by CEMAGREF (France) and running at the daily step with four parameters. A full model description is available in (Perrin et al., 2003). At each time step, a balance of daily rainfall and daily potential evapotranspiration is performed. Consequently, a net evapotranspiration capacity E_n and a net rainfall P_n are computed. If $P_n \neq 0$, a part P_s of P_n fills up the soil reservoir (so, P_s represents infiltration). It is noticeable that this quantity P_s depends on the actual soil moisture content w according to a non linear decreasing function of the w/x_1 where x_1 is the maximum capacity of the reservoir soil (which might represent the field capacity). On the other hand, if the net evapotranspiration capacity $E_n \neq 0$, actual evapotranspiration E_s is computed as a non linear increasing function of the water content involving the ratio w/x_1. Also, this function is parameterized through the ratio E_n/x_1 which refers to the characteristics of climate-soil complex. Furthermore, a leakage component is assumed with a power law function of the reservoir water content w.

$$\text{For } P \geq E_0; \quad P_n = P - E_0 \quad \text{and} \quad E_n = 0 \tag{38}$$

$$\text{For } P < E_0; \quad P_n = 0 \quad \text{and} \quad E_n = E_0 - P \tag{39}$$

$$E_s = w\,(2-(w/x_1))\tanh(E_n/x_1)/\{1+[(1-wx_1)\tanh(E_n/x_1)]\} \tag{40}$$

Where tanh(.) stands for hyperbolic tangent.

As second example, we underline the sub-models adopted in the *HBV* conceptual semi-distributed model proposed by the Swedish hydrological institute (Begström, 1976). The fraction ΔQ of precipitation entering the soil reservoir is assumed as power law function of the ratio (w/FC) of reservoir water content w to a parameter FC representing soil field capacity in HBV model.

$$\Delta Q = P_e[1-(w/FC)^{\beta'}] \tag{41}$$

Where β' is a calibration parameter usually estimated by fitting observed and simulated runoff data. Also, P_e is effective precipitation. In addition, the actual evapotranspiration is a piecewise linear function. The control of actual evapotranspiration rates is performed using a parameter PWP representing a threshold water content. If w< PWP, the evapotranspiration uptake is a fraction of the potential evapotranspiration E_0 otherwise it is at potential rate.

$$\begin{aligned} E_a/E_0 &= w/PWP \text{ for } w<PWP; \\ \text{and } E_a &= E_0 \text{ for } w>PWP \end{aligned} \tag{42}$$

4.2 Model calibration issues

As runoff has been for long time the main targeted response of rainfall-runoff modeling, rainfall-runoff models were often adjusted according to runoff observations. So far, observations from other control variables such as soil moisture content (Lamb et al., 1998), water table levels (Seibert, 2000) and either low flows (Dunne, 1999) have been adopted to enhance runoff predictions. Calibration of model parameters against runoff data is often performed using criteria such as bias and Root Mean Square Error (RMSE), which helps quantifying the discrepancy between observed discharges y_0 and simulated discharges y_i over a fixed time period with N observations.

$$\text{RMSE}=\left(\frac{1}{N}\sum_{i=1}^{i=N}\left(y_{si}-y_{oi}\right)^2\right)^{1/2} \tag{43}$$

The difficulty in the calibration process is that various parameter sets and even model structures might result in similarly good levels of performance, which constitutes a source of ambiguity as out pointed by Wagener et al. (2003) and many other authors before them (see the literature review of Wagener et al. (2003)). Also, it is noticeable that this problem of ability of various model structures and model parameters to perform equal quality with respect to matching observations is not dependent of the calibration process itself. In other words, the use of a performing optimisation tool does not prevent the problem. Another question is related to the single versus multi objective optimization. Wagener et al. (2003) reported that "single objective function is sufficient to identify only between three and five parameters" while lumped R-R models usually adopt far superior number of parameters. Multi-objective approach of calibration using additional output variables such as water table levels or soil moisture observations has been introduced to deal with the problem. Yet, inadequate model structure may be responsible of mismatching between observed and simulated outputs, as related by Boyle et al. (2000).

4.3 Case studies

Three case studies are presented in this section. In the first case, we propose a method for calibrating the empirical parameter k' of Hsuen Chun (1988) (Eq. 28). In the second case, we propose as example of calibrating HBV model using both runoff data and regional evapotranspiration information. In the third case, calibration of BBH model is performed using both runoff data and regional evapotranspiration information.

4.3.1 Fitting empirical models of regional evapotranspiration

This case study is presented in Bargaoui et al. (2008) and Bargaoui & Houcine (2010). It is aimed to calibrate the H.C. model using climatic, rainfall and runoff data from gauged watersheds. Monthly temperature and solar radiation data as well as annual rainfall and runoff data from various locations in Tunisia listed in Table 1 are adopted to calibrate the parameter k' of the empirical Hsuen Chen model (Eq. 28). To this end, 18 rainfall stations and 20 river discharge stations are considered, as well as 8 meteorological stations (Table 1).
On the other hand, the potential evapotranspiration E_0 is computed at monthly scale using Turc formula.

$$E_0 = 0.4\ T_m\ [(R_g/N_j\ )+50]\ /\ [Rg+15] \tag{44}$$

T_m : monthly average temperature in (°C); R_g : global solar radiation (cal.cm^{-2} month^{-1}); N_j : number of days by month
For each river basin, simulated average (long term) annual evapotranspiration is computed using Eq. (28). Then, simulated mean annual runoff is computed as the difference between observed mean annual precipitation and simulated average annual evapotranspiration. The fitting of annual simulated runoff to annual observed runoff using the 20 river discharge stations results in k'= 1.5. The good adequacy of the model is well reflected in the plot of average simulated versus average observed annual runoff (Fig. 1).

River discharge stations			Rainfall stations			Meteorological stations		
Stations	Latitude	Longitude	Stations	Latitude	Longitude	Stations	Latitude	Longitude
Jebel Antra	36°57′18′′	9°27′45′′	Ouchtata	36°57′53′′	8°60′1′′	Sfax	34°43′0′′	10°41′0 ′′
Joumine Mateur	37°2′19′′	9°40′56′′	Cherfech	36°57′0′′	10°3′13′′	Tunis	36°51′0′′	10°20′0′′
Zouara	36°54′15′′	9°7′1′′	Tabarka	36°56′59′′	8°44′50′′	Tabarka	36°57′0′′	8°45′0′′
Barbara	36°40′32′′	8°32′56′′	El Kef	36°10′53′′	8°42′57′′	Bizerte	37°14′0′′	9°52′0′′
Rarai sup.	36°27′36′′	8°21′20′′	Mellègue	36°7′16′′	8°30′2′′	Jendouba	36°29′0′′	8°48′0′′
Mellegue K13	36°7′1′′	8°29′52′′	Tajerouine	36°27′32′′	9°14′57′′	El Kef	36°8′0′′	8°42′0′′
Mellegue Rmel	36°1′1′′	8°37′14′′	Mejez El Bab	36°39′3′′	9°36′17′′	Kairouan	35°4′0′′	10°4′0′′
Haffouz	35°37′58′′	9°39′33′′	Tunis	36°47′23′′	10°10′23′′	Siliana	36°4′0′′	9°22′0′′
Merguellil Skhira	35°44′24′′	9°23′3′′	Feriana	34°56′49′′	8°34′29′′			
Chaffar	34°33′49′′	10°29′14′′	Jendouba	36°30′14′′	8°46′52′′			
Joumine Tine	36°58′3′′	9°43′2′′	Sejnane BV	37°3′35′′	9°14′46′′			
Miliane, Tuburbo Majus	36°23′39′′	9°54′43′′	Ksour	36°45′22′′	9°28′27′′			
M′khachbia aval	36°43′22′′	9°24′24′′	Sers	36°4′19′′	9°1′25′′			
Haidra Sidi Abdelhak	35°56′59′′	8°16′22′′	Ghardimaou	36°27′2′′	8°25′58′′			
Medjerda Jendouba	36°30′40′′	8°46′7′′	Bou Salem	36°36′30′′	8°57′57′′			
Sejnane	37°11′37′′	9°30′16′′	Merguellil H.	35°38′8′′	9°40′36′′			
Tessa Sidi Medien	36°16′44′′	8°57′14′′	Merguellil Skhira	35°44′24′′	9°23′3′′			
Rarai plaine	36°29′16′′	8°32′18′′	Chaffar PVF	34°40′0′′	10°5′0′′			
Ghezala-Ichkeul	37°4′35′′	9°32′12′′						
Douimis	37°12′50′′	9°37′38′′						

Table 1. Location of stations to calibrate H.C. model (after Bargaoui &Houcine, 2010)

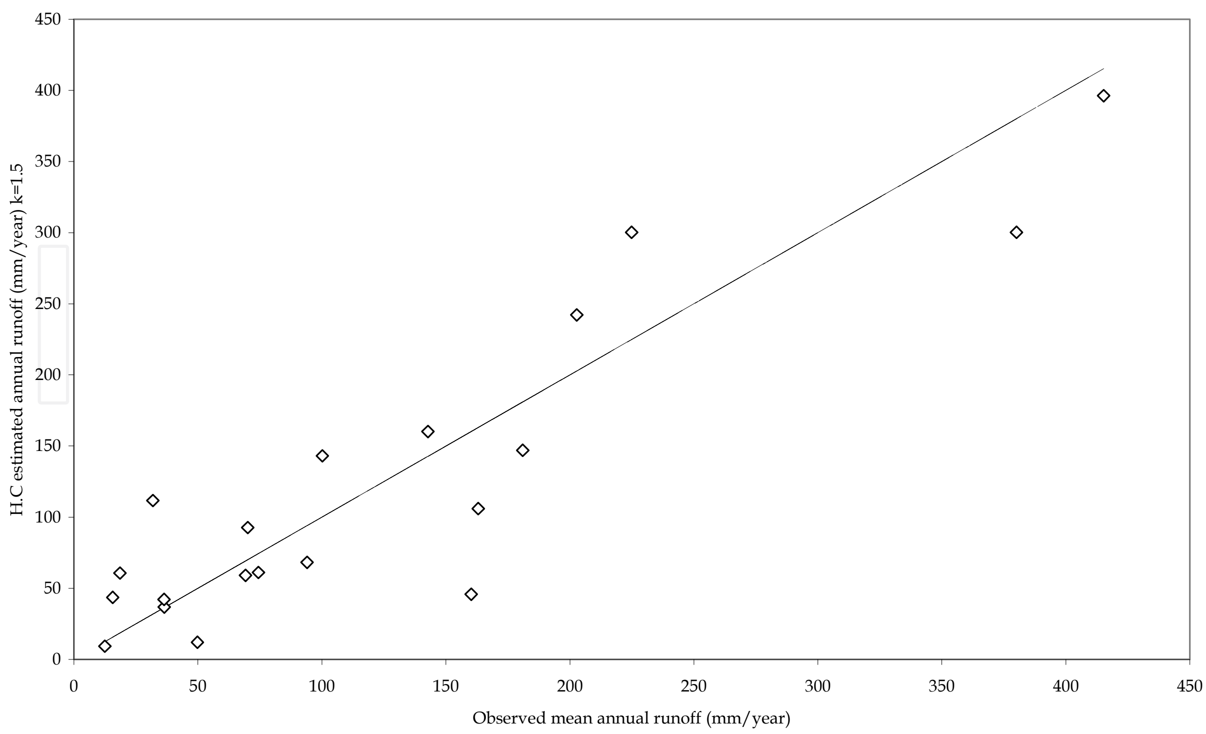

Fig. 1. Comparison of observed and simulated runoff for 20 river basins

4.3.2 Multicriteria calibration of HBV model using regional evapotranspiration information

This application is presented in Bargaoui et al. (2008). The idea is to use the information about the climatic regime as a driver for runoff prediction. Effectively, for a large number of basins with areas in the interval 50 à 1000 km², Wagener et al., (2007) suggested that there is a significant correlation between annual runoff and the ratio of forcing variables P/E_0. In the same way, we seek to use information about average (regional) actual evapotranspiration which is a bio-climatic indicator as means to improve accuracy of runoff predictions. To develop these ideas, the HBV rainfall-runoff model was adopted, coupled to a SCE-UA optimization tool. The calibration method adopts an objective function combining three criteria: minimisation of runoff root mean square error, minimisation of water budget simulation error, minimisation of the difference between mean annual simulated evapotranspiration E_a and regional E_m. The case study is a mountainous watershed of Wadi Sejnane (Tunisia). Mean daily runoff observations from September 1964 to August 1969 are available for a hydrometric station controlling an area of 378 km². Average basin annual rainfall is 931 mm/year. Over 8 years of rainfall observations, the minimum value of the series is 628 mm/year while the maximum value is 1141 mm/year denoting an important rainfall inter annual variability. Mean annual discharge is 2.43 m³/s. Average evapotranspiration computed using HC model (Eq. 28) with k'=1.5 results in E_m=643 mm/year. To calibrate the HBV model parameters, we adopt the period 1964/1967 for calibration and the period 1967/1969 for validation. The minimization of the objective function is performed using SCE-UA algorithm (Duan et al., 1994) in order to adjust 10 parameters (while 7 other HBV parameters have been set constant because they were found insensitive). First, the Nash coefficient of mean daily discharges is chosen as objective function F_0=$Nash_R$. The resulting value F_0=0.81 is quite good. However, for the validation period the ensuing optimal parameter set results in very poor fitting with a negative value of the Nash coefficient ($Nash_R$ = -0.084). Consequently, the objective function was modified to F_1 integrating the average model error (bias) of runoff output. Hence,

$$F_1 = Nash_R - w' \, ER_{RA} \tag{45}$$

Where ER_{RA} is the absolute relative error with respect to annual discharge. The weight coefficient w' = 0.1 is adopted according to Lindström and al. (1997) and helps aggregate the two criteria $Nash_R$ and ER_{RA}. In fact, the adoption of ER_{RA} aims to consider climatic zonality during the calibration process. Resulting optimal solution corresponds to $Nash_R$ =0.81 and ER_{RA} = 5%, which is believed good performance. It is worth noting that this modification of the objective function greatly improved $Nash_R$ also for the validation period ($Nash_R$ =0.55). The mean annual simulated evapotranspiration using HBV model is equal to 728 mm/ year while the H.C. model with k'=1.5 results in 643mm/year. To try to overcome such overestimation, it was proposed to directly include the information about evapotranspiration by adopting a new objective function F_2.

$$F_2 = Nash_R - 0{,}1 \, ER_{RA} - 0{,}1 \, ER_{ETRG} \tag{46}$$

Where ER_{ETRG} is the absolute relative error with respect to mean annual evapotranspiration (simulated by HBV versus estimated by H.C with k'=1.5). The resulting runoff Nash is a little smaller ($Nash_R$ =0.79) than for F_1, but a real improvement is obtained during the validation period ($Nash_R$ = 0.68). Fig. 2 reports HBV estimated annual evapotranspiration obtained with the optimal HBV solution (squares) versus annual rainfall. Comparatively, we also report annual evapotranspiration as evaluated using H.C model with k'=1.5 (interrupted line). Effect of year to year rainfall fluctuation on HBV estimations is well seen in the graph.

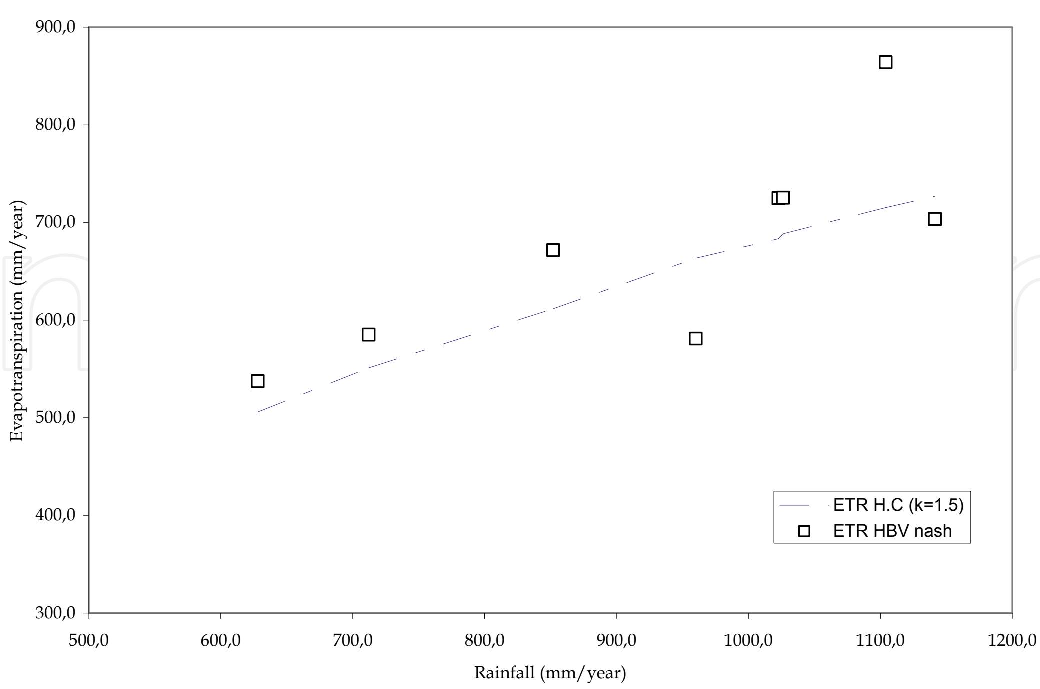

Fig. 2. Comparison of evapotranspiration estimates from HBV and HC models in relation with rainfall

4.3.3 Multicriteria calibration of BBH model using regional evapotranspiration information

In the third application it is aimed to compare BBH model results using the decadal time step. A part of this case study is presented in Bargaoui & Houcine (2011) using monthly data for model evaluation. Here will report results of decadal evaluations. Data are from the Wadi Chaffar watershed (250 km^2) situated under arid climate, South Tunisia. Vegetation cover comprises mainly olives. Meteorological data (solar radiation, air temperature and humidity, sky cloudiness, wind speed and Piche evaporation) are available from September 1989 to August 1999 for computing the daily reference evapotranspiration E_0 according to Allen et al. (1998). E_0 is multiplied by the crop coefficient K_c of olives trees to obtain daily potential evapotranspiration (Allen et al., 1998). Daily average basin rainfalls are available from September 1985 to August 1999. Stream discharge data are available for the basin outlet at the daily time step from September 1985 to August 1999. In the period September 1985 to August 1989, meteorological data are missing and the used E_0 values are the daily long term average computed for September 1989- August 1999. The H.C. model results in an average annual evapotranspiration E_m = 213 mm/year (Bargaoui & Houcine, 2010). BBH model inputs are precipitation and potential evapotranspiration and seven parameters are to be calibrated. To reduce the number of calibrated parameters, we first fix the thickness of active soil layer Z_a (in mm) and the effective soil porosity n (unit less). Also, we undertake a reformulation of leakage component L(s) by using the model of Guswa et al. (2002) where

$$L(s) = K_S \frac{e^{B(s-S_{FC})} - 1}{e^{B(1-S_{FC})} - 1} \tag{47}$$

where s is the degree of saturation (unit less); K_S saturated hydraulic conductivity at soil surface (mm/day); B is the soil water retention curve shape parameter; S_{FC} (unit less) is the field capacity; $W_{max} = nZ_a$ (W_{max} is the total water-holding capacity in mm).

Coupling this expression with pedo-transfer functions it makes it possible after Bargaoui & Houcine (2010), to derive the parameters (a, b, c) as following using pedo-transfer parameters K_S , B and S_{FC}:

$$a = W_{\max}\left[S_{FC} - \frac{1}{B} Ln\left(K_S \frac{1}{e^{B(1-S_{FC})} - 1}\right)\right] \tag{48}$$

$$b = W_{\max}\frac{1}{B} \tag{49}$$

$$c = \left(\frac{1}{e^{B(1-S_{FC})} - 1}\right) K_S \tag{50}$$

In this case, the model by Rawls et al. (1982) is adopted for K_S estimation while S_{FC} is derived according to the Cosby (1984) and Saxton et al. (1986) models recently adopted by Zhan et al., (2008). Finally B = 9 is assumed in agreement with Rodriguez-Iturbe et al. (1999). The dominant soil type is considered to represent the soil characteristics. So, the value n=0.34 corresponding to a sandy soil was adopted; these assumptions result in K_S = 3634 mm/d and S_{FC}= 0.166. Also, after many trials the value Z_a= 0.5 m was adopted. The two remaining parameters σ and η ($0< \sigma <1$; $0< \eta <1$) represent respectively the resistance of vegetation to evapotranspiration and the moisture retaining capacity. The problem is now to fit the parameters σ and η. They are adjusted using two different methods: i.e. using only observed runoff (method 1) and using both observed runoff and regional evapotranspiration information (method 2). Also BBH model has been completed adopting a , contributing area sub-model (Betson, 1964); Dunne et Black (1970). According to this assumption, runoff originates from a part of the watershed (contributing area) contrarily to the assumption of runoff occurring from the entire watershed. For a fixed day j, the contributing area CA_j is herein assumed linked to the soil moisture content according to Dickinson & Whiteley (1969). Additionally, a logistic Probability Density Function as a function of humidity index IH_j is adopted with parameters a_c and b_c (Eq. 51). It means that the mean contributing area is a_c and that the variance of the contributing area is $(b_c\pi)^2/3$. The humidity index takes account for the rainfall accumulated during the actual day and the IX previous days (Eq. 52).

$$CA_j = \frac{e^{((IH_j - a_c)/b_c)}}{(1+e^{((IH_j - a_c)/b_c)})} \tag{51}$$

$$IH_j = W_{j-1} + \omega'' \sum_{l=0}^{IX} P_{j-l} \tag{52}$$

where ω'' is a fixed weight (ω'' =0.1). Then, two cases are considered: case (a) when the total basin area contributes to runoff at the basin outlet; case (b) when only a contributive area gives rise to runoff at the outlet.
After many trials and errors we assumed IX= 90 days, a_c = 20 and b_c = 10 in case (b). The model was calibrated for σ and η using daily hydro meteorological data (solar radiation, air temperature, air humidity, mean areal rainfall) as well as daily runoff records and also average annual evapotranspiration. The decadal, monthly and annual totals are adopted to evaluate model performance.
In each case (a) and (b), a first criterion based on the matching of decadal runoff (Eq. 53) is adopted to delineate adequate solutions for σ and η ($0< \sigma <1$; $0< \eta <1$). A supplementary criterion is based on the matching of long term annual evapotranspiration (Eq. 54).

$$C_y(\sigma,\eta) = \frac{1}{N}\sum_{i=1}^{N}\left|(y_{si}-y_{oi})/y_{oi}\right| \tag{53}$$

$$C_E(\sigma,\eta) = \frac{1}{N'}\sum_{i=1}^{N'}\left|(E_{si}-E_m)/E_m\right| \tag{54}$$

In Eq. (53), y_{oi} and y_{si} are respectively decadal observed and simulated volume runoff and N is the number of simulated decades. In Eq. (54), E_{si} is simulated annual evapotranspiration and N' is the number of simulated years.
For each pair of simulated (σ, η) ($0< \sigma <1$; $0< \eta <1$), candidate solutions verifying the criterion C_y (σ, η) < α (Eq. 53) with α=20% the Nash coefficient R_N is then evaluated. Pairs for which it is found that R_N>0.5, are thus selected. Also, introducing E_m for calibration method 2, the absolute value C_E (σ, η) of the relative error between mean annual simulated evapotranspiration and E_m, is used through the additional selection criterion of (Eq. 54).

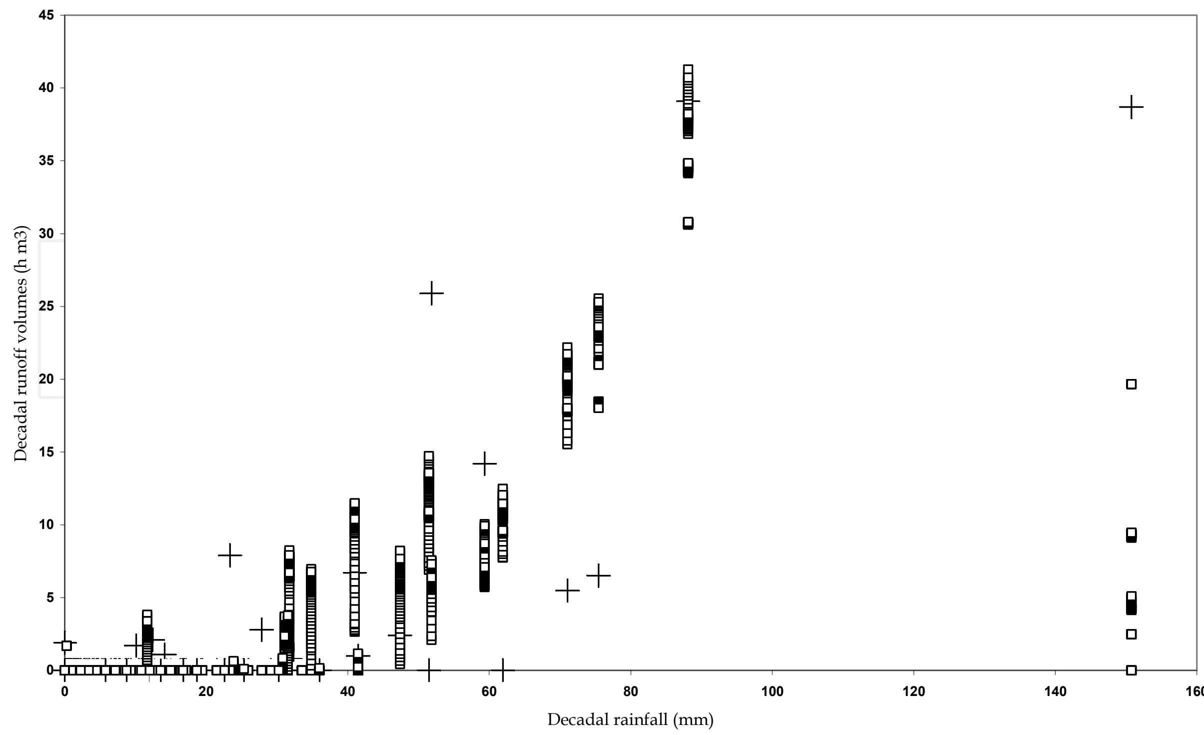

Fig. 3. Estimated decadal runoff versus decadal precipitation with the assumption of total watershed contributing to runoff (+ represent observed volumes and squares represent simulated volume for the selected pairs of (σ, η))

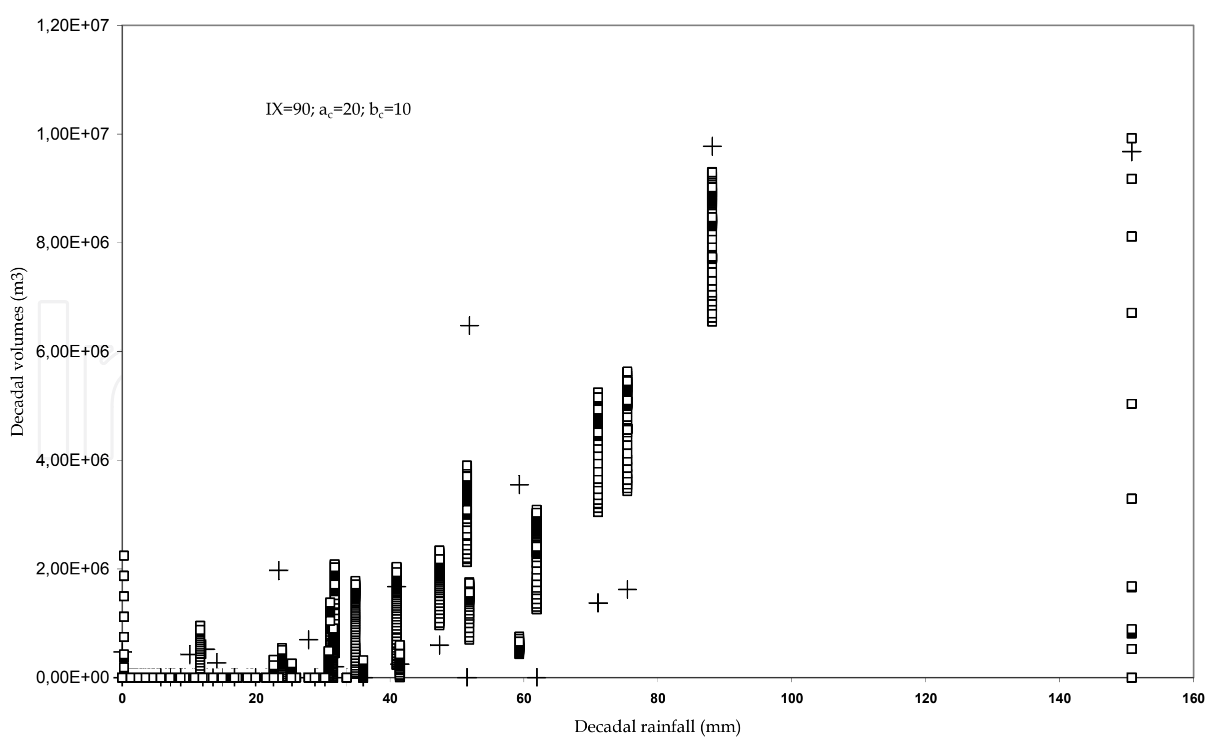

Fig. 4. Estimated decadal runoff versus decadal precipitation with the assumption of contributive area (+ represent observed volumes and squares represent simulated volume for the selected pairs of (σ, η)).

Pairs of simulated (σ, η) ($0< \sigma <1$; $0< \eta <1$) which satisfy both $C_y(\sigma, \eta) < 20\%$; $R_N > 0.5$ and $C_E(\sigma, \eta) < \alpha'$ with $\alpha'=30\%$are finally selected as adequate solutions.
Fig. 3 and 4 report model outputs for sets of (σ, η) fulfilling the above conditions under the assumptions of cases (a) and (b) in case where E_m information is included. Estimated decadal volumes (squares) for the selected pairs of (σ, η) are compared to observed decadal volumes (+) and are reported versus precipitation data. Fig. 3 is related to case (a) corresponding to the assumption of total watershed contributing to runoff. Fig. 4 is related to case (b) assuming a contributing area. The results suggest that the introduction of contributing area outcomes produce outputs which result in a better fitting of the rainfall-runoff evolution. In effect, in the case of total area contributing no solution is found able to simulate the most rainy decade,(squares are far from the symbol + for the Rainiest decade). Conversely, some solutions are found able to reproduce the most rainy decade if we consider contributing area scheme (some squares are located near the +). Also, evapotranspiration information has greatly reduced the interval of acceptable solutions. Effectively, selected solutions are such that $0.15 < \sigma <0.35$ and $0.15 < \eta <0.25$.

5. Conclusions

The simulation of evapotranspiration using the water balance equation is part of hydrological modelling (rainfall-runoff models) and is also important in the framework of global circulation models (Land surface models). A lot of models are now functioning and their formulation is based on different assumptions on soil characteristics in relation with soil moisture, transpiration schemes, as well as infiltration and runoff schemes.
Empirical models for estimating regional evapotranspiration are worth noting for estimating average long term evapotranspiration. They are generally based on climatic information (rainfall and potential evapotranspiration). They often require the adjustment of a single empirical parameter. Under particular climate and soil vegetation, evapotranspiration is controlled by soil moisture dynamics. Thus, Bucket type soil water budget models are worth noting for estimating time series of actual evapotranspiration at smaller time scales (daily to monthly). They involve from one parameter such as in the Manabe model (with parameter W_k) up to six parameters such as in BBH model (with parameters W_{max},a,b,c,σ, η). Parameters are linked to soil, climatic and vegetation characteristics. However, it is generally believed that the temporal variability of soil moisture series is mostly dependent on the rainfall variability especially under conditions of low precipitations. On the other hand, soil parameters such as field capacity, hydraulic conductivity at saturation and wilting point potential are key parameters controlling the evapotranspiration model outputs. One way to derive soil parameters is to adopt pedo transfer functions. Transpiration which corresponds to vegetation uptake is regulated by stomata and driven by atmospheric demand. It is widely represented by a linear piecewise function with parameters depending on vegetation characteristics. Thus, in computing evapotranspiration, a main assumption is the linear piecewise function of evapotranspiration in relation with potential evapotranspiration for taking account for soil water stress. Such an assumption is underlined in several rainfall runoff models (for example the two models GR4 and HBV studied here adopt such analytical form). Model adequacies introduce the question of the choice of the objective function as well as the output variables adopted for model

evaluation. In the case studies presented here, results suggest that the introduction of the information about average (long term) annual evapotranspiration may help improving the accuracy of the water balance simulation results. In effect the runoff Nash coefficient is found to be improved during the validation period in the case where long term evapotranspiration is taking account during the calibration process.

6. Annexe

6.1 Glossary

a: parameter related to the field capacity (mm)
a' : pore size distribution parameter
a_c: Logistic density distribution parameter
$\alpha(\theta)$: the root efficiency function.
A(t) : the fraction of area for which the infiltration capacity is less than i(t)
B : the soil water retention curve shape parameter;
b: parameter representing the decay of soil moisture (mm);
b'' is the infiltration shape parameter.
b_c: Logistic density distribution parameter
β: surface slope angle
β' : a calibration parameter in HBV model
c' : pore disconnectedness index
c: parameter representing the daily maximal capillary rise (mm)
$C_{\%}$: soil percent of clay
CA_j: : the contributing area
cos: the cosinus function
d_1 : depth of near surface soil layer
d_2 : depth of root zone soil layer;
D:soil-water diffusivity parameter
$\Delta\psi$: difference in average matrix potential before and after wetting
$\Delta\theta$: difference in average soil water content before and after wetting
Δw : the change in soil water content
Δz : soil depth.
C_1 : parameter,
C_2 : parameter,
E(s,t): : evapotranspiration
E_a : actual evapotranspiration
EDI : dryness index
E_e exfiltration parameter as function of initial degree of saturation s_0
E_g : bare soil evaporation rate at the surface
E_m average annual evapotranspiration
E_n: net evapotranspiration capacity
E_r: reference evapotranspiration according to FAO model
E_{tr} : transpiration rate from the root zone of depth d_2
ER_{ETRG} : the absolute relative error with respect to mean annual evapotranspiration
E_{pa} average annual potential evapotranspiration

ER_{RA} : the absolute relative error with respect to annual discharge
E_{Turc} : monthly potential evapotranspiration (mm);
E_{ra} average annual surface retention
f : infiltration capacity
F cumulative infiltration for a rainfall event
fc: threshold storage parameter
FC: representing soil field capacity in HBV model
Gd(t): Daily percolation and capillary rise term
$g_r(\theta,z)$: vegetation uptake of soil moisture
I(s,t): infiltration into the soil
$I_{nf}(\theta,z_0)$: precipitation infiltrating into the soil
i(t): infiltration capacity at time t.
i_{max} : maximum value of infiltration capacity
Ω_f (t) : the cumulative infiltration
J(.): evapotranspiration function
k : intrinsic permeability
k(1): intrinsic permeability at saturation
K: hydraulic conductivity
K (1) hydraulic conductivity at saturation
k_v : plant coefficient
K_{av} : average saturated hydraulic conductivity
k' : parameter of HC model
K_c : crop coefficient
K_S : the saturated hydraulic conductivity;
K'_s : correction coefficient of the crop coefficient
κ: shape parameter of the Gamma distribution
l: factor linked to soil matrix tortuosity
L(s,t) :leakage
LAI : Leaf area index
λ: mean storm arrival rate
M_v : vegetation fraction of surface.
μ : dynamic viscosity of water;
n: soil effective porosity
ν : parameter
θ: volumetric water content
θ_f :the volumetric moisture contents of the soil at field capacity
θ_w: the volumetric moisture contents at wilting point
θ_{pwp}: permanent wilting point
θ_g : volumetric water contents of near surface soil layer;
θ_s : saturated soil moisture content
θ_2 : volumetric water contents of root zone soil layer;
θ_{geq} : equilibrium surface volumetric soil moisture content
θ_1 : specific value of soil moisture content
θ_0 : specific value of soil moiqture content
N: number of observations

N_j : number of days by month
$Nash_R$: Nash coefficient of mean daily discharges
P : average annual precipitation
P_e : effective precipitation
PWP : parameter representing a threshold water content in HBV model.
P_g : precipitation infiltrating into the soil;
P_n: net rainfall
q_2 : rate of drainage out of the bottom of the root zone;
R_1 : (s cm^{-1}) a resistance to moisture flow in soil
R_2 : (s cm^{-1}) is vegetation resistance to moisture flow;
R_n : average annual net radiation
R_s : surface runoff
R_g : global solar radiation (cal.cm^{-2} $month^{-1}$)
r(z) : a root density function (cm^{-1})
ρ_w : density of the water;
s: relative soil moisture content or degree of saturation
s*: saturation threshold
s_1 :threshold value of soil saturation
s_w: soil moisture at wilting point.
s_0: initial degree of saturation
S : sorptivity
$S_{\%}$: soil percent of sand
S_{bc}: bucket capacity
S_{FC} : soil field capacity
S_{fc}: storage at field capacity
σ : parameter representing the resistance of vegetation to evapotranspiration;
t: time
T_m : monthly average temperature in (°C);
u(z,t) : local transpiration uptake
w : the actual soil water storage
w_0 : critical soil water storage in Budyko model
W_0: water holding capacity
W_k : a fraction of the soil field capacity
W_{max} : total water-holding capacity (mm);
W_{m0}: mean water holding capacity
ω : a fixed weight
ψ : soil moisture potential (bars)
ψ_p : leaf moisture potential (bars)
$\psi*$: the wilting point potential
$\psi(1)$: the bubbling pressure head which represents matrix potential at saturation.
x_1 : maximum capacity of the reservoir soil
y_i: simulated discharges
y_0: observed discharges
z: the vertical coordinate (z>0 downward from surface)
Z_a: thickness of active soil layer (mm);
z_0 : the vertical coordinate at the surface

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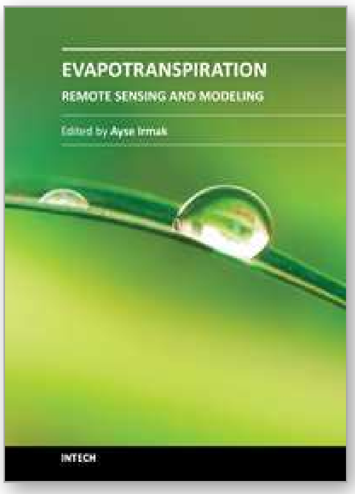

Evapotranspiration - Remote Sensing and Modeling

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This edition of Evapotranspiration - Remote Sensing and Modeling contains 23 chapters related to the modeling and simulation of evapotranspiration (ET) and remote sensing-based energy balance determination of ET. These areas are at the forefront of technologies that quantify the highly spatial ET from the Earth's surface. The topics describe mechanics of ET simulation from partially vegetated surfaces and stomatal conductance behavior of natural and agricultural ecosystems. Estimation methods that use weather based methods, soil water balance, the Complementary Relationship, the Hargreaves and other temperature-radiation based methods, and Fuzzy-Probabilistic calculations are described. A critical review describes methods used in hydrological models. Applications describe ET patterns in alpine catchments, under water shortage, for irrigated systems, under climate change, and for grasslands and pastures. Remote sensing based approaches include Landsat and MODIS satellite-based energy balance, and the common process models SEBAL, METRIC and S-SEBS. Recommended guidelines for applying operational satellite-based energy balance models and for overcoming common challenges are made.