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Critical Review of Methods for the Estimation of Actual Evapotranspiration in Hydrological Models

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1. Introduction

The quantification of a catchment water balance is a fundamental requirement in the assessment and management of water resources, in particular under the impacts of human-induced land use and climate changes. The description and quantification of the water cycle is often very complex, particularly because of the spatial and temporal dimensions, variabilities and uncertainties inherent to the system. The advent of powerful computers, numerical modelling, and GIS is making it possible to describe the complexities of hydrological systems with statistically acceptable accuracy (Duan et al., 2004). Both local (e.g. on-farm) and catchment scale models, physically-based numerical models and simple conceptual balance models are now available to support water resource assessment, management, allocation as well as adaptation to climate change. In particular, the coupling of dedicated atmospheric, hydrological, unsaturated zone and groundwater models is becoming a powerful means of evaluating and managing water resources.

Evapotranspiration (ET) is a key process of the hydrological balance and arguably the most difficult component to determine, especially in arid and semi-arid areas where a large proportion of low and sporadic precipitation is returned to the atmosphere via ET. In these areas, vegetation is often subject to water stress and plant species adapt in different ways to prolonged drought conditions. This makes the process of ET very dynamic over time and variable in space. The focus of this chapter is on the methodologies used in hydrological models for the estimation of actual ET, which may be limited (adjusted) by water or other stresses. The chapter includes: i) a theoretical overview of ET processes, including the principle of atmospheric evaporative demand-soil water supply; ii) a schematic review of methods and techniques to measure and estimate ET; and iii) a review of methods for the estimation of ET in hydrological models.

2. Theoretical overview of evapotranspiration processes

ET is the combination of two separate processes, where liquid water is converted to water vapour (vaporization) from the soil, wet vegetation, open water or other surfaces, as well as from plants by transpiration through stomata (Allen et al., 1998). Evaporation and transpiration occur simultaneously and they are difficult to separate out. ET rate depends on weather conditions, water availability, vegetation characteristics, management and environmental constraints. The main weather variables affecting ET are temperature, solar radiation, wind speed and vapour pressure. The nature of the soil, its hydraulic properties and water retention capacity determine plant available water. Under natural conditions, water stored in the soil is replenished through precipitation, surface and groundwater. The type and developmental stage of vegetation, its adaptation to drought, structure and roughness, albedo, ground cover, root density and depth also affect ET rates. ET rates can be managed through different tillage practices, the establishment of windbreaks, different planting densities and thinning of vegetation, by reducing soil evaporation using, for example, localized irrigation targeting the root zone or mulching, and by reducing transpiration with herbicides or anti-transpirants (substances that induce closing of stomata, envelop vegetation with a surface film or change its albedo). Besides water stress, vegetation may be subject to other types of environmental stresses that are likely to result in a reduction of ET rates and plant growth, like for example pests, diseases, nutrient shortages, exposure to toxic substances and salinization (Allen et al., 1998).

Reference ET is the evaporation from a reference surface of the Earth and it depends on weather conditions. The reference surface can be an open water surface (open pan) or it can be related to weather variables (temperature, radiation, sunshine hours, wind speed, air humidity etc.). Many semi-empirical equations exist that relate reference ET to weather variables. Some of the most commonly adopted are Blaney-Criddle (Blaney and Criddle, 1950), Jensen-Haise (Jensen and Haise, 1963), Hargreaves (1983) and Thornthwaite (1948). Lu et al. (2005) compared the performance of three temperature-based methods, namely Thornthwaite (1948), Hamon (1963) and Hargreaves-Samani (1985), and three radiation-based methods, namely Turc (1961), Makkink (1957) and Priestley-Taylor (1972) for application in large scale hydrological studies in the south-eastern United States. Similarly, Oudin et al. (2005) tested the performance of 27 reference ET models in rainfall-runoff modelling of catchments located in France, Australia and the United States. Both Lu et al. (2005) and Oudin et al. (2005) proposed simple temperature-based methods for calculation of reference ET at catchment scale, in particular when availability of weather data sets is limited.

Theoretical equations that describe the mechanisms of the evaporation process are also available. For example, reference evaporation from an open water surface was first described by Penman (1948) and consisted of a radiation and a vapour pressure deficit term, representing the available energy for the endothermal evaporation process. Priestley and Taylor (1972) proposed the Priestley-Taylor equation, where the radiation term dominates over the advection term by a factor of 1.26, suitable for large forest catchments and humid environments. Based on decades of data and knowledge gained, the FAO (United Nations Food and Agricultural Organization) proposed a grass reference evapotranspiration (ETo) calculated with the Penman-Monteith equation (Monteith, 1965). The FAO Penman-Monteith ETo is defined as the evapotranspiration rate from a reference surface not short of water. The reference surface is a hypothetical grass reference crop with an assumed height of 0.12 m, a fixed surface resistance of 70 s m^{-1} and an albedo of 0.23 (Allen et al., 1998). The Penman-Monteith ETo is a function of the four main factors affecting evaporation, namely temperature, solar radiation, wind speed, and vapour pressure:

$$ETo = \frac{\frac{\Delta}{\lambda}\left(R_n - G\right) + \frac{\rho_a C_p}{\lambda}\frac{e_s - e_a}{r_a}}{\Delta + \gamma\left(1 + \frac{r_s}{r_a}\right)} \tag{1}$$

where λ is the latent heat of vaporization of water (MJ kg^{-1}); Δ is the gradient of the saturation vapour pressure-temperature function (kPa $°C^{-1}$); R_n is the net radiation (MJ m^{-2} d^{-1}); G is the soil heat flux (MJ m^{-2} d^{-1}); ρ_a is the air density (kg m^{-3}); C_p is the specific heat of the air at constant pressure = 1.013 kJ kg^{-1} K^{-1}; e_s is the saturated vapour pressure of the air (kPa), a function of air temperature measured at height z; e_a is the mean actual vapour pressure of the air measured at height z (kPa); r_a is the aerodynamic resistance to water vapour diffusion into the atmospheric boundary layer (s m^{-1}); γ is the psychrometric constant (kPa $°C^{-1}$); and r_s is the vegetation canopy resistance to water vapor transfer (s m^{-1}). Equation (1) uses standard climatic data that can be easily measured or derived from commonly collected weather data. Allen et al. (1998) also recommended procedures for the calculation of missing variables in equation (1).

In equation (1), the type of vegetation is accounted for through canopy resistance to gas exchange fluxes (r_s), vegetation height determining surface roughness (implicitly in r_a) and albedo (implicitly in R_n). Theoretically, the Penman-Monteith equation allows for direct calculation of actual ET through the introduction of canopy and air resistances to water vapour diffusion. However, this one-step approach is difficult to apply because canopy and air resistances are not known for many plant species and they are complex to measure. A two-step approach is then commonly used to determine actual ET, where the potential evapotranspiration (PET) is first calculated using a minimum value of canopy resistance for a specific crop/vegetation and the actual air resistance from weather data and vegetation height. In a second step, actual ET is calculated taking into account reduction in root water uptake due to water (and/or other) stress and reduction in soil evaporation due to drying of the top soil.

ET of crops or other vegetation differs distinctly from ETo because the ground cover, canopy properties, physiological adaptation and aerodynamic resistance of vegetation may be different from grass. These differences can be integrated into a factor Kc, commonly known as the crop coefficient because it is used to calculate crop water requirements (Allen et al., 1998). The FAO-56 model (Allen et al., 1998) provides a means of calculating reference and crop ET from meteorological data and crop coefficients. The effect of climate on crop water requirements is given by the reference evapotranspiration (ETo), and the effect of the crop by the crop coefficient Kc. Crop evapotranspiration under standard conditions (ETc) is the evapotranspiration from disease-free, well-fertilized crops, grown in large fields, under optimum soil water conditions, and achieving full production under the given climatic conditions. ETc can be calculated as:

$$ETc = Kc\,ETo \tag{2}$$

The Kc factor approach is applicable to uniform conditions, e.g. uniform crop fields with adequate fetch distance to minimize micrometeorological effects of field edges. Caution should therefore be exercised in the application of Kc under conditions where spatial variability of soil properties and crop management occur, in natural vegetation etc. The Kc factor can be split into two separate coefficients Kcb + Ke, where Kcb is the basal crop

coefficient referred to crop transpiration and Ke is referred to direct evaporation from the soil (Allen et al., 1998).
The term ETc in equation (2) corresponds to evapotranspiration of vegetation at potential rates (PET) under given climatic conditions. In nature, PET seldom occurs, especially in semi-arid areas. When water is a limiting factor, physiological adaptation of plants occurs, stomata close and ET rates are below potential rates. This mechanism of stomatal control is described schematically in Figure 1.
In the soil-plant-atmosphere continuum (SPAC), water fluxes are driven by atmospheric evaporative demand and limited by soil water supply. Under wet soil conditions, the ratio of actual transpiration (T) and potential transpiration (PT), or relative transpiration (T/PT) is close to 1, showing that the root system is able to supply the canopy with water fast enough to keep up with the atmospheric evaporative demand and thereby preventing wilting. Under these conditions, transpiration is atmospheric demand-limited. As the soil dries beyond field capacity (FC) and beyond a threshold value of water content, T/PT drops below 1. Under these conditions, transpiration is soil water supply-limited as the root system can no longer supply water fast enough to keep up with demand and the soil water can be seen to be less available. Beyond soil water content at permanent wilting point (PWP), transpiration does not occur and T/PT = 0. The same mechanism can be represented for ratios of actual to potential evapotranspiration (ET/PET) as well as actual to maximum yield or productivity (Y/Ym). Plant available water depends on rooting depth, soil texture and structure. A similar mechanism occurs for direct evaporation from the soil surface. Canopy cover is generally used to split evaporation and transpiration, and approximates the available solar energy intercepted by the canopy compared to that reaching the soil surface (Ritchie, 1972).

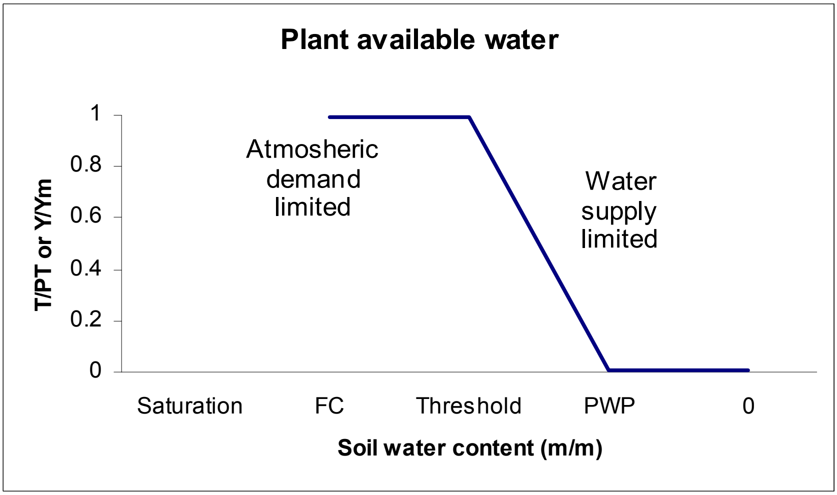

Fig. 1. Schematic representation of the plant available water graph. T – Actual transpiration; PT – Potential transpiration; Y – Actual yield or productivity; Ym – Maximum yield or productivity; FC – Soil water content at field capacity; PWP – Soil water content at permanent wilting point.

The original publication of Denmead and Shaw (1962) included the first scientific evidence of the concept of atmospheric evaporative demand-soil water supply (Figure 2) and this was followed in the last few decades by a large number of research studies on crop productivity-water functions (Doorenbos and Kassam, 1977; Hsiao et al., 2009; Raes et al., 2009; Steduto et al., 2009). This concept is applicable both to wet climates where the limiting factor for ET is generally atmospheric evaporative demand, and to dry climates where the dominant limiting factor is soil water supply.

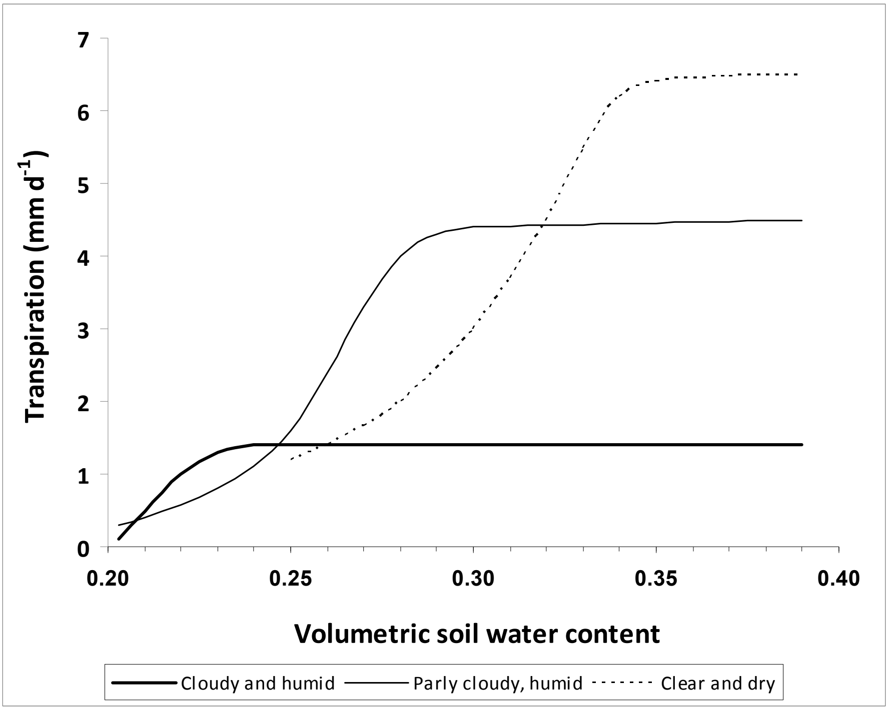

Fig. 2. Graph extracted from the original publication of Denmead and Shaw (1962), supplying scientific evidence of the dependence of transpiration on soil water supply and atmospheric demand.

3. Brief review of methods and techniques to measure and estimate actual evapotranspiration

A large number of methods and techniques for measurement and estimation of ET are available. These can be categorized into the following:

- Lysimeters (Allen et al., 1991): This is the only direct method to measure actual ET.
- Atmospheric measurements
    - Energy balance and micrometeorological methods: These methods are based on the computation of water fluxes based on measurements of atmospheric variables and they are therefore often referred to as direct measurements. Methods and techniques (e.g. Bowen ratio (Bowen, 1926), eddy correlation, scintillometry etc.) were widely discussed by Jarmain et al. (2008).

- Weather data: These methods are based on the calculation of ET from weather data (e.g. Penman-Monteith equation for reference grass ETo).
- Plant measurements
  - Remote sensing from aircraft and satellite: Reflected electromagnetic energy is measured using sensors to generate multi- or hyper-spectral digital images. These data can then be translated into spatial variables such as surface temperature, surface reflectance, and vegetation indices (e.g. the Normalized Difference Vegetation Index NDVI) that describe the vegetation activity and its energy status. These methods were not feasible in the past at large scale and high frequency; however, with the latest technological advances, these techniques show promise (e.g. SEBAL) (Bastiaanssen et al., 1998a and b).
- Soil measurements
  - Soil water balance:

$$ET = P - R - D + \Delta S \tag{3}$$

where P is precipitation; R is runoff or run-on (a component of lateral subsurface inflow/outflow can also be included); D is drainage (or capillary rise), it approximates vertical recharge; ΔS is the change in soil water content, usually measured continuously or manually with a variety of techniques like gravimetric method, soil water sensors, neutron probe, time domain reflectometry etc. (Hillel, 1982). All units are usually expressed in mm per time.

4. Estimation of actual evapotranspiration in hydrological models

Although methodologies for the estimation of ETo and PET are widely adopted, actual (below-potential) ET is difficult to quantify and it usually requires the reduction of PET through a factor that describes the level of stress experienced by plants (two-step approach). The level of stress can be mathematically expressed linearly (slope of line in Figure 1) or through more complex functions. Currently, many models developed for different purposes and operating at different scales apply different functions to reduce PET based on the concept of atmospheric evaporative demand-soil water supply limited ET.

4.1 Field scale models

One-dimensional, field (point) scale hydrological models generally use more detailed functions to predict ET compared to large scale catchment models. The Soil Water Balance (SWB) is an example of a one-dimensional crop model for uniform canopies (Annandale et al., 1999). It is a daily time step model that includes a multi-layer soil water reservoir, where infiltrating water cascades from the top soil layer towards the bottom of the soil profile. Actual transpiration is limited by the evaporative demand (T_{max}) and root water uptake determined by soil wetness (Figure 3). Soil water potential translates into leaf water potential taking into account resistances to water flow in the SPAC (parallel line intersecting the curve in Figure 3) (Annandale et al., 2000).

WATCROS is another example of one-dimensional, cascading water balance and dry matter production simulation model based on climate, soil and plant variables and parameters (Aslyng and Hansen, 1982). It calculates reference ET from grass using a modified formula of Makkink (1957), and it assumes that this grass reference represents any dense, green, growing agricultural crop under Nordic conditions. Such potential evapotranspiration is partitioned into potential evaporation from the soil and crop transpiration using Beer's law.

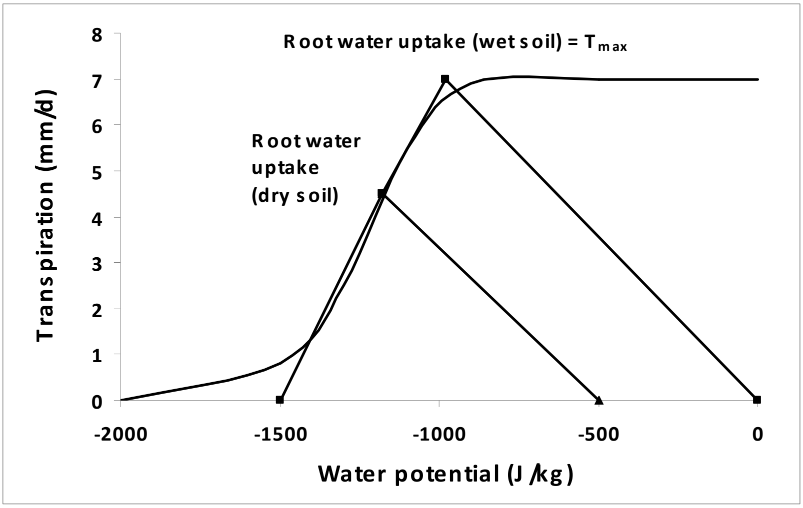

Fig. 3. Schematic representation of the root water uptake function adopted in SWB (adapted from Annandale et al., 2000). T_{max} – Maximum transpiration loss rate (mm d^{-1}).

In order to calculate actual transpiration, water is extracted at a potential rate when the actual soil water content is bigger than half the capacity of the root zone reservoir. Beyond this threshold, actual transpiration is decreased linearly as a function of the remaining water in the reservoir. If no water is left in the root zone reservoir, the transpiration rate equals 0. The size of the root zone reservoir depends on the soil and effective root depth (Hansen, 1984).

GLEAMS (Groundwater Loading Effects of Agricultural Management System) (Knisel, 1993) is also a one-dimensional, piston-flow water balance model used to simulate processes affecting water quality events in agricultural fields. It is the modified version of the well-validated CREAMS model (Knisel, 1980). PET is calculated with the Priestley and Taylor (1972) or with the Penman-Monteith equation (Allen et al., 1998). The model calculates actual soil evaporation and crop transpiration as a function of soil water content and leaf area index.

Cascading soil water balance models based on soil water reservoirs are often employed because of their conceptual simplicity and they are not data intensive. However, soil water movement in porous media can be best described physically with Richards' mass balance continuity equation for unsaturated water flow (Richards, 1931). Richards' equation equilibrates water between specified points (nodes) based on gradients in water energy and hydraulic conductivity:

$$\frac{\partial \theta}{\partial z}=\frac{d}{dz}\left[K(h,z)\frac{dh}{dz}-K(h,z)\right]-S(z,t) \tag{4}$$

where θ is the volumetric soil water content (m^3 m^{-3}); t is time (h); z is soil depth (m, assumed positive downward); h is the soil water pressure head (m); K is the unsaturated hydraulic conductivity (m h^{-1}), a function of h and z; S (z, t) is the sink term (h^{-1}). The conversion of soil water pressure heads into soil water contents and vice versa can be done using different forms of the soil water retention curve (van Genuchten, 1980). The unsaturated hydraulic conductivity-soil water pressure head functions were also described by van Genuchten (1980).

The sink term S(z, t) in equation (4) may include various sinks (or gains with a negative sign) like for example root water uptake. Root water uptake can be calculated with the approach of Nimah and Hanks (1973):

$$-S(z,t)=\frac{\left[H_r+(RRES\cdot z)-h(z,t)-s(z,t)\right]R(z)\cdot K(h)}{\Delta x\cdot\Delta z} \tag{5}$$

where H_r is the effective root water pressure head (m); RRES is a root resistance term; s is the osmotic pressure head (m); Δz is the soil depth increment (m); Δx is the horizontal distance increment; R is the proportion of the total root activity in the depth increment Δz. S cannot exceed potential transpiration.

Richards' equation (4) is non-linear and it can be solved iteratively through a finite-difference solution. It is adopted in several hydrological models to simulate water redistribution in the root zone and for accurate estimates of root water uptake and ET. For example, the RZWQM (Root Zone Water Quality Model) is a physically-based contaminant transport model that includes sub-models to simulate infiltration, runoff, water distribution and chemical movement in the soil (Ahuja et al., 2000). RZWQM simulates PET with a modified Penman-Monteith model and actual ET is constrained by water availability as estimated from Richards' equation.

Soil-Water-Atmosphere-Plant (SWAP) is a 2-D, transient model for water flow and solute transport in the unsaturated and saturated zones (Kroes and van Dam, 2003). It is applied to agrohydrological problems at field scale and it makes use of Richards' equation for soil water redistribution. The relative plant water uptake (T/PT) calculated with this model as a function of soil water potential is shown in Figure 4 (Feddes et al., 1978). The soil water potential values h_1, h_2, and h_4 are inputs. Threshold soil water potentials for reduction in T/PT vary in the range between h_{3h} and h_{3l} and they are applied depending on high (T_{high}) or low (T_{low}) transpiration demand. The h_4 input is wilting point. Reduction in T/PT occurs also in the wet soil range (close to saturation between h_2 and h_1) to simulate the effects of water-logging. The plant water uptake solution in SWAP (Feddes et al., 1978) is also adopted in the HYDRUS unsaturated flow and solute transport model (Simunek et al., 2007) as well as in the SIMGRO (SIMulation of GROundwater and surface water levels) catchment model (van Walsum et al., 2004).

MACRO (Jarvis, 1994) is a deterministic, one-dimensional, transient model for water and solute transport in field soils. It also uses the water uptake function proposed by Feddes et al. (1978). It accounts for conditions that are too wet (close to saturation h_1 in Figure 4) and too dry (close to wilting point h_4 in Figure 4). A dimensionless water stress index ω is used to calculate the ratio of actual to potential root water uptake. This stress index combines two functions describing the distribution of roots and water content in the multi-layered soil profile:

$$\omega=\sum_{i=1}^{i=k} r_i\,\omega_i \tag{6}$$

where k is the number of soil layers in the profile containing roots, and r_i and ω_i are the proportion of the total root length and a water stress reduction factor in layer i. Root length is distributed logarithmically with depth, whilst the stress factor ω_i depends on the soil water content in the particular layer. The root system is usually represented as an inverted cone and its distribution with depth is often non-linear (Yang et al., 2009). The shape of root distribution can therefore be represented with two inputs, namely root depth and an extractable water parameter (Gardner, 1991).

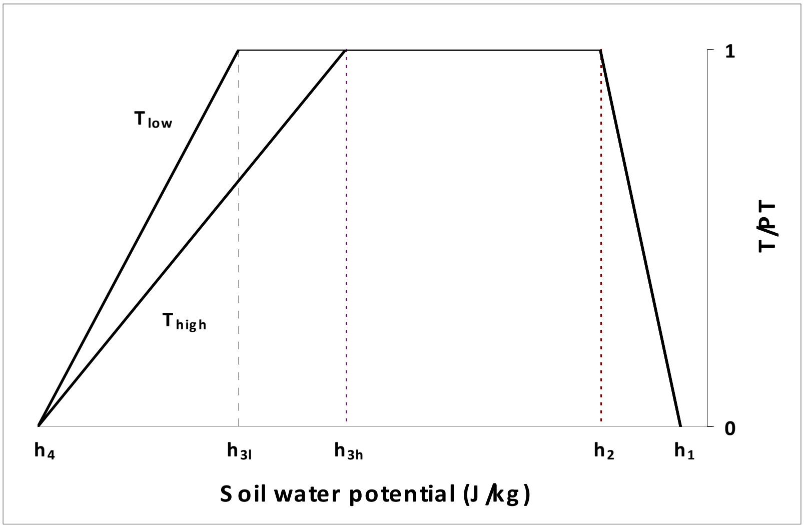

Fig. 4. Schematic representation of the plant available water graph adopted in SWAP (adapted from Feddes et al., 1978). T/PT – Relative plant water uptake; T_{low} – Low transpiration; T_{high} –High transpiration; h_n – Inputs of soil water potential.

The importance of knowing the root depth of vegetation in order to define the size of the soil reservoir and plant available water was underlined by Ritchie (1998) and illustrated in Figure 5. Ritchie (1998) proposed a linear relation between root water uptake and soil water content. Maximum, minimum and usual range of root water uptake are indicated in Figure 5. These depend on root length density Lv and the ability of plants to explore a certain volume of soil.

Another example of a model with a fairly detailed description of root distribution is WAVES (Dawes and Short, 1993; Zhang et al., 1996). WAVES is a water balance model that simulates surface runoff, soil infiltration, ET, soil water redistribution, drainage and water table interactions. Daily transpiration is calculated with the Penman-Monteith equation and reduced using weighting factors determined by the modelled root density and a normalized weighted sum of the matric and osmotic soil water potentials of each layer. The model has been parameterised and used to simulate the water use of various vegetation types in South Africa (Dye et al., 2008).

Feddes et al. (2001) discussed that deep-rooted vegetation and increased water availability may have an effect even on global climate. Deep rooting systems result in large volumes of soil being explored by the roots, large amounts of soil profile available water and large transpiration rates. This is even more prominent in the presence of shallow groundwater. Jovanovic et al. (2004) proved that the contribution of shallow water tables to root water uptake through capillary rise can be a substantial component of the water balance.

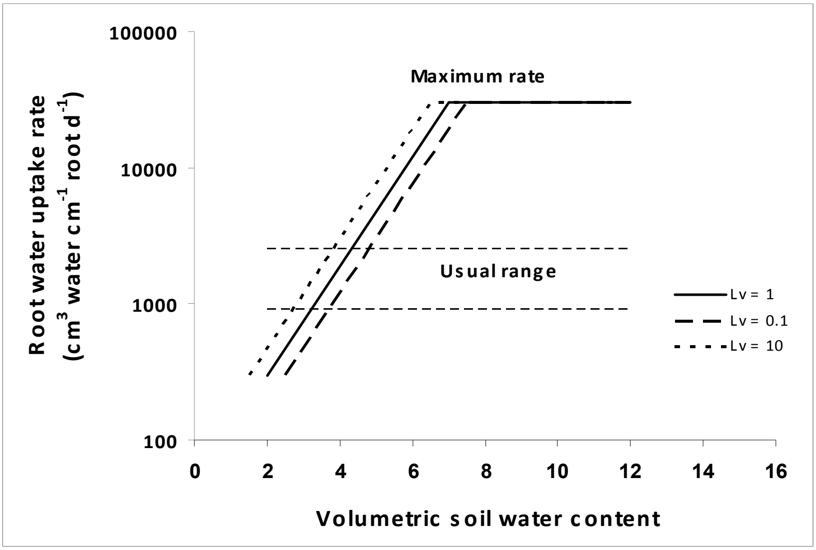

Fig. 5. Relationship between root water uptake rate, volumetric soil water content and root length density (Lv in cm cm^{-3}) (adapted from Ritchie, 1998).

The DRAINMOD computer model was primarily developed to simulate the effects of drainage and associated water management practices on water table depths, the soil water regime and crop yields (Skaggs, 1978). ET is calculated according to the relationship of Norero (1969):

$$ET = \frac{PET}{1 + \left(\frac{h}{h*} \right)^k} \tag{7}$$

where k is a constant that can be defined using methods given in Taylor and Ashcroft (1972) and Norero (1969), h is the soil water potential in the root zone which could be obtained from the soil water characteristics using the average root zone water content, and h* is the value of h when ET = 0.5 PET. Equation (7) is graphically illustrated in Figure 6.

Given the purpose of the DRAINMOD model, direct evaporation from the soil can be estimated using the simplified Gardner (1958) equation relating maximum evaporation rate in terms of water table depth and unsaturated soil hydraulic conductivity:

$$\frac{d}{dz}\left[K(h,z)\frac{dh}{dz} - K(h,z) \right] = 0 \tag{8}$$

The symbols and units are the same as those defined for equation (4). Maximum soil evaporation rate for a given water table depth can be approximated by solving equation (8), using a large negative h value (for example h = -1000 cm) at the surface (z = 0) and h = 0 at the water table depth. An example of solution of equation (8) is shown in Figure 7 for a loamy sand.

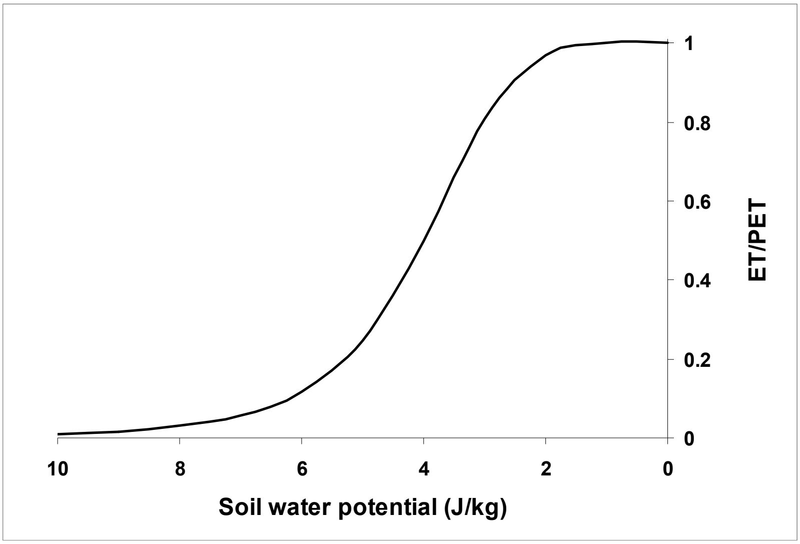

Fig. 6. Schematic of relative evapotranspiration (ET/PET), as affected by soil water potential in the root zone (adapted from Skaggs, 1978)

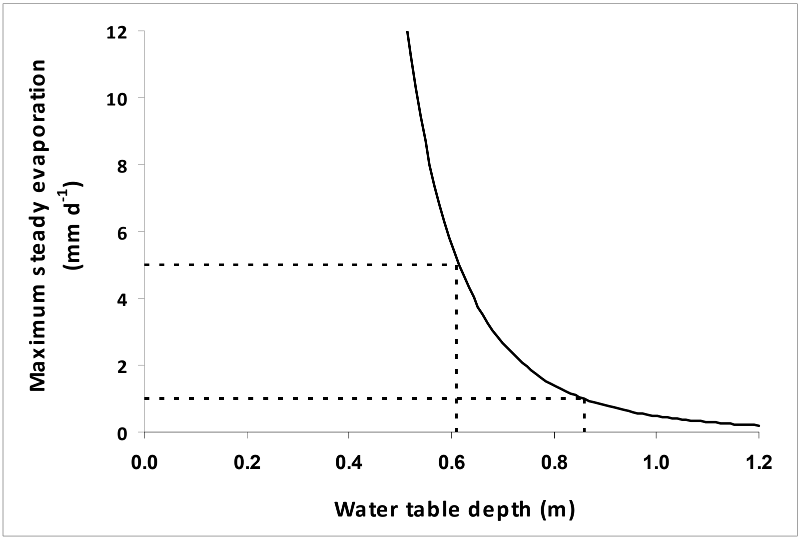

Fig. 7. Relationship between maximum upward movement of water versus water table depth for a loamy sand (adapted from Skaggs, 1978).

4.2 Catchment scale models

Many catchment scale models account for soil moisture in the estimate of ET (Viviroli et al., 2009) using more or less sophisticated approaches. For example, Zhang et al. (2001) developed a semi-empirical water balance model for forested and non-forested catchments in the Murray-Darling basin of Australia. This was based on the assumption that actual ET is equal to precipitation under very dry conditions, and that it equals PET under very wet conditions. On the other hand, Gurtz et al. (1999) applied the PREVAH (PRecipitation-Runoff-EVApotranspiration HRU Model) hydrological model in an alpine basin. They calculated ET using the Penman-Monteith equation by changing the canopy stomatal resistance (equation (1)) below a given threshold of soil moisture.

Barr et al. (1997) reviewed a number of studies where the dependence of ET on soil moisture was evidenced. In their study, they evaluated three methods for estimating ET in the SLURP mesoscale hydrological model (Kite, 1995), namely: i) the complementary relationship areal ET model (Morton, 1983), ii) the Granger (1991) modification of Penman's method and iii) the Spittlehouse (1989) energy-limited versus soil moisture-limited method. The method of Morton (1983) makes use of ET estimated with the Penman (1948) equation and reduced by an amount proportional to vapour pressure deficit, without taking into account the effects of soil moisture on ET. The method of Granger (1991) is a modification to the Penman (1948) equation that includes a relative evaporation variable in the vapour pressure deficit term. The Spittlehouse (1989) method takes into account soil moisture and it calculates actual ET withdrawal from the soil store as the lesser of the soil store and energy-limited rates. The energy-limited rate is calculated with the Priestley-Taylor equation (Priestley and Taylor, 1972) and the soil store-limited rate is calculated as a function of the fraction of extractable soil moisture (Spittlehouse, 1989). The formulation of all three methods was based on forests and grasslands in large catchments. Amongst the three methods tested over a 5-year period in the Kootenay Basin of eastern British Columbia, the Spittlehouse (1989) method including the soil moisture feedback to ET estimates gave the best agreement between simulated and recorded streamflow.

Zhou et al. (2006) used the Shuttleworth and Wallace (1985) model and NDVI to estimate ET from sparse canopies to feed in the BTOPMC distributed hydrological model (Takeuchi et al., 1999). The methodology adopted the Penman-Monteith ETo with an increase of stomatal resistance based on the generic equation:

$$r_s = \frac{r_{s\,\min}}{LAI\, F_i\left(X_i\right)} \tag{9}$$

where $r_{s\,min}$ represents the minimal stomatal resistance of individual leaves under optimal conditions (s m^{-1}), LAI is the effective leaf area index and $F_i(X_i)$ is the stress function for a factor X_i (nutrients, pests, water etc.). The water stress function was expressed as a function of volumetric soil water content θ, in the range between field capacity θ_{fc} and residual soil water content θ_r:

$$f\left(\theta\right) = \frac{\theta - \theta_r}{\theta_{fc} - \theta_r} \tag{10}$$

The Agricultural Catchments Research Unit (ACRU) (Schulze, 1994) is a catchment scale agrohydrological modeling system. It calculates relative evapotranspiration (ET/PET) as a function of plant available water (Figure 8). The reduction of ET/PET on the left side of the graph in Figure 8 describes the effect of water-logging. The threshold f_s is user-specified, or it is calculated as a function of a critical leaf water potential ψ^{cr} and ET/PET:

$$f_s = F\left(user\,specified\,PAW\right) \tag{11}$$

$$f_s = 0.94 + \frac{0.0026\,\psi^{cr}}{ET\,/\,PET} \tag{12}$$

Precipitation-Runoff Modular System (PRMS) is a hydrological modular modeling system for large scale basins (Leavesley et al., 1983, 1996). It calculates actual ET for four types of vegetation/land use and three types of soil texture (Figure 9).

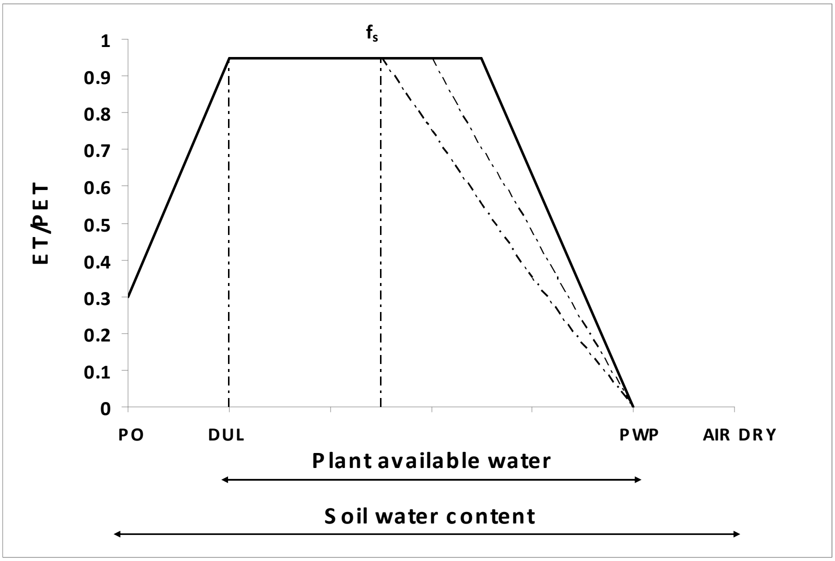

Fig. 8. Schematic representation of the plant available water graph adopted in ACRU (adapted from Schulze, 1994). ET – Actual evapotranspiration; PET – Potential evapotranspiration; PO – Soil water content at saturation; DUL – Drainage upper limit; PWP – Permanent wilting point; f_s – Threshold of reduction of relative evapotranspiration (ET/PET).

MIKE SHE is a physically-based, distributed, integrated hydrological and water quality modeling system (Abbott et al., 1986). ET is calculated based on PET, leaf area index and root depth, soil water content and physical characteristics as well as a set of empirical parameters (Kristensen and Jansen, 1975). Specifically, the ratio of ET to PET is calculated with two functions, the one describing the leaf area index and the other describing the soil water status.

More empirical approaches aimed at describing the hydrological cycle also take into consideration ET. A semi-empirical model called EARTH (Extended model for Aquifer Recharge and moisture Transport through unsaturated Hardrock; Department of Water

Affairs and Forestry, 2006) was developed in South Africa to estimate large scale groundwater recharge by accounting for the variables of the hydrological cycle. EARTH uses modules for vegetation, soil, linear reservoir and saturated flow. The soil module calculates ET as a linear function of soil moisture (Figure 10).

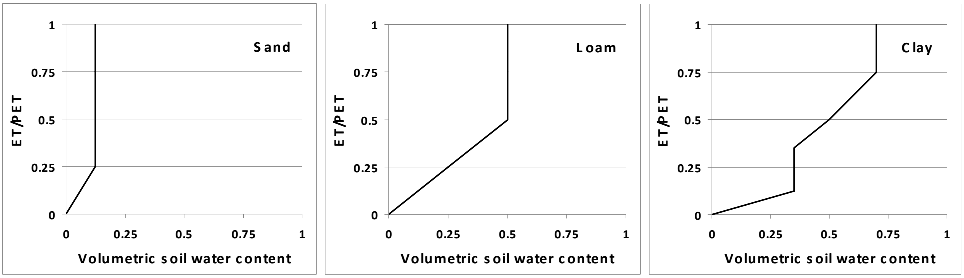

Fig. 9. Schematic representation of the plant available water graph adopted in PRMS for three types of soil texture (adapted from Leavesley et al., 1983). ET – Actual evapotranspiration; PET – Potential evapotranspiration.

The chloride mass balance (CMB) is another method commonly used to estimate groundwater recharge in semi-arid areas (Xu and Beekman, 2003). The estimates of groundwater recharge with CMB refer to long term annual averages, usually over hundreds of years. Implicitly, this technique accounts for the concentrating effects of water by ET in semi-arid regions. Groundwater recharge can be calculated with the following formula:

$$P\,Cl_p = R_T\,Cl_{gw} \tag{13}$$

where P is precipitation (mm a^{-1}); Cl_p is the chloride concentration in precipitation (mg L^{-1}); R_T is total groundwater recharge (mm a^{-1}), approximated with the term D in equation (3);

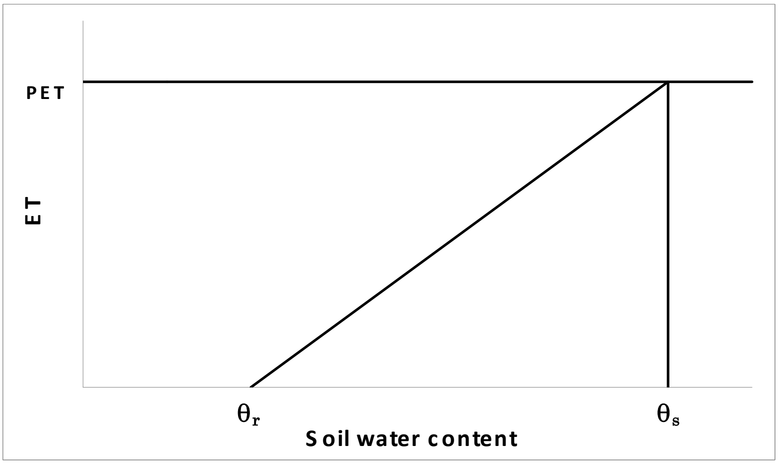

Fig. 10. Schematic representation of the plant available water graph adopted in EARTH (adapted from DWAF, 2006). PET – Potential evapotranspiration; ET – Actual evapotranspiration; θ_r –Soil moisture retained by the soil matrix; θ_s – Maximum soil moisture.

Cl_{gw} is the chloride concentration in groundwater (mg L^{-1}). The source of Cl has to be precipitation solely as other sources may intefere with the interpretation of Cl measurements. Other conservative tracers can also be used. As groundwater recharge can be approximated with term D and ΔS is negligible in the long term, equation (3) can be applied to calculate ET if mean annual runoff data are available.

4.3 Remote sensing applications in the estimate of actual evapotranspiration

The methods discussed above generate point estimates of ET. These values are usually applicable to uniform crop fields, hillslope transects or hydrologically homogeneous areas, and they often need to be upscaled (Oudin et al., 2005). Upscaling can be done through repetitive measurements in all representative areas of interest or through regionalization (Krause, 2002). Due to spatial variations in climate, vegetation, land use and physiographic characteristics, point methods for estimating ET are often too intensive to be applied at large catchment scales. A promising application that may overcome these shortcomings involves areal estimates of ET with remote sensing techniques.

The theory described in the canopy temperature-ET models of Hatfield et al. (1984) was the foundation for surface energy balance approaches based on remote sensing. In these approaches, each pixel of aircraft or satellite images is processed to determine the components of the energy balance equation:

$$\lambda E = R_n - H - G \tag{14}$$

where R_n is net radiation, λE is the latent heat of vaporization, H is the sensible heat flux, G is the soil heat flux and all terms are usually expressed in W m^{-2}. Algorithms such as the Surface Energy Balance Algorithm over Land (SEBAL) use remote sensing imagery, empirical relationships and physical modules to calculate the terms of the energy balance equation and estimate ET (converted from λE in equation (14)) (Bastiaanssen et al., 1998a, 1998b; Tasumi et al., 2005). In particular, SEBAL requires visible, near-infrared and thermal infrared input data obtained from satellite images. Instantaneous net radiation can be calculated from incoming solar radiation measured at ground stations and outgoing thermal radiation estimated from surface albedo, surface emissivity and temperature. Soil heat flux can be computed from surface temperature, albedo and NDVI. The sensible heat flux is calculated with an algorithm of standard heat and momentum transport equations including pixel-based Monin–Obukhov stability corrections. Both wet and dry surface pixels are required because these represent extreme limits in the studied domain at the specific time when the satellite images are taken. The sensible heat flux is constrained by a dry limit (surface with latent heat flux λE = 0; sensible heat flux H = R_n - G) and wet limit (surface with sensible heat flux H = 0; vertical difference in air temperature dT_a = 0). A value of dT_a is assigned to all other pixels assuming it varies linearly between the dry and wet ranges. H is then calculated as a function of dT_a and λE computed as the residual of the energy balance. Instantaneous λE values are extrapolated over time assuming that the instantaneous evaporative fraction in equation (14) is stable for the given time period.

Other remote sensing based methods to estimate ET are also available. The Surface Energy Balance System (SEBS) is an energy balance algorithm for the estimation of ET (Su, 2002) that works on similar principles as SEBAL. The MODIS evapotranspiration (ET – MOD 16) algorithm is based on the Penman-Monteith equation (Allen et al., 1998). Land cover, fraction of absorbed photosynthetically active radiation, leaf area index and global surface meteorology information derived from MODIS are used to estimate daily ET and PET, which is then composited over an 8-day interval. ET is expressed in mm d^{-1} and calculated globally every day at 1 km resolution. METRIC (Mapping EvapoTranspiration at high Resolution with Internalized Calibration) is a computer model that uses LandSat data to compute and map ET. These ET maps (i.e. images) provide the means to quantify ET on a field by field basis in terms of both rates and spatial distribution (Allen et al., 2007). Sinclair and Pegram (2010) implemented a real time platform for supplying satellite-based information on ETo and soil moisture in South Africa. Wang et al. (2003) found a significant correlation between deseasonalized time series of NDVI and soil moisture, from where root zone depth can be indirectly estimated. This procedure, however, requires calibration for specific vegetation and climatic conditions.

Although some studies have been carried out in order to test and compare remote sensing methodologies to conventional methods for estimation of ET (Gibson et al., 2011; Kite & Droogers, 2000), more research is required in order to assess the feasibility of application of remote sensing techniques to improve water use efficiency, irrigation management on farms and catchment management, particularly in arid and semi-arid areas. Given the temporal dynamics of ET and its dependance on soil water supply conditions, the interpolation of instantaneous satellite information to estimate ET over a given time period may require verification (Olioso et al., 2005). Processed information from satellite images needs to be supplied at a required frequency for applications in water management on farms and in large catchments. In addition, cloud-free satellite images are required and these are not always available.

5. Conclusion

This chapter discussed the theoretical principles of some hydrological models as examples. It was not meant to provide a review of all models available. The models described here were extensively evaluated in specific studies. Wagener (2003) proposed models should be evaluated for performance (e.g. by minimizing the objective function which can be the difference between simulated and observed data), uncertainty (e.g. by analyzing reasonable ranges of model inputs, parameters and structure) and realism (e.g. by analyzing how consistent the model output is with our understanding of reality). No unique approach for model evaluation exists and, therefore, there is no easy answer to the question on which model is the most accurate. Models should be used for the purpose that they were developed and evaluated with different techniques and for different conditions.

The quantification of actual ET is of utmost importance for various applications in hydrology and water management, such as resource allocation, water footprinting, quantification of water use efficiency etc. This review has highlighted that a large number of both field (point) scale, one-dimensional models and catchment scale spatial GIS-based models adopt conceptually similar approaches to the estimation of actual ET. These approaches are based on the concept of atmospheric evaporative demand-soil water supply limited ET. Such a concept is applicable both to wet climates (limiting factor is atmospheric evaporative demand) and to dry climates (limiting factor is soil water supply). Some models make use of a one-step approach to increase canopy stomatal resistance directly in the Penman-Monteith equation, which represents a mechanistic and physically sound solution to the estimation of actual ET (e.g. BTOPMC). This methodology is, however, hampered by the difficulty in estimating the canopy resistance term. Other models adopt a more conventional two-step approach to calculate PET and reduce it using a water stress index generally based on soil water content (e.g. WATCROS). Some models make use of the data intensive and physically sound principles embedded in Richards' equation to redistribute water in the root zone (e.g. SWAP). Other models make use of a simplistic soil reservoir-based cascading water balance as finite differences are difficult to apply to complex and large scale systems (e.g. ACRU). In addition, abrupt and large changes in soil water content in space and time may lead to numerical instabilities in the finite difference solution of Richards' equation, or in longer simulation times compared to cascading soil water balance models because equilibrium conditions, usually solved through an iterative process, may not be reached easily.

When applying specific models, it is essential to be aware of the specific assumptions around which they were built, their advantages and limitations. Field scale models are generally more data intensive than catchment scale models. For example, dedicated crop and soil water balance models usually include moving thresholds in the atmospheric demand-soil water supply function (e.g. SWB). Models that estimate leaf area provide the opportunity to partition the energy available for soil evaporation and plant transpiration, and those that calculate root growth and depth facilitate the estimation of plant available water in the soil. If properly calibrated, such models are more accurate in predicting field (point) scale ET, but they are also more data intensive compared to large scale models. Large scale catchment models require ET-related inputs in the spatial domain and make use of less detailed ET calculation sub-routines as trade-off (e.g. PRMS).

Given the principles governing soil water redistribution, the soil water dynamics and ET, it is recommended that a daily time step be used in the calculation of water balance variables. Root depth is a very important variable that determines the volume of soil explored by plant roots. This is not often easily measured resulting in uncertainties in the estimation of ET and the water balance. Promising technologies for large scale spatial estimation of ET, soil moisture, and indirectly root depth include remote sensing. These techniques, however, need to be tested and validated for applicability to a wide range of water management conditions in arid and semi-arid areas. The purpose and applicability of remote sensing methods depend on the spatial resolution of the images and their temporal resolution (frequency).

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