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Hydrological approaches to measure or estimate crop water use – A theoretical background

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ABSTRACT

Hydrological methods are used to retrieve actual evapotranspiration (*ETa*) as an alternative to micrometeorological techniques. This approach consists in performing the mass balance at a well-defined control volume. Despite the apparent theoretical simplicity of these methods, it is very common to witness their incorrect use, due to the lack of consistency in the definition of their terms by different authors, not only in measurements but also and mainly in common estimations for irrigation scheduling. Therefore, this article aims to review the main concepts and basic definitions of the hydrological methods, especially in the context of irrigation applications in climates with dry summer. The main topics addressed consist of: i) Mass balance and control volume; ii) Soil water thresholds for engineering applications; iii) Evapotranspiration or its components as output variables of water balance equation; iv) Applications of the water balance equation using actual evapotranspiration as input variable. This article is meant as a didactic text in the field of irrigation and drainage to support the learning of concepts related to the water balance. It also includes a set of application exercises to improve the comprehension of this subject.

Keywords: water balance, soil water thresholds, evapotranspiration, crop irrigation requirements, irrigation scheduling.

RESUMO

Abordagens hidrológicas para medir ou estimar a utilização da água pelas culturas - fundamentos teóricos

Os métodos hidrológicos são utilizados como alternativa às técnicas micrometeorológicas para medir a evapotranspiração real (*ETa*). Esta aproximação consiste na realização do balanço de massa a um volume de controlo bem definido. Apesar da aparente simplicidade teórica destes métodos, é muito comum assistir-se à sua incorreta utilização devido à falta de consistência na definição dos seus termos por diferentes autores, não só nas medições, mas também e sobretudo nas estimativas comuns para a

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programação da rega. Assim sendo, neste artigo pretende-se rever os principais conceitos e definições de base dos métodos hidrológicos, no contexto de aplicações em gestão da rega, designadamente em climas de tipo mediterrânico ou em climas semi-áridos. Os principais tópicos consistem em: i) Balanço de massa e volume de controlo; ii) Limiares críticos de água no solo para aplicações de engenharia; iii) Evapotranspiração ou as suas componentes como variáveis de saída da equação do balanço hídrico; iv) Aplicações da equação do balanço hídrico usando a evapotranspiração real como variável de entrada. Disponibiliza-se, assim, um texto de carácter pedagógico, na área da rega e drenagem, para apoio à aprendizagem dos conceitos relacionados com o balanço hídrico, contendo um conjunto de exercícios de aplicação que permitem melhorar a compressão desta temática.

Palavras-chave: balanço hídrico, limiares de água no solo, evapotranspiração, necessidades de rega, programação da rega.

1. Mass balance and control volume

As an alternative to micrometeorological techniques, the hydrological methods are often used to retrieve actual evapotranspiration (*ETa*). They are based on the use of a mass balance equation - in this case for water - applied to a certain control volume (CV), at different temporal scales (e.g., minutes to a year) and spatial scales (e.g., from a simple plot to a watershed). A clear definition of the CV is necessary, as only the fluxes that cross the CV boundaries mater for the balance. For irrigation purposes, the CV is usually defined between the soil surface and a parallel plan at a certain depth, often the maximum or average root depth, corresponding to a certain volume of soil exploited by roots per unit of horizontal area. It can be a finite small volume (e.g., container) or be per unit of soil area (extended areas). The variables in the water balance equation depend on the precise definition of the CV and conditions, and on the spatial and temporal scales considered (vide mass balance equation, in 1.1).

In many applications, fluxes near the surface of the earth (energy, mass, momentum) are expressed in flux density units, i.e., fluxes per unit of area. Therefore, in the mass balance, it is preferred to do not consider water flows ($m^3\ s^{-1}$) but water flux densities ($m^3\ m^{-2}\ s^{-1}$ or also mm/day, mm/year), being the volume of water, in the case of the soil, defined per unit of surface area measured in the horizontal plan. One of the variables in the mass balance is the change in storage, dependent on the soil water content (see next paragraphs and 1.2).

As for any balance, water fluxes inside a defined CV have to be ignored, and only fluxes crossing the boundaries of the CV are considered. Descriptive schemes on hydrological cycle, often presented in the literature, can be misleading as they often present all fluxes occurring in every part of the system, and not just those occurring at the boundaries of a certain CV. Besides, if these boundaries are with non-porous media, such as the atmosphere, the fluxes are not quantified in relation to water potential but rather in terms of gradients of water vapour concentration. Notwithstanding, to apply a balance, the fluxes are quantified previously and considered as inputs or outputs, as they are entering or exiting the considered CV. There is an increase in water stored inside the CV, when the total input is higher than total output and the opposite (decrease in storage), if outputs overpass inputs. Consequently, for consistence of units, the variation of water stored (ΔS), in the volume of soil defined as CV, is also expressed per unit of area (e.g., dm^3/m^2 or mm, per day).

1.1. Water balance equation

The water balance equation, adapted to a CV defined between a plan immediately above (or coincident with, adjustments required) the soil surface and a parallel plan at the observed maximum root depth, can include the following terms all expressed as volumetric flux densities ($m^3\ m^{-2} \times 10^{-3}$ or mm / unit of time):

$$P + I + U - ETa - D - R = \Delta S \qquad [1]$$

in which P is precipitation attaining the soil, I is irrigation depth, U is capillary rise (it can be important if the water table is shallow and the soil has a fine texture), ETa is actual evapotranspiration (positive, for up water vapour flux) minus dewfall (down flux, sometimes accounting more than P on an annual basis, Jacobs *et al.*, 2006), D is deep drainage (beyond the root zone, i.e., crossing the lower boundary of the CV), R is runoff (and can be subdivided considering sub-superficial lateral fluxes outside the CV) and ΔS is the change in water stored in the CV per unit of soil surface (Fig. 1).

At the scale of an entire watershed, R becomes the flow of water in the streamline, divided by the watershed area and variables can be different depending on how the CV is defined. If the CV does not include the vegetation canopy, but there is an expressive canopy, water balance needs to incorporate effective P, P_e instead of total P, considering stemflow and throughfall, interception becoming part of ET, all depending again on the lower CV boundary.

The soil, in this context, is seen as a mixture of air, water, mineral particles and other solid compounds, organized as a porous media, with micro, meso and macropores, where major water movements occur according to Darcy's law (macro flux). The corresponding equation states that, in permanent regime, flow density (flow per unit area, $m^3\ m^{-2}\ s^{-1}$) is proportional to the gradient of total water potential. The constant of proportionality is the so-called hydraulic conductivity (in either saturated or unsaturated conditions). This approach can be a starting point to estimate fluxes inside the soil, *U* or *D*, in circumstances where relevant data are available.

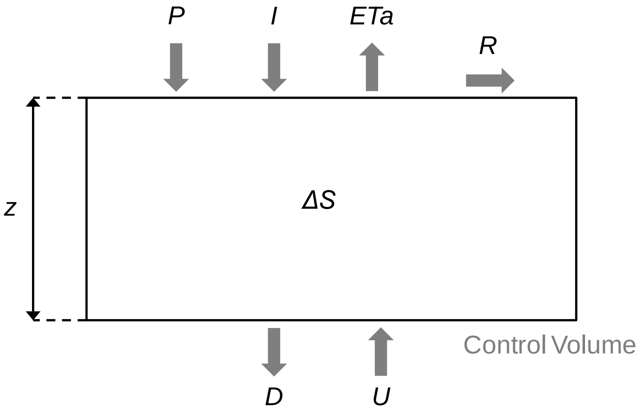


Figure 1. Schematic representation of the control volume, per unit area, and of the main soil water balance terms (see text above). In this simplification, the upper line represents soil surface (bare soil or plants of negligible size), the bottom line, the lower end of the volume considered, usually related to the one occupied by roots, the left and right lines delimitate one square meter or the specific volume.

However, for a certain soil layer, and considering the difficulty to quantify *U*, and the reduced or null value of *R* in most applications, a simplified equation is often considered under irrigation conditions. If the upper limit of the CV is exactly the soil surface, R is not a flux to consider as it happens on the soil surface, outside the CV. Consequently, *P* and *I* are replaced by P_e and I_e, effective irrigation, i.e., the water that entered the soil (CV), obtaining the following:

$$P_e + I_e - ETa - D = \Delta S \qquad [2]$$

Usually, *P* data come from meteorological services or local weather stations. Values for I_e and P_e are quite variable in space, so that reliable values are often not accessible in a relatively large scale. Also, for *D* it is difficult to get sound estimates. During dry summer, main inputs of water come from irrigation; ideally, a very efficient irrigation would reduce *D* to negligible values.

It can be difficult to get precise enough values for ΔS from soil water content (SWC), due to the very high variability of horizontal and vertical water distribution. The replication needed to obtain reliable values of ΔS implies that this approach (to obtain ΔS from SWC) is not very practical, either using direct (gravimetric approach) or indirect methods (soil properties related to SWC, such as the ones applied when using neutron scattering, gamma ray absorption or attenuation, capacitance, among others (Rana and Katerji, 2000; Oliveira, 2011). For application on a very short temporal time scale, ΔS is better evaluated by using special fixed infrastructures, such as weighing lysimeters (section 3.1.).

Finally, any consideration about ETa depends on the aim of the study. Depending on whether *ETa* is an input or an output variable (in equation 1), different applications are considered (details and examples given in sections 3 and 4).

1.2. Concentration in a mixture and water stored in a certain volume of soil

How to obtain ΔS from SWC? The sensors that measure the soil water content (SWC) do not quantify directly the storage, *S*. Usually, *S* is obtained by multiplying the depth of the CV (volume per unit of area) by the soil water content (SWC) expressed in volume, i.e., the water volumetric fraction (θ_v, dimensionless), which is the volume of water (V_w) per unit of volume of the mixture (V_s), i.e., the soil (e.g., $\theta_v = 0.3$ m^3 m^{-3}, meaning 0.3 m^3 of water per m^3 of soil). There is a need for the subscript *v* when using the percentage form, e.g. $\theta_v = 30\%$, in order to differentiate from mass fraction).

Generally, in a homogeneous soil layer $S = \theta_v \times z$ (*z*, being soil depth, details in equation 4). The SWC in the soil profile is rarely evenly distributed either horizontally or vertically (Fig. 2) and the distribution function that would allow the use of an integral function is rarely available. Therefore, the water stored in the soil, till the root depth, in a certain location is, in practice, obtained from $S = \Sigma\ [\theta_{v,i} \times z_i]$, where i is the layer where θ_v measurements are taken along the vertical direction. This description is detailed below (equation 4), in what *z* is concerned.

The concentration of water in the soil can also be expressed in mass fraction (θ_m) which is the mass of water per unit of mass of the mixture, e.g., $\theta_m = 20\%$ or $\theta_m = 0.2$ kg kg^{-1}. As before, the subscript is necessary when using the first form, to differentiate from volumetric fraction.

When using the gravimetric method to obtain the SWC, results come expressed in mass fraction, due to the process of weighing the humid and dry soil samples. Transforming θ_m into θ_v (equation 3) requires data on the so-called soil bulk density of a certain layer (d_s, dimensionless). The soil bulk density corresponds to the relationship between volumetric mass of dry soil (m_s/V_s, kg of soil / m^3 of soil) and volumetric mass of pure water, at reference temperature (4 ºC) and standard atmospheric pressure (m_w/V_w, kg of water / m^3 of water, i.e. 1000 kg/m^3):

$$\theta_v = V_w / V_s = (m_w / m_s)\ [(m_s / V_s) / (m_w / V_w)] = (m_w / m_s)\ d_s = \theta_m\ d_s \quad [3]$$

where V and m are volume and mass, subscripts w and s refer to water and soil, respectively.

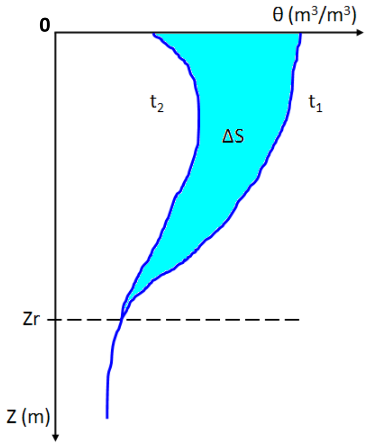


Figure 2. Soil water content (θ) distribution in the soil profile from the surface to maximum root depth, z_r. The change in storage ΔS, for the time interval Δt = $t_2 - t_1$, corresponds geometrically to the area between the two lines. Here it is assumed there were no changes in SWC below root depth.

This value d_s depends on soil texture (particle distribution) and structure, consequently being dependant on soil compaction and closely related with its varying porosity (air volume per unit of soil volume, V_{air} / V_s); it depends on water content due to swelling and shrinking processes (except for organic soils, Hillel, 1980). Soil bulk density is usually lower in upper soil layers. According to Costa (2004), d_s values above 1.50 or 1.75 in fine or coarse soils, respectively, generally imply difficult conditions for root growth. Soil bulk density (d_s) limits for plant and root growth as a function of soil texture classes and compaction conditions are exemplified in Table 1.

As said above, being θ_v the volumetric fraction of water in a certain volume of soil (V_w/V_s), $S = \theta_v\ (V_s / A)$, the total volume of water in the soil (V_w) per unit of soil area (A) is obtained by multiplying θ_v by z, which is the volume of soil (V_s) per unit of soil area ($V_s/A = z$), i.e., the vertical extension (z) of the CV used for the water balance (equation 4). S comes in the same unit used to express z. As P, it represents volume of water per unit of area.

$$S = V_w / A = (V_w / V_s)\ (V_s / A) = \theta_v\ z_r \qquad [4]$$

Table 1. Soil bulk density (d_s) limits for plant and root growth for different soil texture classes and compaction conditions (USDA, 2008)

Soil Texture	**Ideal bulk densities for plant growth**	**Bulk densities that restrict root growth**
Sandy	< 1.60	> 1.80
Silty	< 1.40	> 1.65
Clayey	< 1.10	> 1.47

For irrigation applications, the volume of interest corresponds to the one occupied by roots (z_r), where water is desirably applied. If this volume, in terms of water application and distribution, is not homogeneous in an horizontal plane an average value for z_r should be used or, in case of crops in rows far apart, a partial area where roots develop (see notes on root depth).

Example 1: If θ = 0.20 $m^3\ m^{-3}$ and z_r = 0.4 m, then S = 0.20 (m^3 of water / m^3 of soil) × 0.4 (m^3 of soil / m^2 of surface area) = 0.080 m^3 of water / m^2 of surface area = 0.080 m = 80 mm.

Example 2: If θ = 0.40 $m^3\ m^{-3}$ (θ_v = 40% as commonly presented) and z_r = 1000 mm,
then S = 0.40 × 1000 mm = 400 mm.

1.3. Root depth

The selection of the correct value for z_r is a critical issue for different reasons. Equation 4 refers to a situation where z_r does not change in the horizontal plan, as it occurs in general with continuous low crops completely covering the soil. This can be the case even for woody crops, in certain conditions, For instance, in a rainfed olive stand in South Portugal, water scarcity and specificities of plant roots adaptation determined full and uniform horizontal colonization of the soil by the roots (Conceição *et al.*, 2018).

The determination of appropriate z_r is less simple in some cases, such as the two herewith described (anisotropic and sparse canopies and corresponding root systems) where, if using the equations as above, z_r represents average root depth considering the all area and not the root depth below the plants (examples below).

For instance (case 1, anisotropic), in most woody crops and some herbaceous crops, plants are cultivated in lines separated by a distance often larger than the one between plants in the row. In that case, the stand can be seen as anisotropic, here used in large sense, i.e., applied to two horizontal perpendicular directions. With irrigation applied in the line, a high anisotropy of roots is expected (Fig. 3a). Conversely (case 2, sparse canopies), in irregular, natural or not, sparse stands, plants are too far apart, even if the canopy is isotropic. The roots eventually do not colonize the entire soil volume, but there is no regular pattern geometrically easily defined.

In such cases (1 and 2), z_r depends on location (Fig. 3), assuming higher values below plants or below the places of water distribution (e.g. drippers, which can be placed coincidently with plant lines or not). If, for instance, roots occupy 10% of the total surface area with a certain root depth z_r, an average z_r ($z_{r,a}$) could be simplified as $0.1 \times z_r$, corresponding to a weighted average of each partial area multiplied by the corresponding root depth. However, the roots colonization is not a matter of simply being present or absent, but of progressively decreasing density, either in horizontal or vertical directions (Fig. 3b).

Fine roots almost not visible can have an important role in water uptake (Conceição *et al.*, 2018). Observations of roots based only on clearly visible ones or in criteria related with their volume (or mass) are not entirely appropriate. For irrigation engineering applications, the right compromise is a delicate step requiring experience and fine adjustments (McKeague *et al.*, 1984; Fernández *et al.*, 2013).

2. Soil water thresholds for engineering applications

2.1 Field capacity and permanent wilting point

Not being essential for the water balance application, for irrigation purposes it is very important to be aware of the meaning of specific values of SWC relevant to follow water uptake (≈ water use, on a daily scale). As an example, $\theta_v = 0.2$ has a very different meaning, even opposite, depending on soil texture: it is too low for a clay soil but too high for a sandy soil. The two conditions corresponding to special thresholds used for engineering purposes are the so-called field capacity (FC) and the permanent wilting point (WP).

It should be stressed that FC does not represent a value but a specific condition, occurring, when a soil having received water in excess of its storage capacity drained long enough (soil evaporation or plant water uptake prevented). With time (depending mostly on soil texture), water leaves the macro pores and the drainage rate reduces progressively until it tends to zero so that, for practical purposes, a relatively stable value of θ_v is achieved. Then, at this stage, the soil is considered to be at FC, the value of θ_v observed in such condition being θ_v at FC ($\theta_{v,FC}$).

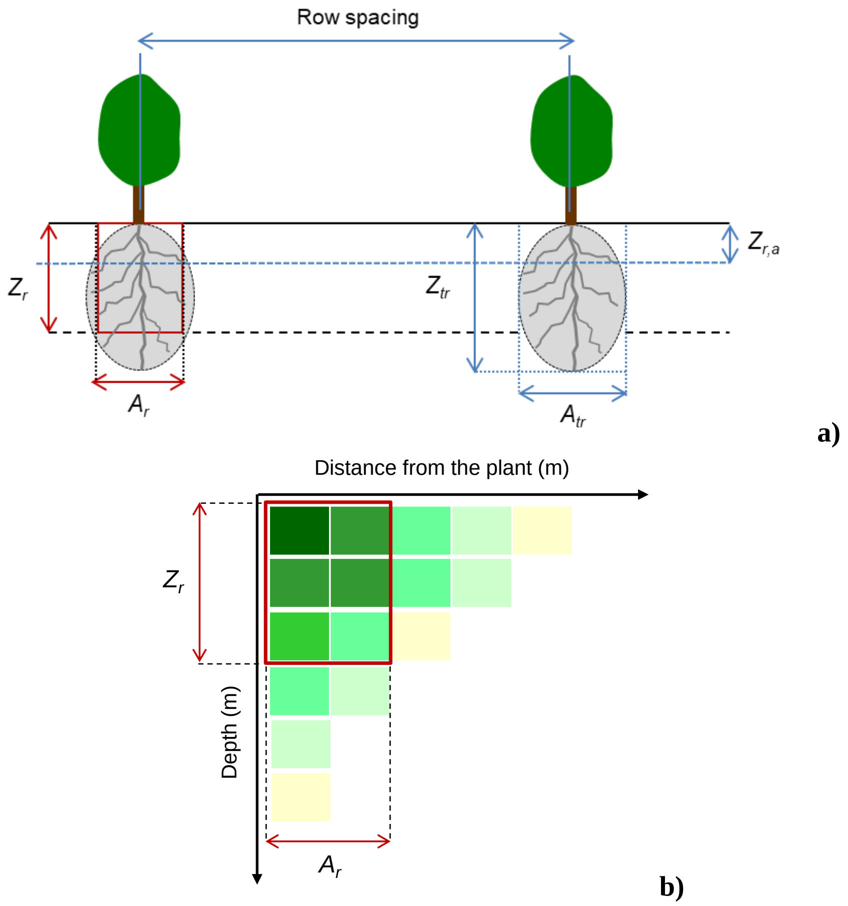


Figure 3. a) Schematic representation of the estimation of the root depth z_r in the case of anisotropic stands, based on the volume of roots in relation to the total soil volume. The grey area (right) represents total root volume $Z_{tr} \times A_{tr}$ (A_{tr} being width × 1 meter in the row), and the red rectangle (on the left) represents the area (A_r, width × 1 meter in the row), where the majority of the roots are located defining a volume of $Z_r \times A_r$. b) Root biomass distribution in the soil profile, with the area (considering 1 m in the row) where the majority of the roots are located highlighted in red. The different tonalities from dark green to light yellow represent decreasing root densities.

WP represents, for engineering purposes, a general condition for which most plants cannot extract water from the soil, because the total pressure (of the water remaining in the soil) became too negative: plants would wilt in a definitive way, with the respective SWC corresponding to θ_v at WP ($\theta_{v,WP}$).

Common values of water volumetric fraction at FC and WP, for different soil textures, are presented (Table 2), for instance, by Allen *et al*. (1998). Both limits, especially the last, depend on other aspects but, even using such simple concepts, these thresholds had proven to be of practical value.

Table 2. Water volumetric fraction at field capacity ($\theta_{v,FC}$) and at permanent wilting point ($\theta_{v,WP}$), for different soil types (Allen *et al*., 1998)

Soil type	**Water volumetric fraction**	
	$\boldsymbol{\theta_{v,FC}}$	$\boldsymbol{\theta_{v,WP}}$
	(m^3/m^3)	(m^3/m^3)
Sand	0.07 - 0.17	0.02 - 0.07
Loamy sand	0.11 - 0.19	0.03 - 0.10
Sandy loam	0.18 - 0.28	0.06 - 0.16
Loam	0.20 - 0.30	0.07 - 0.17
Silt loam	0.22 - 0.36	0.09 - 0.21
Silt	0.28 - 0.36	0.12 - 0.22
Silt clay loam	0.30 - 0.37	0.17 - 0.24
Silty clay	0.30 - 0.42	0.17 - 0.29
Clay	0.32 - 0.40	0.20 - 0.24

The corresponding values of θ_v can be obtained using *in situ* measurements (preferably), laboratory (mainly using specific thresholds for pressure, a convenient but rough approach), algorithms based on soil properties or even simple tables, based on soil texture. As an example of calculation procedures to obtain $\theta_{v,FC}$ and $\theta_{v,WP}$ based on soil physics characteristics easy to measure (such as particle-size distribution, bulk density and organic matter content), pedo-transfer functions can be useful (Gonçalves *et al*., 1997).

2.2. Total available water

Total available water for plants (*TAW*, L/m^2 or mm) is usually estimated for irrigation engineering applications, and corresponds to the difference between *S* at FC ($\theta_{v,FC} \times z_r$) and *S* at WP ($\theta_{v,WP} \times z_r$). In case of a homogenous soil and for a total root depth of $z_{r,a}$:

$$TAW = S_{FC} - S_{WP} = (\theta_{v,FC} - \theta_{v,WP}) \times z_{r,a} \qquad [5]$$

where:

$\theta_{v,FC}$ is the volumetric water fraction in soil at field capacity ($m^3\ m^{-3}$);

$\theta_{v,WP}$ volumetric water fraction in soil (or soil water content) at wilting point ($m^3\ m^{-3}$);
$z_{r,a}$ average root depth for a given stage of plants development (same unit as *TAW*), see Figure 3a.

As z_r progresses with plant growth, *TAW* also increases. If the soil is not homogeneous in the vertical direction, *TAW* changes not only due to z_r but also because roots when colonizing layers of increasing depth arrive to soil layers with different hydraulic properties. The equation will then be modified, to consider each layer of the soil with its specific values of θ_{FC} and θ_{WP}:

$$TAW = \Sigma\ [(\theta_{v,FC,i} - \theta_{v,WP,i}) \times z_i] \qquad [6]$$

where the subscript i (i = 1 to n) denotes the successive layers considered from the soil surface till the total root depth (z_r) in a certain stage of root development.

Again, if the values of these two parameters are not given in volumetric fraction as above (θ_v), but in mass fraction (θ_m), the expected changes in apparent soil density should be taken into account:

$$TAW = \Sigma\ [(\theta_{m,FC,i} - \theta_{m,WP,i}) \times ds_i \times z_i] \qquad [7]$$

In order to avoid ambiguities in relation to the concept of percentage of *TAW*, it is preferable the use specific units for SWC (e.g. 0.2 $m^3\ m^{-3}$).

Example 3:
In a given soil, the weighted average soil water content, expressed as mass fraction (θ_m), is 0.150 and 0.025 ($kg\ kg^{-1}$) at FC and WP, respectively. What is the total available water (*TAW*) in 60 cm of that soil (average soil bulk density $d_s = 1.2$, vide eq. 3).

Answer:
$TAW = (\theta_{m,FC,} - \theta_{m,WP,}) \times d_s \times z = (0.150 - 0.025) \times 1.2 \times 600 = 90$ mm.

Example 4:
The characteristics of one given soil, considering several layers with 20 cm each, are presented in the table below.
Calculate the total available water (*TAW*) for two dates, when $z_r = 0.3$ m (day i), $z_r = 0.8$ m (day i + Δt).

Layer depth, z_r (cm)	**$\theta_{m,FC}$ (kg kg^{-1})**	**$\theta_{m,WP}$ (kg kg^{-1})**	**d_s**
0 - 20	0.29	0.17	1.16
20 - 40	0.31	0.19	1.21
40 -60	0.32	0.18	1.15
60 - 80	0.34	0.20	1.17

Answer:
with $z_r = 0.3$ m, $TAW = \Sigma\ [(\theta_{m,FC,i} - \theta_{m,WP,i}) \times ds_i \times z_i] = [(0.29 - 0.17) \times 1.16 \times 200] + [(0.31 - 0.19) \times 1.21 \times 100] = 27.84 + 14.52 = 42.36$ mm;
with $z_r = 0.8$ m, $TAW = \Sigma\ [(\theta_{m,FC,i} - \theta_{m,WP,i}) \times ds_i \times z_i] = [(0.29 - 0.17) \times 1.16 \times 200] + [(0.31 - 0.19) \times 1.21 \times 200] + [(0.32 - 0.18) \times 1.15 \times 200] + [(0.34 - 0.20) \times 1.17 \times 200] = 27.84 + 29.04 + 32.2 + 32.76 = 121.84$ mm.

2.3. Readily available water and allowable depletion factor

As more subscripts are now necessary to refer the thresholds, for simplification, the v subscript is omitted in the following, whenever the units clearly denote that volumetric fraction are used.

The soil water deficit (*SWD*) in a certain time is defined as:

$$SWD = (\theta_{v,FC} - \theta_v)\, z_r \qquad [8]$$

The way water is available to plants within the range defined above (between FC and WP) differs very much as SWC - here expressed as θ_v - decreases and *SWD* increases with time. It is classically accepted, on average and for engineering purposes, that there is a threshold after which *SWD* affects *ET*, corresponding this threshold to the lower edge (*LRAW*) of the so called readily available water (*RAW*), with its corresponding storage S_{LSAW} and water content θ_{LSWC}:

$$RAW = S_{FC} - S_{LRAW} = (\theta_{v,FC} - \theta_{LSWC})\, z_r \qquad [9]$$

Under this assumption, the fraction of *TAW* for which *ET* is not affected, is called the allowable depletion factor (*p*), being:

$$RAW = p\ TAW \qquad [10]$$

These practical concepts, not necessarily with a strict physical correspondence are nowadays largely used. It is well known that this fraction (depletion factor) depends on different conditions which interact, mainly soil properties, plant properties, *ET* rates (Ferreira, 2017), and even the irrigation method, that affects roots density (Paço *et al.*, 2012), in a quite dynamic process. Nevertheless, this topic, quite controversial since almost a century (e.g. Veihmeyer and Hendrickson, 1927), with important changes of paradigm discussed during the 50s-60s and generally accepted in the 80s, is out of the scope of this chapter. The soil water thresholds and other variables related to soil water are schematically represented in Fig. 4.

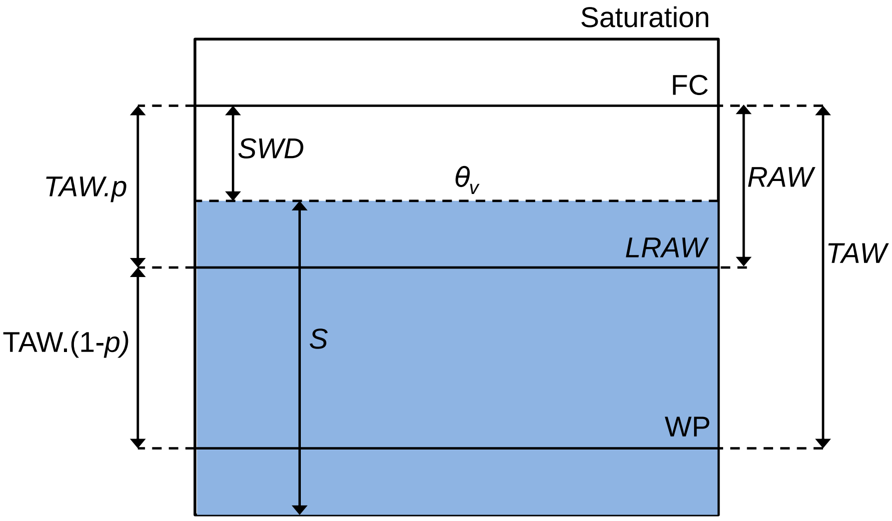


Figure 4. Schematic representation of the soil water thresholds (see text) Field capacity: FC, lower limit of RAW: LRAW, wilting point: WP, which correspond to certain values of SWC and water storage intervals (expressed in mm) defined by those thresholds (RAW: readily available water, TAW: total available water, on the right, and correspondent equations on the left, see equations 9 and 10).

Example 5:

Consider the following soil parameters: $\theta_{FC} = 0.40$ m^3 m^{-3}, $\theta_{WP} = 0.20$ m^3 m^{-3}, $p = 0.45$. The initial SWC is $\theta_i = 0.30$ m^3 m^{-3}.

a) How much water to apply (as an irrigation depth, I) to the root zone (z_r = 600 mm), in order to get that soil in a pre-selected condition defined as 90% of RAW?

b) Determine final SWC (θ_f).

Answer:

a) $SWD_i = (0.40 – 0.30) \times 600 = 60$ mm; $RAW = (0.40 - 0.20) \times 600 \times 0.45 = 54$ mm;

in order to get 90% of RAW,

one should have $SWD_f = (1 - 0.9) \times 54$ mm $= 5.4$ mm;

$I = SWD_f - SWD_i = 60 – 5.4 = 54.6$ mm;

b) $\theta_f = \theta_{FC} - SWD_f / z_r = 0.40 - 5.4/600 = 0.39$ m^3 m^{-3}.

Example 6:
One soil has the average SWC in volume (θ_v), at FC and WP of respectively 38% and 18% (0.38 and 0.18 m^3 m^{-3}) and $d_s = 1.2$. Considering a root depth $z_r = 40$ cm and the allowable depletion factor $p = 0.6$, estimate the soil water content in volume (θ_v) and mass (θ_m) fractions and the storage level of RAW (%), when this soil is at 65% of *TAW*.

Answer:
$TAW = (0.38 - 0.18) \times 400 = 80$ mm; $RAW = 80 \times 0.6 = 48$ mm.
$\theta_v = 0.38 - [(1\text{-}0.65) \times 80]/400 = 0.31$ m^3 m^{-3}; $\theta_m = 0.31/1.2 = 0.26$ kg kg^{-1}; remaining in the soil: $[48 - (1\text{-}0.65) \times 80]/48 = 41.6\%$ of *RAW*.

3. Evapotranspiration or its components as output variable of water balance equation

3.1. Measuring evapotranspiration or its components with lysimeters

Equations [1] or [2] can be used when all variables are measured or carefully estimated and *ETa* (or any of its components, transpiration and evaporation), is the unknown (this is mostly the case in research, in engineering applications *ETa* is often estimated from agrometeorological approaches). As described in section 1.2, a careful estimation of ΔS is then critical, requiring direct measurements (weighing lysimeters or very dense sampling for SWC, deep enough, with high quality measurements), unless ΔS can be roughly assumed as null (as often assumed in drainage lysimeters). In the following, these types of lysimeters are briefly described. We can consider that, if plants are absent, the lysimeters measure soil evaporation (*vide* 3.4) and, if plants are present and soil surface is protected from evaporation, they measure transpiration.

Lysimeters in general are measuring devices used to obtain *ETa*, through the application of the water balance equation (Equation 1), either *ETa* being equal to reference evapotranspiration (*ETo*), or crop evapotranspiration (*ETc*) or even below it. They are essentially a large tank or container, with its upper surface leveled with outside soil surface, full with the same soil of the surrounding land, maintaining the same disposition of the different soil layers as in the outside. It is very important in their installation, that the crop growing in the lysimeter has the same soil conditions in order to ensure similar water status and growth characteristics (height and leaf area index) as the crop in the surrounding field (Fig. 5), for which it is intended to measure *ET*. The border effect should be reduced to a minimum, in order to obtain representative measurements. Therefore, a lysimeter must be installed in an extensive and homogeneous crop parcel with a minimum of obstacles or disturbances between the lysimeter surface and the surrounding ground (Harbeck *et al*., 1966). They are normally used in annual crops, with many plants in each lysimeter, and obviously less commonly used in woody crops.

Relative to the measurement process, lysimeters can be divided into two major groups: weighing lysimeters and drainage or percolation lysimeters.

By using weighing lysimeters, one is able to weight with precision the volume of soil inside it, obtaining a precise value of S, and also of water inputs, *P* and *I*, and therefore *ETa* (Harbeck *et al*., 1966). In fact (see equation 1), *R* is avoided, *U* doesn't occur in most models and *D* is measured directly, being considered the method with greater accuracy. It provides data with a high temporal resolution, e.g. daily readings up to an interval of 10-15 minutes. In addition, this type of lysimeters allows measuring *ETa* under different irrigation strategies including deficit irrigation. This is because, in the case of weighing lysimeters, when using equation 1 simplified (since *R* is always null and *U* is null except in special models), ΔS is directly obtained by the weighing system.

Example 7:
On an automatic weighing lysimeter, with an area of 4 m^2, the following values were recorded for one given day: $P = 5$ mm; $D = 2.5$ L / 4 m^2; remainder mass variation = + 7.5 kg ($\Delta S = 7.5$ L / 4 m^2). Determine the *ET* value that occurred during that day.

Answer: $ET = P - D - \Delta S = 5 - 2.5/4 - 7.5/4 = 2.5$ mm.

Conversely, drainage lysimeters are much simpler installations, in which, by definition, there is no possibility to quantify precisely *S*, allowing *ET* to be determined as the difference between water inputs *I* and *P* and the water that drains out of the lysimeter, *D*. The values of ΔS could be estimated measuring the SWC inside the lysimeter, but the precision is quite limited. Therefore, in order to assume ΔS as null, the soil should be close to FC on a daily time scale, condition that it is also difficult to achieve with precision. The precise measurement of *I* is also critical. Consequently, in order to get reliable averages, this tool only allows measurements of maximal *ET* (*ETc*), usually *ETo* (grass in reference conditions, well irrigated), in a time scale of at least one week.

The use of drainage lysimeters, to obtain *ETo* as in its original definition, simple in theory but difficult to use in practice, has been almost abandoned, after the dissemination of an alternative *ETo* definition based on the outcome from the Penman-Monteith equation with grass parameters (Allen *et al*., 1998), approximately correspondent to the classical Penman equation (Penman, 1948).


a) b)

Figure 5. Installation of drainage lysimeters (a, left) before insulation, visible the thin upper wall and, in distance, the bags containing the soil extracted by layers for refilling and (b, right) installation of the pipes bringing the drained water to the wells where water flow would be recorded (1984, Coruche, Portugal).

When using the water balance approach to measure *ETa*, only weighing lysimeters can be considered for smaller time scales (from minutes).

3.2. Measuring evapotranspiration by means of plot scale measurements

Lysimeters are expensive to install, difficult to maintain and the location is fixed; their use, even restricted to research, has decreased. Another alternative to measure *ETa* through the application of the mass balance equation is by using measurements of SWC in the field, to obtain ΔS, which is generally performed at plot scale. SWC sampling is carried out at various points in the plot and at different depths along the soil profile. To perform the SWC measurements, several methods can be used such as the direct gravimetric approach (highly time consuming), or indirect methods such as neutron probes, capacitive probes, or others instruments, linking different physical properties to SWC. A time scale, of a week at least, is generally considered for the validity of the outputs from the water balance applied from measurements of SWC in an open field.

It is necessary to take into account the spatial heterogeneity of the soil, the plants distribution and the geometric position of the water distribution through the irrigation system. As stated above, there is a general consensus that this approach does not provide rigorous values, when the time scale is less than one week, but can provide insights on the dynamics.

The soil water balance equation when applied to the plot is often simplified assuming that, in a well irrigated plot, $R = 0$, $U = 0$, $D = 0$.

$$ETa = P + I - \Delta S \qquad [11]$$

When this equation is applied to a period between irrigations ($I = 0$) in which there is no precipitation, it can be further simplified, assuming the following form:

$$ETa = - \Delta S \qquad [12]$$

Example 8:

On a sandy loam soil, a crop is installed. At the beginning and at the end of a 5-days period, in which there was no rain or irrigation, the SWC was measured along the soil profile. Disregarding the contribution of the 5 cm surface layer and assuming that the roots do not extract water below 95 cm depth, estimate the average daily evapotranspiration that occurred during this period, considering the soil water measurements shown in the following Table.

Soil layer depths and water content in the soil profile ($m^3\ m^{-3}$)

Measurement Depth (cm)	Soil layer limits (mm)	Day 110	Day 115
10	50 - 150	0.26	0.18
20	150 - 350	0.24	0.19
50	350 - 650	0.24	0.22
80	650 - 950	0.25	0.24

Answer:

Considering the entire 5-day period $ET = -\Delta S$, being $S = \Sigma\ \theta_{v,i} \times Z_i$. Therefore, $ET = - \Sigma\ \Delta\theta_{v,i} \times Zi.$; $ET_{5\ \mathrm{days}} = - [(0.18 - 0.26) \times (150\text{-}50) + (0.19 - 0.24) \times (350\text{-}150) + (0.22 - 0.24) \times (650\text{-}350) + (0.24 - 0.25) \times (950\text{-}650)] = 27$ mm. $\overline{ET}_{daily} = 5.4$ mm d^{-1}.

3.3. Measuring in large time and space scales (small watersheds)

To determine ET on a large scale, one can use small experimental watersheds. These basins should be ideally small and homogeneous. Knowing the precipitation that occurred inside the experimental basin, recorded through udometers, and knowing the flow in a downstream section, an average *ET* that occurred in that basin can be estimated on an annual basis, when it is possible to assume that $\Delta S = 0$. In the case of these basins, equation 1 of the water balance takes the following form:

$$ETa = P + I - R \qquad [13]$$

When ΔS cannot be assumed as null, time consuming SWC measurements would be required at the watershed basis.

3.4. Measuring soil evaporation

The so-called soil evaporation (*Es*) is a component of total *ET* that corresponds to the water evaporated directly from the soil to atmosphere, and not via the plant. It can also be measured using the water balance equation, namely by means of small lysimeters (Fig. 6) from where plants are excluded and where ΔS is quantified by manual weighing, with a number of precautions. In this case, the fluxes at the bottom can be prevented (dry soil) or included in the weighing process by means of an attached isolated lower compartment (Fig. 6a, used for wet soil, e.g. near drippers). In order to reduce the error attributed to the fact that upward fluxes of water vapour from below the lysimeter into it, are not possible, the lysimeters have to be refilled with new soil, on a regular basis, which requires a special strategy, as described in Dammen *et al*. (1995), Conceição (2007) and Tezza *et al*. (2019).

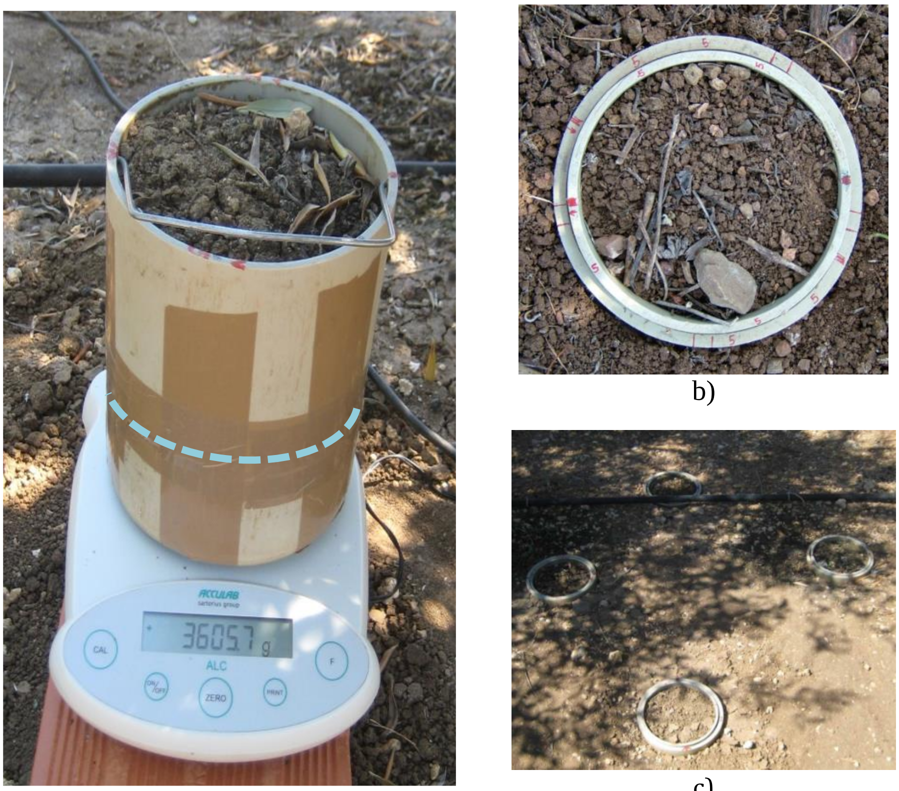
Thought this approach is time consuming, it provides data for modelling, used mostly for research purposes (e.g., Ritchie, 1972; Bonachela *et al*., 1999). For instance, it was used in peach and olive orchards, making it possible to get local adjusted models that allowed good model *Es* estimates on a seasonal basis.

Those values allowed to obtain the difference *ET-Es*, which should be equal to transpiration. Transpiration is often approached via sap flow techniques, much simpler to use than the techniques to obtain precise *ET* and *Es*. Notwithstanding, sap flow estimates can require specific calibrations.

Consequently, the time series for the difference *ET-Es* was then compared with such sap flow outputs, in order to correct long term sap flow data series (e.g., Paço *et al*., 2006; Ferreira *et al*., 2008, 2012; Conceição *et al*., 2017, Tezza *et al*., 2019). This is relevant because complete series (long-term data) are much more difficult to obtain with micrometeorological techniques than with sap flow techniques.

4. Applications of the water balance equation using actual evapotranspiration as input variable

The use of the water balance equation depends upon the aim and context. In normal conditions, it obviously does not make sense to use the water balance equation to determine easily measured variables (such as *P* or even *I*). Consequently, if *ETa* is an input variable, the equation usually serves to estimate ΔS and, this way, control soil water status (mostly applications for irrigation scheduling purposes). In this context, usually *ETa* is estimated from semi-empirical equations (e.g., Penman-Monteith or Hargreaves-Samani).



b)

c)

a)

Figure 6. Small lysimeters for soil evaporation measurements; (a) lysimeter with lower chamber (below blue line) to receive water drained through a metallic fine net , (b) lysimeter placed in its definitive location inside an outer wall in a fixed position and (c) four lysimeters placed at different locations in relation to shade and drippers, in Ferreira do Alentejo (details in Tezza et al., 2019).

Example 9:
For a certain location, initial SWC in the root zone (z_r = 40 mm) was θ_i = 0.30 m^3 m^{-3}. During a time interval of several days, P = 5 mm and I = 20 mm were measured, being *ETa* estimated as 33 mm. Calculate the final SWC (θ_f).

Answer:
$P + I - ETa = \Delta S = (\theta_f - \theta_i)\, z_r$; therefore, 5 + 20 – 33 = (θ_f – 0.30) × 40. Consequently, θ_f = 0.10 m^3 m^{-3}.

Important is to stress that, if the final estimated θ (as in example 9) would have been significantly different from the observed one, there would have been an opportunity to critically revise the parameters of *ETa* estimation, in a so-called self-learning process (e.g., Ferreira, 2017). Examples of this approach are shown in a companion paper.

Soil water balance can be used to perform irrigation scheduling, in which it is intended to determine the timing and depth of irrigations, as well as, to follow the *SWC* along the crop growing season, to identify the level of stress to which the plants are subjected, namely in the case of deficit irrigation (vide 4.1.2). Another use of the water balance in which *ETa* is an input variable includes the calculation of the seasonal irrigation requirements to be used, for instance, at the irrigation project level.

4.1. Irrigation scheduling

To estimate the evolution of the soil water storage (S) over time, enabling irrigation scheduling, equation 1 is modified as follows:

$$S_i = S_{i-1} + P + I + U - ETa - D - R \qquad [14]$$

where S_i is the water stored in the root zone in day i and S_{i-1} is the water stored in the previous day.

The soil water balance terms considered depend on the plot conditions and on the purpose of the calculation; in irrigated plots, it is often possible to consider some of the terms negligible, such as R or U.

When the soil water balance is applied for irrigation scheduling purposes, it is necessary to define the irrigation strategy (e.g. no stress or deficit irrigation) including the logistic options or restrictions relative to the timing at which irrigation can occur, as well as the volumes that it is possible to supply. This information is considered through two variables:

- the irrigation opportunity that determines the moment when irrigation should start and

- the irrigation depth, which defines the volume of water to be supplied to the crop per unit of surface area (irrigation depth, mm = L m^{-2}).

The type of irrigation system has a great influence on irrigation scheduling. While surface irrigation and travelling guns apply high water depths with large intervals between irrigations (if the soil storage and infiltration properties allow), when using center pivots or drip irrigation, a high frequency irrigation is often carried out with reduced irrigation depths. Furthermore, the production objectives also have a high degree of influence. In most crops, the objective is to grow the crop under no water stress aiming for maximum production. However, for some crops, a balance between production and quality is desirable, being achieved through a deficit irrigation strategy. The different irrigation scheduling strategies can be grouped into two main groups: irrigation under no stress and deficit irrigation (DI).

4.1.1. Irrigation under no stress

When irrigation management is aimed for no stress, irrigation is carried out in order to avoid the occurrence of crop water stress, ensuring that the volume of water required to fully satisfy the crop water requirements is available. In this case, the lower limit of the soil water storage is not allowed to go below *LRAW* (when *RAW* is exhausted). Consequently, in a traditional irrigation scheme, the irrigation should occur when the *LRAW* threshold is reached, being applied an irrigation depth that allows the soil to be replenished to the FC (*TAW* fully replenished), being the maximal irrigation depth to be applied given by:

$$I_{max} = (\theta_{v,FC} - \theta_{v,WP}) \times p \times z_r = (\theta_{v,FC} - \theta_{v,LRAW}) \times z_r \qquad [15]$$

We recall that the v subscript will be omitted, for simplification, whenever the units clearly denote that volumetric fraction is used. This irrigation depth I_{max} (mm if z_r comes in mm) corresponds to the largest irrigation volume per unit of surface area that can be applied to the crop without originating deep percolation, implying the least number of irrigation applications. Of course, it is possible to apply smaller irrigation depths with higher frequency between applications and, in that case, lower and upper thresholds (for θ or S) should be adopted with a reduced difference between them, in order to produce such smaller irrigation depths.

In fact, it should be highlighted that the lower management threshold can be set to values other than S_{LRAW}. When it is desired to avoid the occurrence of water stress, in the case of high added value crops, a value above S_{LRAW} must be adopted. When it is intended to carry out deficit irrigation, the lower limit of S should be below S_{LRAW}, causing moderate water stress and reducing crop *ETa*.

Example 10:
Consider the following SWC values (volumetric fraction) measured in a given soil. Calculate when the next irrigation should take place, assuming a root depth of 500 mm, an average *ETa* of 7 mm day^{-1} and assuming that soil water storage (*S*) should not drop below 60 mm.

Measurement depth (cm)	**Soil layer limits (mm)**	**θ_v volume (10^{-2} m^3/m^3)**
5	0-100	13.8
15	100-200	25.5
25	200-300	28.7
40	300-500	22.9

Answer:
$S_{initial} = \Sigma\, \theta_{vi} \times z_i$ = 0.138 × (100 - 0) + 0.255 × (200 - 100) + 0.287 × (300 - 200) + 0.229 × (500 - 300) = 113.8 mm.
Since the lower soil storage is S_{final} = 60 mm, ΔS = 60 - 113.8 = - 53.8 mm. Knowing that $ETa = -\Delta S$, the maximum time interval allowed until the next irrigation is $\Delta t = \Delta S\ /\ \overline{ET}_{daily}$ = 53.8/7 = 7.7 days. Hence, the next irrigation should occur up to 7 days. This is true only if assuming there was no stress, which is impossible to know with the provided data. In order to be able to make an interpretation of this situation, it is necessary to know the soil parameters (SWC at FC and WP) and *p* (next example).

Example 11:
Consider the previous (Example 10) SWC values (volumetric fraction), measured in the soil, and assume the same crop parameters, an average *ETc* of 7 mm day^{-1}, and *p* = 0.4. Furthermore, consider that θ_{FC} and θ_{WP} are, respectively, of 0.27 m^3 m^{-3} and 0.10 m^3 m^{-3} on average. Estimate:
a) the maximal irrigation depth and interval, under no stress;
b) the values of S_{LRAW} (in mm) and the correspondent average SWC (as volumetric fraction);
c) the meaning in percentage of *TAW* of S_{final} = 60 mm (example 10);
d) the amount of water to apply (irrigation depth) to get the soil at FC, if the condition before irrigation is the one indicated in the table of Example 10 ($S_{initial}$ = 113.8 mm);

e) the amount of water to apply (irrigation depth) to get the soil at 75% of *TAW*, if the lowest threshold selected before irrigation is 50% of *TAW*.

Answer:

a) $I_{\max} = RAW = TAW \times p = (\theta_{FC} - \theta_{WP}) \times z_r \times p$, i.e. $I_{\max} = (0.27 - 0.10) \times 500 \times 0.4 = 85 \times 0.4 = 34$ mm;

b) $S_{LRAW} = S_{FC} - RAW = (\theta_{FC} \times z_r) - RAW = 0.27 \times 500 - 34 = 135 - 34 = 101$ mm; then $\theta_{LRAW} = 101/\ 500 = 0.202$ m^3 m^{-3};

c) $S_{final} = \theta_{final} \times z_r$. *So*, $\theta_{final} = 60/500 = 0.12$ m^3 m^{-3}, i.e. between θ_{LRAW} and θ_{WP}. Consequently, there was stress already, with $ETa < ETc$.

This total storage of 60 mm corresponds to 10 mm above WP which is $(\theta_{WP} \times z_r) = 0.1 \times 500 = 50$ mm.

In percentage of *TAW*, it is $(S_{final} - S_{WP})/(S_{FC} - S_{WP}) = 10/85 = 0.117$, i.e., 11.7% of *TAW* remains in the soil;

d) to replenish the soil till FC the irrigation depth (I) should be $I = S_{FC} - S_{initial} = 0.27 \times 500 - 113.8 = 135 - 113.8 = 21.20$ mm

e) $I = (0.75 - 0.50) \times 85$ mm $= 21.25$ mm

4.1.2. Deficit irrigation

When using a deficit irrigation (DI) strategy, the crop is subject to a certain level of water stress, which leads to a reduction in *ETa* possibly causing yield losses. Its adoption may be due to the unavailability of water resources to satisfy the totality of the crop water requirements or be intentional, as in the case of vineyards and olive groves, in which a balance between production and quality is desirable. DI may affect the entire crop cycle, or be applied only during the crop least sensitive periods, with reduced effects on production, as is aimed with controlled DI.

In DI, the lower threshold of the soil storage is below S_{LRAW}. After reaching this management threshold, an irrigation depth is applied, which will increase the water storage in the soil (S) to a value that may be higher or lower than S_{LRAW}.

According to the water stored in the soil and corresponding *SWD*, *ETa* is obtained using the stress coefficient (*Ks*) that, in the lack of locally adjusted function, can be estimated according to equation 16 (Allen *et al.*, 1998).

$$Ks = 1 \text{ if } S \geq S_{LRAW} \text{ and } Ks = \frac{TAW - SWD}{(1\text{-}p)\ TAW} \text{ if } S < S_{LRAW} \qquad [16]$$

being *ETa* given by:

$$ETa = Ks \times ETc \qquad [17]$$

where *ETc* is crop evapotranpiration under standard conditions, namely under optimum soil water conditions, as defined in Allen et al. (1998).

Example 12:
Consider a soil with θ_{FC} and θ_{WP} of 0.32 m^3/m^3 and 0.18 m^3/m^3, respectively, and assume a constant root depth of 500 mm, p = 0.45. Determine *ETa*, knowing that *SWD* was 61 mm, early in the day, and *ETc* = 5.9 mm.

Answer:
First it is necessary to get $TAW = (\theta_{FC} - \theta_{WP}) \times z_r = (0.32 - 0.18) \times 500 =$ 70 mm and
$S_{LRAW} = S_{FC} - RAW = \theta_{FC} \times z_r - p \times TAW = 0.32 \times 500 - 0.45 \times 70 = 160 - 31.5 = 128.5$ mm.
Early in the day *SWD* = 61 mm.
Therefore $S = S_{FC} - SWD = 160 - 61 = 99$ mm.

Since $S = 99 < S_{LRAW} = 128.5$ mm, the crop is under water stress and it is necessary to compute *Ks*.

$$Ks = \frac{TAW - SWD}{(1\text{-}p)\ TAW} = \frac{70 - 61}{(1\text{-}0.45)\times 70} = 0.23.$$

Then, $ETa = Ks \times ETc = 0.23 \times 5.9 = 1.34$ mm (rough estimate).

ETa estimation can be improved, if daily average *SWD* is computed.
Ks would be slightly different:

$$\overline{SWD} = \frac{SWD_{initial} + SWD_{final}}{2} = \frac{61 + (61 + 1.34)}{2} = 61.7 \text{ mm}.$$

$$Ks = \frac{TAW - SWD}{(1\text{-}p)\ TAW} = \frac{70 - 61.7}{(1\text{-}045)\times 70} = 0.22;$$

$ETa = Ks \times ETc = 0.22 \times 5.9 = 1.30$ mm.

Example 13:
Consider the same conditions of Example 12 (θ_{FC} = 0.32 $m^3\ m^{-3}$; θ_{WP} = 0.18 $m^3\ m^{-3}$; z_r = 500 mm, p = 0.45) and the following series of *ETc*.

a) Estimate Ks for the four days after *LRAW* is attained (i.e., *RAW* has been used).
b) Suppose the aim is to let *Ks* decrease down to 0.7 and irrigate up to 30% of *RAW*. Determine the irrigation depth, *I*.

Day after *LRAW* was attained	***ETc* (mm day^{-1})**
1	3.8
2	5.5
3	4.7
4	2.9

Answer:

a) First it is necessary to calculate $TAW = (\theta_{FC} - \theta_{WP}) \times z_r = (0.32 – 0.18) \times 500 = 70$ mm.
Soil water deficit (*SWD*) when *LRAW* is attained is
$SWD_0 = RAW = TAW \times p = 70 \times 0.45 = 31.5$ mm.
In day 1, for example, $SWD_1 = SWD_0 + ETa_1 \approx SWD_0 + ETc_1 \times Ks_0$ (considering $Ks_1 \approx Ks_0$, being the first still unknown).
In order to avoid more steps in an iterative process, a first round is considered enough, for simplification. Consequently,
$SWD_1 = 31.5 + 3.8 \times 1 = 35.3$ and
$Ks = (TAW - SWD_1) / [(1\text{-p})\ TAW] = 70 – 35.3 / [(1\text{-}0.45) \times 70] = 0.90$.
The *Ks* values for the following days are presented in the table below.

Day after *LRAW* was attained	**SWD_i (mm)**	**Ks_i**
1	31.5 + 3.8 = **35.3**	0.90
2	35.3 + 5.5 × 0.90 = **40.3**	0.77
3	40.8 + 4.7 × 0.77 = **44.4**	0.66
4	45.5 + 2.9 × 0.66 = **47.4**	0.59

b) If establishing that critical $Ks = 0.7$, then the lower value for SWD_{lower} is obtained inverting Eq. 15:
$SWD_{lower} = TAW - Ks \times [(1\text{-p})\ TAW] = 70 – 0.7 \times [(1 - 0.45) \times 70] = 43.05$ mm.
If it is intended to irrigate to up to 30% of RAW, then the upper *SWD* is $SWD_{\text{upper}} = (1 - 0.3) \times \text{RAW} = 0.7 \times 31.5 = 22.05$ mm.
Therefore, the irrigation depth $I = (\text{SWD}_{\text{lower}} – \text{SWD}_{\text{upper}}) = 43.05 – 22.05 = 21$ mm.
Again, to improve the accuracy of this *ETa* estimation, the average *SWD* for each day should be computed:

$\overline{SWD} = \frac{SWD_{initial}+SWD_{final}}{2}$ and *Ks* recalculated.

In several cases, e.g., woody Mediterranean species, it has been shown a lack of adequacy of this simplified model (Paço *et al*., 2012) or of its suggested parameters (Ferreira, 2017, 2020). Conversely, in other cases (less deep root systems) and in the range of moderate stress, the model outputs have compared relatively well with measurements (Ferreira, 2017).

4.2. Estimation of seasonal crop irrigation requirements

The soil water balance is also used to estimate the crop seasonal irrigation requirements, or the annual volumes of irrigation that must be supplied to a given crop. The calculation of these annual irrigation requirements is carried out by calculating the water balance throughout the entire irrigation season, frequently using crop parameters obtained through tabulated values (e.g., Allen *et al*., 1998), that can be adjusted using information from literature reporting measurements in similar conditions and/or plant parameters such as vegetation indexes (e.g., Williams and Ayars, 2005).

The annual irrigation requirements are used for irrigation management allowing determining the annual volume of water required and operating time of the irrigation systems and the costs associated with water, energy and labor.

The quantification of the irrigation requirements is normally carried out using soil water balance simulation models (e.g., ISAREG, Teixeira and Pereira, 1992; CROPWAT, Smith, 1998; SIMDualKc, Rolim *et al.*, 2006; IrrigRotation, Rolim and Teixeira, 2008, etc.) that implement the water balance equation (equation 14, defined in 4.1), with a varying degree of detail, in which the evolution of soil water storage over time is simulated.

The irrigation requirements defined in this chapter are net irrigation volumes, not considering losses and the lack of uniformity in the irrigation systems. The gross irrigation requirements are obtained by adding the leaching fraction and the losses due to the irrigation efficiency. This leaching fraction consists of a volume of water per unit surface area that must be added to the net irrigation requirements, in order to leach the salts accumulated in the soil, out of the root zone through deep percolation (Allen et al., 1998; Oliveira, 2011; Machado and Serralheiro, 2017).

The irrigation efficiency quantifies the losses of water that occur in the irrigation system installed in the plot including surface runoff, deep percolation, evaporation and wind drift, etc. Furthermore, transport efficiency in the irrigation network can be considered when appropriate (conveyance and distribution system), including leaks and evaporation of water from channels and reservoirs.

5. Final consideration

As stressed in the introduction to point 4., an important consequence of being able to deal with water balance concepts, is that it is possible to properly operate irrigation scheduling and critically compare results from modelling SWC with its direct measurements.

The time sequence of the possible difference between final estimated values of SWC (θ) in relation to the observed ones, provides an opportunity to adjust the parameters of *ETa* estimation, in a so-called self-learning process. A companion paper that follows this one illustrates the all procedure in low crops.

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