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Water Balance and Climate

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WATER BALANCE AND CLIMATE*

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1.Introduction

Water cycling in a watershed or in a cropped field can be characterized and quantified by a water balance, which is the computation of all water fluxes at the boundaries of the system under consideration. It is an itemized statement of all gains, losses and changes of water storage within a specified elementary volume of soil. Its knowledge is of extreme importance for the correct water management of natural and agro-systems. Gives an indication of the strength of each component, which is important for their control and to ensure the utmost productivity with a minimum interference on the environment.

Let us make a panoramic overview of the SOIL-PLANT-ATMOSPHERE system in relation to agricultural production. The atmosphere rests over the soil and the plant connects both, growing upwards (shoot) and downwards (root). Our interest lies in the plant, more specifically in its yield, which is a function of the available energy, the climate, the soil, the crop management, the genotype, …. (Figure 1). The fundamental equation for crop yield is based on the photosynthesis process, by which carbon dioxide from the atmosphere and water from the soil are combined to produce sugar inside chloroplasts of green plants, using solar energy:

$$\frac{1M\ CO_2}{atmosphere} + \frac{1M\ H_2O}{soil} + \frac{E}{atmosphere} \quad \rightarrow \quad \frac{1M\ CH_2O}{yield} + \frac{1M\ O_2}{atmosphere}$$

The energy source E is the Sun, accounted as global radiation GR, available for the process as net-radiation NR. This energy E defines the air temperature T_{air} and together with the

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rainfall R they are the main definers of the prevailing CLIMATE , which controls crop production. The rate

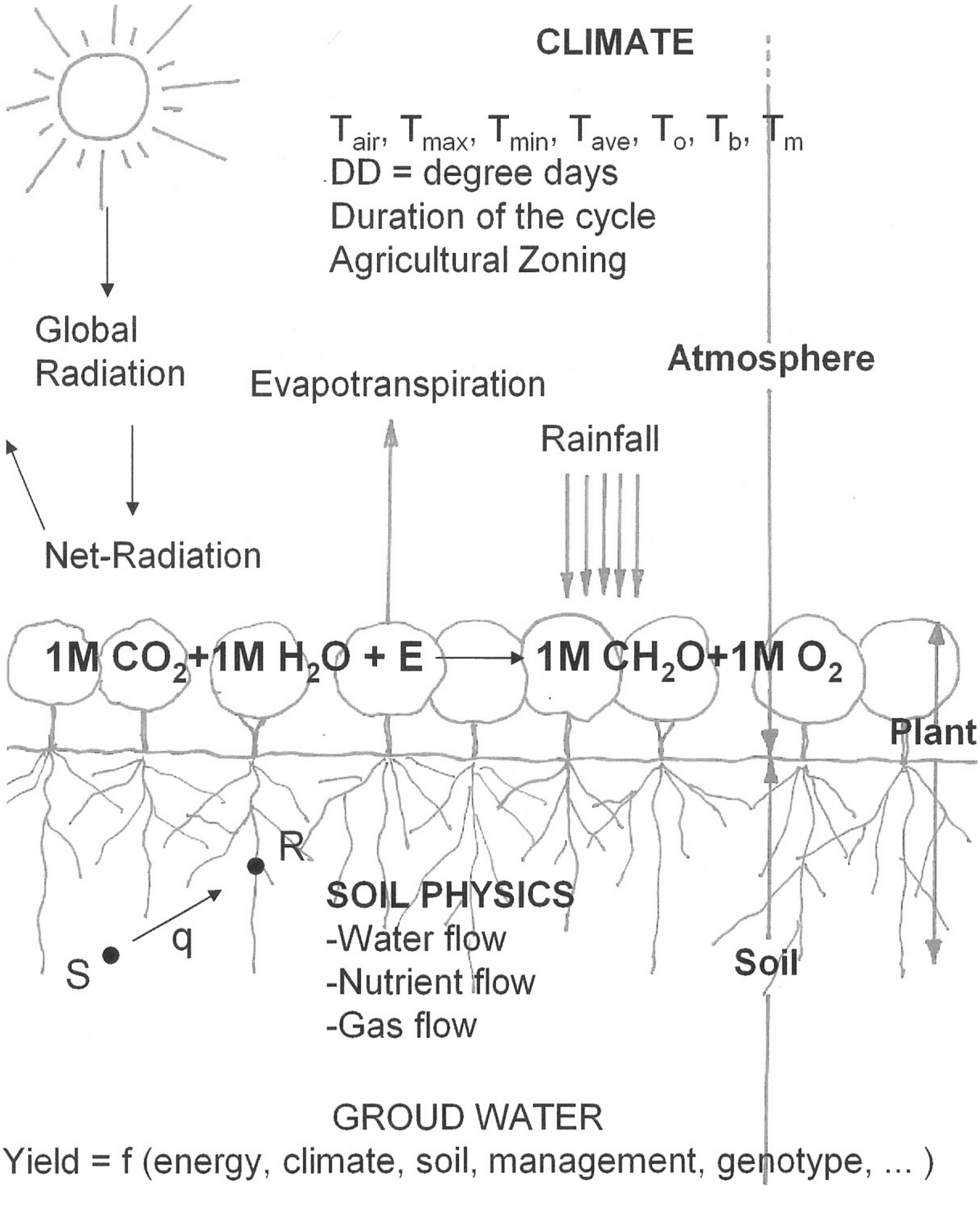
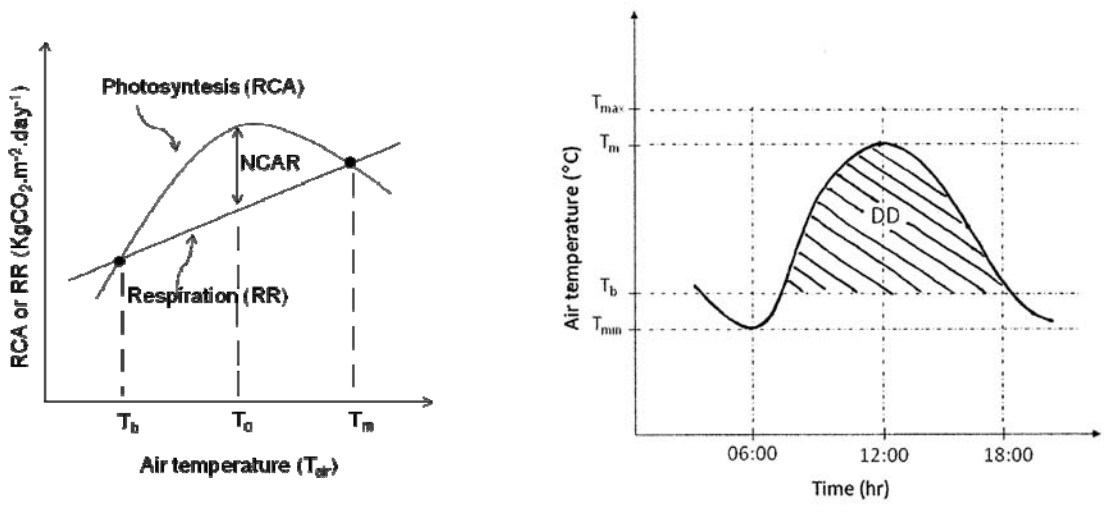


Figure 1: The soil – plant – atmosphere system in relation to crop production.

of assimilation of atmospheric CO_2 by plants RCA will finally determine the yield. It can be defined as the mass or volume of CO_2 that is absorbed per unit of crop projection on the ground, per unit of time, which could be kg. m^{-2} day^{-1}.

In order to grow and live, plants also spend energy through the respiration process, which is essentially the inverse reaction, by which plant sugar is burned by oxygen resulting carbon dioxide and water. For a plant to build up in yield, the sugar production by photosynthesis has to overcome the sugar consumption by respiration, or respiration rate RR. Figure 2 shows schematically the rates of these two processes as a function of T_{air} . The photosynthesis assimilation rate increases with temperature, passes through a maximum and decreases as it becomes too warm. The respiration rate essentially increases linearly with temperature. Each plant species has its own and typical shape of Figure 2a. For temperatures below T_b, the lower basal temperature, the net CO_2 assimilation rate NCAR = [RCA – RR] becomes negative and the plant consumes its energy; for temperatures above T_m, the upper basal temperature the net assimilation rate becomes again negative. Within the temperature interval T_m - T_b , the plant accumulates yield, with an optimum at T_o , when the net assimilation rate is maximum.



a b

Fig 2. Temperature relations for plant growth: a) Photosynthesis rate (RCA), respiration rate (RR), and net carbon assimilation rate (NCAR); b) Degree Day (DD) concept for an equinox day in tropical/subtropical regions.

Plant growth models relate the available solar energy to the net carbon assimilation rate to evaluate crop growth curves. They can be very complex. Some simpler ways use the concept of degree-days DD is based on these temperatures. It represents an area on the daily air temperature graph, limited by T_b and T_m (Fig. 2b). The plant needs a total of energy to complete its life cycle and the DDs are conveniently used to follow the crop growth process. The warmer the climate, the quicker the plant sums-up the DDs necessary to complete its cycle, and the adaptation of a given plant to a given climate.

The soil supplies rainfall or irrigation water (together with nutrients) to the plant. This water flows from the soil through the plant to the atmosphere, where its energy is the lowest. The process is called evapotranspiration and depends on the atmospheric conditions. When the soil reservoir is at high water levels, evapotranspiration is maximal, and depends mostly on the atmospheric water demand. When soil water becomes short, soil physical characteristics play an important role and command the supply of water to the plant.

Soil water reaching the roots carries along the mineral nutrients essential for crop growth and development. Therefore, yield depends also on the rates by which these nutrients are supplied to the plant.

As described above, the process of agricultural production is very complex and several factors affecting it are out of man´s control. Many, however, can be managed in order to maximize the yield of each crop in each region. The water balance gives an important overview of the water regime and is an essential tool for an effective crop management.

2. Elementary Volume and Balance Components

Considering the whole physical environment of a field crop, we define an elementary volume of soil to establish the water balance, having a representative unit surface area (1 m^2), and a height (or depth) ranging from the soil surface (z = 0) to the bottom of the root zone (z = L), where z (m) is the vertical position coordinate (Fig. 3). In practice, the soil surface is never leveled or horizontal, in general presenting an ondulated relief with characteristic slopes in all directions. This complicates the definition of the soil surface plane at z = 0, but for our water balance purposes we consider z = 0 as a moving point A always following the soil surface, which does not mean that this plane is leveled.

We will consider only vertical water fluxes (or better water flux densities) along the vertical coordinate z. They correspond to amounts (volume) of water that flow per unit of cross-sectional area and per unit of time. One convenient unit for agricultural purposes is liters (L) of water per square meter (m^2) per day, which corresponds to $mm.day^{-1}$:

$1\ L = 10^3\ cm^3 = 10^6\ mm^3$

$1\ m^2 = 10^4\ cm^2 = 10^6\ mm^2$

$1\ L / m^2 = 10^6\ mm^3 / 10^6\ mm^2 = 1\ mm$

These fluxes are assumed positive when entering the elementary volume (gain), and negative when leaving (loss).

At the upper boundary plane at the soil surface, of the elementary volume (z = 0), rainfall (**p**) and irrigation (**i**) are considered gains (in particular cases snow, after melted, is also an input, and in general, dew and other minor processes are considered negligible). Reaching soil surface, p or I either infiltrate (gain) or runn-off (-r) the study area (loss) or runn-in (+r) the study area (gain) in cases there is slope and no surface water contol. Water evaporation from the soil surface (**e**), transpiration from plant surfaces (**t**), or evapotranspiration (**et = e + t)**) are losses.

At the lower boundary plane, the bottom of the root zone at z = L, the soil water fluxes ($\mathbf{q_L}$) can be gain (when upwards), sometimes called capillary flow, or loss (downward flow or drainage), representing the deep drainage component. Inside the elementary volume ($L\ m^3 = 1\ m^2 . L\ m$) we consider only soil water fluxes in the z direction q_z, so that lateral fluxes q_x and q_y (Fig. 3) are considered zero (no lateral losses)

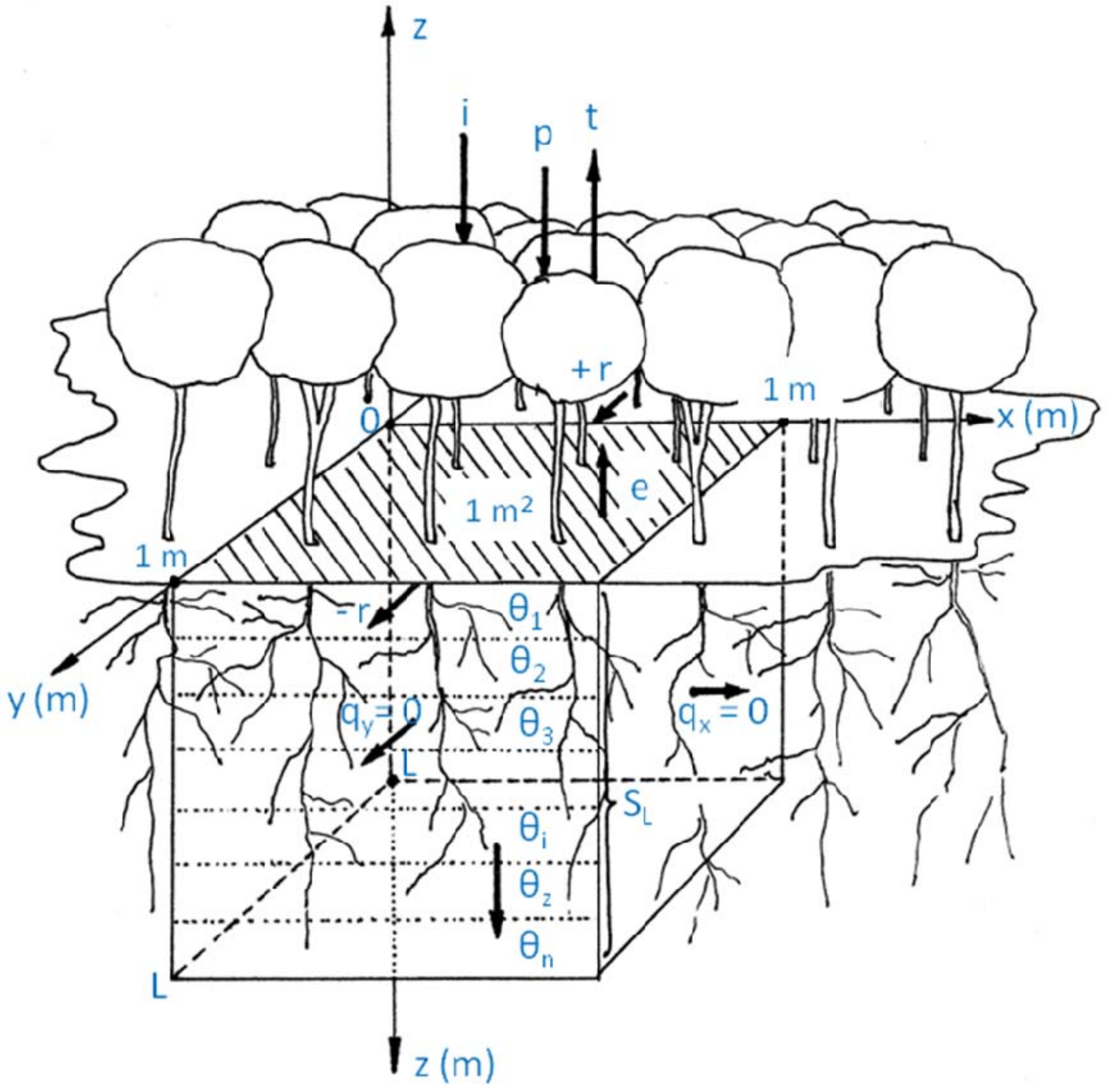


Figure 3: Schematic view of the volume element and of the fluxes that compose the water balance

The change in soil water storage ΔS is the result of the balance, being positive when the profile has a net gain of water, and negative for a net loss. S is defined by equation 2, bellow.

3.The Water Balance

The water balance is an expression of the mass conservation law, which includes the summation of all above discussed flux densities **f** that enter or leave the elemental volume:

$$\sum f = \int_0^L \frac{\partial\theta}{\partial t} dz \qquad (1)$$

where θ is the soil water content ($m^3.m^{-3}$) inside the elementary volume, t the time (day) and **f** stands for the flux densities **p**, **i**, **t**, **e** (or **et**), **r** and **q**. Their sum gives rise to changes in soil water contents ∂θ/∂t in time, which integrated over the depth interval of the elementary volume, z = 0 and z = L, represents the change in soil water storage S. Therefore, equation (1) can be rewritten as:

$$p + i - et \pm r \pm q_L = \frac{\partial S}{\partial t} \qquad (1a)$$

where S is defined by

$$S = \int_0^L \theta dz \qquad (2)$$

which by the trapezoidal rule becomes

$$S = \sum_{i=1}^{n} \theta_i \Delta z = [\theta_1 + \theta_2 + \theta_i + ... + \theta_n]\Delta z = \bar{\theta} L \qquad (2a)$$

and when expressing L in mm, S is also given in mm.

Equation (1a) is an instantaneous view of the balance. When integrated over a time interval $\Delta t = t_f - t_i$, in days, it yields amounts of water (mm):

$$\int_{t_i}^{t_f} (p + i - et \pm r \pm q_L)dt = \int_{t_i}^{t_f}\int_0^L \frac{\partial \theta}{\partial t} dzdt \qquad (3)$$

or

$$P + I - ET \pm R \pm Q_L = \Delta S = S(t_f) - S(t_i) \qquad (3a)$$

Equation 3a is an over time integrated view of the water balance.

The time interval for integration $\Delta t = t_f - t_i$ (equation 3) is chosen according to the objectives of the balance. Since water moves slowly in the soil, the choice of a too small Δt, e.g. less than 1 day, is seldom made. For annual crops common choices are 3, 7, 10, 15 or 30 day intervals. For long term experiments Δt can be of 1 year or more.

When all but one of the above components are known, the unknown is easily calculated algebraically. Five short examples are given below:

1. A soil profile stores 280 mm of water and receives 10 mm of rain and 30 mm of irrigation. It looses 40 mm by evapotranspiration. Neglecting runoff and soil water fluxes below the root zone, what is its new storage?
2. A soybean crop looses 35 mm by evapotranspiration in a period without rainfall and irrigation. It looses also 8 mm through deep drainage. What is its change in storage?
3. During a rainy period, a plot receives 56 mm of rain, of which 14 mm are lost by runoff. Deep drainage amounts to 5 mm. Neglecting evapotranspiration, what is the storage change?

4. Calculate the daily evapotranspiration of a bean crop which, in a period of 10 days, received 15 mm of rainfall and two irrigations of 10 mm each. In the same period, the deep drainage was 2 mm and the change in storage –5 mm.
5. How much water was given to a crop through irrigation, knowing that in a dry period its evapotranspiration was 42 mm and the change in storage was –12 mm? Soil was at field capacity and no runoff occurred during irrigation.

SOLUTIONS

n°	**P**	**+**	**I**	**-**	**ET**	**±R**	**$\pm Q_L$**	**=**	**ΔS_L**	**Answer**
1	10		30		-40	0	0		0	280 mm
2	0		0		-35	0	-8		-43	-43 mm
3	56		0		0	-14	-5		+37	+37 mm
4	15		20		-38	0	-2		-5	-3.8 $mm.day^{-1}$
5	0		30		-42	0	0		-12	+30 mm

As seen in the above discussion, WBs are very straight forward and their principle very simple. We have, however, assumed all components as deterministic values, without considering their variability in space and time. This is not the real case since all of them, when measured, present a stochastic behavior. Soils are not at all homogeneous so that infiltration, runoff, runinn and soil water fluxes vary from site to site, affecting the variability of S and ET. Climate elements like rainfall, air temperature and humidity, wind and solar energy also vary from site to site and in time. Plants vary a lot in spacing, height, leaf are and shape, rooting depth, variety, etc. This variability has been the subject of an enormous number of studies, first using classical statistics tools, involving mean values, medians, modes and variance analyses of all types. Somewhat later researchers started using tools of regionalized variables, geostatistics and state-space analysis. Therefore, the establishment of field WBs is not so straight forward as it is thought at first site. In the following item we will quickly discuss some of the main problems in evaluating WB components in light of their variability in time and space.

4.Discussion of the Components

4.1. Rainfall

Rainfall is easily measured with simple rain gauges which consist of containers of a cross sectional area A (m^2), which collect a volume V (liters) of rain, corresponding to a rainfall depth h (mm) equal to h = V/A. The problem in its measurement lies mostly in the variability of the rain in space and time. In the case of whole watersheds, rain gauges have to be well distributed, following a scheme based on rainfall variability data. For the case of small experimental fields, attention must be given to the distance of the gauge in relation to the water balance plots. Reichardt et al. (1995) is an example of a rainfall variability study, carried out in a tropical zone, where localized thunder-storms play an important role in the variability. Bruno et al. (2008) also discuss aspects of the number of rain gases to be used.

4.2.Irrigation

The measurement of the irrigation depth that effectively infiltrates into a given soil at a given place is not an easy task. Different methods of irrigation (sprinkler, furrow, drip, flooding, etc....) present great space variability in supplying water to the soil, which has to be taken into account.

4.3.Evapotranspiration (ET)

The loss of water through evapotranspiration occurs in the vapor phase, mainly through plant surface (stomata and cuticle) and through soil surface. Energy is needed to convert liquid water into vapor (latent heat that comes from the Sun and from the surrounding air). Some specific definitions are essential:

Reference Evapotranspiration, is the loss of water to the atmosphere of a green, shortly cut grass surface, directly exposed to the prevailing low air layer, growing on a soil with its available water capacity full. Under such well defined conditions it is a reference value for the location, representing the potential water loss of the green surface at each particular climatic condition, without interference of the soil because its water is freely available to the grass;

Pan Evaporation EP, is the loss of a free water surface, in general measured with a 1.2 m diameter pan exposed to the prevailing air conditions. In general, EP > ET_0 because the water is very free for the evaporation process. Since the measurement of is ET_0 is more difficult than EP, a pan coefficient K_p is used to transform one into the other:

$$ET_o = K_p.EP \qquad (4)$$

Maximum Evapotranspiration ET_m , is the potential evapotranspiration of any crop (excluding grass), which follows the definition of ET_0. A crop coefficient K_c is used to obtain ET_m for any crop when ET_0 is known:

$$ET_m\,(\text{crop}) = K_c.ET_o\,(\text{grass}) \qquad (5)$$

Actual or Real Evapotranspiration ET or ET_a , is the evapotranspiration that occurs under any soil-plant-atmosphere condition, and is the one present in the equation (3a) of the water balance. When soil water is readily available $ET_a = ET_m$, and as the soil dries out, the water flow to roots becomes restricted mainly due the reduction in soil hydraulic conductivity, and ET becomes steadily lower than ET_m. The extraction of the soil water by plants decreases, therefore, exponentially, tending to reach the Permanent Wilting Point PWP if there is no addition of water by rain or irrigation. Evapotranspiration can be measured independently using lysimeters or estimated from the balance, if all other components are known. For the measurement of ET_0, a great number of reports can be found in the literature, covering classical methods like those proposed by Thornthwaite, Braney-Criddle and Penmann-Monteith, which are based on atmospheric parameters such as air temperature and humidity, wind, solar radiation, etc. These methods have all their own shortcomings, mainly because they do not take into account plant and soil factors. Several models, however, include aspects of plant and soil, and yield much better results.

The main problem of estimating ET from the balance lies in the separation of the contribution of the components ET and Q_L, since both lead to negative changes in soil water storage ΔS. One important thing is that the depth L has to be such that it includes the whole root system. If there are roots below $z = L$, ET_a is under estimated. If L covers the whole root system and Q_L is well estimated, which is difficult as will be seen below, ET can be estimated from the balance. Villagra et al. (1995) discuss these problems in detail.

4.4.Runoff (R)

Runoff is difficult to be estimated since its magnitude depends on several factors, mainly rainfall intensity and duration, slope of the land, length of the slope, soil type, soil cover, etc. For very mild slopes, runoff is in general neglected. If the soil is managed correctly, using contour lines, even with significant slopes runoff is controlled and can be neglected. In cases it cannot be neglected, runoff is measured using small plots like ramps which are surrounded by a metal frame to maintain the rainfall water inside, about 20 m long and 2 m wide, covering areas from 40 to 50 m^2, with a water collector at the lower end. Again, the runoff depth h (mm) is the volume V (liters) of the collected water, divided by the area A (m^2) of the ramp. Several reports in the literature cover the measurement of R, either directly or through models (equations), and its extrapolation to different situations of soil, slope, cover, etc. This is a very well considered topic in other opportunities of this College.

4.5.Soil Water Fluxes at z = L, Q_L

The estimation of soil water fluxes at the lower boundary z = L, can be estimated using Darcy-Buckingham´s equation, integrated over the time:

$$Q_L = \int_{ti}^{tf} [K(\theta)\partial H/\partial z]dt \qquad (6)$$

where K(θ), (mm.day^{-1}), is the hydraulic conductivity estimated at the depth z = L, and ∂H/∂z (m.m^{-1}) the hydraulic potential head gradient, H (m) being assumed to be the sum of the gravitational potential head z (m), and the matric potential head h (m). Therefore it is necessary to measure K(θ) at z = L and the most common procedures used are those presented by Hillel et al. (1972), Libardi et al. (1980), Sisson et al. (1980), and more recently by Reichardt et al. (2002). These methods present several problems, discussed in detail in Reichardt et al. (1998). The use of these K(θ) relations involves two main constraints: (i.) the strong dependence of K upon θ, which leads to exponential or power models, and (ii.) soil spatial variability.

Two commonly used K(θ) relations are:

$$K = K_o \exp[\beta(\theta - \theta_o)] \qquad (7)$$

and

$$K = a\theta^b \qquad (8)$$

in which β, a and b are parameters obtained by fitting experimental data to the models, K_o the saturated hydraulic conductivity, and θ_o the soil water content at saturation. Reichardt et al. (1993) used model (7), and for 25 observation points of a transect on a homogeneous dark red latosol, obtained an average equation with $K_{oaverage}$ = 144.38 ± 35.33 mm.day^{-1}, and $\beta_{average}$ = 111.88 ± 33.16, obtaining an average equation:

$$K = 144.38 \exp[\, 111.88(\theta - 0.442)] \qquad (7a)$$

. in which $\theta_o = 0.442$ m^3.m^{-3}.

To understand the difficulties in using this average equation in the estimation of soil water fluxes, let's take an example in which the soil water content at the point we are making our calculations is $\theta = 0.4$ m^3.m^{-3}. Applying equation 7a we obtain K = 1.04 mm. day^{-1} . If this value of θ has an error of 2%, which is very small for field conditions, we could have θ ranging from 0.392 to 0.408 m^3.m^{-3}, and the corresponding values of K by applying equation 7a are: 0.43 and 2.55 mm.day^{-1}, with a difference of almost 500%, which means that in our flux calculation we will have very large error. This example shows in a simple manner the effect of the exponential character of the K(θ) relations. The standard deviations of K_o and β, shown above, reflect the problem of soil spatial variability in calculating soil water fluxes in WB studies. Added to this is the spatial variability of θ itself. Therefore, the direct measurement of Q_L using Darcy's equation is a difficult task, and several indirect methods have been suggested in the literature. Again, if we measure well all other water balance components of equation (3a), Q_L could be left as an unknown in the equation.

6.Changes in Soil Water Storage ΔS

Soil water storage S, defined by equation (2) is, in general, estimated either by: (i) direct auger sampling; (ii) tensiometry, using soil water characteristic curves; (iii) using neutron probes;

and (iv) using TDR probes. The direct sampling is the most disadvantageous due to soil perforations left behind after each sampling event. Tensiometry embeds the problem of the establishment of soil water characteristic curves, and neutron probes and TDR have calibration problems.

Once θ versus z data at fixed times are available, S is estimated by numerical integration, the trapezoidal rule being an excellent approach, and in this case, equation (2) becomes:

$$S = \int_0^L \theta dz \cong \sum \theta \Delta z = \bar{\theta} L \qquad (2a)$$

The changes ΔS are simply the difference of S values obtained at the different times t_i and t_f, that is $S(t_f)$-$S(t_i)$ as shown in Equation (3a).

A recent discussion of the establishment of field water balances is found in Silva et al. (2006) and Silva et al. (2007). The same data presented in these two papers was further analysed by Timm et al. (2010), using the state-space or better state-time analysis, indicating that this methodology presents several advantages over the classical statistical analysis..

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