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## Soil water storage as related to water balances

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**Soil water storage as related to water balances**

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Soil water storage (S, mm) is the quantification of the amount of water present in the soil reservoir, at any time t. Soil water at time t might be moving in any direction or be at equilibrium. On several instances soil water movement is relativelly slow and in such situation we will calculate S. It is the main component of water balances, that are the contabilization of the in and out water flows of an elemental soil volume (Figure 1).

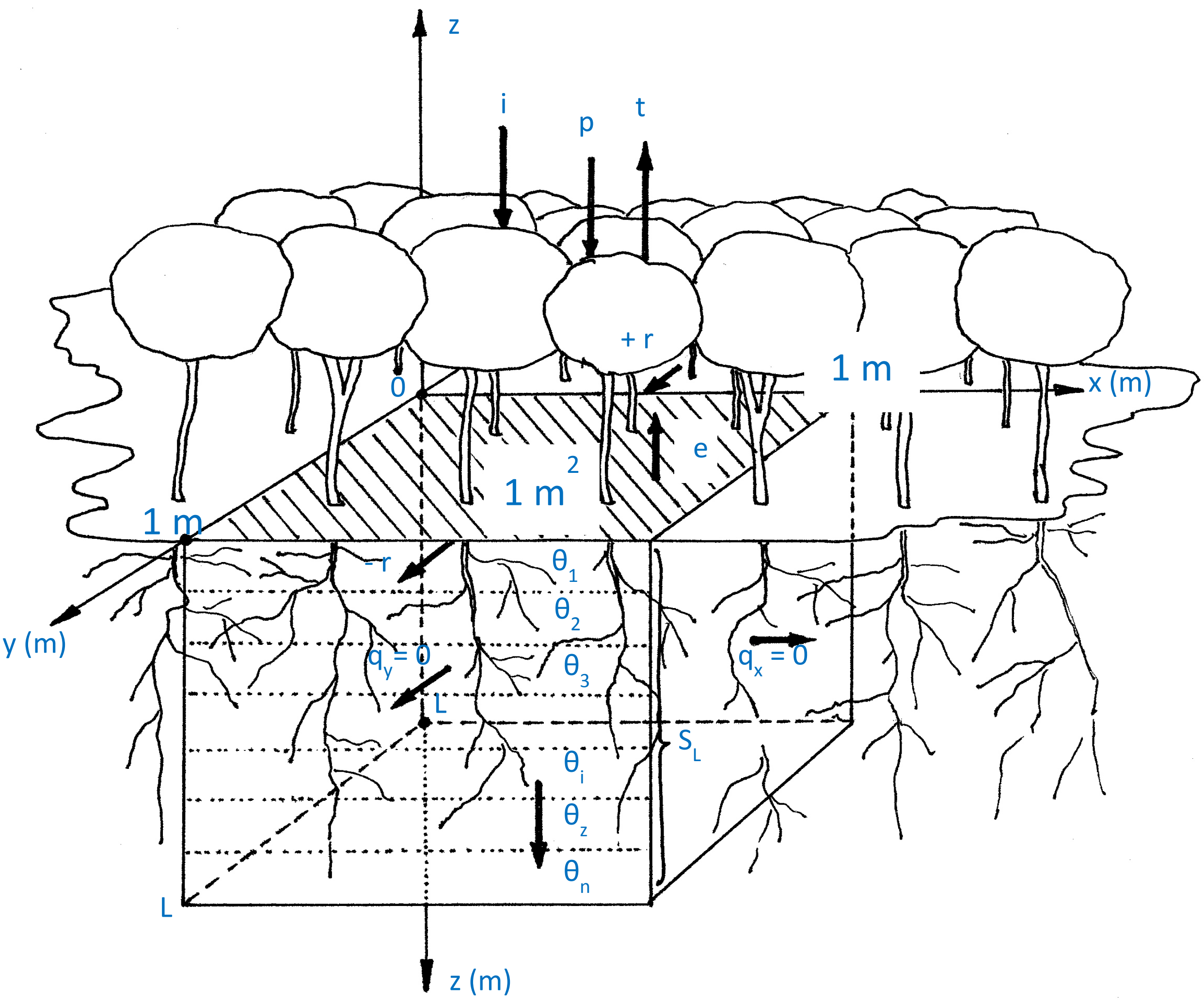


Figure 1. Schematic view of the volume element and of the fluxes that compose the water balance.

Figure 1 depicts a soil volume of surface area 1 $m^2$ and depth L m. This area is representative of a crop, and its size is related to the unit mm of water, which corresponds to 1 L $m^{-2}$. The depth L is the crop rooting depth. This volume contains solid material (soil particles) and pores, each occupying about 50% of the total volume, depending on soil type and compaction. The pore volume is occupied by water and air, and the volume of water per unit soil volume is called volumetric soil water content (ϴ, $m^3$ $m^{-3}$), also shown in Figure 1). We here consider that ϴ does not vary horizontally, only along the vertical coordinate (depth) z, m.

In our previous presentation we saw the WB equation in three equivalent forms

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

$$\int_{t_i}^{t_f} (p + i - et \pm r \pm q_L)dt = \int_{t_i}^{t_f}\int_0^L \frac{\partial \theta}{\partial t} dzdt \qquad (2)$$

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

And they all indicate that S or $\Delta S$ is the result of the contabilization of all in- and out fluxes in the elementary soil volume depicted in Fig. 1.

So, for S calculation, ϴ is measured in depth, at chosen intervals. There are several methods for the field measurement of ϴ. Classical methods extract samples to be taken to the oven for further calculation of ϴ, which are trerefore destructive methods, i.e., they are invasive, destroying soil structure and not allowing the measurent of ϴ at the same positions along time. This is a strong shortcomming because ϴ varies in time, and so does S. More recent methods are almost non destructive, they are neutron probes, gamma or X ray attenuation instruments, TDR and TD? Other methods are called indirect because they etimate ϴ from soil water potential measurements made with tensiometers, and the soil water retention curve.

Once having measured ϴ at several depths of interest, The concept of S is used the transform $m^3$ $m^{-3}$ into mm of water and is represented by the integral below:

$$S = \int_0^L \theta(z)dz \qquad (4)$$

We take as an example five ϴ measurements taken for the soil layers 0-0.2; 0.2-0.4;0.4-0.6; 0.6-0.8 and 0.8-1.0 m, as shown in Figure 2, below:

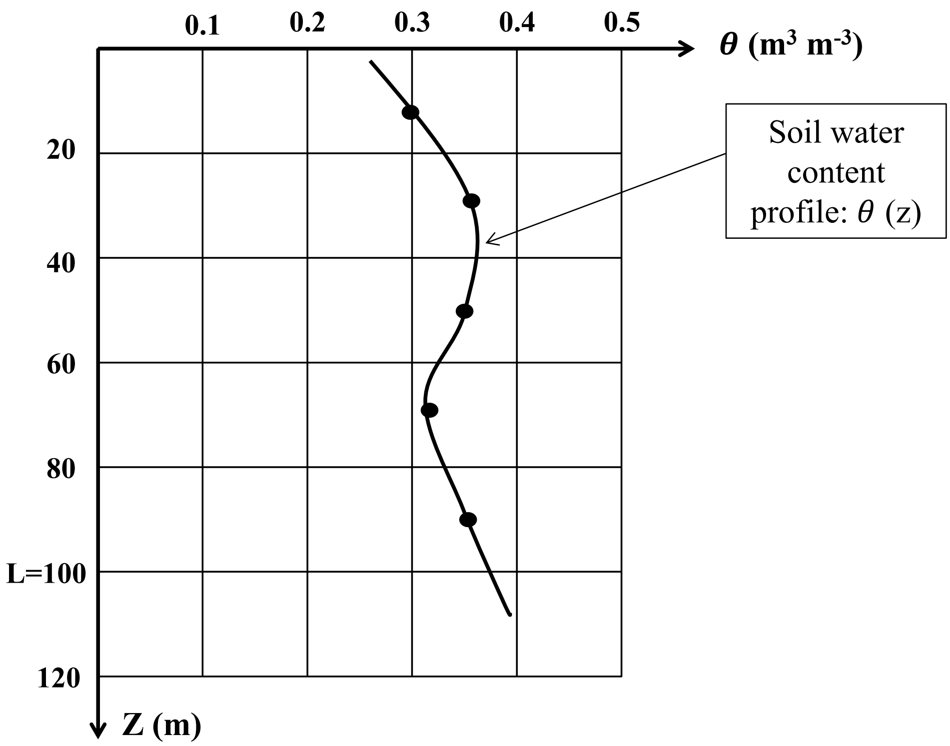


Figure 2. An example of a soil water content profile.

Because in most cases we do not have the function ϴ(z), the integral of equation 4 has to be solved numerically:

$$S = \sum_{i=1}^{5} \theta_i \Delta_z = \theta_1 \Delta_z + \theta_2 \Delta_z + \cdots + \theta_5 \Delta_z$$

$$= (\theta_1 + \theta_2 + \cdots + \theta_5)\Delta_z$$

$$= \left[\frac{(\theta_1 + \theta_2 + \cdots + \theta_5)}{5}\right] = 5\Delta_z$$

$$= \bar{\theta} * L$$

Which means that S is equal to the average of ϴ multipleied by the layer L over which the average $\bar{\theta}$ was taken. Since ϴ is dimensionless, if we express L in mm, S will also be obatined in mm. for the case of Fig. 2, we have:

$$S = (0.3 + 0.35 + 0.37 + 0.31 + 0.34)\ 200$$

$$= 0.344\ m^3\ m^{-3} \times 1000\ mm$$

$$= 334\ mm$$

Such S values are soil water storages at fixed times. However, as already mentioned, S varies in time due to in- and out flows of water in the elemental volume. Processes like infiltration and internal drainage can present very fast changes in S, so that measurements of Θ become difficult in order to apply equation 4. In Fig 3 we give an example of horizontal infiltration of water into a dry homogeneous soil:

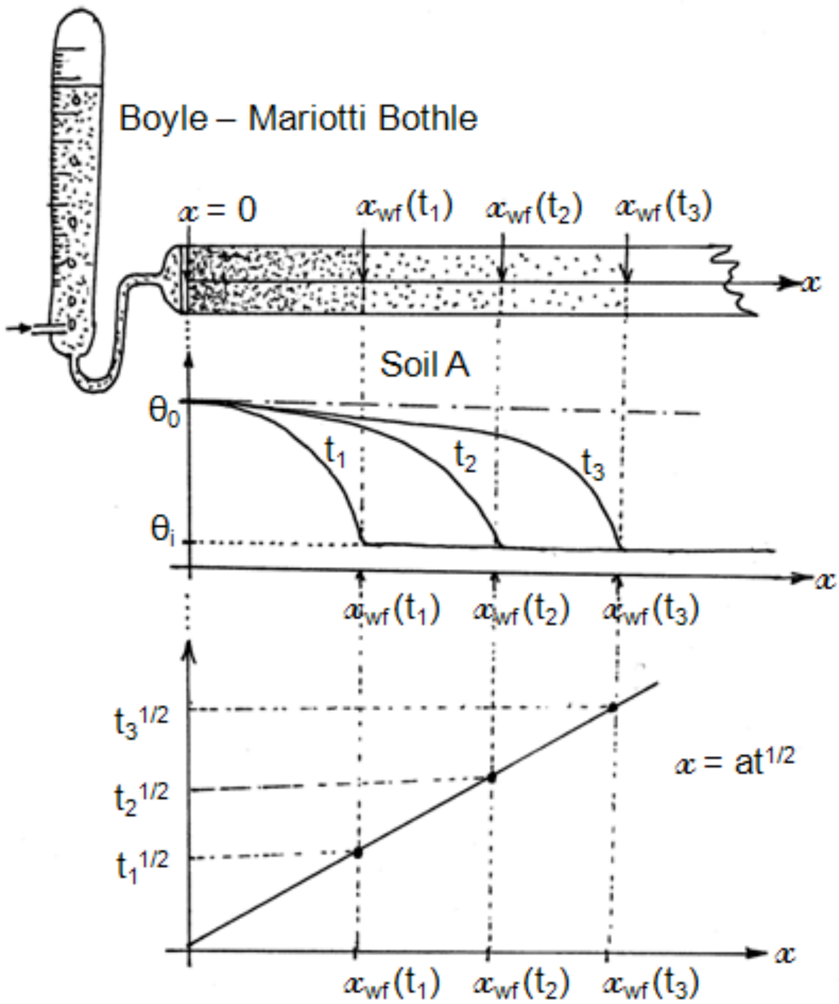


Figure 3. Experimental arrangement to study the horizontal infiltration using transparent acrylic columns for an initially dry soil and the advancement of wetting front as a function of the square root of time.

Note that for each time $t_1$, $t_2$, or $t_3$, the areas between the water contents $\Theta_i$ and $\Theta_0$, and the Θ profile are equal to the integral of equation 4, and therefore the storages $S_1$, $S_2$ and $S_3$ for the respective times. They also represent the cumulative infiltrations I of water into the soil.

In such cases, because water moves relatively fast, measurements of Θ by gamma of X ray attenuation, or by soil tomography, are excellent choices.

In several field water movement studies water moves relatively slow, and daily measurements can be sufficient. One good example of such processes is the internal drainage of a soil profile, during a hydraulic conductivity K(Θ) measurement experiment, when soil surface is covered by a plastic sheet (Figure 4):

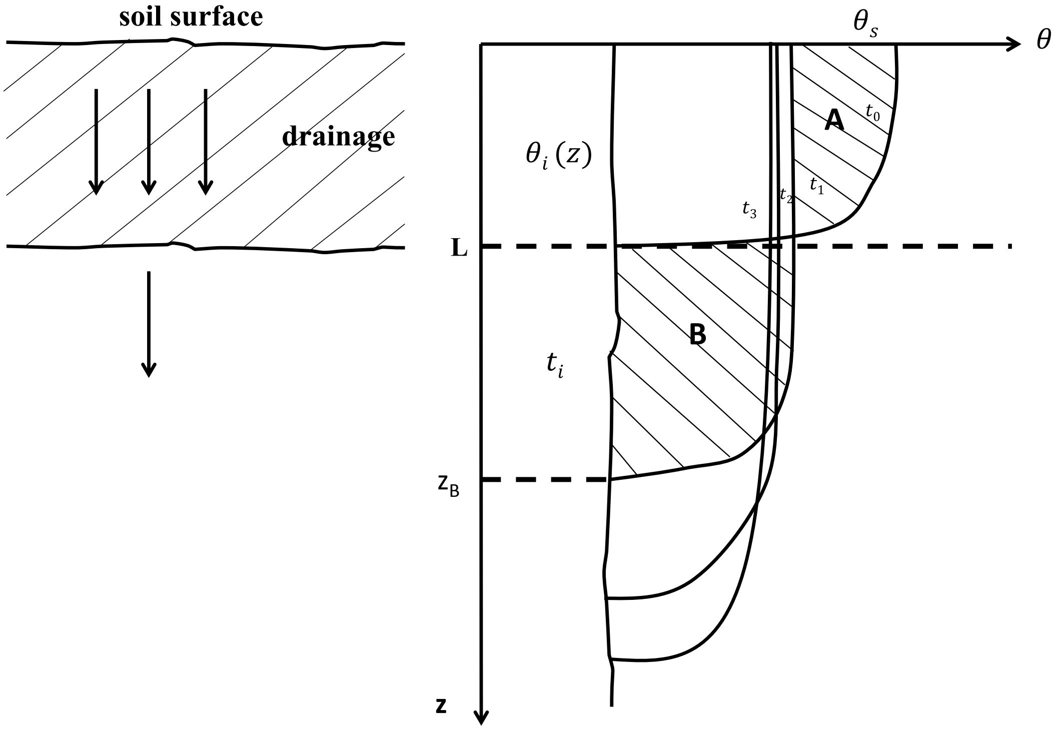


Figure 4. Soil water content profiles during an internal drainage experiment to calculate soil hydraulic conductivity of layer L.

Note that in Fig. 4 the area A represents $\Delta S$ of the soil layer 0 – L m, between times $t_0$ and $t_1$, which is equivalent to area B (also between times $t_0$ and $t_1$) because the soil suface was covered with a plastic sheet so that eventual rainfall or irrigation could not contribute to changes in S, and also no evaporation was possible at soil surface:

$$A = \Delta S_L = S_L(t_0) - S_L(t_1) = \int_0^L \theta(t_0)dz - \int_0^L \theta(t_1)dz$$

$$B = \Delta S_{(z_b - L)} = \int_L^{z_b} \theta(t_1)dz - \int_L^{z_b} \theta(t_i)dz$$

Such Θ measurements are conveniently made with neutron or TDR probes. For the experiment to be discussed below, the following equation was used to estimate S with neutron probe data:

$$S_j(i) = \left[1.5\theta_{i,j}(1) + \theta_{i,j}(2) + \theta_{i,j}(3) + \theta_{i,j}(4) + 0.5\theta_{i,j}(5)\right]\frac{1000}{5} \tag{5}$$

with $\Delta z$ = 0.2m. Soil water contents $\theta_{i,j}(1)$ measured at the depth 0.2m ($k$ = 1) were considered to cover a layer of $1.5\Delta z$ = 0.3m which includes soil surface. The first measurement made at the depth of 0.2 m was evaluated to be deep enough not to lose slow neutrons to the atmosphere. $\theta_{i,j}(5)$ measured at 1.0m ($k$ = 5) covered $0.5\Delta z$ = 0.1m since the lower level of the control volume for water balances was set at 1.0 m, and the total depth $L$ was taken as 1,000mm to obtain data in mm.

In the case of field water balances, the time spell $t_f - t_i$ between two consecutive S measurements is of the order of days, weeks or months, as shown in equation (3). Table 1 shows S data collected in a coffee crop every 14 days, with neutron a neutron probe. They were used to establish WBs with the aim of studying the effects of soil spatial variability on these WBs.

Table 1. Soil water storage $S_L(t_i)$, standard deviations $s(S_L)$, and coefficients of variation (CV) of each period analyzed.

| **Balance** | **Period** | **DAB** | **$S_I$** | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | | | **1** | **2** | **3** | **4** | **5** | $\overline{S_L}$ | **$s(S_L)$** | **CV** |
| 1 | 01/09 to 15/09 | 0_14 | 250.2 | 260.8 | 203.4 | 254.6 | 257.2 | 245.2 | 23.7 | 9.7 |
| 2 | 15/09 to 29/09 | 14_28 | 261.0 | 271.1 | 221.0 | 265.6 | 268.3 | 257.4 | 20.7 | 8.0 |
| 3 | 29/09 to 13/10 | 28_42 | 255.9 | 265.6 | 213.1 | 259.3 | 262.4 | 251.3 | 21.6 | 8.6 |
| 4 | 13/10 to 27/10 | 42_56 | 272.3 | 284.5 | 242.8 | 303.0 | 286.9 | 277.9 | 22.5 | 8.1 |
| 5 | 27/10 to 10/11 | 56_70 | 269.9 | 280.3 | 232.8 | 292.2 | 279.9 | 271.0 | 22.8 | 8.4 |
| 6 | 10/11 to 24/11 | 70_84 | 263.2 | 276.0 | 221.5 | 278.7 | 276.8 | 263.3 | 24.1 | 9.2 |
| 7 | 24/11 to 08/12 | 84_98 | 273.0 | 287.4 | 238.7 | 296.3 | 282.5 | 275.6 | 22.3 | 8.1 |
| 8 | 08/12 to 22/12 | 98_112 | 286.3 | 306.7 | 262.3 | 317.2 | 293.1 | 293.1 | 21.0 | 7.2 |
| 9 | 22/12 to 05/01 | 112_126 | 277.9 | 299.8 | 249.8 | 309.2 | 288.0 | 284.9 | 22.9 | 8.0 |
| 10 | 05/01 to 19/01 | 126_140 | 288.3 | 312.9 | 271.4 | 336.9 | 299.9 | 301.9 | 24.8 | 8.2 |
| 11 | 19/01 to 02/02 | 140_154 | 288.0 | 311.4 | 270.2 | 328.0 | 303.2 | 300.2 | 22.1 | 7.4 |
| 12 | 02/02 to 16/02 | 154_168 | 380.0 | 380.2 | 324.5 | 384.3 | 380.6 | 369.9 | 25.5 | 6.9 |
| 13 | 16/02 to 01/03 | 168_182 | 352.1 | 354.8 | 302.6 | 359.5 | 350.8 | 344.0 | 23.3 | 6.8 |
| 14 | 01/03 to 15/03 | 182_196 | 375.4 | 382.3 | 317.4 | 375.2 | 375.3 | 365.1 | 26.9 | 7.4 |
| 15 | 15/03 to 29/03 | 196_210 | 356.2 | 364.1 | 305.4 | 359.2 | 357.7 | 348.5 | 24.3 | 7.0 |
| 16 | 29/03 to 12/04 | 210_224 | 310.5 | 314.4 | 258.0 | 311.5 | 306.0 | 300.1 | 23.7 | 7.9 |
| 17 | 12/04 to 26/04 | 224_238 | 304.5 | 317.2 | 261.9 | 315.4 | 305.2 | 300.8 | 22.5 | 7.5 |
| 18 | 26/04 to 10/05 | 238_252 | 305.0 | 313.3 | 261.0 | 318.2 | 309.2 | 301.3 | 23.1 | 7.7 |
| 19 | 10/05 to 24/05 | 252_266 | 301.0 | 306.4 | 253.0 | 308.7 | 305.4 | 294.9 | 23.6 | 8.0 |
| 20 | 24/05 to 07/06 | 266_280 | 300.2 | 304.8 | 254.3 | 306.1 | 308.8 | 294.8 | 22.9 | 7.8 |
| 21 | 07/06 to 21/06 | 280_294 | 360.1 | 359.9 | 312.8 | 356.2 | 354.3 | 348.7 | 20.2 | 5.8 |
| 22 | 21/06 to 05/07 | 294_308 | 348.4 | 348.7 | 293.3 | 342.0 | 348.7 | 336.2 | 24.2 | 7.2 |
| 23 | 05/07 to 19/07 | 308_322 | 327.7 | 327.7 | 274.8 | 321.6 | 329.2 | 316.2 | 23.3 | 7.4 |
| 24 | 19/07 to 02/08 | 322_336 | 350.7 | 345.4 | 306.0 | 353.7 | 355.3 | 342.2 | 20.6 | 6.0 |
| 25 | 02/08 to 16/08 | 336_350 | 341.4 | 334.6 | 290.7 | 337.9 | 341.7 | 329.3 | 21.7 | 6.6 |
| 26 | 16/08 to 30/08 | 350_364 | 334.1 | 324.3 | 280.4 | 322.9 | 327.4 | 317.8 | 21.4 | 6.7 |

The daily variation between the five measurements represent the spatial variability of the field, and the 14 day changes represent the time variability due to all WB components, mainly rainfall and evapotranspiration.

## SOIL WATER STORAGE CHANGES MEASURED IN A SOYBEAN CROP IN PIRACICABA, BRAZIL

A soybean (*Glycine max* (L.) Merrill) crop was established on na Oxisol in Piracicaba, Brazil, and for management pouposed the soil water storage S was monitored during the whole cycle. The novelty of the experiment was the continuous measurement of the soil water matric potential h (m) using polimer tensiometers. Readings of h were then transformed into $\Theta$ through the use of a soil water retention curve, to further calculate water storages.

These tensiometers were developed by Bakker et al. (2007) at the University of Wageningen, Holland, with the objective of eliminating the tensiometry problems found with porous cup tensiometers filled with water. These new tensiometers are basically composed by a massive ceramic disc, an innox steel cover containing a polimer of high expansion capacity when absorbing water, and reversibly a high retraction capacity when loosing water. Pressure changes are measured inside the small probe with a pressure transducer, which also has a temperature sensor (Figures 1 and 1a). Sensors are linked to a datalogger that records data every 15 minutes. More details can be found in Bakker et al. (2007). These tensiometers measure h in the interval 0 to -200 m of water (-20 atm or 2 MPa, well below the permanent wilting point (PWP)), while the traditional tensiometers reach only about -0.085 MPa (Durigon et al. ,2011 and Durigon & De Jong Van Lier, 2011).


Figure 1. Polimer tensiometer. Source: Durigon; de Jong Van Lier (2011).

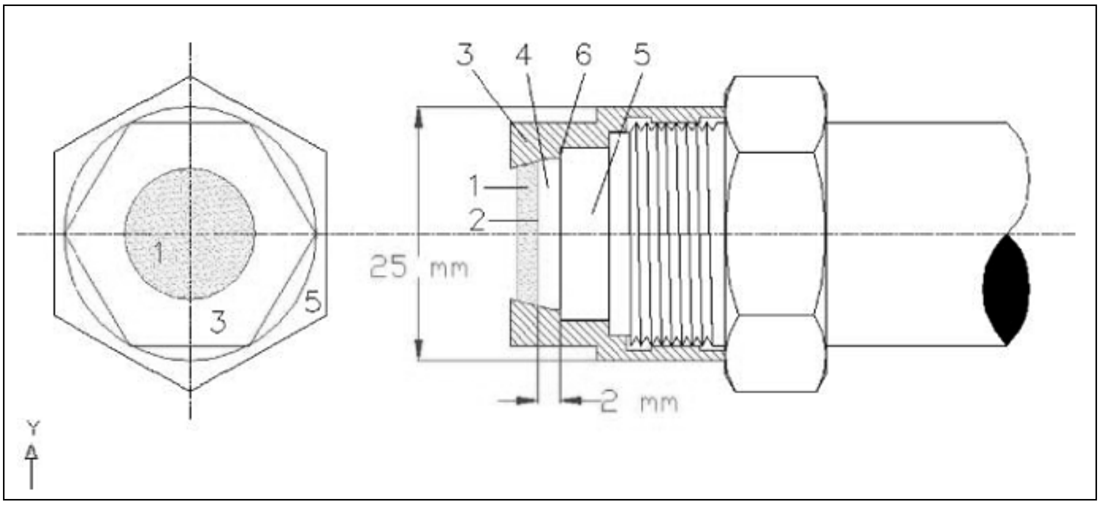


Figure 1a – Details of the polimer tensiometer showing the ceramic disc of α-Al2O3 (1), a membrane of ɤ-Al2O3 (2), inox steel capsule (3), polimer chamber (4), pressure transductor (5), and a synthetic ring (6) . Source: Bakker et al. (2007).

For this soil, MORAES (1991) presents an average soil water retention curve (SWRC), that was obtained from 250 locations of a 3 ha field and adjusted to the van Genuchten (1980) model, with the empirical coefficients $\alpha$, $m$ and $n$, the saturated soil water content $\theta_s$ and the residual water content $\theta_r$.

A view of the initial stage of the soybean crop can be seen in Figure 2, and climatilogic data of the experimental period in Figures 3 and 4,below:


Figure 2. View of the soybean crop at initial growth stage. Piracicaba, 2012.

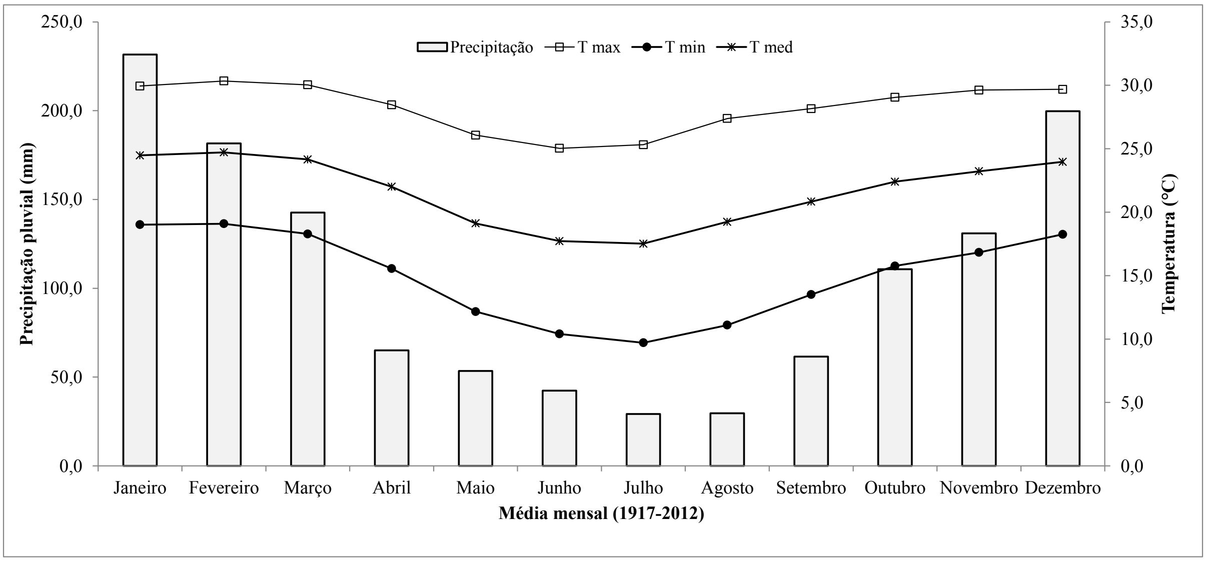


Figure 3. Average values of rainfall P and air temperatures (Tmax, Tmin and Tave) for Piracicaba, covering the period 1917 to 2012.

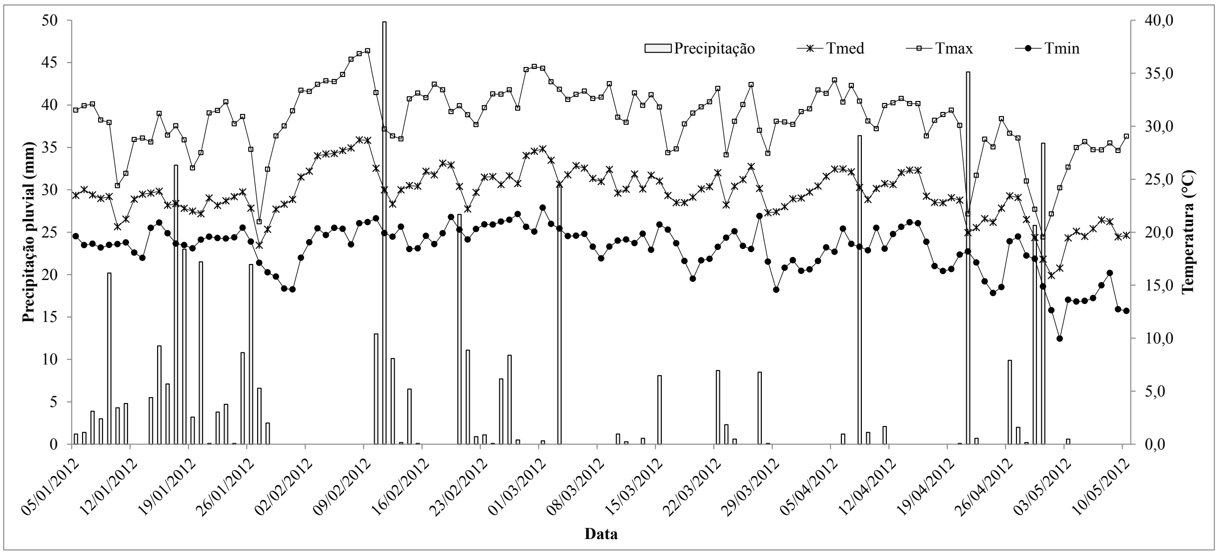


Figure 4. Actual rainfall P anda ir temperature data for the experimental period: 05 January to 10 May, 2012.

Polimer tensiometers were installed in the field on 09 January 2012, at two obsevation points within the experimental area, at three depts: 0.05 m; 0.15 m and 0.3 m to represent the soil profile layers: 0-0.1 m, 0.1-0.2 m and 0.2-0.4 m, respectivelly.

As mentioned above, soil water contents $\theta$ ($m^3$ $m^{-3}$) for each of the three soil layers were calculated by van Genuchten´s (1980) model:

$$\theta = \theta_r + \frac{(\theta_s - \theta_r)}{[1+(\alpha h)^n]^m} \quad (1)$$

With the parameters $\theta_s$ , $\theta_r$ , $\alpha$, $m$ and $n$. Figure 5 shows the experimental points and the model for this particular Oxisol, obtained with the values $\theta s$ and $\theta r$ of 0.468 and 0.296, e of $\alpha$, $m$ and $n$ of 0.024220, 0.330056 and 1.580702, respectively.

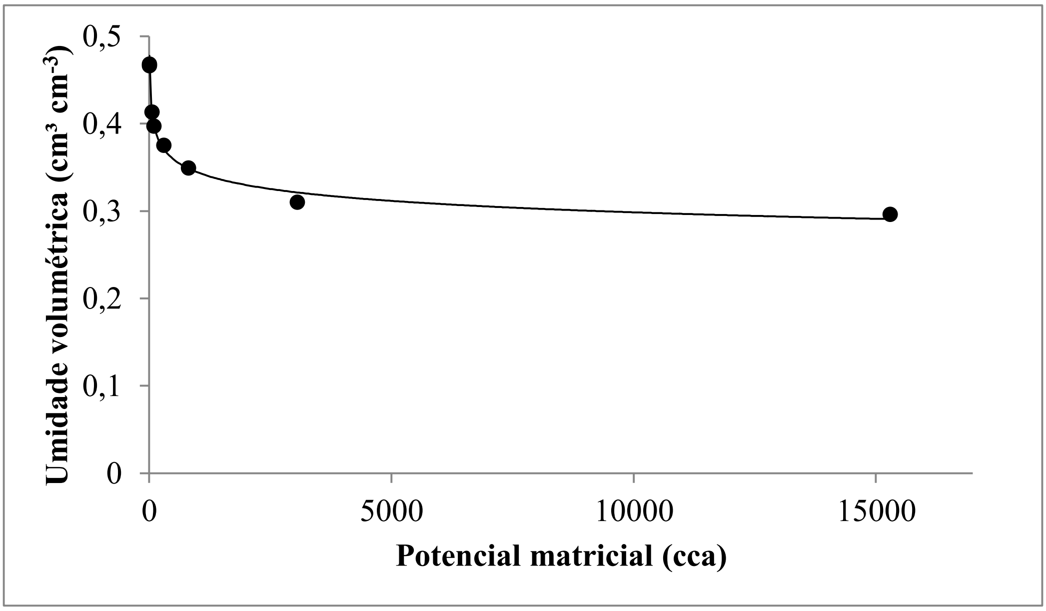


Figure 5 – Average soil water retention curve obtained with 250 points. Source: Moraes (1991).

Another way to look at soil water storage, is using the available water (AW) concept. This concept assumes that the available water to plants lies between a maximum ϴ value, called $\theta_{FC}$ fc, and a minimum $\theta_{PWP}$. The water between $\theta_s$ and $\theta_{FC}$ is considered to be subject to gravitational drainage, not being available to plants. The water below $\theta_{PWP}$ is considered to be at such low matric potentials the plants cannot make use of it. With these concepts, we can define the water holding capacity of a soil (AWC) as $[\theta_{FC} - \theta_{PWP}]$, and soil water storages S can also be defined as:

$$S = \left(\bar{\theta} - \theta_{PWP}\right) * Z_e \tag{2}$$

Where $Z_e$ is the rooting depth, considered in this soybean experiment as 0.4 m or 400 mm. Equation (4) was applied layer per layer to obtain the final value of S.

During the cropping cycle the crop receives water from rainfall P or irrigation I, and every time the soil reaches the AWC, the excess of water is drained below root zone or is lost at the surface as runoff. Within the AWC range, water is eighter evaporated at the soil surface or transpired by plants, resulting the evapotranspiration ET. Water is not equally available in the whole range of the AWC, water extraction becomes more and more difficult as the PWP is reached. This is due to drastic decreases in soil hydraulic conductivity as the soil dries out. There are several models that try to describe the process of water extraction from the soil by plants, and in the previous lecture we mentioned three commonly used models: Thornthwaite & Mather; Rijtema & Aboukhaled; and Dourado & van Lier. For the first, the decrease in S follows the model:

$$S_i = AWC\, e^{-\frac{\theta}{(\theta_{FC}-\theta_{PWP})}} \tag{3}$$

Rijtema and Aboukhaled (1975) take into consideration a water availability factor p for the estimation of S, which decreases as:

$$S_i = (1-p)AWC\ \exp^{\left(\frac{p-\frac{\theta}{(\theta_{FC}-\theta_{PWP})}}{(1-p)}\right)} \tag{4}$$

Dourado and van Lier ( 1993) assume a cossenoidal rate of ET decrease, and S decreases as:

$$S_i = (1-p)AWC\left\{1-\frac{2}{\pi}\, arctg\left[\frac{\pi}{2}\left(\frac{\left(\frac{\theta}{(\theta_{FC}-\theta_{PWP})}\right)-p}{1-p}\right)\right]\right\} \tag{5}$$

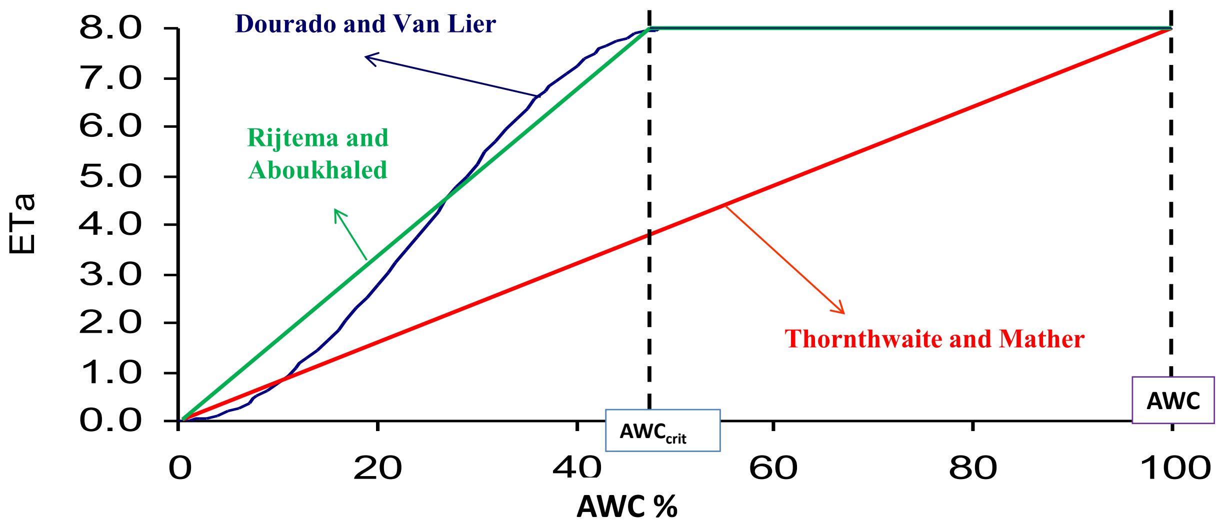


Figure 6. Rate of soil water loss (ETa, mm/period of time) as a function of storage (AWC, mm) for the methods of Thornthwaite and Mather, Rijtema and Aboukhaled and Dourado and Van Lier.

Crop yield is severely reduced by water shortage. Therefore, the concept of water depleated yield $Yr$ was defined by Doorenbos e Kassam (1994):

$$Yr = \left[1-ky\left(1-\frac{ETr}{ETC}\right)\right] * Yo \tag{6}$$

$ky$ being a crop water stress sensitivity factor, that changes as the crop develops; $ETr$ the actual evapotranspiration; and $ETC$ the maximum evapotranspiration of the crop, and:

$$Yo = Fb * C_{IAF} * C_{RESP} * C_{COL} * C_{U} * NDC \tag{7}$$

$Fb$ is the Gross photosynthesis; $C_{IAF}$ a correction factor related to growth phase and leaf area; $C_{RESP}$ a correction factor related to plant respiration; $C_{COL}$ a correction for the harvest índex; $C_{U}$ a correction the water contento f the harvested matter, and $NDC$ the length of the growth period, all proposed by Pereira; Angelocci; Sentelhas (2002).

The following Figures present the experimental data related to the above described considerations.