WATER BALANCE AND CLIMATE

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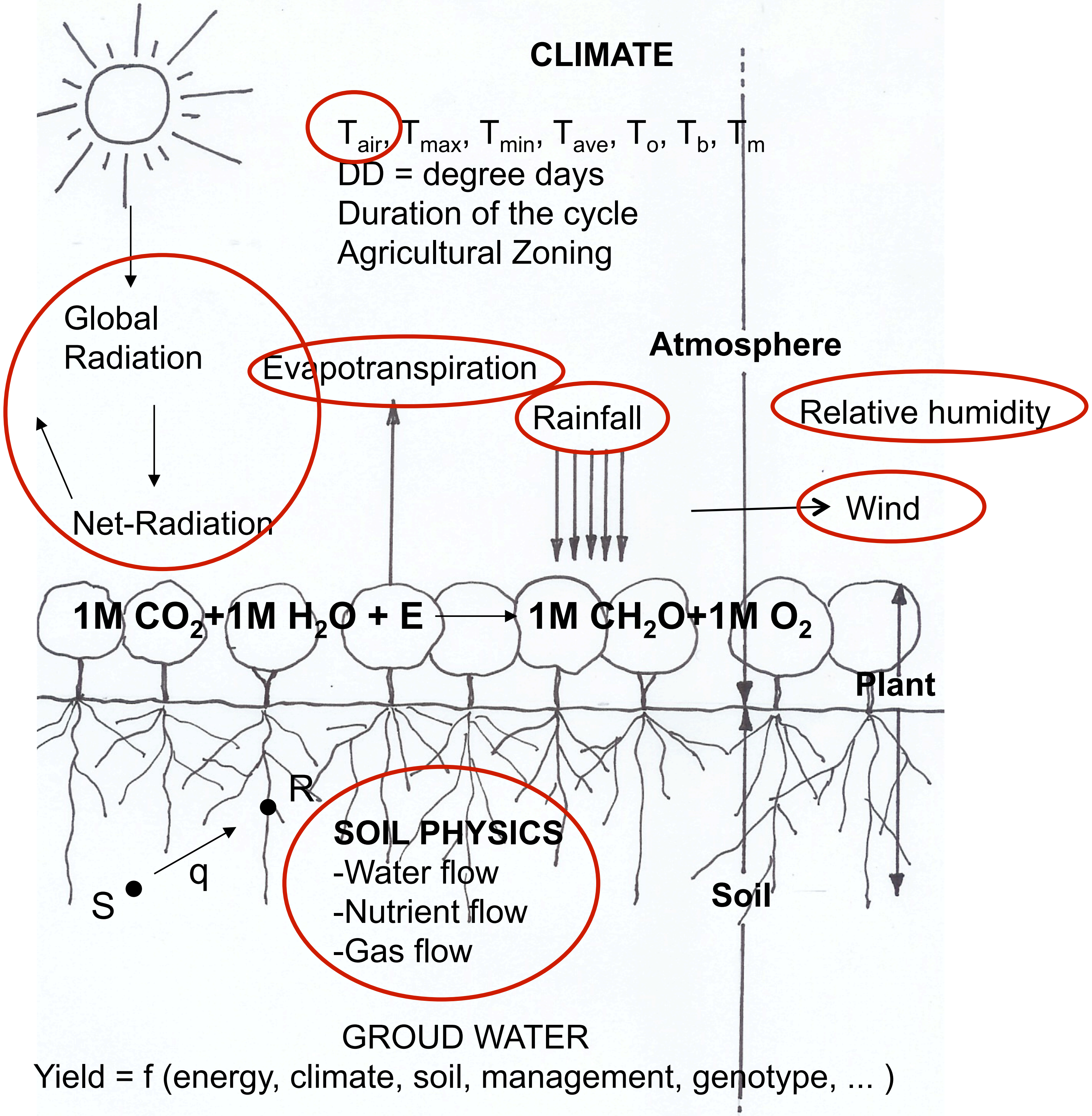
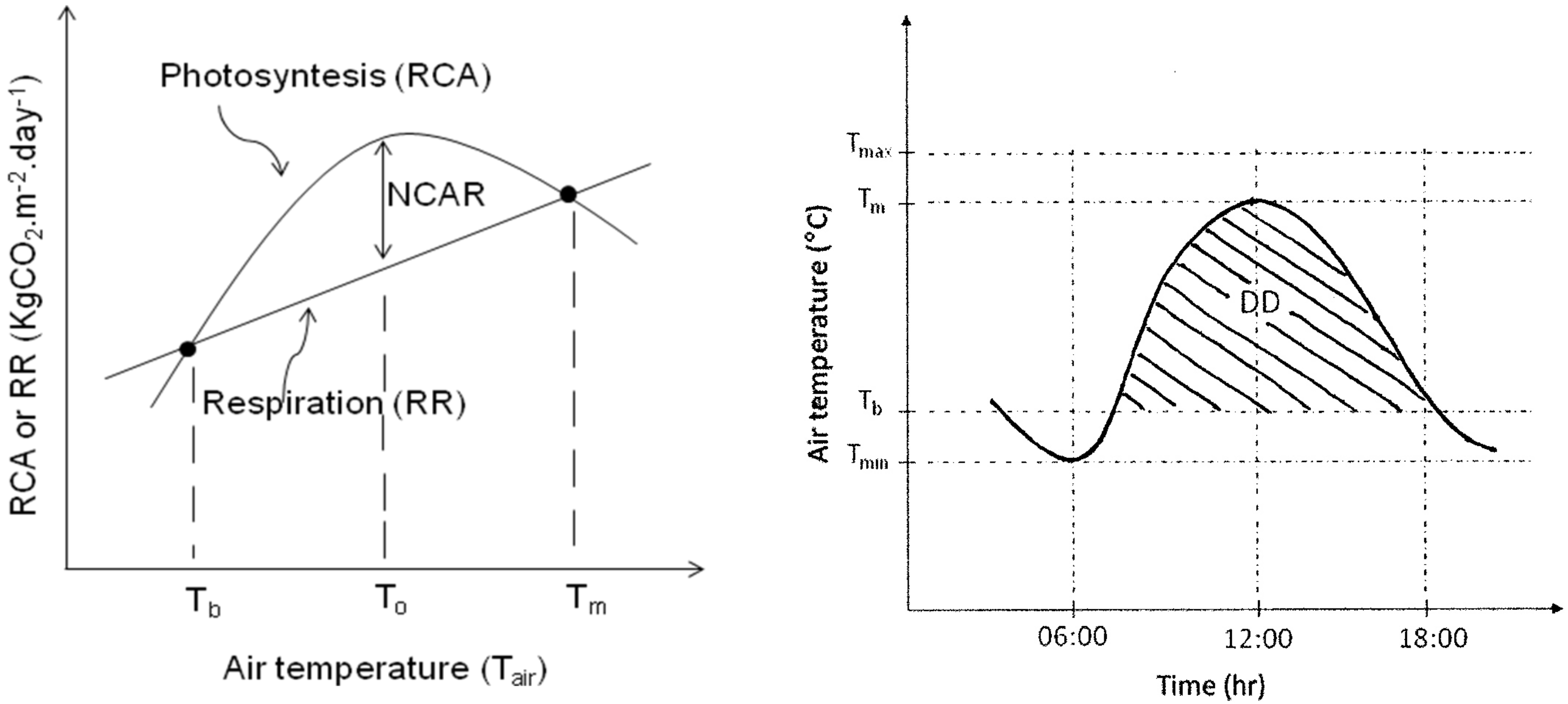
Water cycling in a watershed or in a cropped field can be characterized and quantified by a **water balance, which is the computation of all water fluxes at the boundaries of the system** under consideration. It is an itemized statement of all **gains, losses and changes of water storage** within a specified **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.


CLIMATE
T air, T max, T min, T ave, T o, T b, T m
DD = degree days
Duration of the cycle
Agricultural Zoning
Global Radiation
Net-Radiation
Evapotranspiration
Rainfall
Atmosphere
Relative humidity
Wind
1M CO2+1M H2O + E → 1M CH2O+1M O2
Plant
R
q
S
SOIL PHYSICS
-Water flow
-Nutrient flow
-Gas flow
Soil
GROUD WATER
Yield = f (energy, climate, soil, management, genotype, ...)



A B

Figure 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.

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.

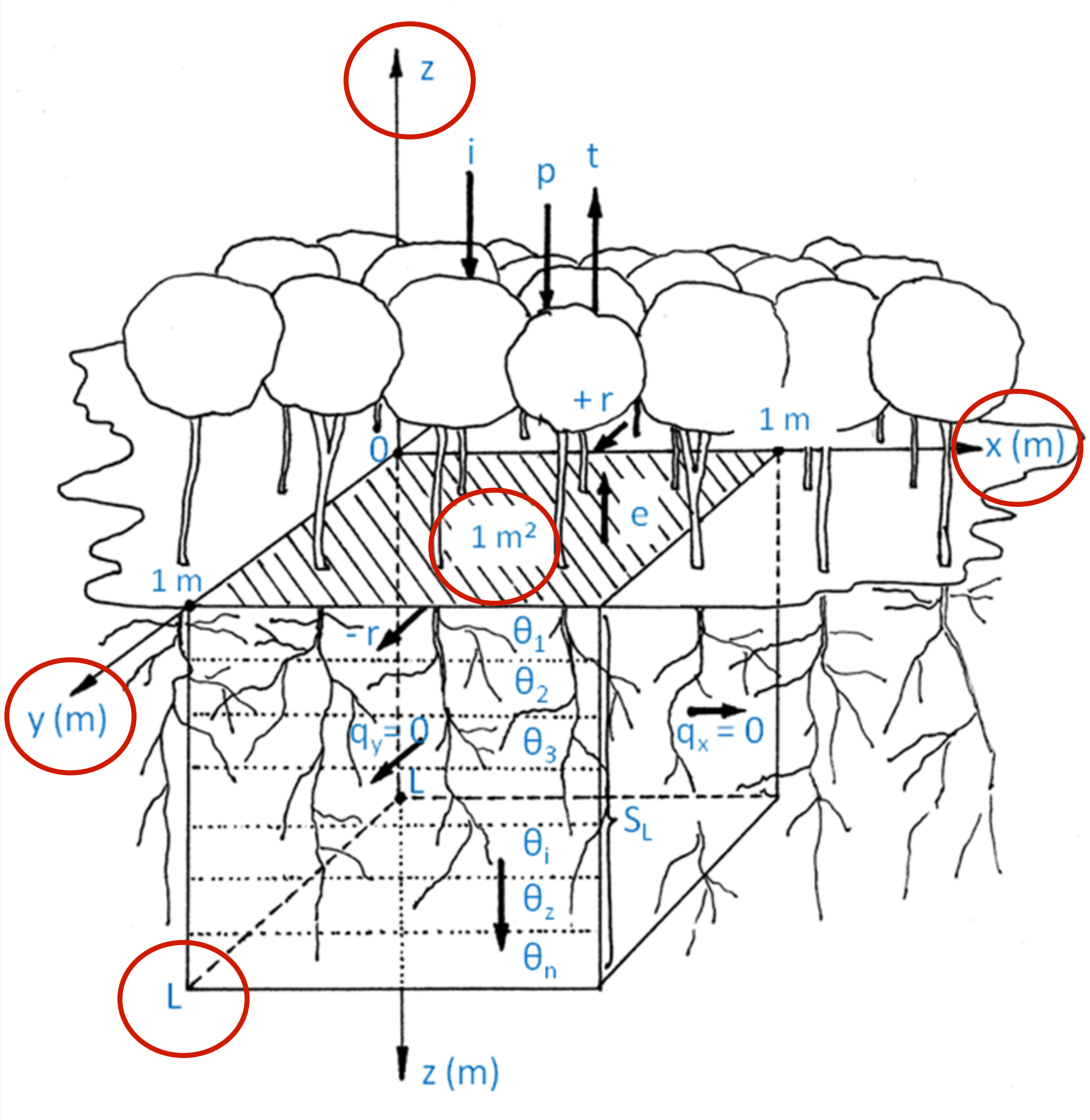


Figure 3. Schematic view of the volume element and of the fluxes that compose the water balance.

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** (**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).

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.

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

where S is defined by

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

which by the trapezoidal rule becomes

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

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.

FIVE SHORT EXAMPLES

When all but one of the above components are known, the unknown is easily calculated algebraically.

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

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.

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 gages to be used.

DAILY RAINFALL VARIABILITY AT A LOCAL SCALE (1,000 ha), IN PIRACICABA, SP, BRAZIL, AND ITS IMPLICATIONS ON SOIL WATER RECHARGE

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DIPLOMA
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IN RECOGNITION TO THE CONTRIBUTION IN RESEARCH
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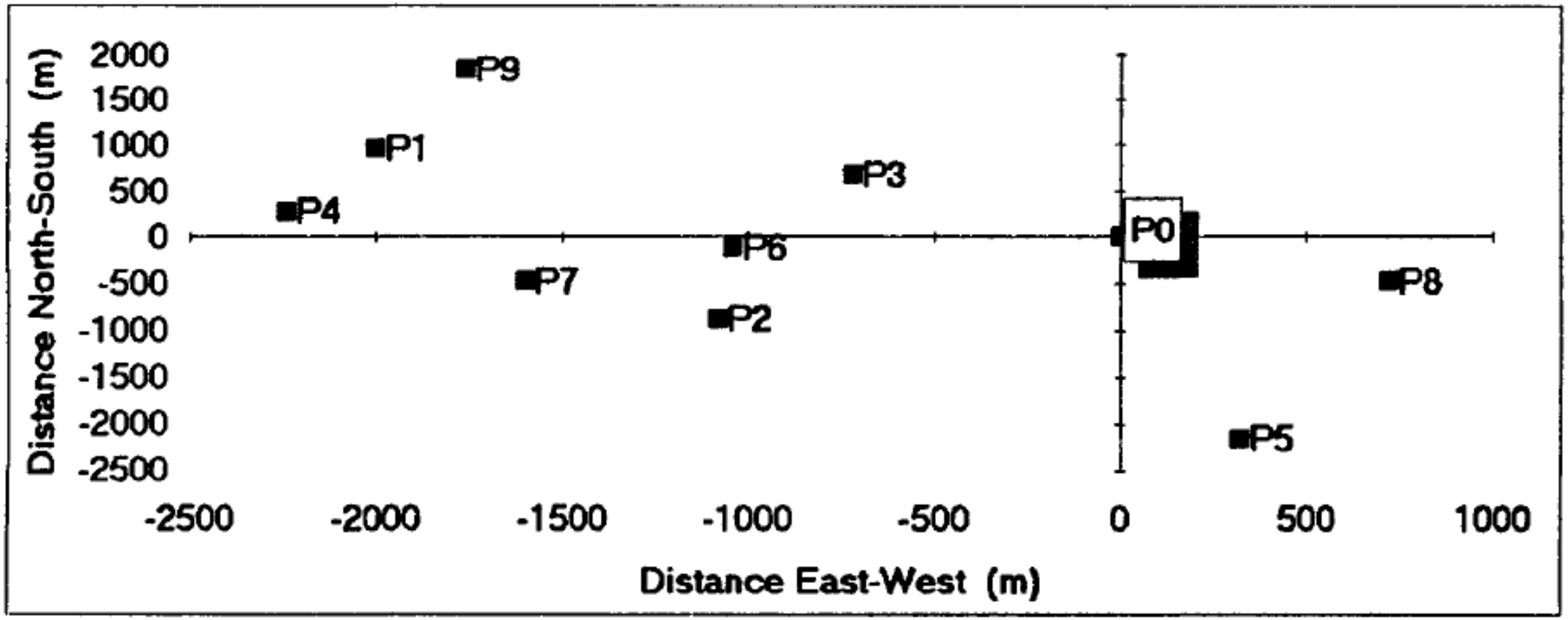

Figure 1 - Location of the obsevation points: P1=Pivot at Fazenda Areão; P2 = Museum "Luiz de Queiroz"; P3=Faculty Club Restaurant; P4= Center for Nuclear Energy in Agriculture; P5=SINFESALQ Recreation Area; P6=Crop Production Experimental Area; P7=Center for Biotechnology; P8=Pasture; P9=Piracicaba Astronomic Observatory; P0=ESALQ Meteorological Station (standard).

Date	1	2	3	4	5	6	7	8	9	0	av.	sd	CV
Jan. 3	3	3,6	3,4	3,7	3,5	3,3	3,3	3,2	3	2,9	3,29	0,27	8,16
Jan. 13	48,6	51,3		49,4	49,3	49,9	50,6	48,5	57	51,9	50,72	2,62	5,17
Jan. 18	5	3,7	3,9	4,1	3,9	3,7	3,3	3,8	5,7	4	4,11	0,71	17,22
Jan. 19	2	1,7	1,8	3	0,2	2,1	4	0,6	1,3	0,3	1,70	1,19	70,21
Jan. 22	3,3	6	1,8	5	3,1	4,3	6,3	1	12,7	2,3	4,58	3,35	73,05
Jan. 25	11,8	8,6	9,5	10,7	9,8	8,8	9,6	10	10,3	8,4	9,75	1,03	10,60
Jan. 27	45,5	48,9	51	46,7	63,7	49,3	50,1	56,3	51	54,4	51,69	5,30	10,26
Jan. 28	2,7	4	4,8	1,5	3,7	5,6	4,2	7	3,1	7,5	4,41	1,88	42,52
Jan. 31	0	0	0,7	0	0,2	0,3	0	2,1	0	1,6	0,49	0,76	154,96

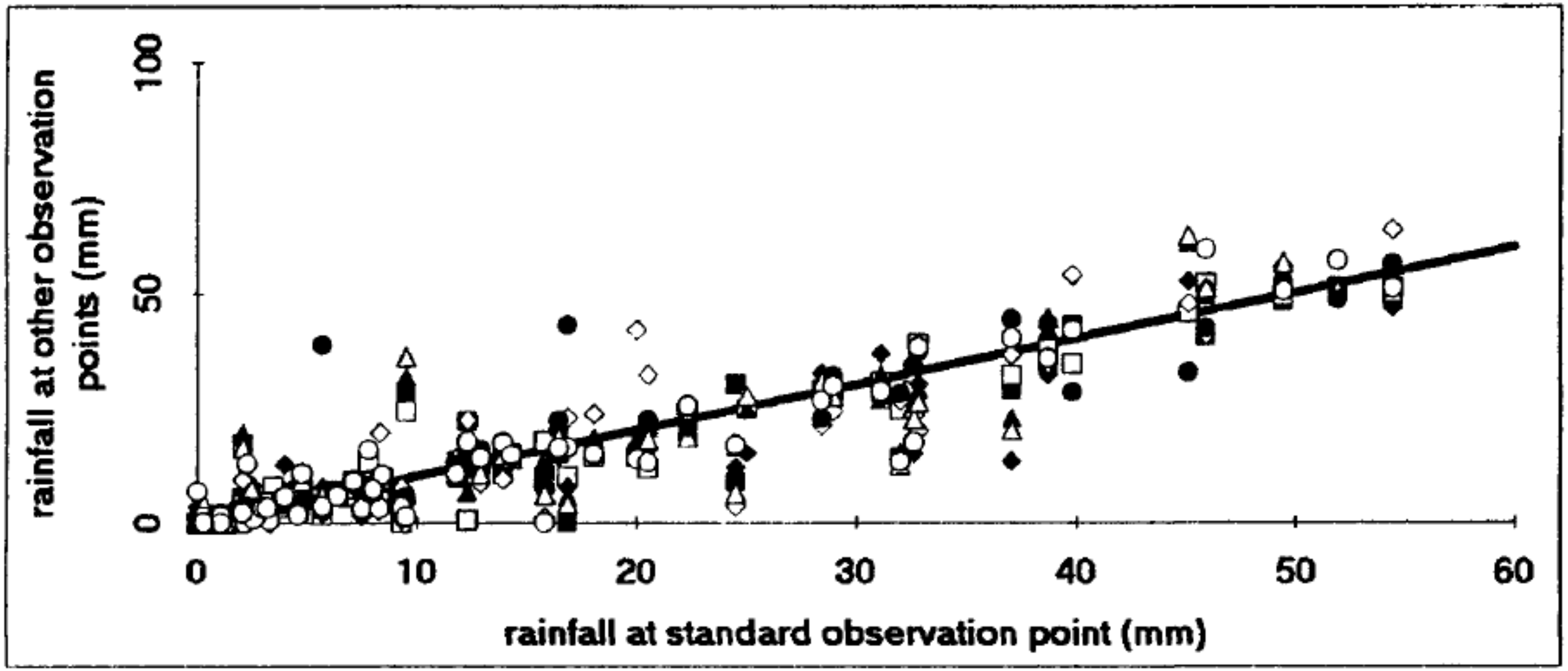

Figure 2 - Regression of rainfall data at each observation point as a function of standard point data

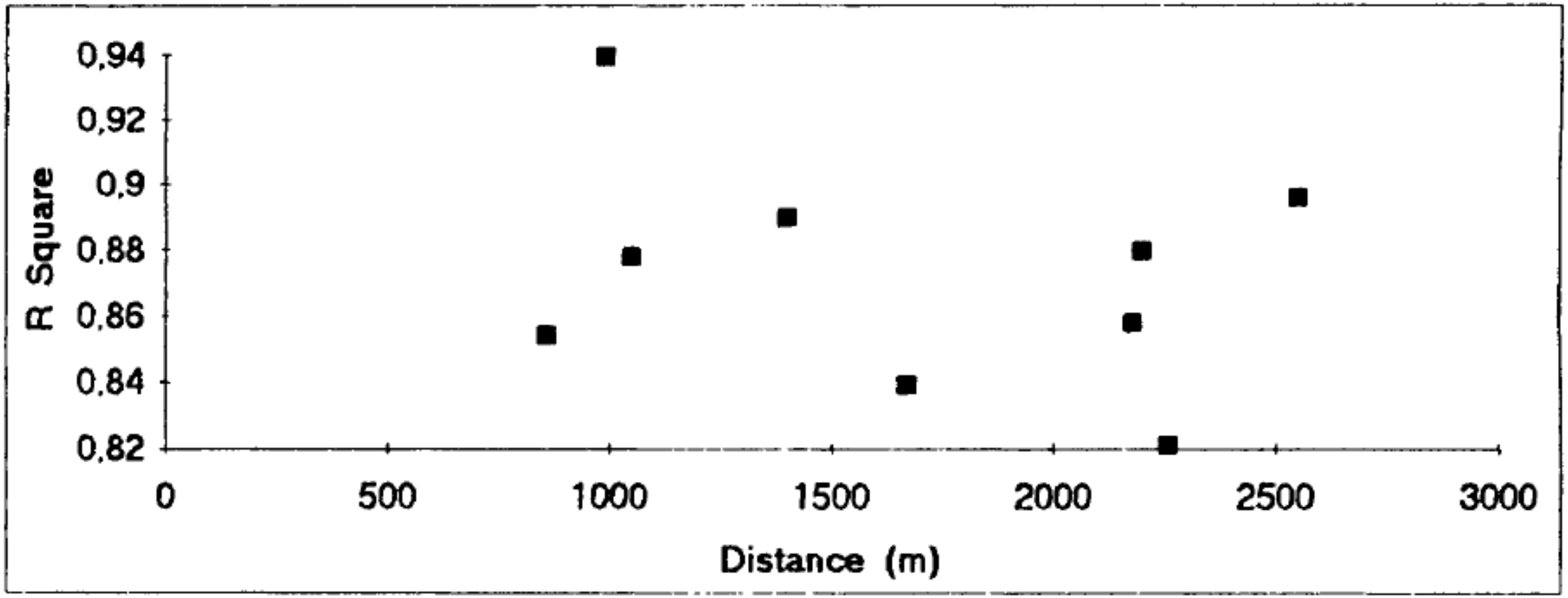

Figure 3 - Correlation between data of pairs of observation points with distances

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.

Evapotranspiration (ET)

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

Reference Evapotranspiration ET_0, 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 potencial 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 genral, 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_n for any crop when ET_o is known:

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

Actual or Real Evapotranspiration ET or ET_a , is the evapotranspiration that occurs under any soil-plant-atmosphere condition. Is the one present in the equation (3a) of the water balance. When soil water is readily available $ET_a = ET_m$, the soil is close to **Field Capacity FC** 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 the deep drainege 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.

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, eighter 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.

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** $q_L = K(\theta)[\partial H / \partial z]$, integrated over the time:

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

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:

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

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 θ_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 **θ = 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 equqtion (3a), Q_L could be left as an unknown in the equation.

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.

VARIABILITY OF WATER BALANCE COMPONENTS IN A COFFEE CROP GROWN IN BRAZIL

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ABSTRACT

The establishment of field water balances is difficult and costly, the variability of its components being the major problem to obtain reliable results. This component variability is here presented for a coffee crop grown in the Southern Hemisphere, on a tropical soil with 10% slope. It is concluded that rainfall has to the measured with an appropriate number of replicates, that irrigation can introduce great variability into calculations, that evapotranspiration calculated from the water balance equation has too high coefficients of variation, that the soil water storage component is the major contributor in error propagation calculations, and that the run-off could be satisfactorily controlled on the 10% slope through crop management practices.

Keywords: water balances; component variability; rainfall; evapotranspiration; soil water storage

1. Experimental Field

The experiment was carried out in Piracicaba, SP, Brazil, (22^{o0}42`S, 47$^{o}$38`W, 580m above sea level) on a soil classified as Rhodic Kandiudalf, locally called “Nitossolo Vermelho Eutroférrico”, A moderate and clayey texture. The climate is Cwa, according to Köppen’s classification, mesothermic with a dry winter, in which the average temperature during the coldest month is below 18^{o}C and during the hottest month, is over 22^{o}C. The annual average temperatures, rainfall, and relative humidity are 21.1^{o}C, 1,257 mm, and 74%, respectively. The dry season is between April and September; July is the driest month along the year. The wettest period is between January and February. The amount of rainfall during the driest month is not over 30 mm (Villa Nova, 1989).

2. Water Balance

Water balances started on September 1, 2003 (DAB-0) and continued to be established for 14 day periods , continually, until August 30, 2004 (DAB-364), completing one year. The classical water balance equation representing the mass conservation law was used, considering water fluxes entering and leaving a soil volume element, integrated over time for 14 day periods, :

$$\int_{t_i}^{t_{i+14}} p\,dt + \int_{t_i}^{t_{i+14}} i\,dt - \int_{t_i}^{t_{i+14}} e\,dt - \int_{t_i}^{t_{i+14}} r\,dt \pm \int_{t_i}^{t_{i+14}} q_L\,dt + S_{i+14} - S_i = 0 \qquad (1)$$

which by solving the integrals results in:

$$P + I - ER - RO - Q_L + \Delta S = 0 \qquad (2)$$

where P=rainfall; I=irrigation; ER=actual evapotranspiration; $\Delta S = S_{i+14} – S_i$ = soil water storage changes in the soil 0–L layer; RO = runoff; and Q_L = deep drainage at the lower boundary of the soil volume at the depth z = L, all expressed in mm.

Rainfall (P) was measured daily and integrated over Δt at each replicate, using traditional rain-gauges ("Ville de Paris") with 0.04047 m^2 collecting areas, installed in the sub-plots 1.2 m above soil surface. Due to the presence of obstacles in the neighborhood of the experimental area, such as, a silo, a warehouse, orchards, and tall trees, the rainfall was measured in each T_2 plot using 5 rain-gauges, opening the possibility of obtaining average values ( ) with standard deviations [s(P)] and coefficients of variation (CV).

Irrigation for coffee in this region of Brazil is supplementary, applied only during periods of severe drought, in our case through the central-pivot system. As mentioned above, the coffee crop plots were at the edge of this irrigation system, which increased the variability of water application. This variable was also measured by the 5 rain-gauges installed for rainfall measurement.

The criteria of amount and time of irrigation were mostly based on physiological aspects of the coffee plant that requires a cold and dry winter to blossom, which starts after the first significant rain. After blossoming, an excessive lack of water may cause flower loss. Therefore, the decision to irrigate was taken by visual observation of the water deficit, trying to apply 30 mm of water depth that approximately would wet a 0.6 m soil layer.

The actual crop evapotranspiration (ER) was estimated by difference from all other components, using equation (2). In wet periods, with a drainage (Q_L) likely to happen and considering it as zero in equation (02), ER, now named ER', was overestimated because it includes Q_L. Thus, in periods in which ER was larger than the potential evapotranspiration (ET), ER was considered equal to ET and the difference ER–ET=Q_L. The potential evapotranspiration was estimated from the reference evapotranspiration (ET_0) corrected by the crop coefficient (K_C). ET_0 was calculated using Penman-Monteith equation (Pereira et al., 1997), with meteorological data collected at the automatic weather-station installed near the experimental area. K_C was calculated by dividing ER by ET_0 along the periods in which the plants were not under stress, when the soil water storage was relatively high and without drainage. The above referred K_C was the average value obtained for these periods.

The soil layer 0-1m (L=1m) was chosen to calculate soil water storages since at this stage of the crop this soil layer contains more than 95% of the root system. was estimated from soil water content measurements ( ) obtained by a neutron probe, using three access tubes installed down to the depth of 1.2 m in each plot, making up a total of 15 tubes. The calibration of this probe, model CPN 503 DR, was made in an area close to the experimental field. The moisture contents were measured at 0.20, 0.40, 0.60, 0.80, and 1.00 m at the selected dates t_i, during the experimental period, which started at t_i (DAI-0) and continued up to t_{i+14}, Δt = 14 days. was calculated using the trapezoidal rule:

$$S(t_i) = \int_0^L \theta(t_i)dz = [\overline{\theta}(t_i)].L \qquad (4)$$

where is the average at time and the soil depth L, in this case taken as 1,000 mm in order to obtain S expressed in mm.

For measuring the runoff, each experimental plot was framed by metal dicks, and the water was collected by gravity in 60L tanks placed downslope.

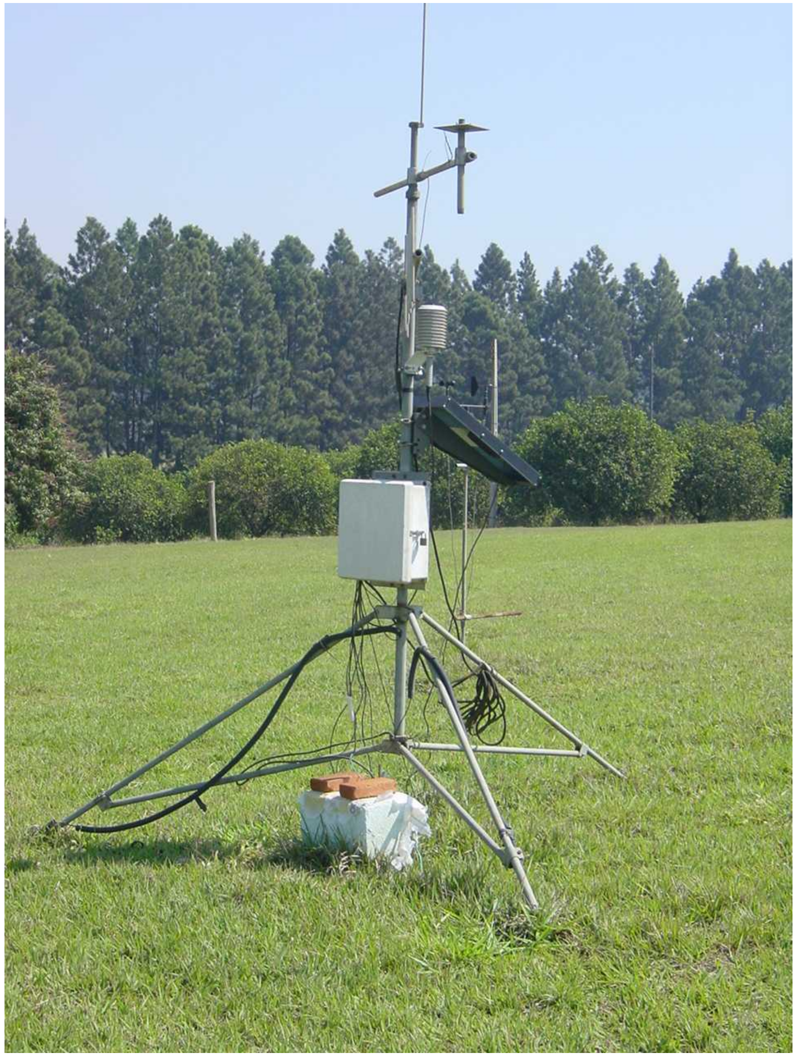



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Table 1: Average rainfall (P), standard deviations [s(P)], and coefficients of variation (CV
of each period.

Balance	Period	DAB	Rainfall (P)							
			1	2	3	4	5	$\overline{P}$	s(P)	CV
1	01/09 to 15/09	0_14	4.0	4.2	4.3	4.2	4.0	4.1	0.1	3.2
2	15/09 to 29/09	14_28	5.8	5.8	6.4	4.8	6.2	5.8	0.6	10.6
3	29/09 to 13/10	28_42	79.0	75.4	80.6	78.0	75.9	77.8	2.2	2.8
4	13/10 to 27/10	42_56	18.2	18.1	18.2	17.6	17.5	17.9	0.3	1.9
5	27/10 to 10/11	56_70	25.4	24.9	26.3	24.5	25.5	25.3	0.7	2.7
6	10/11 to 24/11	70_84	75.7	74.2	78.7	74.2	72.5	75.1	2.3	3.1
7	24/11 to 08/12	84_98	93.9	88.9	91.8	87.4	86.7	89.7	3.0	3.4
8	08/12 to 22/12	98_112	51.0	49.8	49.3	48.5	48.0	49.3	1.2	2.4
9	22/12 to 05/01	112_126	89.2	86.5	85.1	84.4	82.8	85.6	2.4	2.8
10	05/01 to 19/01	126_140	52.4	51.1	50.5	49.6	49.3	50.6	1.2	2.5
11	19/01 to 02/02	140_154	173.7	168.4	165.7	166.7	164.2	167.7	3.7	2.2
12	02/02 to 16/02	154_168	73.9	71.4	69.1	67.9	66.9	69.8	2.8	4.0
13	16/02 to 01/03	168_182	156.6	156.3	153.7	149.2	148.8	152.9	3.7	2.5
14	01/03 to 15/03	182_196	75.9	74.8	72.2	71.4	71.2	73.1	2.1	2.9
15	15/03 to 29/03	196_210	14.4	14.4	14.0	13.8	13.2	14.0	0.5	3.6
16	29/03 to 12/04	210_224	59.4	78.6	62.2	65.0	61.0	65.2	7.7	11.9
17	12/04 to 26/04	224_238	54.7	53.6	51.8	50.9	50.7	52.3	1.7	3.3
18	26/04 to 10/05	238_252	23.9	24.1	22.9	22.3	22.7	23.2	0.8	3.4
19	10/05 to 24/05	252_266	27.4	27.2	25.1	23.9	24.1	25.5	1.7	6.5
20	24/05 to 07/06	266_280	105.5	104.5	101.1	98.5	97.7	101.5	3.5	3.4
21	07/06 to 21/06	280_294	7.6	8.0	7.1	6.7	6.5	7.2	0.6	8.7
22	21/06 to 05/07	294_308	2.4	2.0	1.8	1.6	1.6	1.9	0.3	17.8
23	05/07 to 19/07	308_322	33.2	33.1	32.5	32.2	32.3	32.7	0.5	1.4
24	19/07 to 02/08	322_336	46.8	45.4	43.9	43.6	43.1	44.6	1.5	3.4
25	02/08 to 16/08	336_350	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
26	16/08 to 30/08	350_364	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Sum	01/09 to 30/08	0-364	1350.0	1340.7	1314.3	1286.9	1272.4	1312.9	33.4	2.5

DAB = Days After Begining

Table 2: Average irrigation $(\bar{I})$, standard deviations s(I), and coefficients of variation (CV) for two periods.

Balance	Period	DAB	Irrigation		
			$\bar{I}$	s(I)	CV
1	01/09 to 15/09	0_14	34.2	12.0	35.1
26	16/08 to 30/08	350_364	37.5	15.6	41.7

Table 3: Average actual evapotranspiration (ER'), its standard deviation [s(ER') calculated through equation 03], reference evapotranspiration (ET_0), crop coefficient (K_C), potential evapotranspiration (ET_C), ER and the drainage below root zone (Q_L) for each period.

Balance	DAB	$\overline{ER'}$ (mm)	s(ER')	CV	$\overline{ET_0}$ (mm)	K_C	$\overline{ET_C}$ (mm)	ER=ER'-Q_L (mm)	$\overline{Q_L}$ (mm)
1	0_14	-26.1	33,65	129,0	-45.9	1.1	-50.1	-26.1	0,0
2	14_28	-11.9	29,92	250,9	-56.0	1.1	-61.2	-11.9	0,0
3	28_42	-50.9	31,25	61,3	-53.9	1.1	-58.9	-50.9	0,0
4	42_56	-24.8	32,00	129,1	-65.4	1.1	-71.5	-24.8	0,0
5	56_70	-33.1	33,19	100,3	-47.5	0.7	-33.1	-33.1	0,0
6	70_84	-62.3	32,90	52,8	-60.3	1.0	-62.3	-62.3	0,0
7	84_98	-72.0	30,74	42,7	-50.5	1.4	-72.0	-72.0	0,0
8	98_112	-57.5	31,10	54,1	-62.4	0.9	-57.5	-57.5	0,0
9	112_126	-68.1	33,87	49,7	-57.5	1.2	-68.1	-68.1	0,0
10	126_140	-52.2	33,28	63,7	-63.2	0.8	-52.2	-52.2	0,0
11	140_154	-97.4	33,95	34,9	-39.3	1.1	-42.9	-42.9	-54,4
12	154_168	-95.5	34,66	36,3	-62.0	1.5	-95.5	-95.5	0,0
13	168_182	-130.6	35,80	27,4	-46.8	1.1	-51.2	-51.2	-79,4
14	182_196	-89.3	36,28	40,6	-52.3	1.7	-89.3	-89.3	0,0
15	196_210	-62.4	33,95	54,4	-55.3	1.1	-62.4	-62.4	0,0
16	210_224	-64.2	33,61	52,4	-47.7	1.3	-64.2	-64.2	0,0
17	224_238	-51.7	32,31	62,5	-36.1	1.4	-51.7	-51.7	0,0
18	238_252	-29.6	33,03	111,6	-35.6	0.8	-29.6	-29.6	0,0
19	252_266	-25.6	32,92	128,8	-24.4	1.0	-25.6	-25.6	0,0
20	266_280	-46.8	30,75	65,6	-23.4	1.1	-25.6	-25.6	-21,3
21	280_294	-19.6	31,51	160,5	-29.9	0.7	-19.6	-19.6	0,0
22	294_308	-21.9	33,59	153,2	-35.4	0.6	-21.9	-21.9	0,0
23	308_322	-6.6	31,13	469,1	-27.7	1.1	-30.2	-6.6	0,0
24	322_336	-57.5	30,00	52,2	-35.7	1.1	-39.0	-39.0	-18,5
25	336_350	-11.4	30,48	266,5	-45.1	1.1	-49.3	-11.4	0,0
26	350_364	-46.1	30,26	65,7	-46.7	1.1	-51.0	-46.1	0,0
1_26	0_364	-1315,3	-	-	-1206,0	1.1	-1318,3	-1141,7	-173,6

Table 5: 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

Table 6: Runoff (RO), standard deviations (SD), and coefficients of variation (CV) from each period.

Balance	Period	DAB	RO (mm)					$\overline{RO}$ (mm)	s(RO)	CV
			1	2	3	4	5			
1	01/09 to 15/09	0_14	-	-	-	-	-	-	-	-
2	15/09 to 29/09	14_28	-	-	-	-	-	-	-	-
3	29/09 to 13/10	28_42	0.0	0.0	0.4	0.6	0.1	0.2	0.3	118.9
4	13/10 to 27/10	42_56	0.0	0.0	0.0	0.0	0.1	0.0	0.0	223.6
5	27/10 to 10/11	56_70	-	-	-	-	-	-	-	-
6	10/11 to 24/11	70_84	0.0	0.0	0.6	0.8	0.5	0.4	0.4	94.3
7	24/11 to 08/12	84_98	0.0	0.0	0.2	0.0	0.7	0.2	0.3	173.2
8	08/12 to 22/12	98_112	-	-	-	-	-	-	-	-
9	22/12 to 05/01	112_126	0.0	0.0	0.7	1.8	0.1	0.5	0.8	149.6
10	05/01 to 19/01	126_140	0.0	0.0	0.1	0.3	0.0	0.1	0.1	138.3
11	19/01 to 02/02	140_154	0.0	0.0	1.4	1.4	0.1	0.6	0.7	125.6
12	02/02 to 16/02	154_168	0.0	0.0	0.4	0.9	0.0	0.3	0.4	152.7
13	16/02 to 01/03	168_182	3.0	0.5	1.1	1.3	0.0	1.2	1.1	96.0
14	01/03 to 15/03	182_196	0.0	0.0	0.6	1.5	0.0	0.4	0.7	155.1
15	15/03 to 29/03	196_210	-	-	-	-	-	-	-	-
16	29/03 to 12/04	210_224	0.6	0.2	0.6	0.0	0.0	0.3	0.3	110.6
17	12/04 to 26/04	224_238	0.0	0.0	0.1	0.3	0.0	0.1	0.1	158.8
18	26/04 to 10/05	238_252	0.0	0.0	0.1	0.2	0.0	0.1	0.1	142.6
19	10/05 to 24/05	252_266	0.0	0.0	0.0	0.0	0.0	0.0	0.0	223.6
20	24/05 to 07/06	266_280	0.0	0.0	2.2	1.9	0.0	0.8	1.1	136.6
21	07/06 to 21/06	280_294	-	-	-	-	-	-	-	-
22	21/06 to 05/07	294_308	-	-	-	-	-	-	-	-
23	05/07 to 19/07	308_322	0.0	0.0	0.2	0.1	0.0	0.1	0.1	127.3
24	19/07 to 02/08	322_336	0.0	0.0	0.0	0.0	0.0	0.0	0.0	156.5
25	02/08 to 16/08	336_350	-	-	-	-	-	-	-	-
26	16/08 to 30/08	350_364	1.5	0.2	0.0	0.0	0.0	0.3	0.7	189.9
Sum	01/09 to 30/08	0_364	5,1	1,0	8,7	11,1	1,6	5,5	4,4	80,3

Table 7: Average values of rainfall ($\overline{P}$), irrigation ($\overline{I}$), soil water storage changes ($\overline{\Delta S}$), runoff ($\overline{RO}$), drainage ($\overline{Q_L}$), actual evapotranspiration ($\overline{ER}$), and potential evapotranspiration ($\overline{ET_C}$), for all analyzed periods.

Balance	Period	DAB	$\overline{P}$ (mm)	$\overline{I}$ (mm)	$\overline{S_i}$ (mm)	$\overline{\Delta S}$ (mm)	$\overline{RO}$ (mm)	$\overline{Q_L}$ (mm)	$\overline{ER}$ (mm)	$\overline{ET_C}$ (mm)
1	01/09 to 15/09	0_14	4.1	34.2	245.2	12.2	0.0	0.0	-26.1	-50.1
2	15/09 to 29/09	14_28	5.8	0.0	257.4	-6.1	0.0	0.0	-11.9	-61.2
3	29/09 to 13/10	28_42	77.8	0.0	251.3	26.6	-0.2	0.0	-50.9	-58.9
4	13/10 to 27/10	42_56	17.9	0.0	277.9	-6.9	0.0	0.0	-24.8	-71.5
5	27/10 to 10/11	56_70	25.3	0.0	271.0	-7.8	0.0	0.0	-33.1	-33.1
6	10/11 to 24/11	70_84	75.1	0.0	263.3	12.3	-0.4	0.0	-62.3	-62.3
7	24/11 to 08/12	84_98	89.7	0.0	275.6	17.5	-0.2	0.0	-72.0	-72.0
8	08/12 to 22/12	98_112	49.3	0.0	293.1	-8.2	0.0	0.0	-57.5	-57.5
9	22/12 to 05/01	112_126	85.6	0.0	284.9	17.0	-0.5	0.0	-68.1	-68.1
10	05/01 to 19/01	126_140	50.6	0.0	301.9	-1.7	-0.1	0.0	-52.2	-52.2
11	19/01 to 02/02	140_154	167.7	0.0	300.2	69.8	-0.6	-54.4	-42.9	-42.9
12	02/02 to 16/02	154_168	69.8	0.0	369.9	-26.0	-0.3	0.0	-95.5	-95.5
13	16/02 to 01/03	168_182	152.9	0.0	344.0	21.1	-1.2	-79.4	-51.2	-51.2
14	01/03 to 15/03	182_196	73.1	0.0	365.1	-16.6	-0.4	0.0	-89.3	-89.3
15	15/03 to 29/03	196_210	14.0	0.0	348.5	-48.4	0.0	0.0	-62.4	-62.4
16	29/03 to 12/04	210_224	65.2	0.0	300.1	0.7	-0.3	0.0	-64.2	-64.2
17	12/04 to 26/04	224_238	52.3	0.0	300.8	0.5	-0.1	0.0	-51.7	-51.7
18	26/04 to 10/05	238_252	23.2	0.0	301.3	-6.4	-0.1	0.0	-29.6	-29.6
19	10/05 to 24/05	252_266	25.5	0.0	294.9	-0.1	0.0	0.0	-25.6	-25.6
20	24/05 to 07/06	266_280	101.5	0.0	294.8	53.8	-0.8	-21.3	-25.6	-25.6
21	07/06 to 21/06	280_294	7.2	0.0	348.7	-12.4	0.0	0.0	-19.6	-19.6
22	21/06 to 05/07	294_308	1.9	0.0	336.2	-20.0	0.0	0.0	-21.9	-21.9
23	05/07 to 19/07	308_322	32.7	0.0	316.2	26.0	-0.1	0.0	-6.6	-30.2
24	19/07 to 02/08	322_336	44.6	0.0	342.2	-12.9	0.0	-18.5	-39.0	-39.0
25	02/08 to 16/08	336_350	0.0	0.0	329.3	-11.4	0.0	0.0	-11.4	-49.3
26	16/08 to 30/08	350_364	0.0	37.5	317.8	-8.9	-0.4	0.0	-46.1	-51.0
Sum	01/09 to 30/08	0_364	1312.8	71.6	7931.6	63.7	-5.5	-173.6	-1141.7	-1336.1

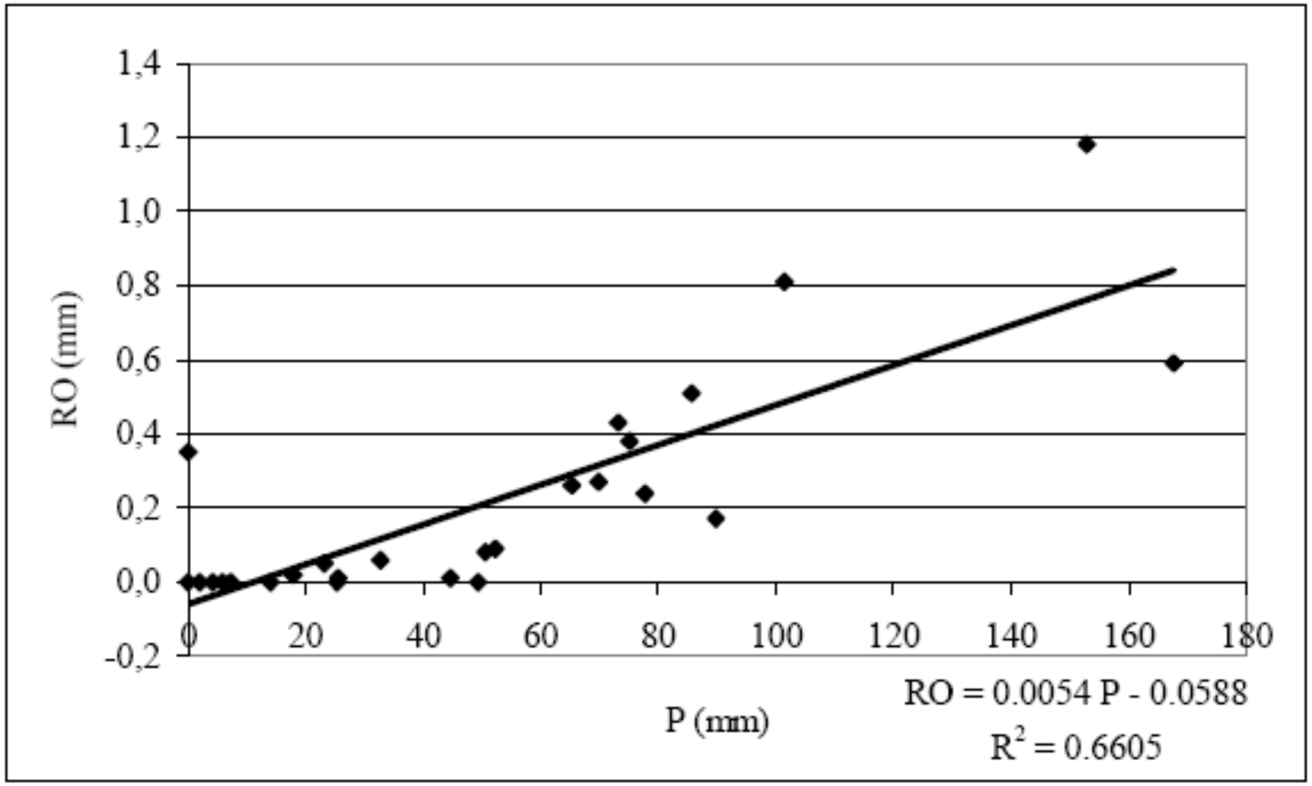

Figure 1: Variations in the runoff, RO (mm), as a function of the rainfall, P (mm).

CONCLUDING REMARKS

1. Rainfall is generally measured only at one point and, in many cases one takes the value of the nearest meteorological station. We verified that in experimental areas having obstacles nearby which affect the dynamics of the wind and, consequently, of the rainfall, the measurement of the rainfall should be made with an adequate number of replicates. In our case, an area of 0.2 ha, with trees, silo, and warehouse located within 100 m of distance, 5 rain-gauges apart from each other by 15 to 100 m, presented CVs up to 17.8%;

2. Irrigation can introduce great variability in water balance calculations when not well controlled, due to operational problems and wind drift;

3. The atmospheric demand of the coffee crop, expressed by its actual evapotranspiration, was 1141.7 mm per year. It was not affected by the parameters that characterize the stadia of growth and development of the crop. Its estimation through water balance calculations is not recommended due to error propagation. Alternative aerodynamic methods are better choices;

4. The soil in question presents a maximum capacity of soil water storage of the order of 125 mm, which represents a backup of water for 25 days, without considering the restrictions on water flux to the roots in drier periods and considering an average demand of 5 mm/day. In this year the rainfall was near to the long term average, and was enough to meet the atmospheric demand of the crop, with restrictions in the period of dry and cold winter, favorable for blossoming. Soils with smaller storage capacity are likely to cause water supply problems and also permit larger values of internal drainage and, consequently, leaching. Soil water storage, although measured carefully, was the component that introduced most variability and error propagation in water balances;

5. The planting of coffee in areas with slopes has to be made in such a way to provide good water infiltration, minimizing runoff losses and the erosion process. Planting made in furrows along contour-lines, reduced considerably the runoff and the erosion was nil. In our case, with an average slope of 10%, the value runoff was very small, of the order from 1.7% in total of the rainfall. As expected, a positive relation between the runoff and the rainfall was observed.

Temporal variability of soil water storage evaluated for a coffee field.

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The state-time analysis characterizes the state of a system (set of *p* unobservable variables) at a time *t* to its state at a time *t-j*, *j* = 1, 2, 3, …, 52, in our study. For *j* =1, the state-space approach is described as follows (called state equation):

$$X_t = \phi X_{t-1} + \omega_{X_t} \qquad (4)$$

X_t and $X_{t\text{-}1}$ being the state vector (a set of *p* unobservable variables) at time *t* and *t* -1; *f* a *p* x *p* matrix of state coefficients, which indicates the measure of the regression; and noises of the system for *t* = 1, 2, 3,…, *j*. Noise values are assumed to have zero mean, not being autocorrelated and being normally distributed with constant variances. If these *X* variables were observable, this would be the usual structure of a vector autoregressive model, in which the coefficients of the matrix *f* could be estimated by multiple regression techniques, taking X_t and $X_{t\text{-}1}$ as the dependent and independent variables, respectively.

In the case of the state-time model, however, the true state of the variables is considered “embedded” in an observation equation:

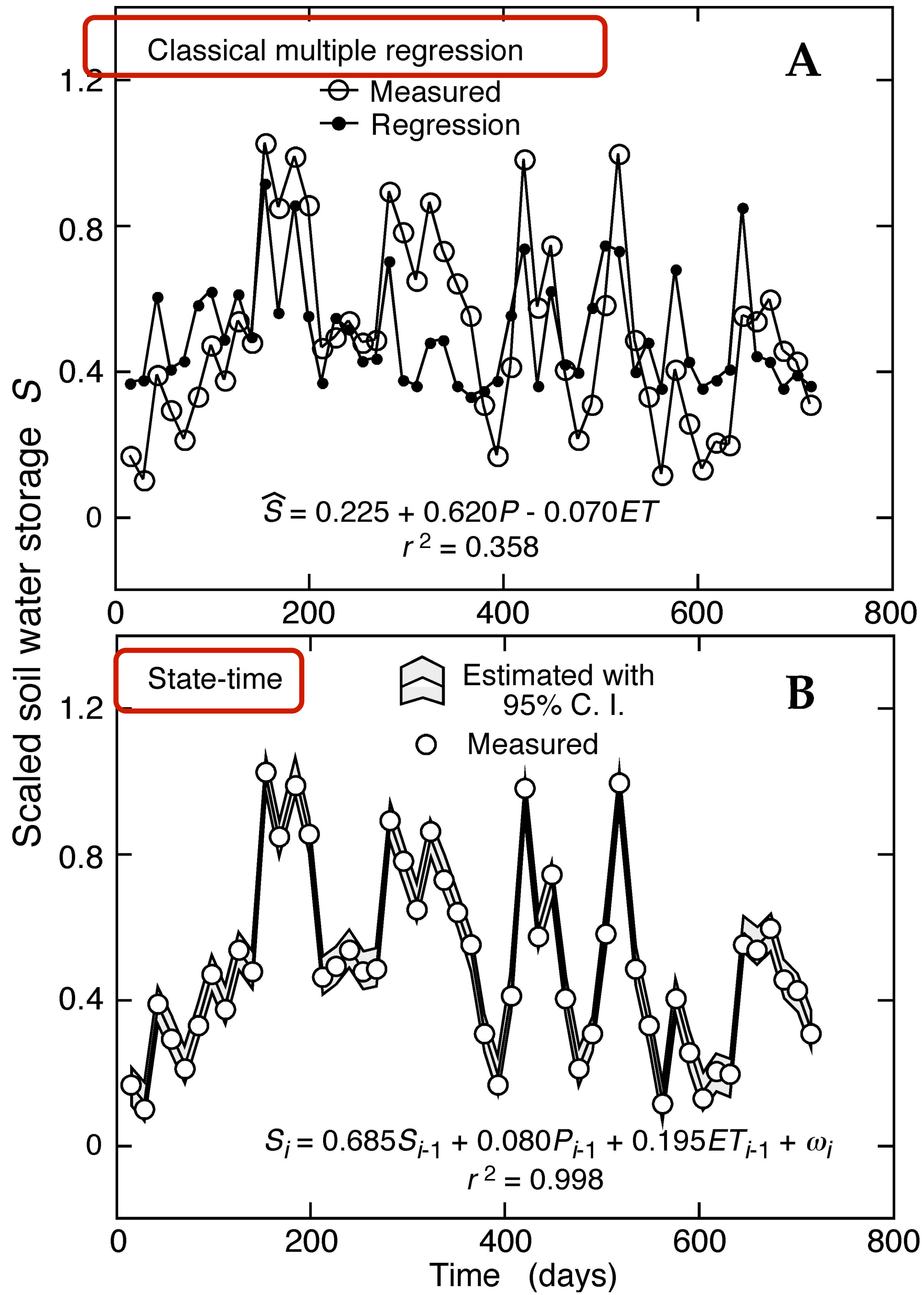
$$Y_t = AY_{t-1} + \nu_{Y_t} \qquad (5)$$

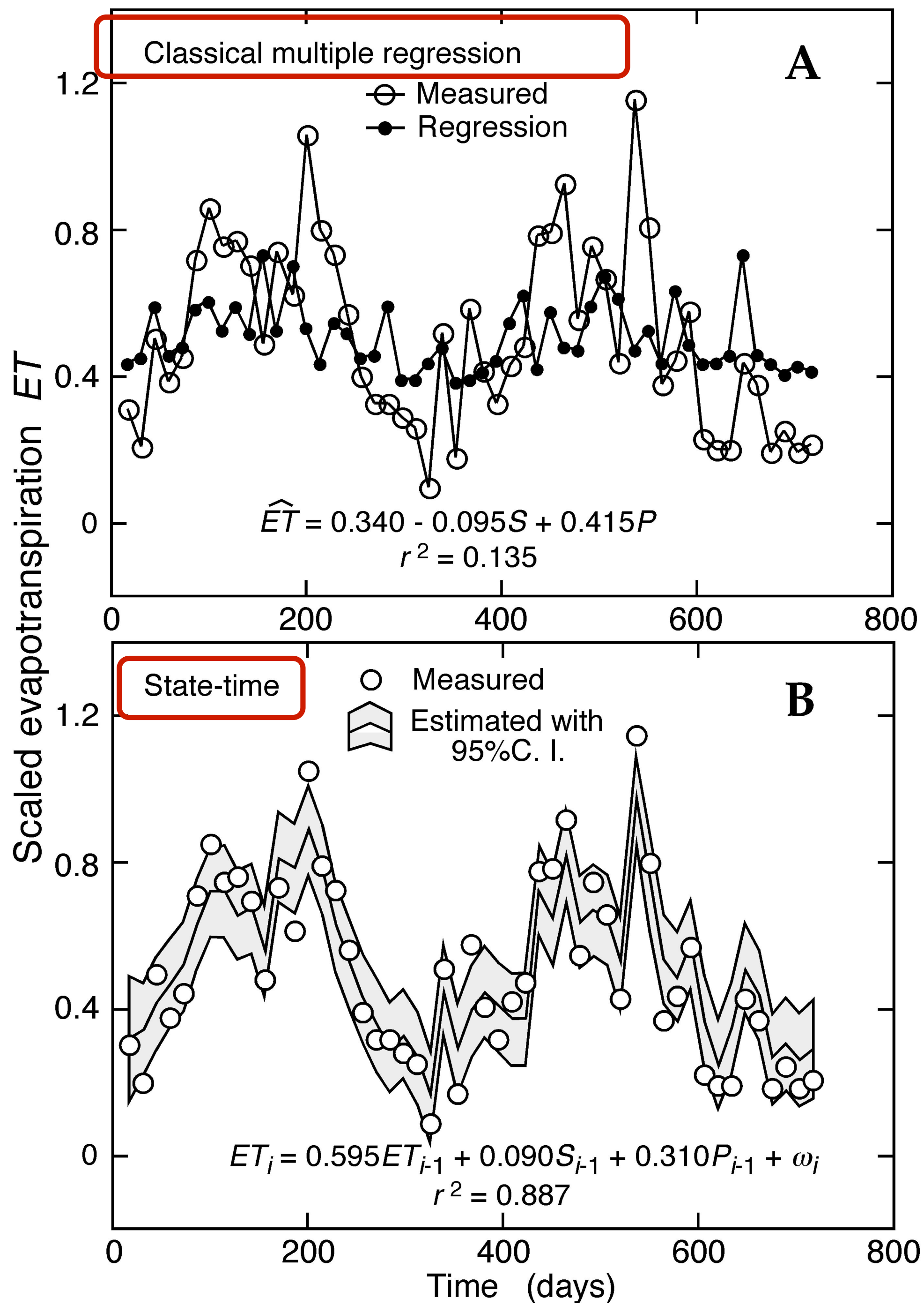
the observation vector Y_t being related to the state vector X_t by an observation matrix A (usually known as, for instance, an identity matrix, p x p) and an observation noise vector , also considered of zero mean, not autocorrelated and normally distributed. The noises and are assumed to be independent of each other. The state coefficients of the matrix *f* and noise variances of equation (4) are estimated through a recursive procedure given by Shumway and Stoffer (1982).

According to Hui *et al.* (1998), if the X_t data are scaled with respect to their mean (m) and standard deviation (s), as follows:

$$x_t = [X_t - (m - 2s)]/4s \qquad (6)$$

the transformed values x_t become dimensionless with mean m = 0.5 and standard deviation s = 0.25. This transformation allows state coefficients of the matrix f have magnitudes directly proportional to their contribution to each state variable used in the analysis. The software Applied Statistical Time Series Analysis (ASTSA) (Shumway 1988) was used for applying the state-space approach.


Classical multiple regression
A
Measured
Regression
$\widehat{S} = 0.225 + 0.620P - 0.070ET$
$r^2 = 0.358$
State-time
B
Estimated with 95% C. I.
Measured
$S_i = 0.685S_{i-1} + 0.080P_{i-1} + 0.195ET_{i-1} + \omega_i$
$r^2 = 0.998$
Scaled soil water storage S
Time (days)
0
200
400
600
800
0
0.4
0.8
1.2



Classical multiple regression
A
Measured
Regression
$\widehat{ET} = 0.340 - 0.095S + 0.415P$
$r^2 = 0.135$
State-time
B
Measured
Estimated with 95%C. I.
$ET_i = 0.595ET_{i-1} + 0.090S_{i-1} + 0.310P_{i-1} + \omega_i$
$r^2 = 0.887$
Scaled evapotranspiration ET
Time (days)
0
200
400
600
800
0
0.4
0.8
1.2

THE END