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EVAPORATION AND EVAPOTRANSPIRARION

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EVAPORATION AND EVAPOTRANSPIRARION

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These two water balance components represent the water loss flux densities, in the vapor phase, from an elemental volume containing water. Evaporation is the loss through non living surfaces, like free water bodies and bare soil surface. Evapotranspiration is the loss though living organisms, in our case, the plant. The passage of water from the liquid phase to the vapor phase, which occurs below the boiling point, depends on the available energy that ultimately comes from the sun, and on other atmospheric conditions like air temperature and humidity, and wind. The average energy/evaporation relation is 245 J mm^{-1} for temperatures in the range 10 to 30^{o} C. The process occurs even under no direct solar radiation presence, and in this case the energy is taken from the surrounding matter, like air, water itself, soil etc.

In the agro-ecosystems, evaporation occurs mainly at the surfaces of free water bodies and bare soil. Whenever plants are present, we talk about evapotranspiration. Some definitions are essential:

1. Potential Evapotranspiration (ET_0, mm) also called Reference Evapotranspiration (with symbols ETR, ET_r, ETP, ET_p), which is the water loss from a large green grass cover that occurs under conditions of no restriction of water availability. Under such conditions the atmosphere, through solar radiation, air temperature and humidity, and wind, regulates the process. It is also taken as an atmospheric potential of evaporation, in the sense that it can be calculated for situations even without the presence of water, e.g. Sahara desert. It characterizes the atmospheric demand of a region. A value of ET_0 = 12 mm, can be seen as a condition under which 12 mm of water would evaporate if water would be freely available.

2. Maximum Evapotranspiration (ET_m, mm) also called Crop Evapotranspiratio (ET_c) is the same definition of ET_0 but for a crop different than grass, i.e., corn, soybean, cotton, forest, etc, because the loss of water depends on the cover. ET_m is related to ET_0 through a crop coefficient Kc:

$$ET_m = K_c \times ET_0 \qquad (1)$$

K_c relates ET of a given crop to the ET of grass under the same atmospheric conditions. So, K_c has to be known, and data on K_c are widely available in the literature (ALLEN et al., 1998), for different crop management systems and growth stages.

3. Actual (or real) Evapotranspiration (ET_a, mm), which occurs at any moment of an agro-ecosystem, with or without water availability restriction. Without restriction $ET_a = ET_m$, and under restriction $ET_a < ET_m$. The soil can restrict the flow of water to its surface and to plant roots. Here Soil Physics plays an important role. Soil water retention and transmission characteristics control water movement in the soil.

There are several methods that estimate ET_0 or ETP from atmospheric data. Thorntwaite presented one of the first methods, based on air temperature only, that is widely used to date. The calculation of ETP_{TH} is based on the equation:

$$ETP_{TH} = f * 16 * \left(10 * \frac{Tn}{I}\right)^a \quad (2)$$

where Tn is the temperature of month n, in °C; I the heat index of the region calculated according to Equation 4; f is a correction factor for latitude and month of the year (Table 1); and a is a regional thermal index calculated by Equation 5, in mm month^{-1}. The f factor is important to correct for the real number of days of each month.

Equation 2 is used for $0 \leq Tn < 26{,}5°C$. For $Tn \leq 26{,}5°C$, ETP_{TH} is given by:

$$ETP_{TH} = -415{,}85 + 32{,}24Tn - 0{,}43Tn^2 \quad (3)$$

Table 1 – Monthly correction factor f for latitude 22° S

Month	*f*	Month	*f*
January	1,14	July	0,94
February	1,00	August	0,99
March	1,05	September	1,00
April	0,97	October	1,09
May	0,95	November	1,10
June	0,90	December	1,16

Source: Thornthwaite (1948); Pereira; Angelocci; Sentelhas (2002)

The value of I depends on the annual rhythm of the temperature and integrates the thermal effect of each month (PEREIRA; VILLA NOVA; SEDIYAMA, 1997; PEREIRA; ANGELOCCI; SENTELHAS, 2002):

$$I = \sum_{n=1}^{12} (0{,}2Tn)^{1{,}514} \tag{4}$$

In the same way as I, a is calculated with the climatologial normals, with characteristic coefficients for each region, independent of the year under study. The a exponent is calculated as:

$$a = 6{,}75 * 10^{-7} * I^3 - 7{,}71 * 10^{-5} * I^2 + 1{,}7912 * 10^{-2} I + 0{,}49239 \tag{5}$$

To estimate ETp by Penmam (ETPp) the following equation is used:

$$ETP_p = \frac{W * R_n}{\lambda} + (1 - W)E_a \tag{6}$$

where λ is the latent evaporation heat (MJ kg^{-1}); W is a weight factor dependent on air temperature do ar (Equation 7); R_n the net radiation (MJ m^{-2} d^{-1}); E_a the evaporative air power (MJ m^{-2} d^{-1}), obtained by Equation 11.

$$W = \frac{\Delta}{\Delta + \gamma} \tag{7}$$

with Δ equal to the slope of the saturation vapor pressure VS air temperature, in kPa °C^{-1} (Equation 10), and γ the psychometric constant related to the atmospheric pressure (Pa) by:

$$\gamma = 0{,}664742 * 10^{-3} * Pa \tag{8}$$

$$\Delta = \frac{4098 * e_s}{(T_{med} + 237{,}3)^2} \tag{9}$$

with e_s equal to the saturation vapor pressure (kPa), calculated by equation 10 and T_{med} the average air temperature (°C).

$${e_s}^{T_{med}} = 0{,}6108 e^{\frac{17{,}27 * T_{med}}{237{,}3 + T_{med}}} \tag{10}$$

The evaporation power of the air (E_a, MJ m^{-2} d^{-1}) is given by:

$$E_a = f(U)DPV \tag{11}$$

where $f(U)$ is given by Equation 13 and DPV is the vapor pressure deficit (kPa), defined as:

$$DPV = e_s - e_a \tag{12}$$

with e_a equal to the actual vapor pressure (kPa). The empirical function of the wind velocity $f(U)$ is given by:

$$f(U) = m(a + bU_2) \tag{13}$$

with m equal to 6.43 MJ m^{-2} d^{-1} kPa^{-1}; a = 1; b = 0.526 s m^{-1} and U_2 the Wind speed 2 m above soil surfece (m s^{-1}).

By the Penman-Monteith method, ETPpm is calculated through:

$$ETP_{PM} = \frac{0{,}408\ \Delta\,(R_n - G) + \gamma \frac{900}{T_{med} + 273{,}16} U_2 * (e_s - e_a)}{\Delta + \gamma(1 + 0{,}34 U_2)} \tag{14}$$

where G is the soil heat flux density (MJ m^{-2} d^{-1}.).

As we have seen in our first lecture, these evapotranspiration definitions and estimations, can be used to calculate water balances (WBs). We give here as examples of climatologic WBs, the methods of Thornthwaite and Mather (THORNTHWAITE-MATHER, 1955), Rijtema and Aboukhaled (RIJTEMA; ABOUKHALED, 1975; DOURADO-NETO; DE JONG VAN LIER, 1993) and the Cossenoidal (DOURADO-NETO; DE JONG VAN LIER, 1993). The main components of these balances are the evapotranspiration ET and the rainfall P. The difference P – ET is called first balance B, when positive indicating water excess EXC and when negative deficit DEF. Under deficit conditions the soil enters as a water source.

First these WB programs calculate ET_0 according to one of the above described methods, and with the crop coefficient Kc these values are transformed into ETp, to thereafter calculate B = P – ETp. We take a monthly WB example, for which a balance sheet is organized including columns of i (month), P_i, ET_{0i}, B_i, the accumulated negative L_i (explained below), the soil water storage S_i, ET_a, DEF and EXC, as shown in Table 2.

Table 2 - An example of a Thornthwaite and Mather climatologic water balance sheet

Month i	ETPi	Pi	Bi	Li	Si	ΔSi	ETai	DEFi	EXCi
1	124.0	300	176.0	0.0	125.0	0.0	124.0	0.0	176.0
2	106.4	250	143.6	0.0	125.0	0.0	106.4	0.0	143.6
3	114.7	70	-44.7	44.7	87.4	-37.6	107.6	7.1	0.0
4	108.0	0	-108.0	152.7	36.8	-50.6	50.6	57.4	0.0
5	108.5	0	-108.5	261.2	15.5	-21.4	21.4	87.1	0.0
6	75.0	0	-75.0	336.2	8.5	-7.0	7.0	68.0	0.0
7	80.6	0	-80.6	416.8	4.5	-4.0	4.0	76.6	0.0
8	86.8	60	-26.8	443.6	3.6	-0.9	60.9	25.9	0.0
9	90.0	120	30.0	164.2	33.6	30.0	90.0	0.0	0.0
10	99.2	150	50.8	49.1	84.4	50.8	99.2	0.0	0.0
11	120.0	190	70.0	0.0	125.0	40.6	120.0	0.0	29.4
12	127.1	280	152.9	0.0	125.0	0.0	127.1	0.0	152.9
Year	1240.3	1420.0					918.1	322.2	501.9

Soil water storage is taken as the amount of water in mm, present in the soil layer chosen for the balance. A saturated soil has a volumetric water content θs, which is subject to drainage up to the field capacity FC, a point with $\theta = \theta_{FC}$. Plants extract soil water up to the permanent wilting point PWP, a point with $\theta = \theta_{PMP}$. Below wilting point water is not available to plants anymore. The interval FC – PMP is called available water capacity AWC, which can be expressed in $m^3\ m^{-3}$ as $\theta_{FC} - \theta_{PMP}$, or in mm, using the concept of S. As already mentioned above, stating with a soil at saturation, ET = ET_0 up to θ_{FC} and thereafter soil water is extracted by plants up to θ_{PMP}. As a soil dries out, the facility of plants to extract water decreases due to water movement reduction from soil to root because the soil hydraulic conductivity is reduced. Under such conditions B becomes negative and ET = ETa, or ETa < ETp. The reduction of ETa as time t passes, equal to ETa, is assumed different for the authors here considered:

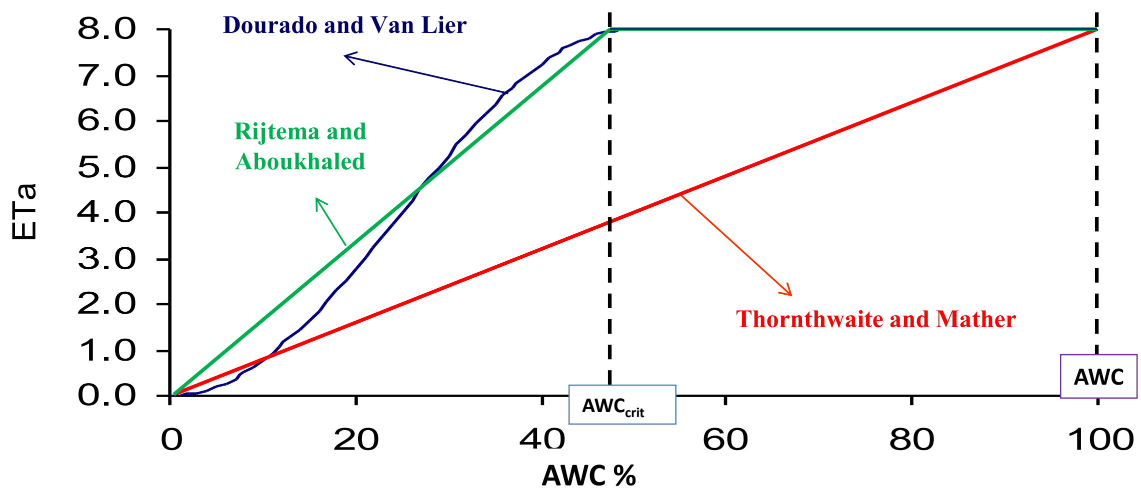

Figure 1 - Rate of soil water loss (ETa, mm/period of time) as a function of storage (AWC, mm) for the methods of Thornthwaite and Mather, Rijtema and Aboukhaled and Dourado and Van Lier.

Thornthwaite and Mather consider a constant rate of ETa decrease in time from the FC to the PWP. This means that the restriction of the soil in allowing the plant to extract soil water begins at the FC and is linear reaching zero at the PWP (Figure 1). In the balance sheet shown in Table 2, when B_i is negative, they consider :

$$L_i = L_{i-1} - B_i \quad (15)$$

When the dry season starts (Table 2) and Bi starts to become negative (B3 in example of Table 2), Li-1 is considered 0 with S = AWC (line 2 in Table 2, AWC = 125 mm). Due to the assumption that dET_a/dt is linear, the changes in S_i decrease exponentially, indicating that as time passes it is more and more difficult to extract water from the soil. It is demonstrated that Si can be calculated as:

$$S_i = AWC\ e^{-\frac{L_i}{AWC}} \tag{16}$$

When the rainy season begins, Bi becomes positive and the soil reservoir is filled up again. In this case,

$$S_i = S_{i-1} + B_i \tag{17}$$

When Si becomes greater than the AWC, there will be EXC of water and S_i is maintained as AWC. In these cases:

$$L_i = -AWC\ ln \frac{S_i}{AWC} \tag{18}$$

Which is again a consequence of the linearity of dET_a/dt.

Rijtema and Aboukhaled (1975) consider that $ET_a = ET_m$ for the initial extraction of the AW, up to a critical point and from there on, the decrease of ET_a is also considered linear as in the case of Thornthwaite and Matter (Figure 1). For this method, a p factor is considered related to water availability, to estimate S and that is tabulated in Allen et al. (1998) for ET_a of 5 mm d^{-1}. Days for which this condition is not observed, p is calculated as:

$$p = 0{,}5 + \big(0{,}04(5 - ETC)\big) \tag{19}$$

If $(1-p)AWC \leq S_{i-1} \leq AWC$, then:

$$S_i = S_{i-1} + B_i \tag{20}$$

$$L_i = 0 \tag{21}$$

If $0 \leq S_{i-1} \leq (1-p)AWC$ and B is less than zero:

$$L_i = L_{i-1} - B_i \tag{22}$$

$$S_i = (1-p)AWC\ \exp^{\left(\frac{p-\frac{L}{AWC}}{(1-p)}\right)} \tag{23}$$

If $0 \leq S_{i-1} \leq (1-p)AWC$ and B is greater than zero:

$$S_i = S_{i-1} + B_i \tag{24}$$

$$L_i = AWC\ [p - (1-p)\ \ln(\frac{S}{(1-p)AWC}) \tag{25}$$

When Si is greater than the AWC, Si = AWC.

For the Dourado and Van Lier method, ET_a is also assumed equal to ET_m up to the critical point, but from there on the reduction of ET_a is considered cosenoidal, i.e., dET_a/dt has a coscenoidal shape (Figure 1). This approach has the advantage of eliminating the sharp beak of the curve at the critical point and also leads ET_a assymptotically to zero. The parameter p is also calculated according to equation 19, and the rate of soil water loss is assumed to be cosenoidal, Li and Si are estimated as:

If $0 \leq S_{i-1} \leq (1-p)AWC$, then:

$$L_i = AWC\left\{p + \frac{2}{\pi}(1-p)tg\left[\frac{\pi}{2}\left(1 - \frac{S}{(1-p)AWC}\right)\right]\right\} \tag{26}$$

And when $(1-p)AWC \leq S_{i-1} \leq AWC$:

$$L_i = AWC - S_i \tag{27}$$

For soil reservoir filling, when $L_{i-1} \geq pAWC$

$$S_i = (1-p)AWC\left\{1 - \frac{2}{\pi}\,arctg\left[\frac{\pi}{2}\left(\frac{\left(\frac{L}{AWC}\right)-p}{1-p}\right)\right]\right\} \tag{28}$$

And when $0 \leq L_{i-1} < pAWC$:

$$S_i = AWC - L_i \tag{35}$$

These WB methods evaluate both EPP and Eta. There are several other methods for their evaluation, as field WBs, Lysimeters, etc.

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Following this presentation of ET definitions and their measurements through climatologic data, we preset an analysis of WB components directly measured in the Field. This work was published as Timm et. al.(2011), with the title:

TEMPORAL VARIABILITY OF SOIL WATER STORAGE EVALUATED FOR A COFFEE FIELD.

Soil water storage (*S*) in agricultural soil profiles is an important parameter for a rational management of any crop, besides giving information on environmental aspects of the water cycle. Spatial variability of *S*, however, imposes serious problems when determining average values over large areas, which are needed to take actions in relation to water availability to crops. The variability of *S* is a consequence of the erratic rainfall input, differences in crop stand, and of natural soil matrix differences that can occur over short distances as well as over large fields due to soil genesis and topography. The knowledge of the characteristics of the variability of *S* helps to understand and predict several hydrologic processes (Western *et al,*. 2004) and to improve soil water sampling strategies (Warrick and Nielsen, 1980).

The variability of soil physical and chemical properties is not a new research topic. Since the first half of last century the way of obtaining the representative sampling of agricultural fields always lead to the development of new sampling schemes. First scientists based their strategies on classical statistics concepts which were later complemented with geoestatistics and time-space series analyses, and more recently using neural networks (Hills and Reynolds, 1969; Mohanty and Mousli, 2000; Western *et al,*. 2002; Timm *et al.,* 2006; Hu *et al.,* 2008).

The temporal stability of *S* measurements was first indicated by Vachaud *et al.* (1985), who statistically determined the presence of locations that systematically presented soil water contents above or below the field average. Kachanoski and De Jong (1988) and Moreti *et al.* (2007) also used this concept to show the temporal persistence of spatial patterns of soil water storage. Reichardt *et al.* (1997) suggested that part of the time stability of soil water content measurements is due to systematic errors introduced by soil water content calibration curves, when indirect methods of measurement are employed, such as neutron probes, time domain reflectrometry (TDR), and frequency domain reflectrometry (FDR). Hu *et al.* (2008) verified the time stability of soil water content measurements made using this last methodology at the soil surface layer of a hill-slope of the Loess Plateau in China, and found significant correlations with several landscape influencing factors. More recently Hu *et al*. (2010) presented a new criterion to identify sites for S determinations based on the mean absolute bias error.

Few studies have analyzed the time variability of S as affected by evapotranspiration and rainfall. A comprehensive report, however, has been presented by Aboitiz *et al.* (1986), who developed a methodology for estimating and forecasting soil water depletion and evapotranspiration in irrigated fields, using a time-varying state-space model, which we here call state-time. We have the aim of contributing to the improvement of water management practices of natural ecosystems and perennial crops such as the coffee crop, analyzing a 2-year series of soil water storage measurements, giving emphasis to the time stability and spatial variability of this set of data. A new perspective and a deeper insight is made through a state-time analysis to better understand the temporal relations between soil water storage, rainfall and evapotranspiration.

This study analyses the temporal variability of soil water storage (S mm) data collected in a coffee crop grown in Piracicaba, SP, Brazil (22^o 42' 30'' S; 47^o 38' 00' 'W, 580 m asl). Soil water contents $\theta(i)$ were measured along a horizontal domain x_i (m) at 15 locations (i = 1, 2,…,15), and at five depths z_k (m), 0.2, 0.4, 0.6, 0.8, and 1.0 m from surface (k = 1, 2,…,5), every 14 days, at times t_j (j = 1, 2, 3,…,52) covering a two year period starting on September 01, 2003. Soil water content measurements obtained with a neutron probe (model CPN 503 DR) were not taken at regular spacings along a leveled contour line of the horizontal domain corresponding to a coffee row, following the distribution of five fertilizer plots arranged within a 0.2 ha coffee field. Details of the fertilizer trial can be found elsewhere (Fenilli *et al.* 2007). Measurements of θ were made using aluminum neutron probe access tubes installed below crop canopies. The coffee (*Coffea arabica* L.), was of the cultivar "Catuaí Vermelho" (IAC-144) and is a perennial crop, 3 to 5 years old during the experimental period, which is the beginning of the yearly coffee production cycles. The spacing between plants was 0.75 m and between rows 1.5 m. Rows were kept bare chemically and manually, as commonly done in coffee plantations.

The soil is a Rhodic Kandiudalf (Soil Survey 1993), locally called "Nitossolo Vermelho Eutroférrico" (Embrapa 2006); and the climate is of the Cwa type (Köppen 1931), with dry winter.

Slow neutron counting data were transformed into soil water contents using calibration curves established as suggested by Reichardt *et al.* (1997), taken as valid over all depths. Soil water storages at times *j* and positions *i*, $S_j(i)$ (mm) for the 0 – 1.0 m soil layer were calculated from $\theta_{t,x}(k)$ data by the trapezoidal rule:

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

with Δz = 0.2m. Soil water contents $\theta_{i,j}(1)$ measured at the depth 0.2m (k = 1) were considered to cover a layer of 1.5Δz = 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 0.5Δz = 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. The coffee root system was assumed not to reach depths below z = 1.0m, which was confirmed by Silva *et al.* (2009).

In order to apply the following statistical procedures, $S_j(i)$ data were tested for normality with respect to space performing cumulative probability plots.

To reduce the number of observation points so that future evaluations of the soil water status of this perennial coffee field could be made more rapidly and without losing accuracy, two approaches were used: i. making a time stability analysis to find out which access tube can represent the overall average of the field, and ii. establishing the minimum number of observation points that would yield an average value within a pre-established coefficient of variation. To verify the time stability of the measurements, the approach proposed by Vachaud *et al.* (1985) was used. For this, the relative deviation $\delta_j(i)$% of each $S_j(i)$ realization in relation to the mean soil water storage $\overline{S_j}(i)$, was calculated as follows:

$$\delta_j(i) = \frac{S_j(i) - \overline{S_j}(i)}{\overline{S_j}(i)} \times 100 \tag{2}$$

According to Vachaud *et al.* (1985), very small time variations of $\delta_j(i)$ indicate a time stability of $S_j(i)$, so that consistently wetter or dryer positions (i) can be selected in the field. Therefore, if time averages $\overline{\delta_i}(j)$ of the $\delta_j(i)$ values are plotted in rank, it is possible to find out which sites present systematically $S_j(i)$ values below or above the position time average $\overline{S_j}$ and also those sites that systematically present a negligible $\overline{\delta_i}(j)$ and, therefore, represent $\overline{S_j}$.

To estimate the number of observations N needed in a new sampling event to obtain a mean value $S_t(i)$, within a chosen deviation (%) of the estimated mean value, the suggestion of Warrick and Nielsen (1980) was applied:

$$N = t_{\alpha}^{2} s^{2} d^{2} \quad (3)$$

where t_{α} is the value of the *t* student distribution considering the level of significance α (for α = 5% the *t* value is 1.96) for infinite degrees of freedom; s^2 is the variance of a previous sampling event $S_t(i)$ made with *n* (15 in our case) replicates, and *d* any desired deviation from the mean for example [0.5, 1, 2%,...of $\overline{S_t}(i)$]. Equation (3) assumes that the samples are independent, the central limit applies and that the true mean deviation σ can be represented by the standard deviation *s*.

In a second step, the time variability structure of the $\overline{S_j}$ data was studied using the state-time approach (Shumway 1988; Nielsen and Wendroth 2003) which provides opportunities for a suitable identification of temporal relations between soil-atmosphere-plant variables taking into account their temporal association. 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):

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

X_t and X_{t-1} being the state vector (a set of *p* unobservable variables) at time *t* and *t* -1; ϕ a *p* x *p* matrix of state coefficients, which indicates the measure of the regression; and ω_{X_t} 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 ϕ could be estimated by multiple regression techniques, taking X_t and X_{t-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:

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

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 ν_{Y_t}, also considered of zero mean, not autocorrelated and normally distributed. The noises ω_{X_t} and ν_{Y_t} are assumed to be independent of each other. The state coefficients of the matrix ϕ 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:

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

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 ϕ 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.

Concomitantly to $S_j(i)$ measurements, Silva *et al.* (2006) evaluated time series of evapotranspiration $ET_j(i)$, rainfall $P_j(i)$, supplementary sprinkler irrigation $I_j(i)$, surface runoff $RO_j(i)$, and soil water drainage fluxes $Q_j(i)$ below the 1.0m depth, to establish complete water balances, which were used in the state-time and multiple regression analyses. Irrigation was applied only during the dry winter, in just a few events when the available water capacity reached about 25% of its maximum. For the analysis, I was added to P. For a few 14 day intervals with no rainfall during the rainy season, a negligible value of P = 0.1 mm was assumed for this variable, so that the state-time analysis could be performed. It is important to mention that classical multiple regression is based on mean values of each variable throughout the time being investigated and that the magnitudes of each variable at a given time compared to their respective values at a previous or future time are neglected.

Coefficients of variation (CV), cumulative probability plots and rank plots were also used in the analysis (SAS and R statistical programs).

Soil water storage $S_t(i)$ data were normally distributed for all 52 measurement dates, as exemplified in Figure 1 through cumulative probability plots for a wet period (January 31, 2005) and for a dry period (September 01, 2003). These spatial data presented space coefficients of variation for fixed times j in the range of 1.1 to 5.9%, indicating that the variability in space can be considered low.

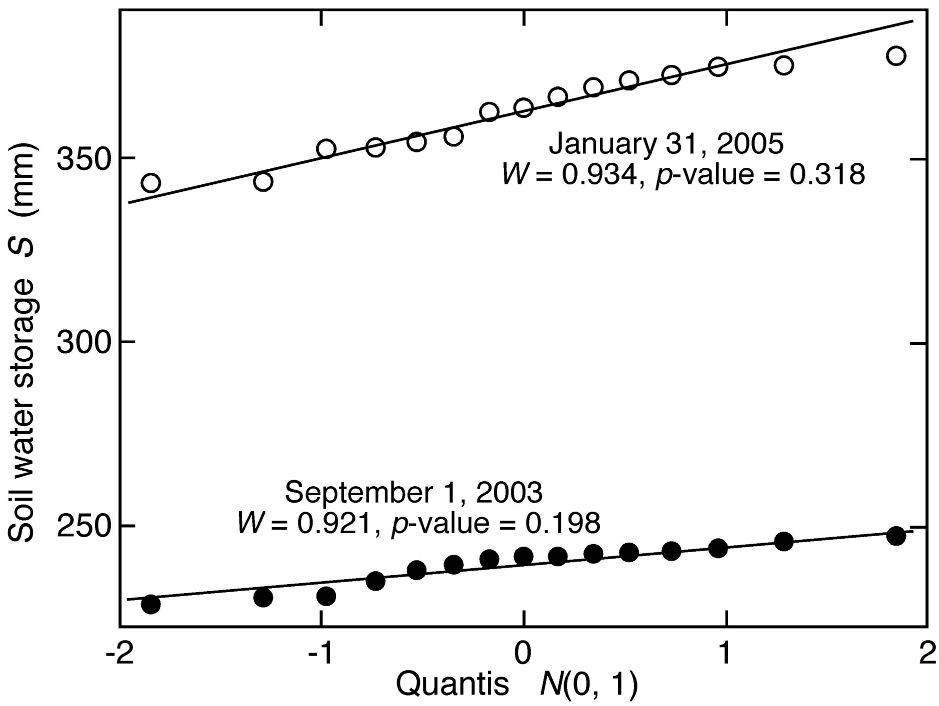

Figure 1 - Cumulative probability plots of soil water storage $S_t(i)$ for two selected dates.

Ranges of soil water storage changes $\Delta S_t(i)$ shown in Figure 2, in which positive values represent maximum soil water recharges occurring in 14-day intervals and negative values represent soil water maximum depletions in 14-day intervals, reflect the great time variability of $S_t(i)$ data observed during the two years in this field. Such plots give a good idea of the spatial variability of soil water storage measurements made in agricultural fields, as in this case for a coffee crop field, justifying the search for good and stable averages of S for water management purposes.

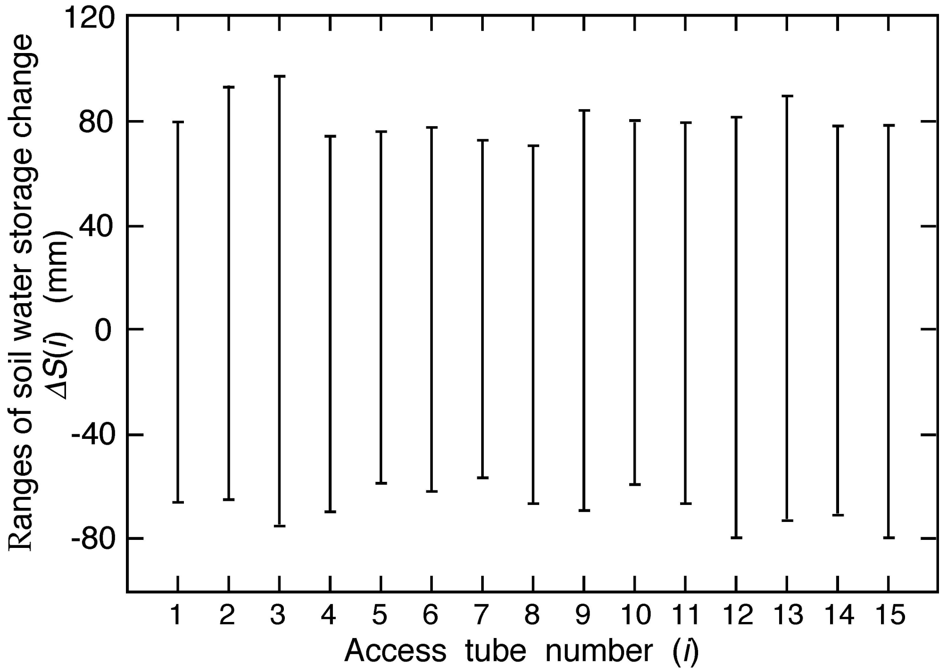

Figure 2 - Ranges of soil water storage changes $\Delta S_t(i)$ observed for the fifteen neutron probe access tubes during the two year observation period, in a coffee crop field.

For future measurements of $S_t(i)$ in the same or other fields of similar condition, the minimum number N of observation points was calculated for chosen precision levels according to equation (3). Selecting three dates for which the $S_t(i)$ value is of the order of 300 mm: j = 10, $S_{10}(i)$ = 302mm, $s_{10}(i)$= 18mm; j = 20, $S_{20}(i)$= 296mm, $s_{20}(i)$= 8mm; and j = 30, $S_{30}(i)$= 285mm, $s_{30}(i)$= 3mm, for which $s_t(i)$ were maximum, medium and minimum, the deviations (*d*) from the mean are 5.9; 2.7; and 1.1%, respectively. For new samplings according to equation (3), if the desired $\overline{S_t}(i)$ of 300mm should be evaluated within 0.5; 1 or 2% of the correct value, with an average $s_t = 8$ mm, the number of samplings would be 56; 14; and 4, respectively. For this example, the only viable choice to reduce the number of sampling points is to accept a deviation of 2% and make future measurements in 4 access tubes.

In terms of time stability of the measurements, the rank plot presented in Figure 3 shows that position 3 best represents the mean $\overline{S}_t(i)$ over the two years of observation, which means that future observations of $\overline{S}_t(i)$ could be performed at this single site or at four sites as discussed above (sites 2, 6, 3, and 10, Fig. 3), with much lower coefficients of variation than 2% used in equation (3) since the chosen four points present the least deviation from the mean. Such measurement would represent the mean soil water storage of the whole field, greatly simplifying future experimental field work. This reduction of observation points is very important for long term experimentation in natural ecosystems or perennial crops like coffee, when $S_t(i)$ is observed over long periods of time (years), e.g. Silva *et al.* (2006) and Moreti *et al.* (2009). It is important to recall that in the establishment of field water balances the soil components are the more laborious measurements.

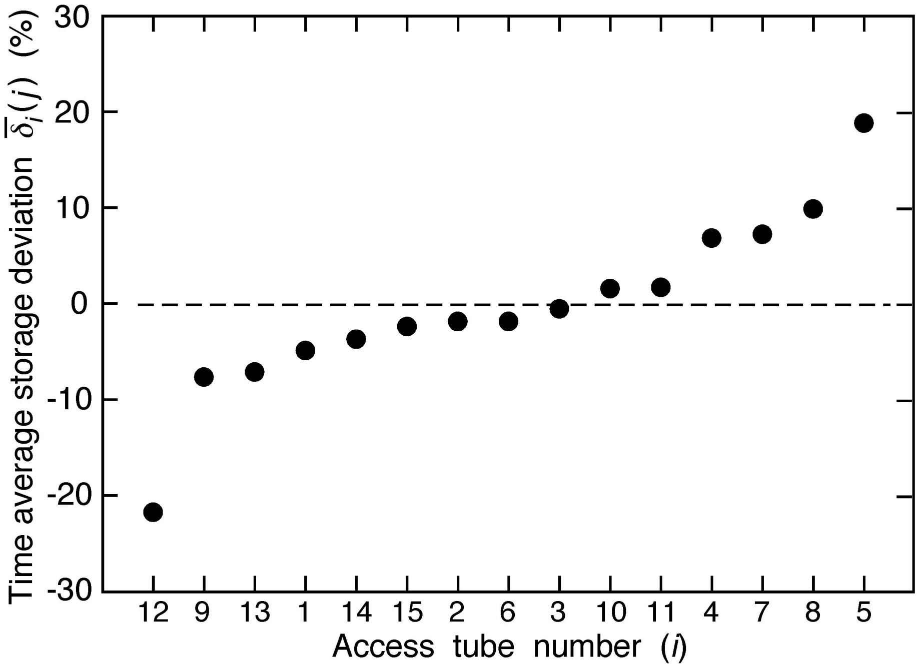

Figure 3 - Rank plots of time average relative spatial storage deviations $\overline{\delta_i}(j)$.

A great shortcoming of the time stability as a criterion to reduce the number of sampling points is the need of representative previous information in space and time, in order to be able to make significant rank plots of mean deviations from the mean. Therefore, the approach presented here is more suitable for long duration experiments in which costly and time-consuming variables are measured.

As discussed below, the state–time analysis is a step ahead of the previous discussion since it allows a better insight of the relations among the climate variables that determine *S*. So, in order to better understand the temporal relations between *S*, *P,* and *ET*, a discussion is made comparing the state-time analysis to the classical multiple regression using the same state variables. Figures 4A and 4B show the multiple regression and state-time equations and the value of their coefficients of determination (r^2) from linear regressions between estimated and measured values of scaled (equation 6) soil water storage. Classical multiple regression is based on mean values of each variable throughout the time being investigated, in which the magnitudes of each variable at a given time compared to their respective values at a previous or future time are neglected, so that no more 35.8% of the variance of the biweekly-measured soil water storage data was explained from the measurements of precipitation and evapotranspiration (Figure 4A).

Estimated values by regression are much less variable than those measured, and consistently underestimate the larger and overestimate the smaller measured values.

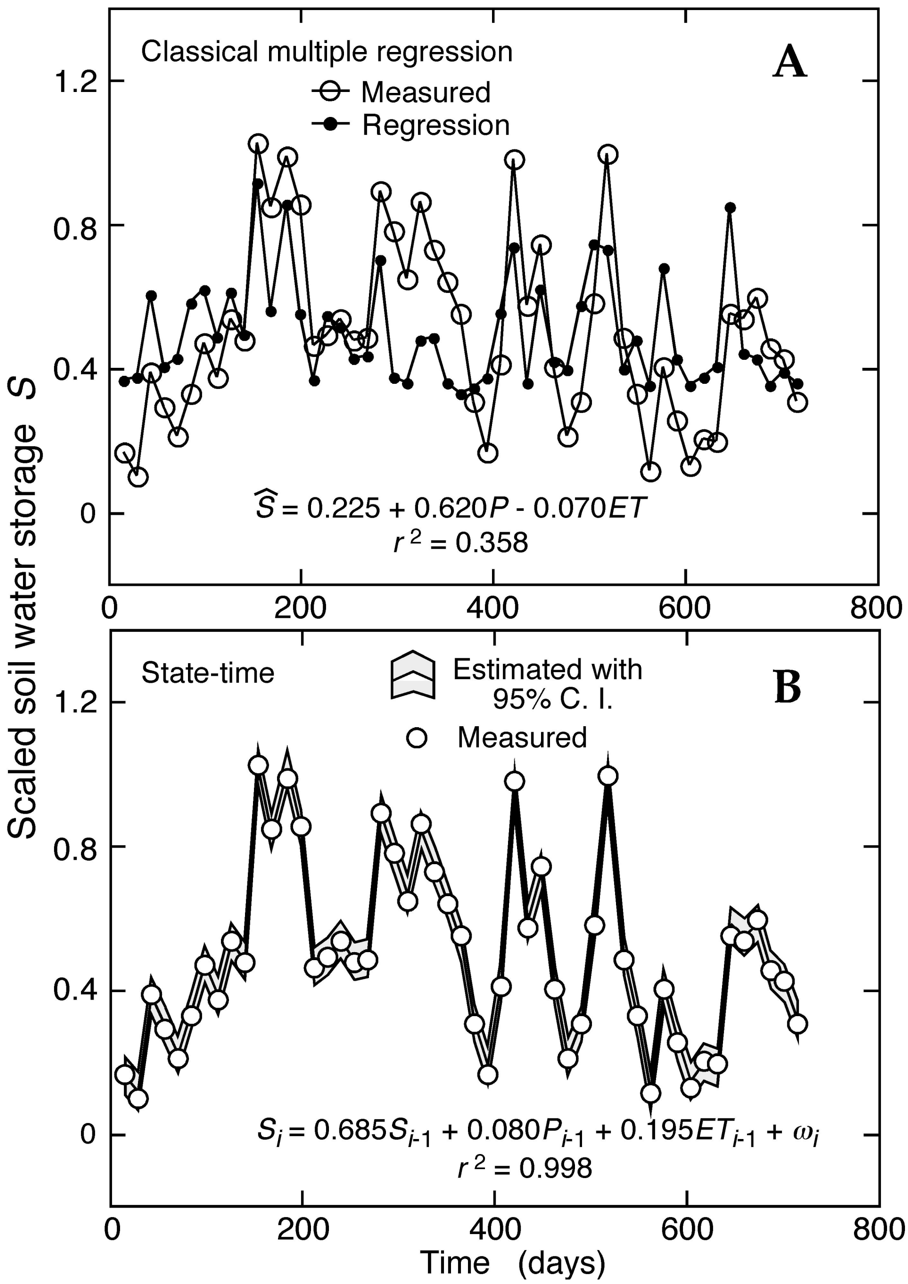

Figure 4 - Estimates of soil water storage measured biweekly for 714 days using A. classical multiple regression and B. state-time analysis.

When the temporal associations among soil water storage, precipitation, and evapotranspiration data were considered, 99.8% of the variance of the soil water storage was explained from the use of the state-time analysis (Figure 4B). We note that nearly 70% of the previous value $S_{i\text{-}1}$ contributes to that of S_i while preceding values $P_{i\text{-}1}$ and $ET_{i\text{-}1}$ contribute only 8 and 20%, respectively.

The major experimental consideration influencing the utility of state-time analyses is the time interval between successive measurements that allows the possibility of state variables to be temporally associated. In other words, measurements taken during very short time intervals will tend to be autocorrelated or cross correlated with each other. However, with increasing time, the state variables change their magnitudes as environmental conditions change. We know that a water balance for a given soil profile is the result of five processes that occur as a function of time – precipitation plus irrigation, surface runoff, evapotranspiration, storage of water in the soil profile and the drainage of water from the soil profile. Each of these processes quantified by Silva *et al.* (2006), who provide data for this study indicated that surface runoff was negligible over the two year period and that the drainage of water from the soil profile has yielded accurate measurements of water storage S_i in the profile. Hence, neglecting surface runoff, the use of only three state variables (S, P and ET) in the state-time analysis accounts for the physical processes responsible for a quantitative estimate of S provided that the amounts of water that eventually drain from the 1-m soil profile from occasional large rainfalls can be robustly accounted for in the state variable P. The temporal autocorrelation and cross correlations functions given in Table 1 indicate that ET, S and P have autocorrelation lengths of 3, about 2 and less than 1 lag, respectively. In other words, values of ET are related to each other during more than 3 consecutive sampling dates (42 days), those of S during no more than 2 consecutive sampling dates (28 days) and those of P are essentially not related to each other between consecutive sampling dates (14 days). All three values of lag are reasonable, including that for precipitation. Indeed, the general nature of rainfall is more seasonal and does not consistently repeat its relative magnitude with a 2-week periodicity through a 2-year period.

Table 1 – Autocorrelation and cross correlation coefficients for state variables soil water storage S, precipitation P, and evapotranspiration ET.

	Autocorrelation Coefficient $r(h)^{\dagger}$			Cross Correlation Coefficient $r_c(h)^{\dagger}$		
lag $h^{\dagger\dagger}$	S	P	ET	S vs P	ET vs S	ET vs P
0	1	1	1	0.595	0.153	0.359
1	0.551	0.163	0.558	0.370	0.005	0.507
2	0.257	0.119	0.444	0.203	-0.053	0.316
3	0.038	0.024	0.344	0.033	0.036	0.375
4	-0.005	0.082	0.185	-0.050	0.159	0.072
5	0.025	0.081	0.099	-0.027	0.126	-0.028

† The 95% significance level of r and r_c is 0.2745. ††A lag of h = 1 is equal to 14 days.

Examining the cross correlation coefficients in Table 1, we are not surprised to find that ET is related to P for more than 3 consecutive sampling dates (42 days) and that S is related to P

for at least 2 consecutive sampling dates (28 days). The fact that *ET* and *S* showed essentially not to be related to each other between sampling dates is not obvious since in many occasions the actual value of *ET* was much below the potential value. However, during the 2-year period, regardless of the daily and biweekly fluctuations of local weather conditions, every effort was made to irrigate the field in a timely manner to provide adequate amounts of water stored in the root zone.

There are several methods available to examine the reliability of state-time analyses (see for example, Shumway and Stoffer, 2000). Here, we choose (on the basis of the information in Table 1) to observe the impact of omitting increasing numbers of observations from the calculations of the state variable being estimated. An example is given in Figure 5 where the soil water storage is estimated with all measurements of *P* and *ET*, but with increasing numbers of its biweekly measurements omitted from the state-time analysis.

Figure 5A illustrates the results when one-half of the observations of soil water storage were not considered in the calculations. Comparing Figures 4B and 5A, it can be seen that the coefficient of determination r^2 decreased slightly from 0.998 to 0.957 and that the width of the confidence intervals increased. At each time step when a measured value of *S* is omitted from the calculation, its forward prediction cannot be compared to its observation, and hence, an update based on its temporal association is precluded and causes a larger confidence interval.

State-time estimates in Figure 5B made while ignoring two out of every three observations of soil water storage are not as good as those illustrated in Figure 5A. Nevertheless, a linear regression between state-time estimated and measured values of *S* yielded a coefficient of determination $r^2 = 0.834$. However, notice that about five values omitted in the calculations fall outside of the confidence interval as a result of the state-time analysis judging they did not belong to the distribution of *S* values used in the calculation.

State-time estimates in Figure 5C made while ignoring three out of every four observations of soil water storage are definitely not reliable. A linear regression between estimated and measured values of *S* yielded a coefficient of determination r^2 of only 0.296 and about sixteen values omitted in the calculations fall outside of the confidence interval. There are two primary reasons why the state-time estimates illustrated in Figure 5C do not agree with reality. First, during a time period of 56 days (4 lags and nearly equal to 2 months), values of soil water storage are no longer temporally related to each other during the 2-year experiment (Table 1) – a requirement of state-time analyses. Second, the amounts of water that eventually drained from the 1-m soil profile from large rainfalls robustly accounted for in the state variable *P* occurring within time spans of 56 days could not be ignored. Note in Figures 4B, 5A, B and C,

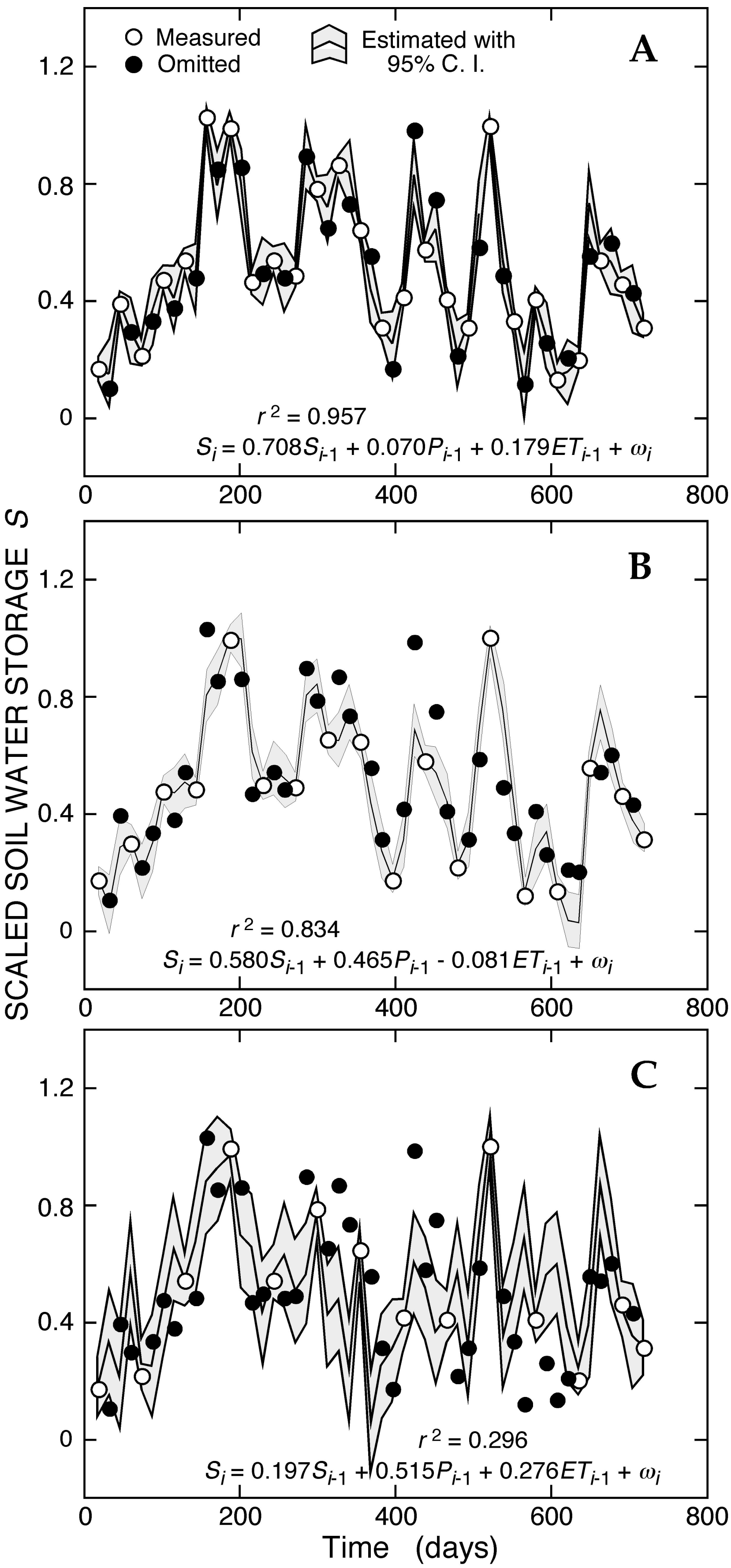

Figure 5 - Soil water storage measured biweekly for 714 days estimated from measurements of precipitation and evapotranspiration with A. one-half, B. two-thirds and C. three-fourths of the soil water storage observations omitted from the state-time analysis.

as the relative number of ignored observations of soil water storage increased, the magnitude of the transition coefficient of $S_{i\text{-}1}$ decreases with estimates of S_i depending progressively on the values of $P_{i\text{-}1}$. In other words, with fewer and fewer temporal observations of S_i available, reliable estimates of S_i depend more and more on the temporal association between soil water storage and precipitation. This dependence is entirely reasonable inasmuch as changes in soil water storage are generally related directly to amounts of precipitation infiltrating the soil surface during relatively short time periods. Notice, however, that no such consistent trend was manifested during these short time periods by the transition coefficients of $ET_{i\text{-}1}$. This fluctuation is reasonable inasmuch as changing local weather conditions can easily cause major shifts in evapotranspiration that do not impose major changes in average soil water storage. We verify the previous statement by examining Figure 6 where the mean values of evapotranspiration throughout the time being investigated are related by simple linear regression to the average amount of water stored in the soil profile. This figure indicates that measurements of evapotranspiration at any given time are not realistically estimated by the amount of water stored within the root zone of the soil profile of the coffee crop. Yet, soil water storage is generally sparingly and inadequately monitored in agricultural fields to assure that there is sufficient water within the root zone for the crop to sustain an adequate transpiration rate for optimal growth and harvestable yield.

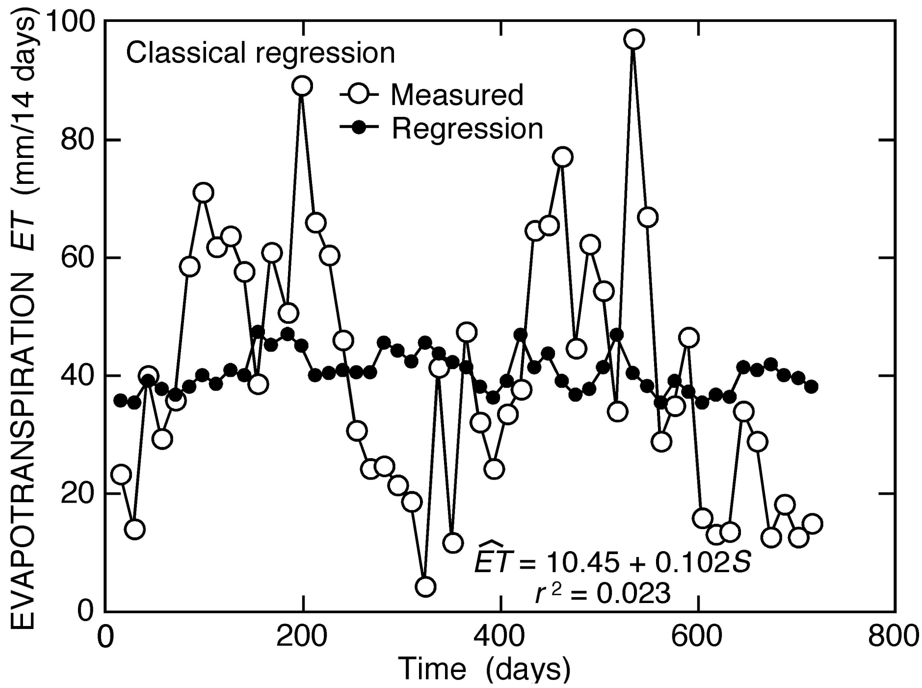

Figure 6 - Evapotranspiration measured biweekly for 714 days estimated using classical linear regression.

Rather than tediously and repetitively measuring the water stored in the soil profile to ascertain evapotranspiration across an agricultural field or even at a location designated as that representing the mean (access tube number 3 according to Fig. 3), a common practice has been the measurement of water lost from a Class A evaporation pan (Allen *et al* 1998). Such a procedure is convenient and inexpensive, but does not necessarily relate to quantitative measures of soil water storage at positions related to mean values for the field, or vice versa.

With measurements of mean values of *ET*, *S,* and *P* laboriously made biweekly in this study, we are able to examine the estimation of *ET* made by classical multiple regression and state-time analyses. Estimations of *ET* using classical multiple regression based on mean values of each variable throughout the time being investigated can be compared with measured values in Figure 7A. We note that no more than 13.5% of the variance of the biweekly-measured evapotranspiration data was explained from the measurements of soil water storage and precipitation. We also note that variations of *ET* with a coefficient of 0.415 were more related to fluctuations of precipitation than those of soil water storage with a coefficient of only 0.095. A similar relationship was also apparent in the state-time analysis presented in Figure 7B where the transition coefficient of *S* was only 0.090 while that of *P* was larger having a value of 0.310. Estimated values of *ET* from the state-time analysis approached those of the measured values, and manifested a coefficient of determination of 0.887. Nevertheless, 8 of the 51 estimated values of *ET* fell outside the 95% confidence interval.

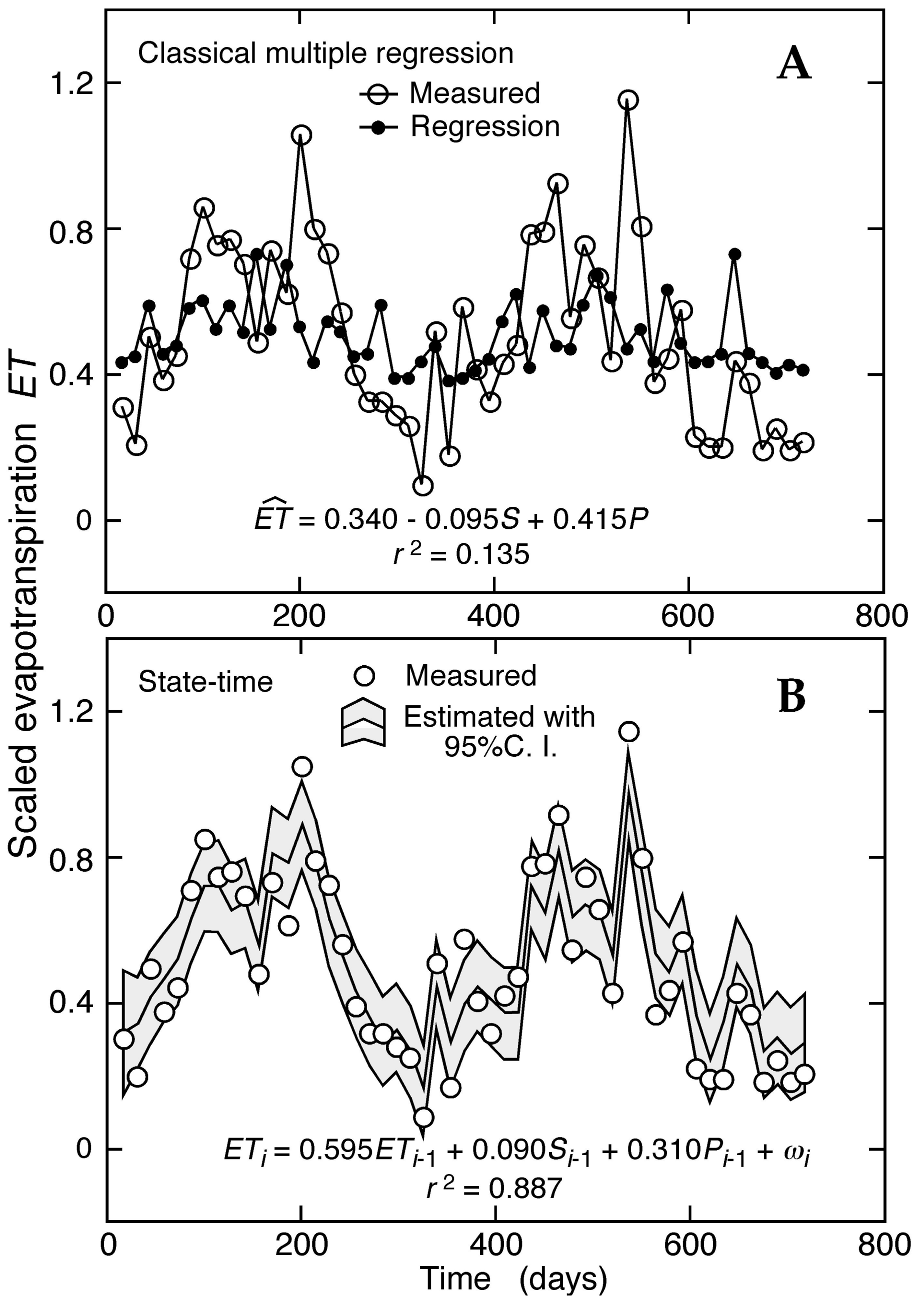

Figure 7 - Estimates of evapotranspiration measured biweekly for 714 days using A. classical multiple regression and B. state-time analysis.

In Figure 8 where the evapotranspiration is estimated with all measurements of *S* and *P*, but with one-half and three-fourths of its biweekly measurements omitted from the state-time calculations, the coefficient of determination decreases to 0.719 and 0.544. Nine of the 51 estimated values of *ET* in Figure 8A and 21 of the 51 estimated values of *ET* fell outside the 95% confidence interval.

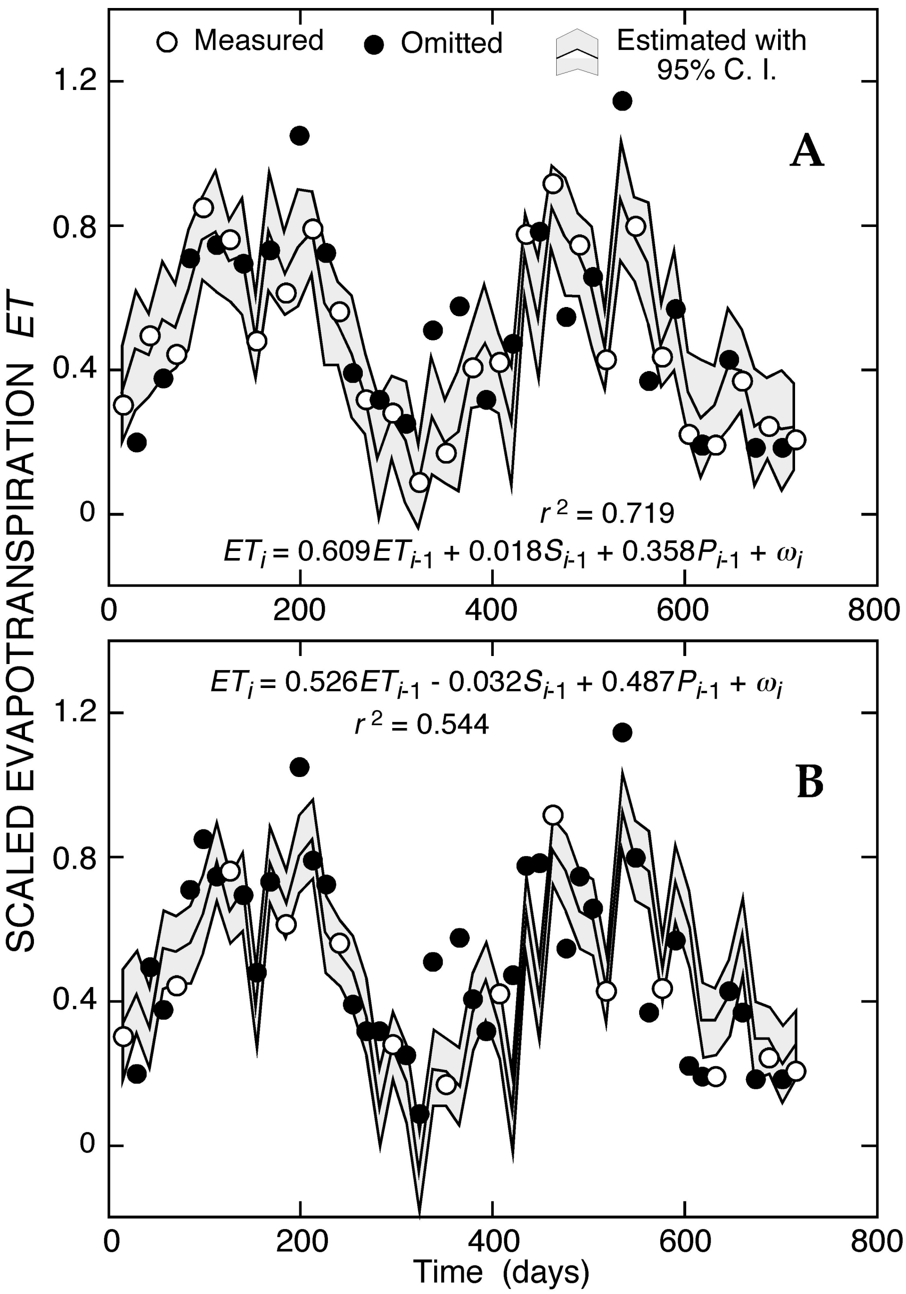

Figure 8 - Evapotranspiration measured biweekly for 714 days estimated from measurements of soil water storage and precipitation with A. one-half and B. three-fourths of the evapotranspiration observations omitted from the state-time analysis.

Noting that the contribution from neighboring values of *S* decreases from 9% in Figure 7B to a mere 2 and 3% in Figures 8A and B, respectively, we learn that for the case of this data set from a coffee crop, the temporal variations in *ET* are not physically caused by variations of *S*. Therefore, we examine the relationships between the two state variables *ET* and *P* in Figure 9.

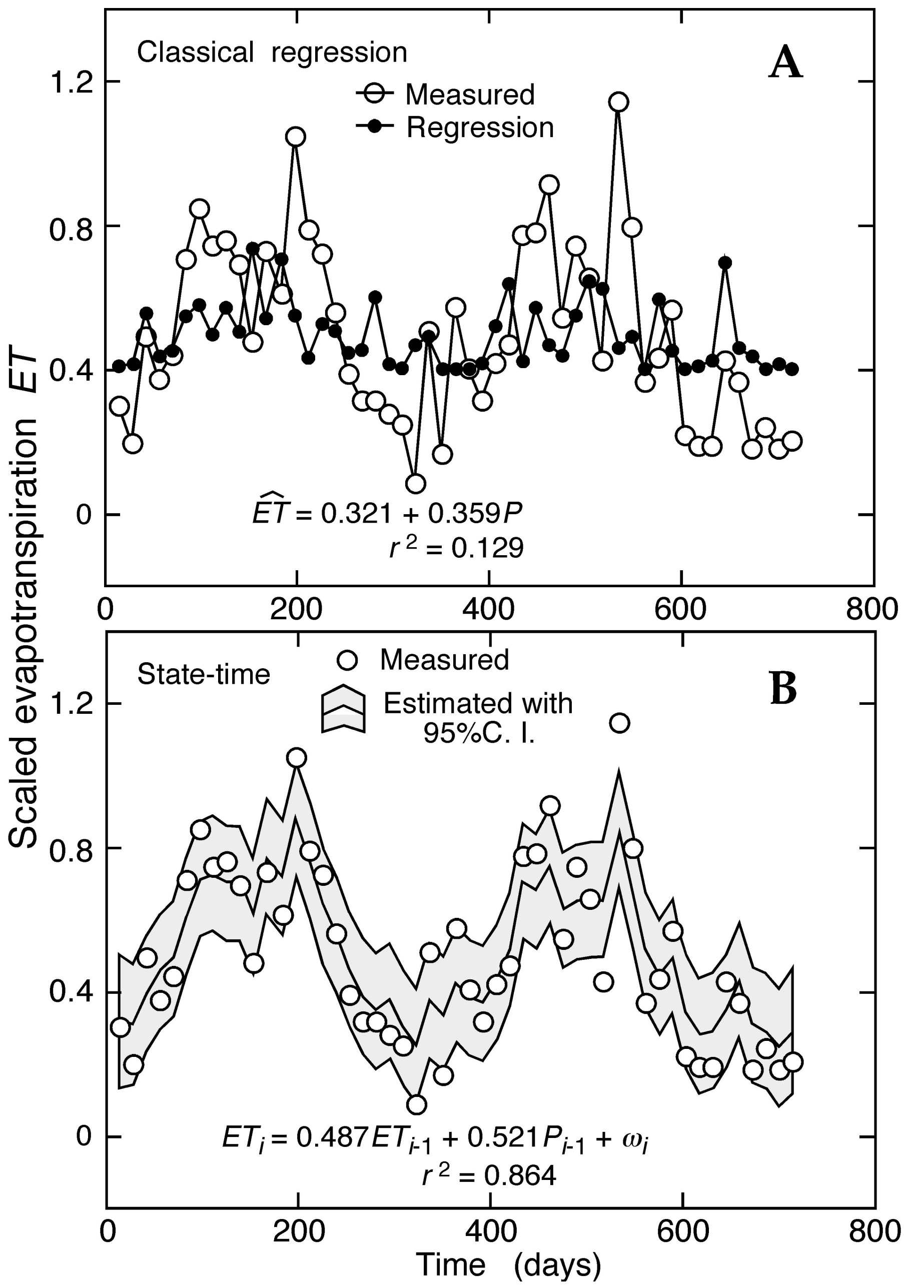

Figure 9 - Estimates of evapotranspiration measured biweekly for 714 days using A. classical multiple regression and B. state-time analysis.

Classical regression between ET and P throughout the time of the investigation yielded a coefficient of determination of only 0.129 (Figure 9A). On the other hand, state-time estimates were much more reliable with a coefficient of determination of 0.864 (Figure 9B). We expected that the state-time analysis would be superior because ET and P are significantly cross correlated to 3 temporal lags and ET has an autocorrelation length of 3 lags. We note that each preceding value of both state variables more or less equally contribute to the estimated value of ET. By omitting one out of two values of measured ET (Figure 10A) and three out of four values of measured ET (Figure 10B) in the state-time analyses, we learn that the coefficient of determination between estimated and measured values of ET reduces from 0.864 (Figure 9B) to 0.694 and 0.554 (Figures 10A and B, respectively). Without having neighboring values of ET for the updating procedure in the calculation, the contribution from the neighboring cross correlated measured P is inadequate to capture estimates of ET within an ever-increasing confidence interval. In other words, state variables physically linked to the cause of ET fluctuations were not monitored.

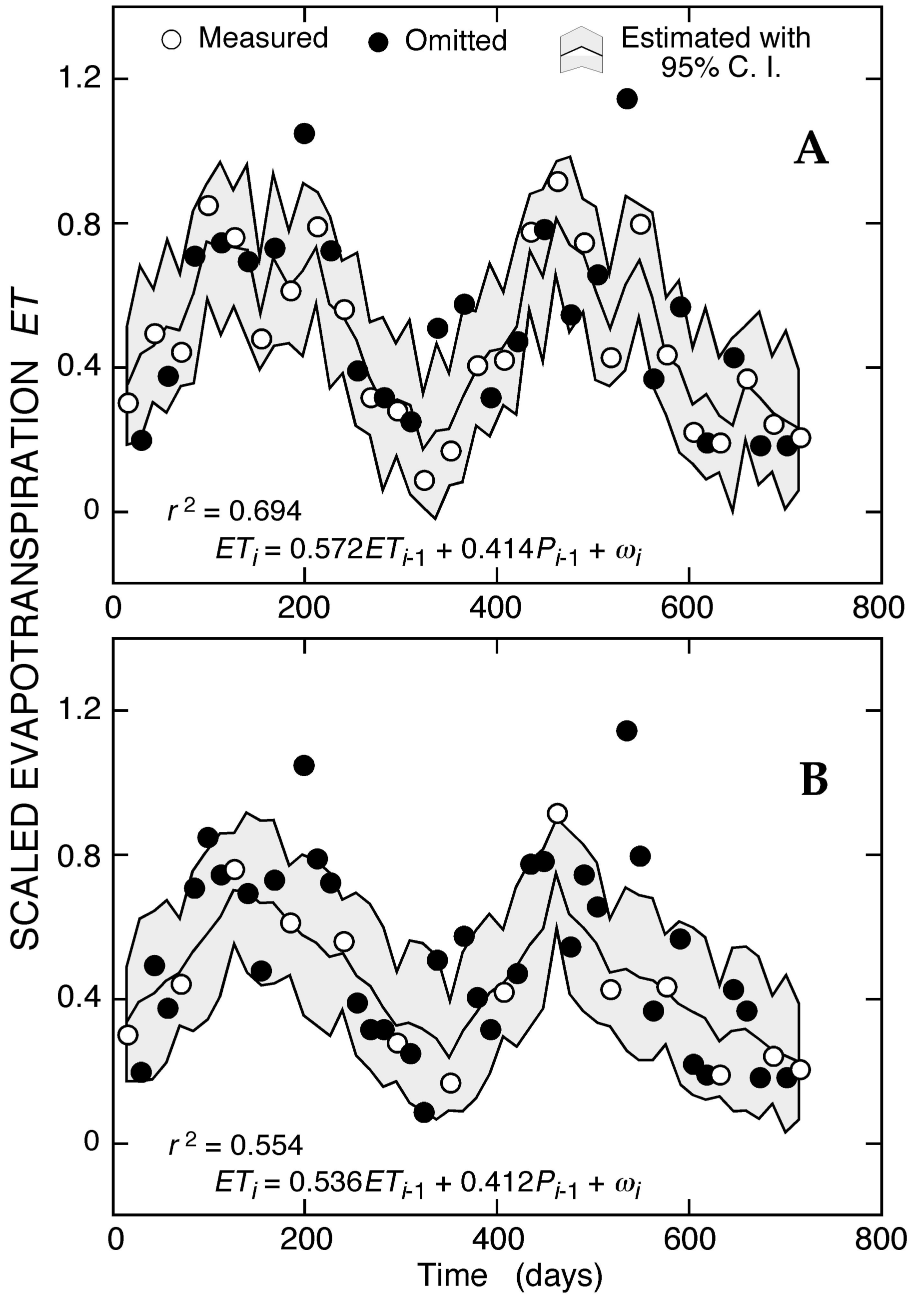

Figure 10 - Evapotranspiration measured biweekly for 714 days estimated from precipitation measurements with A. <u>one-half</u> and B. <u>three-fourths</u> of the evapotranspiration observations <u>omitted</u> from the state-time analysis.

During 14-day intervals, what physical processes in addition to precipitation alter the amount of water transpired from the crop and evaporated from the soil surface? From the above information thus provided rainfall (and to a very limited extent, soil water storage) is the only parameter that accounts for some of the 14-day variability of *ET* throughout each year as illustrated in Figure 11. Patterns of *ET* for both years are very similar, and indeed have similar spectra yielding significant coherence at several temporal frequencies not presented here. In order to identify the cause of this similarity as well as to improve estimates of evapotranspiration as a function of time, it would be necessary to measure at least one other variable or parameter physically responsible or linked to evapotranspiration – e. g., air temperature, relative humidity, cloudiness, wind velocity, soil temperature, distribution of water within the soil profile, vegetative and productive stage of the crop, insect damage, plant diseases, plant nutrient availability, mainly (Penman 1963; Allen *et al.* 1998).

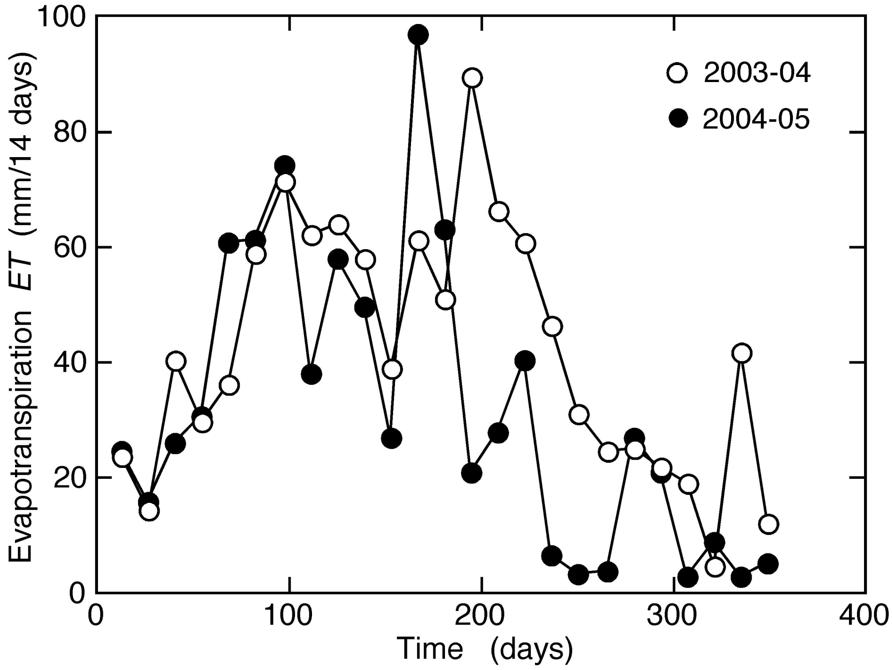

Figure 11 - Biweekly measurements of evapotranspiration during 2003-04 and 2004-05 versus time commencing the first week of September.

Previous and present outlook

As mentioned in the introduction, the estimation of soil water storage is a difficult task due to the spatial and temporal variability of field soils and their local environment. This presentation focused on the characterization of the average amount of water stored in the topsoil across a specific field measured at time intervals of 14 days. Because the majority of coffee plant roots were limited to a depth of 1 m within the soil profile, soil water designated as that available to the coffee crop was calculated from soil water content measurements from the soil surface to 1 m deep. The field was irrigated only when it was deemed necessary, i.e., whenever the stored water in the profile reached less than 20% of its full capacity. This irrigation strategy, embracing the concept that the spatial variation of *S* was invariant in time, allowed the analysis of the distribution of soil water storage measurements within the coffee field to ascertain a unique location consistently manifesting the mean soil water storage regardless of its time of measurement. And, the minimum number of locations sampled to achieve an average value within prescribed level of significance was based on the assumption that the sampled values were normally distributed. This strategy has been suggested during the past 25 years. Various other closely related strategies that include the measurement of a threshold minimum soil water storage, a specified integrated matric potential within the root zone of a plant and minimum soil water content or matric potential at a specified position within the root zone were explored and adopted since 1950 (Nielsen and Kutílek 1994). These strategies ignored the spatial distances between sampling locations, and also ignored the temporal correlations between successive soil water storage sampling campaigns. All of them sought and relied on the ability to find a "good average" to determine when to irrigate a crop. Few strategies were developed to ascertain how seldom measurements could be taken to ascertain when to irrigate for optimal crop production. As a result, published literature will testify that excessive energy and time were spent determining when to irrigate rather than to determine the relative benefit of having irrigated. Hence, the second half of this presentation focused on the temporal association of *S*, *P,* and *ET* at a fixed, hypothetical location assumed to represent the entire field during 14-day time intervals empirically selected for their measurement. The results of this tedious, and time- and energy-consuming sampling program indicate that the amount of water stored in the soil profile during the empirically designated 14-day interval sampling program has little to do with temporal variations of evapotranspiration. However, from locations sampled across the coffee field, it was apparent that the mean values of *ET* during the 14-day sampling intervals are temporally related to *P*, not *S*, and not quantitatively related to infrequent irrigations including those made in September.

After completing this experiment, we are left with the question, "When and where do we take what kind of other measurements to better manage the production of coffee as well as gain information on the environmental impact on its production?" During the past two decades, the concept of site-specific farming, precision farming or precision agriculture has emerged that emphasizes that the quality and quantity of crop production can be improved by simultaneously managing the temporal and spatial variations of crop-dependent processes across an agricultural field during crop growth. In other words, an agricultural field planted to one crop is not considered a unit to be managed or treated uniformly. Instead, based on its local soil and environmental properties, and the nature of physical and biological processes, it is managed as an ensemble of distinct spatial domains each monitored over appropriate scales of space and time. Many methods of statistical analysis (geostatistics, regionalized variable analysis, applied time series, etc.) are available for examining experimental data observed at different points in time and space relative to describing and understanding soil-plant-atmospheric processes within a farmer's field.

In this presentation, we illustrated the utility of state-time analysis to examine the temporal variation of the crop-dependent process of *ET*. We analyzed *ET* considering it to be a random variable and statistically treated its temporal variation as a function of the time between repetitive observations within a 2-year domain. At any given time, its value was considered to be identical at every location within the experimental area. Although such a consideration is not realistic because *ET* actually varies from one location to the next throughout the entire spatial domain, it is consistent with the common practice of irrigating a field with a given amount of water or also assuming that the rain measured at a specific location falls uniformly across the field.

Having briefly illustrated the utility of state-time analysis in this simple experiment to examine the temporal variation of the crop-dependent process of *ET* within a field, it is obvious that many related choices for meaningful field research remain open for immediate application.

One such choice taken by several researchers in the past was to repetitiously make measurements of *S*, *ET* and *P* at the same spatial interval across the experimental area for at least one time. The benefit of state-space analysis to examine the spatial processes of these crop-dependent variables at the time of their measurement should be realized by considering each of them to be a random variable treated statistically with their spatial association and variation being a function of the distance between their measurements. A spatial soil process is the change of a variable or a vector consisting of several variables across a spatial domain caused by localized conditions e.g., the spatial process of soil water storage considered across a field can be mainly influenced by spatial changes in soil type, topography, vegetation, rainfall, evapotranspiration and management.

Obtaining measurements of *S*, *ET,* and *P* made repetitiously across an experimental area at variable spatial intervals for numerous times as presented here provides another choice. Using two-dimensional state analysis in both time and space, a complete analysis of the progression of any or all of the three variables occurring at any location in the field at any time during the 2-year experiment would be highly informative. In other words, the analysis would provide "site-specific" and "time-specific" management information without the disadvantage of considering average values across the field or during each year.

Still more choices could be realized when measurements of coffee plant parameters – those of locally available soil nutrient and micro-environmental conditions related to potential coffee bean yields – are repetitiously and frequently made across the field during each growing season. With this information, a two-dimensional state analysis provides quantitative guidelines during the growing season to better manage the crop within specific local field domains to achieve higher yields without a deleterious impact on soil and water resources. As a result, management of the field would be more efficient and sustainable.

Conclusions

Purposely following the most commonly used classical procedure of randomization to identify sampling locations within a field of small replicated plots, we compared the results of two analyses: classical statistics and one application of applied time series (state-time analysis) to examine the temporal variability of soil water storage in a coffee field.

Classical statistical procedures indicated that randomly spaced estimates of S averaged across the field can be obtained with a deviation of 2% of the mean using only 4 out of the 15 sampled locations. Time stability analysis of S showed that one single specified location would represent the average value of S in relation to the average of the 15 locations, and if a standard deviation is required, 4 specific locations would yield an average with a deviation of only 0.3%.

In contrast to classical multiple regression analysis, the state-time analysis showed that S_i was more dependent on P_{i-1} (52%) than on ET_{i-1} (28%) and S_{i-1} (20%), indicating the low temporal dependence of S in relation to previous measurements. Additionally, the analysis showed that ET_i was not realistically estimated from S_{i-1} measurements inasmuch as it was more dependent on previous estimations ET_{i-1} (59%), than on P_{i-1} (30%) and S_{i-1} (9%). With P and ET easily obtained from automated weather stations, the state-time analysis indicated that S measurements made every 14 days could be reduced to monthly measurements, and that S_i measurements would still be predicted with an r^2 of 0.957 – significantly reducing future field work.

Presently, we as well as others with whom we communicate are conducting field experiments in which measurements of S are being taken at regular intervals in two spatial directions across a cultivated field at specified times that allow a 3-dimensional space-state-time analysis. These experiments should provide improved management without depending on traditional randomly treated small plots supposedly applicable to an entire field without any sort of experimental verification.

References **Professor, as tres referencias em vermelho não estão no texto. A referencia do trabalho citado é:**

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