http://www.ciheam.org/
http://om.ciheam.org/

Pasquale STEDUTO

Ecophysiology Lab
CIHEAM-IAM Bari
VALENZANO (BA)
ITALY

DETERMINATION METHODS of CROP WATER CONSUMPTION

INTRODUCTION

Estimation of water consumption of crops is of fundamental... planning and management... water re... system design, from irrigation scheduling, and environmental assessment.

Strictly speaking, the term *water consumption* of crops refers to... water transpired by the plant (i.e., the evaporation of water passing through the plant interior and undergoing some physiological control), the water evaporated from the soil and the water used by the plant for its metabolic processes. Since evaporation from the soil (E) and transpiration by the plant (T) occur simultaneously in nature, the term *evapotranspiration* (ET) is used to describe the total loss of water vapor from vegetated land surfaces to the atmosphere. Furthermore, since the water used for the plant metabolism is substantially negligible as compared to E...

...evapotranspiration process involves a phase change of water from liquid to gaseous state, with latent heat requirements of about 2.47 MJ per kg of water evaporated, and is one of the major components of the hydrological cycle. A quite general water balance equation for agricultural systems can be written as

$$P + I = ET + R + D \pm \Delta S \tag{1}$$

where P and I (usually the inputs) are precipitation... R and D (usually the outputs) are evapotranspiration, runoff and deep percolation, respectively, and ΔS is changes in soil water storage. All the terms in Eq. (1) are expressed in rates (amount over a given time).

Where runoff (R) and deep percolation (D) can be controlled, Eq. (1) reduces to

$$P + I = ET \pm \Delta S \tag{2}$$

The last simplified equation, changed into the following form

$$ET = P + I \pm \Delta S \tag{3}$$

can be used for direct estimate of ET if other terms are known.

The purpose of this article is to review different methods, direct or indirect, used for estimation of ET and illustrate their major features, indicating main criteria for their choice and applicability.

INTRODUCTION

Estimating the water consumption of crops is relevant to address issues such as regional planning and management of water resources, prediction of crop productivity, irrigation system design, farm irrigation scheduling, and environmental assessment.

Strictly speaking, the term *water consumption* of crops refers to the water transpired by the plant (i.e., the evaporation of water passing through the plant interior and undergoing some physiological control), the water evaporated from the soil and the water used by the plant for its metabolic processes. Since evaporation from the soil (E) and transpiration by the plant (T) occur simultaneously in nature, the term *evapotranspiration* (ET) is used to describe the total loss of water vapor from vegetated land surfaces to the atmosphere. Furthermore, since the water used for the plant metabolism is substantially negligible as compared to E and T, the term *water consumption* is frequently alternative to *evapotranspiration* (ET).

The evapotranspiration process involves a phase change of water from liquid to gaseous state, with latent heat requirements of about 2.47 MJ per kg of water evaporated, and is one of the major components of the hydrological cycle. A quite general water balance equation for agricultural systems can be written as

$$P + I = ET + R + D \pm \Delta S \qquad (1)$$

where P and I (usually the inputs) are precipitation and irrigation, respectively, while ET, R and D (usually the outputs) are evapotranspiration, runoff and deep percolation, respectively, and ΔS is changes in soil water storage. All the terms in Eq. (1) are expressed in rates (amount over a given time).

Where runoff (R) and deep percolation (D) can be controlled, Eq. (1) reduces to

$$P + I = ET \pm \Delta S \qquad (2).$$

The last simplified equation, changed into the following form

$$ET = P + I \pm \Delta S \qquad (3)$$

can be used for direct estimate of ET if other terms are known.

The purpose of this article is to review different methods, direct or indirect, used for estimation of ET and illustrate their major features, indicating main criteria for their choice and applicability.

TWO SIDES IN TACKLING THE PROBLEM

In principle, the problem of determining crop evapotranspiration can be tackled from two sides of the crop stand: the soil (below ground) or the atmosphere (above ground). Although the crop itself is another portion of the soil-plant-atmosphere system to which other methods can be applied to determine T or ET, these are quite close to those used for the above ground side.

Identifying similarities and differences between these two sides, where transfer and storage processes of water take place, are quite instructive in understanding the basic principles behind the different methods which are going to be presented.

The continuity equation, based on the conservation of mass and energy, represents the starting point for analyzing the two sides.

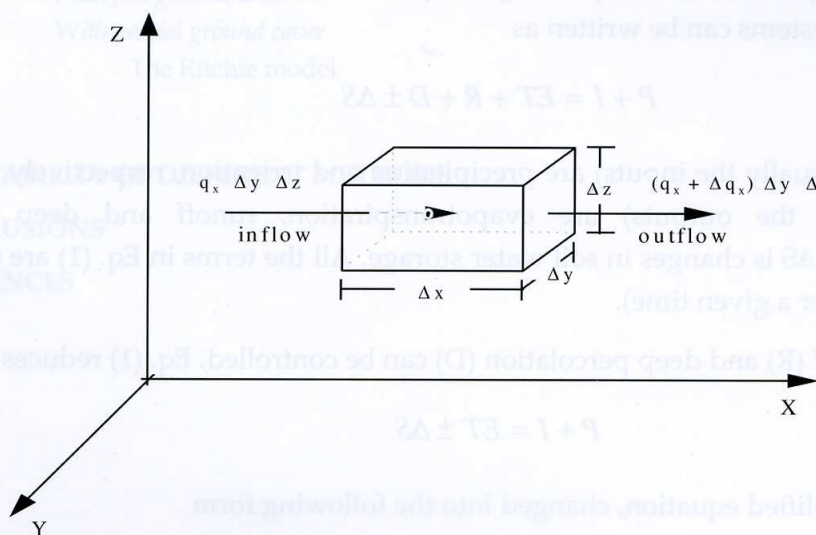Looking at the soil side, let us consider a volume element as in Fig. 1. Assume a flux of

Fig. 1 - *Change in water storage of a soil volume element. Application of the continuity principle (after Hillel, 1980).*

water (q_x in cm s^{-1}) enters the volume element along the x direction and a change in the flux ($q_x+\Delta q_x$) is observed at the exit from the element. The change in water storage inside the volume element is given by

$$\text{inflow} - \text{outflow} = q_x \cdot \Delta y \cdot \Delta z - (q_x + \Delta q_x)\Delta y \cdot \Delta z \qquad (4)$$

with q_x being a Darcy's flux

$$q_x = -k_{(\phi_m)} \frac{\partial \Phi}{\partial x} \tag{5},$$

where Φ is the matrix water potential and $k_{(\phi m)}$ the hydraulic conductivity which varies as a function of soil matrix potential

The differential principle indicates that for small variation of q_x it can be assumed

$$\frac{\Delta q_x}{\Delta x} \cong \frac{\partial q_x}{\partial x} \quad \Rightarrow \quad \Delta q_x \cong \frac{\partial q_x}{\partial x} \Delta x \tag{6}.$$

Then, using Eq. (6) in Eq. (4) yields

$$\text{inflow - outflow} = q_x \cdot \Delta y \cdot \Delta z - (q_x + \frac{\partial q_x}{\partial x} \Delta x) \Delta y \cdot \Delta z = - \frac{\partial q_x}{\partial x} \Delta x \cdot \Delta y \cdot \Delta z \tag{7}.$$

The net difference of fluxes between flux coming in and going out must correspond to the rate of gain in water by the volume element per unit of time (t). By expressing volumetric water content of the element with θ ($cm^3_{water}cm^{-3}_{soil}$), the continuity equation reduces to

$$\left(\frac{\partial \theta}{\partial t}\right) \Delta x \cdot \Delta y \cdot \Delta z = -\left(\frac{\partial q_x}{\partial x}\right) \Delta x \cdot \Delta y \cdot \Delta z \quad \Rightarrow \quad \frac{\partial \theta}{\partial t} = -\frac{\partial q_x}{\partial x} \tag{8}.$$

If the fluxes also in the other two directions y and z (q_y and q_z, respectively) are considered, the generalized continuity equation for soil can be written as

$$\frac{\partial \theta}{\partial t} = -\left(\frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} + \frac{\partial q_z}{\partial z}\right) = -\nabla q \tag{9}$$

where, in shorthand mathematical notation, the ∇ symbol designates the differential operator.

The left and the right sides of Eq. (9) are the *storage* and the *flux* terms, respectively. When these terms are quantified for soil systems, one can visualize that the storage capacity of one m^3 of soil volume can reach about 500 dm^3 of water, and that a storage change of the order of 7 dm^3 can occur in one day, while the hydraulic conductance can reach only a few cm hr^{-1}. In other words, the soil system is characterized by a large storage capacity as compared to flux terms so that ET in the water balance equation (1,3) is conveniently determined by looking at the changes in water storage rather than at the changes in fluxes.

Let's now look at the problem from the atmosphere side. A few preliminary considerations are needed before making the analysis. First of all, the entity being transported and the fluid acting as a carrier are no longer the same. In the soil, water acted both as carrier and entity being carried. In the atmosphere, water is the entity being carried while the carrier is air. Motion is dominantly laminar in the soil while it is turbulent in the atmosphere. Furthermore, water is moving as liquid in the soil and as gas in the atmosphere. It can be inferred that the storage capacity of one m³ volume of air, say at 25 °C, can reach the equivalent of about 0.023 dm³ of liquid water, while the aerodynamic conductance can easily reach 100,000 cm hr⁻¹. In other words, the atmosphere system is characterized with a small storage capacity as compared to flux terms so that ET in the water balance equation (1,3) is conventionally determined by using flux terms rather than changes in water storage. The two sides of the crop stand are schematically represented in Fig. 2, while the differences in boundary conditions between soil and atmosphere are summarized in Table 1.
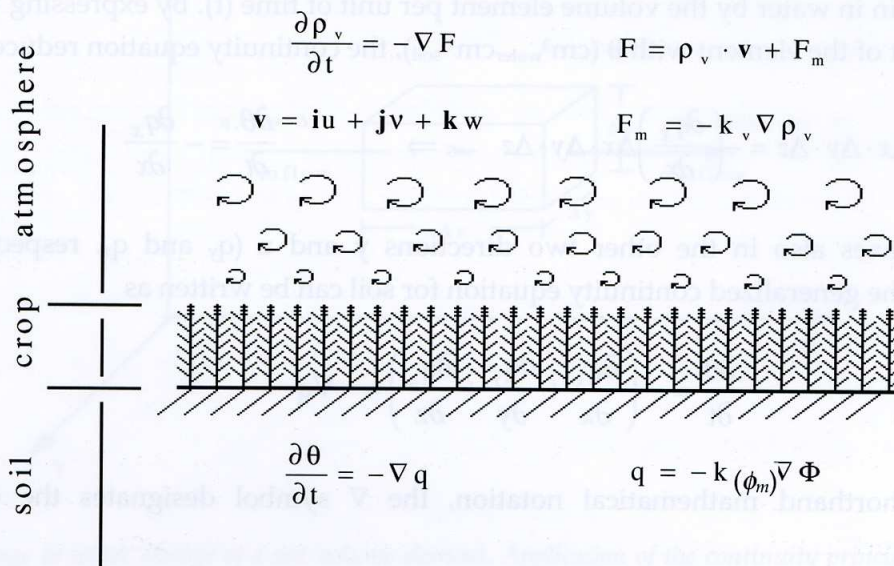
$$\frac{\partial \rho_v}{\partial t} = - \nabla F \qquad\qquad F = \rho_v \cdot v + F_m$$

$$v = iu + jv + kw \qquad\qquad F_m = - k_v \nabla \rho_v$$

$$\frac{\partial \theta}{\partial t} = - \nabla q \qquad\qquad q = - k_{(\phi_m)} \overline{\nabla} \Phi$$

Fig. 2 - *Schematic representation of the continuity equations within the boundary conditions of the soil side and the atmospheric side (see text for explanation).*

Tab. 1 - *Summary of the major characteristics differentiating the soil and the atmosphere systems.*

CHARACTERISTICS	SOIL	ATMOSPHERE
nature of motion	laminar	turbulent
Physical state of the fluid	liquid	gas
carrier fluid	water itself	air
conductivity	very low	very high
storage capacity	very high	very low
Convenience in balance approach	change in storage	change in flux

The continuity equation for water vapor in the atmosphere, in the absence of sinks or sources, is then

$$\frac{\partial \rho_v}{\partial t} = -\left(\frac{\partial F_x}{\partial x} + \frac{\partial F_y}{\partial y} + \frac{\partial F_z}{\partial z}\right) = -\nabla F \tag{10}$$

with ρ_v the water vapor density (g m⁻³) and F the total specific mass flux defined as

$$F = \rho_v \cdot v + F_m \tag{11},$$

where

$$v = iu + jv + kw \tag{12}$$

represents air velocity in unit vectors (**i,j,k**) and the corresponding velocity components in the x,y, and z directions (u,v,w), and

$$F_m = -k_v \nabla \rho_v \tag{13}$$

is the specific mass flux due to molecular diffusion (always in three-dimensional space), with k_v the molecular diffusivity of water vapor in the air. The flux F is somewhat analogous to solute transport in soil, where both diffusion and dispersion are taking place. Thus, water vapor in the atmosphere is transferred by convection with the fluid and through molecular diffusion, largely dominated, though, by convection. Then, knowing the velocity field **v** and the water vapor content at any given point in time and space it may be possible to solve the flux equation. This is possible only in a deterministic approach; however it cannot be accomplished in practice. It can, instead, be tackled statistically using means, deviation from means, and correlation between vectorial and scalar components. The conservation equation for water vapor in the atmosphere becomes then

$$\frac{\partial \overline{q}}{\partial t} + \overline{u}\frac{\partial \overline{q}}{\partial x} + \overline{v}\frac{\partial \overline{q}}{\partial y} + \overline{w}\frac{\partial \overline{q}}{\partial z} = -\left[\frac{\partial}{\partial x}(\overline{u'q'}) + \frac{\partial}{\partial y}(\overline{v'q'}) + \frac{\partial}{\partial z}(\overline{w'q'})\right] + k_v \nabla^2 \overline{q} \qquad (14)$$

where q represents the general flux term, valid not only for water vapor but also for any other scalar measurements (e.g., sensible heat, CO_2, etc.), overbar denotes time average, and the prime (') indicates the deviations around the mean value.

Differences in boundary conditions between soil and atmosphere (Table 1) and the key equations (9) and (14) delineate the major aspects and features in various methods and techniques, used to determine ET term of the water balance equations (1-3).

DETERMINING CROP EVAPOTRANSPIRATION

Measurement Methods

Many methods have been developed for direct and indirect measurements of crop evapotranspiration (ET) for different applications and different spatial and temporal scales. For a broader view of ET determinations, the reader is referred to Sharma (1985) and Itier (1996); here a brief review of methods is illustrated under three categories: water balance, micrometeorological and the physiological methods.

Water Balance

This method is based on previously introduced water balance equation (Eq. 1,3). Depending on the complexity and dynamics of the system, the water balance method applies to areas ranging from small plots (5-10 m²) to large catchments, over periods from weeks to a year. The detailed discussions on this method is covered in the last chapter.

Indirect. If ET is calculated with Eq. (1,3) by measuring or estimating other terms in the equations, this approach is known as "indirect" method. It is used in open fields and is also largely adopted in constant water table or drainage type lysimeters as well as with measurements of soil water content and fluxes. This last method is the main subject developed in the different chapters of this book.

Direct. As opposed to the indirect approach, this method makes use of weighing lysimeters. With these type of lysimeters, hydrologically isolating a soil-vegetation sample, each component of the water balance is measured precisely and ET is directly represented by the loss in lysimeter weight.

Over the years, various lysimeter design, construction, installation, and weighing systems have been developed. The reader is referred to FAO Paper No. 39 by Aboukhaled et al. (1982) for further details on lysimeters.

A properly designed and managed lysimeter can provide an accurate measure of actual ET over a wide range of time (minutes to months), and therefore it can be used as an independent check on micrometeorological and plant physiological methods, in calibrating empirical methods of calculating E, and in verifying ET models.

Micrometeorological

These methods allow evapotranspiration determinations from meteorological data (e.g., temperature, humidity, wind velocity, radiation) measured at and/or above the evaporating surface. In principle, the methods are aimed at estimating natural evapotranspiration with minimal disturbance to the microclimate. They make use of highly advanced instruments and sensors along with data logging and data processing systems, and may therefore sometimes be very costly. ET can be measured over periods of less than an hour to a day or a monthly basis. Most of these micrometeorological methods are suitable for research-oriented programs, while only a few can be used for routine measurements over extended periods of time. While the reader is referred to Kanemasu et al. (1979), and to Baldocchi et al. (1988) for a good understanding of the micrometeorological methods, here a brief description is reported on the major principles used.

Aerodynamic. According to the aerodynamic method, water vapor flux (E) is estimated through measurements of the humidity gradient according to

$$\lambda E = -\lambda \rho_a k_w \frac{\partial q}{\partial z} \tag{15}$$

where ρ_a is the air density, λ is the latent heat of vaporization, k_w is the turbulent transfer coefficient for water vapor, q is the specific humidity, and z is the height above the ground. Assuming a similarity between turbulent transfer coefficients of the different entities, one can use the turbulent transfer coefficient of momentum (k_m) in place of k_w, measuring wind profiles to estimate k_m (Kanemasu, 1979). However, inequalities between these coefficients have been observed due to modification in atmospheric stability, surface heterogeneity, or other causes. An empirical factor ϕ_w, related to atmospheric stability, can be introduced to account for the inequalities between k_m and k_w. Then, Eq. (15) practically becomes

$$\lambda E = -\lambda \rho_a k_m \frac{\Delta q}{\Delta z} \phi_w \tag{16}.$$

This method has been proved to be prone to errors in many conditions and shows to be satisfactory only if accurate determinations of q, wind velocity, k_m and ϕ_w are made.

Bowen-ratio/Energy-balance. The energy balance equation of a crop stand, neglecting minor terms, is expressed as

$$Rn = \lambda E + H + G \tag{17}$$

where Rn is net radiation, λE and H are latent and sensible heat, respectively, and G is the heat flux in the soil. Rn and G are generally measured making use of net radiometers and soil heat flux plates, respectively. Then, the available energy for the crop stand (Rn-G) is partitioned into λE and H. The Bowen ratio is defined as the ratio of sensible to latent heat and is expressed as

$$\beta = \frac{H}{\lambda E} = \frac{c_p}{\lambda} \frac{k_h}{k_w} \left(\frac{\Delta T}{\Delta z}\right) \left(\frac{\Delta q}{\Delta z}\right)^{-1} \tag{18}$$

Similar to the aerodynamic method, if k_h and k_w are identical, and if temperature and humidity gradients are measured over the same height interval (same Δz), then

$$\beta = \frac{H}{\lambda E} = \frac{c_p}{\lambda} \frac{\Delta T}{\Delta q} = \gamma \frac{\Delta T}{\Delta e} \tag{19}$$

where the new variable introduced just for convenience (e) is the actual vapor pressure of the atmosphere. Thus, substituting (19) in (17),

$$\lambda E = \frac{(Rn-G)}{(1+\beta)} = \frac{(Rn-G)}{\left(1+\gamma\frac{\Delta T}{\Delta e}\right)} \tag{20}$$

Contrary to the aerodynamic method, there is sufficient information indicating that $k_h=k_w$ holds under a wide range of conditions. Provided ΔT and Δe (or Δq) can be measured reliably and accurately, this method can properly be applied to field crops. Fig. 3 illustrates an example of a Bowen ratio apparatus.

Eddy correlation. The eddy correlation (or eddy covariance) method provides the most direct means of measuring the actual water vapor flux over the evaporating surface. Instantaneous measurements of vertical wind velocity (w) and humidity (q) are made. Following Eq. 14, the water vapor flux is then determined by

$$E = -\rho_a \overline{w'q'} \tag{21}$$

The summation of the product is usually carried over a period of 1/2 hr. Unlike the previous methods, it does not require equality of turbulent transfer coefficients ($k_m=k_w=k_h$). On the other side, the instrumentation required must be very sensitive and must respond rapidly to fluctuations in wind and humidity. Generally, wind is measured by sonic anemometer and humidity by infrared or ultraviolet hygrometry.
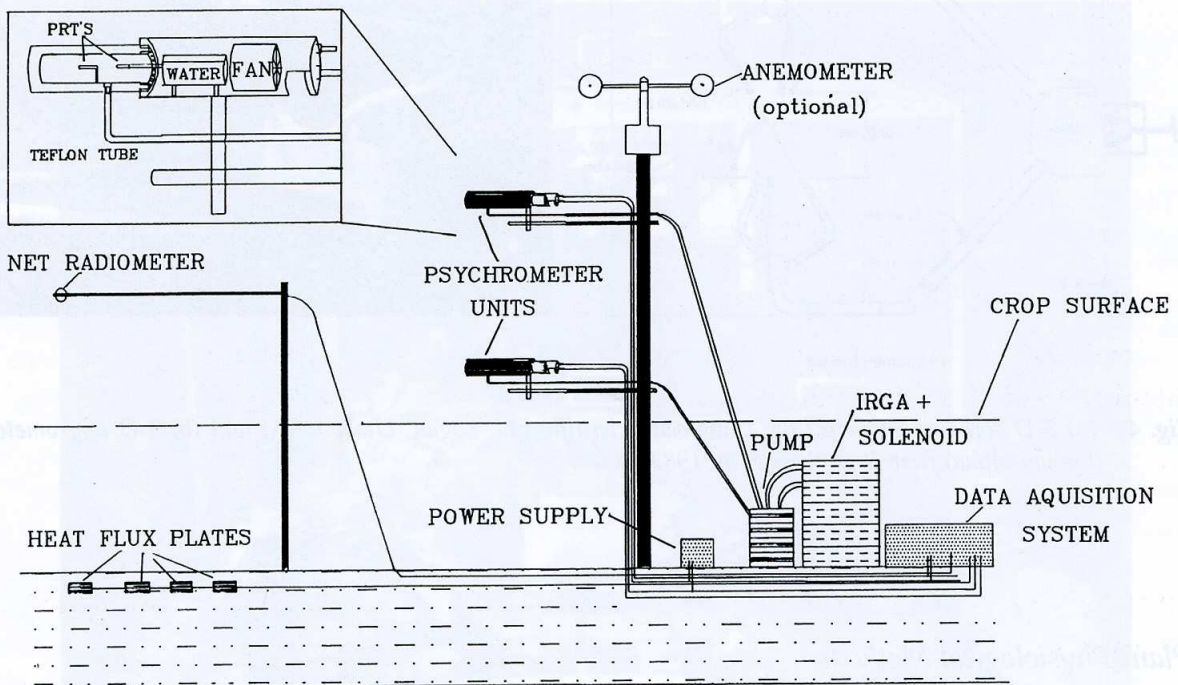

Fig. 3 - *Design and components of a Bowen-ratio/energy balance system (from Held et al., 1990).*

Alternatively, the eddy correlation can be used to determine sensible heat by

$$H = - \rho_a \overline{w'T'} \qquad (22)$$

and then, measuring also Rn and G, use the energy balance equation (17) to derive λE.

Despite several successful applications of this technique, it is not yet considered reliable and satisfactory for routine use with non-specialized people due to the complexity and sophistication of the instruments and the specialized knowledge required by the method. Thus, its use can be recommended in the research context only. However, the advance in technology would make it suitable for wider use in the near future. A schematic view of a

3-D sonic anemometer and 1-D hygrometer typically used with the eddy correlation method is shown in Fig. 4a and b.
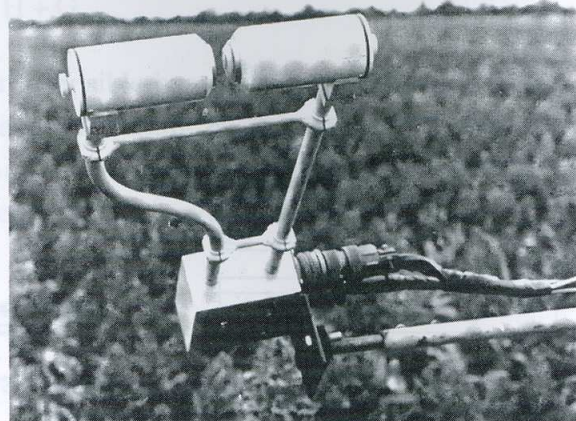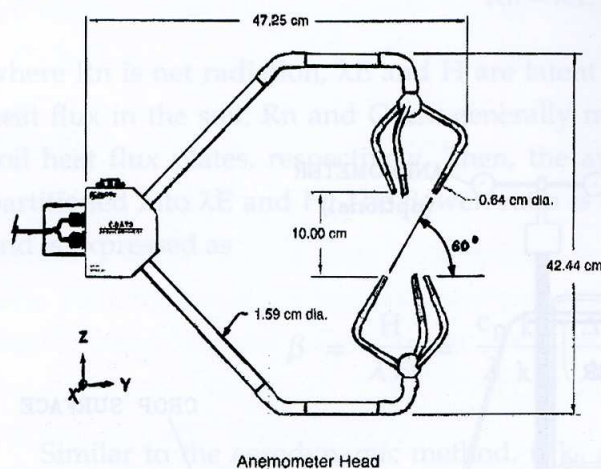

Fig. 4 - (a) *3-D sonic anemometer (by Campbell Scientific Ltd, Logan, Utah, USA) and* **(b)** *1-D hygrometer (Lyman-alpha) from Rosenberg et al. 1983.*

Plant Physiological Methods

Methods in this category have traditionally been developed by plant physiologists and are generally designed to estimate the transpiration components at the crop stand level. Generally, they may apply to trees, plants, parts of plant or small plots.

The major use of plant physiological methods is in acquiring a good understanding of soil-plant-water relations, in order to quantify the role of plant factors in evapotranspiration (e.g., role of stomata, of root density, etc.). Mostly, measurements are over short periods of time: from a few minutes to a daily basis. Major limitations of the physiological methods are that they significantly alter the environment during measurements, thus introducing errors, and there are problems in extrapolating results from leaf or plant scale to canopy and community scale.

<u>Chambers.</u> In this method, the measurements in humidity changes of the air passing through a chamber enclosing plants are taken. Design criteria have been reviewed by Slavik (1974) and evaluated by Pickering et al. (1993). The use of chambers should be restricted to measurements of comparative transpiration under the environmental conditions inside the chamber, and not of absolute transpiration. An example of canopy chamber is shown in Fig. 5.
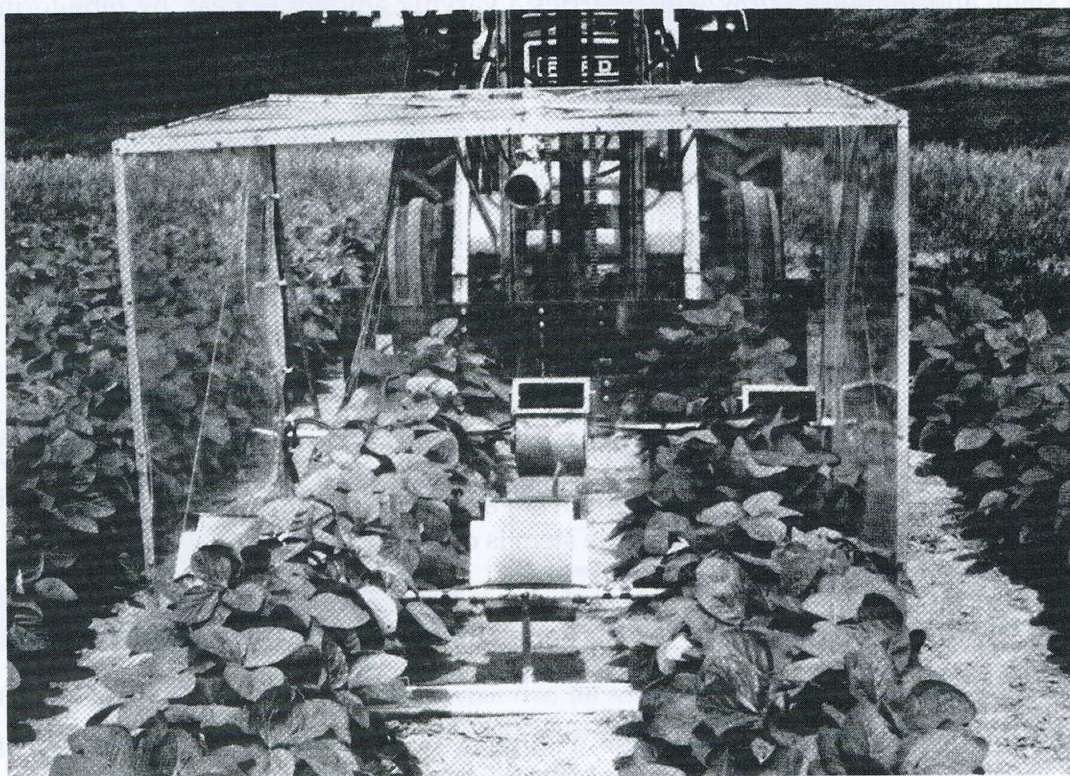
Fig. 5 - *Example of field chamber mounted on a tractor (from Reicosky and Peters, 1977, in Rosemberg et al., 1983).*

<u>Heat pulse and heat balance.</u> These methods apply to stems of trees and plants. The heat pulse technique, mostly used for trees, implies the application of a heat pulse in the stem via an intrusive probe and the measurement of the traveling time of the heat with the transpiring stream at a distance from the point of heating. Water flux is then calculated by solving the equation describing the transport of heat by convection and conduction in the stem, but it requires the knowledge of the dimension of the water-conducting tissue and the thermal properties of the stem. One of the major problems with this method, in fact, is in quantifying

the effective cross-sectional area of the stem involved in water conduction. Appropriate calibrations are then necessary. Additional problems are derived from the fact that the method is intrusive for the stem.

The heat balance technique differs from the heat pulse in that it does not require calibration and does not use intrusive probes. It is based on an energy balance applied to a constant heat source supplied through gauges wrapped around a plant stem. The gauges are precision instruments that sense milliwatt power transfer from a heater strip to the ambient, to the stem, and to the sap flow which carries a varying amount of heat. By monitoring (a) the temperature at two points above and below the heater to measure the stem heat transfer (upward and downward) and the temperature adjacent to the heater and on the outside of the insulation of the gauge to measure the heat lost to the ambient, and (b) through knowing the constant heat supplied to the gauge, the heat carried by the sap flow of the stem can be calculated with the energy balance equation

$$Qf = Pin - Qr - Qv \qquad\qquad (23)$$

where Qf is the sap flow heat transfer, Pin is the power supplied to the gauge, Qr is the radial heat transfer of the gauge insulator, and Qv is the vertical stem heat transfer (compounded by the upward Qu, and downward Qd directions). Dividing Qf by the temperature increase of the sap flow after the gauge heating (dT), and by the heat capacity of water (c_w), the water flow (Fw) through the stem is obtained

$$Fw = \frac{Qf}{c_w\,dT} \qquad\qquad (24).$$

Since the heat supplied to the gauge wrapping the stem is low, preventing damage to the phloem functionality, it will have serious restrictions to be used for large stems. An examples for setting of the gauges, schematic diagram of instrument parts and results from the heat method are reported in Fig. 6 a, b, and c, respectively. Additional information on the method can be found in Baker and van Bavel (1987), Baker and Nieber (1989), Steinberg et al. (1989), Dugas (1990), and Ham et al. (1991).

Modeling Methods

Many models have been developed to estimate ET. Most of those which could consider water stress conditions of the crop are not yet reliably used. Therefore, here only the models with non-limiting water conditions are reported. Two major distinctions are made, regarding crop ground coverage, full or partial coverage.
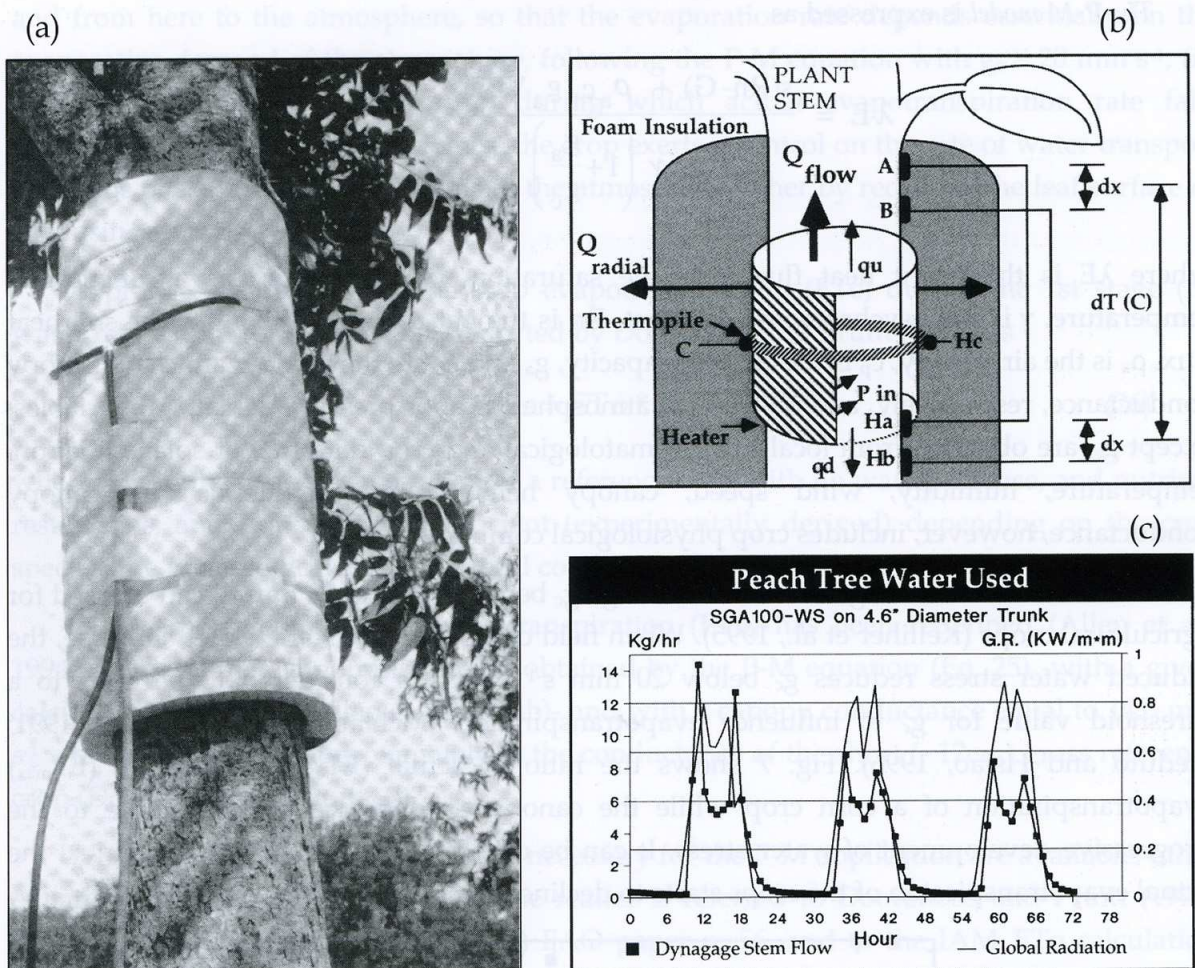
(a)

(b)

(c)

Fig. 6 - (a) *Example of stem-flow gauge on a pecan tree,* **(b)** *schematic heat flow diagram, and* **(c)** *three days of monitoring of incident solar radiation and sap flow of a peach tree (by Dynamax Inc., Houston, Texas, USA).*

With full ground cover

Under a given weather condition, a crop stand with full ground coverage of the ground and not suffering from water shortage, transpires largely as it intercepts the whole incident solar radiation and therefore very minor energy is left available for the soil surface to sustain evaporation process, implying that soil evaporation (E) is generally negligible as compared to plant transpiration (T). Under such conditions, the "big-leaf" model of Monteith (1973) - also indicated as the Penman-Monteith (P-M) model- is well suited to estimate evapotranspiration of field crops, provided the conductance to water vapor of the crop canopy is known.

The P-M model is expressed as

$$\lambda E = \frac{s(Rn-G) + \rho_a c_p g_a \, VPD}{s + \gamma \left(1 + \dfrac{g_a}{g_c}\right)} \qquad (25)$$

where λE is the latent heat flux, s is the saturation vapor pressure as a function of temperature, γ is the psychrometric constant, Rn is the net radiation flux, G is the soil heat flux, ρ_a is the air density, c_p is the air heat capacity, g_a and g_c are the aerodynamic and canopy conductance, respectively, and VPD is the atmospheric vapor pressure deficit. All variables, except g_c, are obtained from local microclimatological and crop data (net or solar radiation, temperature, humidity, wind speed, canopy height and roughness). The canopy conductance, however, includes crop physiological components.

When no water shortage occurs, an average g_c between 20 and 33 mm s⁻¹ can be used for agricultural crops (Kelliher et al., 1995). When field crops undergo some water shortage, the induced water stress reduces g_c below 20 mm s⁻¹ which is considered to be close to a threshold value for g_c to influence evapotranspiration (McNaughton and Jarvis, 1991; Steduto and Hsiao, 1996). Fig. 7 shows the ratio of actual (ET$_a$) to maximum (ET$_{max}$) evapotranspiration of a corn crop while the canopy conductance decreases due to the progressive development of water stress. It can be observed that below $g_c \approx 20$ mm s⁻¹ the actual evapotranspiraton of the crops starts to decline.
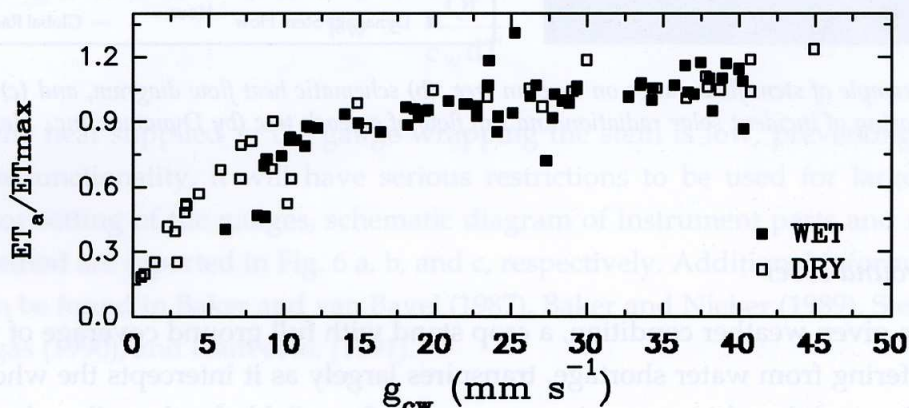

Fig. 7 - *Daily ET$_a$/ET$_{max}$ versus midday canopy conductance (g$_{cw}$) for the irrigated (WET) and non irrigated (DRY) maize crop (from Steduto and Hsiao, 1998).*

Assuming constant weather conditions over the crop stand, also in the transpiration process one can identify, by analogy with soil drying, at least two stages: **(i)** the initial *constant-rate* stage, when the crop does not show any restriction (i.e., shows the minimum resistance) to the supply of water from the soil to the site of evaporation (the leaf epidermis),

and from here to the atmosphere, so that the evaporation rate depends essentially on the evaporative demand of the atmosphere, following the P-M equation with $g_c \geq 20$ mm s^{-1}; (ii) the intermediate falling-rate stage, during which actual evapotranspiration rate falls progressively below potential because the crop exerts a control on the rate of water transport (in vapor phase) from the epidermis to the atmosphere either by reducing the leaf surface or by closing stomata.

A practical estimation of the crop evapotranspiration (ETc) during the 1st stage (no water restriction) remains the one reported by Doorenbos and Pruitt (1977) as

$$ET c = ET o * K c \qquad (26)$$

where ETo is the evapotranspiration of a reference crop with no water, disease, and nutrient restrictions, and Kc is a crop coefficient (experimentally derived) depending on the crop species, the phenological stage and local conditions.

Lately, the reference crop evapotranspiration (ETo) has been redefined (Allen et al., 1994a, 1998) as the evapotranspiration obtained by the P-M equation (Eq. 25) -with a given calculation procedure (Allen et al., 1994b)- and with a canopy conductance equal to 14.5 mm s^{-1} which corresponds approximately to the conductance of the short (~12cm) grass reference crop (Doorenbos and Pruitt, 1977).

When not all the weather variables necessary for the P-M application are available, other equation estimates can be used and the reader is referred to Doorenbos and Pruitt (1977) FAO paper n. 24, to Allen et al. (1998) FAO paper n. 56, and to the IAM_ETo calculation program and *User's Guide* (Steduto and Snyder, 1998) for further information. The crop coefficients (Kc) need to be derived from literature (e.g., Doorenbos and Pruitt, 1977) and/or estimated by local experimentation.

Much more difficult, however, is to assess crop evapotranspiration when water stress occurs. In this case direct measurements are adopted.

With partial ground cover

The case widely encountered in agricultural practices is not based on the two extreme conditions of either complete bare soil or complete crop ground cover, but rather on dynamic circumstances where incomplete ground cover may persist (e.g., tree crops) or a transition from bare soil to partial or full ground cover (e.g., annual field crops) may develop over a period of time.

Under actual conditions, where crops may undergo water stress to some degree, direct measurements of evapotranspiration (e.g., by water balance or micrometeorological

methods) and soil evaporation (e.g., by micro-lysimeters) remain the best performing approach as compared to models. Where, instead, the crops are not suffering from water deficit, i.e., the canopies are "freely evaporating", semi-empirical models may show to be appropriate.

The Ritchie model. The Ritchie model is hereafter illustrated (Ritchie, 1972) as a practical approach for calculating daily evapotranspiration rates from row crops. In this model, full ground cover transpiration from plants (Eo) and evaporation from bare soil (Eso) are considered separately and then, accounting for the degree of ground cover, the actual plant transpiration (Ep) and soil evaporation (Es) are calculated.

To estimate potential evaporation from crops (~max transpiration) at full ground cover (Eo) the combination equation of Penman (1963) is used, with the empirical wind function modified to account for extra roughness of a crop as compared with open water surface (for which the original wind function of Penman was derived)

$$\text{Eo} = \frac{s\,\text{Rno} + \gamma\,0.262(1 + 0.0061\,u)\,\text{VPD}}{s + \gamma} \tag{27},$$

where u is the wind speed at 2m height, Rno is the net radiation over the crop stand (i.e., the available energy neglecting the soil heat flux G), and all other variables are defined as in Eq. (25). To update the Ritchie model, Eq. (25) may replace Eq. 27. In calculating Rno, the net short wave radiation (Rns) accounts for the albedo of the surface (ε) as a combination of soil and plant albedos. The relevant equations are

$$\text{Rns} = (1 - \varepsilon)\,\text{Rs} \tag{28},$$

where Rs is the incoming solar radiation, and

$$\varepsilon = \varepsilon_s + 0.25(\varepsilon_c - \varepsilon_s)\,\text{LAI} \tag{29}$$

with ε_s being the bare soil albedo (variable from soil to soil and with the degree of wetness), ε_c is the full canopy albedo (generally 0.23), and LAI is the leaf area index. To use Eq. (29), the albedo of the given soil conditions needs to be known. Idso and Reginato (1974) presented albedo values for 17 soils at different degrees of wetness.

A subsequent calculation requires the estimate of the potential evaporation (freely evaporating surface) of the soil below the canopy (Eso), considering that the plants provide shade for part of the soil surface. The same combination equation (27) can be used to calculate Eso, but assuming that VPD and wind are negligible below the canopy. Consequently an "equilibrium evaporation" can be considered, and

$$Eso = \frac{s}{s + \gamma} Rns \tag{30}$$

where Rns is the average net radiation at the soil surface (again, G has been neglected).

The relationship between the fractional net radiation at the soil surface (Rns/Rno) and the leaf area index (LAI) has been measured by several workers for various annual crops and the resulting functional relation is formulated as

$$\frac{Rns}{Rno} = e^{(-0.398 \, LAI)} \quad \Rightarrow \quad Rns = Rno \, e^{(-0.398 \, LAI)} \tag{31}.$$

Then, combining (30) with (31) yields

$$Eso = \frac{s}{s + \gamma} Rno \, e^{(-0.398 \, LAI)} \tag{32}.$$

Of course, the assumption that VPD and wind are negligible does not hold at the beginning of the season, when the canopy is small, so that Eq. (32) is not appropriate very early in the season. Furthermore, Eq. (32) is not used during periods when the soil surface is dry.

After calculating the potential Eo and Eso, the actual soil evaporation (Es) and plant transpiration (Ep) are calculated separately. To calculate Es, it is always needed to distinguish stage 1 (Es₁) from stage 2 (Es₂) of the drying process.

For stage 1, it is formulated that

$$Es_1 = Eso \quad \text{for} \quad \sum Es_1 \leq U \tag{33}$$

where U is the cumulative evaporation of stage 1, until beginning of drying stage 2. For the day of the transition between the two stages, Es can be estimated using Es=0.6 Eso. Values of U can be derived experimentally for a given soil. Estimates of U from four types of soil are indicated in Table 2.

For stage 2, Black et al. (1969) have shown that cumulative evaporation of an initially wet and deep soil can be expressed as

$$\sum Es_2 = \alpha \sqrt{t} \tag{34}$$

where α is dependent on the hydraulic properties of the soil. Values for α can be determined experimentally from cumulative evaporation data for a single drying cycle. Estimates of α for four types of soil are indicated in table 2. Generally, the magnitude of the α values is

approximately proportional to the magnitude of the hydraulic conductivity at -0.1 bar soil matric potential. Thus, approximate values of α can be obtained also from knowledge of the hydraulic conductivity data.

Tab. 2 - *Estimated values of the upper limit for the cumulative evaporation (U) of stage 1 and for the coefficient α of stage 2 , given for four soil types (after Ritchie, 1972).*

Soil Type	U (mm)	α (mm day$^{-1/2}$)
clay-loam	12	5.08
Loam	9	4.04
black-clay	6	3.50
Sand	6	3.34

To further refine the Es estimates, special attention has been paid to the period of transition from stage 1 to stage 2 of drying by Jakson et al. (1976) to account for the energy balance changes owing to soil albedo variations due to different degrees of soil wetness. The reader is referred to their paper for details.

To calculate actual plant transpiration Ep (with no stress induced by water shortage) at different degrees of development (different degrees of ground cover), the following empirical function can be used

$$Ep = Eo \, (-0.21 + 0.7 \, \sqrt{LAI}) \quad \text{for} \quad 0.1 \leq LAI \geq 2.7 \tag{35}$$

with Eo calculated with Eq. (27) or (25).

Eq. (35) was empirically derived for cotton and grain sorghum, so that its ability to represent Ep for other climates and crops remain to be assessed. Nevertheless, the equation was found acceptably correct for barley in England, and soybean and sorghum in Arkansas (USA). More recently, Villalobos and Fereres (1990) found a similar expression for sunflower, cotton and corn, valid for semi-arid areas. Their expression is

$$Ep = 1.07 \, Eo \left(1 - e^{-0.52 \, LAI}\right) \tag{36}$$

The validity of the Ritchie approach has shown to be of wide and practical application; it has been utilized extensively. Tests with mini- and micro-lysimeters have demonstrated its validity, and some results are illustrated in Fig. 8 a and b for a Nebraska (USA) and a Patagonian (Argentina) experiment, respectively.
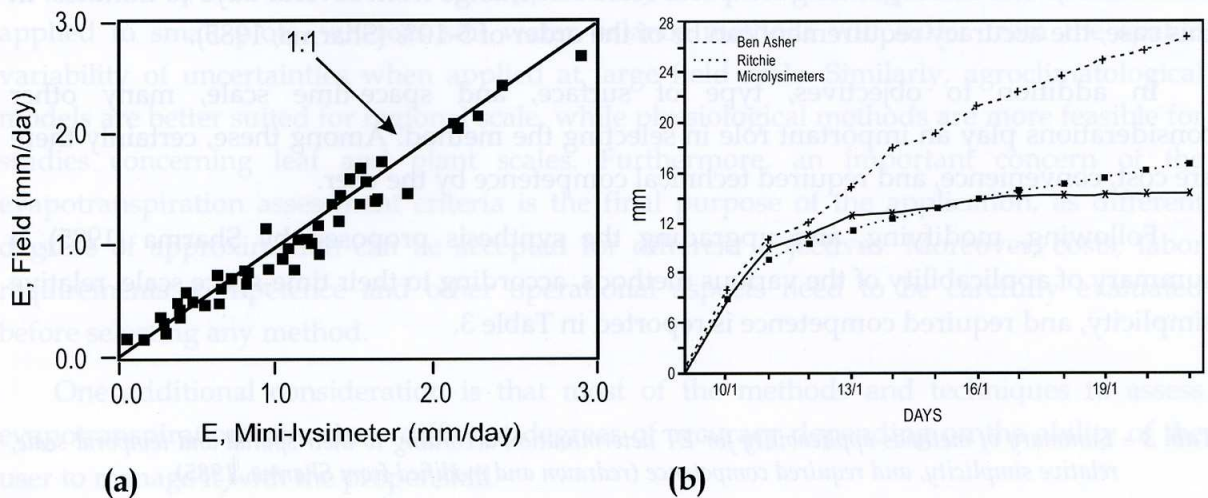
Fig. 8 - (a) Calculated daily evaporation for mini-lysimeters versus surrounding field beneath corn canopy (from Klocke et al., 1990); (b) cumulative evaporation during the drying cycle, estimated in three different ways (from Paruelo et al., 1991).

APPLICABILITY OF THE DIFFERENT METHODS

It should be obvious that no single method to determine crop water consumption can be applicable under all conditions. The criteria for selecting a method is always an optimization problem: the maximization of benefits in accord with general objectives under given constraints. Thus, a clearly defined objective and the corresponding set of boundary conditions are needed. Essential features to be considered are (i) the type of soil-vegetation surface and (ii) the purpose of the information required. This will indicate the spatial and temporal scales and the accuracy requirement for measurements (Sharma, 1985).

As an example, the ET requirement for water management can be used for planning water resources (regional level) and irrigation scheduling (farm level) as well, with changes in spatial scale from thousands of hectares to less than 1 ha, respectively. The corresponding temporal scale may change from a year to a day. What are the accuracy requirements for these purposes? It is difficult to answer this question because it imposes another optimization problem with associated cost and benefit analysis. Though, an estimate of water requirement of about 15% is generally accepted as a good accuracy for water management issues.

If the purpose of ET requirements is to get more insight on the ET process itself and on the factors entering into play, the spatial scale can change from field and to plot scale

(hundreds of square meters) to the single plant or even leaf scale (hundreds of square centimeters). The corresponding temporal scale can change from several days to minutes. In this case, the accuracy requirement can be of the order of 5-10% (Sharma, 1985).

In addition to objectives, type of surface, and space-time scale, many other considerations play an important role in selecting the method. Among these, certainly there are cost, convenience, and required technical competence by the user.

Following, modifying, and upgrading the synthesis proposed by Sharma (1985), a summary of applicability of the various methods, according to their time-space scale, relative simplicity, and required competence is reported in Table 3.

Tab. 3 - Summary of methods applicability for ET determination according to their spatial and temporal scale, relative simplicity, and required competence (redrawn and modified from Sharma, 1985).

METHODS	TIME SCALE min hr day week month year	SPACE SCALE leaf plant plot field catchment region	RELATIVE SIMPLICITY	REQUIRED COMPETENCE
Water Balance				
Lysimeter (weighing)			medium	medium
Lysimeter (nonweighing)			medium	medium
Soil-water balance			medium	medium
Micrometeorological				
Aerodynamic			low	medium
Bowen-ratio			low	medium
Eddy covariance			low	high
Agrometeorological (models)				
Penman-Monteith			high	low
Ritchie			high	medium
Plant Physiological				
Chamber			medium	medium
Heat pulse/balance			low	high

CONCLUSIONS

The field assessment of evapotranspiration remains one of the major issues in arid and semi-arid ecosystems to achieve a proper management and an efficient use of water resources.

Many methods and techniques for measuring and estimating evapotranspiration from crop stands have been presented to illustrate the alternative means to tackle specific problems under specific conditions. It should be clear, in fact, that no single method can be

applicable under all conditions. The major distinction among the methods relates strongly to the different spatial and time scale of applicability. Micrometeorological methods cannot be applied to small plots, whereas soil water balance methods will suffer from the spatial variability of uncertainties when applied at large field scale. Similarly, agroclimatological models are better suited for regional scale, while physiological methods are more feasible for studies concerning leaf and plant scales. Furthermore, an important concern of the evapotranspiration assessment criteria is the final purpose of the application, as different degrees of approximation can be accepted for different objectives. Moreover, costs, labor requirements, competence and other operational aspects need to be carefully evaluated before selecting any method.

· One additional consideration is that most of the methods and techniques to assess evapotranspiration may show different degrees of accuracy depending on the ability of the user to manage it with the proper skill.

REFERENCES

Aboukhaled A., Alfaro A. and Smith M. (1982). Lysimeters. *Irrigation and Drainage Paper n. 39*, FAO Rome, 68 p.

Allen R.G., Smith M., Perrier A. and Pereira L.S. (1994a). An update for the definition of reference evapotranspiration. *ICID Bulletin*, 43(2):1-34.

Allen R.G., Smith M., Perrier A. and Pereira L.S. (1994b). An update for the calculation of reference evapotranspiration. *ICID Bulletin*, 43(2):35-92.

Allen R.G., Pereira L.S., Raes D. and Smith M. (1998). Crop evapotranspiration. Guidelines for computing crop water requirements. *Irrigation and Drainage Paper n. 56*, FAO Rome, 300 p.

Baker J.M. and van Bavel C.H.M. (1987). Measurement of mass flow of water in stems of herbaceous plants. *Plan, Cell and Envir.*, 10:777-782.

Baker J.M. and Nieber J.L. (1989). An analysis of the steady-state heat balance method for measuring sap flow in plants. *Agric. For. Meteorol.*, 48:93-109.

Baldocchi D.D., Hicks B.B. and Meyers T.P. (1988). Measuring biosphere-atmosphere exchanges of biologically related gas with micrometeorological methods. *Ecology*, 69(5):1331-1340.

Black T.A., Gardner W.R. and Thurtell G.W. (1969). The prediction of evaporation, drainage, and soil water storage for a bare soil. *Soil Sci. Soc. Amer. Proc.*, 33:655-660.

Doorenbos J. and Pruitt W.O. (1977). Crop water requirements. *Irrigation and Drainage Paper n. 24*, FAO Rome, 179 p.

Dugas W.A. (1990). Comparative measurement of stem flow and transpiration in cotton. *J. Appl. Climatol.*, 42:215-221.

Ham J.M., Heilman J.L. and Lascano R.J. (1991). Soil and canopy energy balance of a row crop at partial cover. *Agronomy J.*, 83:744-753.

Held A.A., Steduto P., Orgaz F., Matista A. and Hsiao T.C. (1990). Bowen-ratio energy-balance technique for estimating crop net CO_2 assimilation and comparison with a canopy chamber. *J. Theor. Appl. Climatology*, 42:203-213.

Hillel D. (1980). Applications of Soil Physics. *Academic Press*, New York, 385 p.

Idso S.B. and Reginato R.J. (1974). Assessing soil-water status via albedo measurement. *Hydrol. Water Resour. Ariz. Southwest*, 4:41-54.

Itier B. (1996). Measurement and estimation of evapotranspiration. In: Pereira L.S., Feddes R.A., Gilley J.R. and Lesaffre B. (eds). Sustainability of Irrigated Agriculture. NATO ASI Series E, Applied Sciences, vol. 132. Kluwer Academic Publ., The Netherlands, pp. 171-191.

Jackson R.D., Idso S.B. and Reginato R.J. (1976). Calculation of evapotranspiration rates during the transition from energy-limiting to soil-limiting phase using albedo data. *Water Resources Research*, 12(1):23-26.

Kanemasu E.T., Hicks M.L. and Heilman J.L. (1979). Techniques for calculating energy and mass fluxes. In: Barfield B.L. and Gerber J.F. (eds). Modification of the Aerial Environment of Crops. *ASAE*, St. Joseph, Michigan (USA), pp. 156-182.

Kelliher F.M., Leuning R., Raupach M.R. and Schulze E.-D. (1995). Maximum conductances for evaporation from global vegetation types. *Agric. For. Meteorol.*, 73:1-16.

Klocke N.L., Martin D.L., Todd R.W., Dettaan D.L. and Polymenoupoulos A.D. (1990). Evaporation measurements and predictions from soils under crop canopies. *Transaction of ASAE*, 33(5): 1590-1596.

McNaughton K.G. and Jarvis P.G. (1991). Effects of spatial scale on stomatal control of transpiration. *Agric. For. Meteorol.*, 54:279-301.

Monteith J.L. (1973). Principles of Environmental Physics. *Edward Arnold*, London, 241 p.

Paruelo J.M., Aguiar M.R. and Golluscio R.A. (1991). Evaporation estimates in arid environments: an evaluation of some methods for Patagonian steppe. *Agric. For. Meteorol.*, 55:127-132.

Penman H.L. (1963). Vegetation and hydrology. *Technical Communication n. 53*, Commonwealth Bur. of Soils, Harpeden, Herts., England, 124 p.

Pickering N.B., Jones J.W. and Boote K.J. (1993). Evaluation of the portable chamber technique for measuring canopy gas exchange by crops. *Agric. For. Meteorol.*, 63:239-254.

Reicosky D.C. and Peters D.P. (1977). A portable chamber for rapid evapotranspiration measurements on field plots. *Agronomy J.*, 69:729-732.

Ritchie J.T. (1972). Model for predicting evaporation from a row crop with incomplete cover. *Water Resources Research*, 8(5):1204-1213.

Rosemberg N.J., Blad B.L. and Verma S.B. (1983). Microclimate. The Biological Environment. *John Wiley & Sons*, New York, 300 p.

Sharma M.L. (1985). Estimating evapotranspiration. *Advances in Irrigation*, 3:213-281.

Slavik B. (1974). Water exchange between plant and atmosphere. *Ecological Studies*, 9:236-281.

Steduto P. and Hsiao T.C. (1998). Maize canopies under two soil water regimes. II. Seasonal trends of evapotranspiration, carbon dioxide assimilation and canopy conductance, and as related to leaf area index. *Agric. For. Meteorol.*, 89:185-200.

Steduto P. and Snyder R.L. (1998). IAM_ETo Software program & User's Guide. *Options Méditerranéennes, Série B: Etudes et Recherches,* 44 p.

Steinberg S.L., van Bavel C.H.M. and McFarland M.J. (1989). A gauge to measure mass flow in stems and trunks of woody plants. *J. Amer. Soc. Hort. Sci.*, 114:466-472.

Villalobos F.J. and Fereres E. (1990). Evaporation measurements beneath corn, cotton, and sunflower canopies. *Agronomy J.*, 82:1153-1159.