Measurement and estimation of actual evapotranspiration in the field under Mediterranean climate: a review

G. Rana a,*, N. Katerji b

a Istituto Sperimentale Agronomico, via C. Ulpiani, 5, 70125 Bari, Italy
b INRA-Unité de Bioclimatologie, 78850 Thieval-Grignon, France

Received 19 February 1999; received in revised form 13 August 1999; accepted 13 May 2000

Abstract

The Mediterranean regions are submitted to a large variety of climates. In general, the environments are arid and semi-arid with summers characterised by high temperatures and small precipitation. Due to the scarcity of water resources, the correct evaluation of water losses by the crops as evapotranspiration (ET) is very important in these regions. In this paper, we initially present the most known ET measurement methods classified according to the used approach: hydrological, micrometeorological and plant physiological. In the following, we describe the methods to estimate ET, distinguishing the methods based on analytical approaches from the methods based on empirical approaches. Ten methods are reviewed: soil water balance, weighing lysimeter, energy balance: Bowen ratio, aerodynamic method, eddy covariance, sap flow method, chambers system, Penman–Monteith model, crop coefficient approach and soil water balance modelling approach. In the presentation of each method, we have recalled the basic principles, underlined the time and space scale of its application and analysed its accuracy and suitability for use in arid and semi-arid environments. A specific section is dedicated to advection. Finally, the specific problems of each method for correct use in the Mediterranean region are underlined. In conclusion, we focus attention on the most interesting new guidelines for research on the measurement and estimation of actual crop evapotranspiration. © 2000 Elsevier Science B.V. All rights reserved.

Keywords: Actual evapotranspiration; Direct ET measurement methods; Indirect ET measurement methods; ET modelling; Mediterranean climate

1. Introduction

The Mediterranean climate is characterised by a hot and dry season in summer and a mild tem-
does not exceed 2–3 months. In addition, the rainfall and temperature diagrams (Fig. 2) show great differences between the North (autumn rainfall) and the South (winter rainfall) of the basin. During summer, the simultaneous occurrences of high temperatures and small precipitation causes high evapotranspiration (Hamdy and Lacirignola, 1999). For these reasons, 75% of the available water in the Mediterranean area is used for agricultural purposes.

Furthermore, the Mediterranean region is characterised by:
Table 1
Estimation of the percentage of irrigated soils affected by salinity in several Mediterranean countries (Hamdy et al., 1995)

<table>
<thead>
<tr>
<th>Country</th>
<th>Irrigated surface (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Algeria</td>
<td>10–15</td>
</tr>
<tr>
<td>Cyprus</td>
<td>25</td>
</tr>
<tr>
<td>Egypt</td>
<td>30–40</td>
</tr>
<tr>
<td>Spain</td>
<td>10–15</td>
</tr>
<tr>
<td>Israel</td>
<td>13</td>
</tr>
<tr>
<td>Greece</td>
<td>7</td>
</tr>
<tr>
<td>Jordan</td>
<td>16</td>
</tr>
<tr>
<td>Morocco</td>
<td>10–15</td>
</tr>
<tr>
<td>Portugal</td>
<td>10–15</td>
</tr>
<tr>
<td>Syria</td>
<td>30–35</td>
</tr>
</tbody>
</table>

Moreover, semi-arid and, mostly, arid climates have a great impact on crop growth in conditioning yield and product quality. Under these weather conditions, rainfed crops and agricultural fields with limited water resources are often submitted to water stress. Thus, it becomes fundamental to know the exact losses of water by evapotranspiration (i.e. actual evapotranspiration) and the crop water status and its influence on production.

Actual crop ET can be measured (directly or indirectly) or estimated. For example, for research purposes in plant eco-physiology, ET must be precisely measured, while for farm irrigation management it can be estimated. The lower the degree of accuracy in the estimation, the greater will be the water waste by incorrect management of irrigation.

In this paper, we review the most important methods of measuring and estimating actual ET, at plot scale (e.g. at farm level), giving, for each method, the problems, limits and advantages for their use in a field under the Mediterranean climate. This summary cannot be, of course, an exhaustive and complete review of all the existing ET methods; in fact, we will focus attention on the most known and diffuse methods in the international research community, paying particular attention to those used in agricultural research.

The scaling up of ET from plot scale to regional scale requires specific methods and techniques and, therefore, will not be treated in this paper.

2. Measurement and estimation

There is a great variety of methods for measuring ET; some methods are more suitable than others for accuracy or cost or are particularly suitable for given space and time scales. For several applications, ET needs to be predicted, so it must be estimated by model.

It is convenient to discuss the methods of determining ET in considering separately the measurement and modelling aspects.
In general, a ‘measurement’ of a physical parameter is a quantification of an attribute of the material under investigation, directed to the answering of a specific question in an experiment (Kempthorne and Allmaras, 1986). The quantification implies a sequence of operations or steps that yields the resultant measurement. Conventionally, if the value of the parameter is quantified by the use of an instrument, it is ‘directly’ measured and when it is found by means of a relationship among parameters, it is ‘indirectly’ measured (Sette, 1977).

The methods of measuring ET should be divided into different categories, since they have been developed to fulfill very different objectives. One set of methods are primarily intended to quantify the evaporation over a long period, from weeks to months and growth season. Another set of methods has been developed to understand the process governing the transfer of energy and matter between the surface and atmosphere. The last set of methods is used to study the water relations of individual plants or part of plants. Therefore, following Rose and Sharma (1984), it is convenient, when discussing the ET measurement, to place the variety of methods in groups, where the main approach or method depends on concepts from hydrology, micrometeorology and plant physiology, as follow:

- ET measurement
  - Hydrological approaches
    1. Soil water balance
    2. Weighing lysimeters
  - Micrometeorological approaches
    3. Energy balance and Bowen ratio
    4. Aerodynamic method
    5. Eddy covariance
  - Plant physiology approaches
    6. Sap flow method
    7. Chambers system

A physical parameter can be considered as ‘estimable’ if it is expressed by a model. The objective of ET modelling can vary from the provision of a management tool for irrigation design, to the provision of a framework for either detailed understanding of a system or to interpret experimental results. To meet these requirements, it is more convenient to use methods with a sound physical basis, but often the available data only allow the use of either empirical or statistical approaches. Thus, to discuss the ET modelling, it is convenient to divide the ET models as follows:

- ET estimation
  - Analytical approach
    8. Penman–Monteith model
  - Empirical approach
    9. Methods based on crop coefficient approach

This categorisation is, of course, far from completion, but can be considered a good trace in a review of ET determination.

3. Actual crop evapotranspiration measurement

The different methods for directly and indirectly measuring ET are based on the measurement of two classes of factors:

1. The soil water content and the physical characteristics of the evapotranspirative surface (height, plant density, canopy roughness, albedo);
2. Climatic variables: solar radiation, wind speed and thermodynamic characteristics of the atmosphere above the canopy.

In the description of the measurement methods, we will consider as true the hypothesis of ‘conservative flux’, i.e. the flux density does not change with the height above the canopy. So that, in a first approximation, we will consider the different factors (belonging to both (a) and (b) classes) as constant in the space considered (the field).

To say that the factors (a) are constant in the space is the same as saying that evapotranspirative surfaces are perfectly homogeneous along the horizontal scale. In Mediterranean regions, this hypothesis is usually far from being verified. In fact, the fields present crop types, phenological stages and water conditions very changeable in the space. This induces a net discontinuity of surface moisture among fields and, consequently, a non-negligible effect of advection due to energy side exchange between contiguous plots.
The role of the advective regime on the measurement and estimation of ET will be analysed in a separate chapter at the end of the methods presentation.

3.1. Hydrological approaches

3.1.1. Soil water balance

Soil water balance is an indirect method, in fact ET is obtained as a residual term in the water balance equation. This equation is based on the principle of conservation of mass in one dimension applied to the soil, its complete expression is:

\[ P + I + W - ET - R - D = \pm [\Delta S]_0 \]  

(1)

where \( P \) is precipitation, \( I \) is irrigation, \( W \) is contribution from water table upward, \( R \) is surface runoff, \( D \) is drainage and \( \Delta S \) is soil water storage in the soil layer, where the roots are active to supply water to the plant (in m). All the terms are in millimeters per unit time. Soil water storage between two dates (\( i \) and \( i-1 \)) is:

\[ [\Delta S]_0 = [S_i - S_{i-1}]_0 \]  

(2)

with \( S \), the soil water content.

Since it is often very difficult to accurately measure all the terms of Eq. (1), a number of simplifications makes this method unsuitable for precise ET measurements. In fact, irrigation water supply is, in principle, known and precipitation can be measured by rain gauges, but all the other terms need to be measured or, at least, estimated.

The soil water balance method is applicable to small plots (\( \approx 10 \text{ m}^2 \)) or to a large catchment (\( \approx 10 \text{ km}^2 \)); it may cover periods ranging from a week to a year.

Often, for operational application, the soil water balance Eq. (1) is expressed in its simplified form:

\[ P + I = ET \pm [\Delta S]_0 \]  

(3)

The simplifications introduced in Eq. (3) could be critical if applied in dry environments. In arid and semi-arid areas with very small slopes, runoff term \( R \) could be neglected (e.g. Holmes, 1984) but, actually, it depends on the occurrence and characteristics of precipitation (amount, duration and intensity) and can only be neglected for a particular type of soil (Jensen et al., 1990), i.e. coarse (sand and loamy sand) and moderately coarse (sandy loam).

Drainage (\( D \)) is the most unknown of Eq. (1). Some researchers suggest that it can be neglected in dry regions (e.g. Holmes, 1984), but actually it depends on the soil depth, slope, permeability and surface storage (Jensen et al., 1990; Parkes and Li Yuanhua, 1996) and needs to be checked in each particular case (Brutsaert, 1982), depending also on the climate and weather (Katerji et al., 1984). In some situations it is so important that its direct measurement can be used to estimate evapotranspiration on a weekly or greater scale (for an extensive review see Allen et al. (1991)). Moreover, Katerji et al. (1984) demonstrated that it is not simple to establish if the water deep flow can be neglected, in fact it can be a non-negligible fraction of the water balance both in dry and humid seasons and at different time scale. In general, at daily scale it can be neglected if the water supply (\( P \) and/or \( I \)) does not exceed the soil water capacity (Holmes, 1984; Lhomme and Katerji, 1991).

In arid regions, the terms \( W \) and \( \Delta S \) could show problems of correct evaluation. In fact, if the soil system is closed (i.e. shallow soils or soils with a very deep water table), \( W \) can be considered negligible and \( \Delta S \) can be easily determined. In this case, the simplified soil water balance works well, as shown in Fig. 3, where ET measured by Eq. (3) is compared with a reference method (Bowen ratio). The data are relative to a maize crop grown in a Mediterranean region (Southern Italy) and soil water storage is measured by TDR and the gravimetric method.

Vice versa, if the system is open (deep soils or soils with a surficial water table), \( W \) cannot be neglected and \( \Delta S \) is difficult to measure exactly, so that the simplified form of the soil water balance does not work well. This inaccuracy was well demonstrated by Katerji et al. (1984). In fact, they showed that during humid seasons, \( W \) and \( \Delta S \) are of the same order, while during dry seasons, \( W \) can reach 30% of cumulated seasonal actual ET (Fig. 4).
Soil water content needs to be measured accurately and over an adequate depth. It can be measured by a wide number of methods (an extensive review can be found in Stafford (1988) and Phene et al. (1990)). Three methods are the most important: gravimetric sampling, neutron scattering and electrical resistance.

The gravimetric method remains the most widely used, particularly by technicians and consultants; it is well developed at research level and also for irrigation management purposes. It is simple to apply but it can be used only for large time scale (weekly or greater).

Neutron probe remains a useful tool at research level only and its performance has not been well-established in arid environments. In fact, Payne and Bruck (1996), for example, stated that neutron probe tends to lose accuracy under the arid climate of Sahelian Africa, mostly near the surface, wetting fronts and textural discontinuities; conversely, Evett and Steiner (1995) found very good results in a wide range of situations. Furthermore, neutron probe poses particularly difficult and regulatory problems in developing countries.

Electrical resistance matrix type sensors are the least expensive (gravimetry is cheaper) and can be readily used for automatic irrigation control systems (Phene and Howell, 1984; Phene et al., 1989, among many others), but they are not very accu-

---

**Fig. 3.** Comparison between ET measured by Bowen ratio method, soil water balance with TDR and soil water balance by gravimetric method, at daily scale, on corn grown in Southern Italy (Mastrorilli et al., 1998).
rate in estimating actual ET in soil water depletion conditions (see, e.g. Bausch and Bernard, 1996).

The accuracy of soil water balance strongly depends on time and space scales of actual soil moisture measurements (Burrough, 1989) and on the representativeness of the soil sampling (Bertuzzi et al., 1994; Leenhard et al., 1994). In order to meet all these requirements for soil moisture measurement, the technique of time domain reflectometry (TDR) has been introduced quite recently by Topp and Davis (1985) and Topp et al. (1994) among others. It seems that TDR is able to correctly estimate soil moisture both under temperate (Ledieu et al., 1986; Werkhoven, 1993; Plauborg, 1995) and semi-arid climates (Mastrorilli et al., 1998). When TDR probes are used it is possible to have a soil water balance at plot scale and hourly scale.

Problems can be encountered in clay soils when the soil moisture is measured by means of probes (neutron scattering, electrical resistance or TDR). In fact, clay soils submitted to an arid climate (i.e. at low moisture content) crack where they are not adequately irrigated (e.g. Diestel, 1993); the deep cracks do not allow suitable contact between soil and probes (Haverkamp et al., 1984a,b) and the reading of the soil moisture can be affected by large errors (Bronswijk et al., 1995).

3.1.2. Weighing lysimeters

Weighing lysimeters have been developed to give a direct measurement of the term ET in Eq. (1). Historically, they have always been considered as a suitable tool to correctly measure evapotranspiration (Tanner, 1967; Aboukhaled et al., 1982).

In general, it is a device, a tank or container, to define the water movement across a boundary; actually, only a ‘weighing lysimeter’ can determine ET directly by the mass balance of the water as contrasted to non-weighing lysimeter, which indirectly determines ET by volume balance (Howell et al., 1991).

Weighing lysimeters placed in the field contains soil cultivated as the field around it, the sensor is a balance able to measure the weight variation due to ET, often this variation is measured by means of an electronic sensor (i.e. load cell). Under temperate climate, they are able to measure ET at a daily scale with an accuracy of > 10% (Perrier et al., 1974; Klocke et al., 1985b among others) and at an hourly scale with an accuracy of 10–20% (Pruitt and Lourence, 1985; Allen et al., 1991).

However, the weighing lysimeter data are not always representative of conditions of the whole field but, in several situations, they represent only the ET of just one point in the field (Grebet and Cuenca, 1991).

Besides the soil, height and vegetation density differences between the lysimeter and outside vegetation can severely affect ET measurements. In this case, both aerodynamic and radiative transfers to the lysimeter canopy are increasing. If the lysimeter surface and its close area are surrounded
by drier vegetation or bare soil, an oasis effect occurs. Net radiation in excess of latent heat is converted in sensible heat that is advected toward the lysimeter, resulting in net supply of energy to the lysimeter vegetation. All these previously described problems cause an increase of ET relative to the surrounding crop. This overestimation of ET could be particularly important under a high radiative climate, as in the Mediterranean region.

Furthermore, environmental problems can affect measurements of ET with lysimeters (for a review see Allen et al. 1991). One of the most common errors made is the incorrect estimation of an evaporating area of a lysimeter. In fact, this area is calculated by the inner dimensions of the container rather than the true vegetative area, which can be greater than the lysimeter surface when the vegetation from both inside and outside the lysimeter reaches across the rim. This error could be \( \approx 20\% \).

The lysimeter rim can also influence ET measurements. One of the most important effects, mainly in arid environments, is the heating of the metallic rim by radiation resulting in microadvection of sensible heat into the lysimeter canopy. Moreover, if the lysimeter rims are too tall relative to crop, the wind is shielded and the radiative energy balance is modified due to the reflection of solar radiation by the inner wall of the rim toward the crop.

Under arid and semi-arid climates, problems linked to the atmospheric evaporation demand can worsen the performances. In fact, if the soil inside lysimeters has deep cracks (often along the border in contact with the soil), the water evaporation continues from the deepest layers, so that lysimeter can overestimate ET in these periods and underestimate ET in the following periods, when the water depletion inside is greater than that in the field, due to water stress condition of inner plants (Jensen et al., 1990). Conversely, irrigation and rainfall infiltrate into these cracks; this, along with lack of root extraction of water, causes the soil within the lysimeters to be much wetter than the surrounding field soil (Klocke et al., 1991).

Furthermore, the weighing lysimeter is limited in depth in order to make possible the weighing of the soil–plants system; this limitation adds new problems for its use in Mediterranean regions. In fact, it does not take into account the effects of the deeper water fluxes (capillary rising) on crop, which can be very important in water stress periods. Moreover, in areas with deep soil, the weighing lysimeter does not take into account the effects of the total soil available on the plant growth. In this case, i.e. very deep soils, the representativeness of the lysimeter becomes a very difficult technical problem, due to the excessive weight of the system soil–tank–plants. In Fig. 5, three kinds of lysimeter are shown: since the total available water of nearby soil is \( \approx 260 \text{ mm} \), only the first one is well representative of the actual field, but in this case the weight of the system soil–tank is very high (12 t).

3.2. Micrometeorological approaches

From the energetic point of view, evapotranspiration can be considered as the energy employed for transporting water from inner leaves and plant organs to the atmosphere as vapour. In this case it is called 'latent heat' \( (\lambda E, \text{ with } \lambda \text{, latent heat of vaporisation}) \) and it is measured as energy flux density \( (\text{W m}^{-2}) \).

We shall describe in the following sections the main methods based on physical or meteorological principles, by which the latent heat flux can be measured. Three techniques will be analysed: the energy balance and Bowen ratio, the aerodynamic method and the eddy covariance. In general, all
these techniques require accurate measurements of meteorological parameters on small temporal scale (1 h or less). Due to the conservative hypothesis of all the flux density above the crop, they can be applied only in large flat areas with uniform vegetation. There are many situations of practical interest where they cannot be used, mostly in the Mediterranean regions. For example, mixed plant communities, hilly terrain and small plots.

Much care must be taken to analyse local and regional advection under arid conditions (Prueger et al., 1996) and, eventually, to introduce corrections to the measured fluxes (see the section on advection).

3.2.1. The energy balance and Bowen ratio method

The latent heat flux can be obtained from measurements of the energy budget of the surface covered with an active growing crop. In fact, \( \lambda E \) represents the major used part of available energy due to the radiation balance. All the energy involved in the evapotranspiration phenomenon must satisfy the closure of the energy balance:

\[
R_n - G = H + \lambda E
\]

(4)

where we neglect advective processes and where

\[
R_n \text{ (W m}^{-2}\text{), net radiation and } G \text{ (W m}^{-2}\text{), soil heat flux, are directly measurable by net-radiometers and soil heat flux plates, respectively (for an extensive review of environmental instrumentation see Fritschen and Gay (1979)); and } H \text{ is the sensible heat flux density (W m}^{-2}\text{).}
\]

We define the Bowen ratio \( \beta = H/\lambda E \), so Eq. (4) can be rearranged to give

\[
\lambda E = \frac{R_n - G}{1 + \beta}
\]

(5)

\( \beta \) can be measured by the ratio of the air temperature difference between two levels (\( \Delta T \)) and the vapour pressure difference (\( \Delta e \)), with \( e \) (kPa) air vapour pressure, measured at the same two levels:

\[
\beta = \frac{\gamma \Delta T}{\Delta e}
\]

(6)

The Bowen ratio is an indirect method, its accuracy has been analysed by many authors (e.g. Fuchs and Tanner, 1970; Sinclair et al., 1975; Revheim and Jordan, 1976) and can be estimated to be within 10% of the measured value. The Bowen ratio method has been widely studied in a variety of field conditions and it has been proven a standard very accurate method in semi-arid environments (e.g. Dugas et al., 1991; Frangi et al., 1996; Zhao et al., 1996) and for tall crops also (Cellier and Brunet, 1992; Rana and Katerji, 1996).

As previously stated, under arid climates the crops can experience water stress. In this case \( \Delta T \) can be quite high but \( \Delta e \) is very low; so that it is very important to have highly accurate measurements of air vapour pressure (Angus and Watts, 1984). The usual and simplest way is to use differential psychrometry; to meet the requirements of accuracy, for continuous recording, one has to keep the bulbs wet and clean (Fritschen and Gay, 1979). Furthermore, the used thermometers have to be well calibrated, in order to detect temperature differences of 0.05–0.2°C; accuracy can be sensibly improved by fluctuating the psychrometric system (Webb, 1960; Gay, 1988; Fritschen and Simpson, 1989).

Recently, improvements have been obtained by using: (i) just one hygrometer that measures alternatively the humidity of air pumped from two levels (Cellier and Olioso, 1993); (ii) a single cooled mirror dew point hygrometer (some such instruments are briefly described in Dugas et al., 1991); and (iii) spatial averaging systems (Bausch and Bernard, 1992). In any case, technical problems remain in the hygrometers for the measurement of air humidity in arid regions, especially during water stress periods, when its value is usually very low.

3.2.2. The aerodynamic method

If we assume that a flux density can be related to the gradient of the concentration in the atmospheric surface layer (ASL), the latent heat flux by the aerodynamic technique can be determined directly by means of the scaling factors \( u^* \) and \( q^* \), with \( q \) specific air humidity (kg kg\(^{-1}\)), (see e.g. Grant, 1975; Saugier and Ripley, 1978):

\[
\lambda E = - \lambda \rho u^* q^*
\]

(7)
here \( \rho \) is density of air (kg m\(^{-3}\)) and the friction velocity \( u^* \) (m s\(^{-1}\)) is derived from the wind profile measurement:

\[
u^* = \frac{k u}{\ln \left( \frac{z - d}{z_0} \right) - \Psi_m}
\]

where \( k = 0.41 \) is the von Karman constant, \( d \) (m) is the zero plane displacement height, \( z_0 \) (m) is the roughness length of the surface and \( \Psi_m \) is the stability correction function for momentum transport. \( q^* \) is determined similarly from the humidity profile measurement:

\[
q^* = \frac{k(q - q_0)}{\ln \left( \frac{z - d}{z_0} \right) - \Psi_e}
\]

where \( q_0 \) is the air humidity extrapolated at \( z = d + z_0 \) and \( \Psi_e \) is the correction function for latent heat transport.

The calculation of stability functions is made by iterative processes (e.g. Pieri and Fuchs, 1990); users who did not take stability corrections into account have been criticised (Tanner, 1963; Perrier et al., 1974). The problem of taking into account the atmospheric stability is particularly important in arid environments. In fact, the coupling of dry soil and rapid warming or cooling up of air at contact with the earth surface (vegetation or bare soil) leads to extreme conditions of strong stability and instability, with sudden passage from a regime to the opposite one.

The major difficulty with this technique is the correct measurement of the vapour pressure at different heights above the crop. For this reason \( \lambda E \) can be derived indirectly by the energy balance Eq. (4) if the sensible heat flux is determined by the flux-gradient relation:

\[
H = -\rho c_p u^* T^*
\]

where \( T^* \) is deduced by the air temperature profile:

\[
T^* = \frac{k(T - T_0)}{\ln \left( \frac{z - d}{z_0} \right) - \Psi_h}
\]

where \( T_0 \) is the temperature extrapolated at \( z = d + z_0 \) and \( \Psi_h \) is the correction function for the heat transport.

Under this form, the main advantage of the aerodynamic technique consists in avoiding humidity measurements. Nevertheless, the accuracy depends on the number of measurement levels of wind speed and temperature profiles. In fact, Eq. (8) and Eq. (11) require at least three or four levels (Webb, 1965), but accuracy is improved when several levels are used (Legg et al., 1981; Wieringa, 1993). When the stability correction functions of Dyer and Hicks (1970) and Paulson (1970) are used, this method gives good results (Grant, 1975; Pieri and Fuchs, 1990).

A simplified version of the method has been proposed by Itier (1980, 1981) and Riu (1982), based on the measurement of \( \Delta u \) and \( \Delta T \), i.e. wind speed and temperature at two levels only. This method was successfully used to measure actual evapotranspiration of soybean grown in a region of Southern Italy (Rana et al., 1990).

The aerodynamic method does not work with enough accuracy on tall crops, neither in its complete form (Garratt, 1978; Thom et al., 1975) nor in its simplified form (Rana and Katerji, 1996). Correction for making this method applicable for tall crops was attempted by Cellier and Brunet (1992).

3.2.3. The eddy covariance

The transport of scalar (vapour, heat, CO\(_2\)) and vectorial amounts (i.e. momentum) in the low atmosphere in contact with the canopies is mostly governed by air turbulence. The first complete scientific contributions to this topic were given by Dyer (1961) and Hicks (1970); for extensive details of the theory see, for example, Stull (1988).

When certain assumptions are valid, theory predicts that fluxes from the surface can be measured correlating the vertical wind fluctuations from the mean \( (w^e) \) with the fluctuations from the mean in concentration of the transported admixture. So that for latent heat, we can write the following covariance of vertical wind speed (m s\(^{-1}\)) and vapour density \( (q^e) \) in g m\(^{-3}\) :

\[
\lambda E = \rho \dot{w} \dot{q}^e
\]

By making measurements of the instantaneous fluctuations of vertical wind speed \( w^e \) and of humidity \( q^e \) at sufficient frequency to obtain the
contribution from all the significant sizes of eddy and summing their product over an hourly time scale (from 15 min to 1 h), Eq. (12) gives directly the actual crop evapotranspiration.

A representative fetch is required; fetch to height ratios of 10 are usually considered adequate but longer fetches are desirable (Wieringa, 1993). The distribution of eddy size contributing to vertical transport creates a range of frequencies important to eddy correlation measurements (for a brief review of the method requirements, see Tanner et al. (1985)). The sensors must sufficiently respond to measure the frequencies at the high end of the range, while covariance averaging time must be long enough to include frequencies at the low end (Kaimal et al., 1972; McBean, 1972).

To measure ET directly by this method, vertical wind fluctuations have to be measured and acquired contemporary to the vapour density. The first one can be measured by sonic anemometer, the second by fast response hygrometer; both have to be acquired at a typical frequency of 10–20 Hz. The Krypton-type fast hygrometers perform well in the field, but they are expensive and very delicate, so they need particular maintenance. In fact, the commercial fast hygrometers can be severely damaged if moistened, therefore they can only be installed during the day period, which makes it difficult to use this instrument continuously for a period longer than a few days. Furthermore, errors in eddy covariance method can be due not only to possible deviations from the theoretical assumptions, but also to problems of the sensors configuration and meteorological characteristics (Foken and Wichura, 1996).

One important hypothesis to be verified is that time series must be stationary at the scale of the averaging period, which requires some kind of detrending of the original turbulent signals, like linear detrending or more complex high-pass filtering (Kaimal and Finnigan, 1994).

Other known problems are due to the geometrical configuration of the sensors. A distortion of the flow can be caused by the sensor arrangement of the anemometer itself and other sensors. The spatial separation between the sonic anemometer and the hygrometer can cause lack of covariance between the wind speed and the humidity fluctuations. In fact, the typical distance between the measuring path of the vertical wind fluctuation and the hygrometer is 30–40 cm, this spacing acts like a lower-pass filtering process on the measured signals and must be corrected (Foken and Wichura, 1996).

An important aspect to be considered is the density correction (Webb et al., 1980), this correction is particularly important in semi-arid environments, in fact it is proportional to the sensible heat flux and may be quite large for high Bowen ratio (Villalobos, 1997).

To avoid some of the above problems linked to the humidity fluctuations measurements, \( \lambda E \) can be obtained indirectly as residue of the energy budget Eq. (4) if the sensible heat flux is expressed by:

\[
H = \rho c_p \bar{w'} T'
\]  

(13)

where \( c_p \) (J kg\(^{-1}\) C\(^{-1}\)) is the specific heat and density of air. The wind speed and temperature fluctuations are measured by means of sonic anemometer and fast response thermometer, respectively.

Despite problems linked to the correct management of the sensors and data remain, this method has very good performances both at hourly and daily scale, also in semi-arid environments. In Fig. 6, the comparison between ET measured by this method and Bowen ratio method, at daily scale, is shown for a tall crop (sweet sorghum) grown in Southern Italy.

In conclusion, the use of eddy covariance for latent or sensible heat flux is still a useful tool only at research level, even if the recent development of robust sensors could permit in the future its practical application in arid regions.

3.3. Plant physiology approaches

Plant physiology methods either measure water loss from a whole plant or a group of plants. These may include methods such as tracer technique and porometry but here, only the most diffuse methods will be analysed: the sap flow method and the chambers system.
3.3.1. Sap flow method

Sap flow is closely linked to plant transpiration by means of simple accurate models; sap flow can be measured by two basic methods: (i) heat pulse and (ii) heat balance.

In heat pulse method, applied by Cohen et al. (1988) in herbaceous plants, sap flow is estimated by measuring heat velocity, stem area and xylem conductive area. This method seems to be inaccurate at a low transpiration rate (Cohen et al., 1993); moreover, it needs calibration for every crop species.

Therefore, the most popular sap flow method is based on the heat balance method. It is based on the concepts proposed by Cermak et al. (1973), Sakuratani (1981), Granier (1985) and Steinberg et al., (1990). The plant transpiration can be estimated by determining the sap mass flow; this is done using gauges that are attached to or inserted in the plant stem. For the heat balance method, a heater element is placed around the plant stem to provide energy to the system. Thermocouples are used to determine how much heat is lost by conduction up, down and radially in the stem from the heater element. The difference between the heat input and these losses is assumed to be dissipated by convection with the sap flow up the stem and may be directly related to water flow (Kjelgaard et al., 1997). The mass flow rate $F$ (g t$^{-1}$) is expressed by the relationship:

$$F = \frac{Q_h - Q_v - Q_r}{c_w \cdot \delta T}$$  \hspace{1cm} (14)

where $Q_h$ is input heat, $Q_v$ is vertical conductive heat, $Q_r$ is radial heat loss to environment, $c_w$ (J g$^{-1}$ K$^{-1}$) is specific heat of water and $\delta T$ is the temperature difference between the upstream and downstream thermocouples. With the commercial instrumentation, the mass flow rate per plant can be measured at hourly scale.

The micro-engineering problems of the heat emitters and detectors discussed by Cohen et al. (1981) were recently overcome by Kjelgaard et al. (1997), by using constant heat input instead of variable heat input and using surface-mounted thermocouples instead of the inserted ones. These gauges do not disturb the plant root medium but there is still biometric difficulty in extrapolating from one or more plants to the canopy. Furthermore, in spite of the fact that the heat balance method is more dependent upon the sap flow rate than upon stem anatomy (Zhang and Kirkham, 1995), at low flow rates (often the case in a semi-arid and arid environment when the crops are under water stress) important differences were observed between the sap flow determined by the gauges and plant water losses observed by a reference method (Kjelgaard et al., 1997). The best results are obtained at a daily scale, mostly on trees.

For the sap flow method it is necessary do a scaling up of the transpiration measurement from plant to field scale. This is possible only if the canopy structure and the spatial variability of the plants characteristics (density, height, LAI) are precisely known.

Recently, Grime and Sinclair (1999) reported sources of error for the constant power stem heat balance method when commercial gauges are used. Their analysis demonstrated that measurement errors in determining the diurnal pattern of transpiration could be due to the heat storage in
the stem and the pattern of ambient temperature. Moreover, the determining of the sheath conductance by the night-time sap flow measurement can be affected by error due to the stem variation during the season.

These errors are highly dependent on operating conditions and can be minimised by following appropriate recommendations.

Problems linked to the death of the fresh tissues around the heated area, due to silicone application, can be minimised by reducing the period of continuous measurement to 1–2 weeks (Wiltshire et al., 1995).

The sap flow method approaches only transpiration measurements, neglecting soil evaporation. Nevertheless, under the Mediterranean climate evaporation from soil can be a very important fraction (up to 20% of total evapotranspiration) of the soil–plant–atmosphere water budget (Brutsaert, 1982; Klocke et al., 1985a,b).

3.3.2. Chambers system method

The chambers to rapidly measure ET were described for the first time by Reicosky and Peters (1977). They consist of aluminium conduits covered with Mylar, polyethylene or other plastic or glass films; the air within the chamber was mixed continuously with strategically located fans. The first version was portable (by means of a tractor, for example) and ET rate was calculated by a psychrometer before the chamber was lowered on the plot and 1 min later, as latent heat storage. The volume of the chamber can be easily adapted to a herbaceous field crop and the accuracy was ≈10%, compared with weighing lysimeters (Reicosky et al., 1983).

This system was used by Reicosky (1985) to measure, on a daily scale, ET on several different treatments more easily than the weighing lysimeter, but it is not suitable for long term ET measurements. Furthermore, more serious errors may be introduced if chamber results were extrapolated over time as well as spatially (Livingston and Hutchinson, 1994).

Under a semi-arid climate, Dugas et al. (1991) demonstrated that portable chambers, equipped with an infrared analyser (BINOS, Inficon Leybold-Heraeus, NY) for measuring vapour density differences in differential mode, are not representative of the field, particularly on its margins, where ET measurements were affected by surrounding conditions. Furthermore, the last improved version of portable chambers can be very expensive, due to the high costs for key components (data acquisition system and infra-red analyser) and accurate chamber manufacturing that may require some heavy engineering.

Recent developments of portable chamber systems can be found in Wagner and Reicosky (1996); here the equipment also provides CO₂ measurements, but its complexity and costs limit its use for research to a small temporal scale.

The chambers are also suitable for research studies on orchard crops (Katerji et al., 1994). The most serious problems of almost all chambers are related to the modification of the microclimate during the measurement period. First, the solar radiation balance, since the chambers were designed for simultaneous measurement of both CO₂ and water vapour exchanges (Denmead, 1984). These dual requirements are not often compatible because wall materials are usually selected for high transmission in the short wavelengths, neglecting the long-wave exchange (≈20% of the incoming short-wave radiation). Second, the air temperature: its rapid increase inside the chambers could alter the biological control of leaves transpiration process. Finally, the wind speed: inside the chamber it could be strongly reduced with direct consequences on ET measurement accuracy.

4. Actual crop evapotranspiration modelling

Actual ET can be estimated by means of more or less complex models: the accuracy of ET estimation is proportional to the degree of empiricism in the used model or sub-models. Thus, we can categorise the ET estimation into three groups of methods:
1. Methods based on analytical modelling of evapotranspiration;
2. Methods in which actual crop ET is deducted from the evapotranspiration of a reference surface;

4.1. ET analytical models

One-dimensional equations based on aerodynamic theory and energy balance — for this reason called combination models (Penman, 1948; Monteith, 1965, 1973) — have proved very useful in the actual crop ET estimation, because they take into account both the canopy properties and meteorological conditions (Szeicz and Long, 1969; Black et al., 1970; Szeicz et al., 1973).

The most widely used form of the combination equation, called Penman-Monteith equation, can be expressed under the form (Allen et al., 1989):

$$\lambda E = \frac{\Delta A + \rho c_p VPD/r_a}{\Delta + \gamma (1 + r_c/r_a)}$$  \hspace{1cm} (15)

In Eq. (15) we can distinguish weather non-parametric and parametric variables:

- weather non-parametric variables: $A = R_n - G$ (W m$^{-2}$) is available energy, $\Delta$ (kPa C$^{-1}$), is the slope of the saturation vapour pressure versus temperature function, VPD (kPa) is air vapour pressure deficit and $\gamma$ (kPa C$^{-1}$) is the psychrometric constant;

- parametric variables: $r_a$ (s m$^{-1}$) is the aerodynamic resistance and $r_c$ (s m$^{-1}$) is the bulk canopy resistance.

All the terms have to be accurately evaluated (Allen, 1996). While the non-parametric data are standard measurable climatic variables, the parametric data are not directly measurable and they need to be modelled. Indeed, the first one ($r_a$) is the aerial boundary layer resistance and describes the role of the interface between canopy and atmosphere in the water vapour transfer; the second one ($r_c$) is the resistance that the canopy opposes to the diffusion of water vapour from inner leaves toward the atmosphere and it is influenced by biological, climatological and agronomical variables. This model is applied to the whole plant community as if it were a single ‘big leaf’ located at the height of virtual momentum absorption (Thom, 1975).

The degree of empiricism of the Penman–Monteith equation (and consequently its success) mainly depends on the accuracy of the estimation of the canopy resistance (Beven, 1979).

Rana and Katerji (1998) demonstrated that, under a semi-arid climate, the canopy resistance plays the major role, so that its modelling is the most critical point of the Penman–Monteith model to estimate actual crop ET under the Mediterranean climate. In particular, their sensitivity analysis on Penman–Monteith model showed that: (i) in well-watered conditions, $r_c$ modelling is sensible to $R_n$ variations in the case of low crops (i.e. reference grass) and it is sensible to vapour pressure deficit variations, both in the case of medium (Fig. 7) and tall crops (Fig. 8); (ii) in water stress conditions, $r_c$ modelling is sensible to the actual crop water status (Figs. 7 and 8).

4.1.1. Methods of determining canopy resistance

One of the most diffuse methods of estimating $r_c$ is that proposed by Szeicz and Long (1969), in which canopy resistance can be determined as a function of daily mean stomatal resistance of the single leaves ($r_s$) and leaf area index of the leaves effective in transpiration ($\text{LAI}_{\text{eff}}$). These authors assumed that only the surfacial layer of the canopy participates effectively to the transpiration. Usually, such a canopy resistance is assumed to be half the full crop LAI:

$$r_c = \frac{r_s}{\text{LAI}_{\text{eff}}}$$  \hspace{1cm} (16)

This equation is suggested to be used for reference crop ET (Allen et al., 1989) and it is also used, under light different form, for field crop ET (Steiner et al., 1991; Stannard, 1993; Howell et al., 1995).

Since the spatial and temporal variability of the soil water status (and, consequently, of the plant water status) is very high and the microclimate to which they are exposed is very different (Demenad, 1984), to obtain sufficiently accurate canopy resistance values by means of this method, it is necessary to realise a great number of measurements of stomatal resistance in a brief interval time.

Furthermore, from the theoretical and practical point of view, the problems of scaling-up the canopy resistance from leaf to canopy is not
An improvement of this estimating method has been proposed by Monteith (1965) and tested and applied by Katerji and Perrier (1985). In this method, the canopy was divided into layers and canopy resistance was estimated starting from stomatal resistance measured layer by layer and weighted with the LAI of each layer. The expression of the canopy conductance \( g_c \) is given, in this case, by the relationship:

\[
g_c = \sum_i g_{ci} \cdot \text{LAI}_i
\]

where \( g_{ci} \) is the stomatal conductance measured on the leaves belonging to the layer characterised by LAI\(_i\). Usually \( i \) is equal to 2 or 3, following the crop growth: an example is given in Table 2 for an alfalfa crop divided into three layers. This approach is difficult and hard to carry out, giving \( r_c \) values not always accurate. Nevertheless, the estimation of field crop ET can be acceptable, at least for well-watered crops (Katerji and Perrier, 1985).

The most complete model of \( r_c \) should have the following general form (Stewart, 1988, 1989; Itier, 1996):

\[
r_c = f(\text{LAI}, R_g, \text{VPD}, T, \text{crop water status})
\]

where \( R_g \) is global radiation.

Actually, the better way to evaluate the crop water status is by means of measurements of leaf water potential \( \Psi_l \) or root-sources abscisic acid (ABA) (Zhang et al. 1987; Zhang and Davies, 1989). As previously stated, the plant water status varies almost instantaneously, so that to have

Fig. 7. The sensitivity coefficients as function of predawn leaf water potential (PLWP, MPa) for grain sorghum; \( A \) is available energy, \( ra \) is aerodynamic resistance, \( rc \) is canopy resistance and \( \text{VPD} \) is vapour pressure deficit (Rana and Katerji, 1998).
acceptable accuracy of the actual crop water status, a great number of $\psi_f$ or ABA measurements has to be carried out during the entire day. Hence, this method is not suitable for practical application in a field.

In order to avoid the problem of a large number of punctual measurements, a number of models as expressed by the general relationship Eq. (18), have been developed for dry conditions. For example, Hatfield (1985) modelled $r_c$ in the function of global radiation and available soil water by means of an experimental function. Fuchs et

Table 2

<table>
<thead>
<tr>
<th>I</th>
<th>Canopy layer</th>
<th>Stomatal resistance and S.E. (s m$^{-1}$)</th>
<th>LAI and S.E.</th>
<th>Canopy conductance (mm s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Upper side</td>
<td>Lower side</td>
<td></td>
</tr>
<tr>
<td>1</td>
<td>Top</td>
<td>117 ± 29</td>
<td>115 ± 29</td>
<td>1.75 ± 0.35</td>
</tr>
<tr>
<td>2</td>
<td>Medium</td>
<td>199 ± 70</td>
<td>559 ± 196</td>
<td>2.1 ± 0.42</td>
</tr>
<tr>
<td>3</td>
<td>Bottom</td>
<td>1044 ± 365</td>
<td>1200 ± 420</td>
<td>0.85 ± 0.17</td>
</tr>
<tr>
<td></td>
<td>Canopy</td>
<td></td>
<td></td>
<td>4.7</td>
</tr>
</tbody>
</table>

*The resulting canopy resistance is 22 s m$^{-1}$ (Saugier and Katerji, 1991).
al. (1987) stated that \( r_c \) can be modelled by means of photosynthetic active radiation (PAR) and soil moisture deficit, but this model is only valid for mild to medium stress. Again, regressive experimental functions were used by Jarvis (1976) to model \( r_c \) in the function of LAI, \( R_g \), VPD, \( T \) and soil water potential. Unfortunately, all these functions need to be locally calibrated per crop and, often, they need also a time calibration, i.e. they are not valid for the whole growth season (Stewart, 1988).

An original approach to take into account the crop water status to estimate ET by means of Penman–Monteith model was presented by Rana et al. (1997b), in which \( r_c \) is modelled as a function of ‘predawn leaf water potential’, \( \Psi_b \). This parameter represents the crop water status and it does not need to be measured instantaneously, but just once a day. The application of such a model in semi-arid environments gives quite good results. Such \( r_c \) model is valid for crops growing under semi-arid climate (i.e. also for crops under water stress) and needs to be calibrated only ‘per crop’ so that it can be generalised (Rana et al., 1997c). It is possible to use the transpirable soil water as input of the model instead of \( \Psi_b \), but in this case the model looses its generality and must be locally calibrated (Rana et al., 1997a).

4.2. ET empirical models

By these methods, the water consumption of crops is estimated as a fraction of the reference evapotranspiration (ET\(_0\)):

\[
ET = K_c \cdot ET_0 \tag{19}
\]

where \( K_c \) is the experimentally derived crop coefficient and \( ET_0 \) is the maximum evapotranspiration; the latter can be evaluated (i) on a reference crop or (ii) on free water in a pan. The accuracy of such an estimation depends on:
- the reference chosen (grass meadow or free water in a standard pan);
- the method used to evaluate reference ET (measurement or modelling);
- the method used to evaluate the crop coefficient \( K_c \).

Moreover, the data necessary for the determination of reference evapotranspiration \( ET_0 \) are usually collected in standard agrometeorological stations. These stations are usually installed to be representative of the catchment, i.e. an area of several squared kilometres in extension.

4.2.1. \( ET_0 \) reference crop determination

In this case, \( ET_0 \) is the water consumed by a standard crop. In order to make procedures and results comparable world-wide, a well adapted variety of clipped grass has been chosen. It must be 8–16 cm high, actively growing and in well-watered conditions, subjected to the same weather as the crop whose water consumption is to be estimated. \( ET_0 \) can be again measured (e.g. by means of a weighing lysimeter) or estimated. Therefore the comments and observations made for actual crop ET are still valid for \( ET_0 \) determination. In the following, we analysed the more used models of \( ET_0 \).

A semi-empirical approach to estimate \( ET_0 \) was proposed by Penman (1956) as:

\[
ET_0 = \frac{\Delta R_n + \gamma EA}{\Delta + \gamma} \tag{20}
\]

where

\[
EA = f(u)(e - e^*) \tag{21}
\]

and

\[
f(u) = 1 + 0.54 u_2 \tag{22}
\]

with \( u_2 \), wind speed measured 2 m above the soil surface.

The empiricism of this formula is into the wind function \( f(u) \). Actually, this function is just a linear regression adjustment to take into account the differences between estimated and observed \( ET_0 \) for grass grown in the UK. Therefore, it is a pseudo exchange coefficient, which is supposed to take into account the ET grass regulation. This observation explains why it is necessary to adapt the function \( f(u) \) to each site, in order to correctly apply the Penman formula in the Mediterranean regions. Furthermore, it is clear that the wind function decreases going from North toward South (Fig. 9) and these results point out the dependence of canopy resistance on the vapour
pressure deficit, which increases going from North to South Europe (Choisnel, 1988).

Today the most used and suggested method to evaluate reference ET is based on the Penman–Monteith model, therefore, having the same problems encountered in actual crop ET estimation regarding \( r_c \) modelling.

Rana et al. (1994) proposed an ET\(_0\) estimated by means of the Penman–Monteith model, in which canopy resistance is analytically modelled. This model was tested in Mediterranean regions, but it is valid in humid environments too. They demonstrated that \( r_c \) varies with the climate and it is sensitive to any agronomic practice (irrigation date, grass cutting, phytopathological situation); in Fig. 10 hourly \( r_c \) values are shown before and after grass cutting and irrigation.

Recently, Todorovic (1997) proposed a variable canopy resistance, both at hourly and daily scale, modelled in the function of standard meteorological parameters. It was shown that ET\(_0\) calculated by the Penman–Monteith model with such a resistance works well under different climates.

Several authors proposed an ET\(_0\) Penman–Monteith formula using constant value of \( r_c \) (Allen et al., 1989). Steduto et al. (1996) tested such ET\(_0\) estimates with constant \( r_c \) in several Mediterranean regions. Their results demonstrated that this approach does not have good performances in all the experimental sites.

Other simpler methods to estimate reference crop ET are based on statistical–empirical formulas (an exhaustive review of these methods can be found in Jensen et al. (1990)). These methods can

---

**Fig. 9.** Comparison among four \( f(u) \) wind functions in the Penman equation: Penman is \( f(u) = 0.26(1 + 0.54 \ u) \); Businger is \( f(u) = 0.37 \ u \); Doorenbos and Pruitt is \( f(u) = 0.27(1 + 0.86 \ u) \); Seguin is \( f(u) = 0.37 \ u^{0.88} \); Losavio et al. is \( f(u) = 0.12 \ u^{0.67} \) (Losavio et al., 1992).
be used very easily from a practical point of view, above all in rural lands. However, their empiricism may lead to very inaccurate estimations, as clearly shown by Ibrahim (1996), in arid environments (Table 3).

4.2.2. The reference $ET_0$ calculated by the pan evaporation

Pan evaporation data can be used to estimate reference ET, using a simple proportional relationship:

$$ET_0 = K_p \cdot E_{pan}$$  \hspace{1cm} (23)

where $K_p$ is dependent on the type of pan involved, the pan environment in relation to nearby surfaces and the climate. Doorenbos and Pruitt (1977) provided detailed guidelines for using pan data to estimate reference $ET_0$. In the case of a pan surrounded by short green grass, $K_p$ ranges between 0.4 and 0.85. In semi-arid environments its mean value is $\approx 0.7$ (Jensen et al., 1990). The value of $K_p$ to be adopted is strongly dependent on the upwind fetch and on the local advection.

An attempt to model the pan coefficient, $K_p$, has been made by Perrier and Hallaire (1979a,b), on the basis of experimental pan data measured in a humid–tropical area (Baldy, 1978). They expressed $K_p$ in the function of climatic variables and a coefficient $\beta$, the ratio between the exchange coefficient of the pan and the wind function $f(u)$ of the Penman’s formula, as

$$K_p = \frac{1 + \alpha(1 - RH)}{1 + \beta f(u)(1 - RH)}$$  \hspace{1cm} (24)

with RH relative air humidity and $\alpha$ a coefficient, function of air temperature, net radiation and wind speed. This latter coefficient is dependent on

Table 3
Seasonal reference evapotranspiration ($ET_0$) in 3 years and average percent deviation from the mean value in the north central area of the Nile delta region (Egypt) (Ibrahim, 1996).

<table>
<thead>
<tr>
<th>Method</th>
<th>Seasonal $ET_0$ cm</th>
<th>Average seasonal $ET_0$ cm</th>
<th>Percent deviation from mean value</th>
</tr>
</thead>
<tbody>
<tr>
<td>1979 1980 1981</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(a) Blaney–Cridgle</td>
<td>63.2 61.81 60.31</td>
<td>61.71</td>
<td>-4.37</td>
</tr>
<tr>
<td>(b) Radiation</td>
<td>66.78 63.20 66.82</td>
<td>65.60</td>
<td>+1.66</td>
</tr>
<tr>
<td>(c) Modified Penman</td>
<td>62.82 60.60 60.24</td>
<td>61.22</td>
<td>-5.13</td>
</tr>
<tr>
<td>(d) Rijtema</td>
<td>81.64 81.35 82.92</td>
<td>81.97</td>
<td>+27.03</td>
</tr>
<tr>
<td>(e) Thornthwaite</td>
<td>52.16 49.73 48.00</td>
<td>49.96</td>
<td>-22.59</td>
</tr>
<tr>
<td>(f) Jensen and Haise</td>
<td>86.75 80.50 79.41</td>
<td>82.08</td>
<td>+27.20</td>
</tr>
<tr>
<td>(g) Evaporation pan</td>
<td>56.75 54.14 49.24</td>
<td>53.38</td>
<td>-17.28</td>
</tr>
<tr>
<td>(h) Turc</td>
<td>63.47 58.93 58.45</td>
<td>60.28</td>
<td>-6.59</td>
</tr>
<tr>
<td>Average</td>
<td>66.62 63.78 63.17</td>
<td>64.53</td>
<td></td>
</tr>
</tbody>
</table>
4.2.3. The crop coefficient $K_c$

The crop coefficient represents an integration of the effects that distinguish the crop from the reference ET; many such $K_c$ are reported in the literature (e.g. Doorenbos and Pruitt, 1977; Allen et al., 1998) usually derived from soil water balance experiments.

Crop coefficient can be improved for estimating the effects of evaporation from wet soil on $K_c$ on a daily basis (Wright, 1982). In this case, the crop coefficient can be expressed by the equation:

$$K_c = K_s \cdot K_{cb} + K_e$$  \hspace{1cm} (25)

where $K_s$ is a stress reduction coefficient ($0 \leq K_s \leq 1$), $K_{cb}$ is the basal crop coefficient ($0 \leq K_{cb} \leq 1.4$) and represents the ratio of ET and ET$_0$ under conditions when the soil surface is dry, but where the soil water content of the root zone is adequate to sustain full plant transpiration; $K_e$ is a soil water evaporation coefficient ($0 \leq K_e \leq 1.4$). It can be experimentally calculated in the function of soil water storage, $\Delta S$ and depends on the amount of soil available to the plants’ roots. In fact, in open soil systems, when the conditions are favourable for root system development, $K_s$ is almost constant, also when soil humidity decreases considerably, because of an appreciable contribution of the non-rooted soil layer to the water balance. In closed soil systems (pots, for example) $K_s$ is variable following the soil humidity and it begins to decrease appreciably for values of soil water reserve $\approx 60–70\%$ of available soil water to transpiration. This $K_s$ behaviour is reported in Fig. 12, in which the ratio of actual to potential ET of a well-irrigated corn crop is reported in the function of the available soil water, for the above mentioned two conditions (open and closed soils).

In order to avoid the difficulties linked to the available soil water conditions, it is possible to evaluate $K_s$ by means of the predawn leaf water potential, $\Psi_{lb}$. Itier et al. (1992) demonstrated that this relationship (and $K_s$ as a consequence) has general validity, since it does not depend on the soil type and site.

The methods to estimate crop coefficients and reference ET have been recently modified by Allen et al. (1994a), Allen et al. (1994b) and Allen et al. (1996).
The crop coefficient plays an essential role in practice (Pereira et al., 1999) and it has been widely used to estimate actual ET for irrigation scheduling purposes. However, it can be subject to serious criticism regarding the meaning and the use of crop coefficient. Besides obvious variations among different crops, empirical crop coefficients were shown to be affected by crop development and weather conditions (De Bruin, 1987). Furthermore, Stanghellini et al. (1990) demonstrated that no coefficients should be expected to vary according to the conditions of both climate and crop stage under which they are derived. Moreover, these researchers stated that the crop requirements based on a same value of the crop coefficient do not have the same accuracy for different months, different seasons, and, above all at different sites. This inaccuracy of the $K_c$ approach can be found in the discrepancies between local calculated $K_c$ and crop coefficients reported in literature (e.g. Rana et al., 1990; Vasic et al., 1996).

### 4.2.4. Soil water balance modelling

An exhaustive survey of the most diffuse models for soil water balance can be found in de Jong and Bootsma (1995) and Leenhardt et al. (1995). Two classes of models are generally retained for the simulation of soil water balance: (i) mechanistic models and (ii) analogue (or water reservoirs) models.

In the mechanistic approach, the water flux in the soil is controlled by the existence of soil water potential gradients, by means of Darcy’s law and continuity principle. The equations are usually solved by different methods, all involving the splitting of the soil in more or less small layers (de Jong, 1981; Feddes et al., 1988). The difficulties in the application of these models are linked to (i) the accuracy of the used pedo-transfer functions for estimating the water transfer and (ii) the procedures used for estimating the boundary conditions of the soil–plant–atmosphere system.

In the analogue approach, the soil is treated as a collection of water reservoirs, filled by rainfall or irrigation and emptied by evapotranspiration and drainage. They can be based on the two following principles (Lhomme and Katerji, 1991):

1. determination of soil water storage $D_S$ as function of the soil and roots depth (e.g. Lhomme and Eldin, 1985; Brisson et al., 1992)
2. the split of soil water in readily transpirable soil water (RTSW) and total TSW (e.g. Fréteaud et al., 1987). The stress coefficient $K_s$ is supposed to be approximately $=1$ in the RTSW range, then it decreases as the water reserve is in the TSW range.

In a previous paragraph we have analysed the difficulties in determining the soil water storage and the evolution of $K_s$ in function of $D_S$, particularly under a Mediterranean climate. Effectively, Ben Nouna (1995) showed that, in semi-arid regions with closed system soils, the water reservoir approach can be very useful in estimating actual ET only if very accurate measurements of $D_S$ can be realised (Fig. 13).

### 5. ET evaluation in advection regime

All the previously presented methods to measure and/or estimate actual crop ET are based on...
a fundamental critical hypothesis that the considered systems were conservative, i.e. the profiles of air temperature, water vapour concentration and wind speed are constant along the horizontal axe X (e.g. Itier and Perrier, 1976; Stull, 1988).

In the Mediterranean regions, the scarcity of available water resources leads to irrigated lands in a selective way, with the creation of areas characterised by strong discontinuity of water vapour concentration at surface level, also for similar crops. The air thermodynamic characteristics are strongly influenced by the surface conditions, so that the air passing from a dry field toward an irrigated field is cooled as well as its humidity increases. Hence, near the boundary of the irrigated field, ET rapidly increases, then slowly decreases along the wind direction following the thermodynamic characteristics of the air coupled with the vegetation surface.

In general, the latent heat flux density measured or estimated in one point of the plot is, actually, the sum of two terms: (i) a term translating the equilibrium between the evaporative demand of the air determined by the Penman model and crop water conditions; (ii) an additional flux, more or less important, depending on the lateral transfer of energy by advection. From an energetic point of view, latent best flux depends on the modifications of the vapour and thermal characteristics of the air passing through the considered surface.

The theoretical studies on local advection (Philip 1959; Taylor, 1970; Itier and Perrier, 1976; Itier et al., 1994) showed that the advective flux \( \Phi_a \) depends on: (i) the fetch \( F \) (m); (ii) the roughness length of the crop, \( z_0 \) (m); (iii) the wind speed through the friction velocity \( u^* \) (m s\(^{-1}\)); and (iv) the temperature difference between the dry and irrigated fields, \( \Delta T \) (°C).

From the operational point of view such a situation in which an advection regime is verified leads to two main consequences:

1. In order to have a correct representativeness of ET, it is necessary to correct a posteriori the evapotranspiration measured by means of weighing lysimeters, soil water balance, sap flow method and chambers system method when the fetch is not large enough to avoid advection effects. A possible simple correction was proposed by Itier et al. (1978) as:

\[
\Phi_a = 360 \cdot u^* \cdot \Delta T \cdot \frac{6 z_0}{\sqrt{z_0 F}}
\]

This correction can be of 1–2 mm per day for temperature differences between dry and irrigated field of 5–10°C (Itier et al., 1978).

2. The micrometeorological methods must be applied in order to make the measurements inside the equilibrium fully adapted boundary layer. The fetch value assuring an acceptable adapted layer for micrometeorological is reported by Wieringa, (1993) as:

\[
F \approx 2z_0 \left[ \frac{10z}{z_0} \left( \ln \frac{10z}{z_0} - 1 \right) \right]
\]

where \( z \) is the height of the top sensor.

Regarding the Bowen ratio method, it is well known that it can underestimate ET under strong regional advective conditions (Blad and Rosenberg, 1974). This implies an extensive fetch in the upwind direction for the air flowing over the surface of at least 100 times the maximum height of measurements, but Heilman et al. (1989) demonstrated that the Bowen ratio method can be successfully applied up to fetch-to-height ratios as low as 20:1.

6. Conclusions

In Mediterranean areas submitted to arid and semi-arid climates, ET ranges over a large interval depending on water regimes (irrigated or non-irrigated fields). Among the other weather variables, air temperature and humidity experience a large gap between night and day and between winter and summer; the wind is very variable moving from the coast to internal areas. So that, in general, we can state that all the weather variables range in a very large interval of values. Moreover, the variation in one parameter immediately influences all the other variables that are mutually linked. These facts make difficult to correctly evaluate the crop actual evapotranspiration.

In this review, we have presented the most important methods of measuring and modelling actual evapotranspiration in arid and semi-arid
environments. Some methods are more suitable than others in terms of convenience, accuracy or cost for the measurement of ET at a particular spatial and over a particular time scale. In Table 4 the classification of the presented ET methods by space and time scale is shown.

We tried to give, for each method, the advantages and disadvantages for their use in the Mediterranean regions; the summary of this analysis for the measurement methods is reported in Table 5.

In summary, two major observations are possible:

Table 4
Classification by space and time of ET methods to measure or model actual evapotranspiration, adapted by Stewart (1984)

<table>
<thead>
<tr>
<th></th>
<th>Minute</th>
<th>Hour</th>
<th>Day</th>
<th>Month</th>
<th>Growth season</th>
<th>Year</th>
</tr>
</thead>
<tbody>
<tr>
<td>Catchment</td>
<td>--------</td>
<td>--------</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Uniform area</td>
<td>--------</td>
<td>--------</td>
<td></td>
<td></td>
<td>—</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Micrometeorological method —</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td></td>
<td>--------</td>
<td>Crop coefficient method —</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td></td>
<td>Soil water balance —</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td></td>
<td>Soil water balance with TDR —</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td></td>
<td>Penman–Monteith model —</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>Group of plants</td>
<td>Weighting lysimeter; chambers system —</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>Plant</td>
<td>Sap flow —</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>—</td>
</tr>
</tbody>
</table>

Table 5
Summary of the advantages and disadvantages of the ET measurement methods

<table>
<thead>
<tr>
<th>Measurement method</th>
<th>Advantages</th>
<th>Disadvantages</th>
</tr>
</thead>
<tbody>
<tr>
<td>Soil water balance</td>
<td>Soil moisture simple to be evaluated with gravimetric method Not expensive if the gravimetric method is used</td>
<td>Large spatial variability Difficult to be applied when the drainage and capillary rising are important Difficult to measure soil moisture in cracked soils</td>
</tr>
<tr>
<td>Weighing lysimeter</td>
<td>Direct method</td>
<td>Fixed Difficult maintenance It could be not representative of the plot area Expensive</td>
</tr>
<tr>
<td>Energy balance/ Bowen ratio</td>
<td>Simple sensors to be installed Suitable also for tall crops It can be also used when the fetch is 20:1 Not very expensive if psychrometers are used</td>
<td>Difficult to have correct measurement of the wet temperature if psychrometers are used The sensors need to be inverted to reduce bias Difficult maintenance</td>
</tr>
<tr>
<td>Aerodynamic</td>
<td>Simple sensors to be installed</td>
<td>It needs to be corrected for the stability Not suitable for tall crops</td>
</tr>
<tr>
<td>Eddy covariance</td>
<td>Direct method with fast hygrometer</td>
<td>Delicate sensors Difficult software for data acquisition Hygrometer very delicate expensive Difficult scaling-up</td>
</tr>
<tr>
<td>Sap flow</td>
<td>Suitable for small plots It takes into account the variability among plants</td>
<td>The gauges need to be displaced every 1–2 weeks The soil evaporation is neglected</td>
</tr>
<tr>
<td>Chambers method</td>
<td>Suitable for small plots It can be used also for detecting emissions of different gases</td>
<td>It modifies the microclimate Difficult scaling-up</td>
</tr>
</tbody>
</table>
1. The ET measurement methods are based on concepts which can be critical under semi-arid and arid environments for several reasons: (i) representativeness (the weighing lysimeter, for example); (ii) instrumentation (air humidity sensors for example); (iii) microclimate (advection regime); and (iv) hypothesis of applicability (the simplified aerodynamic method for example). Thus, in order to establish the degree of accuracy of the obtained ET measurement and the validity of a method, it is necessary to consider all these parameters.

2. The estimation methods based on an analytical approach could be very accurate but, usually, they are not practical enough. The more operational estimation methods (crop coefficient approach, ET0 calculation, water balance modelling) can be generally affected by very large errors.

Nowadays, a new research trend is being developed in order to integrate these two approaches of actual evapotranspiration estimation. These new models tend to introduce into the analytical approach an acceptable degree of empiricism (Allen et al., 1989; Rana et al., 1994; Pereira et al., 1999). We think that such methods should be accepted and encouraged in the future, since these approaches can greatly improve the water management at farm level, one of the greatest problems of the Mediterranean countries.

References


