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Water Vapor Flux in Agroecosystems Methods and Models Review

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1. Introduction

The water vapor flux in agroecosystems is the second largest component in the hydrologycal cycle. Water vapor flux or evapotranspiration (ET) from the vegetation to the atmosphere is a widely studied variable throughout the world. ET is important for determining the water requirements for the crops, climatic characterization, and for water management. The estimation of ET from vegetated areas is a basic tool to compute water balances and to estimate water availability and requirements. During the last sixty year several methods and models to measure the water flux in agroecosystems have been developed. The aim of this chapter is to provide a literature review on the subject, and provide an overview of methods and models developed which are widely used to estimate and/or measure ET in agroecosystems.

Evapotranspiration constitutes an important component of the water fluxes of our hydrosphere and atmosphere (Conroy *et al*., 2003), and is a widely studied variable throughout the world, due to it applicability in various disciplines, such as hydrology, climatology, and agricultural science. Pereira *et al*., (1996) has reported that the estimation of ET from vegetated areas is a basic tool for computing water balances and to estimate water availability and requirements for plants. Measurement of ET is needed for many applications in agriculture, hydrology and meteorology (Suleiman & Crago, 2004). ET is a major component of the hydrologic water budget, but one of the least understood (Wilson et al., 1992). ET permits the return of water to the atmosphere and induces the formation of clouds, as part of a never-ending cycle. ET also permits the movement of water and nutrients within the plant; water moving from the soil into the root hairs, and then to the plant leaves.

ET is a complicated process because it is the product of the different processes, such as evaporation of water from the soil, and water intercepted by the canopy, and transpiration from plant leaves. Physiological, soil and climatic variables are involved in these processes. Symons in 1867 described evaporation as "*...the most desperate art of the desperate science of meteorology*" (Monteith, 1997). The first vapor flux measurements were initiated by Thornthwaite and Holzman in 1930s, but that works was interrupted by World War II (Monteith, 1997). In the late 1940s Penman (1948) published the paper "*Natural Evaporation from open Water, Bare Soil and Grass*" in which he combined a thermodynamic equation for the surface heat balance and an aerodynamic equation for vapor transfer. The "Penman equation" is one of the most widely used equations in the world. The equation was later modified by Monteith (1965; 1981) and is widely known as the "*The Penman-Monteith Model*". It is also necessary to introduce a review of the work of Bowen, who in 1926 published the relationship between the sensible and latent heat fluxes, which is known as the "*Bowen ratio*". Measurement of the water vapour flux became a common practice by means of the "*Bowen ratio energy balance method*" (Tanner, 1960).

Allen et al. (1998) separated the factors that affect the ET into three groups: *a) Weather parameters*, such as radiation, air temperature, humidity and wind speed. The evapotranspirational power of the atmosphere is expressed by the reference crop evapotranspiration (ET_o) as the Penman-Monteith (FAO-56), or using direct measurements of pan evaporation data (Doorenbos & Pruit, 1977), or using other empirical equations; *b) Crop factors* such as the crop type, variety and developmental stage should be considered when assessing the ET from crops grown in large, well-managed fields. Differences in resistance to transpiration, crop height, crop roughness, reflection, ground cover and crop rooting characteristics result in different ET levels in different types of crops under identical environmental conditions. Crop ET under standard conditions (ET_c) refers to excellent management and environmental conditions, and achieves full production under given climatic conditions (equation 2); and *c) Management and environmental conditions* (ET_{cadj}). Factors such crop water stress, soil salinity, poor land fertility, limited applications of fertilizers, the presence of hard or impermeable soil horizons, the absence the control of disease and pest and poor soil management may limit the crop development etc., and reduce the ET, (ETc_{adj} equation 3).

One of the most common and fairly reliable techniques for estimating ET_0 is using evaporation pan data when adjustments are made for the pan environment (Grismer et al. 2002) using the pan evaporation and the pan coefficient (K_p).

$$ET_o = K_p . E_p \tag{1}$$

Where E_p is the pan evaporation (mmday^{-1}), and K_p is the pan coefficient, and depends on location, so it is important to know or calculate this coefficient before calculating the ET_o. Allen et al., (1998) gave a methodology to know or calculate *Kp,* and is essentially a correction factor that depends on the prevailing upwind fetch distance, average daily wind speed, and relative humidity conditions associated with the siting of the evaporation pan (Doorenbos & Pruitt 1977)

2. Crop water flux using single crop coefficients- The FAO approach

The United Nation Food and Agricultural Organization (FAO) is also well knew as the "*Two steps method*", which is very useful for single crops and when "reference" conditions are available (i.e., no crop water stress). In this case, crop evapotranspiration (ET_c) can be estimated using equation 2 (Doorenbos & Pruitt 1977; Allen at al., 1998):

$$ET_c = K_c . ET_o \tag{2}$$

where K_c is the coefficient expressing the ratio between the crop and reference ET for a grass surface. The crop coefficient can be expressed as a single coefficient, or it can be split into two factors, one describing the affect of evaporation and the other the affect of transpiration.

As soil evaporation may fluctuate daily, as a result of rainfall and/or irrigation, the single crop coefficient expresses only the time-average (multi-day) effects of crop ET, and has been considered within four distinct stages of growth (see. FAO,56. Allen et al., 1998). When stress conditions exist, the effects can be accounted for by a crop water stress coefficient (K_s) as follows:

$$ET_{cadj.} = K_s.\ K_c.\ ET_o \quad (3)$$

2.1 Crop coefficients

Although a number of ET_c estimation techniques are available, the crop coefficient (K_c) approach has emerged as the most widely used method for irrigation scheduling (Hunsaker et al., 2002). As ET is not only a function of the climatic factors, the crop coefficients can include conditions related to the crop development (K_c), and non-standard conditions (K_s).
The K_c is the application of two concepts: a) crop transpiration represented by the basal crop coefficient (K_{cb}), and b) the soil evaporation K_e (Allen et al., 1998) as follow:

$$K_c = K_{cb} + K_e \quad (4)$$

K_c is an empirical ratio between ET_c and ET_o over grass or alfalfa, based on historic measurements. A K_c curve is constructed for an entire crop growing season, and which attempts to relate the daily water use rate of the specific crop to that of the reference crop (Hunsaker et al., 2002).
The FAO paper # 56 (Allen et al., 1998) provided detailed instructions for calculating these coefficients. For limited soil water conditions, the fractional reduction of K_c by K_s depends on the crop, soil water content, and magnitude of the atmospheric evaporative demand (Doorenbos &Pruitt, 1977).
The value for K_c equals K_{cb} for conditions where, first, the soil surface layer is dry (i.e., when $K_e = 0$) and, second the soil water within the root zone is adequate to sustain the full transpiration (non-stressed conditions, i.e., $K_s = 1$). When the available soil water of the root zone becomes low enough to limit potential ET_c, the value of the K_s coefficient is less than 1 (Allen et al., 1998; Hunsaker, 1999, Hunsaker et al., 2002).
The soil evaporation coefficient accounts for the evaporation component of ET_c when the soil surface is wet, following irrigation or rainfall (Allen et al., 1998; Hunsaker et al. 2002). When the available soil water of the root zone become low enough, crop water stress can occur and reduce ET_c. In the FAO-56 procedures, the effects of water stress are accounted for by multiplying K_{cb} (or K_c) by the water stress coefficient (K_s).

$$K_c.\ K_s = (K_{cb}.\ K_s + K_e) = ET_c/ET_o \quad (5)$$

Where $K_s < 1$ when the available soil water is insufficient for the full ET_c and $K_s = 1$ when there is no soil water limitation on ET_c. Thus, to determine K_s, the available soil water within the crop zone for each day needs to be measured or calculated using a soil water balance approach (Hunsaker et al., 2002).
The estimation of K_e using the FAO-56 method, requires the use of the soil field capacity (FC), the permanent wilting point (PWP), total evaporable water (TEW), the fraction of the soil surface wetted (f_w) during each irrigation or rain, and the daily fraction of the soil surface shaded by vegetation (f_c), or conversely the unshaded fraction ($1-f_c$). Hunsaker et al., (2002) reported an exponential relation between $1-f_c$ and height of the Alfalfa crop.

The measurement of K_e and K_{cb} can be made by performing a *daily water balance*, and use of the following equations from FAO Paper 56 (Allen et al., 1998).

$$ET_c = (K_{cb}+K_e)\ ET_o \tag{6}$$

$$K_{cb} = (ET_c/ET_o)-K_e \tag{7}$$

The soil evaporation (E) can be calculated using the equation (8)

$$E = K_e\ ET_o;\ (8) \text{ and } K_e \text{ is equal to } K_e = E/ET_o \tag{9}$$

The soil evaporation (E) can be measured using the water balance (equation 10)

$$E = D_{e,i-1} - (P_i - RO_i) - \frac{I_i}{f_w} + \frac{f_i}{f_{ew}} + Tew_{,i} + DP_{e,i} \tag{10}$$

where: $D_{e,i-1}$ is the cumulative depth of evaporation following complete wetting from the exposed and wetted fraction of the topsoil at the end of day i-1 (mm), P_i is the precipitation on day i (mm); RO_i is precipitation runoff from the surface on day i (mm), I_i is the irrigation depth on day i that infiltrates into the soil (mm), E_i is evaporation on day i (i.e., $E_i=K_e/ET_o$) (mm), $T_{ew,i}$ is depth of transpiration from the exposed and wetted fraction of the soil surface layer on day i (mm), f_w is fraction of soil surface wetted by irrigation (0.01-1), and f_{ew} is the exposed and wetted soil fraction (0.001-1).

The ratio of reference evaporation to reference transpiration depends on the development stage of the leaf canopy expressed as "δ" the dimensionless fraction of incident beam radiation that penetrates the canopy (Cambell &Norman, 1998; mentioned by Zhang et al. 2004).

$$\delta = \exp(-c.LAI) \tag{11}$$

where c is the dimensionless canopy extinction coefficient, and therefore evaporation and transpiration can be calculate how:

$$E_o = \delta.ET_o \tag{12}$$

$$T = (1-\delta).ET_o \tag{13}$$

Hunsaker, (1999) found that ET_c in cotton was higher when the crop was submitted to high depth of irrigation (820-811mm) that when have low depth of irrigation level (747-750mm), similar to the $K_{cb}K_s$ curves, obtaining higher values than the treatment with high frequency (i.e.; $K_{cb}K_s$ =1.5, 90 days after planting) than the low frequency (i.e.; $K_{cb}K_s$ = 1.4, 90 days after planting).

2.2 Limitations in the use of K_c

Katerji & Rana, (2006) reviewed recent literature related to K_c and found differences of ± 40% between K_c values reported in the FAO-56 paper (Allen et al., 1998) and the values experimentally obtained, especially in the mid growth stage. According to the authors, these large differences are attributable to the complexity of the coefficient K_c, which actually integrate several factors: aerodynamic factors linked to the height of the crop, biological factors linked to the growth and senescence of the surfaces leaves, physical factors linked to evaporation from the soil, physical factors linked to the response of the stomata to the vapour pressure deficit and agronomic factors linked to crop management (distance between rows, using mulch, irrigation system, etc.). For this reason K_c values need to be evaluated for local conditions.

The variation in crop development rates between location and year have been expressed as correlations between crop coefficients and indices such as the thermal base index, ground cover, days after emergence or planting, and growth rate (i.e., Wright & Jensen, 1978; Hunsaker, 1999; Brown et al., 2001; Nasab et al., 2004; Hanson & May 2004; Madeiros et al., 2001; Madeiros et al., 2005; and Ramírez, 2007). The K_c is well related with the growing degree grades-GDD and with the fraction of the soil cover by vegetation (f_c) (**Fig. 1**), and depends on the genotype and plant densities (Ramirez, 2007). The equations for two common beans genotypes and two plant densities are:

The equations based on CGDD and f_c for common bean genotype Morales with 13.6 plants.m^{-2} are:

$$K_c = -3x10^{-6} CGDD^2 + 0.0033 CGDD - 0.053; R^2 = 0.76; p < 0.0001 \tag{14}$$

$$K_c = -1.4019 f_c^2 + 2.5652 f_c - 0.2449; R^2 = 0.70; p < 0.0003 \tag{15}$$

The equations based on CGDD and fc for common bean genotype SER 16, with 6.4 plants.m^{-2} are:

$$K_c = -3x10^{-6} CGDD^2 + 0.0034 CGDD - 0.0515; R^2 = 0.60; p < 0.0001 \tag{16}$$

$$K_c = -0.6726 f_c^2 + 1.90086 f_c - 0.2560; R^2 = 0.60; p < 0.0032 \tag{17}$$

2.3 Water stress coefficient (K_s)

The soil water stress coefficient, K_s, is mainly estimated by its relationship to the average soil moisture content or matric potential in a soil layer, and it can usually be estimated by an empirical formula based on soil water content or relative soil water available content (Jensen et al., 1970, cited by Zhang et al., 2004).

The K_s is an important coefficient because it indicates the sensitivity of the crop to water deficit conditions, for example corn grain yield is especially sensitive to moisture stress during tasselling and continuing through grain fill. Roygard et al., (2002) observed that depletion of soil water to the wilting point for 1 or 2 days during tasselling or pollenization reduced yield by 22%. Six to eight days of stress reduced yield by 50%.

Allen et al. (1998), presented the following methodology for estimating K_s:

$$K_s = \frac{TAW - Dr}{TAW - RAW} = \frac{TAW - Dr}{(1-p)TAW} \tag{18}$$

where TAW is total available water and refers to the capacity of the soil to retain water available for plants (mm), Dr is root zone depletion (mm), RAW is the readily available soil water in the root zone (mm), p is the fraction of TAW that the crop can extract from the root zone without suffering water stress.

$$TAW = 1000(\theta_{FC} - \theta_{WP}) Z_t \tag{19}$$

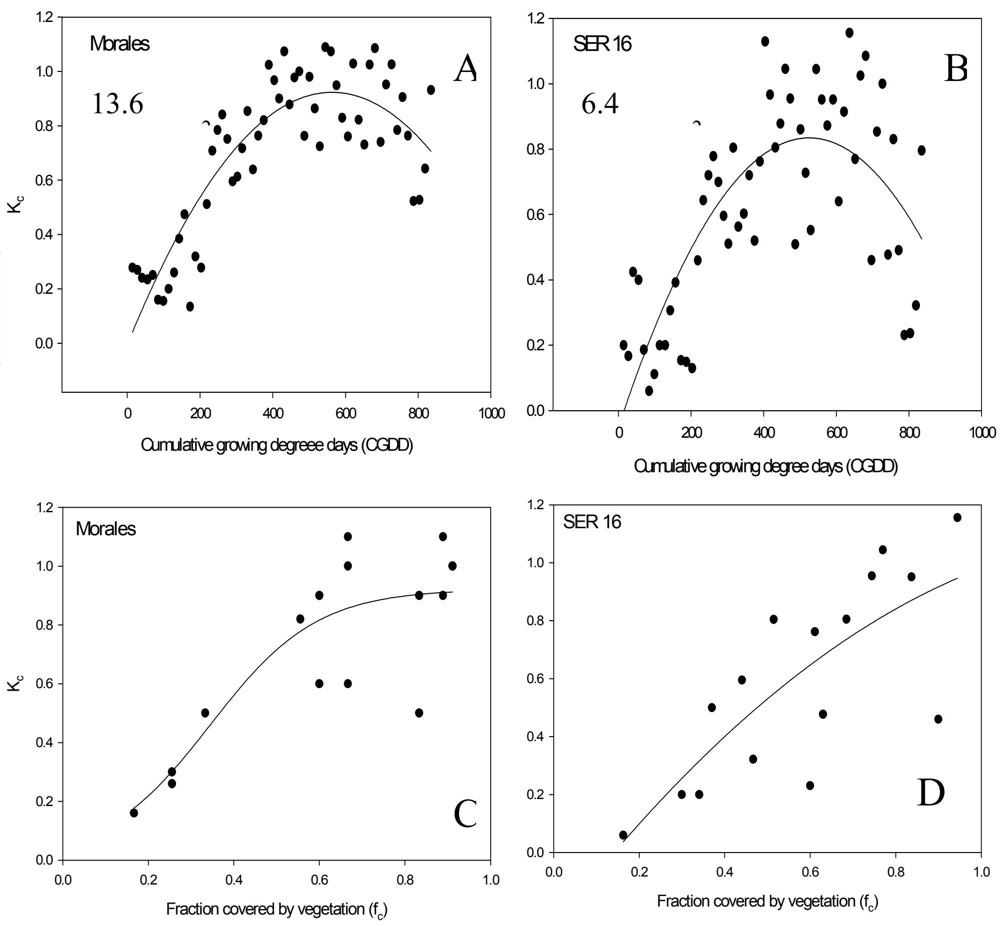

Fig. 1. Crop coefficients (K_c) as related to cumulative growing degree days (CGDD) and fraction covered by vegetation (f_c) for two common bean genotypes: **A**. Morales CGDD vs K_c, **B**. SER 16 CGDD vs K_c, **C**. Morales f_c vs K_c, **D**. SER 16 f_c vs K_c. The curves were fitted from growth periods V1 to R9 (Data from: Ramirez, 2007). *(These data's were obtained under the project sponsored by NOAA-CREST (NA17AE1625), NASA-EPSCoR (NCC5-595), USDA-TSTAR-100,USDA Hatch Project H-402, and University of Puerto Rico Experiment Station)*

where θ_{FC} is the water content at field capacity ($m^3.m^{-3}$), θ_{WP} is the water content at wilting point ($m^3.m^{-3}$), and Z_t is the rooting depth (m)

$$RAW = pTAW \tag{20}$$

Allen et al. (1998) give values to different crops (FAO, 56.p163). Roygard et al. (2002) and Zhang et al. (2004), reported that K_s is a logarithmic function of soil water availability (Aw), and can be estimated as follow.

$$K_s = \ln(Aw+1)/\ln(101) \tag{21}$$

and Aw is calculated according to the equation

$$Aw = 100\left(\frac{\theta_a - \theta_{wp}}{\theta_{FC} - \theta_{wp}}\right) \tag{22}$$

where θ_a is average soil water content in the layers of the root zone depth.
An example of the relationships between K_s and available soil water changes, estimated as a root zone depletion, is presented by Ramirez (2007). The root zone depletion (D_r), can be calculated using the water balance equation:

$$D_{r,i} = D_{r,i-1} - (P - RO)_i - I_i + ET_{c,i} + DP_i \tag{23}$$

where $D_{r,i}$ is the root zone depletion at the end of day i; $D_{r,i-1}$ is water content in the root zone at the end of the previous day, i-1; $(P\text{-}RO)_i$ is the difference between precipitation and surface runoff on day i; I_i is the irrigation depth on day i; $ET_{c,i}$ is the crop ET on day i and DP_i is the water loss from the root zone by deep percolation on day i; all the units are in mm. The root zone depletion associated with a K_s = 1.0 (i.e., no water stress), was up to 10 mm for a root depth between 0 to 20 cm, and up to 15 mm for a root depth of 0 to 40 cm in common beans. Fifty percent of the transpiration reduction was reached for D_r = 22 mm and 25 mm for the common bean genotype Morales and genotype SER 16, respectively. Transpiration ceased completely (K_s = 0) when Dr = 37 mm and 46 mm, respectively, for Morales and SER 16 (Ramírez, 2007).

3. Lysimeters as a direct water vapour flux measurement

The word '*lysimeter*' is derived from the Greek root 'lysis,' which means dissolution or movement, and 'metron,' which means to measure (Howell, 2005). Lysimeters are tanks filled with soil in which crops are grown under natural conditions to measure the amount of water lost by evaporation and transpiration (Allen et al., 1990). A lysimeter is the method of determining ET directly. The lysimeter are tanks buried in the ground to measure the percolation of water through the soil. Lysimeter are the most dependable means of directly measuring the ET rate, but their installation must meet four requirements for the data to be representative of field conditions (Chang, 1968).
Requirement 1. The lysimeter itself should be fairly large and deep to reduce the boundary effect and to avoid restricting root development. For short crops, the lysimeter should be at least one cubic meter in volume. For tall crops, the size of the lysimeter should be much larger.
Requirement 2. The physical conditions within the lysimeter must be comparable to those outside. The soil should not be loosened to such a degree that the root ramification and water movement within the lysimeter are greatly facilitated. If the lysimeter is unclosed on the bottom, precaution must be taken to avoid the persistence of a water table and presence of an abnormal thermal regime. To ensure proper drainage, the bottom of an isolated soil column will often require the artificial application of a moisture suction, equivalent to that present at the same depth in the natural soil (Coleman, 1946)
Requirement 3. The lysimeter will not be representative of the surrounding area if the crop in the lysimeter is either taller, shorter, denser, or thinner, or if the lysimeter is on the periphery of no-cropped area. The effective area of the lysimeter is defined as the ratio of the lysimeter ET per unit area of the surrounding field. The values of this ratio, other than unity, are caused by the in homogeneity of the surface. The maintenance of uniform crop height and density is not an easy task in a tall crop, spaced in rows. If the surface is indeed inhomogeneous, there is no adequate way to estimate the effective area from tank area overlap corrections or plant counts.

Requirement 4. Each lysimeter should have a "guard-ring" area around it maintained under the same crop and moisture conditions in order to minimize the clothesline effect. In arid climates, Thornthwaite in 1954, suggested that a "guard-ring" area of ten acres may or may not be large enough. Where several lysimeters are installed in the same field, the "guard-ring" radius may have to be about ten times the lysimeter separation (Chang, 1968).
Lysimeters surrounded by sidewalks or gravel will not provide reliable data, nor will lysimeters planted to a tall crops if it is surrounded by short grass, or planted to grass and surrounded by a tall crop. Differences in growth and maturity between the lysimeters plants and surrounding plants can result in significant differences in measured ET in and outside the lysimeter (Pruitt & Lourence, 1985; mentioned by Allen et al., 1990). The lysimeters are divided basically in two types: Weighing and Non-weighing, each of which are described below.

3.1 Non-Weighing Lysimeters

Also called *Drainage lysimeters,* they operate on the principle that ET is equal to the amount of rainfall and irrigation water added to the system, minus percolation, runoff and soil moisture changes. Since the percolation is a slow process, the drainage lysimeters is accurate only for a long periods for which the water content at the beginning exactly equals that at end. The length of such a period varies with the rainfall regime, frequency and amount of irrigation water application, depth of the lysimeters, water movement, and the like. Therefore, records of drainage lysimeters should be presented only in terms of a long-period more than one day (Chang, 1968), and they are not useful for estimating hourly ET.
Allen et al. (1990) discusses two types the non-weighing lysimeters: a) *non- weighing constant water-table type*, which provides reliable data in areas where a high water table normally exists and where the water table level is maintained essentially at the same level inside as outside the lysimeters; b) *Non-weighing percolation type*, in which changes in water stored in the soil are determined by sampling or neutron methods or other soil humidity sensors like TDR, and the rainfall and percolation are measured.
General principles of a drainage lysimeter: Provisions are made at the bottom of the lysimeter container to collect and measure volumetrically the deep percolation. Precipitation is measured by rain gauge(s). *Evapotranspiration* is considered as the difference among *water applied, water drainage and soil water change.* (Teare et al., 1973; Xingfa et al., 1999)
When filling-in a lysimeter, the soil dug out from the lysimeter's pit is replaced in the container, special precautions are needed to return the soil to its original status by restoring the correct soil profile and compacting the soil layers to the original density. It is desirable to have a similar soil state inside the lysimeter relative to the outside. However if the roots are well developed and nutrients are available, and as long as the water supply to the roots is unrestricted, dissimilar soil will not give significant variation in water use and yield, provided other conditions are similar.(Xingfa et al. 1999).
Although disturbed soil in filled-in lysimeters does not pose serious problems in ET measurement, the soil can affect plant growth. Breaking up the soil, will change soil structure, aeration, and soil moisture retention characteristics. The lysimeters should provide a normal rooting profile. It should be large enough to render the effect of the rim insignificant. It can give relatively large errors in the ET measurement if the container is small. However, the greater the lysimeters area, the more costly and complicated the installation and operation becomes. (Xingfa et al., 1999).

Installation and walls: The wall can be different materials: reinforced concrete, polyester reinforced with steel, fiberglass or plastic. The installation proceeds in the following steps: Excavation (e.g. 1m*1m*1.2m) in the experimentatal site. Each layer of soil (e.g. 0-30 cm, 30-60cm and 60-100 cm) is separated. Once the excavation it completed, the lysimeter is placed in the excavated hole with 4 wooden boards outside. Before repacking the soil layers, make a V-shaped slope at the bottom and place a 25 mm inside diameter perforated PVC pipe (horizontal). There should be a screen material placed around the perferated pipe to avoid the soil particles from entering the pipe. Connect an access tube (25 mm PVC), approximately 1 m long (vertical). Cover the horizontal pipe with fine gravel approximately 3-5 cm thick. Fill the container with the excavated soil where each layer is repacked inside the lysimeter to match the original vertical soil state. (Xingfa et al. 1999).

3.2 Weighing Lysimeters

A weighing Lysimeter is capable of measuring ET for periods as short as ten minutes. Thus, it can provide more additional information than a drainage lysimeter can. Problems such as diurnal pattern of ET, the phenomenon of midday wilt, the short-term variation of energy partitioning, and the relationship between transpiration and soil moisture tension, can be investigated only by studying the records obtained from a weighing lysimeter. (e.g.; Chang, 1968; Ritchie & Burnett, 1968; Takhar & Rudge, 1970; Parton et al., 1981; Steiner et al., 1991; Allen et al., 1998; Loos et al., 2007; von Unold &Fank, 2008)

Weighing lysimeters make direct measurements of water loss from a growing crop and the soil surface around a crop and thus, provide basic data to validate other water vapor flux prediction methods (e.g.; Dugas et al., 1985; Prueger et al., 1997; López-Urrea et al., 2006; Vaughan et al., 2007). The basic concept of this type of lysimeter is that it measures the difference between two mass values, the mass change is then converted into ET (mm) (Malone et al., 1999; Jhonson et al., 2005).

During periods without rainfall, irrigation and drainage, the ET rate is computed as indicated by Howell (2005), as:

$$ET = \left\lfloor A_l \left[\left(M_i - M_{i-1} \right) / A_l \right] / A_f \right\rfloor / T_i \tag{24}$$

where ET is in units of (mm.h^{-1} or Kg.m^2) for time interval i; M is the lysimeter soil mass, (Kg); A_l is lysimeter inner tank surface area (m^2); A_f is lysimeter foliage area (mid wall-air gap area) (m^2); T is the time period (h). The ratio A_f / A_l is the correction factor for the lysimeter effective area. This correction factor assumes the outside and inside vegetation foliage overlap evenly on all of the sides or edges. If there is no overlap, as occurs in short grass, the A_f / A_l=1.0 (Howell , 2005).

Weighing lysimeters provide the most accurate data for short time periods, and can be determined accurately over periods as short as one hour with a mechanical scale, load cell system, or floating lysimeters (Allen at al., 1990). Some weighing lysimeters use a weighing mechanism consisting of scales operating on a lever and pendulum principle (Harrold & Dreibelbis (1951), mentioned by Malone et al. (1999)). However, some difficulties are very common like: electronic data logger replacement, data logger repair, load cell replacement, multiflexor installation etc. (Malone et al., 1999).

The measurement control in these lysimeters are important because of the following issues: a) re-calibration requirements, b) measurement drift (e.g., slope drift, variance drift), c) instrument problems (e.g., localized non-linearity of load cell, load cell damage, data logger

damage), d) human error (e.g.; incorrectly recording data during calibration) and e) confidence in measurement results (Malone et al. 1999).
A load cell is a transducer that coverts a load acting on it into an analog electrical signal. The electrical signal is proportional to the load and the relationship is determined through calibration, employing linear regressions models (mV/V/mm water), and it is used to determine mass changes of a lysimeter over the period interest (e.g. day, hour, etc.).
The lysimeter characteristics can be different, for example: Malone et al., (1999) built a lysimeter of the following form: 8.1 m^2 in surface area and 2.4 m depth, the lysimeter is constructed without disturbing the soil profile and the underlying fracture bedrock. The soil monolith is supported by a scale frame that includes a 200:1 lever system and a counterweight for the deadweight of the soil monolith. The gap between the soil in the lysimeters and the adjacent soil is between 5.1 cm and 7.0 cm except at the bottom slope where the runoff trough is located, this same author has given instructions for achieving a good calibration for this type of lysimeter.
Tyagi et al., (2000) in wheat and sorghum used two rectangular tanks, an inner and outer tank, constructed from 5-mm welded steel plates. The dimensions of the inner tank were 1.985 x 1.985 x 1.985 m and those of the outer tank were 2.015 x 2.015 x 2.015 m. The lysimeters were situated in the center of a 20-ha field. The size ratio of the outer tank to the inner tank is 1.03, so the error due to wall thickness is minimal. The effective area for crop ET was 4 m^2. The height of the lysimeter rim was maintained near ground level to minimize the boundary layer effect in and around the lysimeter. The lysimeter tank was suspended on the outer tank by four load cells. The load cells were made out of the steel shear beam type with 40,000-kg design load capacity. The total suspended mass of the lysimeter including tank, soil, and water was about 14,000 kg. This provided a safety factor of 2.85. The high safety factor was provided to allow replacement of a load cell without the danger of overloading and also to account for shock loading. A drainage assembly connected with a vertical stand and gravel bedding to facilitate pumping of drainage water was provided. The stand pipe also can be used to raise the water table in the lysimeter.
To calculate the ET using Lysimeter, we need to employ the soil water balance (SWB) equation:

$$ET = R + I - P - Rff +/- \Delta SM \tag{25}$$

Where: R is the rain, I is the irrigation, Rff is the runoff, and +/-ΔSM soil moisture changes, all in mm.
The size of the Lysimeter is an important element to be considered in water vapor flux studies with this method. For example, Dugas & Bland (1989) evaluated small lysimeters (<1.0 m^2) and reported significant differences in the ET estimations, basically associated with the differences in the leaf area index (LAI) inside the lysimeters, which differed among lysimeter, this problem can be addressed using LAI corrections.

3.3 Calibration of the weighing Lysimeter

Seyfried et al.,(2001) made a weighing lysimeter calibration by placing known weights on the lysimeters and then recording the resultant pressure changes. The weights used in that study were as follows: 19.9 kg for supportive blocks placed on the lysimeter, 43.4 kg for the tank which contained the weights, and then twenty-four 22.7 kg sacks of rock added in four-sack increments. The weight of each sack corresponded to about a 13 mm addition of water;

so that weight increments were equivalent to ~52 mm and the total range was~360 mm of water. Measurements were made both as weight was added and removed.
The main arguments against the use of weighing lysimeters for monitoring water balance parameters and measuring solute transport parameters in the soil and unsaturated zone has been the discussion of potential sources of error, such as the well known oasis effect, preferential flow paths at the walls of the lysimeter cylinders due to an insufficient fit of soil monoliths inside the lysimeters, or the influence of the lower boundary conditions on the outflow rates (Fank, 2008).

4. The micrometeorological methods

For many agricultural applications, micrometeorological methods are preferred since they are generally non-intrusive, can be applied on a semi-continuous basis, and provide information about the vertical fluxes that are occurring on scales ranging from tens of meter to several kilometres, depending the roughness of the surface, the height of the instrumentation, and the stability of the atmosphere surface layer. Meyers and Baldocchi (2005) have separated micrometeorological methods into four categories: 1) eddy covariance, 2) flux-gradient, 3) accumulation, and 4) mass balance. Each of these approaches are suitable for applications that depend on the scalar of interest and surfaces type, and instrumentation availability. Some of these methods are described in the following sections of this chapter.

4.1 Humidity and temperature gradient methods

Movement of energy, water and other gases between field surface and atmosphere represent a fundamental process in the soil-plant-atmosphere continuum. The turbulent transport in the surface boundary layer affect the sensible (H) and latent (λE) heat fluxes, which along with the radiation balance, govern the evapotranspiration and canopy temperature (Ham and Heilman, 2003).
Monteith and Unsworth (1990) presented the functional form of the gradient flux equation, and which has been applied by Harmsen et al. (2006), Ramírez et al. (2008) and Harmsen et al. (2009):

$$ET = \left(\frac{\rho_a \cdot c_p}{\gamma \cdot \rho_w}\right) \cdot \frac{\left(\rho_{vL} - \rho_{vH}\right)}{\left(r_a + r_s\right)} \tag{26}$$

where ρ_w is the density of water, ρ_v is the water vapor density of the air, ρ_a is the air density, γ is the psychometric constant, c_p is specific heat of air, r_a and r_s are aerodynamic and bulk surfaces resistances (all these variables are discussed in detail below). L and H are vertical positions above the canopy (L: low and H: High positions), for example in small crops like beans or grass, possible values of L and H could be 0.3 m and 2 m above the ground, respectively.
Harmsen et al. (2006) developed an automated elevator device (ET Station) for moving a temperature and relative humidity sensor (Temp/RH) between the two vertical positions (**Fig.2**). The device consisted of a plastic (PVC) frame with a 12 volt DC motor (1/30 hp) mounted on the base of the frame. One end of a 2-m long chain was attached to a shaft on the motor and the other end to a sprocket at the top of the frame. Waterproof limit switches

were located at the top and bottom of the frame to limit the range of vertical movement. For automating the elevator device, a programmable logic controller (PLC) was used which was composed of "n" inputs and "n" relay outputs. To program the device, a ladder logic was used which is a chronological arrangement of tasks to be accomplished in the automation process. The Temp/RH sensor was connected to the elevator device, which measured relative humidity and temperature in the up position for two minutes then changed to the down position where measurements were taken for two minutes. This process started each day at 0600 hours and ended at 1900. When the elevator moves to the up position it activates the limit switch which sends an input signal to the PLC. That input tells the program to stop and remain in that position for two minutes. At the same time it activates an output which sends a 5 volt signal to the control port C2 in the CR10X data logger in which a small subroutine is executed. This subroutine assigns a "1" in the results matrix which indicates that the temperature and relative humidity corresponding to the up position. At the end of the two minute period, the elevator moves to the down position and repeats the same process, but in this case sending a 5 volts signal to the data logger in the control port C4, which then assigns a "2" in the results matrix. All information was stored in the weather station data-logger CR-10X (Campbell Scientific, Inc) for later downloading to a personal computer.

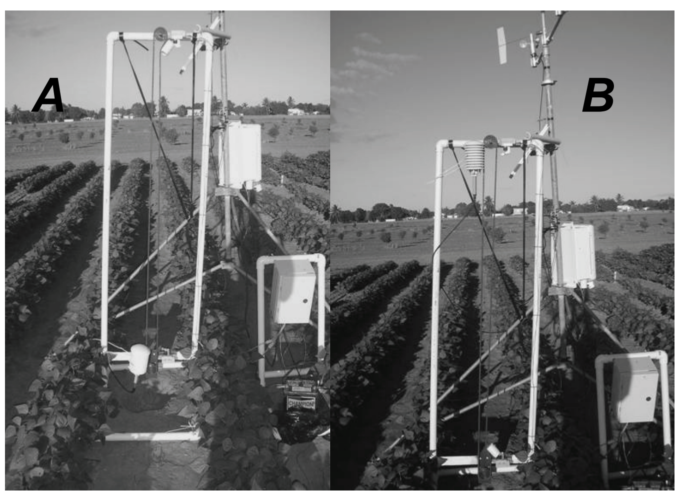

Fig. 2. Automated elevator device developed for moving the Temp/RH sensor between the two vertical positions. **A.** Temp/RH sensor in down position and **B.** Temp/RH sensor in up position. Measuring over common bean (*Phaseolus vulgaris* L.). (*These data's were obtained under the project sponsored by NOAA-CREST (NA17AE1625), NASA-EPSCoR (NCC5-595), USDA-TSTAR-100,USDA Hatch Project H-402, and University of Puerto Rico Experiment Station*)

4.2 The Bowen-ratio energy balance method

The basis for this method is that the local energy balance is closed in such a way that the available net radiative flux (Rn) is strictly composed of the sensible (H), latent (λE), and ground heat (G) fluxes, other stored terms such as those related to canopy heat storage and photosynthesis are negligible (Meyers and Baldocchi, 2005).

This method combines measurements of certain atmospheric variables (temperature and vapour concentration gradients) and available energy (net radiation and changes in stored thermal energy) to determine estimates of evapotranspiration (ET) (Lloyd, 1992). The method incorporates energy-budget principles and turbulent-transfer theory. Bowen showed that the ratio of the sensible- to latent-heat flux (β) could be calculated from the ratio of the vertical gradients of temperature and vapour concentration over a surface under certain conditions.

Often the gradients are approximated from air-temperature and vapour-pressure measurements taken at two heights above the canopy. The Bowen-ratio method assumes that there is no net horizontal advection of energy. With this assumption, the coefficients (eddy diffusivities) for heat and water vapour transport, kh and kw, respectively, are assumed to be equal. Under advective conditions, kh and kw are not equal (Verma et al., 1978; Lang et. al., 1983 cited bay Tomilson, 1997) and the Bowen-ratio method fails to accurately estimate ET.

Based on the assumption that Kh and Kw are equal, and by combining several terms to form the psychometric constants, the Bowen-ratio take the form to the equation 28. Although the theory for this method was develop in the 1920s by Bowen (Bowen, 1926), its practical applications has only been possible in recent decades, due to the availability of accurate instrumentation (Payero et al., 2003). The Bowen ratio initial concept is shown below:

$$\beta = \frac{PC_p Kh \frac{dT}{dz}}{\lambda \varepsilon Kw \frac{de}{dz}} \tag{27}$$

If it is assumed that there is no net horizontal advection of energy, equ. 27 can be simplified as shown below:

$$\beta = \frac{PC_p \frac{dT}{dz}}{\lambda \varepsilon \frac{de}{dz}} \tag{28}$$

where P is the atmospheric pressure (kPa), C_p is the specific heat of air (1.005 J/g°C), ε is the ratio molecular weight of water to air = 0.622 and λ is the latent heat of evaporation (Jg^{-1}). Once the Bowen ratio is determined, the energy balance (equ. 29) can be solved for the sensible-heat flux (H) and latent-heat flux (λE).

$$Rn = \lambda E + H + G \tag{29}$$

where R_n Rn is the net radiation, λE is the latent-heat flux, H is the sensible-heat flux and G is the soil-heat flux ($W.m^{-2}$).

$$H = \beta \lambda E \tag{30}$$

$$\lambda E = \frac{(R_n - G)}{(1+\beta)} \tag{31}$$

The latent heat flux can be separated into two parts: the evaporative flux E (g m^{-1} day-1) and the latent heat of vaporization λ (Jg^{-1}), which can be expressed as a function of air temperature (T-°C) (λ = 2,502.3-2.308 T). The latent-heat of vaporation (λ) is defined as the amount of energy required to convert 1 gram of liquid water to vapour at constant temperature T. Sensible-heat flux (H) is a turbulent, temperature-gradient driven heat flux resulting from differences in temperature between the soil and vegetative surface and the atmosphere.

The soil-heat flux (G) is defined as the amount of energy moving downward through the soil from the land surface, caused by temperature gradient. This flux is considered positive when moving down through the soil from the land surface and negative when moving upward through the soil toward the surface (Tomilson, 1994). The soil heat flux is obtained by measuring two soil heat flux plates below the soil surface at 2 and 8 cm, soil moisture at 8 cm, and soil temperature at 6 cm between the two soil heat flux plates (Campbell Scientific, Inc. 1998).

Because the soil-heat flux is measured below the soil surface, some of the energy crossing the soil surface could be stored in, or come from, the layer of soil between the surface and flux plate located closest to the surface, for this reason a change in storage term, S is added to the measured heat flux (equ. 33). (Cambell Scientific, Inc. 1991):

$$S = \left[\frac{\Delta Ts}{\Delta t}\right] d\rho_b \left(C_s + \left(WC_w\right)\right) \tag{32}$$

where S is the heat flux going into storage (Wm^{-2}), Δt is the time interval between measurement (sec), ΔTs is the soil temperature interval between measurement, d is the depth to the soil-heat-flux plates (0.08m), ρ_b is the bulk density of dry soil, C_s is the specific heat of dry soil (840 J/Kg°C), W is the water content of soil (kg the water/ Kg the soil) and C_w is the specific heat of water (4,190 J/Kg°C). The soil heat flux (G) at the surface is obtained by including the effect of storage between the surface and depth, *d*, using equation 11.

$$G = \left(\frac{FX1 + FX2}{2}\right) + S \tag{33}$$

where FX_1 is the soil-heat flux measured 1 (Wm^{-2}), FX_2 is the soil-heat flux measured 2 (Wm^{-2}).

One of the requirements for using the Bowen-ratio method is that the wind must pass over a sufficient distance of similar vegetation and terrain before it reaches the sensors. This distance is referred to as the fetch, and the fetch requirement is generally considered to be 100 times the height of the sensors above the surface (Campbell, 1977). More detail about determination of the minimum fetch requirement is presented later in this chapter.

Hanks et al. (1968), described by Frank (2003), reported λE/Rn of 0.16 for dry soil conditions and 0.97 for wet soil conditions; On the other hand he found λE/Rn to be lowest in grazed prairie, suggesting that defoliation changes the canopy structure and energy budget components, which may have contributed to increase water loss through evaporation compared with the non-grazed prairie treatment. Hanson and May (2004), using the Bowen Ratio Energy Balance Method to measure ET in tomatoes, found that ET rates decreased substantially in response to drying of the soil surface.

Perez et al. (2008) proposed a simple model for estimating the Bowen ratio (β) based on the climatic resistance factors:

$$\beta = \frac{\Delta + \gamma}{\Delta} . \frac{1 + S}{1 + C} - 1 \tag{34}$$

$$C = \frac{\gamma . r_i}{\Delta . r_a} \tag{35}$$

$$S = \frac{\gamma}{\Delta + \gamma} . \frac{r_c}{r_a} \tag{36}$$

where r_c is the canopy resistance (s m^{-1}) based on the "big leaf" concept, and r_a is the aerodynamic resistance (s m^{-1}). These resistance factors are described in detail in the next section. The factor r_i is the climatological resistance as reported by Montheith (1965):

$$r_i = \frac{\rho_a C_p VPD}{\gamma (R_n - G)} \tag{37}$$

where ρ_a is the air density at constant pressure (Kg.m^{-3}), C_p is the specific heat of moist air at constant pressure (1004 J.Kg^{-1} $^0C^{-1}$), VPD is the vapour pressure deficit of the air (Pa), γ is the pychrometric constant (Pa. $^0C^{-1}$) and R_n and G are in W.m^{-2}. For homogeneous canopies, the effective crop surface and source of water vapour and heat is located at height d + z_{oh}, where d is the zero plane displacement height and z_{oh} is the roughness length governing the transfer of heat and vapour (Allen *et al.* 1998).

4.3 The Penman-Monteith Method

The important contribution of Monteith and Penman`s original equation was the use of resistances factors, which was based on an electrical analogy for the potential difference needed to drive unit flux systems that involve the transport of momentum, heat, and water vapour (Monteth and Unsworth, 1990; Monteith, 1997). The resistances have dimensions of time per unit length, as will be described later. This methodology calculates the latent heat flux using the vapour pressure deficit, the slope of the saturated vapour-pressure curve and aerodynamic resistance to heat, and canopy resistance in addition to the energy-budget components of the net radiation, soil heat flux, and sensible heat flux. Field measurements of air temperature, relative humidity, and wind speed are needed to determine these variables (Tomilson, 1997).

Equation 38 describes the Penman-Monteith (P-M) method to estimate the λE (Allen at al., 1998, Kjelgaard and Stockle, 2001)

$$\lambda E = \frac{\Delta s (R_n - G) + \rho_a C_p \dfrac{VPD}{r_a}}{\Delta s + \gamma \left(1 + \dfrac{r_s}{r_a} \right)} \tag{38}$$

where λE, R_n, and G in Wm^{-2}, VPD is vapour pressure deficit (kPa), $\Delta S$is slope of saturation vapor pressure curve (kPa $\circ C^{-1}$) at air temperature, ρ is density of air (Kgm^{-3}), C_p in J $Kg^{-1} \circ C^{-1}$, γ in kPa $\circ C^{-1}$, r_a is aerodynamic resistant (s m^{-1}) r_s surface resistance to vapour transport (s m^{-1}).

According to Monteith (1997), the appearance of a wind-dependent function in the denominator as well as in the numerator implies that the rate of evaporation calculated from the P-M model is always less dependent on wind speed than the rate from the corresponding Penman equation when other elements of climate are unchanged. In general, estimated rates are usually insensitive to the magnitude of r_a and the error generated by neglecting the influence of the buoyancy correction is often small. In contrast, the evapotranspiration rate is usually a strong function of the surfaces resistance (r_s).

Kjelgaard and Stockle (2001) say the surface resistance (r_s) parameter in the P-M model is particularly difficult to estimate due to the combined influence of plant, soil and climatic factors that affect its value. The magnitude of the stomatal resistance can be estimated in principle from the number of stomata per unit leaf area and from the diameter and length of pores, which is difficult and therefore rarely measured; therefore, the stomatal resistance is usually calculated from transpiration rates or estimated gradients of vapour concentration (Monteith, 1997).

Knowing the value of the a***erodynamic resistance (r_a)*** permits estimation of the transfer of heat and water vapour from the evaporating surface into the air above the canopy. The aerodynamic resistance for a single leaf to diffusion through the boundary layer surrounding the leaf, within which the transfer of heat, water vapour, etc., occurs, proceeds at a rate governed by molecular diffusion. Provided the wind speed is great enough and the temperature difference between the leaf and air is small enough to ensure that transfer processes are not affecting by gradients of air density, the boundary layer resistance depends on air velocity and on the size, shape, and altitude of the leaf with respect to the air stream. In very light wind, the rates of transfer are determined mainly by gradients of temperature and therefore by density, so that the r_a depends more on the mean leaf-air temperature difference than on wind speed. According to Thom (1975), the r_a for heat transfer can be determined by:

$$r_{ah} = \frac{\rho C\rho (T_s - T_a)}{H} \tag{39}$$

At the field level, r_a for homogeneous surfaces, such as bare soil or crop canopies, there is a large-scale analogous boundary layer resistance, which can be estimated or derived from measurements of wind speed and from a knowledge of the aerodynamic properties of the surface as is described later (Monteith, 1997). The r_a can be determined given values of roughness length (Z_o) and zero plane displacement height (d), that depend mainly on crop height, soil cover, leaf area and structure of the canopy (Massman, 1987; Perrier, 1982; Shaw and Pereira, 1982 cited by Alves et al. 1998),

$$r_a = \frac{Ln\left[\frac{(Z_m - d)}{Z_{om}}\right] Ln\left[\frac{(Z_h - d)}{Z_{oh}}\right]}{K^2 u_z} \tag{40}$$

where Z_m is height of wind measurements (m), Z_h is height of humidity measurements (m), d is zero displacement height (m), Z_{om} is roughness length governing momentum transfer of heat and vapour (m) is 0.123h, Z_{oh} is roughness length governing transfer of heat and vapour (m) is $0.1Z_{om}$, K is the von Karman`s constant (0.41), u_z is win speed at height z.

This equation is restricted for neutral stability conditions, i.e., where temperature, atmospheric pressure, and wind speed velocity distribution follow nearly adiabatic conditions (no heat exchange). The application of the equation for short time periods (hourly or less) may require the inclusion of corrections for stability. However, when predicting ET_o

in the well watered reference surface, heat exchange is small, and therefore the stability correction is normally not required (Allen et al., 1998).
Alves et al. (1998) state that though this is the most used expression for r_a, in fact it is not entirely correct, since it assume a logarithmic profile from the source height (d + Z_{oh}) with increasing z in the atmosphere, using the concept to the "big leaf", equ. 40 can be modified as follows:

$$r_a = \frac{Ln\left(\frac{z-d}{h_c-d}\right)Ln\left(\frac{z-d}{Z_{om}}\right)}{K^2 u_z} \quad (41)$$

where h_c is the height of the crop canopy.
According to Tollk et al. (1995), the r_a to momentum transport in the absence of buoyancy effects (neutral stability) follows the equation:

$$r_{a\,m} = \ln[(Zi-d)/Z_{om}]^2 / k^2 u_z \quad (42)$$

Under adiabatic conditions, the equations must be corrected using the Richardson number for stability correction, assuming similarity in transport of heat and momentum, yielding:

$$r_{a\,h} = r_{am}(1+5R_i) \quad (43)$$

The Ri for stability conditions is considered when ($-0.008 \le R_i \le 0.008$) and is calculated by

$$R_i = [g(T_a - T_s)(Z-d)] / T_{av}.u^2{}_z \quad (44)$$

where g is the acceleration of the gravity (9.8 m.s^{-2}), T_a is the air temperature (K), T_c is the plant canopy temperature (K), T_{av} is the average temperature taken as $((T_a+T_c)/2)$. The advantage of the R_i over other stability corrections is that it contains only experimentally determined gradients of temperature and wind speed and does not depend directly on sensible heat flux (Tolk et al., 1995).
The ***bulk surface resistance (rs)*** describes the resistance of vapour flow through transpiring crop leaves and evaporation from the soil surface. Where the vegetation does not completely cover the soil, the resistance factor should indeed include the effects of the evaporation from the soil surface. If the crop is not transpiring at a potential rate, the resistance depends also on the water status of the vegetation (Van Bavel, 1967; Allen et al., 1998), and for this case they proposed the use of the following approximate:

$$r_s = \frac{r_L}{LAI_{active}} \quad (45)$$

where LAI_{ctive} is 0.5 times the leaf area index (m^2 of leaf per m^2 of soil), and r_L is bulk stomatal resistance, which is the average resistance of an individual leaf, and can be measured using an instrument called a porometry, the first stomatal readings were developed by Francis Darwin who develop his horn hygrometer (Turner, 1991).
The r_L readings are highly variable and depend on several factors, such as: crop type and development stage, the weather and soil moisture variability, the atmospheric pollutants and the plant phytohormone balance (Turner, 1991). Typically to determine minimum r_L

using a porometer, fully expanded, sunlit leaves near to the top of the canopy are surveyed during maximum solar irradiance (approximately solar noon under cloudless conditions) and low VPD periods (Kjelgaard and Stockle, 2001). This "standard" value from literature or porometer measurements are hereafter identified as r_{Lmin}. In addition, r_L has been shown to increase with increasing VPD and/or reduced solar irradiance (R_s). Adjustment factors for VPD (f_{VPD}) and R_s (f_{Rs}) were empirically derived and used as multipliers of r_{Lmin}.
The dependence of r_L on VPD can be represented by a linear function (Jarvis, 1976) as

$$f_{VPD} = a + bVPD \tag{46}$$

where a and b are linear regression coefficients, and f_{VPD} is equal to 1 (no adjustment) for VPD ≤ a threshold value, which can be taken as 1.5 kPa. The same authors presented a calibrated form of equation 46 for corn as, f_{VPD} = 0.45+0.39(VPD).
Kjelgaard and Stockle (2001) presented a modified form of the adjustment factor:

$$f_{Rs} = \frac{R_{s\max}}{C_2 + R_s} \tag{47}$$

where R_s and $R_{s\,max}$ are the actual and maximum daily solar irradiance (MJ m^{-2} day^{-1}) and C_2 is a fitted constant.
Taking the maximum of the adjustment factors for VPD and Rs, r_{Lmin} is modified to give the r_L (Kjelgaard and Stockle, 2001):

$$r_L = r_{L\min} \max\{f_{VPD}, f_{Rs}\} \tag{48}$$

where f_{VPD} and f_{Rs}, are equal to or larger than 1.
Alves et al. (1998) indicate that the surface resistance term (r_s) has been the most discussed in the literature. Several components to be considered here include: a) The resistances to water vapour at the evaporating surfaces: plants and their stomates (r_s^c) and soil (r_s^s); b) the resistance to vapour transfer inside the canopy from these evaporating surfaces up to the "big leaf" (r_s^a). The resistance r_s^c, can be approximated using equ. 50.

$$r_s^{\,c} = \frac{\left(\sum_{i=1}^{n} \frac{1}{r_{stj}}\right)^{-1}}{LAI} \tag{49}$$

Where r_{st} is the single leaf stomatal resistance (sm^{-1}), n is a leaf number.
The bulk surface resistance can also be calculated using the inversion of the Penman-Monteith equation with incorporation of the Bowen ratio as follow (Alves et al. (1998) and Alves and Pereira, 2000):

$$r_s = r_a\left(\frac{\Delta s}{\gamma}\beta - 1\right) + \frac{\rho_a C_p VPD}{\gamma \lambda E} \tag{50}$$

Accurate prediction of r_s requires a good estimate of the Bowen ratio (β). Ramirez (2007) has used the following inversion form of the Penman-Monteith equation to obtain estimates of r_s:

$$r_s = r_a.\frac{\left[\frac{\Delta(R_n - G) + \rho_a C_p\left(\frac{VPD}{r_a}\right)}{\lambda E} - \Delta - \gamma\right]}{\gamma} \tag{51}$$

Similarly these authors, analysing the resistance concepts, concluded that the r_s of dense crops cannot be obtained by simply averaging stomatal resistance because the driving force (vapour pressure deficit) is not constant within the canopy.
Saugier (1977) addressed canopy resistance (r_c), stating that it is normally a mixture of soil and plant resistances to evaporation. If the top the soil is very dry, direct soil evaporation may be neglected and r_c is approximately equal to the leaf resistance (r_L) divided by the LAI. Baldocchi *et al. (*1991), indicated that the inverse of the 'big-leaf' model (eg., inverse of the P-M model) will be a good estimate of canopy resistance or surface resistances if certain conditions are met. These conditions include: i) a steady-state environment; ii) a dry, fully developed, horizontally homogeneous canopy situated on level terrain; iii) identical source-sink levels for water vapour, sensible heat and momentum transfer, and negligible cuticular transpiration and soil evaporation.
Szeicz and Long (1969) describe a profile method to estimated r_s as,

$$r_s = \frac{\rho_a.C_p,VPD}{\gamma.\lambda E} \tag{52}$$

These methods can be used in the field when the rate of evapotranspiration is measured by lysimeters or calculated from the Bowen ratio energy balance method, and the temperature, humidity and wind profiles are measured within the boundary layer simultaneously.
Ortega-Farias et al. (2004), evaluated a methodology for calculating the canopy surface resistance ($r_{cv} \approx r_s$) in soybean and tomatoes, using only meteorological variables and soil moisture readings. The advantage of this method is that it can be used to estimate λE by the general Penman-Monteith model with meteorological reading at one level, and without r_L and LAI measurements.

$$r_s = \frac{\rho_a.c_p.VPD}{\Delta.(R_n - G))}.\frac{\theta_{FC} - \theta_{WP}}{\theta_i - \theta_{WP}} \tag{53}$$

Where θ_{FC} and θ_{WP} are the volumetric moisture content at field capacity (fraction) and wilting point (fraction), respectively, and θ_i is a volumetric soil content in the root zone (fraction) measured each day.
Kamal and Hatfield (2004), used the equation 50 to determine the surface resistance in Potato: and stated that the canopy resistance (r_c in s.m^{-1}; "mean stomatal resistances of crops"), can be determined by dividing the r_s by the effective LAI as defined by other authors such as Hatfield and Allen (1996) and for well watered crops, r_c can be can be estimated using equation 54.

$$r_c = \frac{0.3LAI + 1.2}{LAI} r_s \tag{54}$$

Kjelgaard and Stockle (2001) discussed the estimation of canopy resistance (r_c) from single-leaf resistance (r_L) (equation 55), as originally proposed by Szeicz and Long (1969):

$$r_c = \frac{r_L}{LAI_{active}} \tag{55}$$

Kamal and Hatfield (2004) divided the surface resistance (r_s) used in the P-M model into two components, and conceptualized an excess resistance (r_o) in series with the canopy stomatal resistance. This excess resistance was linked to the structure of the crop, particularly crop height.

$$r_s = r_c + r_o \tag{56}$$

Pereira et al. (1999) stated that the surface resistance (r_s) is the sum of two components: one corresponding mainly to the stomatal resistance (r_{st}), the other to the leaf boundary layer and turbulent transfer inside the canopy (r_{ai}) (equation 57), thus, surface resistance is not a purely physiological parameter:

$$r_s = r_{st} + r_{ai} \tag{57}$$

Stomatal resistance can take values from 80 $s.m^{-1}$ to 90 $s.m^{-1}$ as a common range for agricultural crops suggested a value of 100 $s.m^{-1}$ for most arable crops (Monteith, 1981). The **table 1** lists mean average values for various crops under well water conditions.
The r_L is strongly dependent on the time of day (basically due to the temporal nature of climatic conditions), for the soil moisture content and by the genotype. **Fig. 3A**, shows how larger differences in r_L occur, with and without drought stress, after 9:00 am until late in the afternoon, and the most critical point is at 13:00 hours when the highest VPD occurred. For this reason, when this variable (r_L) is not measured, appropriate parameterisation is required for good water flux or ET estimation, especially under drought stress conditions. In **Fig. 3C,** it is possible to see in a common bean genotype under drought stress conditions, lower r_L as compared with less drought resistance during several days with drought stress.
Perrier (1975), as reported in Kjelgaard and Stockle (2001), conceptualized the excess resistance (r_o) as a linear function of crop height and LAI:

$$r_o = ah_c + bLAI \tag{58}$$

where a and b are constants. For corn, Kjelgaard and Stockle (2001) parameterized equ. 58 as follows: r_o= 16.64h_c+0.92LAI.
Canopy resistance can also be determined from leaf or canopy temperature since it is affected by plant characteristics, eg. Leaf area index (LAI), height, and maturity. Soil factors (Available soil water-ASW, and soil solution salinity) and weather factors (R_n and wind speed) also affect the canopy resistance.
Montheith (1965) showed that transpiration rate physically depends on relative changes of surface temperature and r_a, and concluded that r_a depends on the Reynolds number of the air and can be determined from wind speed, the characteristic length of the plant surface, and the kinematic viscosity of the air. An increase in r_c for Wheat was caused by a decrease in total leaf area, by an increase in the resistance of individual leaves due to senescence, or by a combination of both effects; in Sudan grass, r_c increased with plant age and a decrease in soil moisture. Van Bavel (1967) studied Alfalfa throughout an irrigation cycle and found

that canopy resistance increased linearly with decreasing soil water potential. Kamal and Hatfield (2004) found an exponentially inverse relationship between canopy resistance and net radiation, and a linear inverse relationship between canopy resistance and available soil water.

Cover crops	**r_L**	**Source**	**Cover crops**	**r_L**	**Source**
	s/m			s/m	
Corn	200	Kirkham et al. (1985)	Cassava	714 Between 476 to 1428	Oguntunde (2005). This data under limited soil water conditions.
Sunflower	400	Kirkham et al. (1985)	Eucalyptus	200-400	Pereira and Alves (2005)
Soybean and potato	350	Kirkham et al. (1985)	Maple	400-700	Pereira and Alves (2005)
Sorgum	300	Kirkham et al. (1985)	Crops-General	50-320	Pereira and Alves (2005)
Millet	300	Kirkham et al. (1985)	Graim sorghum	200	Pereira and Alves (2005)
Aspen	400	Pereira and Alves (2005)	Soybean	120	Pereira and Alves (2005)
Maize	160	Pereira and Alves (2005)	Barley	150-250	Pereira and Alves (2005)
Alfalfa	80	Pereira and Alves (2005)	Sugar beet	100	Pereira and Alves (2005)
Clipped grass (0.15 m)	100-150	Pereira and Alves (2005)	Clipped and Irrigated grass (0.10-10.12m)	75	Pereira and Alves (2005)
Common beans	170-270	Ramirez et al. (2007)	Sorghum	192	Stainer et al. (1991)
Corn	264	Ramirez and Harmsen (2007). Unpublished data.	Andes Tropical Forestry	132	Ramirez and Jaramillo (2008). (Calculated)
Coffee	149	Ramirez and Jaramillo (2008). (Calculated)	Coffee	150	Angelocci et al. (1983)
Wheat	134	Howell et al. (1994)	Corn	252	Howell et al. (1994)
			Sorghum	280	Howell et al. (1994)

Table 1. Average values of the stomatal resistance (r_L) for several crops.

The Drainage and Irrigation Paper-FAO 56 (Allen et al., 1998) recommends the Szeicz and Long (1969) method for calculating r_s (equation 55), in which an average of r_L for different positions within the crop canopy, weighted by LAI or $LAI_{effective}$ is used. This method seems to give good results only in very rough surfaces, like forest and partial cover crops with a dry soil (Monteith, 1981). Alves et al. (1998) concluded that r_s of dense crops cannot be obtained by simply averaging stomatal resistance (r_L) because VPD, which is the "driving force", is not constant within the canopy. Alves and Pereira (2000) have stated "*The PM model can be used to predict ET if accurate methodologies are available for determining the r_s that take into account the energy partitioning*".

In addition to the lack of r_s values for crops, questions have been raised relative to the appropriateness of using the PM model for partial or sparse canopies because the source/sink fluxes may be distributed in a non-uniform manner throughout the field (Ham and Heilman, 1991; Kjelgaard et al., 1994; Farahami and Bausch, 1995; Ortega-Farias et al.; 2006). Adequate parameterization of the surface resistance makes the P-M model a good estimator of ET (i.e., Saugier and Katerji, 1991; Rana et al., 1997a; Alves and Pereira, 2000; Ortega-Farias et al., 2004).

Ramirez (2007), reported that the daily ET estimation with the P-M model with r_s based on r_L and $LAI_{effective}$ gave a good estimation in two common bean genotypes with variable LAI, without and with moderate drought stress for both years (2006 and 2007).

Ramirez et al. (2008) an reported inverse relation between r_a and r_s and r_L in beans (*Phaseolus vulgaris* L), as well as those reported by Alves and Pereira (2000) (**Fig. 4**), which implies that with low r_a (windy conditions), the r_L (and therefore r_s) increases. The Alves and Pereira (2000) study did not measure the r_L, rather the r_s was estimated based on micrometeorological parameters.

Disparities in the measured r_s using the P-M inverse model arise from: a) imperfect sampling of leaves and the arbitrary method of averaging leaf resistance over the whole canopy, b) from the dependence of r_s on non-stomatal factors such as evaporation from wet soil or stems, or others and c) the complex aerodynamic behaviour of canopies (Monteith, 1995).

Lower LAI index (LAI <1.0) and drought stress also affect the precision in the r_s estimation (eg., Ramirez, 2007). Use of the $LAI_{effective}$ when LAI < 1.0 is not necessary and tends to overestimate the r_s and under-estimate the ET. Katerji and Perrier (1985) found for LAI >1.0 a good agreement between measurement values of evapotranspiration over alfalfa crops using the energy balance method, and values calculated with P-M equation using variable r_s.

Katerji and Perrier (1983) proposed to simulate r_s using the following relation:

$$\frac{r_s}{r_a} = a\frac{r^*}{r_a} + b \tag{59}$$

where **a** and **b** are linear coefficients that need empirical determination, r^*(s.m^{-1}) is a climatic resistance (Katerji and Rana, 2006) giving by:

$$r^* = \frac{\Delta + \gamma}{\Delta\lambda} \cdot \frac{\rho C_p VPD}{(R_n - G)} \tag{60}$$

Table 2 presents values of a and b for several crops.

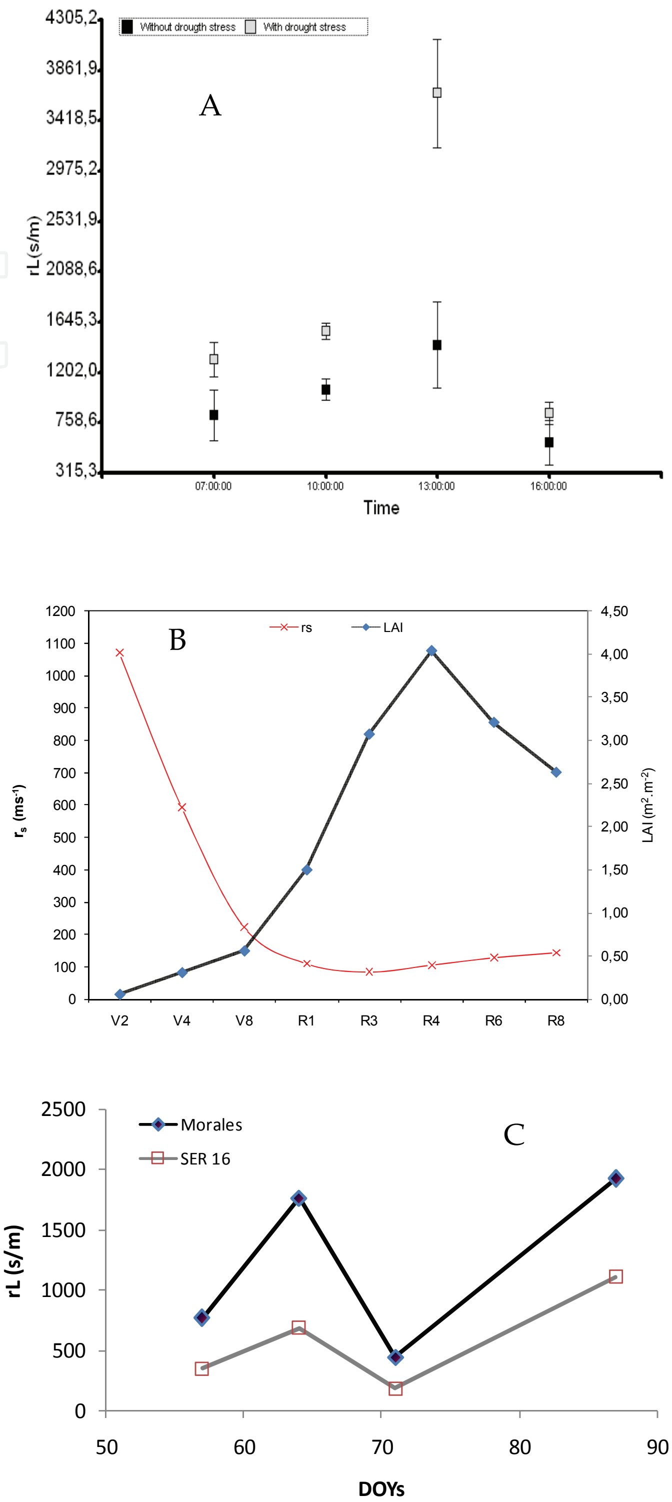

Fig. 3. Relationship between **A.** Changes in the stomatal resistance during the day with and without drought stress in *Phaseolus vulgaris* L. genotype 'Morales'. B. Surfaces resistance and Leaf area index, and **C.** Stomatal behaviour represented in stomatal resistance (r_L) under drought stress conditions for two common bean genotypes -'Morales' lest drought tolerant and 'SER 16' drought stress tolerant.

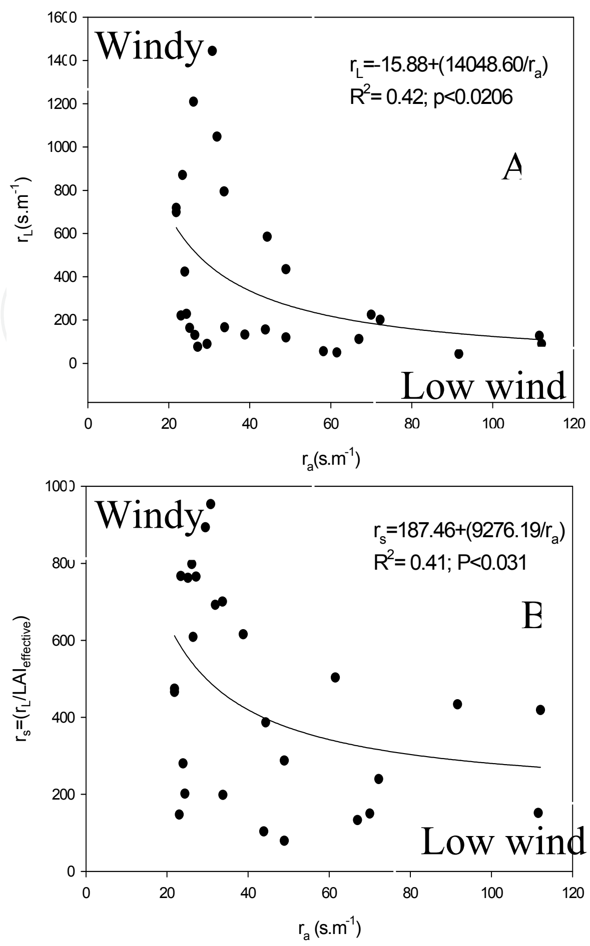

Fig. 4. Aerodynamic resistance (r_a) as a function of: **A**. Stomatal resistance (r_L) and **B.** Measured surface resistance: $r_s = r_L/LAI_{effective}$, (**Datas from: Ramirez, 2007**).

Crop	a	b	Source
Grass	0.16	0.0	Katerji and Rana (2006)
Tomato	0.54	2.4	Katerji and Rana (2006)
Grain sorghum	0.54	0.61	Katerji and Rana (2006)
Soybean	0.95	1.55	Katerji and Rana (2006)
Sunflower	0.45	0.2	Katerji and Rana (2006)
Sweet sorghum	0.845	1.0	Katerji and Rana (2006)
Grass (Tropical climate)	0.18	0.0	Gosse (1976) in Rana et al. (1997)a
Grass (Mediterranean climate)	0.16	0.0	Rana et al. (1994)
Alfalfa	0.24	0.43	Katerji and Perrier (1983) in Rana et al (1997)a
Sorghum	0.94	1.1	Rana et al. (1997)b
Sunflower	0.53	1.2	Rana et a.l (1997)b

Table 2. Coefficients a and b for several crops

The Penman-Monteith model is considered as a 'single-layer' model, Shuttlewoth and Wallace (1985) developed a 'double-layer' model, relying on the Penman-Monteith model concept to describe the latent heat flux from the canopy (λT) and from the soil (λE) as follows:

$$\lambda T = \frac{\Delta(R_n - R_{ns}) + \rho Cp \dfrac{VPDo}{r^c{}_a}}{\Delta + \gamma\left(1 + \dfrac{r^c{}_s}{r^c{}_a}\right)} \tag{61}$$

$$\lambda E = \frac{\Delta(R_{ns} - G) + \rho Cp \dfrac{VPDo}{r^s{}_a}}{\Delta + \gamma\left(1 + \dfrac{r^s{}_s}{r^s{}_a}\right)} \tag{62}$$

where R_{ns} is the absorbed net radiation at the soil surface, $r^c{}_a$ is the bulk boundary layer resistance of the canopy elements within the canopy, $r^c{}_s$ is the bulk stomatal resistance of the canopy, $r^s{}_a$ is the aerodynamic resistance between the soil and the mean canopy height, $r^s{}_s$ is the surfaces resistance of the soil and VPD_o is the vapour pressure deficit at the height of the canopy air stream.

4.4 The double-layer Shulttleworth-Wallace model

The Shulttleworth-Wallace Model (S-W) assumes that there is blending of heat fluxes from the leaves and the soil in the mean canopy airflow at the height of the effective canopy source (Shulttleworth and Wallace,1985). The full expression of the Shulttleworth-Wallace Model (S-W) model is presented by Zhang *et al*. (2008) as follow:

$$\lambda ET = \lambda E + \lambda T = C^S_{SW} PM^S_{SW} + C^P_{SW} PM^P_{SW} \tag{63}$$

$$PM^S_{SW} = \frac{\Delta A_{SW} + \left[\left(\rho CpD - \Delta r^s_a\right)\left(A_{Sw} - A^s_{SW}\right)/\left(r^a_a + r^s_a\right)\right]}{\Delta + \gamma\left[1 + r^s_s/\left(r^a_a + r^s_a\right)\right]} \tag{64}$$

$$PM^p_{SW} = \frac{\Delta A_{SW} + \left[\left(\rho CpD - \Delta r^p_a A^s_{SW}\right)/\left(r^a_a + r^p_a\right)\right]}{\Delta + \gamma\left[1 + r^p_s/\left(r^a_a + r^p_a\right)\right]} \tag{65}$$

$$C^s_{SW} = \frac{1}{1 + \left[R^S_{SW} R^a_{SW} / R^P_{SW}\left(R^S_{SW} + R^a_{SW}\right)\right]} \tag{66}$$

$$C^P_{SW} = \frac{1}{1 + \left[R^P_{SW} R^a_{SW} / R^S_{SW}\left(R^P_{SW} + R^a_{SW}\right)\right]} \tag{67}$$

$$R^S_{SW} = (\Delta + \gamma) r^s_a + \gamma r^s_s \tag{68}$$

$$R^P_{SW} = (\Delta + \gamma) r^p_a + \gamma r^p_s \tag{69}$$

$$Ra_{SW}^{P} = (\Delta + \gamma) r_a^a \tag{70}$$

where λE is the latent heat flux of evaporation from the soil surfaces (W/m2), λT the latent heat fluxes of transpiration from canopy (W/m2), r_s^p the canopy resistance (s/m), r_a^p the aerodynamic resistance of the canopy to in-canopy flow (s/m), r_s^s the soil surfaces resistance (s/m), r_a^a and r_a^s the aerodynamic resistance from the reference height to in-canopy heat exchange plane height and from there to the soil surface (s/m), respectively, A_{sw} A_{SW}^s are the total available energy and the available energy to the soil (W/m2), respectively and defined as follow:

$$A_{sw} = R_n - G \tag{71}$$

$$A_{SW}^s = R_{n\,sw}^{\;s} - G \tag{72}$$

where $R_{n\,sw}^{\;s}$ is the net radiation fluxes into the soil surface (W/m2), and can be calculated using the Beer's law as follow:

$$R_{n\,sw}^{\;s} = R_n . \exp^{(-c.LAI)} \tag{73}$$

where c is the extinction coefficient of light attenuation (e.g.; Sene, 1994 indicate c=0.68 for fully grown plant, c=0 for bare soil; Zhang *et al.*, 2008 use 0.24 for vineyard crops).
The ***surfaces resistance*** is calculated as follow:

$$r_s^p = \frac{r_{st\,\min}}{LAI_{effective} \Pi_i F_i(X_i)} \tag{74}$$

where $r_{st\ min}$ is the minimal stomatal resistance of individual leaves under optimal conditions. $LAI_{effective}$ is: equal to LAI for LAI ≤2.0; LAI/2 for LAI ≥4.0 and 2 for intermediate values of LAI, X_i is a specific environmental variable, and $F_i(X_i)$ is the stress function with 0.0≤ $F_i(X_i)$ ≥1.0 (from: Jarvis, 1976).

$$F_1(S) = \left(\frac{S}{1100}\right)\left(\frac{1100 + a_1}{S + a_1}\right) \tag{75}$$

$$F_2(T) = \frac{(T - T_L)(T_H - T)^{(TH - a_2)/(a_2 - T_L)}}{(a_2 - T_L)(T_H - a_2)^{(TH - a_2)/(a_2 - T_L)}} \tag{76}$$

$$F_{_2}(D) = e^{-a_3 D} \tag{77}$$

$$F_4(\theta) = \begin{bmatrix} 1 - -if - - - - - \theta \geq \theta_F \\ \dfrac{\theta - \theta_W}{\theta_F - \theta_W} - if - - \theta_F < \theta < \theta_W \\ 0 - -if - - - - - \theta \leq \theta_W \end{bmatrix} \tag{78}$$

where S is the incoming photosynthetically active radiation flux (W/m^2), T is the air temperature (^oK), θ_F is the soil moisture at field capacity (cm^3/cm^3), θ_w is the soil moisture at wilting point (cm^3/cm^3), and θ is the actual soil moisture in the root zone. (cm^3/cm^3). T_H and T_L are upper and lower temperatures limits outside of which transpiration is assumed to cease (^{o}C) and are set at values of 40 and 0^{o}C (e.g.; Harris et al., 2004; Zhang et al., 2008). The a1, a2, and a3 are derived by multi-variate optimization, and are 57.67,25.78 and 9.65, respectively (Zhang et al., 2008).

The ***aerodynamic resistances*** r_a^a and r_a^s are calculated from the vertical wind profile in the field and the eddy diffusion coefficient. Above the canopy height, the eddy diffusion coefficient (K) is given by:

$$K = ku*(z-d) \tag{79}$$

where u* is the wind friction velocity (m/s), k is the van-Karman constant (0.41), z is the reference height (m), and d the zero plane displacement (m). The exponential decrease of the eddy diffusion coefficient (K) through the canopy is given as follow:

$$K = k_h.\exp\left[-n\left(1-\frac{z}{n}\right)\right] \tag{80}$$

where k_h is the eddy diffusion coefficient at the top of the canopy (m^2/s), and n is the extinction coefficient of the eddy diffusion. Brutsaert (1982) cited by Zhang et al., (2008) indicate that n =2.5 when h_c <1 m; n = 4.25 when h_c>10 m, linear interpolation could be used for crops with h between those values. k_h is determined as follow.

$$k_h = ku*(h_c-d) \tag{81}$$

The aerodynamic resistance r_a^a and r_a^s are obtained by integrating the eddy diffusion coefficients from the soil surface to the level of the "preferred" sink of momentum in the canopy, and from there to the reference height (Shutlleworth and Gurney, 1990, mentioned by Zhang et al., 2008) as follow:

$$r_a{}^a = \frac{1}{Ku*}\ln\left(\frac{z-d}{h_c-d}\right)+\frac{h_c}{nk_h}\left[\exp\left[n\left(1-\frac{zo+d}{h_c}\right)\right]-1\right] \tag{82}$$

$$r_a{}^s = \frac{h_c\exp^{(n)}}{nk_h}\left[\exp\left(\frac{-nz_o'}{h_c}\right)-\exp\left[-n\left(\frac{z_0+d}{h_c}\right)\right]\right] \tag{83}$$

The bulk boundary layer resistance of canopy is calculated as follow:

$$r_a{}^p = \frac{r_b}{2LAI} \tag{84}$$

where r_b is the mean boundary layer resistance (s/m) (e.g.; Brisson et al., 1998, recommend use 50 s/m).

The **soil surface resistance** r_s^s is the resistance to water vapour movement from the interior to the surface of the soil, and is strongly depending of the water content (θ_s), and is calculated using the Anandristakis et al. (2000) expression:

$$r_s^{\,s} = r_s^{\,s}{}_{\min} f(\theta_s) \tag{85}$$

where θ_s is soil volumetric water content (cm^3/cm^3), and $r_{s\ \min}^{s}$ is the minimum soil surfaces resistance, that correspond with the soil field capacity (θ_{FC}) and is assumed equal to 100 s/m (e.g.; Camilo and Gurney, 1986; Zhang et al., 2008).

The $f(\theta_s)$ is expressed according with Thompson (1981) and mentioned by Zhang et al. (2008) as follow:

$$f(\theta_s) = 2.5\left(\frac{\theta_{FC}}{\theta_s}\right) - 1.5 \tag{86}$$

4.5 Clumping model

The Clumping model is based in the Shulttleworth-Wallace model, this model separate the soil surfaces into fractional areas inside and outside the influence of the canopy, and include the fraction of canopy cover (*f*) in his calculation. Brenner and Incoll (1997) and Zhang et al. (2008) express the model as follow:

$$\lambda E = \lambda E^s + \lambda E^{bs} + \lambda T = f\left(C_c^s PM_c^s + C_c^p PM_c^p\right) + (1-f)C_c^{bs} PM_c^{bs} \tag{87}$$

where λE^s is the latent heat of evaporation from soil under the plant (W/m^2); λE^{bs} is the latent heat of evaporation from bare soil (W/m^2); *f* is the fractional vegetative cover and the other terms are expressed as follow:

$$PM_c^p = \frac{\Delta A_c + \left[\dfrac{\left(\rho CpD - \Delta r_a^p A_c^s\right)}{r_a^a + r_a^p}\right]}{\Delta + \gamma\left[1 + \dfrac{r_s^p}{r_a^a + r_a^p}\right]} \tag{88}$$

$$PM_c^s = \frac{\Delta A_c + \left[\dfrac{\left(\rho CpD - \Delta r_a^s A_c^p\right)}{r_a^a + r_a^s}\right]}{\Delta + \gamma\left[1 + \dfrac{r_s^s}{r_a^a + r_a^s}\right]} \tag{89}$$

$$PM_c^{bs} = \frac{\Delta A_c^{bs} + \left[\dfrac{\left(\rho CpD\right)}{r_a^a + r_a^{bs}}\right]}{\Delta + \gamma\left[1 + \dfrac{r_s^{bs}}{r_a^a + r_a^{bs}}\right]} \tag{90}$$

$$C_c^s = \frac{R_c^{bs} R_c^p \left(R_c^s + R_c^a\right)}{\left[R_c^s R_c^p R_c^{bs} + (1-f) R_c^s R_c^p R_c^a + f R_c^{bs} R_c^s R_c^a + f R_c^{bs} R_c^p R_c^a\right]} \tag{91}$$

$$C_c^p = \frac{R_c^{bs} R_c^s \left(R_c^p + R_c^a\right)}{\left[R_c^s R_c^p R_c^{bs} + (1-f) R_c^s R_c^p R_c^a + f R_c^{bs} R_c^s R_c^a + f R_c^{bs} R_c^p R_c^a\right]} \tag{92}$$

$$C_c^{bs} = \frac{R_c^s R_c^p \left(R_c^{bs} + R_c^a\right)}{\left[R_c^s R_c^p R_c^{bs} + (1-f) R_c^s R_c^p R_c^a + f R_c^{bs} R_c^s R_c^a + f R_c^{bs} R_c^p R_c^a\right]} \tag{93}$$

$$R_c^s = (\Delta + \gamma) r_a^s + \gamma r_s^s \tag{94}$$

$$R_c^p = (\Delta + \gamma) r_a^p + \gamma r_s^p \tag{95}$$

$$R_c^{bs} = (\Delta + \gamma) r_a^{bs} + \gamma r_s^{bs} \tag{96}$$

$$R_c^a = (\Delta + \gamma) r_a^a \tag{97}$$

Where A_c, A_c^p, A_c^s and A_c^{bs} are energy available to evapotranspiration, to the plant, to soil under shrub and bare soil (W/m2) respectively, r_a^{bs} the eddy diffusion resistance from in-canopy heat exchange plane height to the soil surface (s/m), r_s^{bs} the soil surfaces resistance of bare soil (s/m).

The ***Available energy*** for this model, the net radiation (R_n) is divided is divided into net radiation in the plant (R_n^p) and the net radiation in the soil (R_n^s). If the energy storage in the plant is assumed to be negligible, then:

$$R_{nc}^s = R_n \exp^{(-CLAI/f)} \tag{98}$$

$$R_{nc}^p = R_n - R_{nc}^s \tag{99}$$

$$A_c^s = R_{nc}^s - G^s \tag{100}$$

$$A_c^{bs} = R_n - G^{bs} \tag{101}$$

$$A_c^p = R_{nc}^p \tag{102}$$

Where R_{nc}^p and R_{nc}^s are the radiation absorbed by the plant and the radiation by the soil (W/m2) respectively, G^s and G^{bs} are the soil heat flux under plant and bare soil (W/m2) respectively, C is the extinction coefficient of light attenuation according for Sene (1994) is equal to 0.68 for fully grown plant.

The resistance for the bare soil surfaces r_s^{bs} can be calculated equally as in the S-W model, mentioned before. The aerodynamic resistance between the bare soil surface and the mean surfaces flow height (r_a^{bs}) can be calculated assuming that the bare soil surface is totally unaffected by adjacent vegetation so that is aerodynamic resistance equal to r_a^b and defined for:

$$r_a^b = \ln \frac{\left(\frac{Z_m}{Z'_o}\right)^2}{k^2 U_m} \tag{103}$$

Where Z_m is the mean surface flow height (m), and could be assumed equal to $0.75h_c$, and u_m is the wind speed at the Z_m (m/s).

According with Zhang et al. (2008), the aerodynamic resistance (r_a^{bs}) varies between r_a^b and r_a^s as *f* varies from 0 to 1, and the functional relationship of this change is not know.

4.6 Combination model

Theoretical approaches to surface evaporation from the energy balance equation combined with sensible heat and latent heat exchange expressions give the following definition for actual evapotranspiration (Pereira et al., 1999).

$$ET = \frac{\Delta}{\Delta + \gamma}\left[(R_n - G) + \frac{\rho Cp}{\Delta} Hu(VPDa - VPDs)\right] \tag{104}$$

Where R_n - G = available energy (MJ/m²) for the canopy, comprised of net radiation, R_n and the soil heat flux, G; $H(u)$ = exchange coefficient (m/s) between the surface level and a reference level above the canopy but taken inside the conservative boundary sublayer; *VPDs* and *VPDa* (kPa) = vapour pressure deficits (VPD) for the surface level and the reference level, respectively; ρ= atmospheric density (kg/m3); *Cp* = specific heat of moist air (J/kg7C); Δ = slope of the vapour pressure curve (Pa/ºC); and γ= psychrometric constant (Pa/ºC).
To obtain evapotranspiration with (104) the most difficult term to estimate is *VPDs*, representing the vapour pressure deficit at the evaporative surface. If *VPDs* can be associated with a surface resistance term (r_s), then ET can be calculated directly from the flux equation:

$$ET = \frac{\rho Cp}{\gamma}\frac{VPDs}{r_s} \quad (87) \tag{105}$$

and

$$r_a = \frac{1}{Hu} \tag{106}$$

r_a can be calculate using the equations discussed later. Two main solutions can be defined from (104) using climatic data:
1. The case of full water availability corresponding to saturation at the evaporative surface. Then *VPDs* = 0 and r_s becomes null. Eq. (104) then gives the maximum value for ET, the potential evaporation (EP), which depends only on climatic driving forces:

$$EP = \frac{\Delta(R_n - G) + \rho CpF(u)VPDa}{\Delta\lambda} \tag{107}$$

In where F(u) = 1/r_a. The combination the equations can get:

$$ET = \frac{EP}{(1 + \frac{\gamma}{\Delta + \gamma}\frac{r_s}{r_a})} \tag{108}$$

2. The case for equilibrium between the surface and the reference levels corresponds to VP*Ds* = VP*Da*. In this case, the evapotranspiration is referred to as the equilibrium evaporation (*Ee*).

$$Ee = \frac{\rho Cp}{\gamma}\frac{VPDa}{r_e} \tag{109}$$

Where r_s was renamed r_e, termed the equilibrium surface resistance, indicating that the term, in this case, represents the surface resistance for equilibrium evaporation. The value for r_e depends predominately on climatic characteristics although these characteristics are influenced by R_n and G of the vegetative surface. For purposes here, the r_e term can be called the climatic resistance for the surface.

$$r_e = \frac{\rho Cp}{\gamma}\frac{\Delta+\gamma}{\Delta}\frac{VPDa}{R_n - G} \tag{110}$$

EP can be estimate:

$$EP = Ee\left(1 + \frac{\gamma}{\Delta+\gamma} + \frac{r_e}{r_a}\right) \tag{111}$$

and ET can be estimate using:

$$ET = \frac{EP_{(36)}}{\left(1 + \frac{\gamma}{\Delta+\gamma}\frac{r_s}{r_a}\right)} \tag{112}$$

4.7 Pristley and Taylor model

Pristley and Taylor (1972), propose to neglected the aerodynamic term and fix the radiation term by introducing a dimensionless coefficient (α).

$$ET = \alpha\frac{\Delta}{\Delta+\gamma}\left(R_n - G\right) \tag{113}$$

where ET is water flux under references conditions (well watered grass) in mm.day^{-1}; R_n and G are net radiation and soil heat flux respectably in mm.day^{-1}; Δ and γ in kPa.°C^{-1}. The term α is given as 1.26 for grass field in humid weather conditions, and was adopted by Pristley and Taylor (1972) for wet surfaces; however α is ranging between 0.7 to 1.6, over various landscapes (Flint and Childs, 1991).
According with Zhang et al. (2004), the term α can be calculated as follow:

$$\alpha = \frac{\lambda E(\Delta+\gamma)}{\Delta(R_n - G)} = \frac{\Delta+\gamma}{\Delta(1+\beta)} \tag{114}$$

Also de α term sensible at the soil moisture changes (Eg.; Grago and Butsaert, 1992; Grago 1996; and Zhang et al., 2004), that relation can be estimated using a models like:

$$\alpha = k\left[1 - \exp\left(-c\frac{\theta - d}{\theta_{FC}}\right)\right] \tag{115}$$

where k, c and d are parameters of the model, θ is the actual volumetric soil moisture content (cm^3.cm^{-3}) and θ_{FC} is the volumetric moisture content at field capacity (cm^3.cm^{-3}).

4.8 Eddy covariance method

The eddy covariance method is, in general, the most preferred because it provide a direct measure of the vertical turbulent flux across the mean horizontal streamlines, provided by fast sensors (~10 Hz) (Meyers and Baldocchi, 2005). Realizing the limitation of the Thornthwaite-Holzman type of approach, Swinbank (1951) cited by Chang (1968) was the first to attempt a direct measurement by the so-called eddy correlation technique. The method is based on the assumption that the vertical eddy flux can be determined by simultaneous measurements of the upward eddy velocity and the fluctuation in vapour pressure. Actually is a routinely technique for direct measurement of surfaces layer fluxes of momentum, heat, and traces gases (CO_2, H_2O, O_3) between the surfaces and the turbulent atmosphere (Massmam, 2000).

This system recognizes that the transport of heat, moisture, and momentum in the boundary layer is governed almost entirely by turbulence. The eddy correlation method is theoretically simple using an approach to measure the turbulent fluxes of vapour and heat above the canopy surface. The eddy correlation fluxes are calculated and recorded in a 30 min or less temporal resolution. Assuming the net lateral advection of vapour transfer is negligible, the latent heat flux (evapotranspiration) can be calculated from the covariance between the water vapour density (ρ_v) and the vertical wind speed (w).

$$\lambda E = \lambda \overline{w' \rho_v '} \tag{116}$$

where λE is the latent heat flux (W m^{-2}), λ is the latent heat of vaporization (J kg^{-1}), ρ_v' is the fluctuation in the water vapour density (kg m^{-3}), and w′ is the fluctuation in the vertical wind speed (m s^{-1}). The over bar represents the average of the period and primes indicate the deviation from the mean values during the averaging period. According with Weaver (1992) the eddy correlation method depends on the relations between the direction of air movement near the land surface and properties of the atmosphere, such as temperature and humidity.

The sensible heat flux can be calculated from the covariance of air temperature and the vertical wind speed.

$$H = \rho_a C_p \overline{w'T'} \tag{117}$$

Where H the sensible heat flux (W m^{-2}), ρ_a the air density (kg m^{-3}), C_p the specific heat of moist air (J kg^{-1} ^{o}C^{-1}) and T′ the fluctuation in the air temperature (^{o}C).

The fine wire thermocouples (0.01 mm diameter) are not included in the eddy correlation system. The air temperature fluctuations, measured by the sonic anemometer, are corrected for air temperature fluctuations in estimation of sensible heat fluxes. The correction is for the effect of wind blowing normal to the sonic acoustic path. The simplified formula by Schotanus et al. (1983) is as follows:

$$\overline{w'T'} = \overline{w'T_s'} - 0.51\left(\overline{T + 273.15}\right)\overline{w'q'} \tag{118}$$

Where w′T′ is rotated covariance of wind speed and sonic temperature (m ^{o}C s^{-1}), T is air temperature (^{o}C) and q is the specific humidity in grams of water vapour per grams of moist air.

Two Eddy covariance systems are used to measure the water vapour fluxes, the open path and close path. According with Anthoni et al. (2001) the Open-path eddy covariance systems require corrections for density fluctuations in the sampled air (Massman and Lee, 2002; Webb et al., 1980) and in general closed-path system require incorporation of a time lag and corrections for the loss of high frequency information, due to the air being drawn through a long sampling tube (Massman and Lee, 2002; Moore, 1986).

The most common correction in the eddy covariance system is described by Wolf et al. (2008) as: i) Coordinate rotation, ii) Air density correction, and iii) Frequency –dependent signal loss.

Estimation of turbulent fluxes is highly dependent on the accuracy of the vertical wind speed measurements. Measurement of wind speed in three orthogonal directions with sonic anemometer requires a refined orientation with respect to the natural coordinate system through mathematic coordinate rotations (Sumner, 2001). The vector of wind has three components (*u*, *v*, *w*) in three coordinate directions (x, y, z). The z-direction is oriented with respect to gravity, and the other two are arbitrary. Baldocchi et al. (1988) provide procedures to transform the initial coordinate system to the natural coordinate system. Described in details by Sumner (2001), the coordinate system is rotated by an angle η about the z-axis to align *u* into the x-direction on the x-y plane, then rotated by an angle θ about the y-direction to align w along the z-direction. The results force $\bar{v}$ and $\bar{w}$ equal to zero, and $\bar{u}$ is pointed directly to the air stream. When θ was greater than 10 degrees, the turbulent flux data should be excluded based on the assumption that spurious turbulence was the cause of the excessive amount of the coordinate rotation.

$$\cos\theta = \sqrt{\frac{\left(u^2 + v^2\right)}{\left(u^2 + v^2 + w^2\right)}} \tag{119}$$

$$\sin\theta = \frac{w}{\sqrt{\left(u^2 + v^2 + w^2\right)}} \tag{120}$$

$$\cos\eta = \frac{u}{\sqrt{\left(u^2 + v^2\right)}} \tag{121}$$

$$\sin\eta = \frac{v}{\sqrt{\left(u^2 + v^2\right)}} \tag{122}$$

The latent heat and sensible heat fluxes are computed from the coordinate rotation-transformed covariance:

$$\left(\overline{w'\rho_v'}\right)_r = \overline{w'\rho_v'}\cos\theta - \overline{u'\rho_v'}\sin\theta\cos\eta - \overline{v'\rho_v'}\sin\theta\sin\eta \tag{123}$$

$$\left(\overline{w'T_s'}\right)_r = \overline{w'T_s'}\cos\theta - \overline{u'T_s'}\sin\theta\cos\eta - \overline{v'T_s'}\sin\theta\sin\eta \tag{124}$$

After the coordinate rotation, the final latent heat flux can be estimated from equation (116) plus the following correction of air density (C_{air}) (Webb et al., 1980) and correction of oxygen (CO_2) (Tanner and Greene, 1989).

$$C_{air} = \frac{\overline{\rho_v}\,\overline{H}}{\rho C_p (T + 273.15)}\lambda \tag{125}$$

$$C_{O2} = \frac{F K_o \overline{H}}{K_w (T + 273.15)}\lambda \tag{126}$$

Where F is the factor used in krypton hygrometer correction that accounts for molecular weights of air and oxygen, and atmospheric abundance of oxygen, equal to 0.229 g °C J^{-1}, K_o is the extinction coefficient of hygrometer for oxygen, estimated as 0.0045 m^3 g^{-1} cm^{-1}, K_w is the extinction coefficient of hygrometer for water, from the manufacture is 0.149 m^3 g^{-1} cm^{-1}.
With the measured four flux components from the energy balance equation, the energy balance should be closed, however, this is not practically the case. A tendency to underestimate energy and mass fluxes has been a pervasive problem with the eddy covariance technique (Ham and Heilman, 2003). Ham and Heilman (2003) reported clousure of 0.79 for priarine locations and 0.96 for forest. Ramirez and Harmsen (2007-Data without publication) indicate 0.71 for grass and 0.75 for corn.

- The errors in eddy covariance method could be associated with:

1. Accuracy of the R_n and G measurements (errors are often 5 to 10%).
2. The length scale of the eddies responsible for transport (if is larger, the frequency response and sensor separation error may have been smaller)
3. Sensor separation and inadequate sensor response (can generate 15% of Underestimation of λE by (Ham and Heilman, 2003) and 10% reported by Laubach and McNauhton (1999).
4. Ham and Heilman (2003) conclude *"The inherent tendency to underestimate fluxes when using eddy covariance may be linked to the errors caused by sensor separation and inadequate frequency response of the sensors. The correction proposed by Massman and Lee (2002) is difficult to implement for the non-specialist because they require calculation of cospectra using high-frequency (10Hz) data, and also is required expertise experience to interpret the cospectra properly"*

The "energy balance closure" is corrected using the Bowen ratio (Kosugi and Katsuyama, 2007) as follow:

$$H = \beta * \lambda E \tag{127}$$

$$\lambda E = Rn - G - H \tag{128}$$

Where: β and λE came from eddy covariance system, R_n and G are measured.
The Massman analytical formulae for spectral corrections to measuremed momentum and scalar fluxes for eddy covariance systems. Massman (2000) develop an analytical method for frequency response corrections, based in the Horst's (1997) approach develops as follow:
For *Stable atmospheric conditions* (0<ζ≤ 2).

a. Fast-response open path system

$$\frac{Flux_m}{Flux} = \left[\frac{ab}{(a+1)(b+1)}\right]\left[\frac{ab}{(a+p)(b+p)}\right]\left[\frac{1}{(p+1)}\right]\left[1 - \frac{p}{(a+1)(a+p)}\right] \tag{129}$$

b. Scalar instrument with 0.1-0.3s response

$$\frac{Flux_m}{Flux}=\left[\frac{ab}{(a+1)(b+1)}\right]\left[\frac{ab}{(a+p)(b+p)}\right]\left[\frac{1}{(p+1)}\right]\left[1-\frac{p}{(a+1)(a+p)}\right]\left[\frac{1+0.9p}{1+p}\right] \quad (130)$$

c. Unstable atmospheric conditions (ζ≤0)

$$\frac{Flux_m}{Flux}=\left[\frac{a^\alpha b^\alpha}{(a^\alpha+1)(b^\alpha+1)}\right]\left[\frac{a^\alpha b^\alpha}{(a^\alpha+p^\alpha)(b^\alpha+p^\alpha)}\right]\left[\frac{1}{(p^\alpha+1)}\right]\left[1-\frac{p^\alpha}{(a^\alpha+1)(a^\alpha+p^\alpha)}\right] \quad (131)$$

where the subscript m refers to the measurement flux, $a=2\pi\int x\tau_h$; $b=2\pi\int x\tau_b$; $p=2\pi\int x\tau_c$, and τ_h and τ_b are the equivalent time constant associated with trend removal (τ_h) and block averaging (τ_b). For relatively broad coespectra with relatively shallow peaks, such as the flat terrain neutral/stable flat terrain coespectrum α=0.925, and for sharper, more peaked coespectra, such as the stable terrain coespectra α=0.925 (Kaimal et al., 1972). These approximations are clearly easier to employ than numerical approaches and are applicable even when fluxes are so small as to preclude the use of *in situ* methods. Nevertheless, this approach is subject at the next conditions: i) horizontally-homogeneous upwind fetch, ii) the validity of the co-espectral similarity, iii) sufficiently long averaging periods, and preferably, iv) relatively small corrections (Massman, 2000).

4.9 The infrared surface temperature methods

The surface temperature has also been used for the estimation of the sensible heat flux (H) using the resistance model (Alves et al., 2000)

$$H=\rho.Cp\frac{To-Ta}{r_a} \quad (132)$$

Where ρ is air density (Kgm^{-3}), Cp specific heat at constant pressure ($Jkg^{-1}{}^oC^{-1}$), To is the temperature at surface level (°C), Ta is the temperature at the reference level (°C), and r_a is the aerodynamic resistance to heat flux between the surface and the reference level (sm^{-1}), the latent heat flux (λE) can be computed as the residual term in the energy balance.

$$\lambda E=Rn-G-H=Rn-G-\rho.Cp\frac{To-Ta}{r_a} \quad (133)$$

Alves et al. (2000) say the radioactive surface temperature has a several drawbacks. Thermal radiation received by the instrument can originate from the leaves but also from de soil, and the measurement can be highly dependent on crop cover, inclination of radiometer and sun height and azimuth, especially en partial cover crops, the first one lies in the use of an adequate value of r_a .

Where d is zero plane displacement height (m), Z_{oM} and Z_{OH} are the roughness lengths (m) for momentum and heat respectively, k is the von Karman constant, u_z is the wind speed (ms^{-1}) at the reference height z (m), and ψM and ψH are the integrated stability functions for describing the effects of the buoyancy or stability on momentum transfer and heat between the surface and the reference level.

The necessary instruments are: Wind speed and direction sensor at (0.85 and 1.46m), psychrometer at the same height that wind sensor, a net radiometer placement a 1.5 m and infrared thermometer perpendicular to the rows the crop, and positioned at an angle of 60° below horizontal to view the top leaves of the plants at 0.40 m distance. (Alves et al., 2000)
Sensible heat flux, H is calculated with the flux applied to levels Z_1 and Z_2.

$$H = \rho Cp \frac{T1 - T2}{[ra]_1^2} \tag{134}$$

$[ra]_1^2$ is the aerodynamic resistance to heat flux between the two levels, and is computed using the equation

$$[ra]_1^2 = \frac{\ln\left(\frac{Z2 - d}{Z1 - d}\right)}{ku*} \tag{135}$$

with u* the friction velocity, obtained in the process of determining aerodynamic parameter d and Z_{oM} from the win profile measurements.
The air temperature at the surface level (To) is calculate using:

$$To = Ta + \frac{Hr_a}{\rho Cp} \tag{136}$$

The stability conditions can be calculated using the Richardson number

5. Fetch requirements

The air that passing over a surface is affected by the field surfaces feature (Rosenberg et al., 1983); the minimal fetch requirement can be estimated based on the thickness of the internal boundary layer (δ in m) and a roughness parameter (Z_o in m) considering the minimal and maximal crop height during the grown season. The δ can be calculated using the relation proposed by Monteith and Unsworth (1990).

$$\delta = 0.15.L^{4/5}.Z_o^{\ 1/5} \tag{137}$$

where L is the distance of traverse (fetch) across a uniform surface with roughness Z_o. The Z_o for crops is approximately one order of magnitude smaller than the crop height h, and can be calculated according with Rosenberg et al. (1983) as follow:

$$\mathrm{Log}_{10} Z_o = 0.997 \log_{10} h - 0.883 \tag{138}$$

As a factor of safety a height to fetch of 1:50 to 1:100 is usually considered adequate for studies made over agricultural crop surfaces (Rosenberg et al., 1983, Allen et al., 1998) but may be too conservative and difficult to achieve in practice. Alves et al. (1998) obtained full profile development using a 1:48 fetch relation in Wheat and lettuce. Heilman et al. (1989) found that for Bowen-Ratio estimates a fetch 1:20 was sufficient over grass, and Ham and Heilman (1991) and Frithschen and Fritschen (2005) obtained similar results.

6. Stability correction

The gradient method need a stability correction, one of the most used is the Monin-Obukhov stability factor (ζ) described by (Rosemberg et al., 1983; Campbell, 1985; Prueger and Kustas;2005).

$$\xi = \frac{(-k.z.g.H)}{\left(\rho_a.C_p.T_a.u*^3\right)} \tag{139}$$

where k is von Karman's constant, z is height of wind and air temperature measurements (m), g is the gravitational constant (9.8 m.s^{-2}), H = β.λE, T_a is air temperature (^{o}K), u* is the friction velocity given by Kjelgaard et al. (1994) without the stability correction factor:

$$u* = \frac{k.u_z}{\ln\left(\frac{z - d + Z_{om}}{Z_{om}}\right)} \tag{140}$$

flux with a negative sign for ζ indicate unstable conditions and needs to be exclude, in flux under unstable conditions the λE is over R_n (**Fig.5a**), when the flux with negative ζ are exclude, λE is low that R_n (**Fig.5b**).

Payero et al. (2003) indicate that fluxes with incorrect sign and β ≈ -1 should not be considered when estimated the energy balance components by the energy balance Bowen ratio method. The negative ζ are corresponded with negative β (**Fig.6**).

The Richardson number (Ri) represented by the equation 44, also is well know as stability factor (e.g.; Alves et al., 2000; Tolk, et al., 1995) and represent the ratio of the buoyancy – "thermal effect" to mechanical –"wind shear" (Prueger and Kustas, 2005). Negative values indicate instability conditions where surfaces heating enhances buoyancy effects, and positive Ri values indicate a stable conditions where temperature near the surfaces are cooler than away from the surfaces.

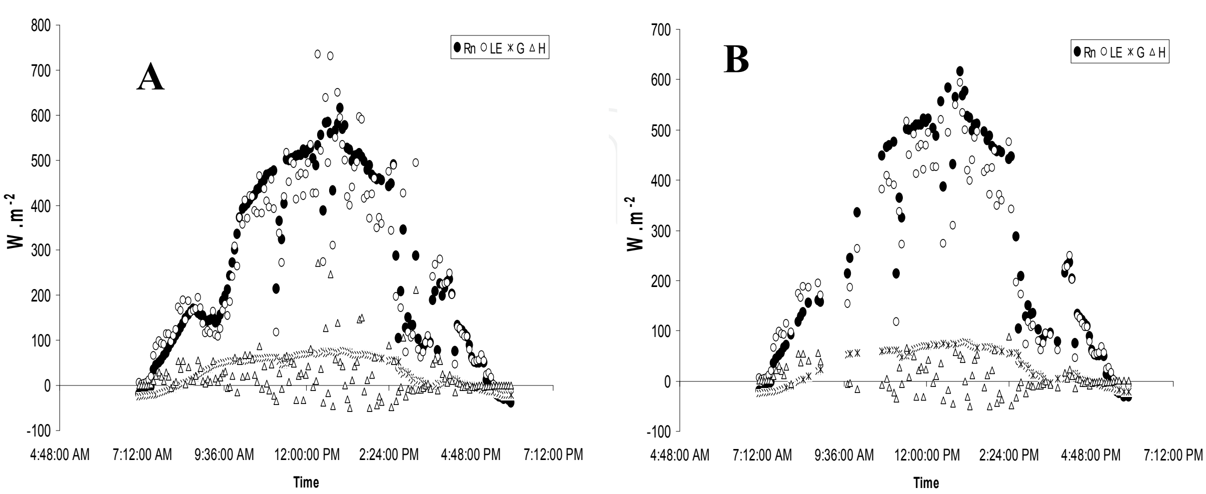

Fig. 5. Energy balance components measured by Bowen ratio method in grass **A**. without stability correction and **B.** with stability correction.

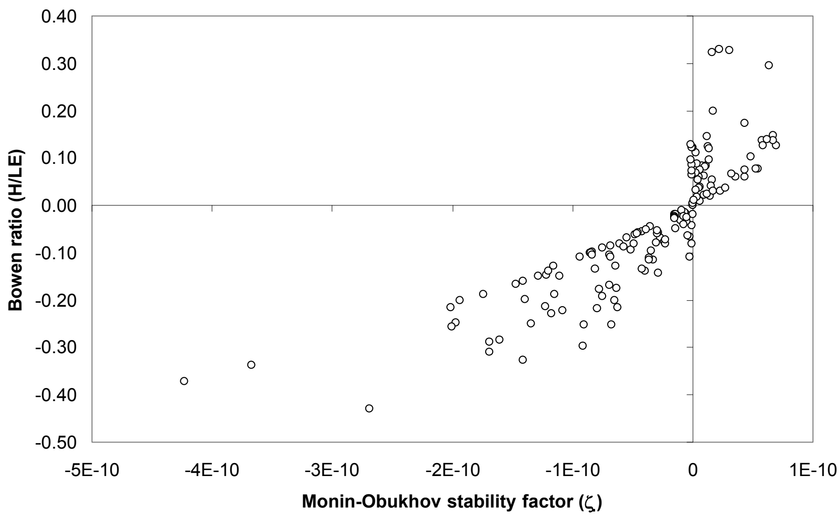

Fig. 6. Relationship between Bowen ratio (β)) and the Monin-Obukhov stability factor

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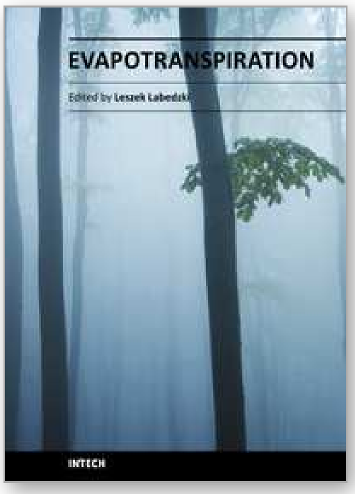

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Evapotranspiration is a very complex phenomenon, comprising different aspects and processes (hydrological, meteorological, physiological, soil, plant and others). Farmers, agriculture advisers, extension services, hydrologists, agrometeorologists, water management specialists and many others are facing the problem of evapotranspiration. This book is dedicated to further understanding of the evapotranspiration problems, presenting a broad body of experience, by reporting different views of the authors and the results of their studies. It covers aspects from understandings and concepts of evapotranspiration, through methodology of calculating and measuring, to applications in different fields, in which evapotranspiration is an important factor. The book will be of benefit to scientists, engineers and managers involved in problems related to meteorology, climatology, hydrology, geography, agronomy and agricultural water management. We hope they will find useful material in this collection of papers.

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