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CHAPTER I

INTRODUCTION

Evaporation is important to crop growth since it dictates to a large degree the soil moisture that is available. From a practical viewpoint of scheduling irrigation a knowledge of day-to-day water loss is desirable. Several approaches to the evaluation of evaporation are currently in use. Lysimetric and water balance approaches estimate the amount of water lost from the soil. Large, expensive lysimeters have provided very reliable estimates (King, Tanner and Suomi 1956, Pruitt and Lourence 1966) while other soil moisture techniques (neutron probe and gravimetric) have provided less reliable results due mainly to sampling problems. Microclimatological methods provide alternative approaches. Most of these methods require precision measurements, sometimes at several heights in the lower atmosphere, which cannot be attained routinely. In principle, the combination model, first proposed by Penman (1948) and since revised by Penman (1961), McIlroy (Slatyer and McIlroy 1961), Monteith (1965) and Tanner and Fuchs (1968), provides an attractive and practical solution to evaporation assessment. It combines equations dealing with the energetics of evaporation and atmospheric turbulence to provide a method of calculating evaporation using measurements at only one height. Because it includes all the factors which influence evaporation, it is useful in demonstrating the interaction and relative
importance of these. However, it requires parameterization of the underlying surface and turbulent transfer mechanism.

This study examines the performance of several combination model forms and the control exerted by the underlying surface during soil moisture drying cycles.

A field experiment was carried out in 1971 in which two grass sites were instrumented identically. Over one, the control site, evaporation varied only in response to natural precipitation and crop growth. Over the other, the experimental site, evaporation varied with moisture and growth conditions produced by irrigation. This procedure allows the effects of soil moisture and crop variables to be isolated. Data from both sites collected before the first irrigation are used to compare instrumental performance under similar crop and soil moisture conditions. Data collected following the first irrigation are used to isolate the effects of soil moisture since the crop conditions had not changed in response to increased soil moisture on the experimental site. The effects of different crop conditions are shown using data when soil moisture at both sites was similar. Such conditions were satisfied after long drying periods and after precipitation. In addition, data from either site are used to show the effects of drying cycles on energy exchange and the combination models.
CHAPTER II

THEORY

In this chapter, the underlying theory of the energy balance approach to evaporation and the combination model are developed. The use of an Ohm's Law analogy to describe diffusion processes is also introduced.

The Energy Balance Approach to Evaporation:

The conservation of energy principle is the basis of the energy balance approach to climatology. At a cropped surface, the principle can be expressed as

\[ R_n - H - \lambda E - G - P = 0 \]  

Here, energy is considered as a flux density (\( Wm^{-2} \)). In the equation, \( R_n \) is the net energy received at the surface from the balance of the radiative energy fluxes, \( P \) is the energy used in photosynthesis, \( G \) is the heat flux into the ground, \( H \) is the convective, sensible heat flux into the air and \( \lambda E \) is the latent heat flux into the air. The energy used in photosynthesis is often less than 5% of the net radiation (Yocum, Allen and Lemon, 1964) and is usually neglected.

\[ \text{(1)} \]

1 A complete list of symbols and units is given in Appendix A.
Direct evaluation of the fluxes of sensible and latent heat is still a difficult problem. However, they can be calculated from microclimatological measurements using the Bowen ratio solution to the energy balance equation (Bowen 1926). Equation (1), neglecting $P$, can be rearranged to give

$$R_n - G = \lambda E (1 + H/\lambda E),$$  \hspace{1cm} (2)

in which $H/\lambda E = \beta$, the Bowen ratio. Hence,

$$\lambda E = (R_n - G)/(1 + \beta).$$  \hspace{1cm} (3)

From mass transfer theory,

$$H = - \rho C_p K_h (\partial T/\partial z)$$  \hspace{1cm} (4)

and

$$\lambda E = - \rho C_p/\gamma K_w (\partial e/\partial z),$$  \hspace{1cm} (5)

in which $C_p$ is the heat capacity of air, $\rho$ is the density of air, $K_h$ and $K_w$ are the eddy diffusivities of heat and water vapour respectively, $\partial T/\partial z$ and $\partial e/\partial z$ are the vertical gradients of temperature and vapour pressure, and the bar denotes that the gradients are time-averaged.

Then, the Bowen ratio can be expressed as

$$\beta = (\gamma K_h \partial T/\partial z)/(K_w \partial e/\partial z).$$  \hspace{1cm} (6)
Using finite difference approximations instead of differentials and assuming that $K_h = K_w$ (Swinbank and Dyer 1967 and Dyer 1967),

$$\lambda E = \frac{(R_n - G)}{(1 + \gamma \Delta T / \Delta e)} \quad .$$  \hspace{1cm} (7)

The evaporative fluxes may then be calculated if the temperature and vapour pressure at two levels, net radiation and soil heat flux are measured. Sensible heat flux can be calculated as a residual in the energy balance.

This procedure is only valid within the boundary layer since it assumes that the fluxes of latent and sensible heat are constant with height.

**The Combination Model**

The model combines the energy balance approach with turbulent transfer theory.

Using an Ohm's Law analogy, a flux ($F$) may be regarded as a potential difference ($\Delta X$) divided by a resistance ($R$);

$$F = \frac{\Delta X}{R} \quad .$$

Assuming that the diffusivities of heat and water vapour are equal, the potential differences from the surface to a height $z$ (Figure 1) across which the fluxes of sensible and latent heat act ($\Delta T = T_o - T_z$ and $\Delta e = e_o - e_z$) are opposed by the same aerodynamic resistance $r_a$. The fluxes of sensible and latent heat can then be defined as
Figure 1 - Resistance Network Model Including Aerodynamic Resistance ($r_a$), Dry Surface Resistance ($r_d$), and Stomatal Resistance ($r_l$) which is part of the total plant resistance ($r_p$). The surface resistance ($r_s$) is composed of $r_p$ and $r_d$ in parallel.

Figure 2 - Saturation Vapour Pressure - Temperature Curve
\[ \lambda E = \left( \frac{C_p}{\gamma} \right) \left( \Delta e / r_a \right) \]  

and

\[ H = \frac{C_p(\Delta T)}{r_a} \]  

where the aerodynamic resistance is defined as the time (secs.) for a unit volume of air \( m^3 \) to exchange heat or water vapour with a unit area of surface \( m^2 \). From the saturation vapour pressure-temperature curve and the psychrometric equation (Figure 2),

\[ \Delta e = S\Delta T_w - \gamma \Delta D \]  

and

\[ \Delta T = \Delta T_w + \Delta D \]  

where \( \Delta D = D_0 - D_z \), and \( D = T - T_w \), the wet bulb depression, and \( T \) and \( T_w \) are the dry and wet bulb temperatures respectively. Substituting equations (10) and (11) into equations (8) and (9),

\[ \lambda E = \left( \frac{C_p}{\gamma r_a} \right) \left( S \Delta T_w - \gamma \Delta D \right) \]  

and

\[ H = \left( \frac{C_p}{r_a} \right) \left( \Delta T_w + \Delta D \right) \]  

Combining equations (12) and (13) with the energy balance equation (1) gives the combination model for actual evaporation,
Potential evaporation may be defined simply as the evaporation from any thoroughly wet surface. In this state, the air next to the surface is saturated, \( D_0 = 0 \) and equation (14) reduced to

\[
\lambda E = \frac{S(R_n - G)}{(S + \gamma)} - \frac{\rho \Delta D}{r_a} \tag{14}
\]

This is a simpler and clearer definition of \( \lambda PE \) than that proposed by Penman (1948).

McIlroy considered the special case, which he called equilibrium evaporation (Slatyer and McIlroy 1961), when \( D_0 = D_z \). Then, equation (14) reduces to

\[
\lambda PE_s = \frac{S(R_n - G)}{(S + \gamma)} \tag{15}
\]

This case can occur in two different sets of conditions. Firstly, if an evaporating surface is located downwind from a very long fetch over a very wet surface, then the air above that surface may become saturated and \( D_0 = D_z = 0 \). This is referred to as the "mid-oceanic" case. Secondly, it occurs if \( D_0 = D_z \neq 0 \). It will be shown later that this second state is closely approximated over very dry, cropped surfaces.

In all cases, the disappearance of the aerodynamic resistance from the equation means that the equilibrium rate of evaporation is independent of wind speed.

The underlying surface influences all three forms of the
combination model through net radiation and soil heat flux. In addition both λPE and λE models contain $r_a$ which is surface dependent and the λE model contains $D_o$. At present, the inclusion of $D_o$ in the equation for λE prohibits the use of that model, although Tanner and Fuchs (1968) have made some progress in operationalizing it over unvegetated surfaces by reducing the problem to a measurement of surface temperature.

Monteith (1965) avoids the evaluation of $D_o$ by an extension of the Ohm's Law analogy to the surface. Initially, he considered evaporation from a single leaf. In Figure 1, water evaporating from the leaf moves across the potential difference $(e_i - e_o)$, where $e_i$ and $e_o$ are the vapour pressures at the stomatal wall and the leaf surface respectively, against a stomatal resistance $(r_i)$. It is then subject to another potential difference $(e_o - e_z)$ and the aerodynamic resistance $(r_a)$. Since the flux of vapour across the two potentials must be equal,

$$\lambda E = \left(\frac{gC_p}{\gamma r_i}\right)(e_i - e_o) = \left(\frac{gC_p}{\gamma r_a}\right)(e_o - e_z) \quad (17)$$

Assuming that the vapour pressure at the stomatal wall is the saturation vapour pressure at the temperature of the wall, then

$$e_i = e_s T_i$$

Since within a thin leaf, the temperatures of the stomatal wall and the surface $(T_o)$ can be assumed equal,

$$e_i = e_s T_o$$
and equation (17) becomes

$$\lambda E = \left(\frac{\chi C_p}{\gamma r_1}\right) (e_s T_o - e_o) = \left(\frac{\chi C_p}{\gamma r_a}\right) (e_o - e_z)$$  \hspace{1cm} (18)

Therefore,

$$e_s T_o - e_z = (1 + r_1/r_a)(e_o - e_z)$$  \hspace{1cm} (19)

For a canopy, Monteith replaces $r_1$ with a surface resistance ($r_s$) and $T_o$ and $e_o$ become the surface temperature and vapour pressure. Using

$$D = (e_s T - e)/(S + \gamma)$$

to rewrite equation (14) in terms of vapour pressures and substituting for $(e_s T_o - e_z)$ gives

$$\lambda E = \frac{S(Rn - G) + \chi C_p (e_s T_z - e_z)/r_a}{S + \gamma + \gamma r_s/r_a}$$  \hspace{1cm} (20)

Surface control has been shifted from the $D_o$ term in McIlroy's formulation to the surface resistance $r_s$. Under potential conditions it is assumed that $r_s = 0$, and

$$\lambda PE = \frac{S(Rn - G) + \chi C_p (e_s T_z - e_z)/r_a}{S + \gamma}$$  \hspace{1cm} (21)

The Aerodynamic and Surface Resistances

Aerodynamic Resistance

Under conditions of neutral stability, the total kinetic energy
will be solely due to mechanical or forced convection. In this case, the vertical gradient of wind is given by

\[ \frac{\partial u}{\partial z} = \frac{u^*}{kz} \quad , \quad (22) \]

where \( u^* \) is the friction velocity and \( k \) is Von Karman's constant. The logarithmic wind profile model is obtained by integrating with respect to height

\[ u = \left( \frac{u^*}{k} \right) \ln \left( \frac{z}{z_0} \right) \quad . \quad (23) \]

Here \( z_0 \), the constant of integration, is usually referred to as the surface roughness parameter. In the case of vegetated surfaces, the wind profile is effectively displaced upwards by a length \( d \) (referred to as the zero-plane displacement) and equation (23) becomes

\[ u = \left( \frac{u^*}{k} \right) \ln \left( \frac{z-d}{z_0} \right) \quad . \quad (24) \]

Re-arranging,

\[ u^* = \left( ku \right) \ln \left( \frac{z-d}{z_0} \right) \quad . \quad (25) \]

From the mass transfer equation for momentum (\( \tau \)),

\[ \tau = \mu \frac{K_m}{z} \frac{\partial u}{\partial z} \quad , \quad (26) \]

where \( K_m \) is the eddy diffusivity for momentum, the aerodynamic resistance
may be stated as

\[ r_a = \int_0^z \frac{dz}{K_m + \nu} \quad (27) \]

in which \( \nu \) is the kinematic viscosity of air. Since \( \nu \) is negligible compared to \( K_m \), it will be ignored. From equations 22 and 26,

\[ K_m = k \, z \, u^* \]

and substituting in equation (27),

\[ r_a = \frac{1}{ku^*} \int_0^z \frac{dz}{z} = \frac{1}{ku^*} \ln(z/z_0) \]

Substituting for \( u^* \) from equation (25),

\[ r_a = \frac{[\ln(z-d)]^2}{k^2u} \quad (28a) \]

This equation is only valid when the atmosphere is in a state of neutral equilibrium, but Tanner and Fuchs (1968) have shown that it may be adjusted for departures from neutral stability by using the KEYPS correction \( \phi \) (Kazanski and Monin 1956, Ellison 1957, Yamamoto 1959, Panofsky 1961 and Sellers 1962),

\[ \phi = \frac{1}{[1 - 18R_d]^{-1/4} - 1]} \, d\ln(z-d) \]

where \( R_d \) is the gradient form of the Richardson number which defines
atmospheric stability. The aerodynamic resistance may then be calculated under most conditions of stability from

\[ r_a = \frac{[\ln\left(\frac{z-d}{z_o}\right) + \phi]^2}{k^2 u} \]  

(28b)

**Surface Resistance**

Using equation (18), Monteith evaluated surface resistance by extrapolating the profiles of temperature, vapour pressure and wind speed to the plane of zero wind speed. This he interpreted as the surface. Philip (1966) and Tanner and Fuchs (1968) objected to this strongly on the basis that the sources of heat and water vapour would not normally correspond to the level of the momentum sink. Szeicz and Long (1969) countered that to derive a working value of the surface resistance, it is unnecessary to assign any specific level to the measurement of \( T_o \) and \( e_o \) or to define the actual heights of sources and sinks. In addition, they showed for a grass surface that the extrapolation method gave values to within 1-2°C of values obtained independently.

If \( \lambda E \) is measured, the surface resistance may also be obtained from equations (20) and (21) as

\[ r_s = r_a (1 + S/\gamma)(\frac{PE}{E} - 1) \]  

(29)

The use of this approach raises a problem which, hitherto, has received little attention. It is difficult to obtain an accurate measure of \( \lambda PE \). This has commonly been obtained using data collected when \( D_o \neq 0 \). The results from this method can be misleading since from the definition,
potential evaporation can only be evaluated when field conditions closely approximate a very wet surface. It will be shown later that equation (21) grossly overestimates potential evaporation in non-potential conditions.

The surface resistance is interpreted as the resistance to the diffusion of water vapour from the vapour source to the effective surface (Figure 1); that is the plane where aerodynamic resistance begins to act. Over a cropped surface, the total surface resistance is composed of two individual resistances in parallel. Water evaporating directly from the soil surface is opposed by a resistance created by decreased capillary conductivity in the dry surface layer between the surface and the vapour source (Tanner and Fuchs 1968). Water transpiring from the crop is opposed by a plant resistance which, in detail, is composed of several individual resistances (Cowan and Milthorpe 1968). Cowan and Milthorpe (1968) have shown that the plant resistance is not linearly related to soil moisture, but that the resistance remains constant until soil moisture becomes limiting. Under these conditions, plant resistance increases when stomates close to prevent water loss. Apparently, the plant is able to extract enough water from deeper sources to maintain a constant transpiration rate.

Unless water limits transpiration, variations in the total surface resistance are then due to variations in the resistance to water vapour diffusion through the dry surface layer. This means that surface soil moisture plays an important role in determining the surface resistance.
CHAPTER III

SITE, INSTRUMENTATION AND CALCULATION

Site

The study was conducted at the Ontario Horticultural Experimental Station near Simcoe, Southern Ontario, during the months June to October of 1971. The site was a flat plot (122 x 134 m) surrounded on three sides by mixed crops and on the western side by a road (Figure 3). It was planted in perennial ryegrass and red clover in early May. The ryegrass constituted the bulk of the crop while the clover was planted to obtain a rapid ground cover. At the start of the field program, the ground cover was approximately seventy-five percent.

The plot was divided into southern and northern halves, each 122 x 67 m. The southern half (field A) was irrigated. Before commencing each run, both fields were mowed to a height of 0.10 m and field A was irrigated.

Instrumentation

Net radiation and soil heat flux was measured independently on each field as were dry and wet-bulb temperatures at three levels (0.25, 0.50, and 0.75 m). Wind speeds were measured at four levels (0.25, 0.50, 1.00 and 2.00 m) at a location on the boundary between the two fields. In addition, rainfall on field B and rainfall and irrigation on field A were measured with standard rain gauges.
LEGEND

- Fence
- Instrument Hut
- Shrubs
- Temperature and Humidity
- Net Radiation
- Soil Heat Flux
- Rainfall
- Windspeed

FIGURE 3 - The Site

SCALE 1:1130
Figure 3 shows the locations of all the instruments.

Net Radiation

Net radiation was measured over both sites with shielded net radiometers (Swissteco, Type S-1) mounted 1.25 m above the surface (Figure 4). In order to avoid shadows cast by the mast, the radiometers were pointed towards south. To keep the polyethylene domes inflated and to prevent condensation within them, the radiometers were purged with dry air using small aquarium pumps and a desiccant (Silica Gel). The return air flow from the domes was led into a bubbler. A bubble rate of 50 to 60 per minute was attained to conform with the findings of McCaughey (1968).

Soil Heat Flux

Soil flux was measured with two soil heat flux plates (Middleton Pty, Ltd.) wired in series and a divergence thermopile (Figure 5). The transducers were suspended by wires on a small stainless steel frame to maintain them in the same plane. Legs were attached to the frame such that they projected 0.05 m above the plane of the frame and 0.25 m below that plane. These served a two-fold purpose. Firstly, they ensured that when the system was installed in the soil, the frame would not lose its position parallel to the surface. Secondly, as long as the tops of the legs were visible at the surface, the plane would be precisely 0.05 m below the surface.

To compensate for vertical heat flux divergence in the soil ($\Delta G$) between the depth of the plates and the surface, the method suggested by Van Wijk (1965) and Fuchs and Tanner (1968) was used. Here the surface soil heat flux ($G_o$) is given by,
FIGURE 4 - SWISSTECO NET RADIOMETER
FIGURE 5 - THE SOIL HEAT FLUX SYSTEM
\[ G_o = G_z + \Delta G = G_z + C(\Delta T/\Delta t) \ z, \]  

(30)

in which \( G_z \) is the soil heat flux measured at depth \( z \) by the transducers, \( (\Delta T/\Delta t) \) is the temperature change in the top 0.05 m soil layer in time \( \Delta t \) and \( C \) is the heat capacity of the soil. \( \Delta T \) was obtained from temperature measurements with the thermocouples wired in series and set at constant depth intervals between the surface and the soil heat flux plates.

The heat capacity of the soil was determined from

\[ C = C_m X_m + C_o X_o + C_w X_w + C_a X_a, \]  

(31)

where \( X \) is the volume of the material in the soil and the subscript refer to mineral (m), organic (c), water (w) and air (a). The heat capacity of air is small enough to be neglected. De Vries (1963) suggested values of \( C_m = 0.46 \), \( C_o = 0.60 \) and \( C_w = 1.0 \) cal cm\(^{-3}\) °C\(^{-1}\), so that equation (31) reduces to

\[ C = 0.46X_m + 0.60X_o + X_w. \]  

(32)

Using the values of \( X_m = 0.459 \) and \( X_o = 0.024 \) determined from previous work at Simcoe (Wilson and McCaughey, 1971),

\[ C = 0.225 + X_w. \]

Hence heat capacity depends only on volumetric measurements of soil moisture.
Soil Moisture

The surface soil moisture in the top 0.05 m was sampled gravimetrically using 5 samples of approximately 100 grams from each field. Samples were taken at the same time each day 0830 mean solar time (MST). This time was selected because dew, which may have produced overestimates of true soil moisture, had generally evaporated by then. In order that the soil moisture measurements would refer specifically to the soil heat flux measurements, samples were always taken within five meters of the soil heat flux systems. However, care was taken to sample far enough away to ensure that the sensors would not be disturbed.

Temperature and Humidity Profiles

a) Construction of Probes

The temperature and humidity sensors were constructed following a modified Pruitt and Laurence (1969) design. Five-junction thermopiles were constructed from 30-gauge copper-constantan thermocouple wire. The tips of these thermopiles were passed through 4.76 mm diameter stainless steel tubing and potted with polyester resin within 4.76 mm diameter aluminum tubing. The stainless steel provided the probes with rigidity while the aluminum ensured good conduction of heat to the junctions. The exposed wire between the two ends of the thermopile was sheathed with 6.25 mm PVC plastic tubing. This tubing was sealed onto the stainless steel portion of the probe, ensuring that the entire thermopile was waterproof. The PVC tubing also prevented radiative heating of the thermocouple wire and subsequent conduction to the sensing tips.

The probes were calibrated against a platinum thermometer over
a temperature range from 0°C to 30°C, both before and after the field season. This calibration remained the same at

\[ \Delta T = 0.023 + 5.1238 \text{ (mV)} - 0.0238(\text{mV})^2 \]

where \( \Delta T \) is the temperature difference between the hot and cold junctions and mV is the millivolt output of the thermopile. Since all thermopiles were constructed in the same manner, using wire from the same spool, the same calibration was applied to all of them. During the calibrations, the response time of the probes was determined to be approximately one minute.

The reference junctions of each thermopile were potted in silicon sealant inside a small vacuum flask which was emersed in an ice and water bath. This ensured that once the reference pot reached 0°C, it would remain at that temperature and no temperature differences between the reference junctions of the probes would exist. The temperature of the pot was monitored by a platinum resistance thermometer.

b) Construction of the Temperature and Humidity Masts

The vertical masts were constructed of 25.40 mm ID, 35.0 mm OD galvanized pipe (Figure 6). The horizontal arms were constructed of 35.0 mm OD PVC pipe. One end of each horizontal arm was fitted with a plastic T-junction and the other end was fitted with a small Rotron fan which was used to aspirate the system. The sensors were held in position projecting horizontally and at right angles to the cross-arms by a plug which was inserted inside the T-junction. The plugs were constructed of 35.0 mm PVC pipe which had been turned down along most of its length so
**Figure 6 - Construction of Temperature and Humidity Mast**

- **25.4 mm ID Galvanized Pipe**
- **23 mm Power Cable**
- **35.0 mm OD PVC Pipe**
- **35.0 mm PVC Pipe**
- **50.8 mm OD Cast Acrylic Tubing**
- **50.8 mm Cast Acrylic Tubing**
- **35.0 mm PVC Pipe**
- **End Plug**
- **Reservoir**
- **ROTTRON FAN**
- **THERMOPILE**
- **PVC T-JUNCTION**
- **CAST METAL T-JUNCTION**
- **ADJUSTMENT CLIP**
- **TYGON TUBING**
- **INNER RADIATION SHIELD (35.0 mm PVC PIPE)**
- **OUTER RADIATION SHIELD (50.8 mm OD CAST ACRYLIC TUBING)**
that it would fit inside the T-junction, and were filled with polyester resin. A hole 4.76 mm diameter was drilled lengthwise through the center of the plug to accommodate the wet bulb sensor. A second hole of the same size was drilled along the length of the plug to accommodate the dry bulb sensor near the side of the plug. A third hole was drilled obliquely from the side of the plug to intersect the center hole and fitted with a small piece of stainless steel tubing. This tubing conducted wicking through the plug to the wet bulb. The center hole was then enlarged to 12.7 mm diameter from the inside end of the plug to the intersection of the wicking tube. This ensured that the wick would not be contaminated by contact with the polyester resin. Figure 7a illustrates the construction of the end plug.

The water reservoir was constructed of polyethylene plastic tubing sealed at either end by a rubber stopper. A small piece of stainless steel tubing was inserted through the stopper at one end of the reservoir. To minimize radiant heating of the water in the reservoir, it was covered with mylar tape. This was necessary to ensure that the water reaching the wet bulb was not warmer than the air. The reservoir was mounted on the horizontal arm by a small clip which allowed vertical adjustment.

The wicking was led from the reservoir through a short piece of tygon tubing to the metal tube in the plug. The wick was slit in order to allow the wet bulb probe to be inserted inside it. Figure 7b shows the final arrangement of the sensors and wicking in the end plug.

To shield the sensors from radiant heating, double radiation shields were used. The outer radiation shield was constructed of 50.80 mm
FIGURE 7a - CONSTRUCTION OF END PLUGS

FIGURE 7b - THE BASIC DRY AND WET-BULB SYSTEM
cast acrylic tubing and the inner shield of 35.0 mm (OD) PVC pipe. Both shields were sprayed on the inside with flat black enamel paint and covered on the outside with mylar tape. Since mylar tape has a high reflectivity and emissivity, direct radiant heating of the probes was minimized. The flat black paint minimized reflection of any radiation which entered the shields onto the probes. Conduction of heat from the radiation shields to the sensors was prevented by aspiration. Air was drawn past the sensors inside the inner shield and through the space between the inner and outer shields.

The wet-bulb sensor was centered in the shields by the plug which held it. The dry-bulb sensor projected beyond the wet-bulb and was centered by a small plastic spacer. Figures 8 and 9 show the components of the mast and the mast in situ, respectively.

Before the experiment commenced, it was realized that if the temperature and humidity masts were centered on the two fields, there would be inadequate fetches to them. From previous work at Simcoe, it was known that wind blew most frequently from the southwesterly sector of the field. For these reasons, the instruments were located in the northeasterly sectors of each field. This meant that the experiment could only be operated when southerly wind blew. Figure 10 presents wind direction data collected between June 3 and August 17. It shows clearly that this locational procedure was warranted. Figure 11 shows fetches for different directions on the two fields.

**Wind Profiles**

Half-hourly means of wind speed were measured at 0.25, 0.50, 1.00
FIGURE 8 - COMPONENTS OF THE
THE TEMPERATURE AND
HUMIDITY MAST

FIGURE 9 - TEMPERATURE AND HUMIDITY
MAST IN SITU
FIGURE 10 - WIND ROSE (values in percentages)
FIGURE II - HEIGHT - FETCH RATIOS TO TEMPERATURE MASTS
and 2.00 m using a Thornthwaite Wind Profile System (Figure 12). However, early in the field season, the 2.00 m level ceased operation and only the remaining three could be used.

Relative calibrations of the system were performed periodically to check that all levels behaved the same. This was accomplished by placing all anemometers at one level and using one anemometer as a reference against which all other anemometers were calibrated. The corrections required were never greater than two percent.

Figure 13 shows the height to fetch ratios for the various wind directions. Adequate height to fetch ratios may be determined by examining wind profiles which have been derived under conditions of neutral stability. Under such conditions, adjusted profiles will be confirmed if the relationship between wind speed and the logarithm of the height at which the wind speed is measured is linear. Figure 14 shows four of these profiles. The profiles were selected because they represent times when the wind blew from the direction of minimum fetch (ESE). These data indicate that they were derived within the boundary layer. This means that adjusted profiles could be expected during southerly winds.

**Crop Parameters**

Crop height was measured simply with a meter scale. In order to obtain values which were representative of the field, a systematic walk was used which covered most sections of the field (Figure 15). Twenty to thirty measurements were taken on each field and meaned. Samples were taken approximately every three days.

Crop cover was estimated using a meter-square grid which was
FIGURE 13 - HEIGHT-FETCH RATIOS TO WIND MAST
FIGURE 14 - ADJUSTED WIND PROFILES

A - JUNE 11
B - JUNE 17
C - JUNE 25
D - AUGUST 15
FIGURE 15 - SYSTEMATIC PATTERN USED IN SAMPLING CROP PARAMETERS
divided into 25 sections. The same walk pattern was used as for crop height but fewer locations were sampled. At each location, the percentage of ground covered with grass to the total ground was estimated. The mean of these twenty-five estimates was considered representative of that location. The mean of all the location estimates served as the crop cover estimate for the field. As the crop cover was slower to change than crop height, it was sampled less frequently.

Recording and Analysis

All signals were conducted through shielded cable to a junction box inside the instrument hut. All signals were grounded to a copper rod inserted to a depth of approximately 1.50 m in the ground outside the hut. Millivolt signals were measured every minute using a Solartron type LY - 1771 Compact Data Logging System. These measurements were output onto standard teletype copy as well as on punched paper tape. The system is shown in Figure 16. The Solartron provided 2.5 microvolt resolution for temperature probes and soil heat flux and 10 microvolt resolution for net radiation signals.

The punched paper tape was read directly into the CDC 6400 computer in the McMaster University Computer Center and copied onto magnetic tape for analysis. The analysis included an editing routine which eliminated sporadic measurement errors occasionally generated in the components of the recording and output system.

Error Analysis of the Bowen Ratio Solution to the Energy Balance

An error analysis was performed to determine the accuracy of
FIGURE 16 - DATA ACQUISITION AND RECORDING SYSTEM
estimating λE. The sensitivity of evaporation estimates to each of the variables in the Bowen ratio solution was determined (Figure 17). Evaporation is most sensitive to changes in net radiation since a 1% change in Rn results in a 1.11% change in λE. Evaporation was also fairly sensitive to changes in vapour pressure (a 1% change in λE results in a change of 0.47% in λE), and slightly less sensitive to changes in ΔT and Δe. Changes in soil heat flux result in only minor changes in evaporation.

The absolute error in evaporation was calculated following procedures discussed by Cook and Rabinowitz (1963), from

\[
\delta\lambda E = \frac{\partial \lambda E}{\partial Rn} \delta Rn + \frac{\partial \lambda E}{\partial G} \delta G + \frac{\partial \lambda E}{\partial \Delta T} \delta \Delta T + \frac{\partial \lambda E}{\partial \Delta e} \delta \Delta e ,
\]  

(33)

in which δλE, δRn, δG, δΔT and δΔe are the absolute errors in λE, Rn, G, ΔT and Δe respectively, and \(\frac{\partial \lambda E}{\partial Rn}\), \(\frac{\partial \lambda E}{\partial G}\), \(\frac{\partial \lambda E}{\partial \Delta T}\), \(\frac{\partial \lambda E}{\partial \Delta e}\) are the partial derivatives of λE with respect to Rn, G, ΔT, and Δe obtained from the energy balance equation.

The relative error in evaporation was calculated by dividing equation (33) by λE and computing the root mean square:

\[
\frac{\delta \lambda E}{\lambda E} = \pm \sqrt{\left(\frac{\delta \lambda E}{\partial Rn} \frac{\delta Rn}{\lambda E}\right)^2 + \left(\frac{\delta \lambda E}{\partial G} \frac{\delta G}{\lambda E}\right)^2 + \left(\frac{\delta \lambda E}{\partial \Delta T} \frac{\delta \Delta T}{\lambda E}\right)^2 + \left(\frac{\delta \lambda E}{\partial \Delta e} \frac{\delta \Delta e}{\lambda E}\right)^2} .
\]  

(34)

For the calculation, absolute errors of 5% and 10% were prescribed for net radiation and soil heat flux. Absolute and relative errors for ΔT and Δe were calculated over the ranges 0.1°C to 2.0°C and 0.1 mb to 1.0 mb (Table 1). Equations (33) and (34) were then evaluated to give
the absolute and relative errors in $\lambda E$ (Table 2).
**TABLE 1**

**ABSOLOUTE AND RELATIVE ERRORS IN ΔT AND ΔE**

<table>
<thead>
<tr>
<th>ΔT (°C)</th>
<th>Absolute Error</th>
<th>Relative Error (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.10</td>
<td>0.026</td>
<td>25.60</td>
</tr>
<tr>
<td>0.20</td>
<td>0.026</td>
<td>12.28</td>
</tr>
<tr>
<td>0.50</td>
<td>0.026</td>
<td>5.10</td>
</tr>
<tr>
<td>1.00</td>
<td>0.026</td>
<td>2.56</td>
</tr>
<tr>
<td>1.50</td>
<td>0.026</td>
<td>1.71</td>
</tr>
<tr>
<td>2.00</td>
<td>0.026</td>
<td>1.28</td>
</tr>
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</table>

<table>
<thead>
<tr>
<th>Δe (mb)</th>
<th>Absolute Error</th>
<th>Relative Error (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.10</td>
<td>0.031</td>
<td>30.80</td>
</tr>
<tr>
<td>0.25</td>
<td>0.031</td>
<td>12.34</td>
</tr>
<tr>
<td>0.50</td>
<td>0.031</td>
<td>6.17</td>
</tr>
<tr>
<td>0.75</td>
<td>0.031</td>
<td>4.12</td>
</tr>
<tr>
<td>0.90</td>
<td>0.031</td>
<td>3.42</td>
</tr>
<tr>
<td>1.00</td>
<td>0.031</td>
<td>3.08</td>
</tr>
<tr>
<td>$\lambda E \text{ (ly/min)}$</td>
<td>Absolute Error</td>
<td>Relative Error (%)</td>
</tr>
<tr>
<td>-----------------</td>
<td>----------------</td>
<td>--------------------</td>
</tr>
<tr>
<td>0.10</td>
<td>0.007</td>
<td>9.00</td>
</tr>
<tr>
<td>0.20</td>
<td>0.008</td>
<td>6.16</td>
</tr>
<tr>
<td>0.30</td>
<td>0.011</td>
<td>4.90</td>
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<tr>
<td>0.40</td>
<td>0.015</td>
<td>4.36</td>
</tr>
<tr>
<td>0.50</td>
<td>0.017</td>
<td>4.00</td>
</tr>
<tr>
<td>0.60</td>
<td>0.019</td>
<td>3.87</td>
</tr>
<tr>
<td>0.70</td>
<td>0.021</td>
<td>3.60</td>
</tr>
<tr>
<td>0.80</td>
<td>0.023</td>
<td>3.32</td>
</tr>
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CHAPTER IV

RESULTS

In this chapter surface control of the energy balance components, the three forms of the combination model and surface resistance are examined. Before comparing the results from the two sites under different soil moisture regimes, it must first be shown that the same results were obtained under similar soil moisture conditions on both sites.

Comparison of Results from the Two Sites Under Similar Soil Moisture Regimes

Surface volumetric soil moisture (S.M.) over the field season varied from 0.007 to the field capacity value of 0.225 (Figure 18). Periods of irrigation and rainfall are shown on the diagram. On June 5 and September 25, soil moisture on the two fields were similar and energy balance comparisons were made. Figure 19 shows the diurnal variation of net radiation, evaporation and soil heat flux for both fields on these two days. The total energy used in evaporation $\lambda E$ on June 5 was 2.50 kWh m$^{-2}$ day$^{-1}$ over field A and 2.52 kWh m$^{-2}$ day over field B. On an hourly basis, maximum differences in $\lambda E$ were 6.4% but, in general, they were within instrumental accuracy. Maximum differences in $Rn$ amounted to 5.7% but, in general, $Rn$ differences were within 5%, the accuracy of the net radiation measurements. Slight differences in soil heat flux existed on both days, but evaporation is insensitive to such small changes.
FIGURE 18 - DAILY SOIL MOISTURE VARIATIONS
FIGURE 19 - COMPARISON OF Rn, λE and G OVER THE TWO SITES
The Effect of Soil Moisture Differences

Sample Days

Fifteen of forty-five days were selected for study. These are indicated in Table 3 along with corresponding soil moisture values for each site and cloud cover descriptions. The days were selected because the wind directions were southerly and substantial differences in soil moisture between fields existed.

Energy Balance Response to Soil Moisture Variation

The components of the energy balance were examined under different soil moisture conditions on an hourly and a daily basis. These have been made dimensionless by dividing by net radiation, which allows comparison between data measured when net radiation was dissimilar.

Diurnal Relationships between Energy Balance Components and Soil Moisture

Two typical days (June 23 and June 30) were selected from the sample to demonstrate the effect of soil moisture on the energy balance components. Wind direction on June 23 was mainly southwest, but was northerly before 0745. The sensitivity of the experiment to fetch direction is apparent. Before 0745, evaporation on the dry field showed a marked fluctuation which is not reflected in the radiation curve (Figure 20A). After 0745, the evaporation curve follows the radiation curve more closely.

Two days had elapsed since A had been irrigated but large differences in evaporation rates still existed. Over A, \( \lambda E/Rn \) was
### TABLE 3

**SAMPLE DAYS**

<table>
<thead>
<tr>
<th>Date</th>
<th>Soil Moisture</th>
<th>Sky Cover</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Field A</td>
<td>Field B</td>
</tr>
<tr>
<td>June 12</td>
<td>0.155</td>
<td>0.034</td>
</tr>
<tr>
<td>17</td>
<td>0.130</td>
<td>0.076</td>
</tr>
<tr>
<td>18</td>
<td>0.100</td>
<td>0.060</td>
</tr>
<tr>
<td>23</td>
<td>0.125</td>
<td>0.024</td>
</tr>
<tr>
<td>30</td>
<td>0.206</td>
<td>0.014</td>
</tr>
<tr>
<td>July 15</td>
<td>0.121</td>
<td>0.007</td>
</tr>
<tr>
<td>16</td>
<td>0.086</td>
<td>0.010</td>
</tr>
<tr>
<td>21</td>
<td>*0.225</td>
<td>0.043</td>
</tr>
<tr>
<td>22</td>
<td>0.159</td>
<td>0.019</td>
</tr>
<tr>
<td>23</td>
<td>0.120</td>
<td>0.010</td>
</tr>
<tr>
<td>August 10</td>
<td>0.153</td>
<td>0.011</td>
</tr>
<tr>
<td>14</td>
<td>0.182</td>
<td>0.034</td>
</tr>
<tr>
<td>16</td>
<td>0.114</td>
<td>0.025</td>
</tr>
<tr>
<td>17</td>
<td>0.110</td>
<td>0.035</td>
</tr>
<tr>
<td>18</td>
<td>0.062</td>
<td>0.017</td>
</tr>
</tbody>
</table>

* - Irrigated at noon - before irrigation S.M. = 0.078
FIGURE 20a - ENERGY BALANCE COMPONENTS (JUNE 23)

FIGURE 20b - BOWEN RATIOS (JUNE 23)
0.90 but it was only 0.74 over B. Large differences in G/Rn also existed; soil heat flux into the drier surface (0.12) was substantially larger than into the wetter surface (0.07).

Mean Bowen ratios are shown in Figure 20b for June 23. Ranges of values for the three intervals (0.25-0.50 m, 0.50-0.75 m, and 0.25-0.75 m) are indicated. Until 0745, agreement between intervals was poor. With southerly winds, however, the agreement improved over most of the day. Toward the end of the day (after 1500) the scatter of values again increased though this was typical of all days and resulted from measurement problems when ΔT and Δe were small. The mean daily Bowen ratios differed markedly between the two sites. Over the wetter surface the ratio was only 0.06 while over the drier surface it was 0.21.

Field A was irrigated again on June 29. Because the soil was almost saturated (S.M. = 0.206) while field B was drier than on June 23, data collected on June 30 (Figure 21a) show even larger differences between the two fields. Mean daily values of λE/Rn were 0.91 on A and 0.71 on B. The value of G/Rn on the wet field (0.051) had decreased to less than half that of the dry field (0.117), although the magnitude of the dry field value had not changed substantially. Bowen ratios (Figure 21b) had increased over the dry field (0.22) and decreased over the wet field (0.04). The very low half-hourly values of 6 over the wet field indicated near saturation conditions.

Net radiation was virtually identical over the two fields on both days. However, the net radiative energy was distributed differently on each field between the components of evaporation, sensible heat and soil heat. Since water was abundant on A, most of the net radiation
FIGURE 21a - ENERGY BALANCE COMPONENTS (JUNE 30)

FIGURE 21b - Bowen Ratios (June 30)
was used in evaporation and less than 10% was shared between sensible
heat flux and soil heat flux. Since the water supply was smaller on B,
less energy was used in evaporation and more energy was distributed between
sensible heat and soil heat, which resulted in larger values of $\beta$ and
$G/R_n$.

The form of the evaporation trace for June 30 on the dry field
is of interest. $\lambda E$ is well correlated with $R_n$ until 1100. Between 1100
and 1300 $\lambda E$ remains virtually constant. Later, the correlation with $R_n$
is resumed. The corresponding wet field trace exhibits no such curtail-
ment of evaporation rate. This feature is attributed to partial stomatal
closure in response to limited water supply and high radiation loads. A
similar feature was reported by McCaughey (1968).

The soil heat flux trace over both fields on both days shows a
peak before the net radiation maximum. This is noticeable also in work
by Wilson and McCaughey (1971). Logically, the greatest flux of heat
into the surface should occur at the maximum rate of increase in $R_n$
since the largest temperature gradients in the soil occur at that time.
However, there is a time lag between this maximum and the maximum measured
soil heat flux which probably results from the slow conduction of heat
to the sensors. On June 30 the maximum rate of increase in $R_n$ occurs
between 0830 and 0930 and the maximum soil heat flux occurs at 1015.

Relationships between Energy Balance Components and Soil
Moisture Drying Cycles

Figure 22 and 23 illustrate the effect of one irrigation on the
evaporation rate and the Bowen ratio on July 21 and the subsequent drying
FIGURE 22 - EFFECT OF IRRIGATION ON EVAPORATION AND DRYING
Figure 23 - Effect of Irrigation and Drying on Bowen Ratio
on the 22nd and the 23rd. Data from July 21 show a dramatic change in the evaporation rate after the irrigation. On field A values of $\lambda E/R_n$ changed from 0.72 ($\beta = 0.27$) before the application to 1.16 ($\beta = -0.09$) afterward. Energy used in evaporation exceeded that supplied by net radiation due to the development of a temperature inversion and downward sensible heat flux.

Subsequent drying after the irrigation was evident in data from July 22 and July 23, when the evaporation rates over the two surfaces began to converge. For field A values of $\lambda E/R_n$ decreased to 0.929 ($\beta = 0.05$) on July 22 and 0.866 ($\beta = 0.10$) on July 23 and remained constant at 0.635 ($\beta = 0.35$) over field B. Fluctuations in $\beta$ were due to difficulties in measurement of the small temperature gradients over the still saturated surface under cloudy-bright conditions.

**Daily Soil Moisture Relationships**

The results of the diurnal comparisons suggest that daily relationships between energy balance components and soil moisture should exist. Values of $\lambda E/R_n$, $G/R_n$ and $\beta$ from fourteen days were related to soil moisture. July 21 was excluded due to the difficulty in obtaining representative daily values under the very different soil moisture regimes that existed before and after the midday irrigation.

The relationship between $\lambda E/R_n$ and soil moisture (Figure 24) shows that evaporation varies almost solely in response to net radiation at soil moisture values greater than about 0.1, but becomes increasingly dependent on surface soil moisture at smaller values. The same data are plotted on
Figure 24 - Comparison of $\frac{\lambda E}{Rn}$ under different soil moisture conditions.
semi-logarithmic paper in Figure 25. The strength of the relationship is indicated by the high correlation coefficient (0.922). Approximately 85% of the variance in $\lambda E/R_n$ is explained by variations in soil moisture. Some of the scatter is due to errors in soil moisture sampling. The rest is attributed to changes of surface resistance in response to small water supplies and large evaporative demands (illustrated in Figure 4.4a) and changes in surface resistance due to changes in crop cover. The relationship shows that when surface soil moisture approaches zero, $\lambda E/R_n$ still has a value of 0.60. Since evaporation from the bare soil would be negligible, this represents the transpiration rate from plants which tap deeper sources of moisture. The application of this relationship is attractive since it requires only measurements of soil moisture and net radiation to predict evaporation. Recent developments in remote sensing of soil moisture and calculation of net radiation from relatively simple meteorological data may allow such a relationship to be used in evaluating $\lambda E$ for catchments or larger areas.

The relationship between soil heat flux and soil moisture is not as strong (Figure 26). Some of the scatter is again due to measurement and sampling errors in soil heat flux and moisture. However, the effects of crop growth are also important. Over a well-covered surface, much of the solar energy may be intercepted by the crop leaves and less will actually reach the soil. The absolute magnitude of the soil heat flux will be less over a well-cropped surface than over a sparser one. This may be illustrated by comparing data derived under different crop covers on days when soil moisture is similar. On June 18, the crop was not
\[ \frac{\lambda E}{R_n} = 1.053 + 0.212 \log_{10} (\text{SM}) \]

\[ R = 0.922 \]

**Figure 25** - The relationship between \( \lambda E/R_n \) and soil moisture
$G/R_n = 0.116 - 0.388 (SM)$

$R = -0.826$

**Figure 26 - The relationship between $G/R_n$ and soil moisture**
fully established on the dry field and volumetric soil moisture was 0.060. On August 18, the volumetric soil moisture was 0.062 on the wet field and the crop was densely rooted and covered the entire surface. The ratios of G/Rn were 0.085 on June 18 and only 0.073 on August 18.

Soil heat flux is controlled by thermal conductivity as well as the temperature gradient in the soil. The increased thermal conductivity of a wetted soil tends to increase soil heat flux. However, temperature gradients in wetter soils are typically smaller than in drier soils because of evaporative cooling at the surface. This tends to decrease the soil heat flux. The inverse nature of the linear relationship in Figure 26 indicates that in this experiment soil heat flux is dominated by temperature gradients rather than thermal conductivity. Despite the scatter in values, 68% of the variance in G/Rn is explained by variations in soil moisture.

Mean daily values of the Bowen ratio were also examined under different soil moisture regimes (Figure 27). The correlation coefficient (-0.86) was not as high as was expected from the evaporation relationship (Figure 24). However, it seems physically correct since it predicts that the Bowen ratio is zero when soil moisture is 0.21, (virtually at field capacity). Thus at field capacity, the sensible heat flux is very small and evaporation may be calculated from net radiation reduced by soil heat flux. This supports work by Budyko (1958), Tanner (1960), House, Rider and Tugwell (1960) and Davies (1967) who state that \( \Delta \text{PE} \approx \text{Rn} - G \). Some of the scatter in the relationship is a result of attaching equal weight to the values of \( \beta \) at the extremities of days (Figure 23) or on other occasions when \( \Delta T \) and \( \Delta e \) are difficult to evaluate accurately.
Figure 27 - The relationship between $\beta$ and soil moisture

$\beta = 0.147 - 0.263 \log_{10}(SM)$

$R = 0.860$
Errors in evaporation calculated from these Bowen ratios were small since net radiation is typically small at such times. Despite this weakness, 74% of the variance in the Bowen ratio may be attributed to variation in soil moisture.

Daily comparisons of net radiation from the two sites indicated that variations were not related to soil moisture fluctuations. This was confirmed by regression analysis of half-hourly net radiation values from fifteen days from both fields (Figure 28). The linear relationship (correlation coefficient = 0.995) was very close to the 1:1 line with a slope of 0.993 and an intercept of 0.003. This relationship supports the statement of Monteith and Szeicz (1961) that "... the radiation balance over grass is independent of height or management."

Equilibrium and Potential Evaporation

The performance of the potential and equilibrium forms of the combination model were compared with actual evaporation under different soil moisture regimes.

Equilibrium Evaporation

Denmead and McIlroy (1970), Wilson (1971), and Wilson and Rouse (1972) have shown good agreement between equilibrium and actual evaporation over dry surfaces. Half-hourly values of equilibrium evaporation agreed with actual evaporation for July 15, 16, 22 and 23 when volumetric soil moisture did not exceed 0.02 (Figure 29). The dry field data shows almost perfect agreement. Under wet field conditions, however, equilibrium evaporation accounts for only 85% of actual evaporation and departs
FIGURE 28 - COMPARISON OF Rn OVER THE TWO SITES
Figure 29 - Performance of the Equilibrium Model
from the 1:1 line as evaporation increases. These results are in general agreement with work by Wilson (1971) and Davies (in press).

This relationship obscures the gradual change of $\lambda$PES toward $\lambda$E. Figure 30 shows diurnal traces of equilibrium and actual evaporation on four days when different soil moisture regimes existed. On June 18, $\lambda$PES/$\lambda$E was 0.733. As soil moisture decreased, $\lambda$PES/$\lambda$E approached unity. On July 16, $\lambda$PES/$\lambda$E was 0.836; on August 10 it increased to 0.897; and on July 15 it was 1.010.

This relationship was examined in more detail by comparing ratios of $\lambda$PES/$\lambda$E to soil moisture for all fifteen days. Figure 31 shows that $\lambda$PES/$\lambda$E increased as soil moisture decreased, and approached unity when soil moisture was approximately 0.01. In addition, equilibrium evaporation agreed with actual evaporation to within 10% when soil moisture was less than 0.07 (approximately 30% of field capacity).

**Potential Evaporation**

Values of aerodynamic resistance required for calculation of $\lambda$PE were obtained from wind profile measurements. Zero-plane displacement ($d$) and surface roughness parameters ($z_o$) were determined by Lettau's iterative procedure as described by Tanner (1963). The analysis is sensitive to small errors in wind speed measurements. Large errors in $z_o$ may occur. Figure 32 shows that a 3% overestimate in wind speed at one level results in a 50% overestimate in $z_o$, while a 3% underestimate results in a 20% underestimate in $z_o$. Wind speed errors also produce marked errors in the zero-plane displacement. Tanner (1963) demonstrates similar errors in the calculation of $z_o$ and $d$. He also recommends that
FIGURE 30 - DIURNAL COMPARISON OF $\lambda E$ AND $\lambda$PEs UNDER DIFFERENT SOIL MOISTURE REGIMES
$\text{LOG}_{10} \left( \frac{\text{APEs}}{\lambda E} \right) = -0.186 - 0.0914 \text{LOG}_{10} \text{SM}$

$R = -0.85$

**Figure 31** - The relationship between $\frac{\text{APEs}}{\lambda E}$ and soil moisture.
CORRECT WINDSPEEDS

3% OVERESTIMATE IN BOTTOM LEVEL

3% UNDERESTIMATE IN BOTTOM LEVEL

FIGURE 32 - SENSITIVITY OF $Z_o$ TO ERRORS IN WINDSPEED MEASUREMENT
the analysis only be used when measurements at more than three levels are available. Since this was not always possible in this study, the values of \( z_0 \) could not always be accepted confidently. To obtain reliable values on all days, a relationship was sought between crop height and \( z_0 \), using data from wind measurements at four levels when available but also from three levels when optimum fetches and neutral stability were achieved. This type of relationship has been shown by Lettau (1969), Tanner and Pelton (1960) and Szeicz, Endrodi and Tajchman (1969) to be acceptable.

Eighteen values of \( z_0 \) were selected from the data. These were regressed against crop height (Figure 33). The strength of the relationship (correlation coefficient = 0.922) means that it is certainly good enough to be used for the calculation of \( z_0 \) from crop height. Values of \( z_0 \) from the relationship were then applied to days when the wind profiles did not provide acceptable values.

The same procedure was applied to zero-plane displacement (d). Figure 34 shows that the regression of d against h did not provide as strong a relationship as \( z_0 \) against h. However, the results were acceptable for this study since computed values of aerodynamic resistance are not sensitive to small errors in zero-plane displacement (Tanner 1963). The procedure of employing crop height measurements to determine \( z_0 \) and d is attractive since it allows \( r_a \) to be estimated from measurements of crop height and wind speed at a single level.

Careful estimation of \( z_0 \) may yield very reliable estimates of aerodynamic resistance. In the data used in Figure 32, a 50% error in \( z_0 \) produced only a 17% error in \( r_a \). As Tanner and Fuchs (1968) point out,
Figure 33 - The relationship between $Z_0$ and crop height ($h$)

$Z_0 = 0.125h - 0.002$

$R = 0.922$
Figure 34 - The relationship between $d$ and crop height ($h$).

$$d = 0.661 \times h - 0.017$$

$R = 0.731$
high accuracy in $r_a$ is not essential in the correct evaluation of the combination model in humid and subhumid conditions since the radiative term $S \frac{S}{S_{HFV}} (Rn-G)$ is usually larger than the aerodynamic term $\rho C_p \Delta D/ra$. This meant that values of aerodynamic resistance could be confidently accepted.

Potential evaporation results indicated that the method of estimating potential evaporation by assuming that surface resistance in the combination model is zero is in gross error for land surfaces. Comparison of dry and wet field data (Figure 35) showed that for June 18, total evaporation over the wet field was 3.77 kWh m$^{-2}$ day$^{-1}$, which exceeded the value of 3.42 kWh m$^{-2}$ day$^{-1}$ calculated for the dry field. This indicated that since more soil moisture was available, the potential evaporation rate on the wet field should equal or exceed that of the dry field. However, the mean potential rate was only 5.95 kWh m$^{-2}$ day$^{-1}$ on the wet field and 6.52 kWh m$^{-2}$ day$^{-1}$ on the dry field. The difference became very much larger as soil moisture differences increased. On July 22 the mean potential rate on field A was only 5.58 kWh m$^{-2}$ day$^{-1}$ but was 6.90 kWh m$^{-2}$ day$^{-1}$ on field B. Assuming that the wet field potential evaporation rate was correct, dry field results overestimated by 10% on June 18 and by 24% on July 22. It will be shown later than, in fact, the potential rate was even overestimated on the wet field which resulted in larger overestimates of the correct potential rate.

The aerodynamic term may only exceed the radiative term when dry air is advected across a wet surface. In this experiment, these conditions could not exist since its operation was limited to times when
Figure 35 - Diurnal comparison of $\lambda E$ and $\lambda Pe$ over the two sites
winds blew from the south across the irrigated field before encountering the drier field. Figure 36 shows the relative sizes of these two terms on June 30 for the two surfaces. The radiative terms differ by approximately 15% with the wet field values exceeding those of the dry field. The aerodynamic terms, however, differ by approximately 100% with the dry field values exceeding those of the wet field. Since aerodynamic resistances differed only by the small stability correction for each field (Figure 37), the differences in the aerodynamic term were due mainly to the large differences in the wet-bulb depressions. The potential model assumes that the wet-bulb depression is independent of the state of saturation of the surface. This is obviously incorrect.

The dependence of the radiative term (the equilibrium model) and $D_2$ on soil moisture means that the potential rate of evaporation is dependent on soil moisture. This relationship (Figure 38) is inverse. Since the maximum rate of evaporation from a surface occurs at field capacity, it was expected that $\lambda PE/\lambda E$ would have a value of unity when complete saturation existed. The relationship predicted a value of 1.4 at field capacity. However, it predicted a value of unity when volumetric soil moisture is 1.0, which is in fact a water surface. This shows that the potential model does not predict the potential rate of evaporation over a land surface but the rate of evaporation over a water surface which is under the same radiative and advective conditions. This is a result of the assumption that surface resistance is zero when the land surface is saturated. It will be demonstrated later that surface resistance was not zero at field capacity.
FIGURE 36 - COMPARISON OF THE RADIATIVE AND AERODYNAMIC TERMS OVER THE TWO SITES
FIGURE 37 - AERODYNAMIC RESISTANCES OVER THE TWO SITES
The relationship between $\frac{\lambda PE}{\lambda E}$ and soil moisture is shown in Figure 38. The equations for the relationships are:

$$\log_{10} \frac{\lambda E(A)}{\lambda E(B)} = 0.070 - 0.091 \log_{10} SM$$

with $R = -0.967$

and

$$\log_{10} \frac{\lambda PE}{\lambda E} = 0.006 - 0.243 \log_{10} SM$$

with $R = -0.881$.

The graph also includes data points and a field capacity indication.
It has been shown (Figure 24) that in this experiment evaporation varied almost solely with net radiation when volumetric soil moisture was greater than 0.10. Under these conditions, \( \lambda E/Rn \) was greater than 0.85. Several workers have stated that potential conditions prevail when \( \lambda E/Rn \) exceeds that value. They included Pruitt and Angus (1961), Tanner and Lemon (1962) and Fritschen and Van Bavel (1963). On eleven days, wet field volumetric soil moisture exceeded 0.10. Using evaporation measured over field A on these days as a measure of potential evaporation, \( \lambda PE/\lambda E \) was again examined in relation to soil moisture (top line Figure 38). The relationship (correlation coefficient = 0.967) predicted a value of \( \lambda PE/\lambda E \) close to unity (0.98) at field capacity. This indicates that potential evaporation is more correctly defined as the evaporation over a thoroughly wetted surface.

Hence the validity of potential evaporation calculated over land surfaces using the combination model is in question.

Surface Resistance and Soil Moisture

In this experiment, surface resistance was evaluated from equation 29. \( \lambda E/Rn \) values were plotted against \( r_s \) (Figure 39). Results of a similar plot by Monteith (1965) are also included. There is good agreement between the two sets of data. Surface resistance was compared to soil moisture (Figure 40); a strong relationship was evident. This relationship shows that at field capacity, there was still a surface resistance of 0.23. This confirms the hypothesis that the combination model for potential evaporation does not predict the potential evapora-
FIGURE 39 - THE RELATIONSHIP BETWEEN $\frac{\lambda E}{R_n}$ AND $r_s$
\[
\log_{10} r_s = -1.086 - 0.656 \log_{10} SM
\]

\[R = -0.973\]

FIGURE 40 - THE RELATIONSHIP BETWEEN \( r_s \) AND SOIL MOISTURE
tion rate over land surfaces since some resistance is always present. The relationship in Figure 40 has considerable application since the surface resistance may now be predicted from soil moisture measurements, and used in the combination model to calculate actual evaporation.

The use of this relationship was tested for eleven days of data. Surface resistances were obtained from Figure 40. These values were used in the combination model for actual evaporation and the predicted values were compared with measured values (Figure 41). The plotted points lie close to the 1:1 line; most predicted values are within 10% of the measured values.

Figure 42 shows that percentage changes in surface resistance result in much smaller percentage changes in evaporation. This means that small errors in the evaluation of surface resistance (say 10%) allow evaporation to be calculated within 5%. This would explain the good agreement between predicted and measured values of evaporation, since errors in surface resistance resulting from errors in soil moisture sampling would produce minimal errors in evaporation.
FIGURE 41 - COMPARISON OF MEASURED $\lambda E$ AND $\lambda E$ PREDICTED FROM $r_s$ VALUES TAKEN FROM FIGURE 40.
FIGURE 42 - SENSITIVITY OF $\lambda E$ TO CHANGES IN $r_s$
CHAPTER V

CONCLUSIONS

The study has shown that soil moisture exerts considerable control over the components of the energy balance. Variation in all components of the energy balance are dependent on soil moisture variations except net radiation. The relationship between evaporation and soil moisture is very strong and has considerable application to both irrigation scheduling and hydrologic studies. The Bowen ratio-soil moisture relationship showed that $\beta$ approaches zero when soil moisture approaches field capacity. This lends support to the hypothesis that $\lambda\text{PE} \approx (Rn - G)$. It has also been shown that the plant has a limiting effect on the hourly rate of evaporation through partial stomatal closure when radiation loads and evaporative demands are high and water supplies are low.

Examination of the combination model showed that the equilibrium rate of evaporation is directly related to soil moisture as is actual evaporation but that the potential rate of evaporation is inversely related to soil moisture. Equilibrium evaporation agrees well with actual evaporation when soil moisture is less than 30% of field capacity but underestimates by increasing degrees as soil moisture increases. The potential evaporation results showed that potential evaporation calculated from the combination model does not predict the evaporation from a
thoroughly wet surface but the evaporation over a water surface which is under the same radiation, temperature structure and wind conditions. This is due to the assumption inherent in the derivation of the combination model for potential evaporation that the surface resistance is zero. The relationship between surface resistance and soil moisture confirmed that surface resistance does not become negligible over a grass surface until volumetric soil moisture reaches 100% (a water surface). This combined with the relationship between $\beta$ and soil moisture indicate that potential evaporation is better defined as $\lambda E = Rn - G$. Alternatively, it may be defined as the evaporation over a surface when the value of $\lambda E/Rn$ becomes insensitive to changes in soil moisture. In this experiment, potential evaporation would then be defined as $\lambda E/Rn \geq 0.85$.

The surface resistance-soil moisture relationship also provides an attractive alternative method for estimating surface resistance. The study confirms previous findings (e.g. Tanner and Pelton (1960)) that $z_o$ can be calculated from a knowledge of crop height. Hence, aerodynamic resistance may be estimated from measurements of crop height and wind speed at one height. With these two resistances and measurements of net radiation and soil heat flux as well as temperature and humidity at only one level, the combination model could be operationalized. However, the relationship between soil moisture and surface resistance must be more clearly established for several crop surfaces before this procedure may be accepted.
# APPENDIX A

## LIST OF SYMBOLS

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Explanation</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>C</td>
<td>Heat capacity of soil</td>
<td>joule kg(^{-1}) °K(^{-1})</td>
</tr>
<tr>
<td>D</td>
<td>Wet-bulb depression</td>
<td>°K</td>
</tr>
<tr>
<td>λE</td>
<td>Evaporation</td>
<td>Wm(^{-2})</td>
</tr>
<tr>
<td>G</td>
<td>Soil heat flux</td>
<td>Wm(^{-2})</td>
</tr>
<tr>
<td>H</td>
<td>Sensible heat flux</td>
<td>Wm(^{-2})</td>
</tr>
<tr>
<td>i</td>
<td>Subscript designating the stomatal wall</td>
<td></td>
</tr>
<tr>
<td>K(_H)</td>
<td>Eddy diffusivity of sensible heat</td>
<td>m(^2) sec(^{-1})</td>
</tr>
<tr>
<td>K(_W)</td>
<td>Eddy diffusivity of latent heat</td>
<td>m(^2) sec(^{-1})</td>
</tr>
<tr>
<td>K(_m)</td>
<td>Eddy diffusivity of momentum</td>
<td>m(^2) sec(^{-1})</td>
</tr>
<tr>
<td>P</td>
<td>Energy used in Photosynthesis</td>
<td>Wm(^{-2})</td>
</tr>
<tr>
<td>λPE</td>
<td>Potential Evaporation</td>
<td>Wm(^{-2})</td>
</tr>
<tr>
<td>λPE(_S)</td>
<td>Equilibrium Evaporation</td>
<td>Wm(^{-2})</td>
</tr>
<tr>
<td>Ri</td>
<td>Richardson number</td>
<td>dimensionless</td>
</tr>
<tr>
<td>Rn</td>
<td>Net radiation</td>
<td>Wm(^{-2})</td>
</tr>
<tr>
<td>S</td>
<td>Slope of the saturation vapour pressure-temperature curve</td>
<td>bar °K(^{-1})</td>
</tr>
<tr>
<td>T</td>
<td>Dry-bulb temperature</td>
<td>°K</td>
</tr>
<tr>
<td>T(_w)</td>
<td>Wet-bulb temperature</td>
<td>°K</td>
</tr>
<tr>
<td>t</td>
<td>Time</td>
<td>sec</td>
</tr>
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<td>Symbol</td>
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<td>Units</td>
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<tr>
<td>--------</td>
<td>-------------</td>
<td>-------</td>
</tr>
<tr>
<td>Cp</td>
<td>Specific heat of air at constant pressure</td>
<td>joule kg(^{-1}) °K(^{-1})</td>
</tr>
<tr>
<td>d</td>
<td>Zero-plane displacement</td>
<td>m</td>
</tr>
<tr>
<td>e</td>
<td>Vapour pressure of the air</td>
<td>bar</td>
</tr>
<tr>
<td>e(_s)</td>
<td>Saturation vapour pressure</td>
<td>bar</td>
</tr>
<tr>
<td>k</td>
<td>Von Karman's constant</td>
<td>dimensionless</td>
</tr>
<tr>
<td>o</td>
<td>Subscript designating the surface</td>
<td>m</td>
</tr>
<tr>
<td>ra</td>
<td>Aerodynamic resistance</td>
<td>sec m(^{-1})</td>
</tr>
<tr>
<td>r(_s)</td>
<td>Surface resistance</td>
<td>sec m(^{-1})</td>
</tr>
<tr>
<td>rp</td>
<td>Total plant resistance</td>
<td>sec m(^{-1})</td>
</tr>
<tr>
<td>rd</td>
<td>Dry surface resistance</td>
<td>sec m(^{-1})</td>
</tr>
<tr>
<td>rl</td>
<td>Stomatal resistance</td>
<td>m sec(^{-1})</td>
</tr>
<tr>
<td>(\nu)</td>
<td>Kinematic viscosity of air</td>
<td>m sec(^{-1})</td>
</tr>
<tr>
<td>u</td>
<td>Horizontal wind speed</td>
<td>m sec(^{-1})</td>
</tr>
<tr>
<td>x</td>
<td>Volume</td>
<td>m(^3)</td>
</tr>
<tr>
<td>z</td>
<td>Height in the atmosphere. Also used as a subscript to designate a reference height</td>
<td>m</td>
</tr>
<tr>
<td>z(_o)</td>
<td>Surface roughness</td>
<td>m</td>
</tr>
<tr>
<td>(\beta)</td>
<td>Bowen ratio</td>
<td>dimensionless</td>
</tr>
<tr>
<td>(\gamma)</td>
<td>Psychrometric constant</td>
<td>bar °K(^{-1})</td>
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<tr>
<td>(\lambda)</td>
<td>Latent heat of vapourisation</td>
<td>joule Kg(^{-1})</td>
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<tr>
<td>(\xi)</td>
<td>Density of air</td>
<td>kg m(^{-3})</td>
</tr>
<tr>
<td>(\phi)</td>
<td>KEYPS wind profile corrections for stability</td>
<td>dimensionless</td>
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REFERENCES


McCaughey, J.H., 1968: A test of the Penman Combination Model for potential evaporation. Publication in Climatology, 1, 91 pp, McMaster University, Hamilton.


