THE EFFECT OF INTERCEPTED RAINFALL ON EVAPOTRANSPIRATION

THE EFFECT OF INTERCEPTED RAINFALL ON EVAPOTRANSPIRATION RATES OVER A MIXED HARDWOOD FOREST IN SOUTHERN QUEBEC

by

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ABSTRACT

This thesis investigates the difference in latent heat transfer for a wetted as opposed to an unwetted forest canopy. Previous work has shown that the rate of evaporation (LE_w) of intercepted rainfall from vegetation is several times that of the transpiration (LE_d) of soil moisture, were the canopy dry, assuming the same weather conditions. The present investigation uses the results of two seasons of field observations conducted in a mixed hardwood forest. Field measurements were designed so as to measure or estimate all the components of the hydrologic cycle.

The results substantiate the view that LE_w is several times greater than LE_d. Furthermore, the magnitude of LE_w/LE_d or $LE_w - LE_d$ is in turn dependent upon soil and plant factors and weather conditions. Because LE_w is several times greater than LE_d, rainfall interception by vegetation constitutes a significant loss of water to the soil and hence affects the computation of the water balance.

L'EFFET DE L'INTERCEPTION DE LA PLUIE AU NIVEAU DU FEUILLAGE SUR LE TAUX D'EVAPOTRANSPIRATION DANS UNE FORÊT DECIDUE MIXE DU SUD DU QUEBEC

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RÉSUMÉ

Cette thèse examine les differences de transferts de chaleur latente entre un feuillage forestier sec et un feuillage forestier mouillé. Des études précédentes ont démontré que, pour des conditions atmosphériques semblables, le taux d'évaporation (LE_w) à l'intérieur d'un feuillage mouillé est, par suite de l'interception de la pluie à ce niveau, plusieurs fois supérieur au taux de transpiration (LE_d) de la vapeur d'eau par ce même feuillage à l'état sec. Cette étude est le résultat d'observations effectuées, dans une forêt décidue mixte, sur une période de deux années. Tous les termes de bilan hydrologique y ont été soit mesurés, soit estimés.

Les résultats appuient le point de vue stipulant la supériorité de LE_w sur LE_d. De plus, le rapport de grandeur LE_w/LE_d ou $LE_w - LE_d$ est directement relié à des variables concernant le sol, la végétation et les conditions atmosphériques. Du fait de la supériorité de LE_w sur LE_d résultant de l'interception de la pluie au niveau du feuillage, ce dernier fait doit être tenu compte de dans les calculs de bilans hydrologiques.

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TABLE OF CONTENTS

CHAPTER 5 - PREDICTIVE MODEL FOR CANOPY RESISTANCE

CHAPTER 6 - EFFECT OF CANOPY WETNESS ON EVAPOTRANSPIRATION RATES

CHAPTER 7 - THE EFFECT OF WEATHER CONDITIONS

CHAPTER 8 - CONCLUSION

LIST OF FIGURES

LIST OF TABLES

Table Number Page

CHAPTER I

Introduction

1.1 The Problem

At the present time there is a good deal of controversy regarding the effect of intercepted rainfall on the water balance of forested watersheds. The core of the argument is whether rain retained by vegetation represents a total loss, a partial loss, or no loss of moisture beyond the normal evapotranspiration of the canopy. The disagreement stems largely from the way in which intercepted rainfall is viewed. If interceptional loss is viewed as a reduction in rainfall reaching the ground because of water retention by the aerial parts of vegetation, then it can be considered as a total loss. But if the entire water cycle of the soil-vegetation complex is considered, then intercepted rainfall may not constitute a total loss, in that when the foliage is wetted the transpirational withdrawal of soil moisture is subdued.

The magnitude of transpirational saving during the evaporation of intercepted rainfall is critical in determining how much of a moisture loss interception constitutes. If the rate of water vapor loss is the same whether the vegetation is wet or dry, then it matters very little whether the evaporative demand of the atmosphere is satisfied by intercepted water or soil moisture. If on the other hand, the evaporation of intercepted rainfall (LE_w) proceeds at a faster rate than the transpiration of soil moisture (LE_d) for an unwetted canopy, then, assuming the same weather conditions, there would be some amount of transpirational saving. The magnitude of this saving depends on the ratio LE_w/LE_d or the difference $LE_w - LE_d$.

Both the ratio LE_w/LE_d and the difference $LE_w - LE_d$ depend mainly on soil moisture availability, plant physiology and weather conditions. Soil and plant factors control the degree of stomatal opening through which water vapor must diffuse into the free atmosphere, which in turn regulates the amount of water which becomes available for transpiration. Weather conditions are important in that they dictate the amount of evaporation from the canopy during rainfall and the rate at which the wetted canopy dries out following rainfall. Light intensity rainfalls are usually more conducive to the evaporation of intercepted water than heavy intensity showers. Also, the weather after the canopy has been wetted controls the rate of evaporation of the water retained by the canopy. After a shower followed by clear, windy conditions for instance, the canopy can be dried out rapidly, thereby restricting any significant transpirational saving.

1.2 Approach to Problem

To fully comprehend the effect of intercepted rainfall on the water balance of a forested watershed, the movement of moisture in all phases and for all components of the hydrologic cycle must be examined. In the present research two full growing seasons (1974 and 1975) of field experimentation were carried out at Mont St. Hilaire, Quebec, to examine the role of intercepted rainfall on evapotranspiration rates over a beech-maple forest.

Rain in the open, together with throughfall and stemflow were measured for each fall of rain so as to obtain a measure of the amount of intercepted rainfall. Also evapotranspiration estimates were made for both dry and wet canopy conditions in order to gauge the magnitude of the ratio LE_w/LE_d and the difference $LE_w - LE_d$. Over the course of both growing seasons soil moisture content was also measured. Soil moisture depth not only provided a measure of a component of the water balance, but also served as the basis for the development of a predictive model for mean stomatal resistance to vapor diffusion. To provide an overall check on the accuracy of the other calculations of the water balance, runoff was measured in a small experimental basin near the main site. These various measurements were then analyzed so as to highlight the effect of intercepted rainfall on evapotranspiration and soil moisture withdrawal.

CHAPTER 2

Background Objectives and Methodology

2.1 Views on the Role of Intercepted Rainfall

The literature regarding the effect of intercepted rainfall on the hydrologic cycle contains a number of different, often conflicting, arguments. These conflicting viewpoints have arisen from either a change in thinking as greater understanding of the subject developed, or from research being conducted over varying vegetation types in different environments. There seem to be three basic points of view. Firstly, it is argued by some people that intercepted rainfall constitutes a direct loss of moisture to the canopy-soil complex unless some of it is absorbed by leaves before it is evaporated. Horton (1919) for example, maintains that interception represents a loss of precipitation which would otherwise be available to the soil. This viewpoint is shared by several other researchers (Kittredge, 1948; Law, 1957; Delfs, 1967; Helvey and Patric, 1965 and Patric, 1966). The basis for this first line of argument is that rainfall interception is viewed only in terms of the input segment of the hydrologic cycle. As a result, the direction of earlier studies has mainly been to examine variations in interceptional loss in terms of vegetation type and season of the year.

More recently, however, it has become apparent that the interception of rainfall by vegetation also affects the consumption of water by regulating the water available for runoff and also the storage or soil moisture component of the hydrologic cycle. Jones (1957) for example, discovered that during the evaporation of intercepted rainfall transpirational water loss by the plant was subdued, hence resulting in a saving on soil moisture. Similar results were obtained by Rutter (1959) and Thorud (1967). Out of this awareness has stemmed further points of view.

Some investigators argue that evaporation of intercepted water fully compensates for transpiration that would otherwise have occurred if the canopy were dry and is therefore not a loss beyond the normal evapotranspiration. Burgy and Pomeroy (1958) found that in vigorously growing laboratory grass plots the evaporation of a given amount of intercepted moisture was accompanied by an equal reduction in the amount of transpiration from the plants. In other words, total moisture loss was approximately the same in plots with wet and dry leaf surfaces. Field studies conducted by McMillan and Burgy (1960) gave similar results. Thus it makes no difference to water loss whether the evaporative demand of the atmosphere is satisfied by soil moisture or water withheld by the leaves. Leyton and Carlyle (1959) maintained that since a given supply of thermal energy will only evaporate a certain quantity of water, then the evaporation of water retained by the foliage must be compensated by a like reduction in transpiration.

Obviously the basis for this form of argument is that the amount of energy available for evapotranspiration for both a dry and a wetted canopy under a given set of atmospheric conditions is a constant fraction of the net available energy. In other words, surface conditions are always assumed to be potential, which is very likely for a well-watered grass cover.

There are however, other twists to this viewpoint. Goodell (1963) argued that if the transpiration rate of the entire plant is reduced when leaves are wetted by rain, the essential effect might be a period of replenishment of water within the plant tissues followed, after drying of the leaves, by a higher rate of transpiration. He maintains that this sequence may be especially likely if rain occurs during the night or early in the morning. The morning rate of transpiration may be reduced while the leaves dry, but this may simply shorten or eliminate a midday period of reduced transpiration with little or no effect on transpiration loss for the day as a whole. He even extends this form of reasoning to a seasonal pattern claiming that rainfall interception during the growing season may simply prolong the period of high rates of transpiration.

There also exists a third school of thought. Experimental results from a variety of researchers show that the rate of evaporation of intercepted rainfall from a wetted canopy is greater than the rate of transpiration from an unwetted canopy, where potential conditions are not satisfied, under similar weather conditions. Rutter (1959) found that the weight of cut, wet branches decreased much faster than the weight of transpiring branches in a laboratory experiment. From this he concludes that the rate of evaporation of intercepted water exceeds the transpiration rate by several times. Subsequent investigations (Rutter, 1967) showed that the rate of evaporation of intercepted water is, on the average, about four times as great as the transpiration rate in the same environmental conditions. Similarly Frankenberger (1960) using turbulent transfer methods found that the total evaporation from tree stands immediately after a rain was greater than that from the same stands after the foliage had dried.

Penman (1963) too remarked that the evaporation of intercepted rainfall from a wetted canopy is effectively the same as from an open water surface. However, a forest surface is rougher than a normal water surface. Consequently the rate of evaporation from a forest will be greater, during the day, than that of the transpiration of soil water, and can even go on at night. Leyton et al (1967) supported this conclusion by observing that the rate of evaporation of intercepted water from shrubs and trees occurs at a greater rate than that of transpiration under similar environmental conditions.

Thorud (1967) working with small potted ponderosa pine trees found that water applied to foliage reduces the transpiration rate by an average of 14 percent, or 9 percent of the water applied and thereby conserved soil moisture. However the net effect is small when compared with the amount of water applied. Shindel (1963) and Harr (1966), also working with small potted trees, found similar results. Also Waggoner et al (1969) found that the rate of evaporation from a wetted corn crop is at least twice the transpiration rate when the same crop is dry. Stewart and Thom (1973) further remarked that assuming identical weather conditions, intercepted rainfall can be expected to evaporate at about five times the corresponding transpiration rate. Finally McNaughton and Black (1973) found that evaporation from a wetted young Douglas fir forest in coastal British Columbia proceeds at a rate 20 percent faster than the expected transpiration from the same canopy when it is well supplied with water, but dry. Evidence of this kind has led Rutter (1968) to remark that a high proportion of the water intercepted by a forest is evaporated without any saving of stored water in the soil.

In the wake of these developments came attempts to explain, in physical terms, how it is that the rate of evaporation of intercepted rainfall is greater than the rate of transpiration under the same weather conditions.

Monteith (1965) envisages this difference as being attributable to changes in surface moisture conditions. He proposed that the latent heat flux over a plant cover could be examined in terms of potentials and resistances, where the potentials of the system are the net available energy and the evaporative demand of the atmosphere, and the resistances are the aerodynamic resistance to vapor transfer and the stomatal resistance to vapor diffusion. When the canopy is dry there exists a finite stomatal or plant resistance, depending mainly on light and soil moisture conditions, but when the canopy is wetted, the plant resistance reduces to zero, since the moisture demand of the atmosphere can now be satisfied by the readily available film of water on the leaf surfaces. In other words, he saw the difference as a case of potential (wet) as opposed to non-potential (dry) evapotranspiration conditions.

Rutter (1968) sees the fact that the evaporation of intercepted rainfall proceeds at a faster rate than the transpiration of soil moisture as being due to increased available energy for a wetted as opposed to a dry canopy. He maintains that there is a sizeable transfer of sensible heat to wetted vegetation at the expense of the surrounding region especially if the vegetation occupies a small area. More recently Murphy (1970) and Murphy and Knoerr (1975) have suggested that the increased latent heat exchange for a wetted canopy occurs at the expense of long-wave terrestrial radiation and sensible heat transfer.

It would seem that these latter viewpoints in fact concur with Monteith's earlier observation in that added energy through advection or through the reduction of sensible heat transfer and the suppression of terrestrial long wave radiation, would be evident in the form of an increased vapor pressure deficit of the ambient air.

2.2 Aims of Research

The foregoing discussion indicates that there is general agreement with the view that the evaporation of water from a wetted canopy proceeds at a rate that is greater than that from an unwetted canopy. This belief may be attributable either to a greater abundance of energy for latent heat transfer from a wetted canopy or to the greater availability of surface water, which thereby creates potential evapotranspiration conditions, or to both.

The intent of the present research is mainly to provide some experimental evidence for these arguments so as to redefine and elaborate upon some of the earlier conclusions. Firstly, following the lead of Monteith (1965) and Rutter (1968), an attempt will be made to substantiate the view that the rate of evaporation of intercepted rainfall (LE_w) is several times that of transpiration (LE_d), assuming the same weather conditions. A mixed hardwood forest in Southern Quebec is to be used as the experimental site. Furthermore, it will be shown that the magnitude of the ratio LE_w/LE_d is not constant, as implied in earlier investigations

(Rutter, 1967; McNaughton and Black, 1973 and Stewart and Thom, 1973), but rather undergoes both diurnal and seasonal changes.

Diurnal changes are mainly related to weather and soil moisture conditions. It can be shown that the ratio LE_w/LE_d is mainly a function of the ratio of the canopy resistance (r_c) to the aerodynamic resistance (r_a) (see section 6.3). Also the canopy resistance (r_c) has a diurnal regime which is characterized by stomatal behavior (see section 5.2) and has a large amplitude, whereas the aerodynamic resistance that is mainly a function of wind speed, is by contrast conservative. The ratio LE_w/LE_d is therefore to a large extent controlled by the parameter (r_c), which in turn is regulated by radiant energy and soil moisture depth (see chapter 5).

It can also be demonstrated that diurnal changes in the ratio LE_w/LE_d are related to weather conditions, namely the distribution, duration and intensity of rainfall and post-wetting synoptic conditions (see chapter 7). In other words, it can be demonstrated that the magnitude of the evaporation of intercepted rainfall (LE_w) during a light prolonged shower is greater than during a short intense rain-storm (see section 7.1). Furthermore, intermittent rainfalls separated by periods when the canopy is allowed to dry out partially or completely through evaporation can be conducive to greater interceptional losses. Also rainfall periods followed by sunny or windy conditions or both would allow intercepted water to be evaporated at a faster rate than under calm or cloudy conditions or both, since the former conditions would enhance energy receipt and/or turbulent transfer.

Seasonal changes of the ratio LE_w/LE_d on the other hand are mainly related to soil moisture and vegetal conditions. As will be demonstrated later (see section 5.4), the canopy resistance (r_c) is controlled to a large extent by soil moisture depth. When soil moisture availability at the root of the plants becomes limiting canopy resistance (r_c) generally increases, thereby increasing the ratio LE_w/LE_d. The natural life cycle of deciduous hardwoods also creates seasonal differences in the parameter (r_c). Usually, stomatal resistance becomes higher because of senescence of the leaves at the end of the growing season. This effect creates higher r_c and hence LE_w/LE_d values.

The magnitude of the ratio LE_w/LE_d is however not the critical factor in evaluating the amount of water loss for a wetted as opposed to a dry canopy. As will be shown subsequently (see section 6.3), the value of the difference $LE_w - LE_d$ is a better gauge for comparing the evaporation of intercepted rainfall to transpiration, under the same weather conditions. It can be shown that the difference $LE_w - LE_d$ is usually greatest when the ratio LE_w/LE_d is least. The reason for this is that although canopy resistance (r_c) is lowest when solar radiation is greatest as around mid-day, thus giving lower r_c/r_a and hence LE_w/LE_d values, the greater available energy and the increased saturation deficit of the ambient air causes the rate of latent heat transfer to increase. But this increased evapotranspiration rate is greater for a wet canopy where potential surface conditions are satisfied, than it is for an unwetted canopy, especially if a shortage of soil moisture makes surface conditions highly non-potential. Also, as will be further demonstrated (see section 6.2) a wetted canopy acts as a strong sink for both incoming solar energy and for advected energy, and the rate of evaporation of intercepted rainfall thus often exceeds the rate of supply of net radiant energy. As a result the magnitude of the difference $LE_w - LE_d$ is greatest when available energy and the vapor pressure gradient, between the canopy and the ambient air, are greatest.

As noted earlier however (see section 2.1) transpirational water loss is suppressed during the evaporation of intercepted rainfall. This effect causes a certain amount of soil moisture saving, the magnitude depending upon the ratio LE_w/LE_d or the difference $LE_w - LE_d$. It will be shown later (see section 6.3) that on the average only a small percentage of intercepted rainfall goes towards soil moisture conservation while the remainder is lost to the atmosphere. In terms of the present experimental basin, therefore, water retained by the aerial parts of the vegetation can be considered as more of a loss than a saving of soil moisture. Consequently, when computing the water balance, the effect of the intercepted rainfall must be considered, since neglect of this component can lead to a sizeable underestimation of evapotranspiration.

2.3 Methodology

In order to attain the goals set forth in the preceding section the research will lead through several phases. Since the intent of this thesis is to highlight the effect of intercepted rainfall on evapotranspiration rates over a beech-maple forest in Quebec, it is necessary to devise a method whereby comparisons between evapotranspiration estimates for an unwetted (LE_d) as opposed to a wetted (LE_w) canopy can be made.

Different methods are available for the measurement of latent heat transfer over either wetted or unwetted vegetation. As will be explained later (see section 3.1) however, a general combination-type model that is attributable to Monteith (1965) will be used in this research. One of the main problems encountered in using this variant of the combination model is being able to derive hourly estimates of the canopy resistance (r_c) to vapor diffusion when the leaves are dry. As a solution to this problem, a model that predicts hourly values of the parameter (r_c) is proposed (see chapter 5).

To derive this model and to be able to make meaningful and objective comparisons of latent heat transfer over a wetted as opposed to an unwetted canopy for the forest under consideration, field measurements are necessary. In addition to evapotranspiration estimates for both wet and dry periods, other components of the water balance, namely precipitation, interceptional loss, channel runoff and soil moisture depth are to be measured. These supporting measurements will be used mainly to check the evapotranspiration calculations. The field methods and procedures utilized will be discussed elsewhere (see chapter 4).

Various forms of analysis are then to be performed on the field data. The results however will concentrate upon the particular effect of intercepted rainfall on evapotranspiration rates, and under different weather conditions (see chapters 6 and 7). Before any further discussion on the topics itemized in this section, some further theoretical considerations on the evapotranspiration - interception link are in order. This will accordingly form the subject matter of the next chapter.

CHAPTER 3

The Interception - Evapotranspiration Relationship: Theory and Measurement

3.1 Measurement of Evapotranspiration

Numerous methods are available for the measurement or estimation of evaporation and transpiration (W.M.O. Tech. Note No. 83, 1966; Federer, 1970). However the measurement techniques for the evaluation of the evaporative flux over natural surfaces and under natural conditions are still very much in the developmental stage, especially for non-potential surface conditions.

There are basically two main types of methods that are used to measure latent heat transfer, namely direct and indirect. Direct measurement techniques consist of two categories. The first group operates on the principle of the conservation of matter in that the change in volume of water over a defined time period forms the basis of the calculation. There are three types of instruments that adopt this principle, namely atmometers, evaporation pans and lysimeters. The first two types are designed to estimate the evaporative flux over a free water surface. But failure to fully represent natural conditions have limited their use. Lysimetry on the other hand, though suited for use over a crop or bare soil cover does not lend itself for use over tall vegetation. Besides, cost factors make other methods such as micrometeorological techniques more attractive.

The other direct method is the eddy - correlation technique that is attributable to Swinbank (1951). This method is based on the instantaneous fluctuations in the vertical component of the wind velocity and specific humidity about their mean values. However the extreme sophistication of the instruments and techniques required for the adoption of this method has limited its use.

Indirect methods also consist of two groups: the soil-water budget method and micrometeorological methods. The water budget method is simple in principle. Ignoring the contributions of lateral inflow or outflow and ground water seepage, evaporation is computed as:

$$E = P - R \pm \Delta Sm \qquad (3.1)$$

where

E = evaporation

P = precipitation

R = channel runoff

Δ Sm = change in soil moisture or lake storage.

This technique however is only suited for use over long time periods such as a growing season or year. Micrometeorological methods on the other hand are mainly theoretical in that they are attempts to measure the rate of latent heat transfer in terms of the physical principles controlling the vaporization process. The magnitude of the evaporative flux is basically governed by three processes: one is the rate of energy input necessary to change liquid water into vapor; another is the ease with which the vapor is removed from the evaporating surface;and the other is the rate of water supply to the vaporizing surface. For an open water surface only the first two conditions apply. The last parameter is however extremely important for non-potential surfaces such as vegetation or bare soil.

The most common and widely used micrometeorological techniques are the aerodynamic, the energy budget and the combination methods. The aerodynamic or mass transfer approach is a profile technique concerned with the turbulent transfer of water vapor between two levels in the ambient air, a small distance above the evaporating surface. The eddy motions associated with the turbulent flow of the wind give rise to the transport of water vapor and heat, the concentrations of which vary with height above the evaporating surface. By analogy with molecular and other transfer processes, basic transport equations for water vapor, heat and momentum may be stated as:

$$LE = \frac{-\rho\epsilon L}{p} K_w \frac{\Delta e}{\Delta z}, \tag{3.2}$$

$$H = -\rho c p K_H \frac{\Delta t}{\Delta z}, \tag{3.3}$$

$$\tau = \rho K_m \frac{\Delta u}{\Delta z}, \tag{3.4}$$

where

LE, H and τ = the vertical fluxes of water vapor, heat and momentum respectively

ρ = air density

cp = the specific heat of air at constant pressure

p = atmospheric pressure

ϵ = the ratio of the molecular weight of water to the molecular weight of dry air

L = latent heat of vaporization

K_w, K_H and K_m = the eddy diffusivities of water vapor, heat and momentum respectively

$\frac{\Delta e}{\Delta z}$, $\frac{\Delta t}{\Delta z}$ and $\frac{\Delta u}{\Delta z}$ = the vertical gradients of vapor pressure, temperature and horizontal air velocity, with height (z) respectively.

The problem with this approach is that ~~it~~ is generally not possible to measure K_w or K_H directly. However it is possible to calculate K_m using wind profile theory, and on the assumption that $K_m = K_w$ it is possible to determine the evaporative flux as:

$$LE = \frac{\rho \epsilon L k^2 \Delta u \Delta e}{p[\ln(z/z_o)]^2} \qquad (3.5)$$

This equation (3.5) is the classical Thornthwaite and Holtzman (1942) relation

where

k = von Karman's constant

z_o = the roughness length

ln = the base of natural logarithm

and all other terms are as previously defined.

The major drawback of this method is that it is only valid under conditions of neutral stability, when the lapse rate is adiabatic and when the assumption $K_m = K_w$ holds true (Munn, 1961). For non-adiabatic conditions corrections for stability must be made. However for heights of about one metre and less from the evaporating surface these corrections can be ignored (Sellers, 1965).

The energy balance method is also a profile technique, and is based on the principle of the conservation of energy. A measure of the amount of water transferred to the atmosphere is obtained by assessing the heat flux associated with the energy used to convert liquid water to vapor. Ignoring horizontal divergences of the heat fluxes and the storage of energy in the biomass, in air within the plant community, and in photosynthesis, the energy balance for a vegetated surface can be written as:

$$Rn = LE + H + G \qquad (3.6)$$

where

Rn = the net radiation flux

LE = the latent heat flux

H = the sensible heat flux

G = the soil heat flux.

Ordinarily LE is solved as a residual in the above equation. Both Rn and G are readily measurable but there is no simple method of measuring H. This difficulty is overcome by using the ratio H/LE = β (Bowen, 1926) to solve for LE. Rearranging equation (3.6) and dividing by LE gives:

$$LE = \frac{Rn - G}{1 + H/LE} . \qquad (3.7)$$

Now from the mass transfer equations (3.2 and 3.3) presented earlier, and assuming that ΔT and Δe are measured at the same heights one can deduce:

$$\frac{H}{LE} = \frac{C_p p}{L \epsilon} \frac{K_H}{K_w} \frac{\Delta T}{\Delta e} \qquad (3.8)$$

The term $C_p p/L\epsilon$ is known as the psychrometric constant, γ (0.66 mbar $^\circ C^{-1}$). The Bowen ratio (β) can then be calculated by:

$$\beta = \gamma \frac{K_H}{K_w} \frac{\Delta T}{\Delta e} . \qquad (3.9)$$

If it is assumed that $K_H = K_w$, it is possible to reduce equation (3.7) to:

$$LE = \frac{Rn - G}{1 + \gamma \frac{\Delta T}{\Delta e}} . \qquad (3.10)$$

As in the case of the aerodynamic approach, the short-coming of the energy budget approach lies in the assumption $K_H = K_w$, which strictly speaking holds true only for neutral conditions. However, Swinbank and Dyer (1967) and Dyer (1967) have shown that this assumption is valid for a wide range of atmospheric stability.

Both the aerodynamic and the energy-budget methods are basically profile techniques and they require measurements of the fluxes of momentum, heat and water vapor at more than one height. For tall vegetation such as a forest, the setting-up of instruments at more than one level is an extremely difficult task. Also because of the strong turbulent mixing caused by the rough forest canopy differences in the fluxes of temperature and vapor pressure with height are very small near the canopy top. As a result extremely sensitive and hazard-prone instrumentation is required. A method that requires measurements at only one height above the canopy is therefore desirable. Such a method is provided by the combination model, and this is the technique to be used in this thesis.

3.2 Combination Model

The success of both the energy-balance and aerodynamic methods depend on the measurement of surface temperature and vapor pressure, both of which are very difficult to measure (Penman 1955). In order to eliminate the measurement of surface parameters Penman (1948) combined the energy-budget and aerodynamic approaches, hence the origin of the term combination model. Penman (1948) started out with the simplest case, an open water surface where potential conditions readily apply. Restating the aerodynamic equation (3.5) in its simplest form and which he called the "sink strength" term, Penman (1948) wrote:

$$E = (e_s - e_d)\ f\ (u) \qquad (3.11)$$

where

E = evaporation rate in unit time

e_s = saturation vapor pressure at the evaporating surface

e_d = actual vapor pressure in the air at some height above (equivalent to the saturation vapor pressure at dew point temperature)

$f(u)$ = a function of wind speed

Then, ignoring soil heat storage, he restated equation (3.6), which he labelled the energy term as:

$$H = E + K \qquad (3.12)$$

where

H = the heat budget (net radiation)

E = evaporation (latent heat transfer)

K = heating of the air (sensible heat transfer)

Again, assuming that the transfer coefficients for heat and water vapor are the same, the Bowen ratio is stated as:

$$\beta = \gamma (T_s - T_a)/(e_s - e_d) \qquad (3.13)$$

where

β = Bowen ratio

T_s = surface temperature

T_a = air temperature

γ = psychrometric constant

and e_s and e_d are as defined previously (3.11). From equations (3.12) and (3.13) then, it follows that:

$$E = H/(1 + \beta) \qquad (3.14)$$

If it can be assumed that air temperature (T_a) near to the evaporating surface is equal to the surface temperature (T_s), then by substituting E_a for E in equation (3.11) and replacing e_s by e_a (the saturation vapor pressure at (T_a)), one can derive:

$$E_a = (e_a - e_d)\, f(u). \tag{3.15}$$

Then from (3.11) and (3.15):

$$E_a/E = 1 - (e_s - e_a)/(e_s - e_d), \tag{3.16}$$

which can be simplified to:

$$E_a/E = 1 - \phi \tag{3.17}$$

where

$$\phi = (e_s - e_a)/(e_s - e_d).$$

Now from equations (3.13) and (3.14)

$$E = H/(1+\beta) = H/(1 + \gamma\,(T_s - T_a)/(e_s - e_d)). \tag{3.18}$$

By setting $(T_s - T_a) = (e_s - e_a)/S$, where S is the slope of the saturation vapor pressure curve at T_a, then:

$$H/E = 1 + \gamma\,(e_s - e_a)/S(e_s - e_d) = 1 + \gamma\phi/S, \tag{3.19}$$

From equations (3.16) and (3.19) Penman (1948) finally derived the expression:

$$E = (SH + \gamma E_a)/(S + \gamma), \tag{3.20}$$

which can be further simplified to:

$$E = \frac{(S/\gamma)H + E_a}{(S/\gamma) + 1}. \tag{3.21}$$

Note that in equation (3.21) S is the slope of the saturation vapor pressure curve at air temperature (T_a). Also the aerodynamic component (E_a) of the equation is a measure of the drying power of the air, in that the term ($e_a - e_d$) in equation (3.15) is the saturation deficit of the air at mean air temperature. Thus the measurement of surface temperature and vapor pressure is eliminated. Air temperature and humidity both of which are elements that are easy to measure can now be used.

Equation (3.21) is the well-known potential evaporation formula since it was designed for use over surfaces for which water supply is non-limiting, namely a free water surface. For surfaces, such as vegetation or bare soil, where the supply of water for latent heat transfer can be a limiting factor, alternative methods have to be used. Different methods for the estimation of evaporation or transpiration have been presented by Slatyer and McIlroy (1961), Tanner and Fuchs (1968), Fuchs et al (1969) and Monteith (1965). Since the primary aim of the present research is to compare evaporation over a wetted canopy to transpiration over an unwetted canopy, the Monteith variant of the combination model, commonly known as the "canopy resistance" model is to be used since it can be readily applied to wetted or unwetted vegetation.

Furthermore, the method allows for an examination of how the soil-plant complex controls latent heat transfer. Plants ought to be treated as aerodynamically rough and mainly dry surfaces. The concept of potential evaporation being a water loss that encounters no restriction to vapor flow even from unwetted leaves, must therefore be re-examined. When a leaf surface is not wet, it is realistic that the rate of evaporation is less than the potential rate because of stomatal control that introduces a resistance to the diffusion of vapor. This latter aspect forms the basis of the "canopy resistance" model.

3.3 Canopy Resistance Model

The theory of the canopy resistance model is based on the assumption that exchanges of matter and energy between plant communities and the air can be described by an Ohms law relationship of the type:

Flux = Potential Difference/Resistance,

where the potentials of the system are the concentrations of diffusing gases and the heat content of unit volume of air, and the resistances are both external, characterizing the aerodynamic properties of the plant-atmosphere system, and internal, simulating the physiologic properties of the soil-plant complex (Monteith, 1963).

The concept of diffusive resistances in the plant atmosphere system was first introduced by Penman and Schofield (1951) in a crop model describing transpiration and the flux of carbon dioxide in photosynthesis. In an attempt to extend Penman's (1948) original combination method so as to consider non-potential evaporating conditions, they hypothesized that over plant communities, transpiration will proceed at a rate that is less than the potential evaporation rate because of a stomatal factor which impedes vapor diffusion and because of a daylength factor, since stomates are closed during the night. This was expressed by Penman (1952) in the following form:

$$E_T = \frac{S\,H + \gamma(e_a - e_d)\;f(u)}{(S + \gamma)/S^*D} \qquad (3.22)$$

where

E_T = transpiration rate in appropriate units

S^* = stomatal factor

D = daylength factor,

and all other terms are as previously defined.

It was Monteith (1965), however, who developed the "resistance" combination model. He maintained that evaporation over non-potential plant surfaces is sustained by a supply of heat from the atmosphere and by a movement of water within the plant preventing the desiccation of the leaf tissue. Furthermore, the path for the diffusion of water from leaf cells to the free atmosphere is divided into two parts: across a canopy resistance (r_c) determined primarily by the size and distribution of stomatal pores and by the leaf area index, and across an air resistance (r_a) determined by the wind speed and the aerodynamic properties of the plants.

Restating equation (3.21) in terms of potential differences and diffusive resistances Monteith (1965) wrote:

$$LE_w = \frac{S\ Rn + \rho c\left\{e_s\ (T) - e\right\}/r_a}{S + \gamma} \qquad (3.23)$$

where

LE_w = potential evaporation rate (cal., cm^{-2}, $sec.^{-1}$)

Rn = net available energy (cal., cm^{-2}, $sec.^{-1}$)

$e_s(T)$ = saturation vapor pressure of air temperature at height Z (mbar)

e = actual vapor pressure of air temperature at height Z (mbar)

r_a = aerodynamic resistance to vapor transfer at height Z (secs $cm.^{-1}$)

ρc = volumetric heat capacity of the air (2.9×10^{-4} cal. cm^{-3} $^{o}C^{-1}$)

S = slope of the saturation vapor pressure curve at air temperature T (mbar ^{o}C)

γ = psychrometric constant (o.66 mbar^{o}C).

It is readily apparent that equation (3.23) is exactly the same as equation (3.21) except that $\rho c\ (e_s(T) - e)/r_a = (e_a - e_d)\ f(u) = \gamma E_a$, or more precisely $\rho c/r_a = f(u)$.

Neither equations (3.21) nor (3.23) are valid for a surface where the vapor pressure is less than the saturation vapor pressure at surface temperature. For an unwetted leaf, water evaporates at the surface of the cell walls surrounding the sub-stomatal cavities and reaches the outer surfaces of the leaf by molecular diffusion through stomata and cuticle. When a leaf has a uniform temperature T_o, the vapor pressure of air in contact with the stomatal cell walls is usually very close to the saturation vapor pressure $e_s(T_o)$. At the dry leaf surface however, outside the cell walls, the vapor pressure (e_o) during transpiration is always less than $e_s(T_o)$, and surface air in contact with the leaf is never saturated. As a result, the unsaturated air at the leaf surface has a finite wet - bulb depression, a parameter which is difficult to measure. Following the lead of Penman (1948), Monteith (1965) resolved this difficulty by eliminating surface parameters and replacing them by ambient air conditions.

By assuming that the transfer coefficients and hence the resistance to diffusion (r_a) of heat and water vapor are the same, the transpiration rate can now be given as $(e_o - e)/r_a$, where eo is the vapor pressure at the leaf surface. Similarly the rate of vapor diffusion within the stomates of the leaves is proportional to $(e_s(T_o) - e_o)/r_c$ where r_c is the internal stomatal resistance. Since within a thin leaf, the temperature of the stomatal wall and the surface can be assumed equal, the rates of diffusion within the leaf and external air are then also equal and:

$$E(\gamma L/\rho c) = \frac{e_s(T_o) - e_o}{r_c} = \frac{e_o - e}{r_a} \qquad (3.24)$$

where the constants ($\gamma L/\rho c$) preserves the consistency of units. By rearranging terms:

$$e_s(T_o) - e = \left\{1 + \frac{r_c}{r_a}\right\} (e_o - e) \qquad (3.25)$$

Equation (3.25) shows that a leaf with internal resistance (r_c) can be treated formally as a free water surface if the total vapor pressure drop ($e_o - e$) is replaced by $e_s(T_o) - e/(1 + r_c/r_a)$.

Returning to equation (3.23) and replacing surface temperature (T_o) by air temperature (T) as suggested by Penman (1948), the latent heat of transpiration from a leaf becomes:

$$LE_d = \frac{S\ Rn + \rho c(e_s(T) - e)/r_a}{S + \gamma\ (1 + r_c/r_a)} \ . \qquad (3.26)$$

It should be noted that equation (3.26) applies to both a wetted and a dry canopy. When leaves are wet $(e_o - e) = (e_s(T_o) - e)$. In other words the stomatal resistance (r_c) becomes zero and equation (3.26) is reduced to equation (3.23). Both equations (3.23) and (3.26) show that the rate of evaporation over a vegetated surface depends on three weather parameters: net available energy, humidity of the ambient air and wind speed at some fixed height above the canopy, and a plant parameter, namely canopy resistance to vapor diffusion. The methods used to measure or estimate these components in the present research will be discussed at length in the next chapter.

3.4 Measurement of Interception

As in the case of evaporation and transpiration, different methods can be used to measure interception. By definition the interceptional loss is the amount of precipitation that is prevented from reaching the ground by the aerial parts of vegetation. This loss is accounted for by two processes: evaporation of water from the canopy during the period of wetting or rainfall, and evaporation of intercepted rainfall following wetting. Interceptional loss therefore depends on weather conditions during wetting since they control the amount of evaporative loss and on canopy characteristics which determine the retention capacity of the stand.

Although the woody parts of vegetation such as branches and stems retain some moisture, by far the greater part of intercepted rainfall is withheld by leaves. Generally a leaf aborbs little, if any, water from its surface (Rutter, 1963). Its storage capacity may therefore be considered to be the amount of water it can retain on its surface. This amount is a function of leaf size, its configuration and composition, together with the viscosity of the water and the external pressure on the liquid, as well as the amount of precipitation (Leonard, 1967). The alignment of branches, canopy density and smoothness of the woody parts, such as the bark, are other determining plant factors.

Weather factors or more precisely rainfall characteristics also affect interceptional loss, in that evaporation of intercepted rainfall can occur during wetting. Wilm and Niederhof (1941) observed that about 19 percent of each storm is lost to the ambient air by evaporation from the canopy during rainfall. As will be seen later (see chapter 7) intermittent and light intensity rainfalls allow for greater evaporation of intercepted rainfall during wetting than moderate and heavy intensity rainfalls. This is attributable to the longer duration of the former category and also to the greater vapor pressure deficit of the ambient air that accompany these rainfalls. Windiness that produces shaking of the branches, and the impact of velocity of raindrops during severe storms also affect the interceptional loss, because they lessen the retention capacity of the canopy.

In view of the preceding, both direct and indirect techniques have been evolved for the measurement of interceptional loss. The most commonly used direct method is a water-balance technique based on the conservation of matter: interceptional loss in considered to be a residual after accounting for water input and output. Written in its simplest form, the equation is:

$$I_L = P - (T + S_f) \tag{3.27}$$

where

I_L = interceptional loss

P = precipitation above the canopy

T = throughfall

S_f = stemflow.

Other direct methods are basically extensions of equation (3.27). They however regard the interceptional loss as dynamic rather than passive in nature and as a result are more process-oriented. Horton (1919) observed that the amount of interceptional loss is equal to the sum of water stored on the plant surfaces at the end of a storm and subsequently returned to the atmosphere by evaporation, plus evaporation from the wet plant surfaces during the storm. He expressed this relationship in an equation in the form:

$$I_L = S + R E T \tag{3.28}$$

where

I_L = interception loss in unit depth over the projected area of the canopy

S = the water stored on vegetation in unit depth over the projected area of the canopy

R = the ratio of vegetation surface area to the projected area of the canopy or leaf area index

E = evaporation rate in unit depth per unit time during the storm

T = duration of storm in unit time.

Alternatively, he expressed the relationship in the form:

$$I_L = S + K\,P \qquad (3.29)$$

where

K = (RET/P), i.e. the fraction of precipitation lost by evaporation to the ambient air and absorption by the plant, if any during the storm

P = storm precipitation in unit depth and I_L and S are as defined above.

This second formula (3.29) only applies when P is greater than S. Otherwise I_L is nearly equal to P.

Linsley et al (1949), however, argued that equation (3.28) yielded a value of interception which was independent of the amount of precipitation, since it assumed that the rainfall in each storm filled interception storage. As an alternative it was suggested that if it could be assumed that the interception loss given by (3.28) could be treated exponentially in terms of rainfall increasing from zero to some higher value, for a specified duration then:

$$I_L = (S + RET)\ (1 - e^{cP}) \qquad (3.30)$$

where

e = the base of Naperian logarithms

c = constant

and all other terms were as defined before.

Merriam (1960) however questioned the application of an exponential factor to the evaporation (RET) portion of the equation. Alternatively he suggested the following equations:

$$I_L = S(1 - e^{cP}) + RET \text{ for } P \leq S \qquad (3.31)$$

and

$$I_L = S(1 - e^{cP}) + KP \text{ for } P > S \qquad (3.32)$$

He further differentiated both equations with respect to P, and equated $\delta I_L / \delta P$ to unity as P approached zero, and from this he derived that c is equal to 1/S and the general equation thus becomes:

$$I_L = S(1 - e^{P/S}) + RET \text{ for } P \leq S \qquad (3.33)$$

$$I_L = S(1 - e^{P/S}) + KP \text{ for } P > S \qquad (3.34)$$

where all values are as defined previously.

More recently Rutter et al (1971/72, 1975) proposed a running water-balance technique to measure the interceptional loss. The canopy is regarded as having a surface storage capacity, S, which is charged by rainfall and discharged by evaporation and drainage. The rate of water inflow, P_I, to the canopy, is given by:

$$P_I = (1 - p)\, P \qquad (3.35)$$

where p is the fraction of the precipitation (P) which falls through gaps in the canopy. The rate of water outflow via evaporation, E, when the canopy is wet is given by:

$$E = E_p(C/S) \qquad (3.36)$$

where

E_p = evaporation rate as given by equation (3.23)

C = the amount of water present on the canopy

S = canopy storage capacity.

When $C = S$ then, $E = E_p$.

The rate of drainage, D, from the canopy is given by:

$$\ln D = a + bc \qquad (3.37)$$

where

b = a drainage coefficient

a = constant

ln = base of natural logarithm.

In terms of a water balance it follows that:

$$P_I = E + D. \qquad (3.38)$$

The interceptional loss I_L can then be said to equal E in equation (3.36).

Indirect methods on the other hand are basically empirical techniques whereby the interceptional loss I_L, is expressed as a function of precipitation, P, above the canopy (Zinke, 1967). The general form of the relationship is:

$$I_L = bP + a \qquad (3.39)$$

where b and a are the slope and intercept respectively of the linear regression equation.

It is implicit that the constants and coefficients in all of the equations for measuring or estimating the interceptional loss presented thus far refer to a particular vegetation stand. Since the vegetal cover being examined in the present thesis is in a sense unique in terms of regional climate and canopy characteristic, new formulations must be developed. Also the emphasis in the present study is not to describe the interception process, but rather to compartmentalize it so as to be able to compare evaporation of intercepted rainfall during and after the period of wetting. The effect of wetting on water consumption and hence the water balance of the forest basin is also of major relevance. The equations that follow then are the ones used in this thesis.

If the interceptional loss, (I_{Lp}) were to be viewed as being a passive process in the sense that the amount of interception is only a function of the retention capacity of the foliage, then the following expression could be written:

$$I_{LP} = CS \qquad (3.40)$$

where

I_{LP} = passive interceptional loss in mm depth of water

CS = canopy retention storage in mm depth of water.

It should be noted that in this equation canopy storage (CS) is different from Merriam's (S) in that CS more precisely refers to canopy retention, or "residual storage" (Grah and Wilson, 1944) which could be equal or less than the saturation value.

But as suggested earlier (Horton, 1919) interceptional loss is more a dynamic process, in that evaporation of intercepted water can occur during a rainfall. Thus the dynamic interceptional loss (I_D) can be found from:

$$I_D = \int_{T_1}^{T_2} LE_w \, \delta T \qquad (3.41)$$

where

I_D = dynamic interceptional loss in mm depth of water

T_1 = time of beginning of rain event

T_2 = time of ending of rain event

LE_w = evaporation of intercepted rainfall during the rain event, as given by equation (3.23).

From equations (3.40) and (3.41) it follows that canopy detention storage (CS) can be calculated from:

$$CS = \int_{T_2}^{T_3} LE_w \, \delta T \qquad (3.42)$$

where

T_3 = time of complete disappearance of intercepted rainfall from canopy

and all other terms are as previously defined.

Finally, the total interceptional loss during and following each rain event can be calculated as:

$$I_T = I_D + CS \qquad (3.43)$$

where

I_T = total interceptional loss for each rain event in mm depth of water.

Equation (3.43) can then be simplified to give:

$$I_T = I_D + \sum_{T = T_2}^{T_3} LE_w \qquad (3.44)$$

and finally, by considering equations (3.41 and 3.42), to:

$$I_T = \sum_{T = T_1}^{T_3} LE_w \qquad (3.45)$$

Equations 3.41 to 3.45 can be used to measure interceptional loss both during and after the wetting period. The method for estimating LE_w has been discussed to some extent (see section 3.3). To estimate T_1 and T_2 traces from the rainfall intensity gauge (see section 4.3) are utilized. T_3 can be measured by means of a moisture sensor whose method of use will be described later (see section 4.7).

3.5 Interception - Evaporation Relationship

So far discussion on interception and evaporation has been kept somewhat apart. If a plant canopy is wetted by rainfall however, interesting inter-relationships between these two processes emerge. When a leaf canopy becomes wet, the plant no longer needs to act as a pump to supply the water for evapotranspiration in the sense that transpirational cooling is effected through evaporation of intercepted rainfall. In general, therefore, evaporation of intercepted water reduces transpirational losses from the soil, or, looked at in another way, the transfer of water from soil to air is actually short-circuited.

As mentioned previously (see chapter 2) however, and as will be seen later (see section 6.3), the evaporation of intercepted rainfall usually proceeds at a faster rate than the transpiration rate for an unwetted canopy and for which non-potential conditions apply, assuming the same weather conditions. The approximate ratio of evaporation of intercepted water to transpiration can be gauged from equation 3.46 given below. The equation (3.46) assumes that the net radiation and the saturation deficit of the ambient air do not change appreciably following a period of wetting. Writing equation (3.23) for r_c = zero (LE_w) and r_c = finite (LE_d) then:

$$\frac{LE_w}{LE_d} = \frac{(S/\gamma) + 1 + r_c/r_a}{(S/\gamma) + 1} \qquad (3.46)$$

where

LE_w = evaporation rate when the canopy is wet

LE_d = transpiration rate when the canopy is dry

and all other terms are as defined previously.

When r_c is zero, the fraction LE_w/LE_d is unity. But when the plant resistance (r_c) takes on a finite value, the ratio LE_w/LE_d becomes greater than unity, the magnitude depending on the value of the ratio r_c/r_a and S, which is a function of air temperature.

Rutter (1967) following the lead of Rijtema (1965), suggested an alternative method for estimating evapotranspiration from vegetation, which takes account of the separate contributions of intercepted and transpired water. In order to express total evapotranspiration in a form consistent with equation (3.46) so as to appraise the significance of interception in the water consumption of forests, he suggested another approach.

In his formulation Rutter (1968) viewed interception as a potential loss. He accepted that equation (3.26) which contains an internal diffusion resistance, is strictly an expression of a transpiration rate (LE_d). Also, if r_c (canopy resistance) is set equal to zero, the equation becomes an expression for the rate of evaporation of intercepted water (LE_w), which is the same as equation (3.23). Furthermore, Rutter (1967) found that on the average LE_w was four times as great as LE_d. If it is assumed, therefore, that there is no transpiration while intercepted water is being evaporated, then one quarter of the intercepted water is equivalent to transpiration that would otherwise have occurred in the same atmospheric conditions, while three quarters was evaporation that would not have occurred in the absence of precipitation and interception. Rutter (1968) therefore contended that the additional evaporation consequent on interception (I_N) which is the same as the net interception loss of Burgy and Pomeroy (1958), can be stated as:

$$I_N = I(LE_w - LE_d)/LE_w. \qquad (3.47)$$

where

I = the unit depth of water intercepted.

When $I = LE_w$, equation (3.47) simply becomes:

$$I_N = (LE_w - LE_d) \qquad (3.48)$$

Equation (3.47) can also be simplied to

$$I_N = (1 - LE_d/LE_w)I \qquad (3.49)$$

Remembering that the inverse of equation (3.46) can also be written as:

$$LE_d/LE_w = \frac{S + \gamma}{S + \gamma((r_a + r_c)/r_a)} \qquad (3.50)$$

where all terms to the right are as in equation (3.26), then:

$$(1 - LE_d/LE_w)\,I = \frac{\gamma r_c I/r_a}{S + \gamma((r_a + r_c)/r_a)} \qquad (3.51)$$

From equations (3.49 and 3.51) the net interceptional loss (I_N) can then be written as:

$$I_N = \frac{\gamma r_c I/r_a}{(s + \gamma)r_a + r_c} \qquad (3.52)$$

It is apparent from the above equation that the net interceptional loss is dependent upon the ratio r_c/r_a and S which is temperature dependent. In other words, when r_c is large relative to r_a and S is small, the fraction of I approaches 1, and the net interceptional loss is greatest.

The preceeding sections have dealt at large with the theory and methods of measurement or estimation of evaporation and interception, and the relationship between these two components. In order to test and to validate the concepts that highlight the effect of intercepted rainfall on evapotranspiration rates field experimentation must be attempted. A mixed-hardwood forest in Southern Quebec is to be used as the test site. The measurement techniques and methods of experimentation will be discussed in the next chapter.

CHAPTER 4

Site, Instrumentation and Experimental Procedure

4.1 Physical Characteristics of Site

The field research for this experiment was conducted at Mont St. Hilaire (45° 33'N, 73° 10'W) Quebec, about twenty miles east of Montreal. The mountain is one of a group of landforms known collectively as the Monteregian Hills which are believed to be igneous intrusives that have been exposed by various erosional processes (Phillips, 1972). Mont St. Hilaire rises very sharply, in most places, from the St. Lawrence Lowlands up to a height of approximately 410 metres above sea level, which is about 370 metres above the surrounding plain. Seven distinct peaks on the perimeter of the mountain enclose a central basin, the lowest part of which is occupied by a shallow lake named Lac Hertel. The local relief of the mountain is about 250 metres.

The main experimental site chosen was at the southern gently sloping base of Lake Hill (see figure 4.1). A small tributary basin of West Creek was also selected for testing the water balance.

Soil types in the region range from a clay to a sandy or gravelly loam texture belonging to the Dystric Brunisol group (Canada Department of Agriculture, 1974). Soil depths vary considerably from place to place, but seldom exceed 1 metre (Wilson, 1968).

With the exception of the lake, various steep rock surfaces on some of the outer slopes and an orchard in the interior basin, the mountain is completely forested. The forest is an undisturbed mixture of deciduous hardwoods, the dominant species being American beech (Fagus Grandifolia Ehrh.) and sugar maple (Acer Saccharum Marsh).

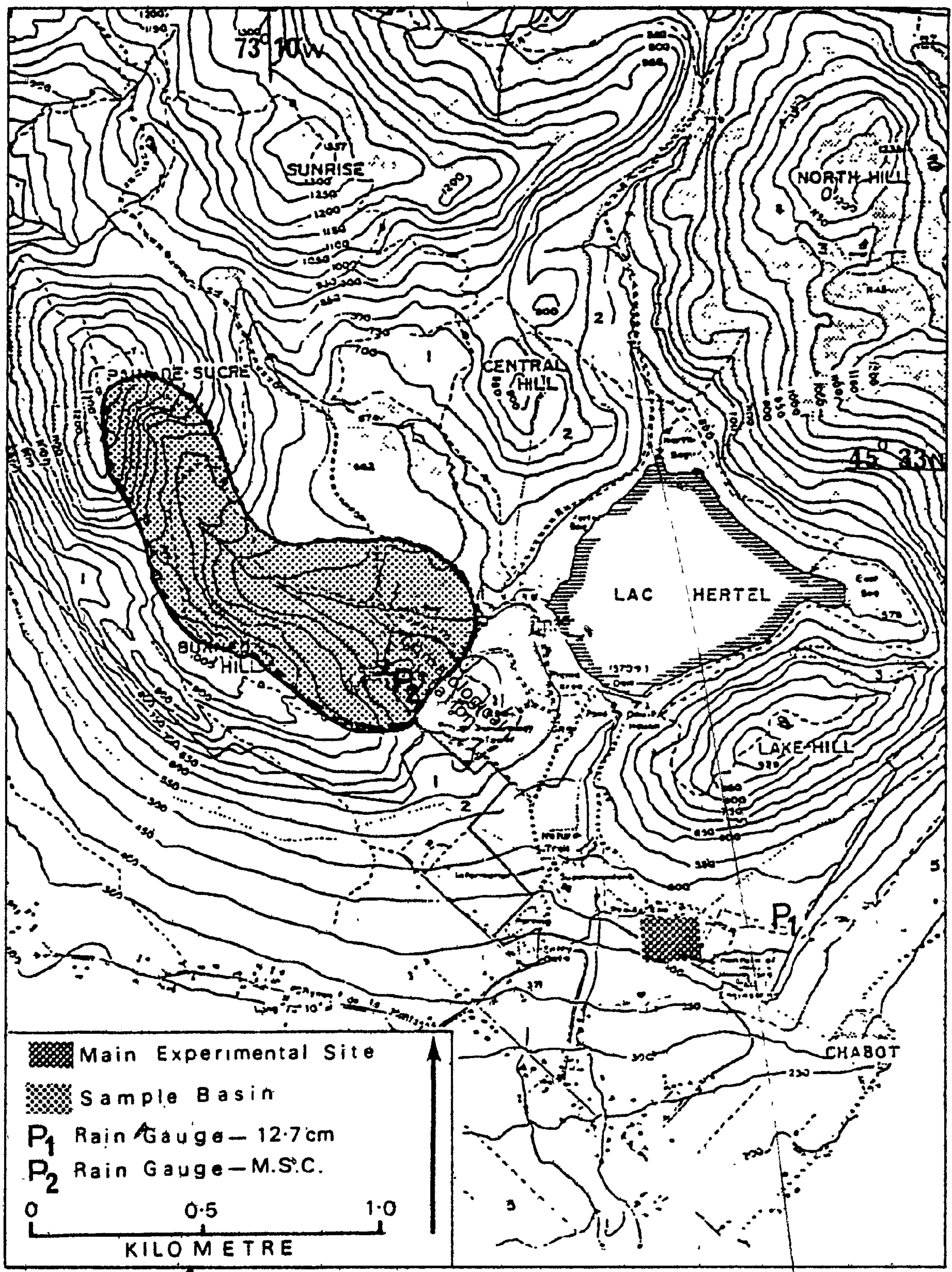

Figure 4.1 General site

4.2 Measurements and Procedures

The period of field measurements spanned two growing seasons, namely May to October, 1974 and 1975. Data collection during the first summer was delayed by instrumentation problems. During the following season, however, measurements were not only uninterrupted but were also refined based on the previous season's experience. Since the ultimate aim of the research was to evaluate the effect of intercepted rainfall on the water balance, all components of this balance were measured.

4.3 Precipitation Measurements

Incoming precipitation in all forms over and within the forest, namely open, throughfall and stemflow and interception were measured. All measurements, except interception, were made directly using different forms of gauges.

(a) Rain: Above Canopy

Because of the extreme difficulty in installing gauges above the canopy, two open sites were used to measure incoming precipitation (see figure 4.1). One gauge (12.7 cm diameter) was placed in an opening in the forest near the main experimental site. Appropriate exposure standards were satisfied by the location chosen. The other gauge was located in the open orchard where a climatological station is in operation. This was an M.S.C. pattern tipping-bucket rain gauge, with a 25.4 cm diameter receiver, and a recorder to keep a track of rainfall intensities. Rain records were measured or checked following each rainfall.

(b) Throughfall

Throughfall can be described as that portion of the rainfall which reaches the ground directly through the vegetative canopy, through openings and as drip from leaves, twigs and stems. Two methods are commonly used to calculate throughfall. The first technique estimates throughfall from gross or above-canopy rainfall by means of a simple linear regression equation (Leonard, 1961; Helvey and Patric, 1965; Zinke, 1967). The general form of the relationship is:

$$T = bP - a \qquad (4.1)$$

where

T = throughfall in appropriate units

P = gross rainfall in the same units,

and a and b are the regression coefficients.

The other method is measuring throughfall directly by means of a number of gauges. Since the regression coefficients in equation (4.1) are not universal, in that they are unique for each forest stand, the latter technique was used.

Throughfall was accordingly measured by means of 6 gauges located within the main experimental area (see figure 4.2). Because of the lengthy and arduous task of measuring large volumes of water by hand the number of gauges was kept at a minimum. The selection of gauge locations can best be described as systematic random, in that gauges were placed so as to sample the whole range of throughfall values: at least one gauge was placed where the canopy was thick, one where it was thin, and one where there was an opening.

The type of gauge used was a table with a 1.2m X 0.6m

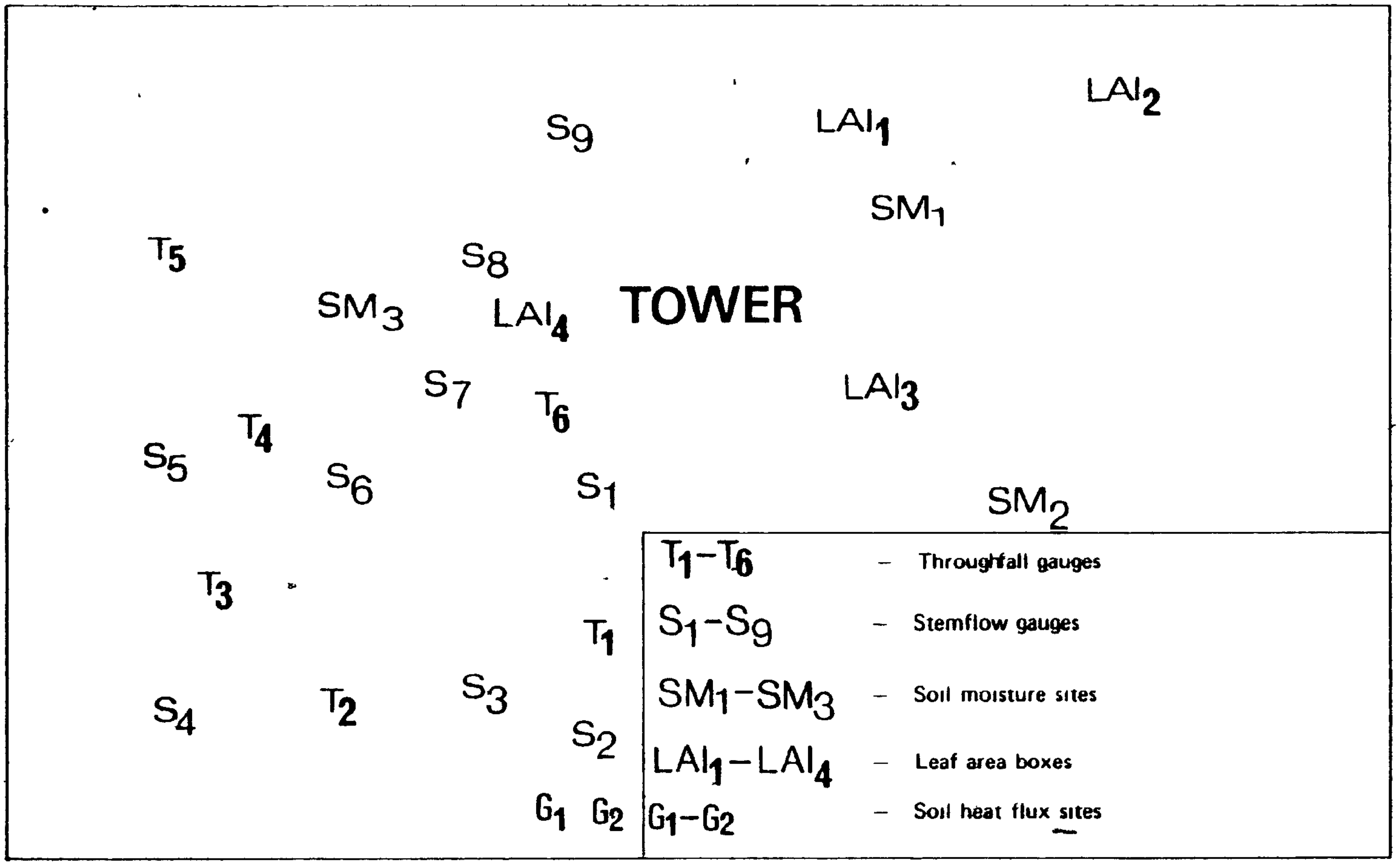

Figure 4.2 Plan of main research site

corrugated plastic top that sloped at an angle, from the horizontal, of about 10 degrees, and drained into metal eaves troughs that in turn directed the water into plastic garbage pails (0.6 m tall, 0.45 m upper diameter) (see figure 4.3). A measuring plastic jug, calibrated against a measuring jar whose capacity was the equivalent of 12.7 cm of throughfall was used to facilitate measurements. As with rainfall measurements in the open, the measurement of throughfall was made after each period of rain.

(c) Stemflow

According to Zinke (1967) stemflow can be described as that portion of the rainfall which, having been intercepted by the canopy, reaches the ground by running down the stems, and branches, and draining down the trunks. As in the case of throughfall, two methods are available for the measurement of stemflow. The indirect technique consists of using a regression equation whose form is:

$$S = bP - a \tag{4.2}$$

where

S = stemflow in appropriate units and

P = gross rainfall in the same units

and a and b are the regression coefficients. For the same reasons as given for throughfall, the following direct measurement technique was used.

Stemflow was measured by taking a sample of trees that varied both in species (4 maple and 5 beech) and in trunk diameter. Again, sample size was restricted by the time-consuming task of measuring large volumes of water. The stemflow gauges consisted of ordinary 2.54 cm diameter

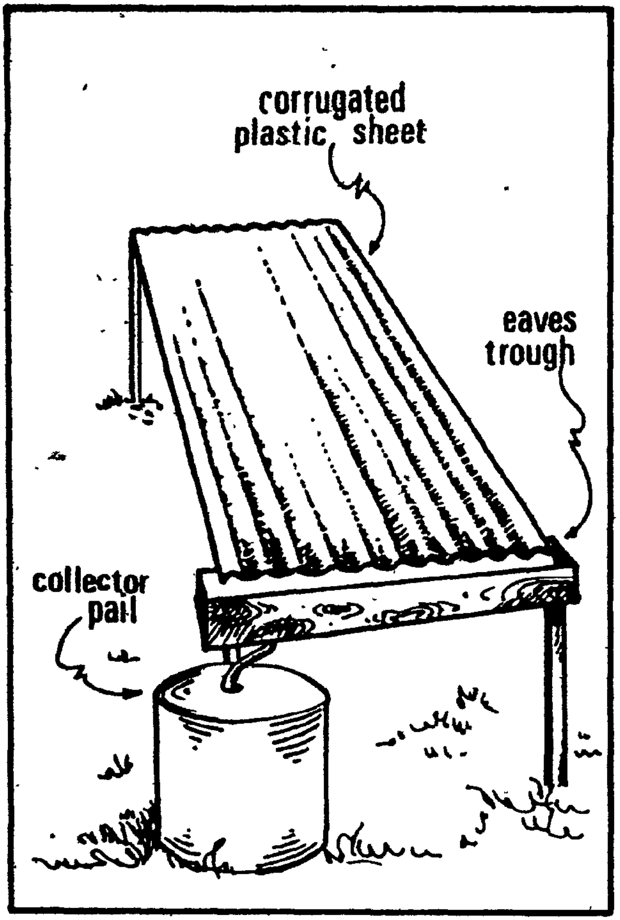

Figure 4.3 Throughfall gauge

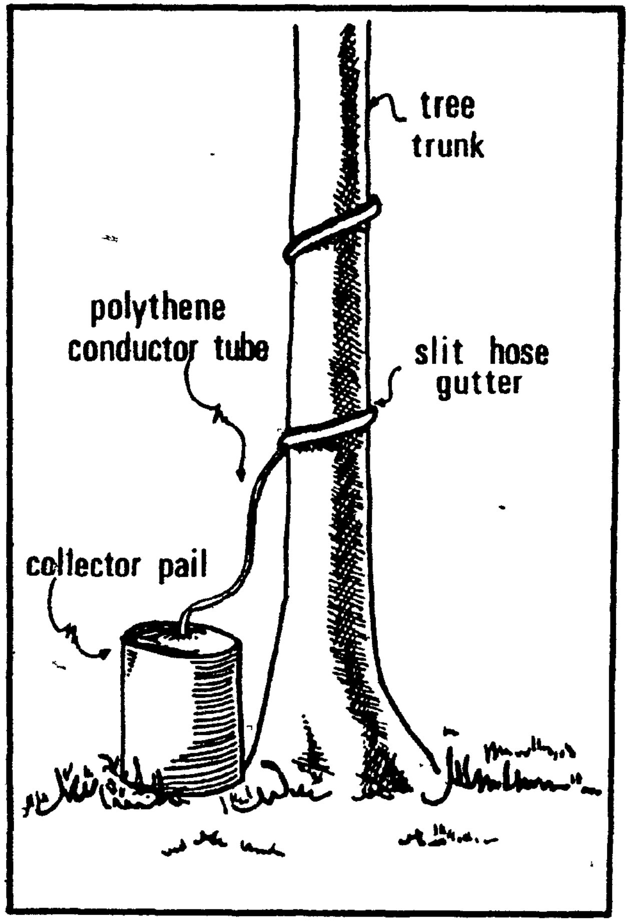

Figure 4.4 Stemflow gauge

garden hose slit into two and wrapped around the trunks (see figure 4.4). The semihose was secured to the trunk with small tack nails and the remaining spaces were sealed with rubber calk. The lower end of the hose was left intact and drained into covered garbage pails that varied in size (0.6 m tall X 0.45 m diameter to 0.91 m tall X 0.61 m diameter) according to tree size.

Judging from visual checks both during and after rain events, this design of gauges performed remarkably well under most weather conditions. Only during rainfall of very high intensities was there a small amount of overflow from sections of the rubber collars. During heavy rainfall accumulation (> 1.27 cm) gauges for the bigger trees had to be measured and emptied occasionally during the rainfall to avoid overflows.

The area of each stemflow gauge was derived by tracing the outline of the tree canopy as accurately as possible on the ground, and then estimating the area of the outline. This area was also calibrated against the area of a 12.7 cm diameter gauge and the same plastic jug as used for throughfall was used to facilitate measurements.

(d) Interception

The definition of interception used in the present work is that suggested by Zinke (1967): rainfall retained on standing vegetation and evaporated without dripping off or running down the stems. This is the same as the "residual storage" of Grah and Wilson (1944).

Several methods are also available for the measurement of interceptional loss. For the same reasons given earlier (see section 3.4) interceptional loss, for the purposes of this research, is calculated firstly by subtracting the sum of throughfall plus stemflow from gross rainfall (see equation 3.27) and secondly by estimating the total evaporative loss when the canopy is wet (see equation 3.45).

4.4 Evapotranspiration Measurements

As was mentioned earlier the surface resistance version of the combination model as expressed by equations (3.23) and (3.26) can be used to estimate evapotranspirational losses over the forest.

The model is ideally suited for an experiment of this nature in that it only requires measurements at a single height. This is justified in terms of the difficulty of setting up profile methods above tall vegetation such as would be necessary for the energy balance technique. Also because of the very rough nature of the forest extremely sensitive and hence hazard-prone instrumentation would be required. Most importantly, however, the model can be used to estimate evapotranspiration losses for both a wet and a dry canopy. The greatest problem with using this model, however, is deriving estimates of the aerodynamic and canopy resistances. These problems will be discussed in greater detail subsequently. As is evident from the model the meteorologic parameters that need to be measured are net radiation, soil heat flux, wet and dry-bulb air temperature and the aerodynamic resistance to vapor diffusion at an appropriate height above the canopy. The surface parameter which must be obtained is basically the degree of wetness of the canopy, a feature which is characterized by either a canopy resistance to vapor diffusion (r_c finite) or the presence of intercepted rainfall (r_c zero).

Except for the measurement of the soil heat flux all variables need to be measured at some height either within or above the forest canopy.

To achieve this end two triangular television towers consisting of 2.5 metre steel sections and rungs 0.7 metre apart were constructed 2.5 metres apart. These towers were secured by guy wires attached to neighbouring trees and protruded to a height of about a half-metre above the forest top. A three metre mast with instrumentation for measuring above-canopy parameters, namely net radiation, dry and wet-bulb air temperatures, and wind speed, was then attached to one of the towers, with its top at a reference height of about 1.5 metres above the mean height of the canopy. Above this was another smaller mast with a pulley attachment which was used to lower the instrument mast on occasions when instrument repairs had to be performed. A lightning rod with an insulated ground wire was also installed to protect the equipment from voltage overload during thunderstorms.

In order to provide a working base for instrument installation and repair, and for taking leaf resistance measurements, as well as for adding further stability to the towers, two wooden platforms (2.5 m X 20.32 cm) that connected the towers were erected at heights of 12 metres and 15 metres. Wooden railings were also mounted about 1.5 metres above each platform, and these served as safety devices as well as supports for permanent clip-boards for writing data. An electrician's climbing belt was also used to ensure safety and to facilitate manoeuvrability. A detailed description of the towers, platforms and mounted equipment is given in figure 4.5.

Signal outputs were led from the equipment on the mast to a control box at the bottom of the tower by a series of wires where fuses were installed for safety purposes. From here the signals were fed over a distance of a couple hundred metres to a laboratory where a further

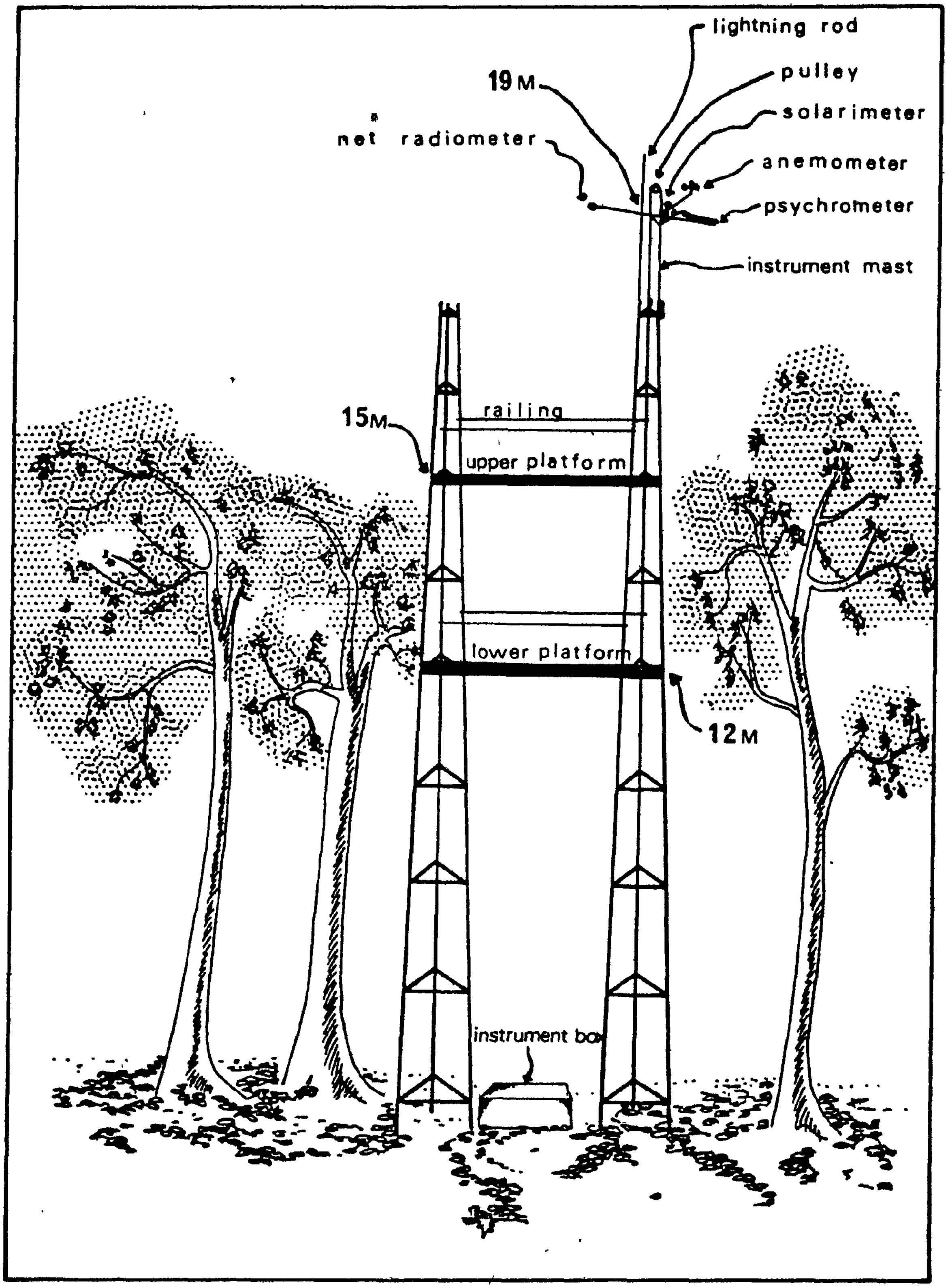

Figure 4.5 Tower design and instrumentation

set of fuses was used before the wires were connected to continuous chart recorders. Continuous recordings were restricted mainly to the duration of daylight.

Checks at the beginning and end of each season's experimentation showed no depreciation in signal output from tower site to laboratory.

(a) Net Radiation

Net radiation was measured as stated previously at a height of 1.5 m above the canopy, using an S.R.I. net radiometer. The sensing element was on a long enough arm so as to avoid tower interferences. In order to equalize convective heat losses from both sides of the thermopile plate the polyethylene domes were kept inflated by an aquarium pump housed at the bottom of the tower. The air was however first blown past a reservoir of silica gel so as to prevent internal condensation. The proper pressure adjustment of the pump was made through manipulation of the bubble rate (4 - 5 per minute) in a water bottle, into which the back pressure was fed. The signal was continuously recorded on an I.C.A. model 400 strip chart recorder running at a speed of 2.54 cm per hour. Mean hourly values of these traces were then derived, using the constant calibration factor of the net radiometer of 55.8 MV ly min^{-1}.

(b) Wet and Dry-Bulb Air Temperature

The dry and wet-bulb temperatures of the air at reference height were measured by means of home-made copper-constantan thermocouples, with a calibrated sensitivity of 39.5 mv/°C. The thermocouples were shielded and insulated by 2 sizes of P.V.C. pipes, the outer one being coated with aluminum foil to restrict radiation absorption. Both sensors were held tightly within and away from the inner tube by two wooden shafts that were anchored in a tightly sealed rubber cork at one end of the tube. Artificial ventillation was used to aspirate the thermometers by drawing air past the bulbs at a rate of about 4 metres sec.$^{-1}$ with a vacuum fan.

The thermocouples were set up in such a way that the dry bulb acted as the common hot junction. The cold junctions consisted of the wet bulb thermocouple and a zero °C reference point. The ice point reference consisted of a Zeref chamber located at the foot of the tower that continuously maintained an oil bath at zero °C.

Maintaining the proper water supply to the wick of the wet-bulb thermometer was rather problematical, in that the water reservoir, at the top of the mast needed to be refilled about every three weeks. The wick encasing the wet bulb consisted of a white cotton shoe-lace that was fed from a one-litre volume plastic bottle through a 1 cm diameter plastic tubing.

In order to avoid having to lower the mast, which was practically impossible without disrupting the other instruments, the following bottle refilling procedure was devised. One end of a plastic Y-junction was fitted to a 1-litre volume squeeze bottle. Another outlet was fitted to a 1 cm diameter plastic tubing that led down from the top of the reservoir to the upper platform. The remaining outlet was fitted to an ordinary hand pump with an extension of 1 cm-diameter plastic tubing. With the squeeze bottle filled and sealed tightly and all other outlets air-tight, pressure was applied using the hand pump. The pressure buildup

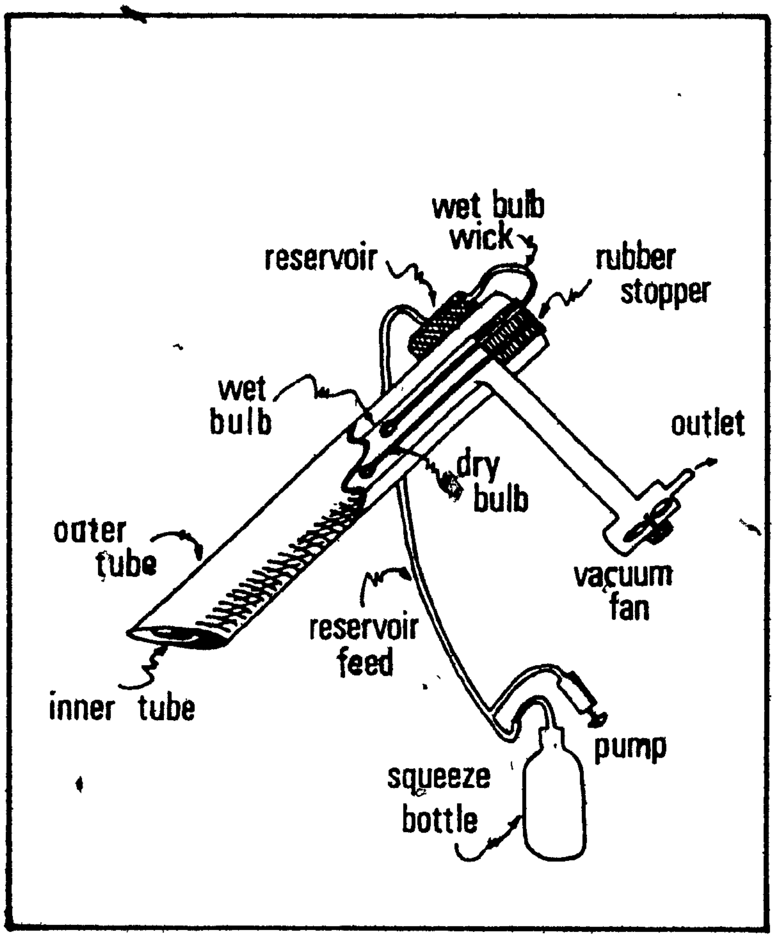

Figure 4.6 Psychrometer design

in the squeeze bottle was sufficient to take the water all the way up to the reservoir, a head of about 4 metres. A tiny hole at the top of the reservoir was utilized to release the back-pressure. It took only about ten minutes to refill the reservoir using this technique. The only drawback of the method, however, was that with excess air pressure in the reservoir, the wet-bulb wick became saturated, so that it sensed water temperature rather than evaporative cooling. Fortunately, this anomalous condition only lasted about 10-15 minutes. Greater details of this and the psychrometric design are shown in figure 4.6.

(c) Soil Heat Flux

The soil heat flux represents changes in energy storage in the ground, and theoretically this flux must be determined at the soil surface. Energy storage in the biomass, and that used in photosynthesis were ignored since these together account for a minimal amount of net radiation (Allen et al, 1964).

A pair of home-made copper-constantan thermopile flux plates calibrated against a commercially manufactured (Middleton and Co.) instrument, in the laboratory, were used to measure the flux of soil heat. These were placed at a depth of about 5 cm in two locations about 5 metres apart adjacent to the tower site (see figure 4.2), and were connected in series thereby giving an average sensitivity of 1 mv/0.07 ly min^{-1}. These signals were recorded on a Speedomax H continuous recorder at a chart speed of 5.08 cm per hour.

A soil heat flux plate is subject to considerably different radiative, thermal and water-conducting properties from the soil around it, and this produces a certain amount of flux divergence (Tanner and

Fuchs, 1968). However, since the soil heat component accounts for only about 5 percent of net radiation entering a forest, it was assumed that the net effect of any flux divergence error on evapotranspiration rates would be minimal. As a matter of fact the rather shallow depths of the flux plates already precluded any significant flux divergence.

(d) Aerodynamic Resistance (r_a)

By definition the aerodynamic resistance (r_a) is the resistance to molecular and turbulent diffusion of water vapor between leaf surfaces and the air above the canopy at a reference height (Robins, 1974). It is given by:

$$r_a = \frac{c_s' - c}{E} \qquad (4.3)$$

where

c_s' = vapor concentration at the leaf surface

c = vapor concentration at the reference height

E = the vapor flux.

However, since the vapor flux is a parameter which aerodynamic resistance is being used to evaluate, the expression given above cannot be used to calculate r_a. Therefore, an alternative method must be adopted. Monteith (1965) suggested that an estimate of r_a can be derived from the following relationship:

$$r_a = \frac{1}{K^2 u} \left[\ln \left(\frac{Z - d}{Zo} \right) \right]^2 \qquad (4.4)$$

where

K = Von Karman's constant

d = zero plane displacement

Zo = roughness length

u = wind speed at reference height Z.

If it is assumed that the diffusion coefficients of momentum and water vapor are equal, then equations (4.3) and (4.4) are expressions of the same resistance, which in turn is inversely related to wind speed. The problem with equation (4.4) is deriving representative estimates of d and Zo, which vary depending on canopy characteristics and stability conditions. Estimates of these parameters are available for other vegetation types (Thom et al 1975, Hicks et al 1974), but not for a hardwood forest.

In an independent study (Singh, 1976) of the canopy under consideration, wind profile measurements, at three levels above the canopy were made between August 19 and 28, 1975. Selected profiles for periods when stability conditions were assumed to be near-neutral were used to solve for d and Zo from the logarithmic wind profile equation. The method of solution was graphical, and the results gave values of $d = 0.83(H)$ and $Zo = 0.072(H)$, where H is the mean canopy height.

Equation (4.4), however, strictly describes the resistance to the transfer of momentum between wind and the canopy in conditions of neutral stability. Stewart and Thom (1973) mentioned that a stability correction factor should be applied to r_a for conditions other than neutral. Also Chamberlain (1966) argued that a further quantity should be added to r_a to account for the incomplete analogy between the transfer of mass and momentum at rough surfaces. Szeicz et al (1969) however maintained that in climatological and hydrological investigations restricted to temperate climates, buoyancy corrections can be safely neglected. In this study therefore r_a is calculated using equation (4.4), with the values of d and Zo equal to those quoted above.

(e) Canopy Resistance (r_c)

When leaves transpire, water evaporates from cell walls and escapes to the atmosphere by diffusing into substomatal cavities, through stomatal pores and finally through the leaf boundary layer into the free atmosphere (Monteith, 1973). The resistance of stomatal pores depends on the geometry, size and spacing of the pores, on associated anatomical features, and on environmental conditions (Meidner and Mansfield 1968). The canopy resistance of a plant cover is composed mainly of the stomatal resistance of all the leaves in parallel.

Several methods can be used to estimate the surface or canopy resistance. Szeicz and Long (1969) have suggested several meteorological techniques, all of which yield satisfactory results. The problem with these methods however is that they not only require a prior estimate of the transpiration rate (LE_d) and the potential evaporation rate (LE_w) but also either surface or profile measurements of temperature, humidity, wind speed, and in one case the net available energy. Since the primary use of the canopy resistance in this thesis is to derive an estimate of the latent heat flux, the applicability of these methods is precluded.

An alternative method, which is utilized in the present research, makes use of the fact that when evaporation from the soil is negligible, the surface or canopy resistance of a plant cover is estimated from the compound resistance of all its leaves in parallel. In other words, if the effective leaf area index of the vegetation (LAI) and the mean stomatal resistance of a single leaf of the same species (r_s) is known, then the following relationship can be used to calculate the canopy resistance (r_c):

$$r_c = r_s/LAI \quad \text{secs cm}^{-1} \tag{4.5}$$

The difficulty in using equation (4.5) arises from the fact that in a fully developed canopy, the lower leaves may not be illuminated well enough to fully open their stomates, and as a result the leaf area index effective in transpiration (LAI_{eff}) may always be less than the total (LAI). Also, deriving hourly measurements of the parameter r_s, especially for tall vegetation whose leaves are not readily accessible, can be rather arduous. As will be seen later however (see section 4.7 and chapter 5) both these problems can be overcome.

There are basically two methods, namely, indirect and direct, that can be used to calculate the value of a single leaf resistance (r_s). Indirect methods are basically attempts at expressing leaf resistance as a function of some forcing parameter such as solar radiation.

Gaastra (1959) has shown that the total leaf resistance (r_s) of well watered plants can be derived from the following empirical relationship, on the assumption that there is a constant relationship between solar radiation (Q + q) and the bands to which stomates respond:

$$r_s = r_{sm} + \left[A/\left\{(Q + q) + B\right\}\right] \tag{4.6}$$

where r_{sm} is minimum leaf resistance, and A and B are constants.

Szeicz et al (1973) working in terms of conductances found that the response of sorghum leaves to solar radiation (Q + q) can be given by:

$$(K - K_c) = (Km - K_c)\left\{1 - e^{-\alpha(Q + q)}\right\} \tag{4.7}$$

where K_c is the cuticular conductance, and k and km are the epidermal conductances (stomatal and cuticular) in normal and in maximum irradiance, e is the base of natural logarithm and α is a constant.

Another model, that is developed for the vegetal cover under consideration, namely a mixed hardwood forest, and that is suited to the aims of this research will be discussed later (see chapter 5).

Direct measurements of single-leaf resistance, either in the field or in the laboratory are also possible. Barrs (1968) suggests a number of techniques that can be used to measure stomatal aperture, but not necessarily stomatal resistance. Kanemasu _et al_ (1969) however, following the lead of van Bavel _et al_ (1965), have designed a stomatal diffusion porometer that permits direct field measurements of leaf resistances. The principle of operation in porometry is simply recording the time required for a given quantity of water vapor to diffuse from the stomatal walls into a sensor cup, when placed over a leaf, and become absorbed by a humidity sensing element housed inside the cup.

In the present study, a Diffusive Resistance Meter, manufactured by LAMBDA Instruments Corporation (Brochure A-174) was used. Since the canopy type being sampled is hardwood deciduous, the horizontal type sensor (Model LI 15S) with the accompanying meter (Model LI 60) was obtained. The sensor was calibrated in the laboratory using the specially-prepared resistance plate at a room temperature of 25°C. For field use, the meter was fitted with straps so that it could be hung around the neck of the experimenter so as to facilitate climbing of the tower and reading of the meter while in use, and to ensure safety of the equipment. An electrician's climbing belt was also used during measurements to ensure safety of the experimenter.

Preliminary test runs were made hourly at the beginning of the 1974 growing season to determine the final frequency of measurements. Sampling periods, that included getting up and down the tower, usually lasted from 15 to 30 minutes. The task of taking hourly leaf resistance measurements daily at heights of 12 metres and over proved to be rather arduous and dangerous. Also the test results showed that bi-hourly sampling periods for selected days were adequate for the purpose at hand. As a result, sampling periods were restricted to bi-hourly measurements twice weekly for daylight periods, and at least once daily, usually between 10.00 a.m. and 2 p.m. EST., for the other days.

Measurements of leaf resistances as described above were taken at two heights within the canopy, for both maple and beech trees from the two wooden platforms mentioned earlier. One leaf from each species was sampled during each measurement. The sampled leaf or leaves during the course of the day were always from the same branch. This was both because of accessibility and the desire to keep measurements as uniform as possible. Besides, preliminary measurements taken from different sized leaves from different branches for both species did not show any significant dissimilarity. The higher level measurements were taken at a height of about 17 metres which is close to the canopy top. Because of the oval outline of the canopy (see figure 4.5) these leaves were fully exposed to the sky and were thus fully illuminated during bright sunshine. The lower level measurements were taken at a height of about 14 metres. This height corresponded to the shaded lower portion of the canopy. It was found that shaded leaves had higher resistance values than fully illuminated leaves, as might be expected in view of varying light intensities within the canopy.

Since evapotranspiration was to be calculated for hourly periods, it was necessary to devise some method for filling in for the times when stomatal resistances were not measured. For this purpose an empirical model was developed using the measured values and relating them to both net and global radiation. The discussion of this model is reserved for the next chapter since its nature and reliability form an integral and important part of the research project, which needs to be discussed separately.

4.5 Runoff Measurements

The preceding section has outlined the methods used for measuring evapotranspiration. As a means of checking the accuracy of these estimates, in terms of the water balance equation (3.1), surface runoff measurements are needed. For this testing, a tributary of the West Creek basin (see figure 4.1) which is located within the experimental area, was chosen.

Discharge was measured at the mouth of the tributary. The method used for measuring the volume of streamflow was a 90 degree V-notch weir of the kind suggested by Gibson (1957). During the first season of measurement a dam consisting of 2.54 cm thick wood was used. The V-notch itself however, consisted of 1.59 mm thick ply-wood. This method was found to be inadequate, however, in that leakage at the sides and bottom was common, especially following periods of intense rainfall.

During the second experimental season, therefore, a concrete dam, reinforced with steel rods and wire mesh, was constructed (see figure 4.7). To prevent leakage at the sides, the dam was extended well into the channel

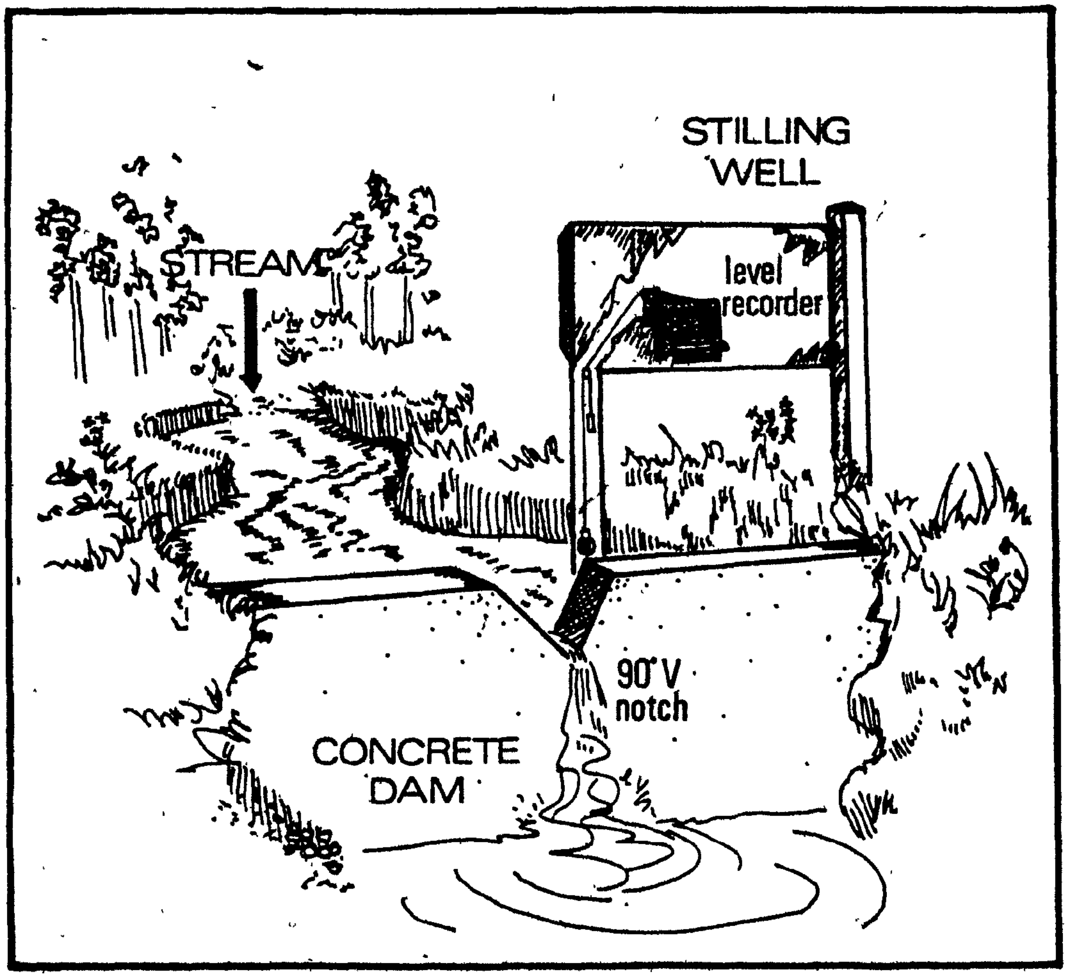

Figure 4.7 Weir design for runoff measurement

embankments. Leakage at the base of the structure was prevented by lining the channel bottom with a heavy plastic sheet that extended about 3 metres upstream. The actual V-notch itself was made of a 12 mm thick iron sheet that was casted into the concrete structure. The lip of the V-notch was ground to a fine sharp edge. The base of the V-notch was about 40 cm above the bottom of the channel.

According to Gibson (1957) the volume of flow for a 90 degree V-notch weir is

$$Q = 0.305\ H^{2.5} \qquad (4.8)$$

where

Q = streamflow in cubic feet per minute

H = height in inches of the water head above the vertex of the V-notch

For daily periods equation (4.8) can be written as:

$$Q = 439\ H^{2.5} \text{ cubic feet per day} \qquad (4.9)$$

From (4.9) it follows that hourly discharge rates can be found from:

$$Q = 18.29\ H^{2.5} \text{ cubic feet per hour} \qquad (4.10)$$

For calculating the water balance the depth equivalent of discharge was obtained by dividing the volume of discharge by the total area of the watershed as:

$$Q = (18.29\ H^{2.5}/A)\ 304.8 \text{ mm/hr.} \qquad (4.11)$$

where A is basin area (510,950 sq. metres) that was calculated by means of planimetry and H is in the same units as in equation (4.10).

The height of the head of water was measured at a point upstream by means of a portable liquid level recorder (Belfort Instrument Company, Catalog No. 5-FW-1), that was positioned about 2 metres from the V-notch.

The recorder was fitted with a gear wheel that made the drum, on which the chart paper was mounted, complete one revolution every 8 days. The tracing pen had a range of zero to 10 inches. The latter height was never exceeded during the experiment. The recorder was housed in a pad-locked wooden box placed above a stilling well which had a height of 1-metre and a diameter of 30 cm. The float mechanism of the recorder was then housed into the stilling well.

The height on the recording drum was calibrated against a yard-rule divided into sixteenths of inches. To ensure the proper calibration, height measurements were checked at least twice weekly, and after each rain event in that the calibration changed with changes in the stream bed level following sediment deposit. Both stilling well and stream bed were cleared occasionally, after a significant amount of sediment build-up.

By means of this technique a continuous record of discharge for the sample basin was available for all of the 1975 and part of the 1974 growing seasons. The data obtained were in close agreement with previous measurements (Rouse, 1965) of a similar nature.

4.6 Soil Moisture Measurements

The need for soil moisture measurements is desirable for two purposes. Firstly it serves as a check on evapotranspiration estimates through the use of equation (3.1). Secondly it provides a measure that can be used to gauge stomatal behavior (see chapter 5).

In view of the rocky nature of the soil in the experimental area, soil content in depth units was calculated by means of a gravimetric technique suggested by Hoover (1962).

To utilize this method, the percent moisture by weight and the bulk density of the soil has to be known. The product of these two quantities gives the percent of soil volume occupied by water. Percent moisture by volume multiplied by the depth of the layer considered, divided by 100 gives the depth of soil moisture in the layer. The total moisture content in the soil profile in unit depth of water is then found by summing the depths of the individual layers.

During the first experimental season (1974) soil moisture measurements were taken at 3 different sites (see figure 4.2). Each site consisted of a plot of about one metre square, and successive samplings were taken in an organized manner at distances of approximately 30 cm. Soil moisture content was measured at successive depths of 15 cm down to a total depth of 60 cm which is close to the total depth of the soil profile (Wilson, 1968). Cores of soil samples were extracted by means of an auger, and after each boring the holes were refilled so as to avoid future uneven water accumulation.

Measurements were taken on a regular basis, namely every seven days, except when rain was falling or when the soil did not have sufficient time to drain out completely. On these exceptional occasions, measurements were taken on the next day, and the following week the initial 7-day cycle was returned to.

Soil samples were placed in tightly sealed and labelled plastic bags in the field and transferred the same day to the laboratory where they were placed in tin-foil containers then weighed and dried overnight in an oven set at about 105°C. The following morning the dry soil samples were then weighed. From these measurements the percent moisture by weight was derived.

The values of soil density used were the same as those which had been found by Wilson (1968) at adjacent and similar sites. Correction for rock content was also taken into consideration. The values of soil density used for the different sites at different depths are given in table 4.1.

TABLE 4.1 Soil Density (gm cm^{-3}) at Different Sites and Depths

Soil Layer cm	Density gm cm^{-3}		
	Site 1	Site 2	Site 3
0 - 15	0.50	0.52	0.44
15 - 30	0.57	0.60	0.45
30 - 45	0.65	0.59	0.43
45 - 60	0.68	0.77	0.47

During the second year (1975) only 2 sites (1 and 2) were sampled since data from the previous year (1974) showed that these gave a mean value not significantly different from that utilizing three sites (see figures 5.3 and 5.8). The net change in soil moisture (Δ sm) for evaluating the water balance (see equation 3.1) was simply taken as the difference of soil moisture content between successive time periods. The results obtained by using these methods gave values of soil moisture content that corresponded to those measured by Wilson (1968).

4.7 Supporting Measurements

The preceding sections have outlined the methods and procedures used to measure or estimate the various components of the water balance. As mentioned earlier however (see section 4.4) hourly sequences of the canopy resistance (r_c) are required to derive the evapotranspiration rates. Since these were not measured on a regular hourly basis, a method had to be developed to estimate them. In the next chapter a model that predicts mean leaf resistance for dry conditions from net or global radiation will be presented. Equation (4.5) which requires a measure of leaf area index is then used to compute the canopy resistance. For wet conditions however a technique that shows the presence of water on the leaves is required. To get an overall picture of the canopy resistance then further measurements, namely global radiation, leaf area index and leaf wetness are needed.

(a) Global Radiation

Global radiation was measured at a height of about 1.5 metres above the forest, using a Lintronic Dome solarmeter (Serial No. 5540). This instrument is sensitive to short-wave radiation in the range of 0.3 - 3.5μm. The sensor consists of an 80-junction thermopile plate, without temperature compensation and housed in a frosted glass dome. The instrument was calibrated against another commercially calibrated brand (Talley Sol-a-meter), and it showed a sensitivity of 22.8 mv/ly min^{-1}.

The instrument was positioned at the top of the mast, and was free of obstacle effects. Signals were continuously recorded during the daylight period on an I.C.A. model 400 strip chart recorder at a speed of 3.8 cm. per hour. Mean hourly values were then derived using the same method as that used for net radiation.

(b) Leaf Area Index

Since the resistance to the vapor flux over the forest is treated as being from single leaves in parallel a measure of leaf area index is desireable. This index is the ratio of total leaf area to unit ground area. Several methods can be used to estimate the leaf area index (Chang, 1974). For the purposes of this research however, the following simple and convenient method was adopted.

Four collection wooden boxes, each one metre square in area and 15 cm deep and underlain by a heavy plastic material, were randomly placed around the tower site (see figure 4.2). These were anchored to the forest floor by means of wooden spikes. The boxes were installed in September 1974 towards the end of the first experimental season, which was just before leaf fall had started. By the first week of November, 1974, the trees had completely shed their leaves. The contents of each box were then emptied into labelled plastic bags and taken to the laboratory for analysis.

Firstly, samples of 30 leaves were randomly selected from each bag. These were then pasted unto sheets of blank white paper and their images were xeroxed. The photocopied images were then planimetered to give the total area (AS) of each sample. These sampled leaves were then oven-dried overnight at 105°C and subsequently weighed to give their dry weight (WS).

The remaining leaves in each bag were also oven dried to give their dry weight (WR), care being taken to remove twigs, nuts and other extraneous material.

Then the area of the remaining leaves in the boxes (AR) were derived by means of the following relationship:

$$\frac{AS}{WS} = \frac{AR}{WR} \qquad (4.12)$$

where

AS = area of sample square cm

WS = weight of sample gm

AR = area of remaining leaves (unknown) square cm

WR = weight of remaining leaves gm.

Equation (4.12) was then transformed to solve for AR, which was then added to AS, to give the total area of all leaves in each box. To finally derive the leaf area index (L.A.I.), the sum of AS and AR was divided by the area of each box (10,000 square cm). The following table (table 4.2) gives a summary of the various measurements obtained by the method just described.

TABLE 4.2 <u>Leaf Area Index Calculations</u>

Box No.	Area of Sample (AS) sq. cm	Weight of Sample (WS) gm	Weight of Remaining Leaves (WR) gm	Area of Remaining leaves (AR) sq. cm	Leaf Area Index L.A.I.
1	1089.33	4.7	212.49	50283.47	5.03
2	1427.82	5.5	188.43	50344.93	5.03
3	1455.13	5.9	209.80	51744.42	5.17
4	1406.74	5.1	184.01	50755.17	5.08

The table shows that the values of leaf area index did not vary considerably among boxes. The range of values was 5.03 - 5.17. The mean value of the four boxes (5.08) was therefore taken as the final value of leaf area index for the whole canopy.

With this method no attempt was made to differentiate between beech and maple species since their resistances were similar (see chapter 5). Besides, working on the premise that maple leaves were greater in area than beech leaves, but were fewer in number, a method whereby the area of beech and maple leaves were calculated separately, did not give a final leaf area index that was very different from the values presented above. The derivation of the values in table 4.2 are therefore based on the assumptions that unit area of beech leaf has the same dry weight as unit area of maple leaf and that the sample ratio of maple to beech leaves represented the sample ratio for each box as a whole. Also it was assumed that the leaves collected by each box fell from unit area within the canopy.

(c) Leaf Wetness

Because one of the primary aims of this research is to compare evapotranspiration rates for an unwetted as opposed to a wetted canopy some method that differentiates between these two states is necessary. An objective method is provided through the use of a home-made moisture sensor that was designed by St. Laurent (1973) (see figure 4.8). The instrument is an excellent device for detecting the presence of water on a leaf surface. The principle of operation is that whenever there is moisture present between the pair of probe tips consisting of high-conductance wire and connected to a pair of resistors, a low-level positive voltage passes through the two resistors to the gate of a Silicon controlled rectifier (SCR), which when fired triggers an alarm or light signal.

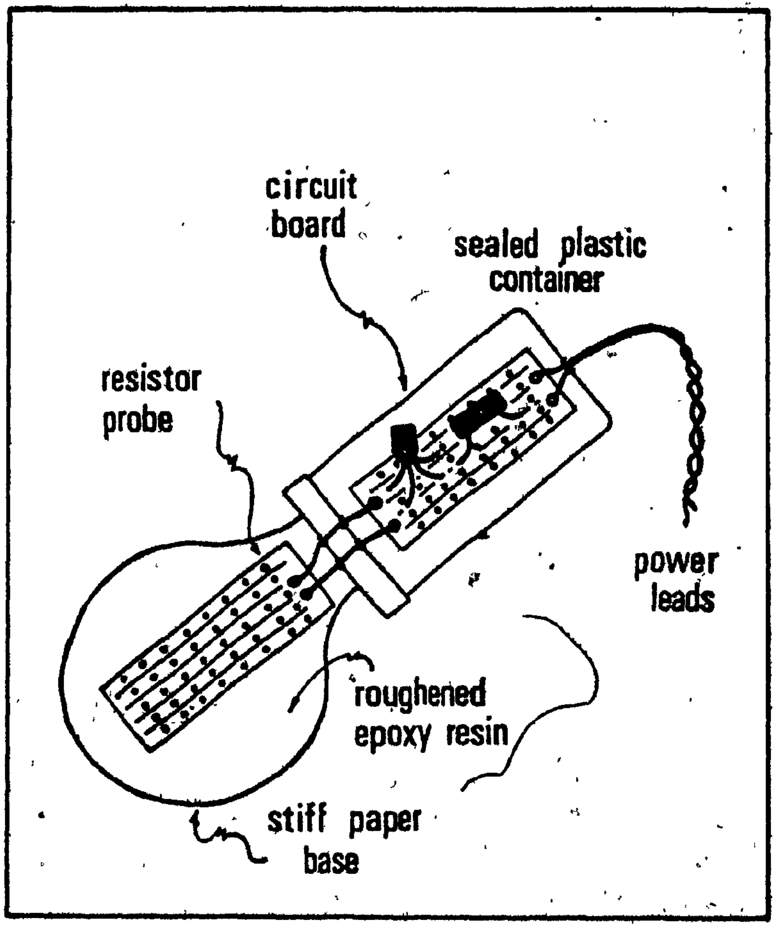

Figure 4.8 Moisture sensor

To avoid having to reset the sensor everytime the SCR was fired, stepped-down AC voltage (12 volt) was used, utilizing a miniature transformer. The power leads were then hitched to a light bulb fitted to a rectifier to obtain DC output. This in turn was connected to an I.C.A. model 400 recorder with a chart speed of 2.54 cm/hour, via a voltage divider, so that when the sensor was triggered a deflection was recorded.

To obtain data for the needs of the experiment two sensors were positioned at heights corresponding to the levels at which leaf resistances were measured. These were connected in series, so that when the canopy was completely wetted and both sensors were fired a full scale deflection of 60 mv was recorded. When only one of the sensors was fired, as was the case when the top of the canopy dried before the shaded portion or when only the leaves at the top were wetted following a light drizzle, a deflection of 30 mv was recorded.

The circuit board of the sensor itself was placed into a tightly sealed plastic jar, while the protruding probes were connected to a tortuous pair of fine copper wires that sat on a paper base that resembled a leaf (see figure 4.8). The pair of wires was held in place and kept separate by epoxy cement. Visual checks showed that the paper base with a layering of roughened expoxy cement dried out at approximately the same rate as a leaf. The moisture sensor therefore, provided a good estimate of the duration of wetness for both levels of the canopy.

CHAPTER 5

Predictive Model for Canopy Resistance

5.1 Background

The previous chapter (section 4.4(e)) has described the methods and principles that can be used to derive the canopy resistance (r_c). For reasons given earlier the leaf resistance - leaf area index concept, as stated in equation (4.5), has been adopted. Also direct measurements of single-leaf resistance (r_s) through the use of porometry have been utilized. As mentioned previously (see section 4.4 (e)) however, obtaining hourly measurements of single-leaf resistances for tall vegeatation is an extremely arduous and difficult undertaking. Since hourly measurements could not be taken on a regular basis then, and since one of the aims of the research was to obtain mean hourly evapotranspiration rates, some method had to be devised whereby mean canopy resistance for hourly periods could be derived. In the present chapter, a model that fulfills this need, is presented. The model is based on two of the main physical factors that control stomatal behavior, namely irradiance and soil water availability.

5.2 Response of Stomates to Radiation and Soil Moisture

Leaf resistance to vapor diffusion is basically a function of stomatal opening. The degree and extent of stomatal opening and closing, however is a rather complex phenomenon, because it depends not only on a wide variety of weather conditions, but also on such internal factors as moisture supply and type and age of species. Physically however, it appears to be well established that increases in guard cell turgor, absolute or relative to that of the adjoining epidermal cells, causes opening of the stomatal pore, while decrease in turgor produces closing (Meidner and Mansfield, 1968).

Guard cell movements on the other hand seem to be intricately related to CO_2 concentration in the intercellular spaces of leaves. Below a certain critical value of CO_2 opening is initiated and above a similar limit closing commences. These critical limits vary according to species and lie between 0.01 percent (100 ppm), which is the lower limit for leaves, and 0.03 percent (300 ppm), which is the normal atmospheric concentration (Slatyer, 1967). Furthermore, internal CO_2 concentration is related to and is controlled by such physical factors as light intensity, leaf temperature and moisture stress. In the absence of soil-water deficit, light is probably more important in determining stomatal behavior than any other component of the environment. On a quantum basis, blue light seems to be the most important in this respect (Virgin, 1956). In most plants stomata normally open during the day and close at night, but the speed of response to light varies considerably among species. In some cases the opening is gradual and may take several hours, whereas in others the movement is more rapid, and may be nearly complete in about half an hour.

Opening increases with light intensity up to a saturation value that is often of the same order of magnitude as moderate sunlight (Meidner and Mansfield, 1968). The higher the light intensity, the higher the CO_2 concentration required to produce a given degree of stomatal closure, in that there is a greater consumption of CO_2 in photosynthesis at higher light intensities. The effect of light on stomatal aperture therefore appears to be mediated by the photosynthetic reduction of intercellular CO_2 concentration.

The degree and extent of stomatal opening in response to light is however controlled by soil moisture availability at the roots of the plant. Water deficits can clearly exert a direct effect on stomatal aperture by their effect on relative and absolute turgor levels in guard cells and surrounding cells (Meidner, 1965). Two types of reaction are recognizable, one being a transient change in stomatal aperture as a result of changes in guard cell turgor relative to that of adjacent cells, and the other a longer term change associated with severe stress.

The first type of response is the well-known midday closure which results from water stress caused by extremely high evaporative demands. The second type of response occurs at different values of soil water potential for different species and for plants grown under different environmental conditions (Ehlig and Gardner, 1964). It appears that water deficit, in itself, may not affect stomatal aperture until a critical value is reached, and then, as the water deficit increases, progressive decreases in stomatal aperture occur until almost complete closure exists. Both these responses are also mediated to some degree by internal CO_2 concentration, since reduced photosynthesis caused by water stress lead to increased internal CO_2 levels which in turn influence aperture (Slatyer, 1967).

It appears then, that above a certain critical level of soil moisture, light is the dominant factor that controls stomatal aperture. Beneath this critical level however, the effect of soil moisture availability becomes increasingly important. In a sense then, soil moisture exerts the final control on stomatal aperture.

In the light of the preceding factors, preliminary tests were conducted during the 1974 growing season, in an attempt to formulate a relationship between leaf resistance, irradiance and soil moisture depth. At first, instantaneous values of leaf resistance for individual maple and beech leaves for both levels of the canopy were plotted against net radiation (see figure 5.1). Because of instrumental problems net radiation was the only irradiance parameter measured. The results show that except during periods of midday closure, there exists a well defined inverse relationship between stomatal resistance (r_s) and net radiation (Rn). Also, stomatal response to increasing or decreasing radiation is fairly rapid, attaining maximum opening or complete closure in less than a couple of hours. Note also that except for periods of closure shaded leaves consistently display higher resistance values than sunlit leaves because of uneven illumination.

If the mean resistances of both species ($r_{\bar{s}}$), for upper leaves, are plotted against net radiation above the canopy, for selected days spanning part of the growing season (see figure 5.2), then the scatter of points again show the existence of an inverse non-linear relationship. Stomatal opening is initiated at a net radiation value of about 0.0 ly min^{-1}. Also, the saturation point of lowest resistance or greatest opening, depending on soil moisture depth, is attained at a net radiation value of approximately 0.25 ly min^{-1}, which is equivalent to moderate sunshine.

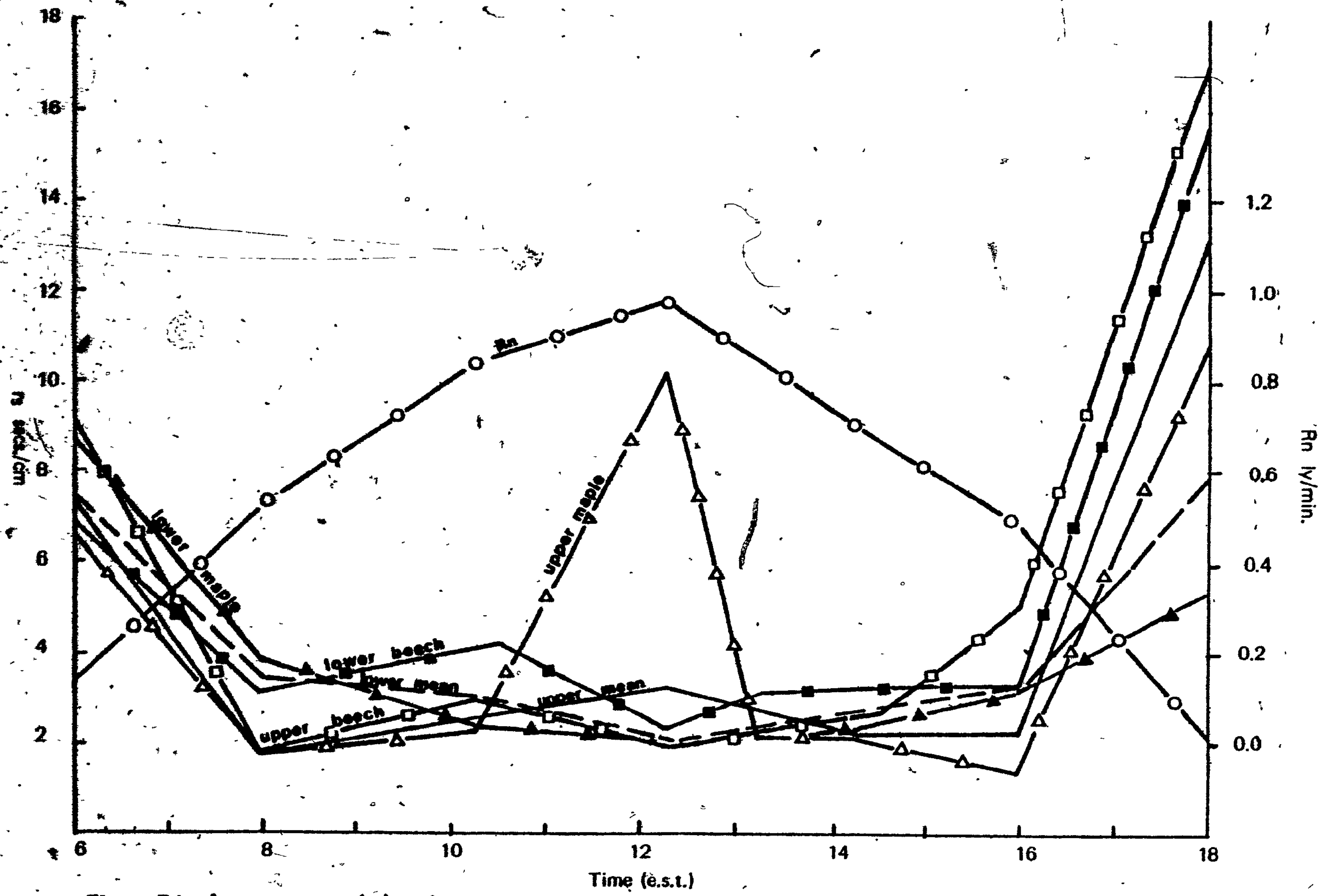

Figure 5.1 Instantaneous (r_s) and mean leaf resistance ($\bar{r}_s$) vs net radiation (Rn), 12th August. 1974.

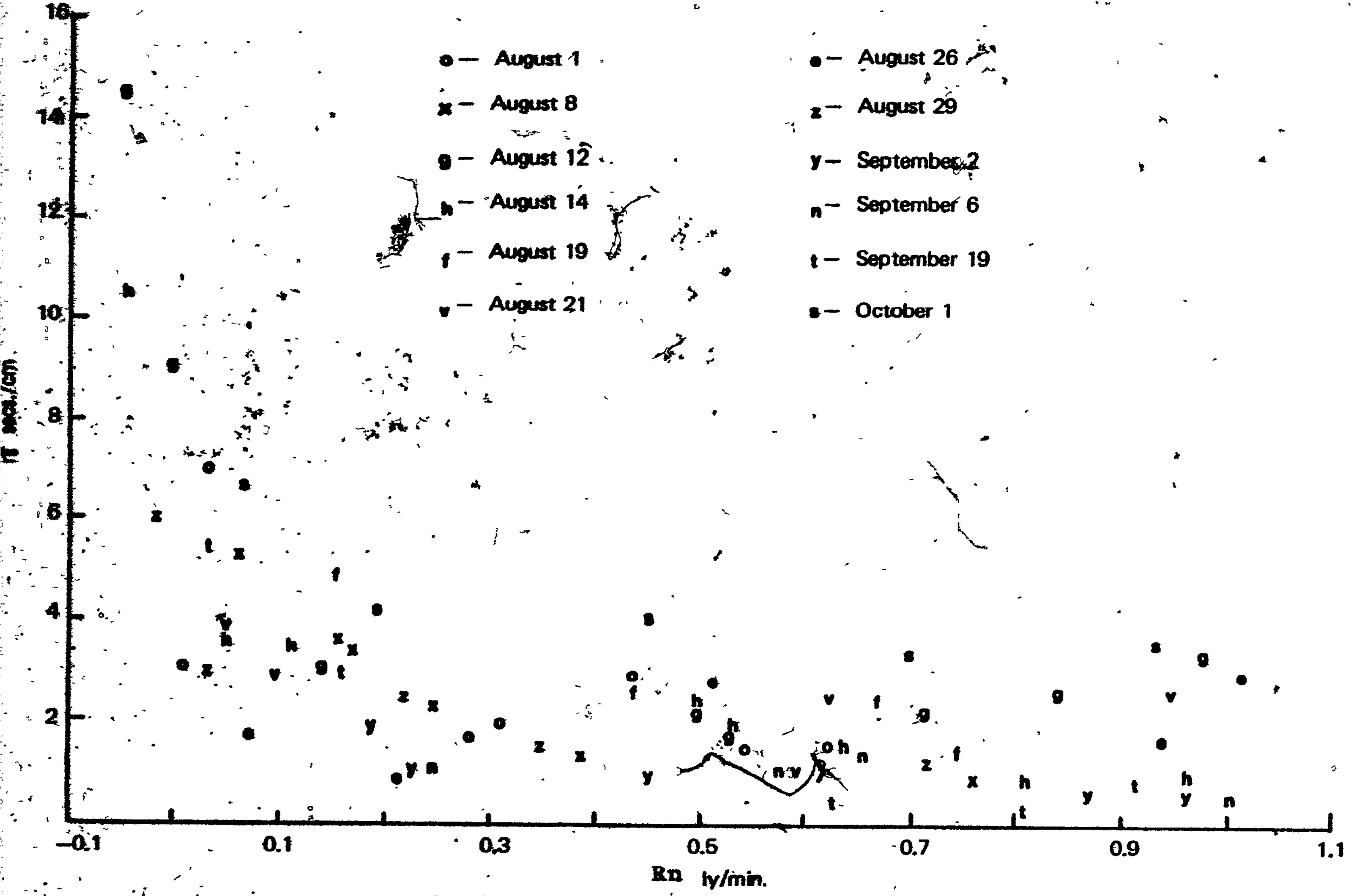

Figure 5.2 Mean stomatal resistance ($\bar{r}_s$) (upper level) vs net radiation (Rn) for selected days, 1974.

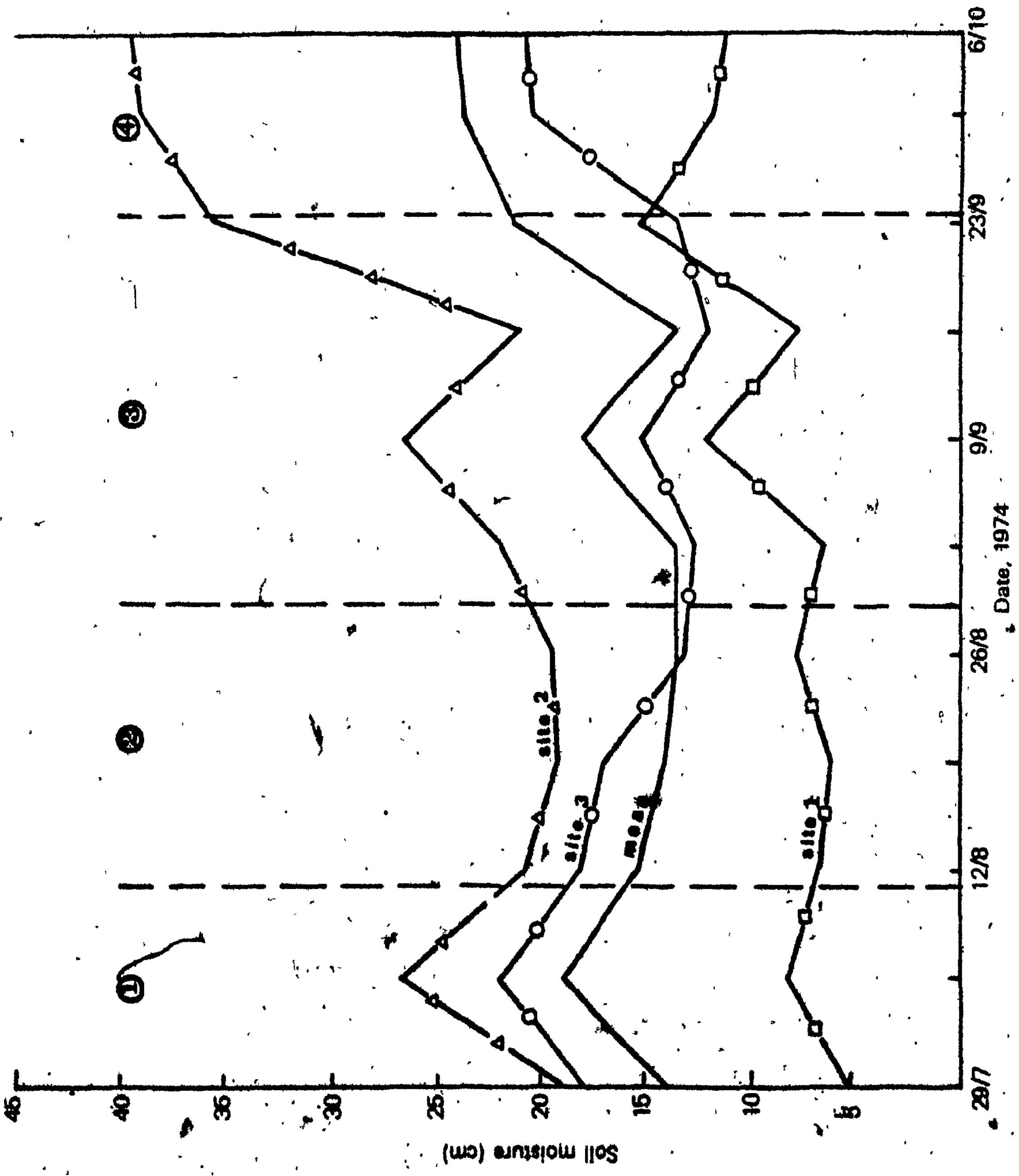

Figure 5.3 Soil moisture depth and periods, 1974.

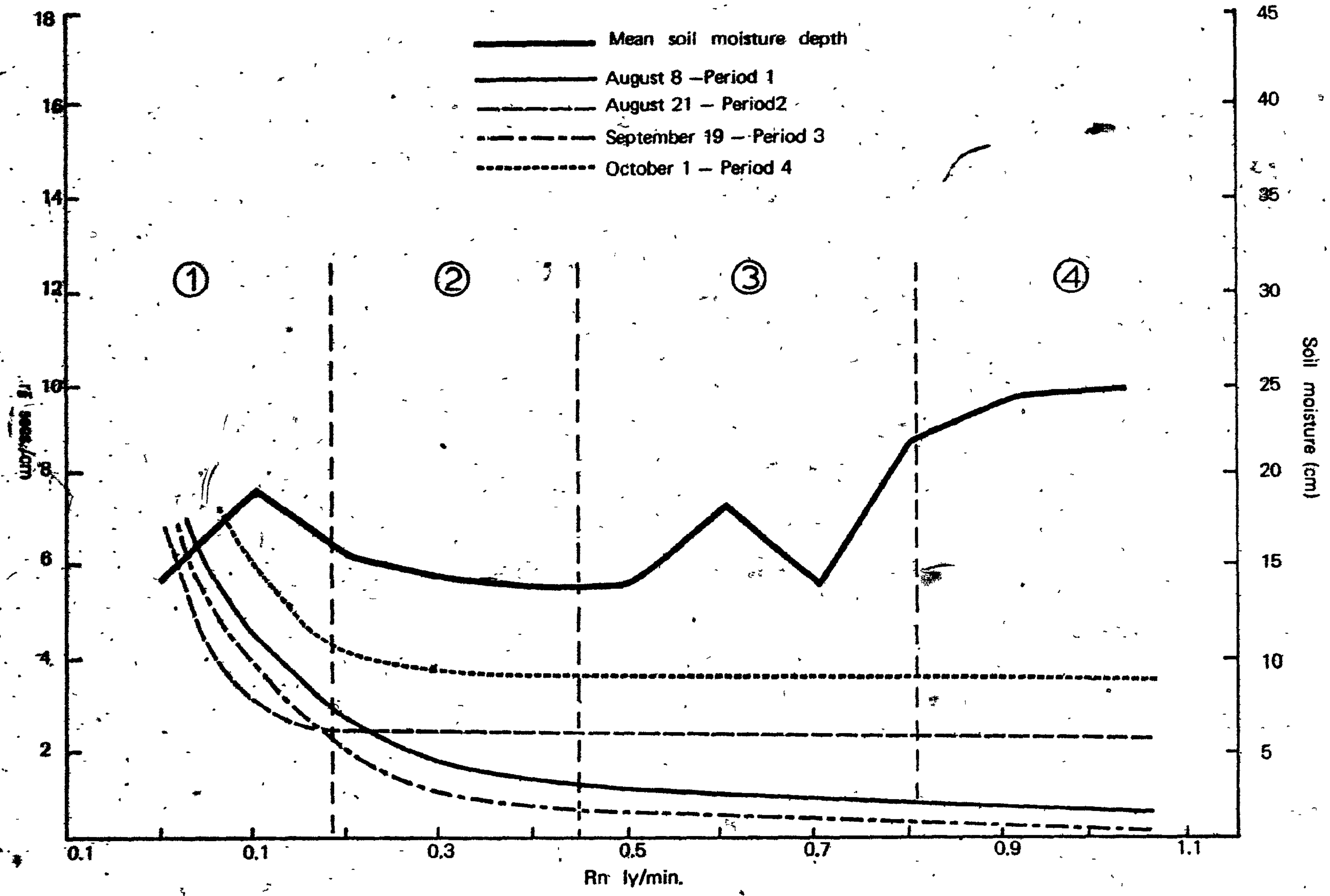

Figure 5.4 Change in r_s with change in mean soil moisture content, 1974.

Furthermore, it can be seen that the approximate level of the minimum resistance, as characterized by the horizontal asymptotes (see figure 5.4) shows marked seasonal changes that are relatable to soil moisture depth conditions (see figure 5.3). It is therefore possible to divide the growing season into separate periods, each of which is characterized by a certain mean value of minimum resistance, which in turn is affected by soil moisture content. Generally minimum stomatal resistance is lowest when soil moisture depth is greatest. An exceptional condition arises during leaf senescence (period 4 in figure 5.4), when wilting of the leaves naturally gives higher resistance values, regardless of soil moisture conditions.

From the foregoing then, it is possible to formulate a model that predicts mean hourly stomatal resistance, for each level of the canopy, based on radiation and soil moisture conditions.

5.3 The Predictive Model

Based on the preliminary tests described in the preceding section, the following empirical model that predicts mean stomatal resistance ($r_{\bar{s}}$) from radiative and soil moisture conditions, is proposed.

$$\mathrm{Ln}\,(Y - C_1) = \mathrm{Ln}\,a + b\,\mathrm{Ln}(X + C_2) \qquad (5.1)$$

where

Y = the mean stomatal resistance ($r_{\bar{s}}$) (secs cm^{-1})

X = net or global radiation above the canopy (ly min^{-1})

C_1 = an estimate of the minimum mean stomatal resistance that can be obtained under a certain range of soil moisture depths (secs cm^{-1})

C_2 = an approximation of the critical level of radiation that initiates stomatal opening (ly min^{-1})

and a and b are constants, and Ln is the base of natural logarithms.

Equation (5.1) is an extension of a linear regression equation. A constant (C_2) is added to the X-axis and another constant (C_1) is subtracted from the y-axis to attain symmetry. Log transformations are then applied to both dependent and independent variables to obtain linearity. The equation (5.1) is essentially empirical since the constants (C_1 and C_2) were determined by trial and error, so as to give the best fit line, as characterized by the highest correlation coefficient (r), and delimitation of their actual values is imprecise.

The value of Y or $r_{\bar{s}}$, the mean stomatal resistance for each level of the canopy, in equation (5.1) was derived by taking the mean conductance of single maple and beech leaves, so as to appropriately weight the leaves that were more actively transpiring. In mathematical terms then, this relationship can be expressed as:

$$r_{\bar{s}} = 1/\left\{ (1/r_{sm} + 1/r_{sb})/2 \right\} \qquad (5.2)$$

where

$r_{\bar{s}}$ = mean stomatal resistance (secs cm^{-1})

r_{sm} = stomatal resistance of a single maple leaf (secs cm^{-1})

r_{sb} = stomatal resistance of a single beech leaf (secs cm^{-1}).

The predicted value of Y or $r_{\bar{s}}$ is then derived from radiation once the regression coefficients have been calculated from equation (5.1), by first exponentiating Y and then adding the appropriate constant (C_2). Mathematically this is expressed as:

$$r_{\bar{s}} = (\exp.\ Y) + C_2 \qquad (5.3)$$

where exp. has natural base, and all other terms are as defined in equations (5.1 and 5.2).

In order to derive a final value of the surface or canopy resistance (r_c) of the vegetation as a whole, which can be used in equation

(3.26) an extension of equation (4.5) is utilized. As indicated earlier (see section 4.4 (e)), the canopy is considered as being bi-level, consisting of fully illuminated and shaded leaves. Because of the near-oval nature of the outline of individual trees (see figure 4.5), it is assumed that 2 units of total leaf area (5.08) are fully illuminated and the remainder (3.08) are shaded. Again, conductances are used so as to weight the resistance of the generally more vigorously transpiring sunlit leaves appropriately. The formulation of the relationship then becomes:

$$r_c = 1/\left\{2(1/r_{\bar{s}u}) + 3.08\ (1/r_{\bar{s}l})\right\} \qquad (5.4)$$

where

r_c = mean canopy resistance (secs cm^{-1})

$r_{\bar{s}u}$ = mean stomatal resistance of sunlit leaves (secs cm^{-1})

$r_{\bar{s}l}$ = mean stomatal resistance of shaded leaves (secs cm^{-1}).

Both $r_{\bar{s}u}$ and $r_{\bar{s}l}$ are derived according to equation (5.2).

5.4 Verification of Model

(a) Estimates from Net and Global Radiation

In equation (5.1), it was mentioned that the mean single-layer leaf resistance ($r_{\bar{s}}$) could be predicted from either net (Rn) or global radiation (Q + q) above the forest. This is understandable in view of the fact that there exists a linear relationship between both parameters, as demonstrated by figure 5.5. Similar results were also found by Davies (1967) and Polavarapu (1970).

As will be seen later, there seems to be a stronger relationship between global radiation, and the radiation bands to which stomates respond than between these bands and net radiation (see tables 5.1 and 5.2).

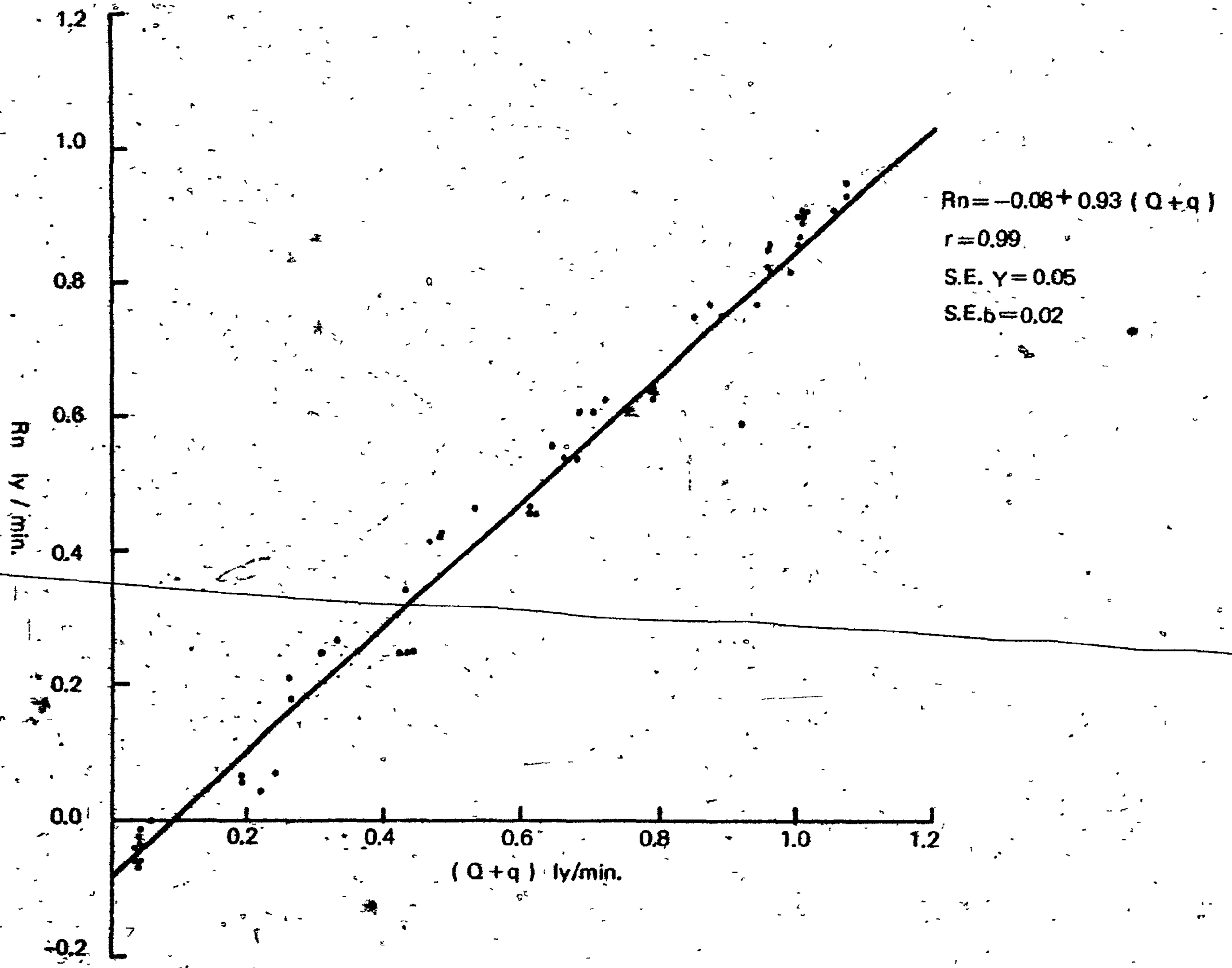

Figure 5.5 Net (Rn) vs global (Q + q) radiation, June 25 - 28, 1975.

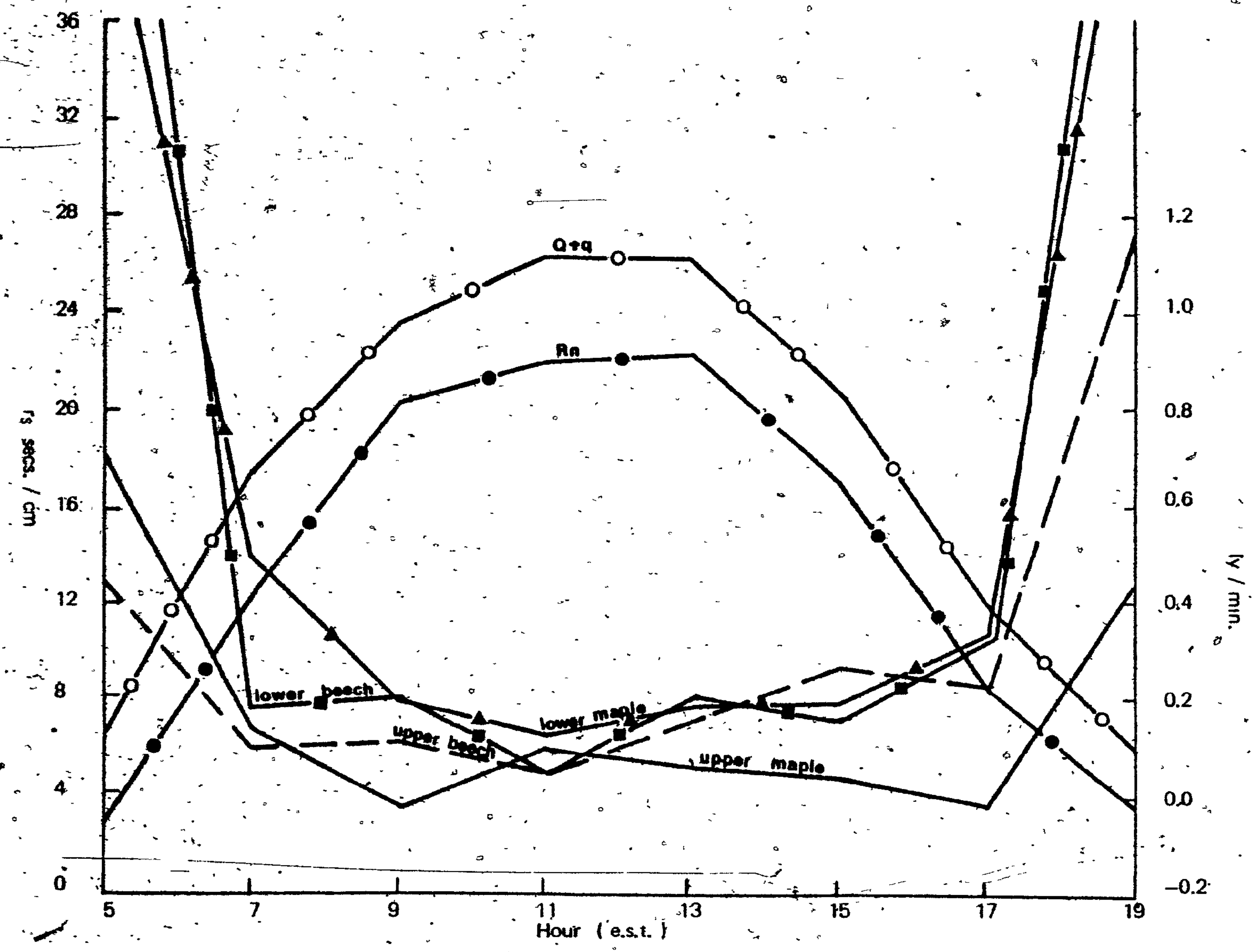

Figure 5.6 Variation of individual leaf resistances with radiation, June 10, 1975.

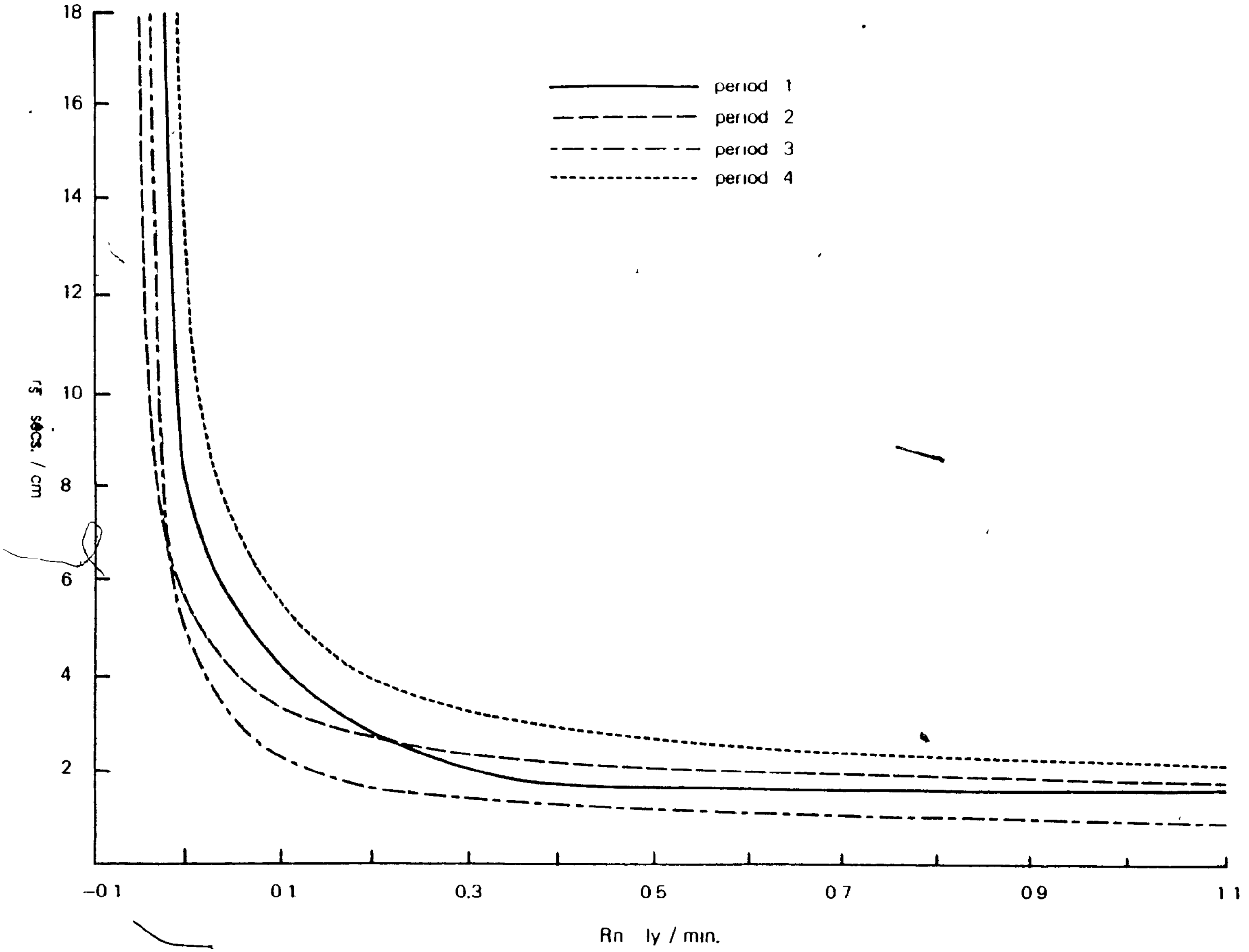

Figure 5.14 Predicted rξ (upper level) vs Rn, 1974.

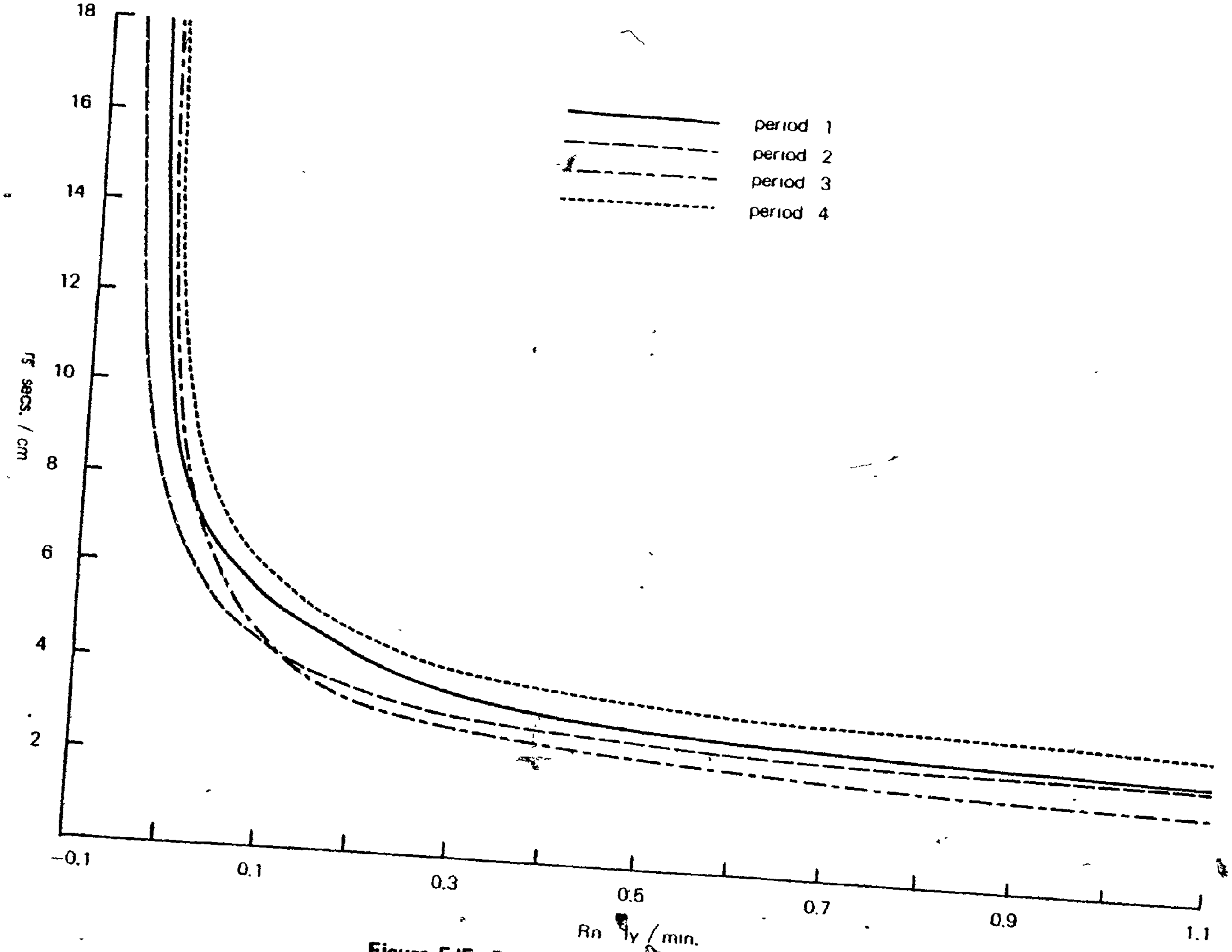

Figure 5.15 Predicted r_s (lower level) vs Rn, 1974.

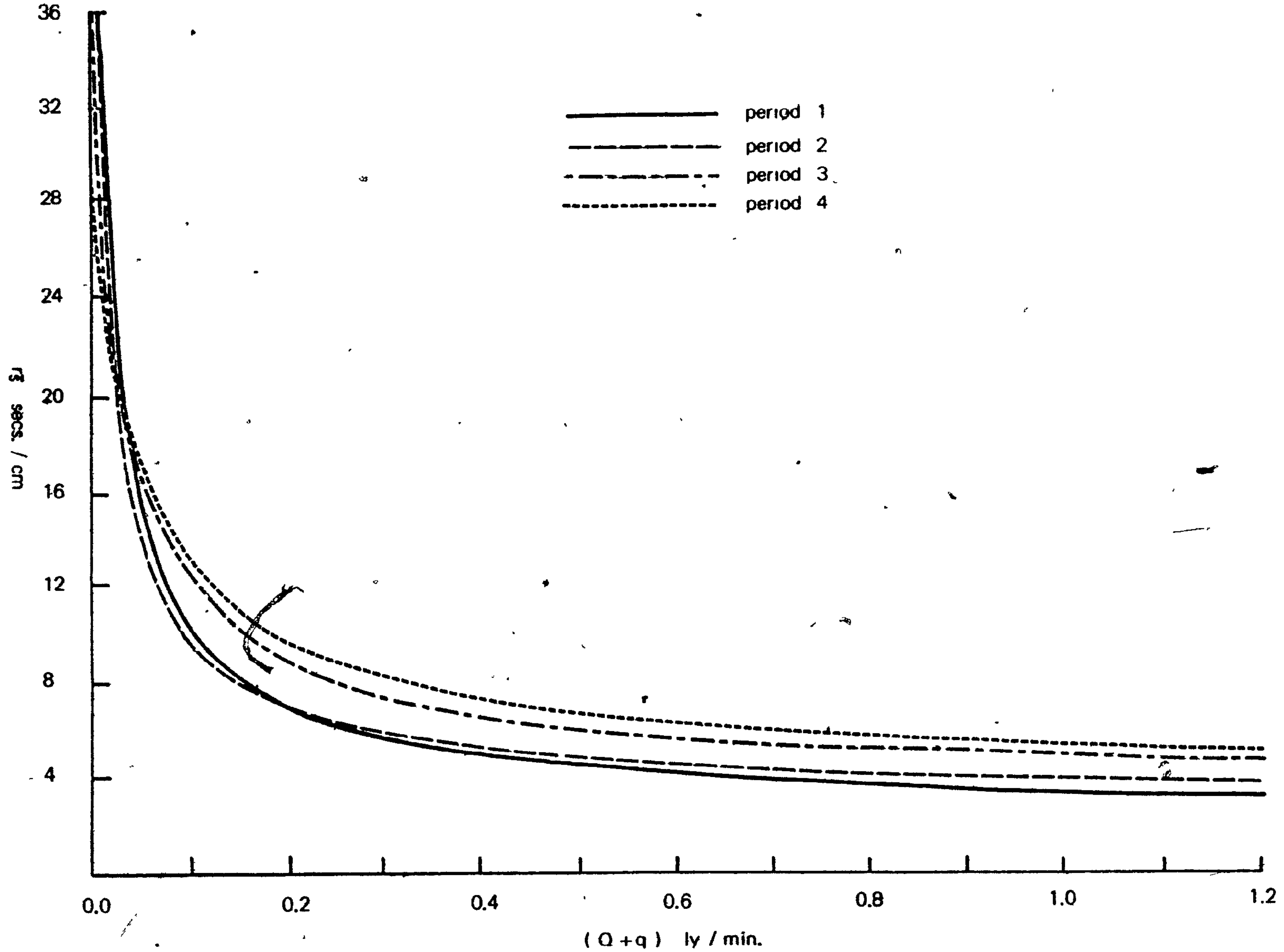

Figure 5.16 Predicted r_ξ (upper level) vs (Q + q), 1975.

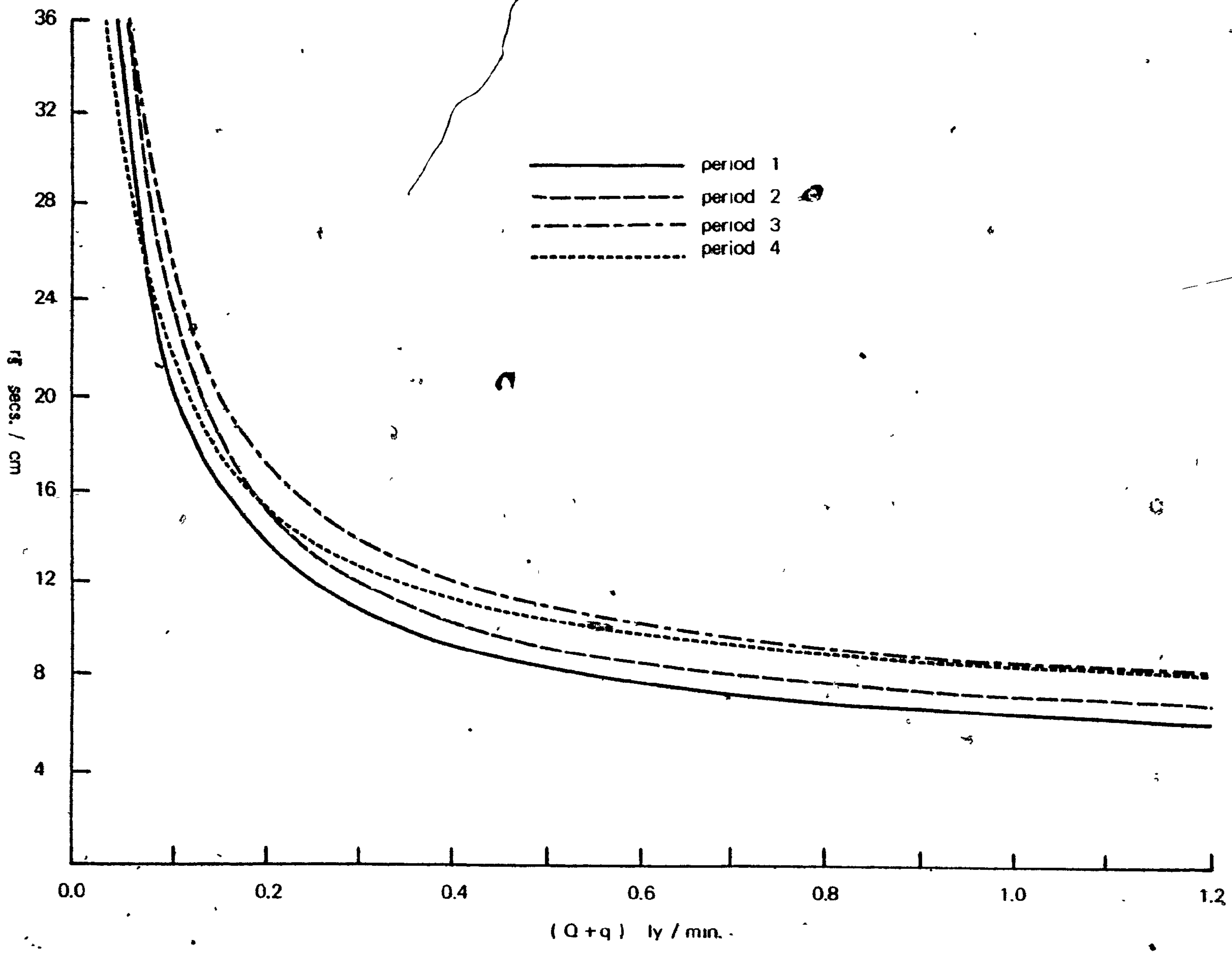

Figure 5.17 Predicted r_s (lower level) vs (Q + q), 1975.

both levels of the canopy are inversely related to soil moisture depth, in that resistances are generally high when soil moisture depth is low. The rather low values of $r_{\bar{s}}$ during period 3 are not only attributable to greater depths of soil moisture, but may also be due to the decreased daily evaporative demand of the atmosphere at this time of year, as caused by shorter daylengths and generally cloudier skies, resulting from more disturbed weather. The obvious anomaly during period 4 is because of the fact that the leaves were wilting because of senescence.

During the exceptionally dry 1975 growing season however the correspondence between mean stomatal resistance ($r_{\bar{s}}$) and soil moisture depth was somewhat anomalous. Apart from the period of senescence (5) there existed another discrepancy in this case, namely period 4. During this period the mean stomatal resistance was slightly higher than for period 3 when in fact soil moisture depth was greater during the former period. A possible explanation for this occurrence is that frost conditions at night were experienced at the beginning of period 4 and this might have caused a certain amount of tissue damage to the stomatal pores. Or it might be that drought conditions may have existed in the memory of the plants and as a result they did not respond accordingly to increased soil moisture conditions (Iljin, 1957). Also this could have in fact been a period of premature senescence as was evident from distinct discolorations on the leaf surfaces.

Soil water availability at the roots of the plants during certain times of day is also critical to stomatal response. It is very common to have the plant suffering from stomatal or hydro-active closure during periods when it is vigorously transpiring. Closure has been found

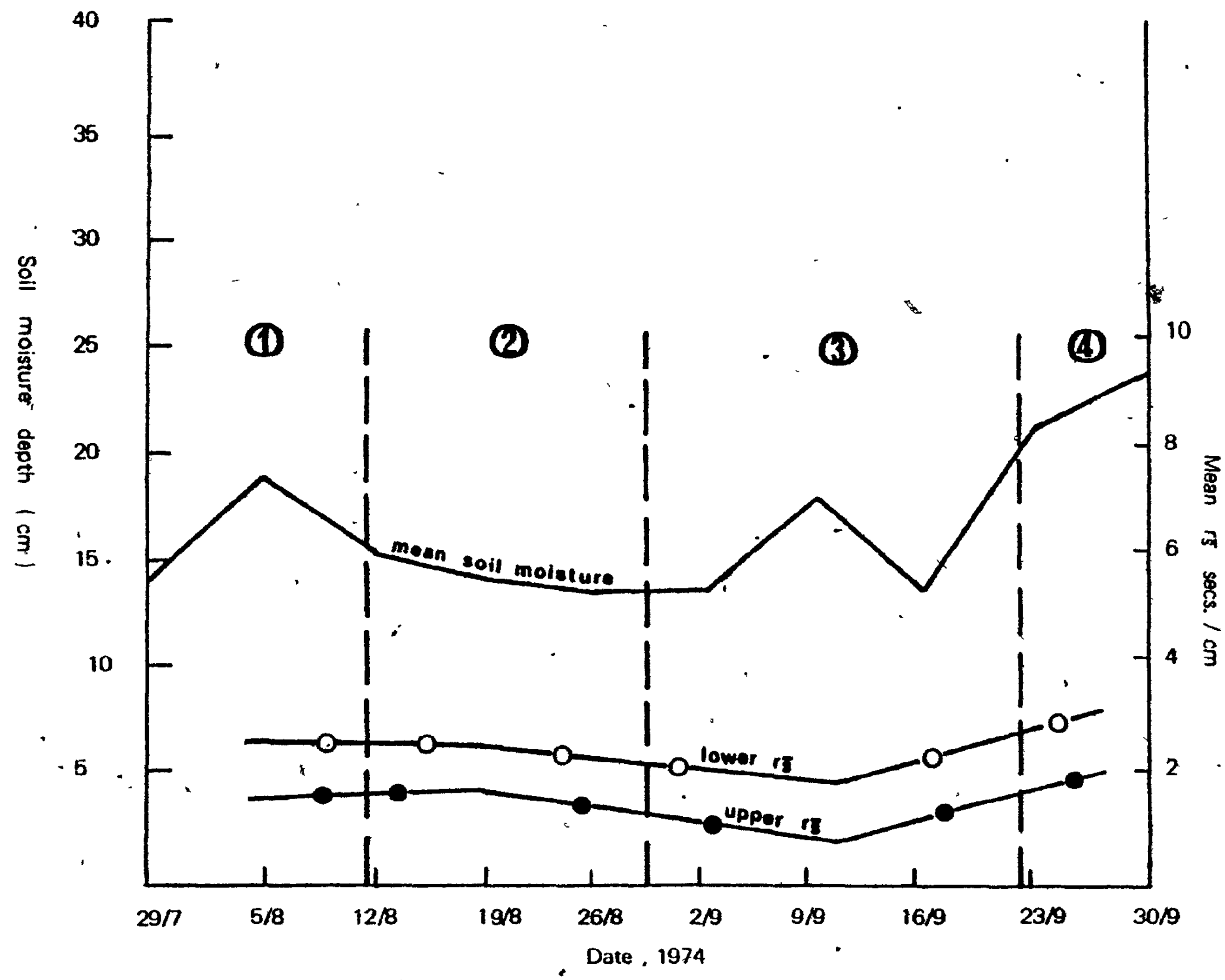

Figure 5.18 Mean stomatal response ($r_{\bar{s}}$) to soil moisture depth, 1974.

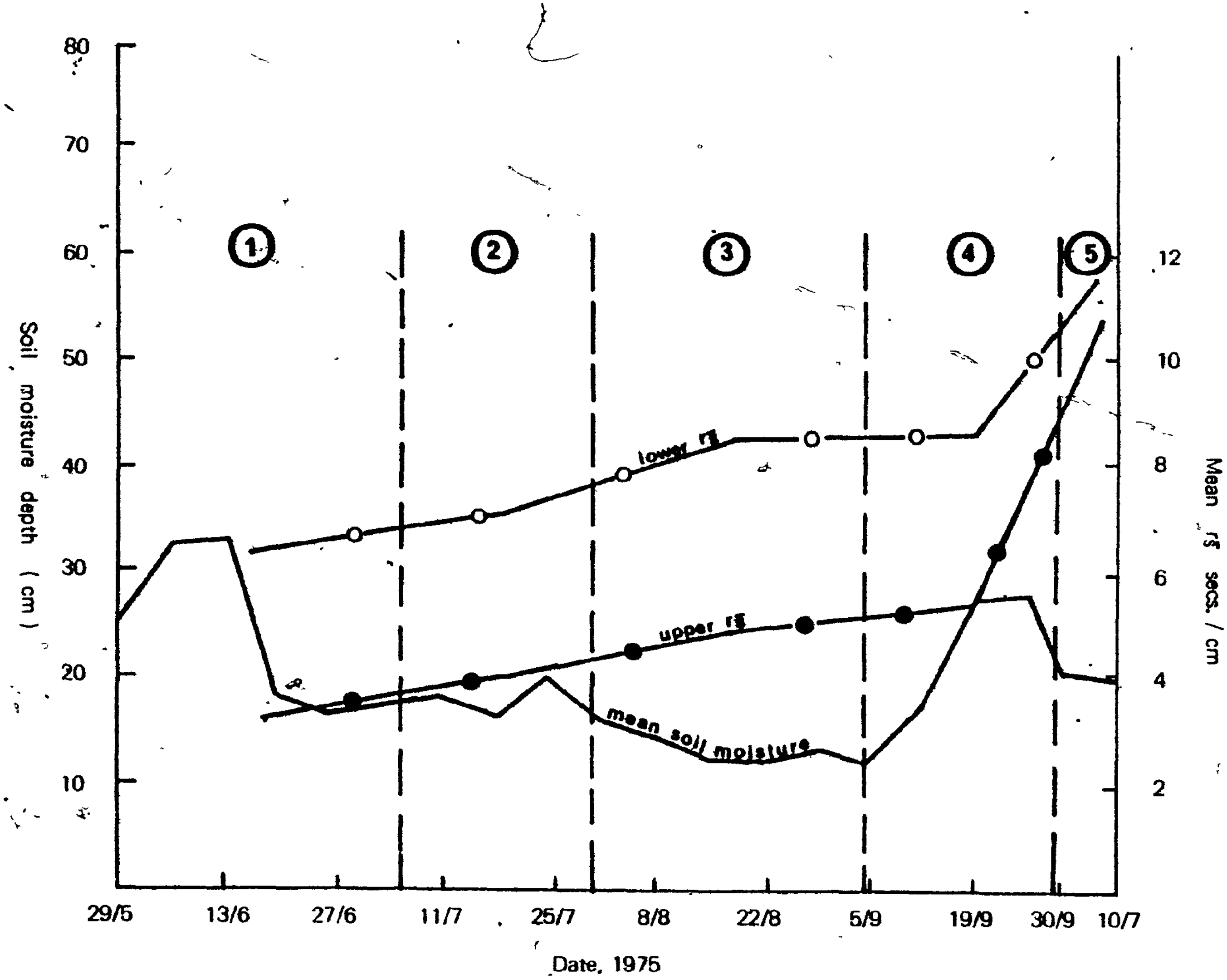

Figure 5.19: Mean stomatal response (r_s) to soil moisture depth, 1975.

to last for an hour, at most, around noon and more often affected upper level maple leaves. This could be due to the greater sensitivity of maple leaves to increasing light stimulation or their greater leaf area index or both. Also as was mentioned earlier, mid-day stomatal closure was more common when soil moisture was nonlimiting and when the plants were actively transpiring. It was also more frequent when the days were longer, since total evaporative demand was greater.

Accounting for this latter phenomenon, namely hydro-active closure, is the major drawback of the model suggested here. Since the wet-bulb depression of the atmosphere increases substantially during periods of closure, it was thought feasible to account for stomatal behavior by monitoring the vapor pressure deficit of the air. This technique however proved inadequate in that the vapor pressure deficit of the air continued to rise after mid-day despite cessation of stomatal closure, because of the diurnal temperature regime. It is suggested that if the leaf surface temperatures are available, the vapor pressure gradient between leaf and atmosphere could probably be used to gauge periods of stomatal closure.

In the final analysis therefore, anomalous conditions such as occurred during periods of shading for upper level leaves and closure, are neglected in the derivation of the mean stomatal resistance ($r_{\underline{s}}$) as given by equation (5.1). This is justified on the basis of the rather infrequent occurrence of these anomalous conditions. Also only sunlit maple leaves seemed susceptible to this effect. In the next chapter, the performance of the model, when applied to equation (3.26) will be examined.

Chapter 6

Effect of Canopy Wetness on Evapotranspiration Rates

6.1 Preliminary Comments

The preceding chapter has discussed a model whereby hourly estimates of the canopy resistance (r_c) can be derived from measurements of either net (Rn) or global (Q + q) radiation. Since the model gave better results when using global radiation, this will be the parameter used to get hourly estimates of r_c. In the present chapter these predicted values of r_c will be used to calculate both the actual transpiration flux when the canopy is unwetted, and the assumed transpirational loss when the canopy is wetted by intercepted rainfall. Comparison of this flux (LE_d) with the evaporation of intercepted rainfall (LE_w) will be stressed. Data for the 1975 growing season only, when refined measurements were made, will be used for calculating the water balance.

6.2 Transpiration during Dry Periods

As a test of the model used for deriving surface or canopy resistance (r_c) of the vegetation, transpiration rates were calculated according to equation (3.26). The weather variables, namely net radiation, soil heat flux, vapor pressure deficit of the ambient air and aerodynamic resistance, were measured as described in chapter 4, while the surface or canopy resistance (r_c) is derived in accordance with equations (5.1) to (5.4).

Figures 6.1 and 6.2 show the diurnal regimes of net radiation (Rn), soil heat flux (G), the vapor pressure deficit of the ambient air (V.P.D.) and the latent heat of transpiration (LE_d) for two sunny days

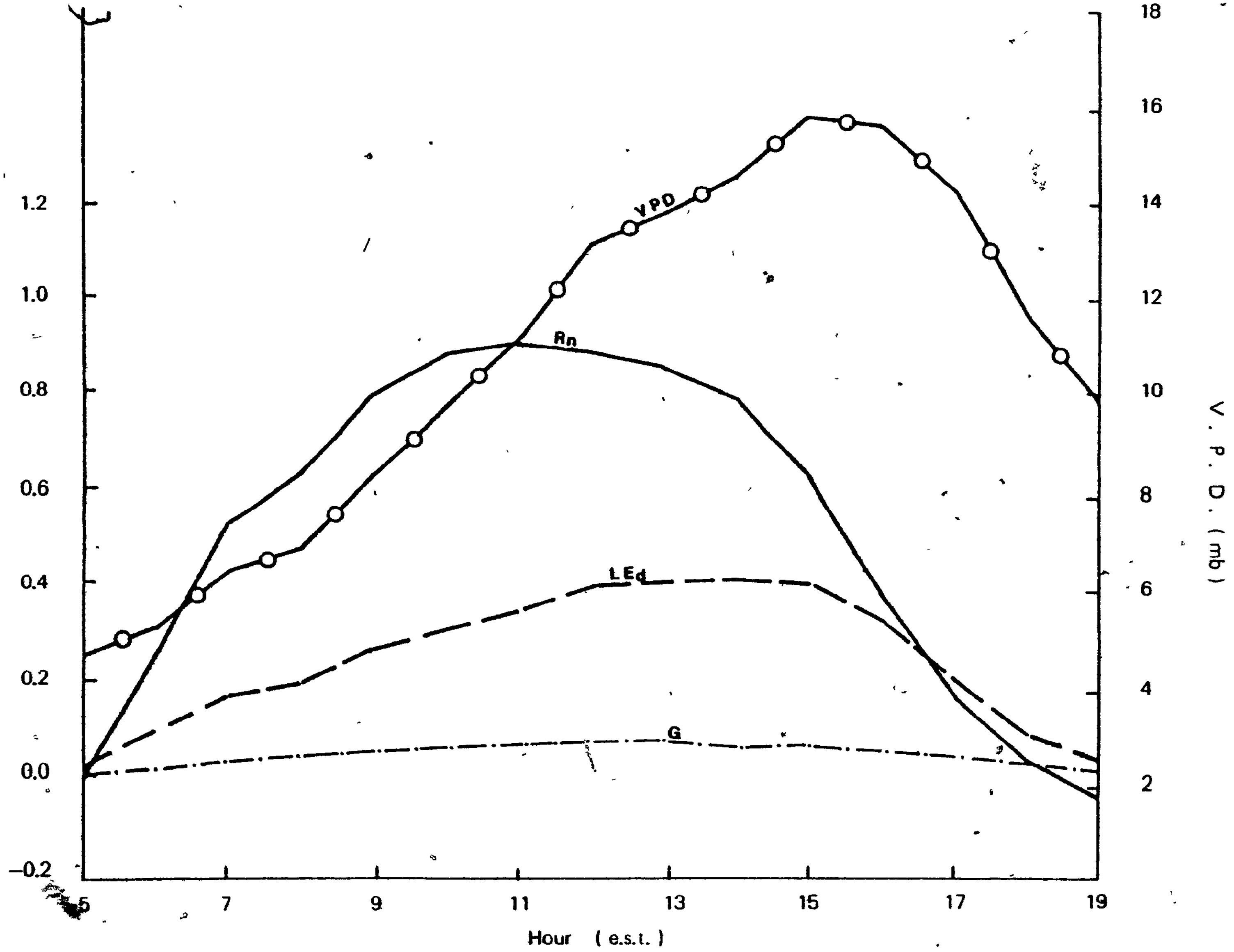

Figure 6.1 LEd for a mainly sunny day June 11, 1975.

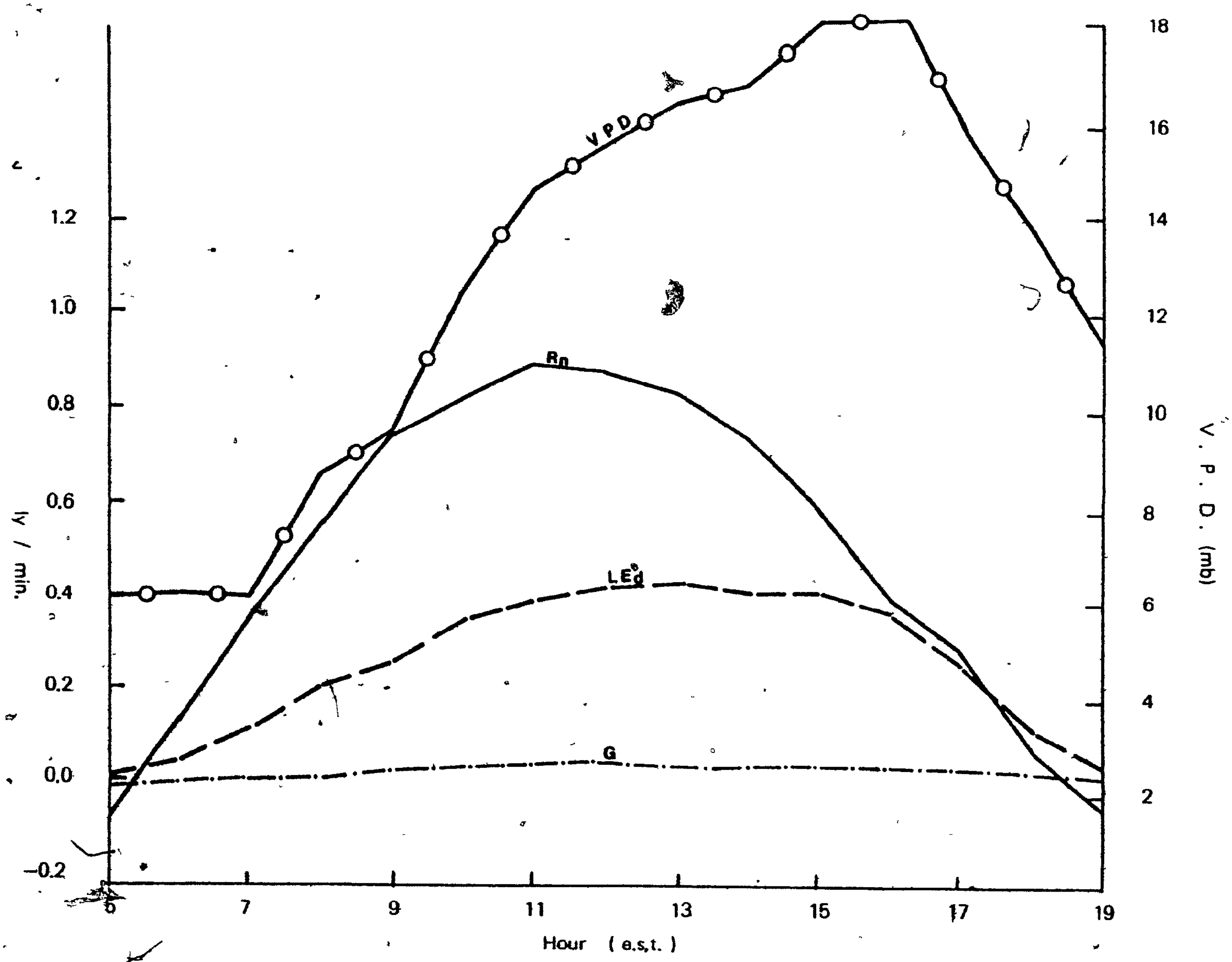

Figure 6.2 LEd for a mainly sunny day August 8, 1975.

during the 1975 growing season. It is noticeable that the latent heat flux (LE_d) is strongly dependent on the vapor pressure deficit (V.P.D.) in the afternoon period. It seems likely that under these conditions the leaves more or less tended to function as wet-bulbs in that in the absence of stomatal closure there was a substantial transfer of latent heat at the expense of sensible heat. This result is in agreement with the observation of Stewart and Thom (1973), namely, that transpiration from the forest occurs at rates much less dependent on net radiation than on ambient vapor pressure deficit, provided that the latter is not less than a few millibars.

The daily totals of transpirative water loss (3.9 to 4.0 mm), if the LE_d curves are integrated in figures 6.1 and 6.2, are also within the range of values found by Szeicz et al (1969) for pine forests in more extreme environments. They derived transpiration rates for Southern England, a relatively wet climate, that ranged from 1.0 to 2.7 mm/day. Similarly for the relatively dry climate of Southern California, transpiration estimates ranged from 2.10 to 5.05 mm/day, during the growing season.

The proportion of the daily total of net available energy that goes into latent heat transfer ($LE_d/(Rn - G) \approx 0.50$) is also consistent with the values mentioned by other researchers. For instance, in Southern England, Stewart and Thom (1973) found that the fraction of net available energy used to evaporate water seldom exceeds 0.4 over a pine forest. On the other hand, Monteith (1965) found that this ratio varies, on an annual basis, between 0.69 in the Thames Valley of Southern England and 1.00 in the Sacramento Valley of Northern California, over a pine forest in both cases.

The model that is proposed (see chapter 5) for estimating the canopy resistance (r_c), when used in equation (3.26), therefore, appears to give reasonable estimates of the transpirative latent heat flux. It is now necessary to examine the evaporative flux, when the canopy is wetted and r_c is reduced to zero.

6.3 Evaporation of Intercepted Rainfall

When the canopy is wetted by intercepted rainfall, the rate and amount of evaporation can be calculated according to equation (3.23). As was mentioned in the previous section, the weather variables were measured directly. Surface conditions, as characterized by the presence or absence of water on the leaves, were measured as described in section 4.8. To provide a check on the magnitude of the evaporative loss, when the canopy was wet, the amount of interceptional loss, calculated as described in section 4.3 was derived.

Figures 6.3 to 6.6 describe the diurnal regimes of latent heat transfer during completely wet, partially wet, dry and assumed dry canopy conditions, for selected days during the 1975 growing season. It is readily apparent from these diagrams that the evaporation rate (LE_w), when the canopy is wet, is several times the transpiration rate (LE_d) which would occur if the canopy were assumed to be dry in the same weather conditions. In fact the ratio LE_w/LE_d was found to range between 4.0 and 35.0, the magnitude depending on the ratio of the canopy to the aerodynamic resistance (r_c/r_a) and weather conditions, mainly the temperature and saturation deficit of the ambient air. Figure 6.4 shows that this latter effect also holds true when a wet as opposed to a unwetted canopy is examined: the rate of evaporation following

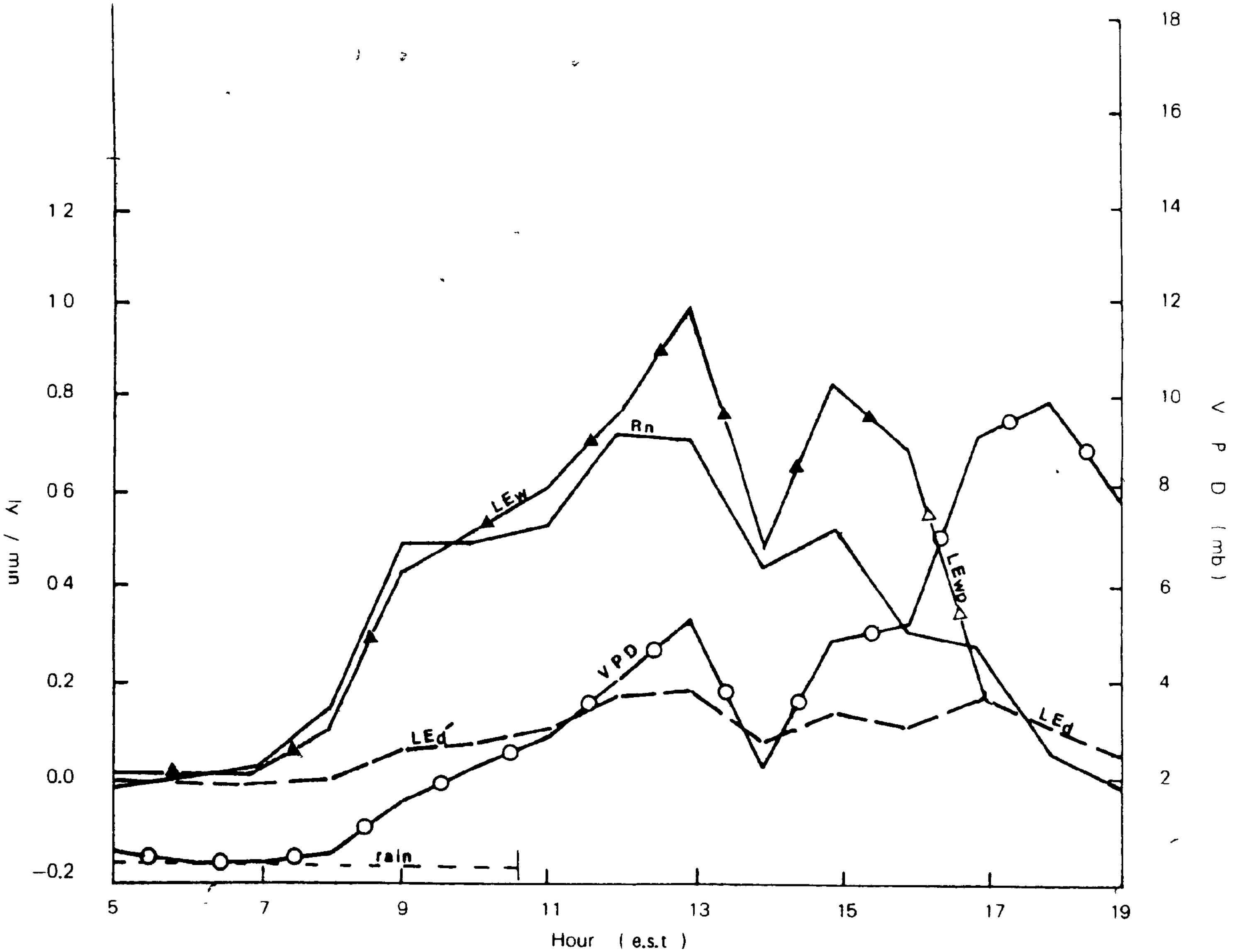

Figure 6.3 Latent heat transfer for wet (LEw), partially wet (LEwp) assumed dry (LEd´) and dry canopy conditions, May 31, 1975

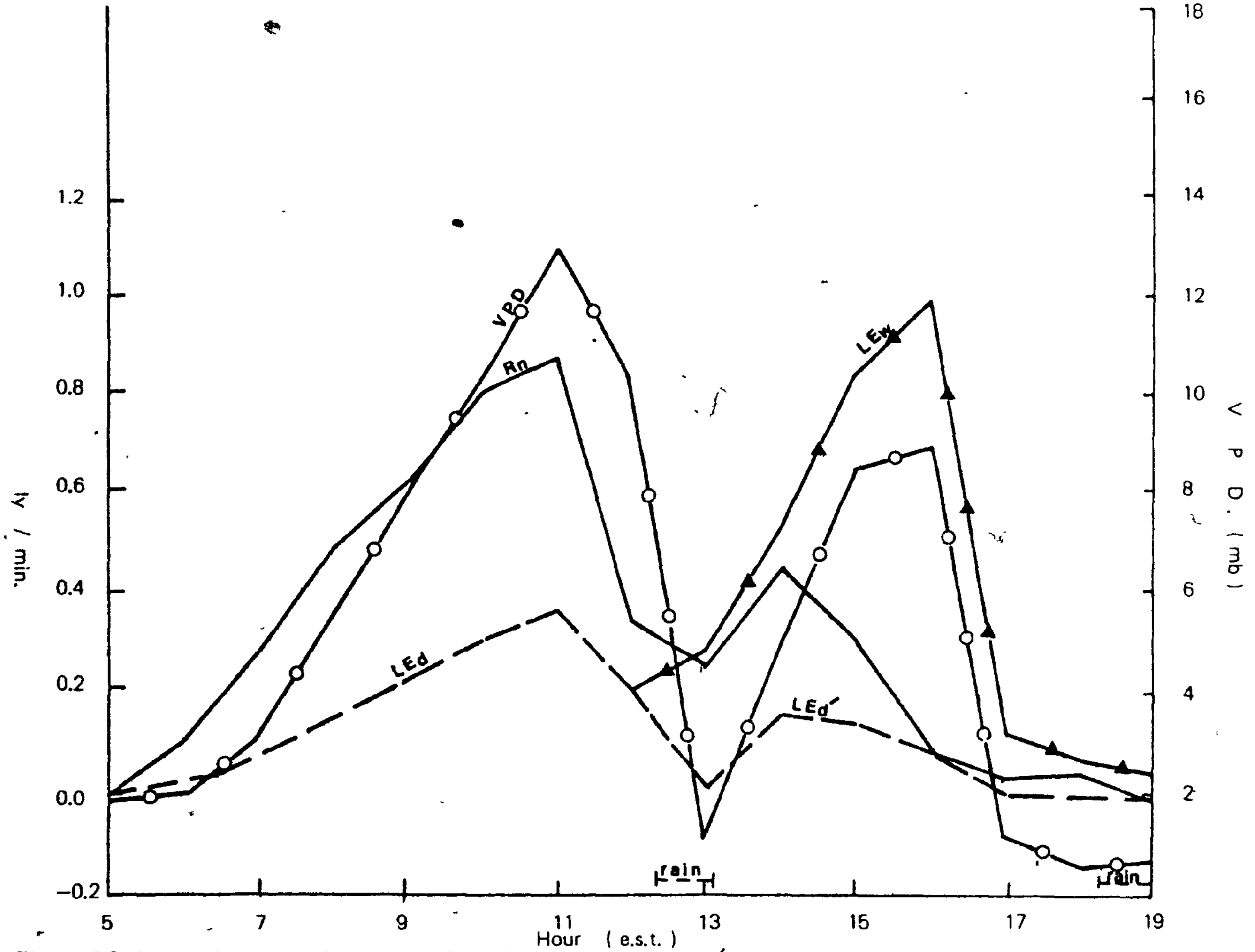

Figure 6A Latent heat transfer for wet (LE_w), assumed dry (LE_d') and dry (LE_d) canopy conditions, July 19, 1975

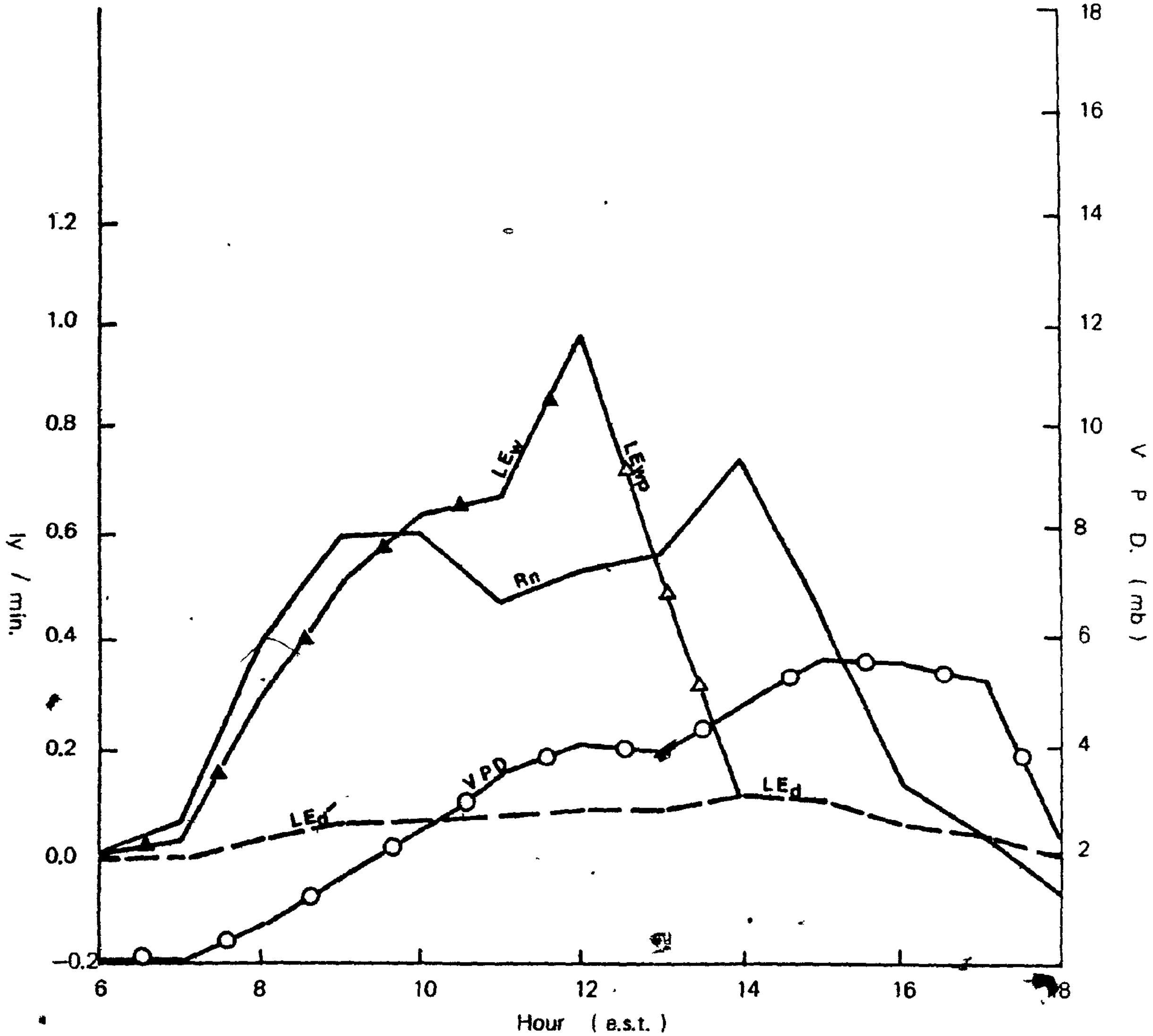

Figure 6.5 Latent heat transfer for wet (LE_w) partially wet (LE_{wp}) assumed dry (LE_d') and dry (LE_d) canopy conditions, Sept. 14, 1975

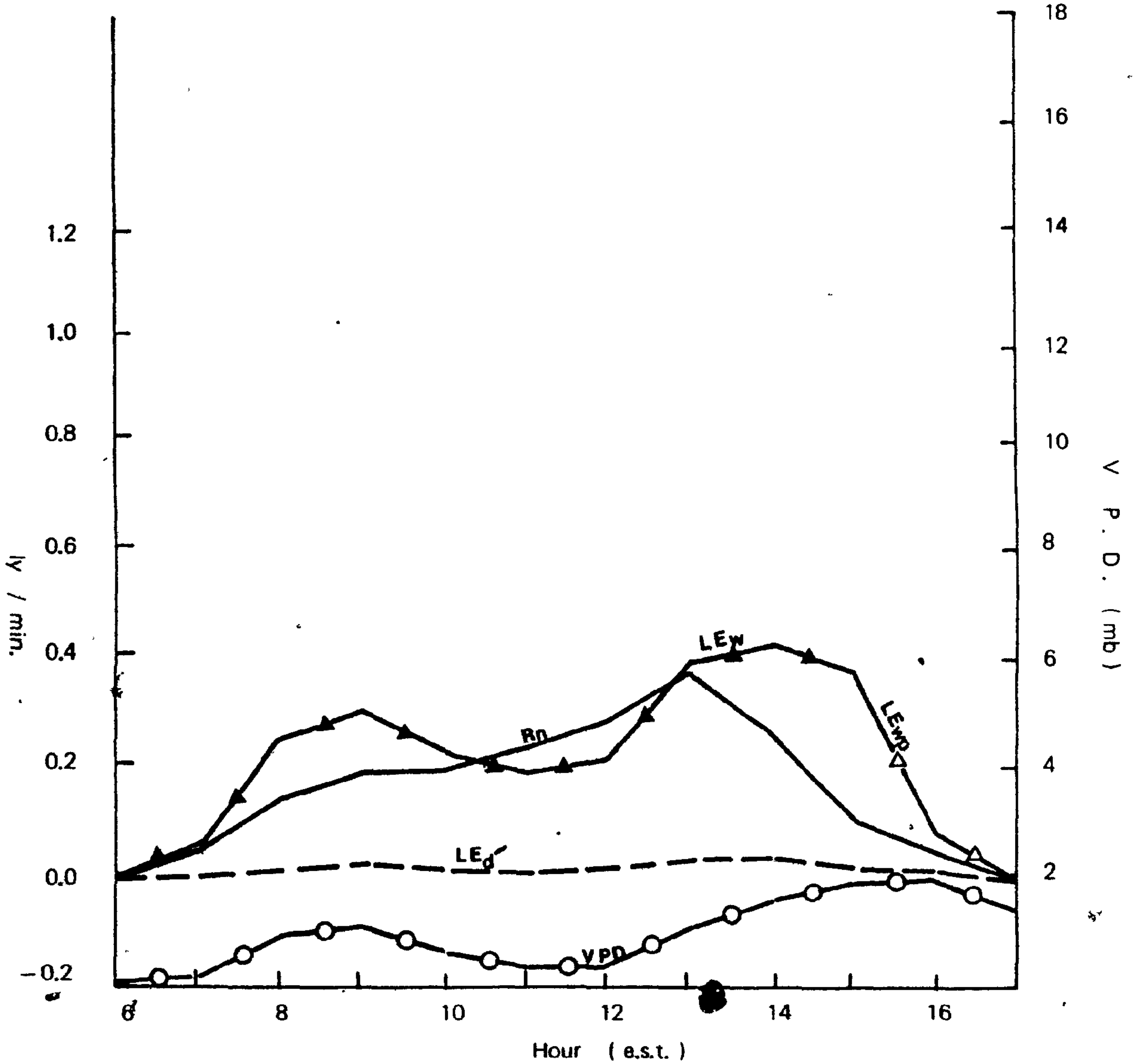

Figure 6.6 Latent heat transfer for wet (LEw) partially wet (LEwp) and assumed dry (LEd′) canopy conditions, Oct. 2, 1975

interception in the afternoon is observed to be several times the transpiration rate in the morning when the canopy was dry, although radiative and ambient-air conditions were more favorable for latent heat transfer in the morning.

Furthermore, during the majority of instances when the canopy was wet, the rate of latent heat transfer (LE_w) to the ambient air was found to be greater than the supply of net radiant energy (Rn) (see figures 6.3 to 6.6). This would imply that when the forest is wetted by intercepted rainfall, it becomes a strong sink for advected energy. This conforms with the findings of Rutter (1968) who suggests that there is a sizeable transfer of sensible heat to wetted vegetation at the expense of the surrounding area, especially where the vegetation occupies a small part of a much larger enclosing area, as is very much the case at the present site. McNaughton (1976a, 1976b) refers to this effect as "advective enhancement". It could also be, as suggested by Murphy and Knoerr (1975), that a wetted canopy conserves most of the incident radiant energy for latent heat transfer by suppressing long-wave radiative loss and sensible heat exchange.

In order to assess the accuracy of the estimated amount of latent heat exchange during wet periods, the total amount of evaporation of intercepted water during and after each rainfall, as calculated by equation (3.45), is compared with the amount of interceptional loss, measured as described in section 4.3 (see figure 6.7).

It is readily noticeable in figure 6.7 that except for the higher values of evaporation or interceptional loss (i.e. > 10 mm), there exists a close correspondence between estimated evaporation of

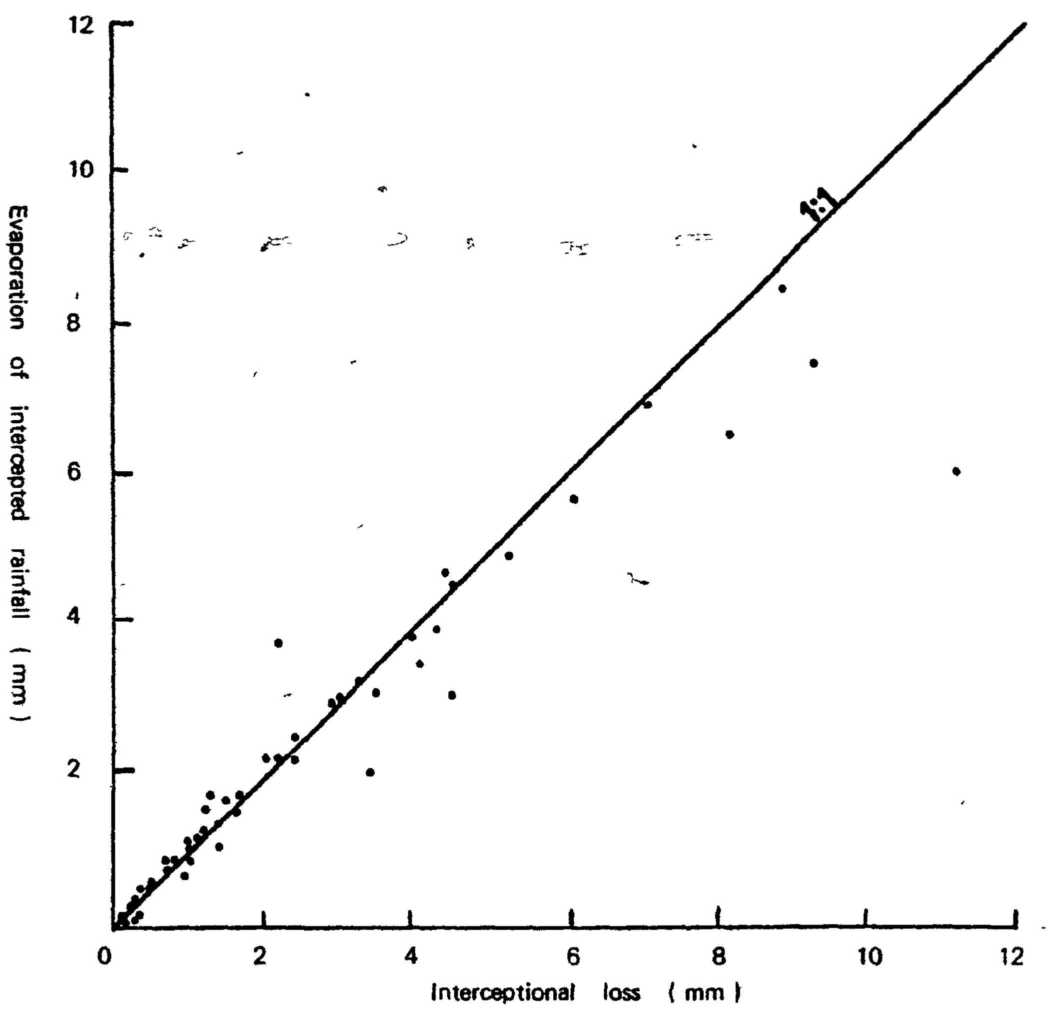

Figure 6.7 Evaporation of Intercepted rainfall vs measured interceptional loss, 1975

intercepted rainfall and measured interceptional loss. The discrepancy for the higher values is mainly attributable to the fact that evaporative losses were not measured during the night. As remarked by Penman (1963), there can however occur a substantial amount of evaporation at night, in the presence of a pronounced vapor pressure deficit in the ambient air and a relatively low aerodynamic resistance, as occurs during periods of strong windiness. As a result if night-time conditions are omitted, there will be a significant underestimation of the evaporation of intercepted rainfall. This effect was especially common when the forest remained wet at night, following a late afternoon or early evening shower. Other inconsistencies arose mainly from measurement errors for both variables, especially for the evaporation of intercepted rainfall where the delimitation of a wet or partially wet, as opposed to a dry canopy, was somewhat imprecise. In the calculation of the water balance however, interceptional loss is gauged by means of equation (3.45).

The results given above therefore show that the evaporation of intercepted rainfall (LE_w) proceeds at a much faster rate than the transpiration of soil moisture, when the canopy is dry or assumed to be dry, under similar weather conditions. In the next section, the interrelationships of these varying rates of vaporization will be discussed.

6.4 Wet Versus Dry Canopy

In order to gauge the ratio of the rate of evaporation of intercepted rainfall (LE_w) to that of transpiration (LE_d), assuming the same weather conditions, equation (3.46) is utilized. Reference to figures 6.8 and 6.9 shows that the magnitude of the ratio LE_w/LE_d is ~~closely~~ related to that of the ratio r_c/r_a. Also the greatest values

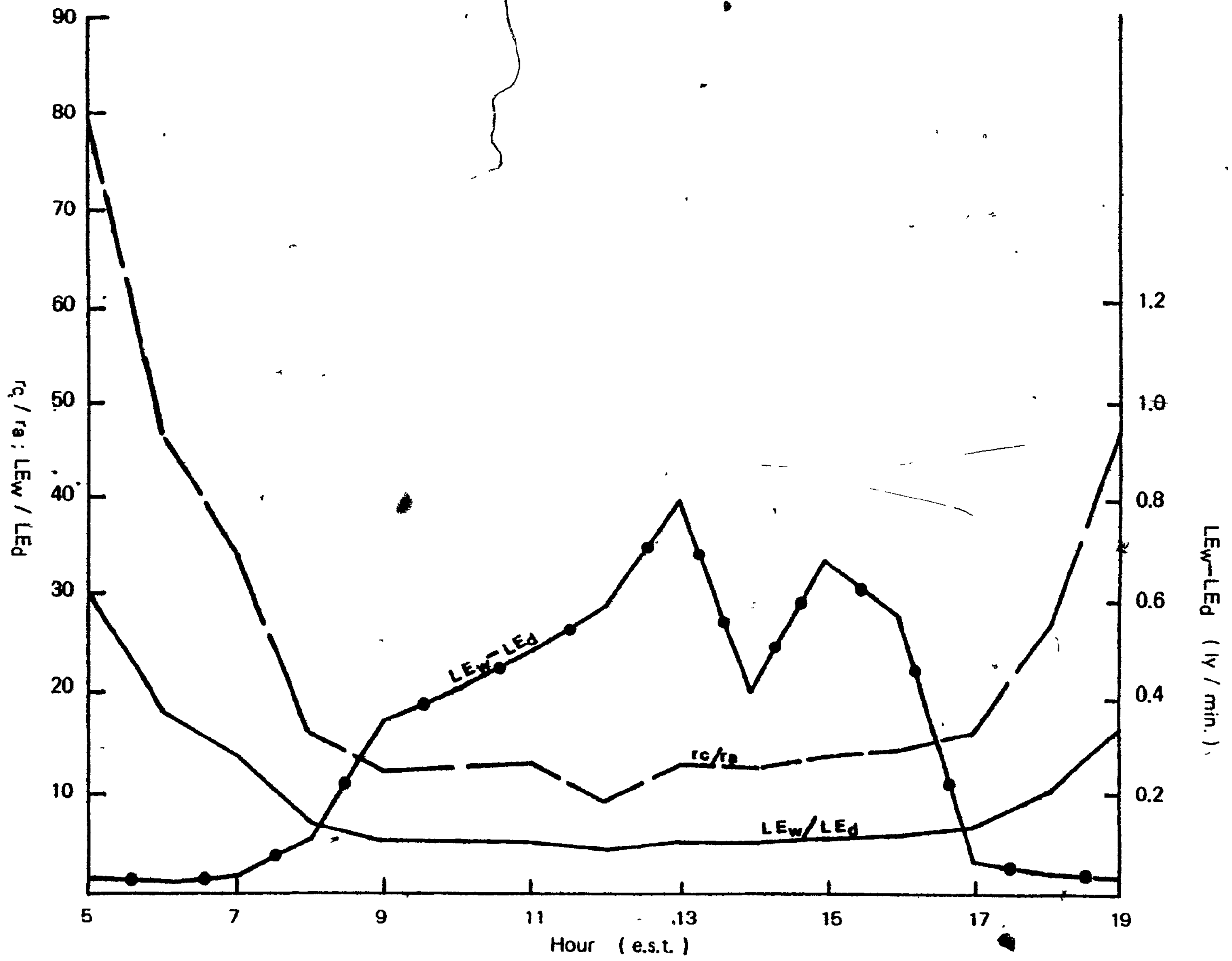

Figure 6.8 LEw/LEd as a function of rc/ra, and the diurnal regime of LEw– LEd, May 31, 1975.

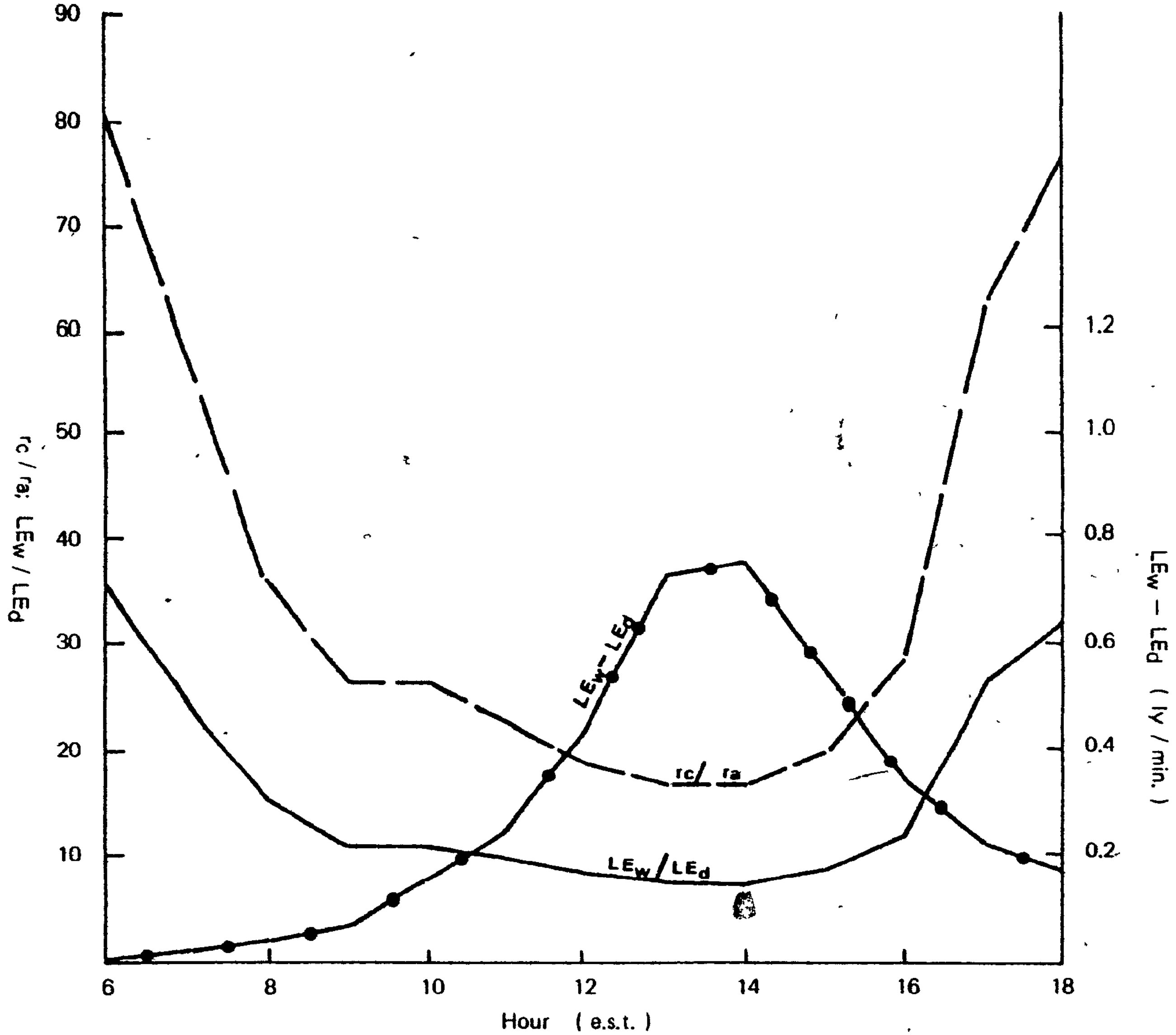

Figure 6.9 LE_w/LE_d as a function of r_c/r_a, and the diurnal regime of $LE_w - LE_d$ Sept. 27, 1975.

of the proportionality LE_w/LE_d occur during the early morning and late afternoon. This diurnal trend is due to the fact that during these periods of low radiative inputs, the stomates of the leaves are either closed or approaching closure, thereby giving the highest canopy resistance (r_c) values. Since the aerodynamic resistance (r_a) which is normally of a lesser magnitude than r_c, is rather conservative in value, stomatal behavior is the dominant factor, and as a result the ratio r_c/r_a, and hence LE_w/LE_d, is greatest during the low sun periods.

The ratio LE_w/LE_d by itself, however, does not truly reflect the differential in water loss through evapotranspiration for a dry as opposed to a wetted canopy, assuming the same weather conditions. Both figures 6.8 and 6.9 show that the value of the difference $LE_w - LE_d$, as given by equation (3.48) is more convenient as a means of exploring the magnitude of water loss consequent upon interception. In actuality $LE_w - LE_d$ is greatest when the ratio LE_w/LE_d is least. This result is obviously related to the diurnal regime of latent heat transfer. Despite the fact that the differential between potential (wetted) and non-potential (unwetted) surface conditions is least when radiant energy is greatest because of stomatal behavior, the accompanying increase in the evaporative demand of the atmosphere, as evidenced mainly by a greater saturation deficit, causes greater latent heat transfer. But the water loss via latent heat exchange is greater for potential than for non-potential surface conditions, since surface resistance (r_c) to vapor diffusion is reduced to zero in the former situation. This effect causes the greatest difference in $LE_w - LE_d$ to occur during the time of peak evaporative demands by the ambient air. The magnitude of this difference depends, in turn, on the ratio r_c/r_a, as observed earlier, and prevailing weather conditions. Because of the unique control of weather conditions on the evaporation of intercepted rainfall, a fuller discussion of their effect is reserved for elsewhere (see chapter 7).

It is also noticeable in figure 6.10 that the ratio LE_w/LE_d is subject to seasonal changes. The ratio is seen to be consistently greater in late September than in late May because of higher stomatal resistances in the former period, as was observed earlier (see section 5.4). However the latter effect is slightly reduced by the higher temperatures in May (see figure 6.11) when higher saturation vapor pressure conditions (S), resulting from higher temperatures, give more weight to the denominator of equation 3.46, which in turn causes LE_w/LE_d to be lower.

The preceding sections have shown that the evaporation of intercepted rainfall proceeds at a much faster rate than the transpiration of soil moisture, were the canopy assumed to be dry. Also this difference in the rate of latent heat exchange undergoes both diurnal and seasonal changes.

6.5 Effect on Soil Moisture Withdrawal

If it is assumed that the transpiration of soil moisture is suppressed during the evaporation of intercepted rainfall, then a certain amount of soil moisture saving can result. Furthermore, if the evaporation of intercepted rainfall proceeds at a much faster rate than the transpiration of soil moisture, were the canopy not wet, in the same weather conditions, then the evaporation of water consequent upon interception constitutes a greater water loss than if the moisture were to

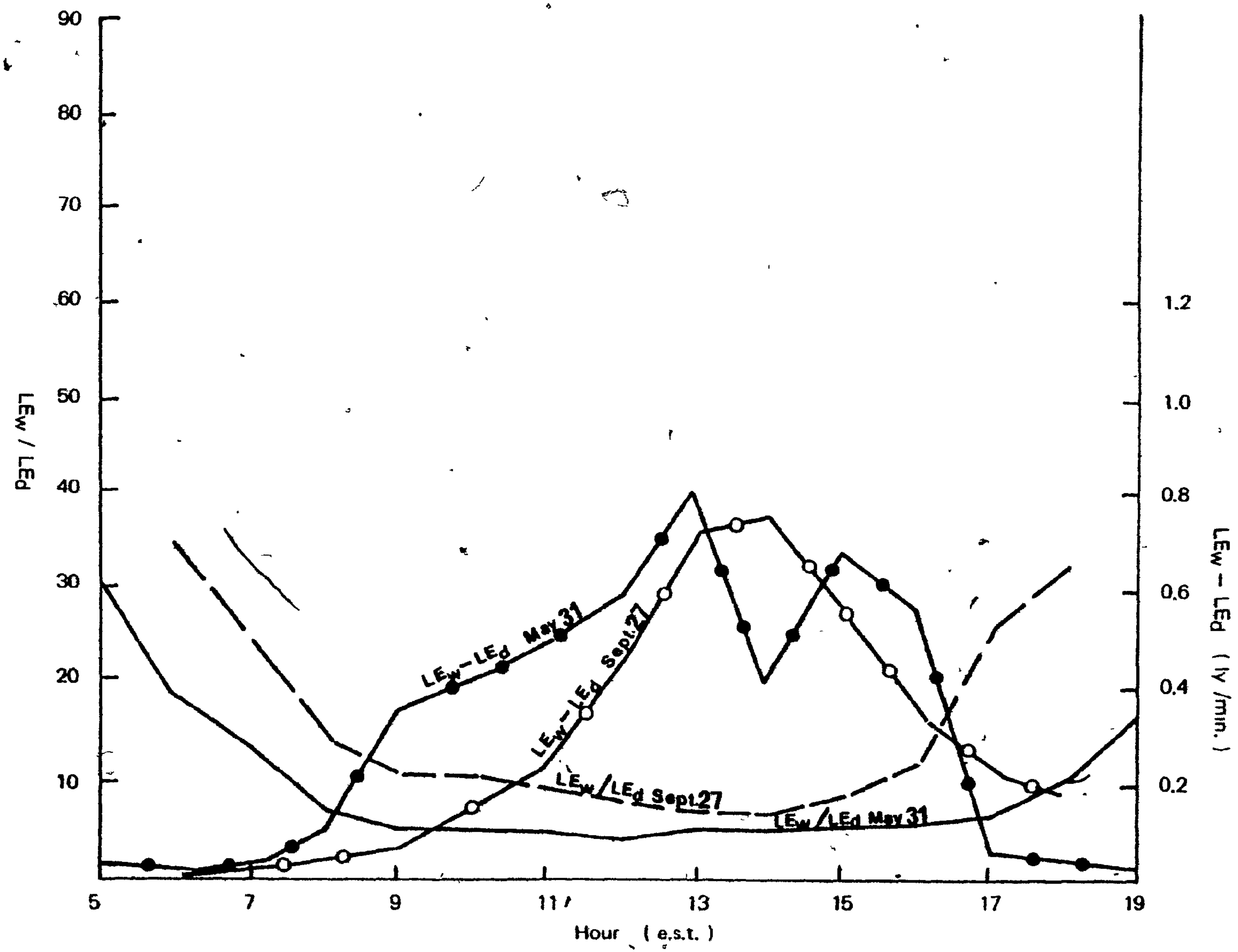

Figure 6.10 Variation of LE_w/LE_d and $LE_w - LE_d$ with season, 1975.

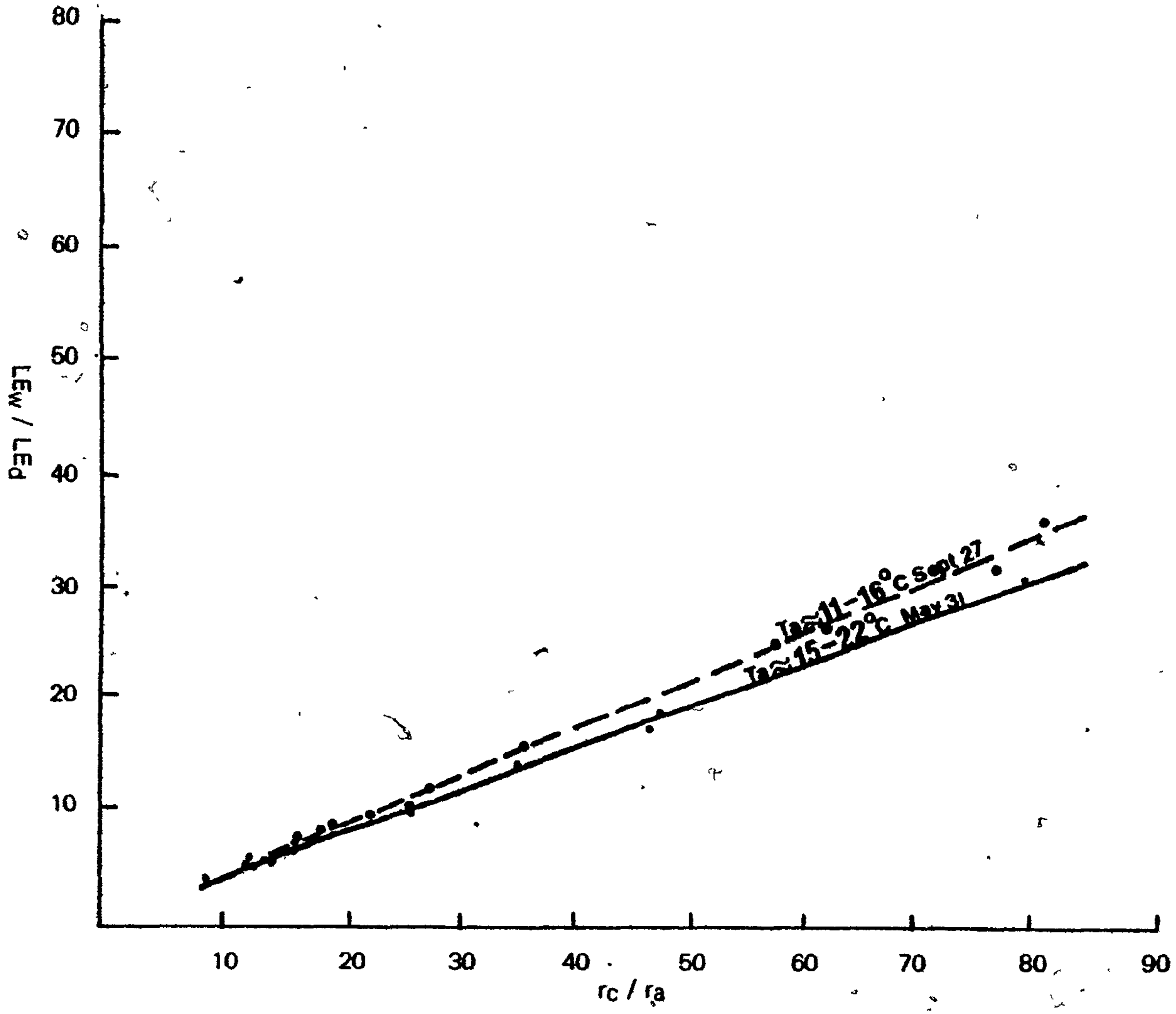

Figure 6.II Effect of air temperature on LE_w/LE_d, 1975.

transpire through the leaves of the vegetation.

An examination of figure 6.10 shows that, on the average, the rate of evaporation of intercepted rainfall (LE_w) ranges from about 4 to 30 times the rate of transpiration (LE_d), under the same weather conditions. Assuming, therefore, that there is no transpiration while intercepted water is being evaporated, it can be shown through use of equations 3.50 to 3.52 that about a quarter to one thirtieth (25 percent to 3.3 percent) of intercepted rainfall is equivalent to transpiration that would otherwise have occurred in the same atmospheric conditions, were the canopy dry. On the other hand, about three-quarters to twenty-nine thirtieths (75 percent to 96.7 percent) of intercepted water is attributable to evaporation that would not have occurred in the absence of precipitation and interception.

Equation (3.52) is used to calculate the additional evaporation consequent on interception (I_N), for selected days during the 1975 growing season. The values derived are shown in table 6.1. It is readily apparent from this table that on the average the additional evaporation consequent on wetting is greater than 70 percent of the total interceptional loss, and that less than 30 percent of this total goes towards conserving soil moisture through the supposed suppression of transpiration. A sizeable portion of intercepted rainfall (>70 percent) is therefore lost to the soil beneath. This figure may be somewhat high as a generalization for the forest in question in that the summer of 1975 was unusually dry thereby giving higher than average r_c/r_a and hence LE_w/LE_d ratios.

TABLE 6.1 Net Interceptional Loss (I_N) for Selected Days, 1975

Period-Date 1975	LE_d' mm	LE_w mm	I mm	I_N	
				mm	%
1 - June 6	0.324	2.260	3.403	2.1955	85.7
1 - June 12	0.815	1.819	1.752	0.9674	55.2
1 - June 19	0.487	3.112	3.073	2.5924	84.3
2 - July 19	1.462	6.831	6.706	5.2667	78.6
2 - July 24	0.687	6.988	7.138	6.4363	90.1
3 - August 4	0.660	2.318	2.363	1.6903	71.5
3 - August 29	0.370	1.422	1.422	1.0520	73.9
4 - September 6	0.471	2.317	2.261	1.8013	79.7
4 - September 19	0.471	1.693	1.524	1.1000	72.2
4 - October 1	0.225	2.616	2.362	2.1589	91.4

LE_d' = Transpiration rate assuming a dry canopy (mm)

LE_w = Evaporation of intercepted rainfall (mm)

I = Interceptional loss (mm)

I_N = Net interceptional loss (mm).

It can be concluded therefore that although intercepted rainfall by the forest under consideration contributes to soil moisture saving, the bulk of it is lost to the soil beneath. Intercepted rainfall can therefore be viewed as a significant loss of moisture to the forest basin, and its effect, when calculating the water balance, must be considered.

6.6 Effect on the Water Balance

It is quite conceivable that the much greater rate of evaporation of intercepted rainfall as compared to transpiration, under the same weather conditions, can significantly affect water consumption over the forest and hence the water balance.

In order to highlight the effect of intercepted rainfall, the water balance, for each of the periods mentioned earlier (see section 5.4) and for the 1975 growing season as a whole, are calculated by taking account of interception on the one hand and by ignoring its effect on the other.

The form of the water balance equation can be written as:

$$Ppt = E + Ro \pm \Delta Sm \qquad (6.1)$$

where

Ppt = precipitation in the open (mm)

E = evapotranspiration (mm)

Ro = runoff (mm)

ΔSm = change in depth of soil moisture (mm)

Rainfall in the open (Ppt) and runoff were measured as described previously (see sections 4.3a and 4.5 respectively) while soil moisture change (Δsm) was found by taking the difference of soil

TABLE 6.2: Water Balance Calculations, 1975 (all values mm water)

Period	Ppt	Evapotranspiration		Ro	ΔSm	Contribution of Intercepted Rainfall (I_N)		Difference			
		Including Wet Periods	Transpiration Only				%	Including Wet Period	%	Transpiration Only	%
1 - May 29 -July 1	120.520	129.080	104.201	58.063	-70.50	24.879	19.3	-3.877	-3.2	-28.756	-23.8
2 - July 7 -July 30	96.770	92.504	56.451	22.439	-13.60	34.053	36.8	4.570	4.7	-29.480	-30.5
3 - July 31 -Sept. 5	84.074	101.563	79.730	24.218	-43,50	21.833	21.5	-1.794	-2.1	-23.626	-28.1
4 - Sept. 6 -Sept. 29	133.858	42.180	17.647	19.606	80.05	24.533	18.3	7.978	6.0	-16,555	-12.4
5 - Sept. 30 -Oct. 7	9.906	7.671	5.294	8.007	-5.90	2.377	24.0	1.872	10.8	-2.505	-25.2
Total Season	445.128	372.998	265.323	132,333	-52.50	107.675	28.9	7.703	1.7	-99.972	-22.5

moisture depth between the beginning and end of each period. Evapotranspiration was calculated according to equations 3.23 and 3.26 depending on surface wetness conditions.

Table 6.2 summarizes the results of these calculations. The table also shows the magnitude of the errors when total evapotranspiration on the one hand is estimated by including the effect of intercepted rainfall and by ignoring its effect on the other.

From table 6.2 it can be seen that, for all periods, by including the additional evaporation consequent upon the interception of rainfall, the magnitude of the differences between the left hand side of equation (6.1) (input) and the right hand side (consumption plus yield plus storage change) ranges between 2.1 percent (underestimation) for period 3 and 10.8 percent (overestimation) for period 5. For the growing season as a whole the value is only 1.7 percent (overestimation. These values are entirely satisfactory for the water balance in that they lie within the bounds of a reasonably allowable measurement error of about 10 percent.

By disregarding the additional evaporative loss caused by interception on the other hand, there results a consistent underestimation of water consumption that ranges between 12.4 percent (period 4) and 30.5 percent (period 2). Over the entire growing season the underestimation is 22.5 percent. These values are definitely outside the limits of a reasonable allowable error which if we follow acceptable limits of measurement can be placed at ± 10 percent (Gray, 1970; Helvey and Patric, 1965).

The added contribution to evaporative loss by intercepted rainfall therefore ranges from 18.3 percent (period 4) to 36.8 percent (period 2) of total evapotranspiration. The value of the extra water loss consequent on interception, for the growing season as a whole is 107.67 mm of water or 28.9 percent of the total water lost through evaporation and transpiration.

The calculations presented in table 6.2 then, substantiate the view that the evaporation of intercepted rainfall (LE_W) proceeds at a faster rate than the transpiration of soil water (LE_d), assuming the same weather conditions. Because of the faster rate of LE_W than that of LE_d a substantial percentage (>70 percent) of intercepted rainfall is lost to the soil beneath. Neglect of the additional evaporative loss produced by intercepted rainfall can therefore lead to sizeable underestimations of water consumption over the forest and can thus significantly affect the calculation of the water balance.

Chapter 7

The Effect of Weather Conditions

7.1 Preliminary Remarks

In the previous chapter it was shown that surface conditions, as characterized by canopy wetness, affect the rate of evapotranspiration over the forest being discussed. It was also indicated that the ratio LE_w/LE_d and the difference $LE_w - LE_d$ were largely determined by the ratio r_c/r_a and weather conditions. This chapter will attempt to demonstrate the fact that the rate of evaporation of intercepted rainfall is determined by such weather characteristics as rainfall amount, intensity duration and frequency, and post-wetting synoptic conditions.

7.2 Evaporation During Rainfall

In section (3.4) it was shown that total interceptional loss through vegetation is more a dynamic rather than a passive process. In other words, the amount of precipitation withheld from the ground depends not only on the intercepting capacity of the forest stand but also on the amount of concurrent evaporation from the canopy during the period of rainfall. As mentioned previously, Wilm and Niederhof (1941) observed that about 19 percent of each rainfall is lost through evaporation from the canopy during wetting. Rutter et al (1971/72; 1975) found similar values. The degree and amount of evaporation during rainfall is, on the other hand, a function of rain characteristics and prevailing weather.

An examination of figure 7.1 shows that for a very light rain that fell throughout the day, evaporation of intercepted rainfall occurred throughout the period of wetting. Saturation of the ambient air actually

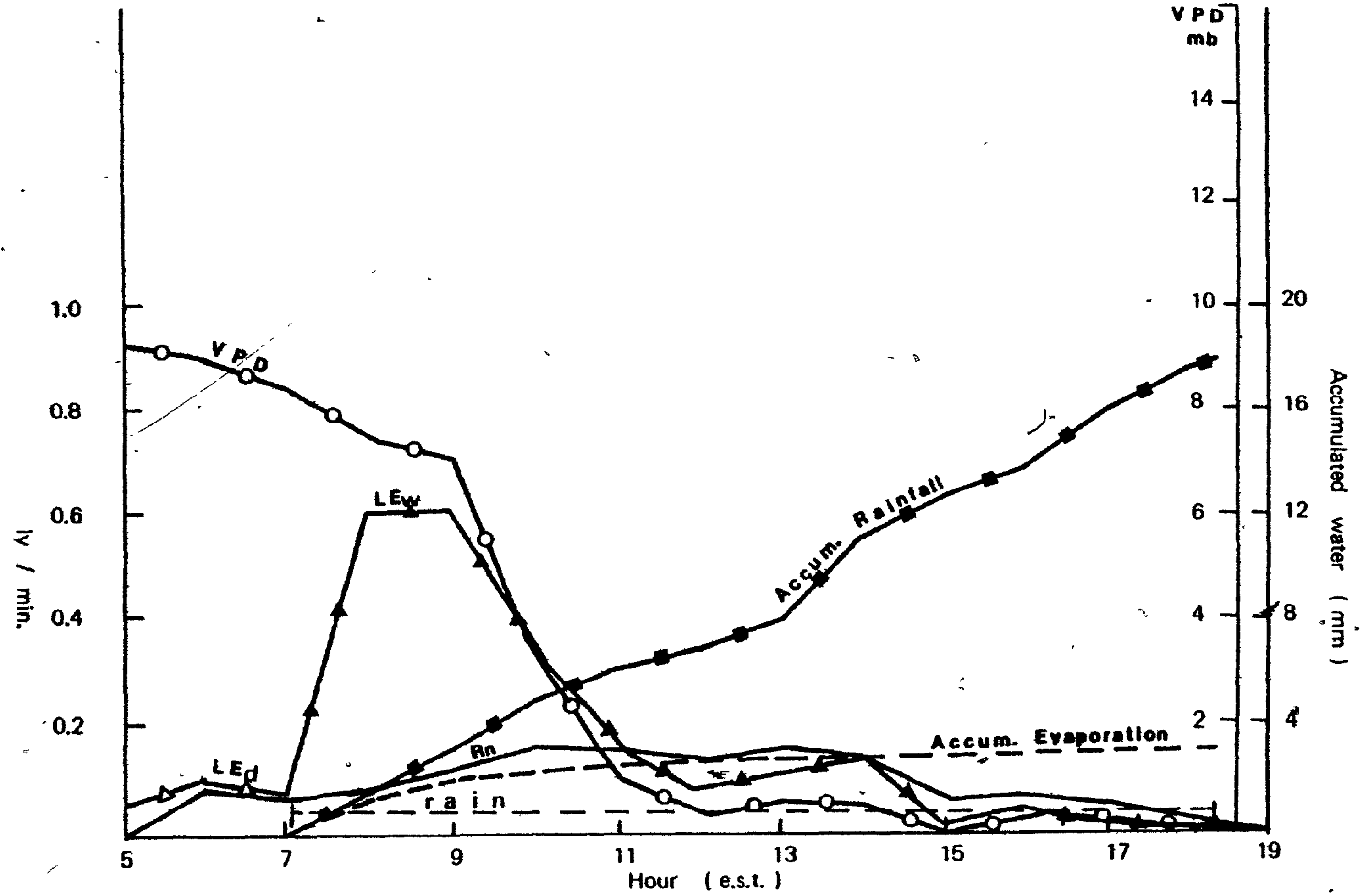

Figure 7.1 Evaporation of intercepted water during light intensity precipitation, May 30, 1975.

never took place because of windy conditions that provided continuous ventillation. Over a period of about 11 hours an accumulated total of 3 mm of evaporation occurred. Most of this amount was accounted for during the initial stages of the rain, when the saturation deficit of the air was still relatively high.

During rainfalls of medium and heavy intensity (see section 7.3) however, evaporation of intercepted rainfall was found to be minimal (see figure 7.2). On these occasions saturation or near-saturation of the ambient air was quickly obtained. The rather short durations of these showers also inhibited sizeable evaporative losses during wetting. In exceptional cases however a small amount of evaporative loss was observed to occur during the initial stages of these rainfall types. These occasions occurred either when the rainfall arrived suddenly, thus allowing the saturation deficit of the ambient air to be high initially, or when extremely windy conditions advected unsaturated air, at least during the early stages of the rainfall.

From the foregoing then, it is conceivable that a significant amount of interceptional loss through evaporation, that sometimes exceeds the intercepting capacity of the canopy, can occur during rainfall. The amount of interceptional loss during wetting is in turn dependent upon the nature of the rainfall.

7.3 Rainfall Characteristics

Rainfall characteristics, namely amount, duration, intensity and frequency of occurrence, can affect the amount of evaporation of intercepted rainfall. The regime of evaporative loss varies in terms of rainfall characteristics both during and after the period of wetting.

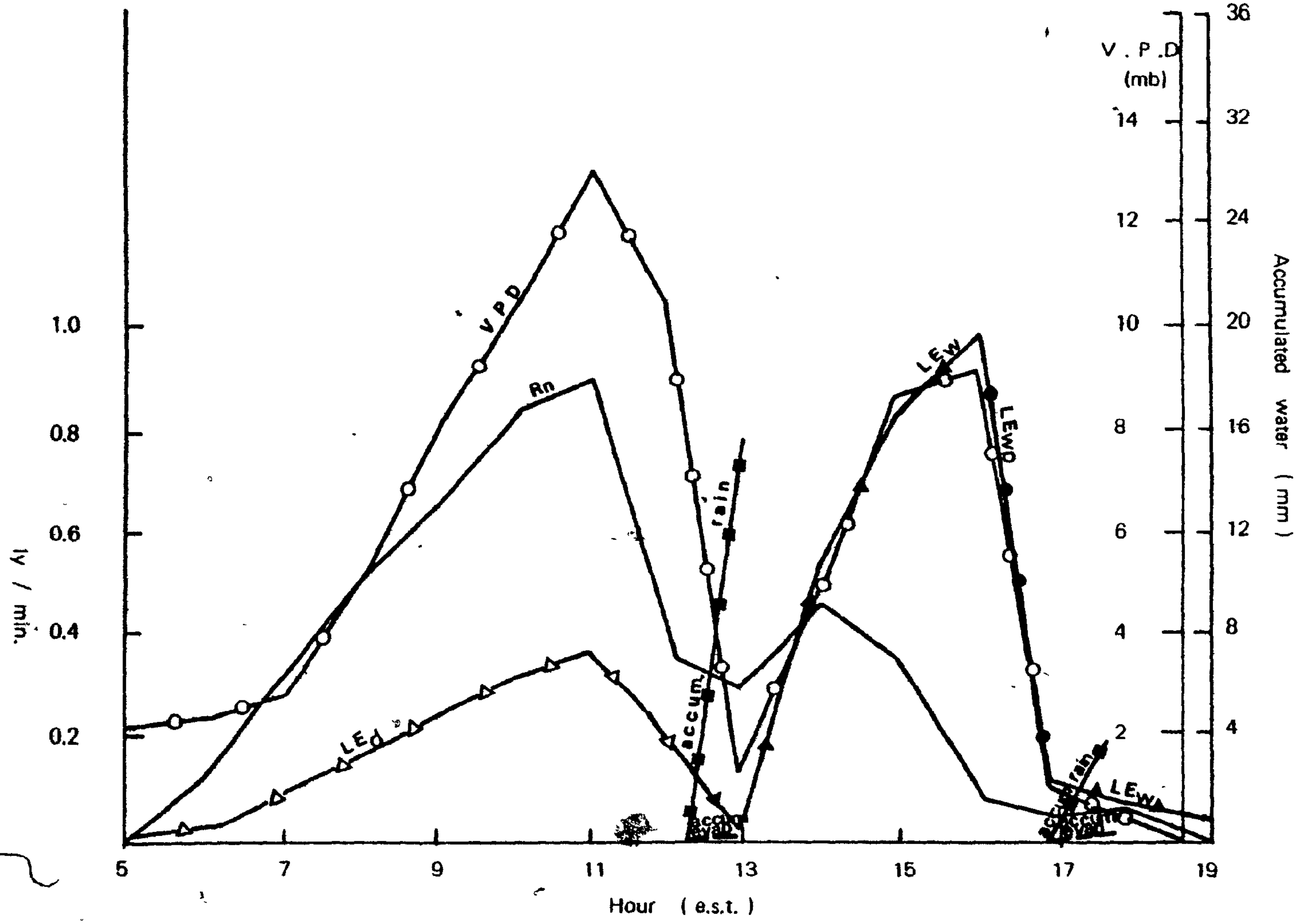

Figure 7.2 Evaporation of intercepted water during heavy intensity precipitation, July 19, 1975.

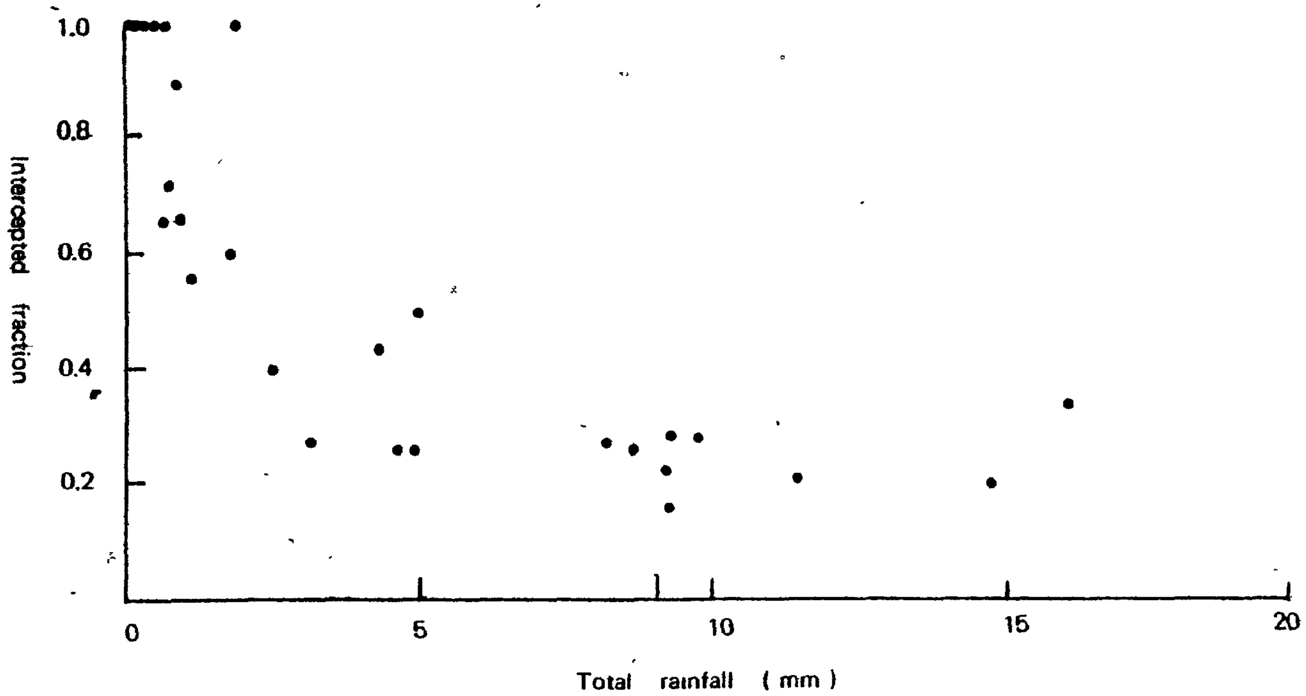

Figure 7.3 Intercepted fraction vs total rainfall, August 1 to October 2, 1974.

Reference to figure 7.3 shows that, on the average, the fraction of total precipitation intercepted is about 20 to 30 percent for rainfall amounts greater than about 5 mm. Similar percentages were observed by Horton (1919), Zinke (1967), Rogerson and Byrnes (1968), and Bultot et al (1972), for comparable vegetation types. For decreasingly lesser amounts (< 5 mm) the intercepted fraction increases exponentially and a saturation point, at which almost all of the precipitation is intercepted, is reached, at rainfall amounts of about 2 mm and less. It must be observed that throughfall can occur, before saturation of the intercepting capacity is reached, since some raindrops can penetrate the canopy through open spaces, or can splash off the edges of leaves. This would suggest that the intercepting capacity of the vegetal cover being discussed is in the vicinity of 2 mm of water.

In order to delimit the intercepting capacity (passive) of the canopy, gross precipitation (above canopy) was plotted against net precipitation (throughfall plus stemflow)(see figure 7.4). The data points used are for medium and heavy intensity rainfalls (>26 cm/hr) of short duration (less than a couple of hours), so as to subdue the effect of evaporation during rainfall. Also only rainfall amounts greater than 4 mm were used since, as observed in figure 7.3, and as noted in previous experiments (Horton, 1919; Rowe and Hendrix, 1951; Leyton et al, 1967), the slope of the regression of net against gross precipitation changes because net rainfall approaches zero beneath this approximate critical value. The intercept of the best-fit line with the gross precipitation axis gives an estimate of the intercepting capacity of the canopy, which in this case happens to be 2.4 mm (see figure 7.4).

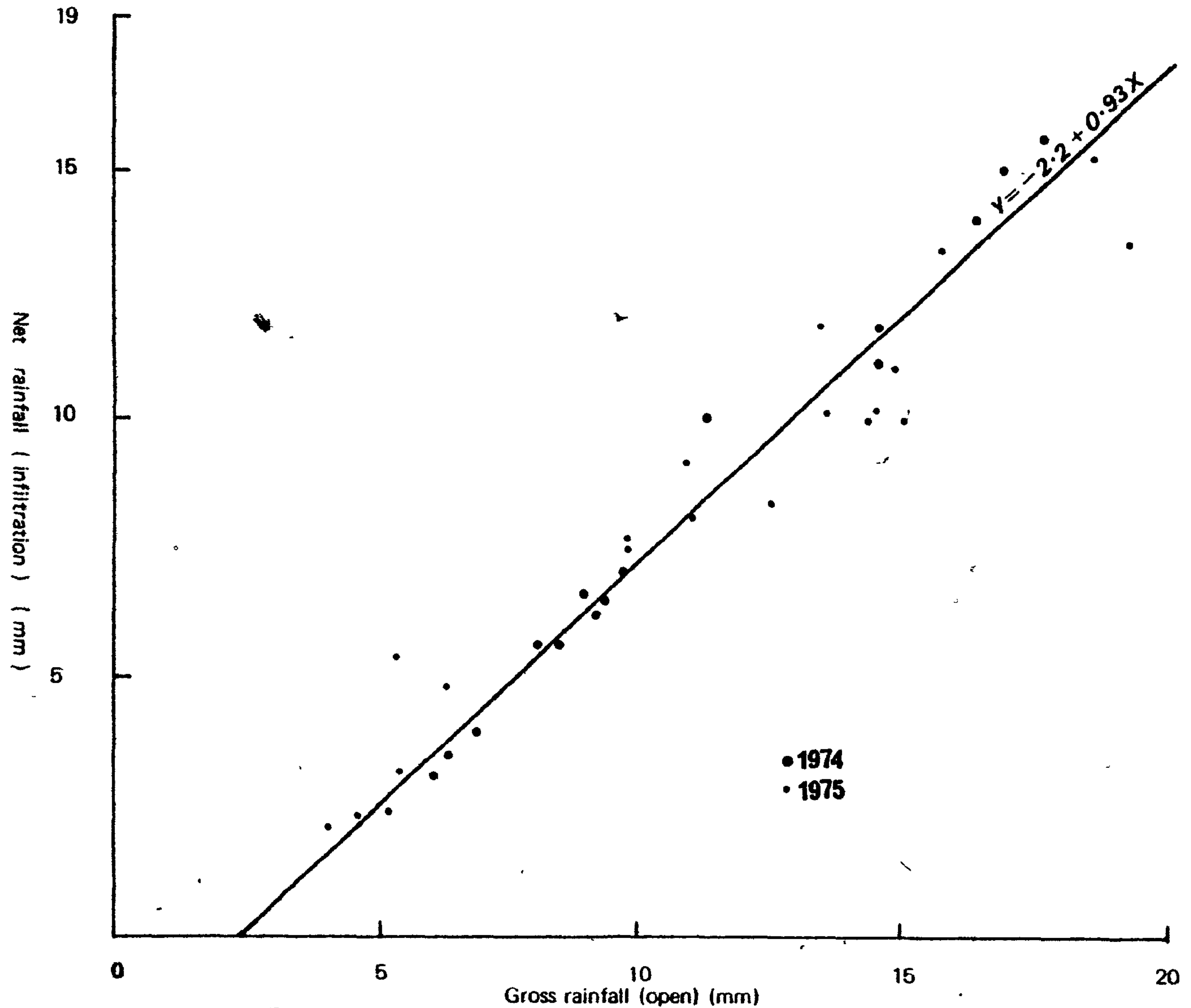

Figure 7.4 Determination of storage capacity (2.4 mm) of canopy.

This value is similar to that observed by Zinke (1967), and Bultot et al (1972), for a similar type of vegetation. The intercepting capacity however varies according to windiness, reaching its maximum under still conditions, since wind-induced shaking of the branches causes mechanical removal of water from the leaf surfaces.

Apart from amount, rainfall intensity can also be critical in determining the amount of precipitation that is intercepted. Rainfalls for both growing seasons (1974 and 1975) were classified into 3 intensity categories following the classification of the Atmospheric Environment Service of Canada, as set out in Manobs (1961): light (< 0.25 cm/hr), moderate (0.26 to 0.76 cm/hr) and heavy (> 0.76 cm/hr). Figure 7.5 shows that there is some relationship between rainfall intensity and the fraction of precipitation intercepted. The most apparent result is that, except in cases where excessive windiness reduced the storage capacity of the leaves, all rainfall amounts less than the canopy storage capacity (2.4 mm), were intercepted, regardless of intensity. When precipitation totals exceeded the intercepting capacity however, light intensity rains usually had a greater proportion of the total amounts intercepted, than moderate and intense rainfalls, since a greater amount of evaporation occurred during light intensity rainfalls, as was seen in the preceding section. On exceptional occasions however, as when windy conditions advected drier air to the measurement site, a sizeable amount of evaporation of intercepted rainfall occurred during moderate and heavy intensity showers. Similarly, on some occasions light intensity rainfalls were preceded by extended cloudy and humid conditions, so that by the time the rain commenced, the vapor pressure deficit of the ambient air was already

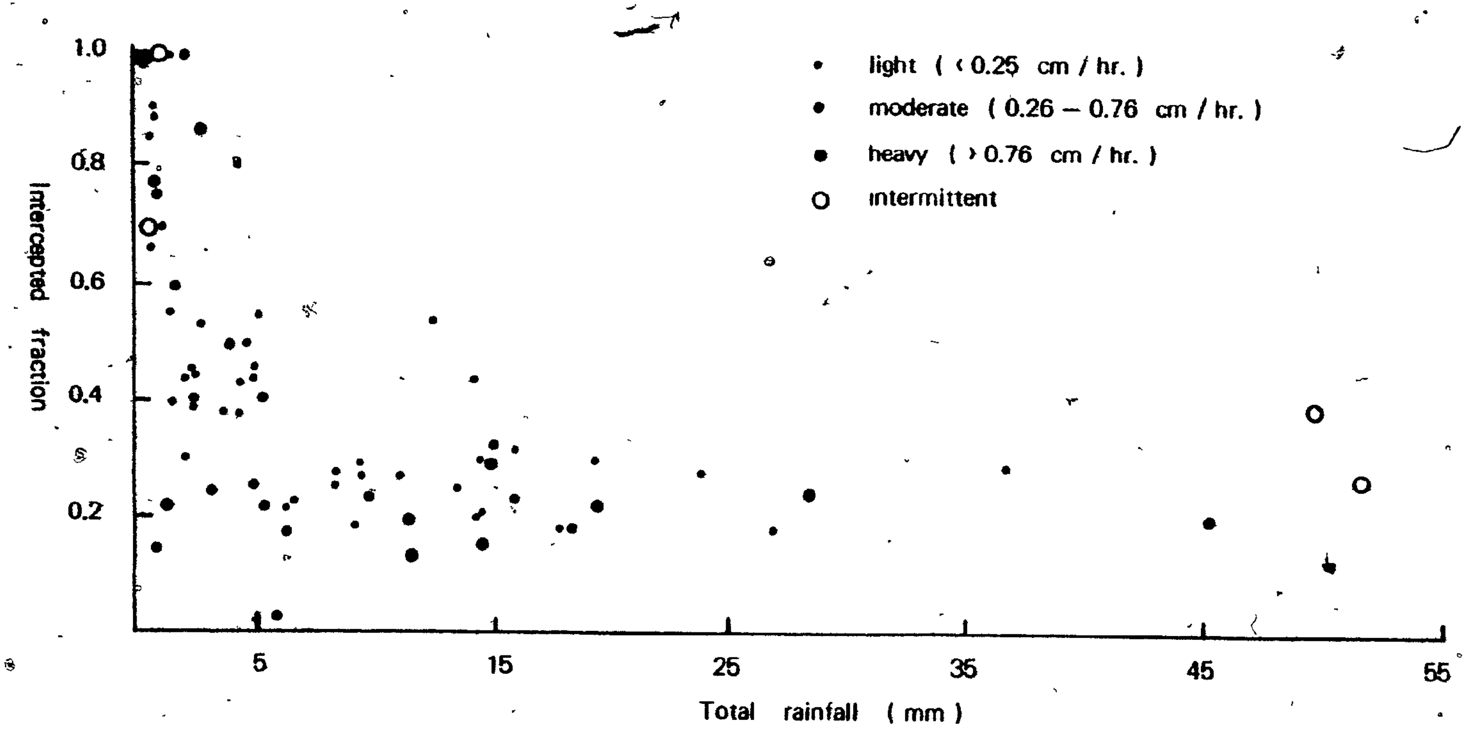

Figure 7.5 Intercepted fraction as a function of rainfall amount and intensity, 1974, 1975.

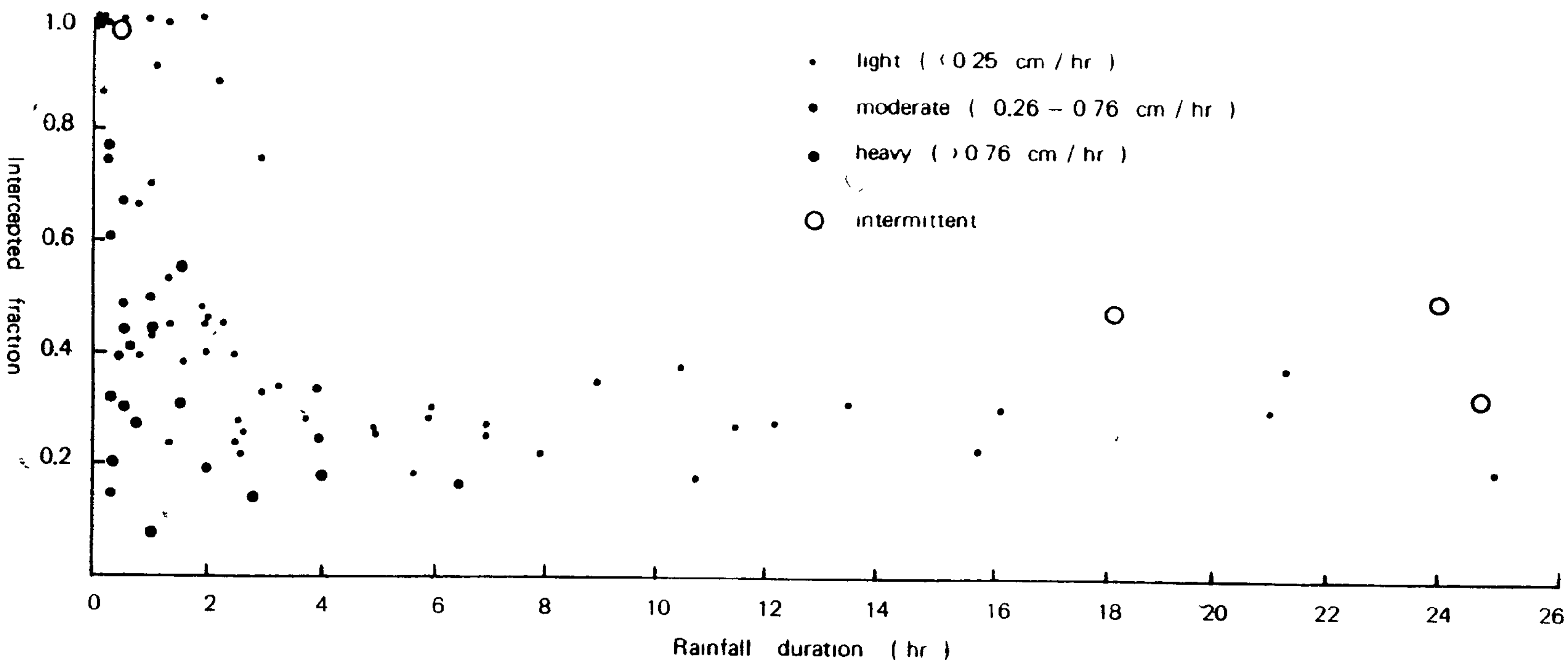

Figure 7.6 Intercepted fraction as a function of rainfall duration and intensity, 1974, 1975.

quite low and reached saturation in a short time period, thereby restricting substantial evaporative losses, and hence interceptional loss, during the rainfall.

Duration of rainfall is also critical in determining the interceptional loss, especially if non-saturation of the ambient air were to prevail throughout the period of wetting. Figure 7.6 shows that in most cases, for rainfalls exceeding storage capacity, the longer lasting light-intensity type storms have a greater portion of the totals intercepted. The interceptional fraction, however, was greatest for very light intermittent showers in that under these circumstances not only was the canopy allowed to dry out partly, but also the evaporating power of the air, as characterized mainly by its saturation deficit, was successively increased, upon temporary cessation of wetting. The short-lasting medium and heavy intensity storms on the other hand usually satisfied canopy capacity almost instantly, at least for storms whose amount exceeded canopy storage, and most of the precipitation reached the ground as throughfall and stemflow.

Rainfall characteristics therefore, namely amount, duration and intensity, can affect the amount of interceptional loss by the way they control evaporation ~~rates~~ during wetting. Rain types however also tend to have characteristic synoptic conditions following rainfall cessation, and these post-wetting weather conditions also tend to have characteristic effects on evaporation of intercepted rainfall.

7.4 Effect of Post-Wetting Weather

Post-wetting weather conditions dictate the rate at which canopy detention storage is depleted. Upon the cessation of rain both the input of incoming solar energy and the vapor pressure deficit of the ambient air normally increase, at least for the daylight period. Clearing following rainfall in the study area, was however found to be rather slow, in that cloudy conditions prevailed, following wetting, for extended periods in the majority of instances. The vapor pressure deficit of the ambient air, especially when windy conditions continually advected warmer and drier air, was therefore found to be more critical than radiation receipt in controlling post-wetting evaporative losses.

An examination of figure 7.7 shows that the evaporation of intercepted rainfall following wetting proceeds at a faster rate than during rainfall. The amount of water to be evaporated is usually of the same order of magnitude as the storage capacity of the vegetation. This intercepted moisture can however be evaporated in from about one to several hours, depending on post-wetting weather conditions.

On some occasions, as can be observed in figure 7.8, calm, cloudy conditions lasted for several hours, upon the cessation of rainfall. As a result, because of limited energy receipt and restricted ventillation, the evaporation of intercepted rainfall proceeded at a rather slow rate, so that the canopy remained wet for an extended period of time.

On other occasions however, especially when the rain had fallen overnight or early morning, relatively clear skies together with unsaturated ambient air, followed wetting of the canopy (see figure 7.9). On these occasions, drying of the canopy was achieved in a relatively short time period, especially when wind conditions provided a continuous supply of moderately dry air.

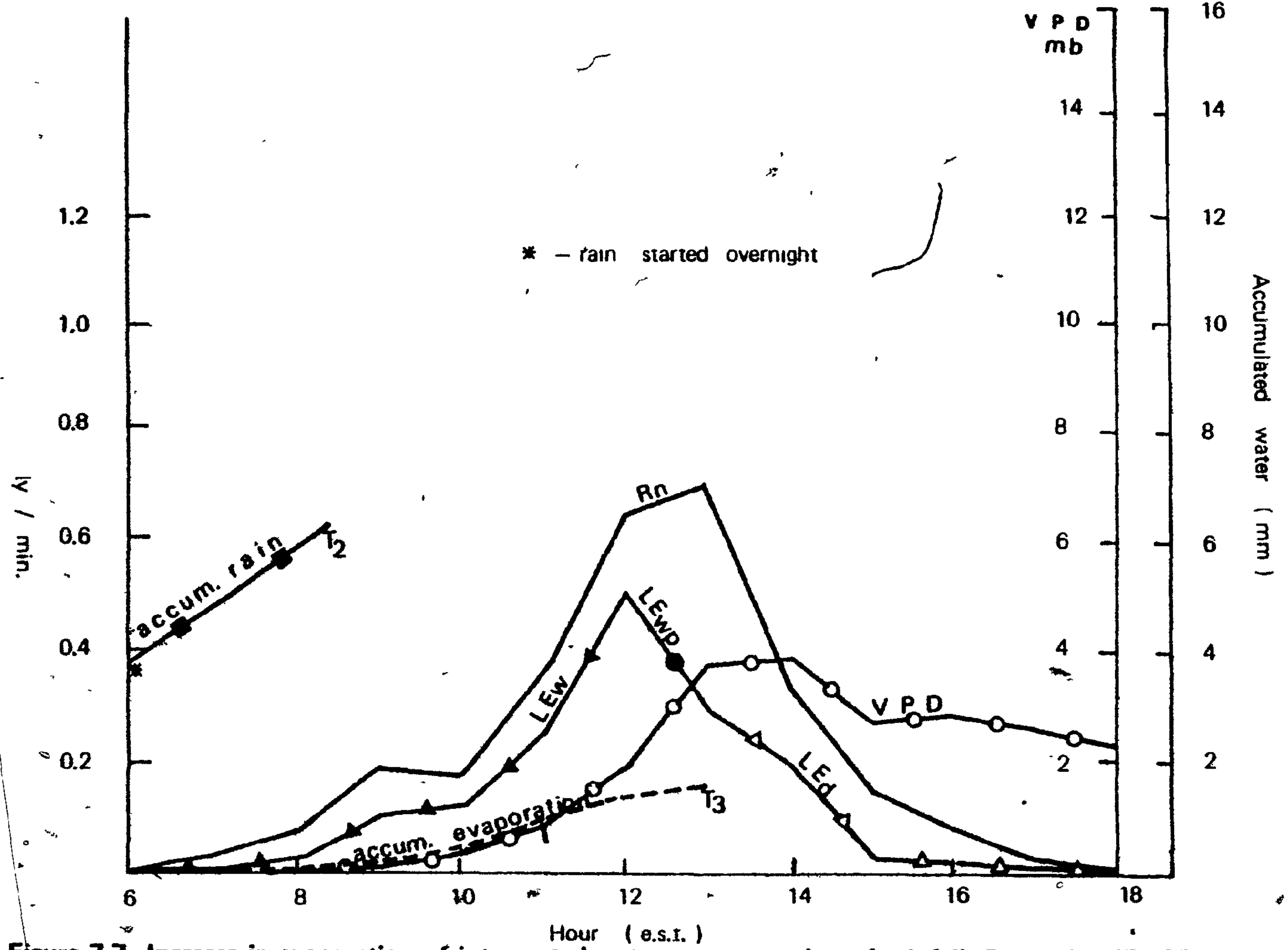

Figure 7.7 Increase in evaporation of intercepted water upon cessation of rainfall, September 17, 1975.

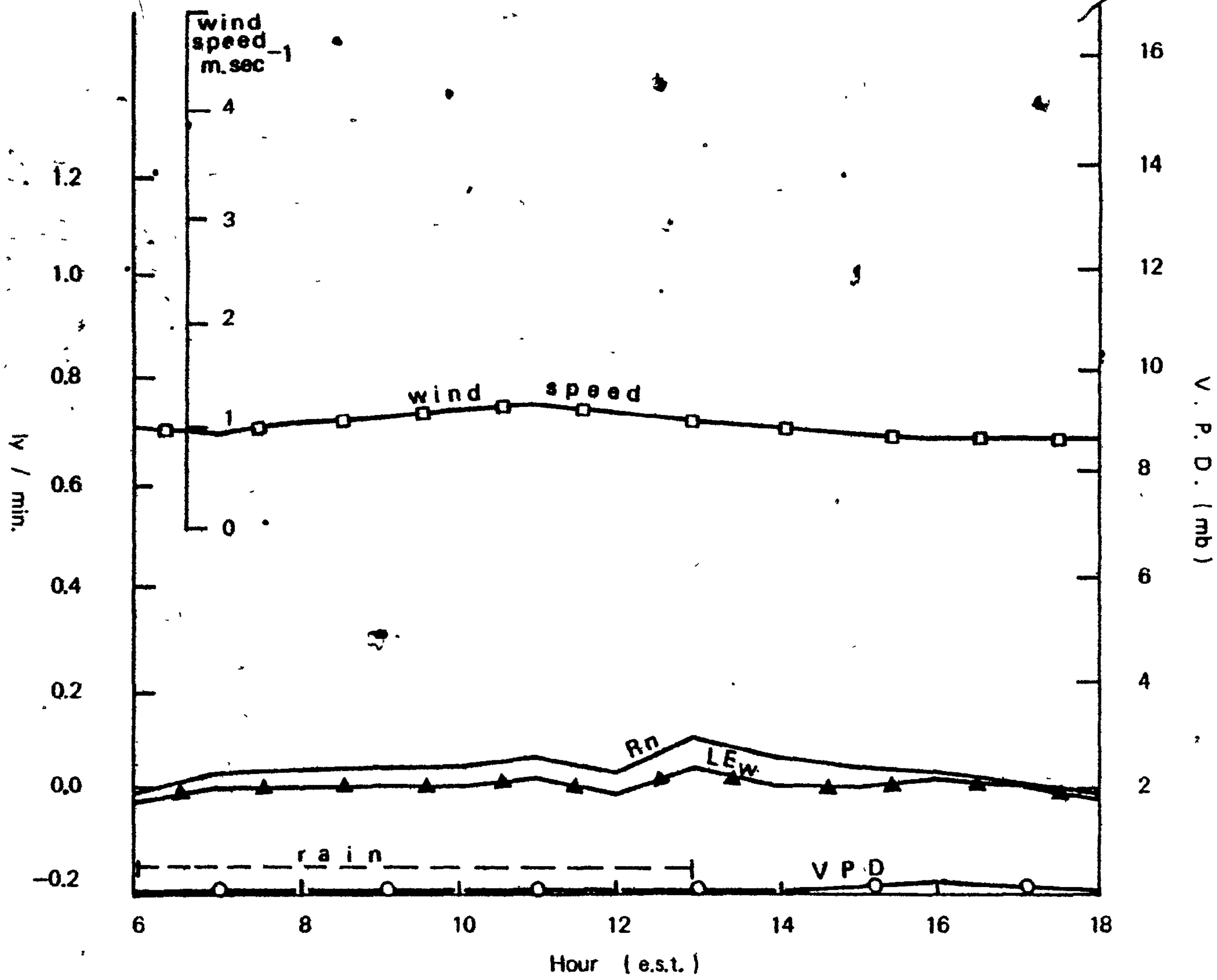

Figure 7.8 Evaporation of intercepted rainfall under calm, cloudy weather, September 19, 1975.

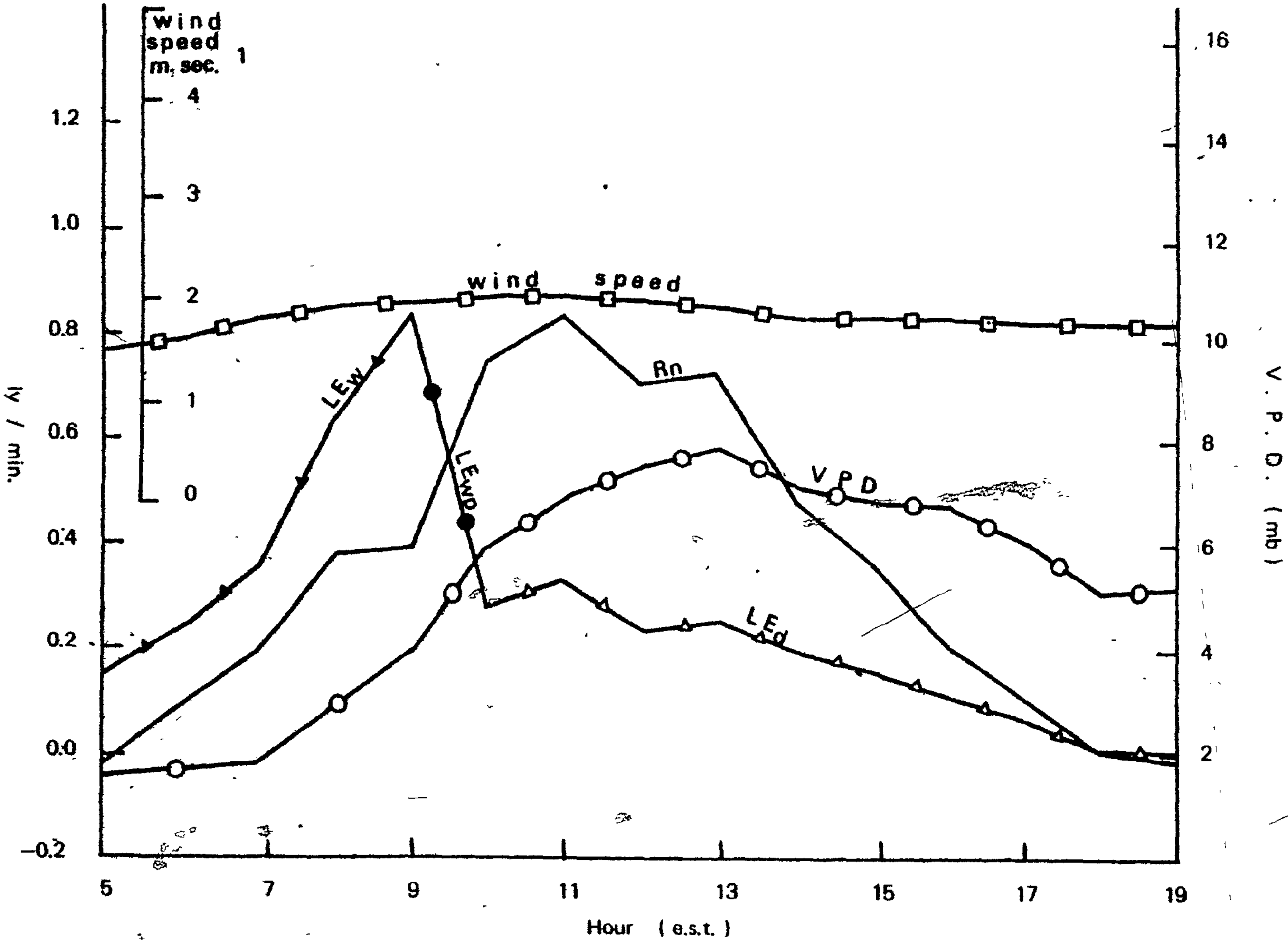

Figure 7.9 Evaporation of intercepted water for a mainly sunny day, following overnight rainfall, June 15, 1975.

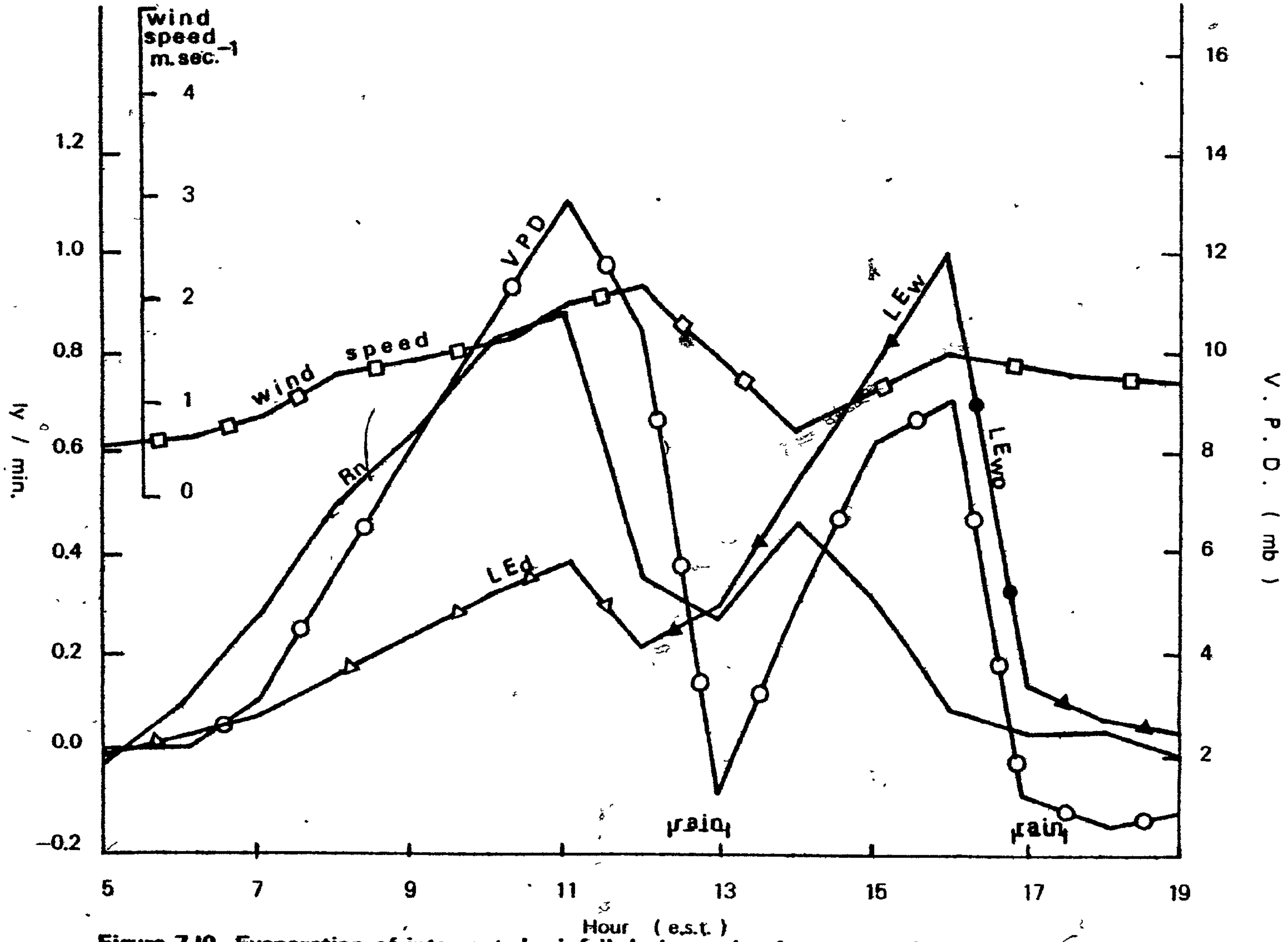

Figure 7.10 Evaporation of intercepted rainfall during early afternoon and early evening, July 19, 1975.

It follows from the above then that the time when wetting occurs is also important. Both figures 7.9 and 7.10 show that when rain cessation occurred during the morning or early afternoon, the canopy usually dried out before nightfall, the same day. But when the rain stopped during the early evening or overnight (see figure 7.10), in most cases, it was not until the following day that the canopy became dry. As can be seen from the early morning hours in figure 7.10, the greater amount of overnight rains were usually preserved on the canopy until the next morning, except in cases where strong winds advected a strong amount of energy for latent heat transfer. Also, in exceptional cases, where separate showers were closely spaced, the canopy was sometimes not allowed to dry out completely before the onset of the following rainfall. In these instances then most of the subsequent rain reached the ground in that canopy storage was already satisfied by the previous rainfall.

It follows from the foregoing discussion that post-wetting weather conditions, as characterized mainly by the saturation deficit of the ambient air, can also affect the rate of evaporation of intercepted rainfall. These weather conditions are also controlled by the timing of wetting, mainly in terms of a day and night basis. The varying rates of evaporation of intercepted rainfall that result then can in turn influence the withdrawal of soil moisture.

7.5 Effect on Soil Moisture

The preceding sections have shown that rain characteristics and weather conditions can significantly affect the nature and magnitude of the evaporation of intercepted rainfall, and hence soil moisture withdrawal. In instances where the canopy remained wet for extended durations, as during and following light intensity rains (see figures 7.1 and 7.8) there was a substantial amount of transpirational saving by the canopy. At the same time however, it must be remembered that there can occur a fair amount of evaporation during this type of rainfall (see figure 7.1). Thus, although some soil moisture is conserved, a lesser proportion of the total rainfall reaches the ground. On the other hand, for short-lasting, moderate and heavy intensity rainfalls (see figure 7.2) there is very little time for evaporation of intercepted rainfall during the rain. Consequently most of the precipitation reaches the ground provided that the rainfall amount is greater than the canopy storage.

Moreover, depending on the post-wetting weather, the duration of soil moisture saving can last from one to several hours. On occasions when the intercepted rain is quickly evaporated there is little saving of soil water. But when the canopy remains wet for an extended time period there is a substantial amount of daily transpirational saving. Short, heavy intensity rains followed by a period of slow evaporation may therefore be the most conducive to the conservation of soil water in that they not only provide a greater supply of soil moisture, but also subdue transpirational withdrawal for an extended period.

In the case of overnight and late evening rains, although evaporation of intercepted moisture is restricted at night, there is nevertheless very little transpirational saving in that stomates are closed at night.

In summary, then it can be stated that rain characteristics, namely amount, duration, intensity and frequency of occurrence contribute to how much of precipitation is withheld from the ground by the vegetation. The efficiency of water loss from a wetted canopy however, is also dictated by post-wetting weather. Both of these factors in turn affect soil moisture consumption and hence the water balance.

Chapter 8

Conclusion

8.1 Summary

In this thesis the problem concerning the effect of intercepted rainfall on evapotranspiration rates, and hence the water balance, over a mixed hardwood forest, has been examined. The primary aim of the research was to compare the amount of intercepted rainfall evaporated from the canopy, with the amount of soil moisture transpired through the leaves, under similar weather conditions. To this end, a model which could be used to gauge latent heat transfer from a wetted or an unwetted canopy, as proposed by Monteith (1965), formed the basis of the method of analysis.

To provide experimental evidence, field measurements were undertaken. All the components of the water balance were either measured directly, or estimated by means of indirect techniques. Since evapotranspiration estimates were to be stressed, and since the surface resistance version of the combination model was being utilized, regular estimates of the canopy resistance (r_c) to vapor diffusion, when the forest was unwetted, were required. To satisfy this need, a model as outlined in chapter 5 was formulated.

The results of the experiment show that under the same weather conditions, and assuming a dry canopy, the rate of evaporation (LE_w) of intercepted rainfall would be several times that of the transpiration (LE_d) of soil water, even when soil moisture is non-limiting. Also the magnitude of this ratio (LE_w/LE_d) is primarily dependent upon the ratio of the canopy to the aerodynamic resistance (r_c/r_a) and prevailing weather conditions. Further analysis of the data shows that because of this differential in latent heat transfer, a substantial amount (> 70 percent) of intercepted rainfall is lost to the soil beneath, and a lesser proportion (< 30 percent) goes towards conserving the transpirational withdrawal of soil moisture.

8.2 Conclusions

In the light of the results from this research, the evaporation of intercepted rainfall can be viewed as constituting more of a direct loss of moisture to soil water depth, than of a saving to the transpirational withdrawal of soil moisture. If the effect of intercepted rainfall is neglected therefore, there can arise a significant underestimation of water consumption, via evaporation, over the forest basin. This neglect can in turn affect the computation of the water balance.

As a result of the present study, therefore it is suggested that there are several improvements or refinements that can be made to certain aspects of the research. In particular, it has been shown that the model for predicting leaf stomatal resistances (r_s) breaks down during periods of hydro-active closure. Thus the need for a more sophisticated model. As mentioned earlier, a possible solution to this problem may lie in examining the temperature and vapor pressure differences between the leaf surface and the ambient air, since leaf temperature can be expected to rise substantially when evaporative cooling ceases.

Also, during periods when the canopy was wet, it was very common for the rate of latent heat transfer to exceed the rate of net radiant energy input. It can be argued either that a wetted canopy acts in such a manner as to promote latent heat transfer at the expense of sensible heat transfer and terrestrial radiation loss, or that a wetted canopy acts as a veritable sink for energy advected from the surrounding area. The relative contribution of these two mechanisms of energy hoarding by wetted vegetation is well worth further investigation.

Finally, a more comprehensive and rigorous treatment needs to be given to the effect of rain characteristics and post-wetting weather on the evaporation of intercepted rainfall than was possible here. This will require taking sample measurements over several growing seasons so as to arrive at more typical parameters for the region in question. Stochastic models can then be developed so that meaningful forecasts of water consumption in forested areas can be attempted in terms of weather conditions and rainfall characteristics.

REFERENCES

Allen, Jr., L. H.; Yocum, C. S. and Lemon, E. R. 1964. Photosynthesis under field conditions VII. Radiant energy exchanges within a corn crop canopy and implications in water use efficiency. Agron. J. 56: 253-259.

Atmospheric Environment Service, Department of Transport, Canada. July, 1961. Manobs, 5th edition.

Barrs, H. D. 1968. Determination of water deficits in plant tissues. In Water deficits and plant growth: 235-368. T. T. Kozlowski (editor). Academic Press.

Bowen, I. S. 1926. The ratio of heat-losses by conduction and by evaporation from any water surface. Phys. Rev. 27: 779-789.

Bultot, F.; Dupriez, G. L. and Bordeaux, A. 1972. Rainfall interception by a forested vegetation; estimation of the daily interception using a mathematical model. J. Hydrol. 17: 193-223.

Burgy, R. H. and Pomeroy, C. R. 1958. Interception losses in grassy vegetation. Trans. Amer. Geophys. Union 39(6): 1095-1100.

Canada Department of Agriculture. 1974. Rev. The system of soil classification for Canada. Publication No. 1455: 255 p.

Chamberlain, A. C. 1966. Transport of gases to and from grass and grass like surfaces. Proc. Roy. Soc. A. 290: 236-260.

Chang, Jen-Hu. 1974. Climate and agriculture; an ecological survey. Chicago: Aldine Publishing Company. 304 p.

Davies, J. A. 1967. A note on the relationship between net and solar radiation. Quart. J. Roy. Met. Soc. 93: 109-115.

Delfs, J. 1967. Interception and stemflow in stands of Norway spruce and beech in West Germany. In Forest hydrology: 179-185. W. E. Sopper and H. W. Lull (editors). Oxford: Pergamon Press.

Dyer, A. J. 1967. The turbulent transport of heat and water vapour in an unstable atmosphere. Quart. J. Roy. Met. Soc. 93: 501-508.

Ehlig, C. F. and Gardner, W. R. 1964. Relationship between transpiration and the internal water relations of plants. Agron. J. 56: 127-130.

Federer, C. A. 1970. Measuring forest evapotranspiration: Theory and problems. U.S.D.A. Forest Service Research Paper NE-165. 25 p.

Frankenberger, E. 1960. Beitrage zun internationalen geophysikalischen Jahr. Ber Deut. Wett. 73(10). 84 p.

Fuchs, M.; Tanner, C. B.; Thurtell, G. W. and Black, T. A. 1969. Evaporation from drying surfaces by the combination method. Agron. J. 61: 22-26.

Gaastra, P. 1959. Photosynthesis of crops as influenced by light, carbon dioxide, temperature and stomatal diffusion resistance. Meded. Land. Wag. 59: 1-68.

Gibson, A. H. 1957. Hydraulics and its applications. London: Constable and Company Ltd.. 813 p.

Goodell, B. C. 1963. A reappraisal of precipitation, interception by plants and attendant water loss. J. Soil and Water Conserv. 18: 231-234.

Grah, R. F. and Wilson, C. C. 1944. Some components of rainfall interception. J. of Forestry 42(12): 890-898.

Harr, R. D. 1966. Influence of intercepted water on evapotranspiration losses from small potted trees. Ph.D. thesis, Colorado State University, Fort Collins.

Helvey, J. D. and Patric, J. H. 1965. Canopy and litter interception of rainfall by hardwoods of the eastern United States. Water Resour. Res. 1(2): 193-206.

Hicks, B. B.; Hyson, P. and Moore, C. J. 1974. A study of eddy fluxes over a forest. J. Appl. Met. 14: 58-66.

Hoover, M. D. 1962. Water action and water movements in the forest. In Forest Influences: 31-80. Rome: U.N., F.A.O Publ..

Horton, R. E. 1919. Rainfall interception. Monthly Weath. Rev. 47(9): 608-623.

Iljin, W. S. 1957. Drought resistance in plants and physiological processes. Ann. Rev. Plant Physiol. 8: 257-274.

Jones, R. L. 1957. The effect of surface wetting on the transpiration of leaves. Physiol. Plant. 10: 281-288.

Kanemasu, E. T.; Thurtell, G. W. and Tanner, C. B. 1969. Design, calibration and field use of a stomatal diffusion porometer. Pl. Physiol. 44(6): 881-885.

Kittredge, J. 1948. Forest influences. New York: McGraw-Hill. 394 p.

Leonard, R. E. 1961. Interception of precipitation by northern hardwoods. U. S. Forest Serv., N. E. Forest Expt. Sta., Paper 159. 16 p.

______________. 1967. Mathematical theory of interception. In Forest hydrology: 131-136.

Leyton, L. and Carlisle, A. 1959. Measurement and interpretation of interception by forest stands. Ass. Int. Hydrol. Sci. 1: 111-119. Hannoversch-Munden.

______________, Reynolds, E. R. C. and Thompson, F. B. 1967. Rainfall interception in forest and moorland. In Forest hydrology: 163-177.

Linsley, R. K. Jr.; Kohler, M. A. and Paulhus, J. L. 1949. Applied hydrology. New York: McGraw-Hill. 689 p.

Law, F. 1957. Evaporation losses in a plantation of sitka spruce. Int. Ass. Sc. Hydrol. 2: 397-411. Toronto.

Meidner, H. 1965. Stomatal control of transpirational water loss. Symp. Soc. Exp. Biol. 19: 185-203.

______________ and Mansfield, T. A. 1968. Physiology of stomata. London: McGraw-Hill. 179 p.

Merriam, R. A. 1960. A note on the interception loss equation. J. Geophys. Res. 65(11): 3850-3851.

McMillan, W. D. and Burgy, R. H. 1960. Interception loss from grass. J. Geophys. Res. 65: 2389-2394.

McNaughton, K. G. 1976a. Evaporation and advection I: Evaporation from extensive homogeneous surfaces. Quart. J. Roy. Met. Soc. 102: 181-191.

______________. 1976b. Evaporation and advection II: Evaporation downwind of a boundary separating regions having different surface resistances and available energies. Ibid: 193-202.

______________ and Black, T. A. 1973. A study of evapotranspiration from a Douglas fir forest using the energy balance approach. Wat. Resour. Res. 9(6): 1579-1590.

Monteith, J. L. 1963. Gas exchange in plant communities. In Environmental control of plant growth: 95-112. T. L. Evans (editor). New York: Academic Press.

______________. 1965. Evaporation and environment. In The state and movement of water in living organisms. Symp. Soc. Expt. Biol. 19: 205-234. Swansea.

______________. 1973. Principles of environmental physics. London: Edward Arnold. 241 p.

Munn, R. E. 1961. Energy budget and mass transfer theories of evaporation. Proc. Hydrol. Symp. No. 2: 8-30. Department N. Affairs and Natl. Res..

Murphy, C. E. 1970. Energy sources for the evaporation of precipitation intercepted by tree canopies. Ph.D. thesis, Duke University, Durham, N.C..

______________ and Knoerr, K. R. 1975. The evaporation of intercepted rainfall from a forest stand: An analysis by simulation. Wat. Resour. Res. 11(2): 273-280.

Patric, J. H. 1966. Rainfall interception by mature coniferous forests of S. E. Alaska. J. Soil Wat. Conserv. 21: 229-231.

Penman, H. L. 1948. Natural evaporation from open water, bare soil and grass. Proc. Roy. Soc. London A 193: 120-145.

______________ and Schofield, R. K. 1951. Some physical aspects of assimilation and transpiration. Symp. Soc. Expt. Biol. 5: 115-129.

______________. 1952. The physical bases of irrigation control. Rept. 13th Int. Hort. Congr. 12 p.

______________. 1955. Evaporation: An introductory survey. Neth. J. Agr. Sci. 4(1): 9-29.

______________. 1963. Vegetation and hydrology. Tech. Comm. No. 53. England: Farnham Roy., Comm. Bur. of Soils.

Phillips, W. J. 1972. An evaluation of the relative importance index to the study of forest vegetation on Mont St. Hilaire, Quebec, Canada. M.Sc. thesis, McGill University, Montreal, Canada

Polavarapu, R. J. 1970. A comparison study of global and net radiation measurements at Guelph, Ottawa and Toronto. J. Appl. Meteor. 9: 809-813.

Rijtema, P. E. 1965. An analysis of actual evapotranspiration. Agr. Res. Repts. Wageningen 659. 107 p.

Robins, P. C. 1974. A method of measuring the aerodynamic resistance to the transport of water vapor from forest canopies. J. Appl. Ecol. 11: 315-325.

Rogerson, T. L. and Byrnes, W. R. 1968. Net rainfall under hardwoods and red pine in central Pennsylvania. Wat. Resour. Res. 4(1): 55-57.

Rouse, W. R. 1965. Aspects of a forest microclimate. Ph.D. thesis, McGill University, Montreal, Canada.

Rowe, P. B. and Hendrix, T. M. 1951. Interception of rain and snow by second-growth ponderosa pine. Trans. Amer. Geophys. Un. 32(6): 903-908.

Rutter, A. J. 1959. Evaporation from a plantation of Pinus Sylvestris in relation to meteorological and soil conditions. Ass. Int. Hydrol. Sci. 48: 101-110. Hannoversch-Munden.

______________. 1963. Studies in the water relations of Pinus Sylvestris in plantation conditions. J. Ecol. 51: 191-203.

______________. 1967. An analysis of evaporation from a stand of scots pine. In Forest hydrology: 403-417.

______________. 1968. Water consumption by forests. In Water deficits and plant growth 2: 23-84.

______________, Kershaw, K. A.; Robins, P. C. and Morton, A. J. 1971/72. A predictive model of rainfall interception in forests; 1. derivation of the model from observations in a plantation of Corsican pine. Agric. Meteor. 9: 367-384.

______________, Morton, A. J. and Robins, P. C. 1975. A predictive model of rainfall interception in forests; II. generalization of the model and comparison with observations in some coniferous and hardwood stands. J. Appl. Ecol. 12(1): 367-380.

Shindel, H. L. 1963. The effects of intercepted water on the transpiration of red oak seedlings at different levels of soil moisture. M. S. thesis, Pa. St. Univ., Univ. Park, Pa..

Sellers, W. D. 1965. Physical climatology. Chicago and London: U. of Chicago Press. 272 p.

Singh, B. 1976. Wind profile estimates for a hardwood foest. Climatol. Bull. (19): 17-22.

Slatyer, R. O. 1967. Plant-water relationships. London and New York: Academic Press. 366 p.

______________ and McIlroy, I. C. 1961. Practical microclimatology. CSIRO, Melbourne, Australia. 310 p.

St. Laurent, H. 1973. Build a moisture sensor; detect presence of water in any location. In Electronic experimenter's handbook: 22-24. Park Avenue, New York: Ziff-Davis Publishing Company.

Stewart, J. B. and Thom, A. S. 1973. Energy budgets in pine forest. Quart. J. Roy. Met. Soc. 99: 154-170.

Swinbank, W. C. 1951. The measurement of vertical transfer of heat and water vapor by eddies in the lower atmosphere. J. Meteor. 8: 135-145.

______________ and Dyer, A. J. 1967. An experimental study in micrometeorology. Quart. J. Roy. Met. Soc. 93: 494-500.

Szeicz, G. and Long, I. F. 1969. Surface resistance of crop canopies. Wat. Resour. Res. 5(3): 622-633.

______________; Endrodi, G. and Tajchman, S. 1969. Aerodynamic and surface factors in evaporation. Wat. Resour. Res. 5(2).

______________; Van Bavel, C. H. M. and Takami, S. 1973. Stomatal factors in the water use and dry matter production by sorghum. Agric. Meteor. 12: 361-389.

Tanner, C. B. and Fuchs, M. 1968. Evaporation from unsaturated surfaces, a generalized combination method. J. Geophys. Res. 73: 1299-1304.

Thom, A. S.; Stewart, J. B.; Oliver, H. R. and Gash, J. H. 1975. Comparison of aerodynamic and energy budget estimates of fluxes over a pine forest. Quart. J. Roy. Met. Soc. 101: 93-105.

Thornthwaite, C. W. and Holzman, B. 1942. Measurement of evaporation from land and water surfaces. Tech. Bull. (817). Washington, D. C.: U. S. Dept. of Agric. 75 p.

Thorud, D. B. 1967. The effect of applied interception on transpiration rates of potted ponderosa pine. Wat. Resour. Res. 3(2): 443-450.

Van Bavel, C. H. M.; Nakayama, F. S. and Ehrler, W. L. 1965. Measuring transpiration resistance of leaves. Plant Physiol. 40: 535-540.

Virgin, H. I. 1956. Light-induced stomatal movement in wheat leaves recorded as transpiration; experiments with error hygrometer. Plant Physiol. 9: 280-303.

Waggoner, P. E.; Begg, J. E. and Turner, N. C. 1969. Evaporation of dew. Agric. Meteor 6: 227-230.

Wilm, H. G. and Niederhof, C. H. 1941. Interception of rainfall by mature lodgepole pine. Trans. Amer. Geophys. Un. 22: 660-665.

Wilson, R. G. 1968. Topographic influences on a forest microclimate. Ms.C. thesis, McGill University, Montreal, Canada.

World Meteorological Organization. 1966. Measurement and estimation of evaporation and evapotranspiration. Tech. Note (83). Geneva, Switzerland. 121 p.

Zinke, P. J. 1967. Forest interception studies in the United States. In Forest hydrology: 137-161.

Addendum

Gray, D. M. (ed.). 1970. Handbook on the Principles of Hydrology. Secretariat, Can. Nat. Comm. Int. Hydrol. Dec. Sec. VII.