Environmental Research Center Papers

Number 4

1984

Environmental Research Center
The University of Tsukuba

STUDY ON EVAPOTRANSPIRATION FROM PASTURE*

By Shinji Nakagawa

Environmental Research Center
The University of Tsukuba
305 Ibaraki, Japan
(received 21 December, 1983)

ABSTRACT

Since the observation of evapotranspiration over a long period is difficult, several estimation formulae and concepts for evapotranspiration have been proposed. Indiscriminate applications of these concepts and formulae lead to confusion concerning definitions and their use. Intensive micrometeorological observation data and long term evapotranspiration data were collected and analyzed.

Firstly, potential evapotranspiration, equilibrium evaporation, and potential evaporation were taken as the representative concepts of evapotranspiration. The conditions to which they can be applied were investigated, based on the evapotranspiration data obtained over a field of actively growing pasture with no soil water shortage. The potential evapotranspiration, defined by a vapor-saturated surface condition, can be applied only when the vegetation canopy behaves as a completely wetted surface during dew evaporation, or when the over-passing air is humid. Otherwise, the potential evapotranspiration by this definition becomes greater than the potential evapotranspiration from actively growing pasture under an ample soil water condition. The potential evapotranspiration, defined by a vapor-saturated surface condition, and that defined by a well-watered soil condition cannot be considered to have the same meaning.

Hourly evapotranspiration from pasture is found to fall in a range of 1.0 to 1.26 times as large as the equilibrium evaporation, and on the average, the former is 1.16 times as large as the latter. The potential evaporation, which is defined as 1.26 times as large as the equilibrium evaporation, can be applied only to the pasture canopy which is in a completely wetted condition caused by dewfall. The upper and the lower limits of evapotranspiration from actively growing pasture, without soil water shortage, are found to be expressed by the potential evaporation and by the equilibrium evaporation, respectively.

Secondly, a simple equilibrium evaporation model was developed and tested using evapotranspiration data measured by a weighing lysimeter. In summer, the daily evapotranspiration from pasture can be estimated by the equilibrium evaporation model, with the proportional

* Doctor of Science Thesis in the Institute of Geoscience, the University of Tsukuba.

constant equal to 1.14. The distinctive seasonal trend, which may reflect the activity of pasture, is found in the relationship between the daytime evapotranspiration and the daytime equilibrium evaporation. The equilibrium evaporation model is shown to estimate the annual evapotranspiration within 10% accuracy if the appropriate values of the proportional constant are used in the model.

Thirdly, the Thornthwaite and the Penman methods for potential evapotranspiration were tested, using the actual evapotranspiration data measured by a weighing lysimeter. The Thornthwaite method is shown to overestimate summer evapotranspiration and underestimate winter evapotranspiration. Also the method overestimates annual evapotranspiration by about 30%. On the other hand, the potential evapotranspiration by the Penman method agrees well with the actual evapotranspiration in summer and autumn, but the former is greater than the latter in winter and spring. The annual evapotranspiration estimated by the Penman method exceeds the actual evapotranspiration by around 20%. Annual growing cycle of vegetation and climatic conditions play important roles in the determination of evapotranspiration.

CONTENTS

LIST OF FIGURES

LIST OF TABLES

LIST OF SYMBOLS

a, a', a''	nondimensional constant
b, b', b''	nondimensional constant
c_p	specific heat of air at constant pressure
D	day-length factor
D_a	wet-bulb depression in the air
D_o	wet-bulb depression at an evaporating surface
d	zero-plane displacement height
E	evapotranspiration
E_a	drying power of the air
E_a'	ventilation term in the equation of potential evapotranspiration defined by a vapor-saturated surface condition
E_{eq}	equilibrium evaporation
E_{es}	estimated evapotranspiration
E_{ly}	evapotranspiration measured by a weighing lysimeter
E_o	hypothetical open water evaporation
E_p	potential evaporation
E_{Pen}	potential evapotranspiration by the Penman method
E_T	potential evapotranspiration
E_{Th}	potential evapotranspiration by the Thornthwaite method
e_a	vapor pressure of the air
$e_a{}^*$	saturation vapor pressure of the air
f	reduction factor in the Penman method
G	soil heat flux
H	sensible heat flux
h	turbulent transfer coefficient
K_E	turbulent diffusivity for water vapor
K_H	turbulent diffusivity for heat
k	Karman's constant
$L\downarrow$	downward long-wave radiation flux
$L\uparrow$	upward long-wave radiation flux
L^*	net long-wave radiation flux
N	number of daylight hours
n	number of hours of bright sunshine
PE	potential evapotranspiration defined by a vapor-saturated surface condition
p	atmospheric pressure
R'	decreasing rate of daily available energy by nighttime outgoing radiation
R_A	extraterrestrial radiation
R_n	net radiation flux
$R_n - G$	available energy

$R_n - G^*$ available energy in evaporation equivalence
R_{no} net radiation flux at a hypothetical open water surface in the Penman method
$R_{no}{}^*$ net radiation flux at a hypothetical open water surface in the Penman method in evaporation equivalence
$R_{nT}{}^*$ net radiation flux at a vegetation surface in evaporation equivalence
r_{H} aerodynamic resistance for the transfer of heat
r_{M} aerodynamic resistance for the transfer of momentum
r_{a} aerodynamic resistance
r_{c} canopy resistance
r_{cm} minimum canopy resistance under an ample soil water condition
r_{s} surface resistance
S stomatal factor
$S\downarrow$ downward short-wave radiation flux
$S\uparrow$ upward short-wave radiation flux
T temperature
T_{a} temperature in Kelvin
T_{g} soil temperature
u wind speed
u_* friction velocity
X_{mea} measured quantity
X_{es} estimated quantity
z height
z_0 roughness length

α proportional constant in the equilibrium evaporation model
α_{a} albedo
α_{d} proportional constant in the equilibrium evaporation model using daytime data
α' increasing rate of the proportional constant in the equilibrium evaporation model
$\bar{\alpha}$ proportional constant in the equilibrium evaporation model using daily data
$\bar{\alpha}_*$ proportional constant in the equilibrium evaporation model using daily data under the condition $\lambda E < R_n - G$
β Bowen's ratio
γ psychrometric constant
Δ slope of the saturation vapor pressure curve
δ relative estimation error
ϵ ratio of mole weights of water vapor to air
λ latent heat for vaporization
ρ density of the air
σ Stefan-Boltzmann constant
τ shearing stress

CHAPTER 1

INTRODUCTION

Evaporation is the physical process by which a substance is converted from a liquid or solid phase into a vapor state. In the case of a solid substance, the process is referred to as sublimation. The process of vaporization of water that has passed through the stomata of vegetation is called transpiration. Over a large land area such as a basin, transpiration from plants and direct evaporation from soil, water surface, and intercepted water on plant leaves occur simultaneously, so that it is difficult to separate them in computation. Therefore, the term evapotranspiration is used to describe the combined processes of water transfer from land surface into the atmosphere.

These distinctions are all useful at times. Recently, however, the term evaporation has often been used to describe all processes of vaporization unless specified otherwise.

In natural environments, evaporation of water is one of the main components of the hydrological cycle. Water entering into the evaporation phase of the hydrological cycle becomes unavailable for further use by plants and human activities. Therefore, accurate knowledge of its consumptive use through evaporation is indispensable for the planning and management of water resources.

In addition, evaporation is a connecting link between the water budget and the energy budget. Probably the most important physical effect of evaporation is the cooling that occurs at an evaporating surface. This prohibits an extreme temperature rise. Evaporation as latent heat flux provides one of the mechanisms of dissipation of heat received at the earth's surface and it plays a crucial role in governing weather and climate.

1-1 Review of recent studies

Potential evapotranspiration

Direct measurement of evapotranspiration over long time periods is difficult because of the lack of routinely usable instruments. As a result, many estimation methods for evapotranspiration have been developed and several concepts of evapotranspiration have been proposed. Of these concepts of evapotranspiration, the earliest one is the "potential evapotranspiration" proposed by Thornthwaite in 1944 (Thornthwaite and Mather, 1955). At first, Thornthwaite (1948) considered that when water supply to vegetation increases, evapotranspiration rises to a maximum that depends only on climate. He defined this evapotranspiration as "potential evapotranspiration". Later Thornthwaite and Mather (1955) defined "potential evapotranspiration" as the amount of water which will be lost from a surface completely covered with vegetation if there is sufficient water in the soil at all times for use by vegetation. In his study on the classification of climate, Thornthwaite (1948) proposed an empirical formula for potential evapotranspiration, based on catchment area data and controlled experiments. Since the Thornthwaite formula for potential evapotranspiration is very simple, that is, only air temperature and day-length factor are necessary as climatic data, it has been discussed in detail on a number of occasions. Under conditions where air temperature and net radiation availability are closely related, the Thornthwaite formula works quite effectively; on other conditions it is more or less unsatisfactory (e.g., Sellers, 1964; Kayane and Kobayashi, 1973).

Another semi-empirical approach for the estimation of "potential evapotranspiration" was developed by Penman (1948). Analyzing evapotranspiration data measured by water-filled and turf-planted cylinders (120 cm in depth and 75 cm in diameter), Penman (1948) obtained an evaporation rate from turf with adequate water supply as a fraction of that from open water. The fraction for turf showed a seasonal change attributed to the annual cycle of length of daylight. Penman tried to link the evaporation rate to the available energy at a surface, and to the effective ventilation of a surface by air moving over it. Then, he proposed the so-called "combination equation", in which both energy balance and aerodynamic terms appear in a single relationship.

Penman (1950) used the term "potential transpiration" describing evapotranspiration that would take place from an area if it were covered by green vegetation with no shortage of water. Penman (1963) stated as follows: the terms "consumptive use" by Blaney and Criddle (Blaney, 1954), "potential evapotranspiration" by Thornthwaite (1948), and "potential transpiration" by Penman (1950) have, implicitly or explicitly, a common concept, namely that when a full crop cover is kept plentifully supplied with water, the rate at which the water is transpired is dictated primarily by the weather, with plant and soil factors playing only secondary roles. Hence, the term "potential evapotranspiration" is most commonly used nowadays.

The Penman method for potential evapotranspiration requires two steps. The first step is the determination of a hypothetical open water evaporation E_o,

$$E_o = \frac{\Delta R_{no}{}^* + \gamma E_a}{\Delta + \gamma} \qquad (1\text{-}1)$$

where $R_{no}{}^*$ is the net radiation at a hypothetical open water expressed in evaporation equivalence, E_a the expression for the drying power of the air, Δ the slope of the saturation vapor pressure curve, and γ the psychrometric constant. At first, Penman (1948) proposed the following form of E_a :

$$E_a = (e_a{}^* - e_a)\, f(u) \qquad (1\text{-}2)$$

where $e_a{}^*$ is the saturation vapor pressure at air temperature (mb), e_a the vapor pressure of the air (mb), and $f(u)$ the wind function of the form

$$f(u) = 0.26\ (1 + 0.54u) \qquad (1\text{-}3)$$

where u is the wind speed measured at a height of 2 m (m/s). Later, using results from the Lake Hefner studies, Penman (1956) modified the wind function $f(u)$ to

$$f(u) = 0.26\ (0.5 + 0.54u) \qquad (1\text{-}4)$$

Although Penman later felt that Eq. (1-3) was preferable over Eq. (1-4), Eq. (1-4) is still widely used in hydrological practice (Brutsaert and Stricker, 1979).

The second step is a conversion of E_o to potential evapotranspiration (E_T) by a factor $f = E_T/E_o$ based on experiments. Values of f obtained in southern England are shown in Table 1-1.

Table 1-1 Reduction factor (f) in the Penman method for potential evapotranspiration.

	f
Mid-winter (November ~ February)	0.6
Spring and autumn (March ~ April, September ~ October)	0.7
Mid-summer (May ~ August)	0.8
Whole year	0.75

To account for differences between hypothetical open water evaporation and evapotranspiration from a vegetated surface, Penman (1949) introduced an empirical 'root constant' to be specified for each vegetation type. Penman and Schofield (1951) considered the reasons for the differences as (1) the higher albedo of vegetation, (2) closure of stomata at night, and (3) diffusion impedance of stomata. They tried to evaluate these effects by introducing new parameters, that is, stomatal factor (S) and day-length factor (D). According to Penman (1963), Penman developed the following equation in 1952 :

$$E_T = (\frac{\Delta}{\gamma} R_{nT}^* + E_a) / (\frac{\Delta}{\gamma} + \frac{1}{SD}) \tag{1-5}$$

where R_{nT}^* is the net radiation over vegetation expressed in evaporation equivalence. Businger (1956) derived the wind function which accounts for the surface roughness and the atmospheric stability, and indicated that when a wind function with a rational basis is used, there is no need for the empirical factor f or the additional terms S and D in Eq. (1-5). Tanner and Pelton (1960) used a wind function suggested by Businger (1956), based on the aerodynamic roughness of lucerne, to obtain good estimates of evapotranspiration from irrigated lucerne. Van Bavel (1966) re-emphasized the need to treat crops as aerodynamically rough surfaces and developed a combination method for estimating potential evapotranspiration, with a theoretically based wind function. Van Bavel tested the method against lysimetrically measured evaporation from open water, wet soil, and well-watered alfalfa to obtain good agreement between calculated and measured values for 24-hr totals as well as for hourly basis. Van Bavel stated that his combination equation yielded good results in spite of large amounts of advection that occurred over alfalfa. Davies and McCaughey (1968) also showed that the Penman equation can be applied to the estimation of hourly or daily evapotranspiration, with measured net radiation over vegetation and appropriate turbulent transfer expressions.

Penman (1963) stated that the direct estimation of potential evapotranspiration can be made by Eq. (1-5) with $SD = 1$. As the basis for the above opinion by Penman, Rijtema (1965) demonstrated that both the magnitude of f and its seasonal behaviour are consistent with the difference in albedo between an open water surface and ground covered with short vegetation. A method using net radiation over vegetation, instead of that over open water, has also been widely used for the calculation of potential evapotranspiration.

Soil water content and evapotranspiration

A close relationship between the amount of evapotranspiration and plant growth has been pointed out. For example, Tanner and Pelton (1960) stated that maximum yield of many crops occurs under non-limiting soil water conditions. Davies and McCaughey (1968) demonstrated a simple relationship between dry matter productivity of crop and cumulative evapotranspiration. The term "potential evapotranspiration", therefore, has been widely used for crop irrigation planning and control. Many studies have been done to determine the relationship between evapotranspiration and soil water content. These studies used the evaporation ratio, the ratio of actual evapotranspiration to potential evapotranspiration, as the basis for the analyses.

Several different opinions have been presented concerning the relationship between the evaporation ratio and soil water content. These opinions are that evapotranspiration is not limited by soil water content until either a permanent wilting point (Veihmeyer and Hendrickson, 1955a, 1955b; Veihmeyer, 1972), field capacity (Thornthwaite and Mather, 1955), or some water content between permanent wilting point and field capacity (Budyko, 1956; Holmes and Robertson, 1959; Marlatt et al., 1961; van Bavel, 1967; Priestley and Taylor, 1972; Ritchie et al., 1972; Ritchie, 1973) is reached. Decreasing patterns of the evaporation ratio with decreasing soil water content can be divided into two groups, that is, linearly and non-linearly decreasing patterns (Holmes and Robertson, 1959; Kayane, 1967; Slatyer, 1967; Rutter, 1975).

Denmead and Shaw (1962) first showed that the evaporation ratio depends on the value of potential evapotranspiration, as well as soil water content. They found that evapotranspiration is maintained at the potential rate even in dry soil if the potential evapotranspiration is low enough. In contrast with this, it may be less than the potential rate, even in wet soil, if the potential rate is high enough. Shepherd (1972) and McNaughton and Black (1973) found stomatal control of evapotranspiration even in wet soil. On the other hand, Ritchie (1973) found that the magnitude of potential evapotranspiration over corn does not influence the evaporation ratio.

It has been proved from many studies that species, canopy structure, growing stage, root development, and soil type cause different responses of the evaporation ratio to the soil water content. Therefore, it can be considered that different species growing on different soils in different climates would limit evapotranspiration at different values of soil water content.

Physiological control of evapotranspiration

As advances in knowledge of physiological feature of vegetation occurred, the physiological parameters came to be included in the equation of evapotranspiration. By an analogy to Ohm's law in electricity, Monteith (1963) introduced stomatal resistance (r_s) and aerodynamic resistance (r_a) for the water flow from leaf stomatal cavities to the atmosphere. By considering stomatal resistance of all leaves acting in parallel, Monteith (1965) derived the combination formula :

$$\lambda E = \frac{\Delta(R_n - G) + \rho c_p \,(e_a{}^* - e_a) \,/\, r_a}{\Delta + \gamma\,(1 + r_c \,/\, r_a)} \qquad (1\text{-}6)$$

where λ is the latent heat for vaporization, E the evapotranspiration, R_n the net radiation, G the soil heat flux, ρ the density of air, c_p the specific heat of air at constant pressure, $e_a{}^*$ the saturation vapor pressure at air temperature, e_a the vapor pressure of air, and r_c the canopy resistance. The potential evapotranspiration from freely transpiring vegetation can be obtained with $r_c = 0$ (Thom, 1975).

Equation (1-6) has important significance in the understanding of the interactions of the soil-plant-atmosphere system. The term "surface resistance" is often used as a synonym for "canopy resistance" in many occasions, but surface resistance has a more general meaning as the parameter which represents the surface control on evapotranspiration.

The important control of canopy resistance on evapotranspiration has been pointed out (e.g., Monteith et al., 1965; Thom, 1972; Bailey and Davies, 1981). Canopy resistance has been shown to be largely of a physiological origin and vary substantially within a plant (Szeicz and Long, 1969; Szeicz et al., 1969; Black et al., 1970; Szeicz et al., 1973; Sinclair et al., 1976; Tan and Black, 1976). Szeicz et al. (1973), Brady et al. (1975), Tan and Black (1976), Federer (1979), and Bailey and Davies (1981) related canopy resistance to environmental factors. Thom and Oliver (1977) re-analyzed the ventilation term in the Penman equation (Eq. 1-1) and proposed a modified equation for regional evaporation by adding the surface resistance.

Through the accumulation of field data, the nature of canopy resistance has been found to differ not only between different species, but also between different genetic strains of the same species (Shimshi and Ephrat, 1975; Jones, 1976) and the stage of crop development (Nkemdirim, 1976). The canopy resistance plays an important role in explaining the relationship between the evaporation ratio and soil water content. The response of canopy resistance to environments, however, has not been fully understood, especially for a long period such as a month or a year.

New concepts on evapotranspiration

As the detailed understanding of the evapotranspiration mechanism has progressed, some complicated models, which treat the water transfer in the soil-plant-atmosphere system as a catenary process, have been proposed (Goldstein et al., 1974; Swift et al., 1975; Deardorff, 1978; Federer, 1979). With an increase in model complexity, the data requirements to drive the equations often make the model useless for field applications. Consequently, there are continual efforts to make empirical substitutes to satisfy local conditions and to develop more simple, physically based models.

The concept and model of the "equilibrium evaporation" by Slatyer and McIlroy (1961) is one example. They proposed another type of combination equation for evapotranspiration, that is,

$$\lambda E = \frac{\Delta}{\Delta + \gamma}(R_n - G) + \rho c_p h (D_a - D_o) \qquad (1\text{-}7)$$

where h is the turbulent transfer coefficient and D_a and D_o the wet-bulb depressions, in overlying air and at the evaporating surface, respectively. Slatyer and McIlroy (1961) considered that over a very large homogeneous and moist surface, under well established steady state condition, an air mass passing over the evaporating surface adjusts its temperature and humidity profiles to a new surface, resulting in the air becoming saturated. They named the evapotranspiration in this case "equilibrium evaporation", which can be expressed by setting $D_a = D_o$ in Eq. (1-7). Hence, equilibrium evaporation (E_{eq}) is given by

$$\lambda E_{eq} = \frac{\Delta}{\Delta + \gamma}(R_n - G) \tag{1-8}$$

The equilibrium evaporation is considered to occur in two very different environments: (1) in a very dry environment when $D_a = D_o \mp 0$ and (2) in a very humid environment when $D_a = D_o = 0$. Monteith (1965) noticed these two possibilities. Tanner and Fuchs (1968) stated that Eq. (1-8) represents evapotranspiration into a moist air. Slatyer and McIlroy (1961) and Denmead and McIlroy (1970) concluded that Eq. (1-8) represents the lower limit of the potential evapotranspiration.

Priestley (1959) also developed the same equation as Eq. (1-8) in a different way. Priestley (1959) assumed that, when the air over a moist surface is vapor saturated, and the ranges of variations of temperature and saturated vapor pressure are not too large, saturated vapor pressure is linearized, so that Bowen's ratio (β) in the form of

$$\beta = \frac{\gamma}{\Delta} \tag{1-9}$$

is obtained. Equations (1-8) and (1-9) represent the same phenomena.

Although Eq. (1-8) was anticipated to have limited applications, Denmead and McIlroy (1970), Davies (1972), and Wilson and Rouse (1972) have shown that the equilibrium evaporation gives a satisfactory approximation to the evapotranspiration from fairly dry surfaces.

Another simple formula was proposed by Priestley and Taylor (1972). They took the equilibrium evaporation as the basis for the estimation of "potential evaporation", the evapotranspiration from a horizontally uniform saturated surface with a minimal advection. They proposed the following equation for potential evaporation (E_p) :

$$\lambda E_p = \alpha \frac{\Delta}{\Delta + \gamma}(R_n - G) = \alpha \lambda E_{eq} \tag{1-10}$$

where α is a constant. Analyzing evapotranspiration data obtained over several open water and saturated land surfaces, they found an overall mean of $\alpha = 1.26$. As the method proposed by Priestley and Taylor (1972) is physically sound and very simple, many applications have been made over a variety of surfaces.

Table 1-2 summarized the results of these applications. Although $\alpha = 1.26$ has been reported universally for open water surfaces (Stewart and Rouse, 1976, 1977; de Bruin and Keijman, 1979), $\alpha = 1.26$ has not necessarily been obtained over bare soil or vegetated surfaces with unlimited soil water supply. For example, Barton (1979) for bare soil surface, McNaughton and Black (1973), Stewart and Thom (1973), Black (1979), and Spittlehouse and Black (1981) for forested surfaces, and Rouse and Stewart (1972), Yap and Oke (1974), and McNaughton et al. (1979) for cropped surfaces reported values of α less than 1.26. On the other hand, values of α larger than 1.26 have been obtained by Jury and Tanner (1975) and Kanemasu et al. (1976). They considered that such large values of α resulted from advective conditions. Davies and Allen (1973), Williams et al.

Table 1-2 Parameter(α)in Priestley and Taylor potential evaporation.

surface type	place	α	reference
bare soil			
plowed bare soil	Gurley, N.S.W., Australia	1.08	Dyer and Hicks (1970) cited in Priestley and Taylor (1972)
clay loam soil	Phoenix, Ariz., U.S.A.	1.37	Jackson et al. (1976)
burnt area	Deniliquin, N.S.W., Australia	1.04	Barton (1979)
open water			
Indian Ocean		1.20	Paulson (1967) cited in Priestley and Taylor (1972)
Indian Ocean		1.25 – 1.31	Deacon and Stevenson (1968) cited in Priestley and Taylor (1972)
Atlantic Ocean		1.30	Hoeber (1970) cited in Priestley and Taylor (1972)
Lake Eucumbene	N.S.W., Australia	1.25	Webb (1960) cited in Priestley and Taylor (1972)
Perch Lake	Ont., Canada	1.26	Ferguson and den Hartog (1975) cited in Stewart and Rouse (1977)
Perch Lake	Ont., Canada	1.26	Barry and Robertson (1975) cited in Stewart and Rouse (1977)
shallow lake	Ont., Canada	1.26	Stewart and Rouse (1976, 1977)
Lake Flevo	Netherlands	1.25	de Bruin and Keijman (1979)
forest			
Douglas fir	Vancouver, B.C., Canada	1.03	McNaughton and Black (1973)
Douglas fir	Vancouver, B.C., Canada	0.8	Black (1979)
Douglas fir	Vancouver, B.C., Canada	1.1	Spittlehouse and Black (1981)
pine trees	Thetford, Norfolk, England	0.6 – 0.7	Stewart and Thom (1973) cited in Black (1979)
crop			
grass	Hay, N.S.W., Australia	1.33	Clarke et al. (1971) cited in Priestley and Taylor (1972)
grass	Ladner, B.C., Canada	1.02	Yap and Oke (1974)
grass	Tronto, Ont., Canada	1.29	Mukammal and Neumann (1977)
grass	Asa-NLH, Norway	1.25	Hansen and Hegg (1981)
ryegrass	Simcoe, Ont., Canada	1.27	Davies and Allen (1973)
wheatgrass	Kamloops, B.C., Canada	1.26	Williams et al. (1978)
pasture	Aspendale, Vic., Australia	1.34	Priestley and Taylor (1972)
sedge meadow	Pen Is., Ont., Canada	1.26	Stewart and Rouse (1977)
sedge meadow	Thor Lake, Sask., Canada	1.26	Stewart and Rouse (1977)
alfalfa	Hancock, Wis., U.S.A.	1.42	Jury and Tanner (1975)
snap bean	Madison, Wis., U.S.A.	1.30	Black et al. (1968) cited in Priestley and Taylor (1972)
soybean	Manhattan, Kan., U.S.A.	1.45	Kanemasu et al. (1976)
potato	Hancock, Wis., U.S.A.	1.28 – 1.43	Jury and Tanner (1975)
winter wheat	Great Plains, U.S.A.	1.35	Heilman et al. (1977)
oat	Palmerston North, New Zealand	1.22	McNaughton et al. (1979)
sorghum	Manhattan, Kan., U.S.A.	1.28	Kanemasu et al. (1976)
radish	Chiba, Japan	1.26	Nakayama and Nakamura (1982)
tundra vegetation (lichen, mosses, and stunted shrubs)	Hudson Bay lowland, Ont. Canada	0.96	Rouse and Stewart (1972)
pasture, paspalum, and lucerne	Palmerston North, New Zealand	1.16	McNaughton et al. (1979)
variety of crops		1.35	Tanner and Jury (1976)

(1978), and Barton (1979) related α to soil water conditions, but their equations for this relationship are not sufficient for general use, and further studies are needed. Potential evaporation has been used in the evaporation models (e.g., Heilman et al., 1977; Brutsaert and Stricker, 1979).

The fact that, even over completely wetted surfaces like open water, the value of α is not equal to unity, but to 1.26, shows that advection-free conditions, in the sense of Slatyer and McIlroy (1961), do not occur. Brutsaert (1982) reasoned that this is due to the fact that the atmospheric boundary layer is never a truly homogeneous boundary layer, but that it is continually responds to large scale weather patterns, involving condensation and unsteady three-dimensional motion. McNaughton (1976) developed an analytical model to examine effects of changes in surface resistance or available energy on the local evaporation. He suggested that $\alpha > 1$ represents mesoscale advective enhancement of evapotranspiration and that $\alpha < 1$ represents advective suppression or strong surface control of evapotranspiration.

The Priestley and Taylor model for potential evaporation has been applied indiscriminately. As a result, the values of α have been found to vary widely even under wet soil conditions. Therefore, it is necessary to clarify the conditions to which "potential evaporation" by Priestley and Taylor can be applied.

Confusion of the interpretation of potential evapotranspiration

Through the advance in the theory and the accumulation of field data on evapotranspiration, there appears to be confusion both as to the interpretation of several concepts on evapotranspiration and of the application of evaporation formulae. In the earlier definition of "potential evapotranspiration" by Thornthwaite or "potential transpiration" by Penman, full-cover cropped surfaces with no shortage of water were treated as a necessary condition for the potential rate. Although Penman (1956) restricted the vegetation condition for potential evapotranspiration to a fresh green crop of fairly uniform height, the vegetation condition has not been well specified. While the ambiguous features inherent in the definition of potential evapotranspiration have been discussed (Chang, 1965; Rosenberg, 1974), many applications have been done without considering the ambiguous features in detail. As a result, there is no unanimity about the surface conditions or the calculation procedures for potential evapotranspiration. One researcher refers potential evapotranspiration to the actual vegetation of interest, whereas others consider potential evapotranspiration from a hypothetical reference crop (which is mostly a short green grass). These different treatments result from an ambiguity concerning crop type. The obscurity about the soil water condition has caused two types of definitions of potential evapotranspiration; one is the externally wetted surface condition corresponding to $r_c = 0$ in Eq. (1-6), and the other is the well-watered soil condition corresponding to $r_c = r_{cm}$ (the minimum canopy resistance under an ample soil water condition). For a cropped surface, r_a is large and the difference between the above definitions is considered so small that the above two types of definitions have been used alternatively. In contrast to this, for a forest, r_a is small and the difference between the above definitions is considered to be large (Federer, 1975).

Federer (1975) proposed that the term "potential evapotranspiration" should be used in the meaning of externally wetted condition and that the term "unstressed evapotranspiration" should be used in the meaning of the non-limiting soil water condition. De Bruin (1981) proposed the use of the term "crop water requirement" and "reference crop evapotranspiration" to eliminate confusion. He stated as follows: "crop water requirement" is defined as the amount of water needed to meet the water loss through evapotranspiration of a disease-free crop, growing in a large field under non-restricting soil conditions including soil water and fertility and achieving full production potential under a given environment; on the other hand, "reference crop evapotranspiration" is defined as the rate of evapotranspiration from an extensive surface 8 to 15-cm tall, green grass cover of uniform height, actively growing, completely shading the ground without shortage of soil water. Brutsaert (1982) also pointed out the ambiguities in the term "potential evapotranspiration" and considered the term "potential evaporation" as probably preferable. He defined "potential evaporation" as the evaporation from any large uniform surface which is sufficiently moist or wet so that the air in contact with it is fully saturated. He stated that such conditions prevail usually only after the occurrence of precipitation or dew.

The above proposals have not been fully verified by field observations. From the above discussion, the main points in the ambiguities are the treatment of the surface wetness and the surface roughness. Therefore, it is necessary to clarify the conditions to which concepts on evapotranspiration can be applied.

1-2 Objectives of the study

Details in the evapotranspiration process have been determined progressively by recent studies and several evaporation equations have been proposed. There appears, however, confusion in the definition and the usage of concepts on evapotranspiration. Since little attention has been paid to the evaporation process in wet regions, especially in Japan, there remain some unsolved problems in water balance studies (e.g., Kayane and Takeuchi, 1971).

Three objectives are set in this study.

The first objective is to investigate the conditions to which concepts on evapotranspiration can be applied. Potential evapotranspiration, equilibrium evaporation, and potential evaporation are taken as representative concepts.

The second objective is to make clear the effects of wetness of the evaporating surface on evapotranspiration.

The third objective is to investigate the validities of the several evaporation equations most commonly used for hydrological purposes in Japan.

1-3 Procedures used

To investigate applicable conditions of several concepts on evapotranspiration, a direct measurement of evapotranspiration is necessary. For this purpose, a micrometeorological observation system was designed to obtain basic data. Evapotranspiration data obtained by a large weighing lysimeter were also used. These observation and data acquisition methods are discussed in Chapter 2.

In Chapter 3, the applicable conditions of concepts on evapotranspiration are discussed on the basis of hourly evapotranspiration data. A similar analysis, based on daily evapotranspiration data is made in Chapter 4. Also, a simple evaporation model, based on the equilibrium evaporation, is presented and tested in this Chapter.

Seasonal patterns of evapotranspiration have been obtained directly by a weighing lysimeter. The investigation of applicabilities of the equilibrium evaporation model and some estimation formulae for evapotranspiration is made in Chapter 5.

CHAPTER 2

METHODS

2-1 Experimental site

The study was conducted at the heat balance and water balance experimental field of Environmental Research Center (ERC), the University of Tsukuba, Ibaraki Pref., Japan (36°05′ N, 140°06′ E). The University of Tsukuba is located in the core of the Tsukuba Science City about 60 km northeast of Tokyo (Fig. 2-1).

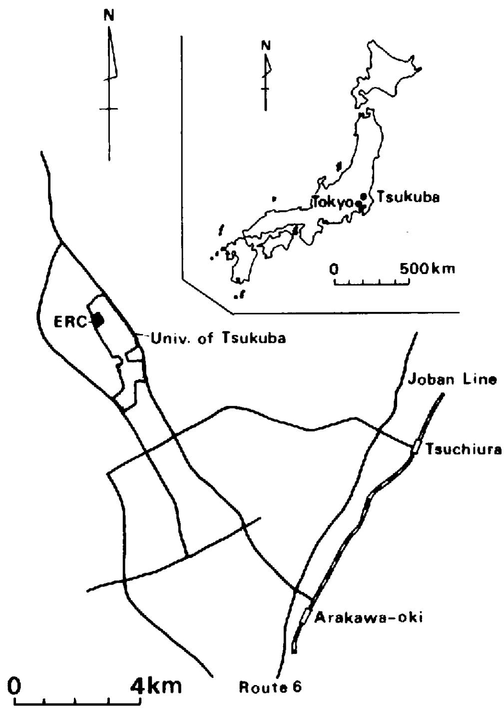

Fig. 2-1 Location map of the experimental site.

The experimental field is a circular plot with a radius of 80 m and has a 30-m high meteorological observation tower at its center. The vegetation of the field consists of mixed pasture (*Poa pratesis* L., *Eragrostis curvula* Nees, etc.). The pasture approaches maturity in summer and mowing is done in early winter.

The surroundings of the field are not homogeneous, being interrupted by some buildings and pine trees (Fig. 2-2). To the north of the field, there exists a large and long building which contains a large scale experimental flume, 188-m long and 9.5-m high. Pine trees with a height of about 10 m exist to the northwest and the southeast of the field.

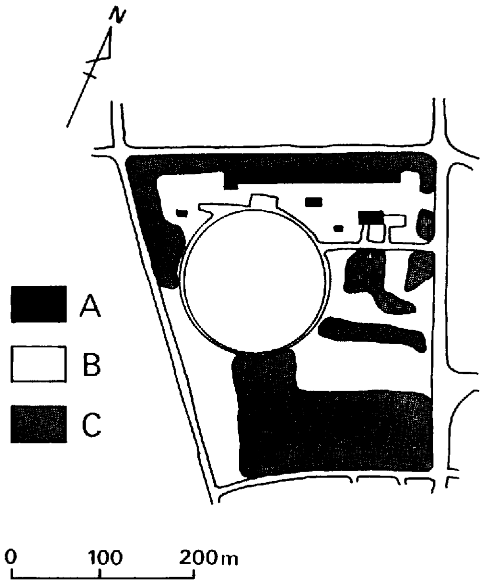

Fig. 2-2 Experimental site and its surroundings.
A : Building
B : Experimental field
C : Pine trees

2-2 Instrumentation

All the measurements were made near the center of the experimental field to ensure enough fetch for the development of a surface boundary layer. Figure 2-3 shows the locations of the instruments in the field. In summer, the prevailing wind direction is between southeast and southwest. Therefore, data are collected with fetch between 60 and 90 m. In winter, however, the dominant wind direction changes to northeast and northwest, so that fetch becomes less than 60 m.

In ERC, long-term basic data for heat balance and water balance are routinely measured, but these data are insufficient for a detailed study of evapotranspiration. Therefore, some other instruments were installed for the research. Figure 2-4 shows the instrumentation in the study.

Net radiation was measured with a Funk type shielded net radiometer (Eko Instruments Trading Co., Ltd., Type CN-11) mounted 1.5 m above the ground surface.

Soil heat flux was measured with a soil heat flux plate (Eko Instruments Trading Co., Ltd., Type CN-9) buried at a depth of 2 cm below the surface.

Solar radiation was measured with the upper part of a solari-albedo meter (Eko Instruments Trading Co., Ltd., Type MR-21) mounted 1.5 m above the surface.

Reflected solar radiation was measured with the lower part of the solari-albedo meter as described above.

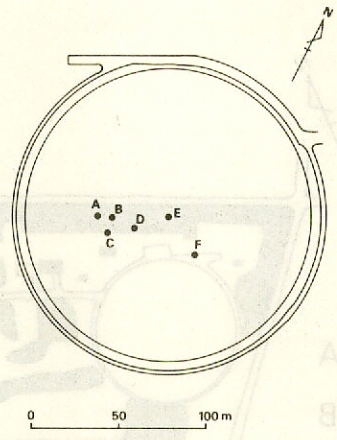

Fig. 2-3 Location map of the instruments in the experimental field.

A : Anemometers and ventilated psychrometers
B : Tensiometers
C : Albedometer, net radiometer, soil heat flux plate, and thermometers for soil temperature
D : Weighing lysimeter
E : Ventilated thermometer
F : Net radiometer

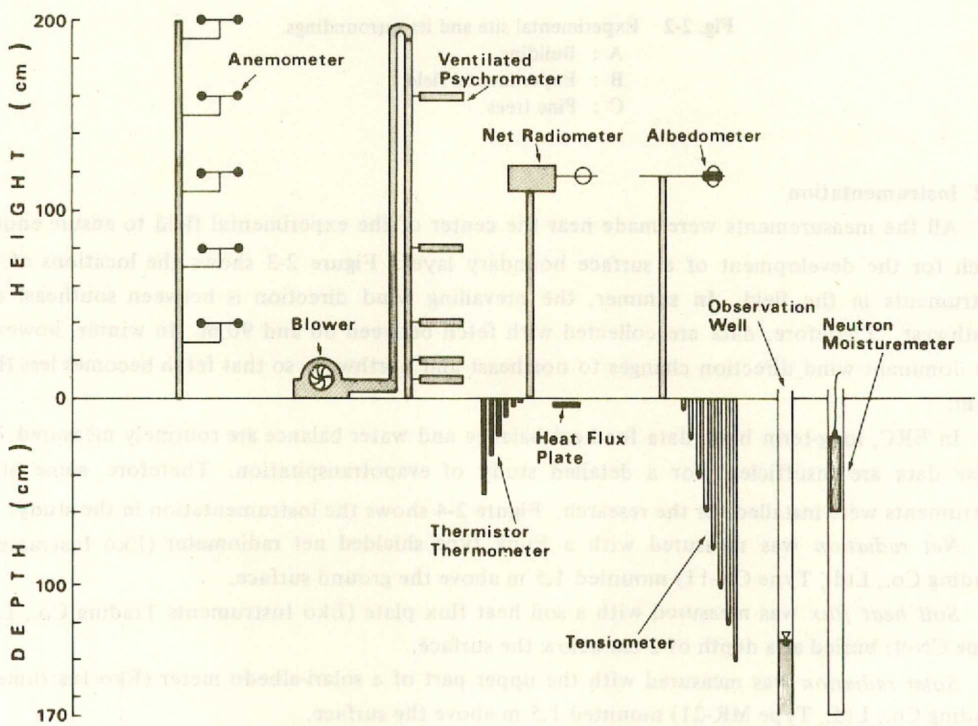

Fig. 2-4 Schematic diagram of the observation system.

Vertical profiles of air temperature and vapor pressure were determined from measurements at heights of 0.1, 0.2, 0.4, 0.8, and 1.6 m above the surface by using ventilated psychrometers. Several psychrometers are necessary for the determination of the vertical profiles of temperature and humidity. Since budgetary and size limitations prohibited the use of commercial instruments, ventilated psychrometers, which met the essential demands for profile determination, were designed and constructed.

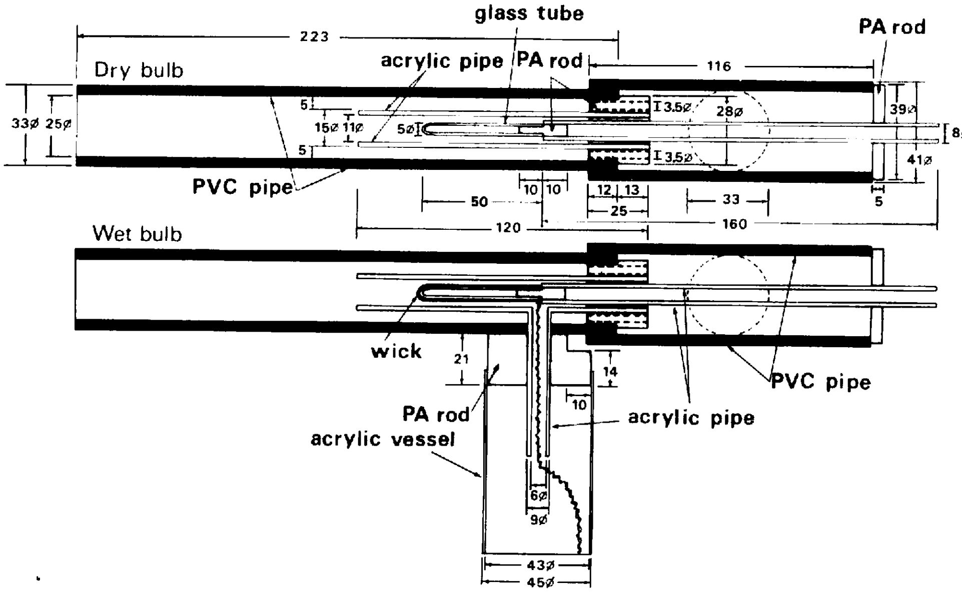

Fig. 2-5 Schematic diagram of the ventilated psychrometer (side view). (unit: mm)

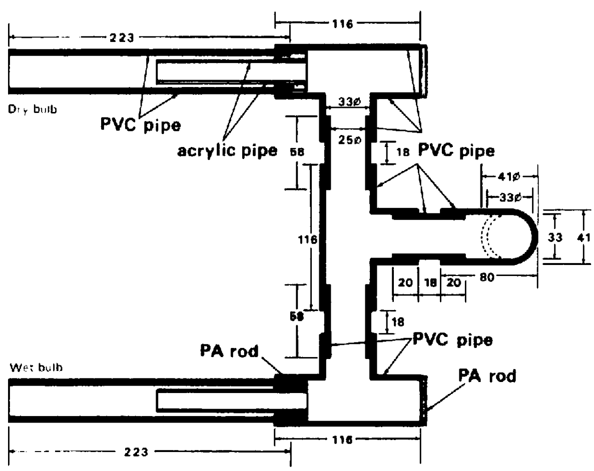

Fig. 2-6 Schematic diagram of the ventilated psychrometer (plan view). (unit: mm)

An outline of the psychrometer is shown in Figs. 2-5 and 2-6. To reduce radiation effects, outer and inner shields were used and the outer shield was painted white. As shown in Figs. 2-5 and 2-6, the outer shield was made of a 22.3-cm long polyvinyl chloride (PVC) pipe (3.3 cm O.D. and 2.5 cm I.D.) supported by a PVC tee. The inner shield was made of an acrylic pipe (1.5 cm O.D. and 1.1 cm I.D.), which was supported by a short polyacetal (PA) rod attached to the outer shield. The shields for dry- and wet-bulbs were identical except that a wet-bulb shield had a water supply pipe and a water tank holding apparatus. Water was conveyed to a wet-bulb sensor from the water tank through a wick. PA rods connecting inner and outer shields had 8 small holes (0.35 cm in diameter) which allowed air to pass through.

Temperature sensors were made by inserting copper-constantan thermocouple into a grass tube (5 cm in length, 0.5 cm O.D., and 0.3 cm I.D.) filled with liquid paraffin. A sensor assembly was made by attaching a sensor to an acrylic pipe (160 cm in length) with a PA connector. A set of psychrometer was constructed by connecting dry-bulb and wet-bulb parts with PVC tees (Fig. 2-6).

Five units of psychrometers were mounted on a single PVC pipe. Aspiration was made simultaneously for five psychrometers by a blower connected with a flexible tube. The aspiration

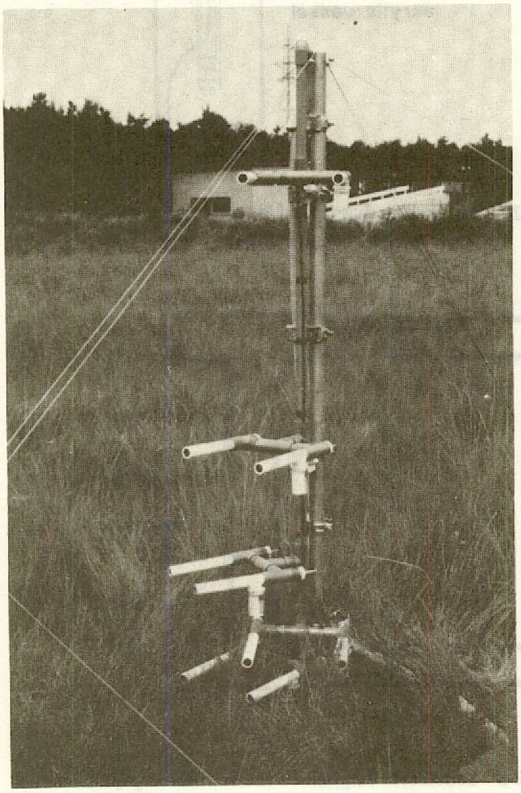

Fig. 2-7 Ventilated psychrometer system in the field.

ability of the blower provided a 500 cm/s aspiration rate. Figure 2-7 shows the ventilated psychrometer system.

Wind speed profile was obtained with the use of three-cup anemometers (Makino Applied Instruments, Inc., Type AC750P). Measuring heights were 0.4, 0.8, 1.2, 1.6, and 2.0 m above the ground.

Soil temperature profile was obtained with thermistor thermometers (Takara Thermistor Instruments Co., Ltd., Type THR A03) buried at 2, 5, 10, 20, 30, and 50 cm depths.

Soil water potentials were measured with tensiometers at 5, 20, 40, 60, 80, 100, and 140 cm depths. Readings of soil water pressure heads were made with U-tube mercury manometers.

Groundwater level was measured manually in an observation well dug to 2 m below the surface.

Soil moisture content profile was obtained by a neutron method. A thin-wall aluminum access tube, 4.9 cm O.D. and 200-cm long, was installed vertically. The equipment used to measure soil moisture content included a moisture probe with a 50 μCi ^{252}Cf source (Type SX-8n), a paraffin filled standard box, a line-operated portable scaler (Type CX-1A), and an AC adapter (Type AX-1). The equipment was provided by Nippon Earthwork & Electro Technic Lab. Measurements of soil moisture content were made once a day from 10 cm below the soil surface to a groundwater table position at 10 cm increments.

Actual evapotranspiration was routinely measured by a large weighing lysimeter (Shimazu Seisakusho Ltd.). The soil containers consist of two cylindrical vessels. The outer container is made of stainless steel with a radius of 1 m and a length of 2 m. A 5-mm thick circular stainless steel drainage board with about 400 small holes (radius of 20 mm) is set at 18 cm from the bottom of the container. The inner container is made of iron, 0.9 m in radius, 1.8-m long and 9-mm thick. Undisturbed Kanto loam soil was packed in the inner container by putting it into the ground. The packed soil is the same as field soil. An iron filter padded with grass wool was attached to the bottom of the inner container. Then the inner container was set inside the outer one. Disturbed Kanto loam soil was packed in the gap between the inner and outer containers. These containers were rested on the platform of the balance (2.2 x 2.3 m) in the basement (Fig. 2-8).

Drainage from the soil column is caused by the force of gravity and drained water is pooled in a reserve tank. If drained water reaches a given depth, the level switch in the reserve tank operates to open the pinch bulb and water is drained quickly into a water tank (Fig. 2-9). The water level in the lysimeter can be controlled in the range of 1.5 to 1.7 m from the soil surface. It has been fixed to 1.7 m for 4 years.

The mechanical weighing system is shown in Fig. 2-10. First, the weight of the container is balanced by counterweights (e). The change in weight of the container is measured by the automatical electromagnetic balance system using a feedback mechanism, which consists of photoelectrical elements and feedback coils. The change of weight is transmitted to the balance beam (d) and causes a downward or upward displacement. The displacement is detected by photoelectrical elements (f) which receive a light beam from the lump (h), through a slit (g), so that the electrical current equivalent to the degree of displacement is supplied to the feedback coils (j) after amplification (k). As a result, an electromagnetic force occurring in the magnet (i) operates to put the balance beam back to equilibrium position. The magnitude of the electric current supplied to maintain the equilibrium position is proportional to the change of weight.

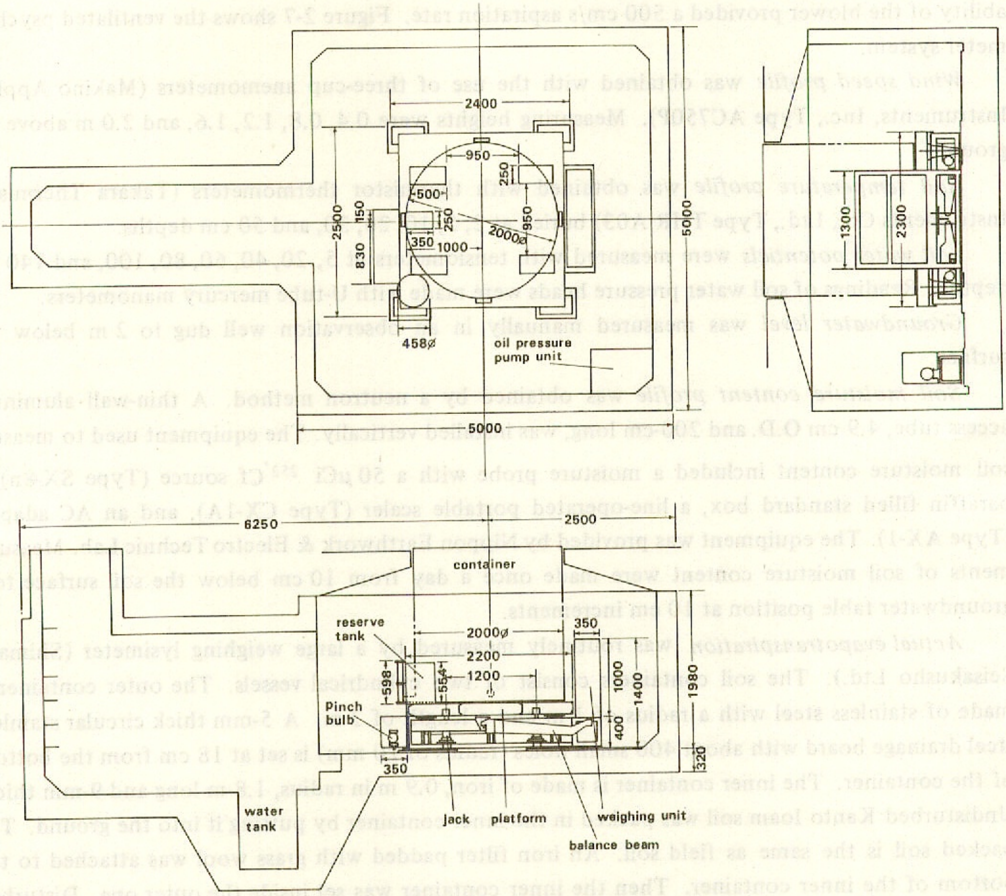

Fig. 2-8 Diagram of a weighing lysimeter. (unit: mm)

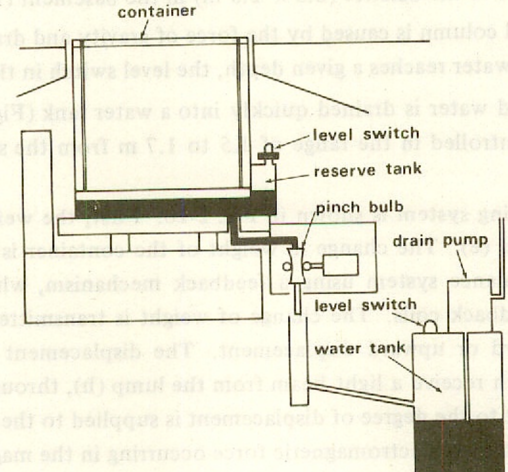

Fig. 2-9 Drainage system of a weighing lysimeter.

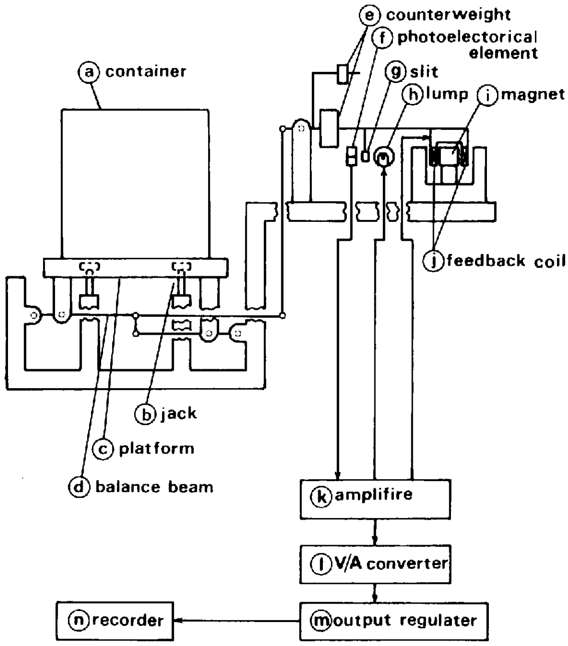

Fig. 2-10 Measuring system of a weighing lysimeter.

The maximum weighing capacity of this system is 1500 kg and the weighing range is 0 to 250 kg. Total weight of the lysimeter is about 9000 kg and the weighing sensitivity is 500 g, which is equivalent to 0.16 mm of water.

2-3 Data collection

Intensive observations were conducted from July 20 to August 31, 1980. From September 1, 1980 until April 17, 1981, net radiation, solar radiation, reflected short-wave radiation, soil heat flux, air temperature, and actual evapotranspiration were measured. In the latter case, air temperature was obtained by the ventilated resistance thermometer at a height of 1.6 m.

Figure 2-11 shows a block diagram for the data collection system. Instantaneous and 1-hr total values of net radiation, solar radiation, and reflected short-wave radiation were recorded by a potentiometric dotted-line recorder. One hour total values were obtained electrically by means of analog integrators (Eko Instruments Trading Co., Ltd., Type MP-20). Instantaneous values of soil heat flux were recorded by the same dotted-line recorder as described above. A 1-hr total value was obtained by measuring the area on a recording chart with an automatic area meter (Hayashi Denko Co., Ltd., Type AAM-7).

Dry-bulb and wet-bulb temperatures were logged and printed out every 10 min. by using a data logger (Yokogawa Electric Works, Ltd., Type 3874-41). One hour average values of dry-bulb and wet-bulb temperatures were calculated from each 10 min. data from which 1-hr averages of vapor pressure were obtained through use of a psychrometric equation.

Soil temperatures were recorded by a dotted-line recorder after linearization of the thermistor outputs.

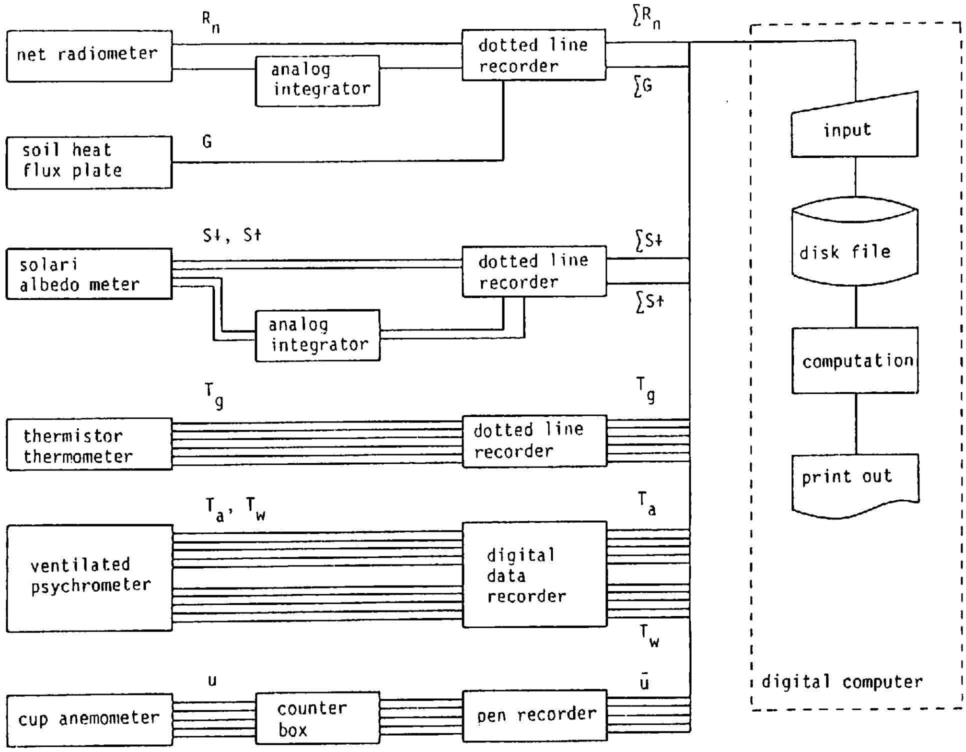

Fig. 2-11 Block diagram for the data acquisition and computation.

One hour average wind speeds were obtained by counting the number of pulses from cup anemometers recorded on a chart. A rotation of cup was produced at every 1.2 m wind run and an electrical pulse was generated at each rotation. The original pulses were decreased in number to 1/50, using a counter box and deducted signals were recorded on a chart.

Soil water potentials were measured several times a day by reading the values of mercury manometers.

Soil moisture content profile and groundwater level were measured once a day.

CHAPTER 3

ANALYSIS OF HOURLY EVAPOTRANSPIRATION

A direct measurement of evapotranspiration is very difficult and many estimation methods and models have been developed. Several concepts on evapotranspiration also have been proposed. The field tests of these estimation methods have been extensively done in arid and semi-arid regions, but are very scarce in humid regions, especially in Japan.

In this Chapter, potential evapotranspiration, equilibrium evaporation, and potential evaporation are chosen as representative concepts of evapotranspiration. The conditions to which they can be applied are discussed.

3-1 Determination of actual evapotranspiration

Field observations were carried out from July 20 to August 31, 1980. At the time of observation, the field crop had attained maturity and its height was about 40 cm.

For a short time period such as 1-hr, the use of a weighing lysimeter for the determination of actual evapotranspiration is very difficult, because of the measurement errors caused mainly by wind. Therefore, an energy budget with Bowen's ratio (EBBR) method was used for the determination of latent heat flux. The energy balance equation at the surface can be written as

$$R_n - G = \lambda E + H \tag{3-1}$$

where R_n is the net radiation, G the soil heat flux, λ the latent heat for vaporization, E the evapotranspiration rate, λE the latent heat flux, and H the sensible heat flux. The quantity of the left hand side in Eq. (3-1), i.e., $(R_n - G)$, is called as the available energy.

The transfer equations for the sensible heat and latent heat fluxes are expressed as

$$H = -\rho c_p K_H \frac{\partial T}{\partial z} \tag{3-2}$$

$$\lambda E = -\frac{\rho c_p}{\gamma} K_E \frac{\partial e}{\partial z} \tag{3-3}$$

where ρ is the density of the air, c_p the specific heat of the air at constant pressure, γ the psychrometric constant, K_H the turbulent diffusivity for heat, K_E the turbulent diffusivity for water vapor, T the air temperature, e the vapor pressure, and z the height.

The ratio of these two flux terms is called Bowen's ratio (β) and defined as

$$\beta = \frac{H}{\lambda E} = \frac{\gamma K_H \partial T/\partial z}{K_E \partial e/\partial z} \tag{3-4}$$

In this equation, the differential forms are expressed by the difference forms and K_H is set equal to K_E, then Eq. (3-4) is reduced to

$$\beta = \gamma \frac{T_1 - T_2}{e_1 - e_2} \tag{3-5}$$

where subscripts, 1 and 2, refer to lower and higher levels of temperature and vapor pressure measurements, respectively. By combining Eqs. (3-1) and (3-5),

$$\lambda E = \frac{1}{1+\beta}(R_n - G) \tag{3-6}$$

is obtained.

As mentioned, temperature and vapor pressure were measured at three levels above the crop surface, namely 0.8, 1.2, and 1.6 m. The combinations of 1.2 with 1.6 m and 0.8 with 1.6 m were used to calculate Bowen's ratio and these values were referred to as BW1 and BW2, respectively.

However, it sometimes became difficult to obtain the latent heat flux by the EBBR method. If there is only a small gradient of either temperature or vapor pressure at either sunrise or sunset, an accurate measurement can not be obtained. Also, on occasion there may be a drying of the wet-bulb. Therefore, the following energy budget with wind and scalar profile (EBWSP) method was used, when the EBBR method could not be applied. Since the temperature profile could be obtained more accurately than the vapor pressure profile, temperature was used as a scalar in the EBWSP method.

In the surface boundary layer, the wind profile under neutral conditions is expressed as

$$u(z) = \frac{u_*}{k} \ln\left(\frac{z-d}{z_0}\right) \tag{3-7}$$

where u is the wind speed, u_* the friction velocity, k the Karman's constant, d the zero-plane displacement height, and z_0 the roughness length. The aerodynamic resistance for the transfer of momentum from height z_2 to z_1 is expressed as

$$r_M(z_1, z_2) = \frac{\rho(u_2 - u_1)}{\tau} \tag{3-8}$$

where r_M is the aerodynamic resistance for the transfer of momentum, τ the shearing stress and subscripts 1 and 2 refer to height. The shearing stress is defined as

$$\tau = \rho u_*^2 \tag{3-9}$$

From Eqs. (3-7), (3-8), and (3-9), $r_M(z_1, z_2)$ is also expressed by

$$r_M(z_1, z_2) = \frac{1}{k^2} \frac{[\ln\{(z_2 - d)/(z_1 - d)\}]^2}{u_2 - u_1} \tag{3-10}$$

The representation of sensible heat flux by resistance form can be written as

$$H = \rho c_p \frac{T_1 - T_2}{r_H(z_1, z_2)} \tag{3-11}$$

where r_H is the aerodynamic resistance for the transfer of heat. By the assumption $r_M = r_H$, Eqs. (3-10) and (3-11) produce the following equation for the sensible heat flux:

$$H = \rho c_p k^2 \frac{(T_1 - T_2)(u_2 - u_1)}{[\ \ln\{(z_2 - d)/(z_1 - d)\}\ ]^2} \tag{3-12}$$

In the EBWSP method, first the sensible heat flux is calculated by Eq. (3-12), and then the latent heat flux is obtained as a residual of the heat balance equation (Eq. 3-1).

The value of zero-plane displacement height in Eq. (3-12) was calculated by a procedure proposed by Robinson (1962). As a result, averages of 21.4 cm and 6.6 cm were obtained for the zero-plane displacement height and the roughness length, respectively. It has been reported that the zero-plane displacement height and the roughness length depend not only on the crop height, but also on the wind speed and friction velocity (Tani, 1960; Udagawa, 1966; Maki, 1969, 1975, 1976; Kotoda, 1979; Hayashi, 1979; Kotoda and Hayashi, 1980). However, the average value of zero-plane displacement height was used to calculate the sensible heat flux in Eq. (3-12).

Since the observations were carried out near the ground, i.e., below a height of 2 m, the influence of atmospheric stability on the transfer of fluxes was ignored.

In the analysis, the calculated 6-hr totals of latent heat fluxes by the EBBR and the EBWSP methods were compared with those measured by a weighing lysimeter. Erroneous cases were omitted from the analysis.

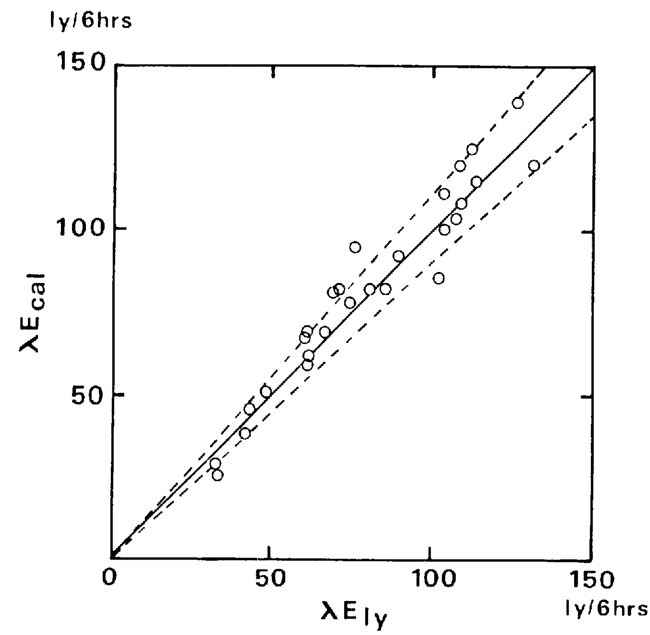

Fig. 3-1 Relationship between measured latent heat flux (λE_{ly}) by a weighing lysimeter and calculated latent heat flux (λE_{cal}). Broken lines in the figure represent ±10% error.

Figure 3-1 shows the relationship between the calculated and the measured latent heat fluxes. Generally, the discrepancy between them is as large as 10% and systematic errors are not detectable.

3-2 Relationship between actual evapotranspiration and potential evapotranspiration

As mentioned in Chapter 1, there are two definitions of potential evapotranspiration. One is based on a vapor-saturated condition of an evaporating surface and the other is based on a well-watered soil condition. Analyses were done to make clear the following: whether or not these two definitions of potential evapotranspiration can be used interchangeably; and if not, to what conditions they can be applied.

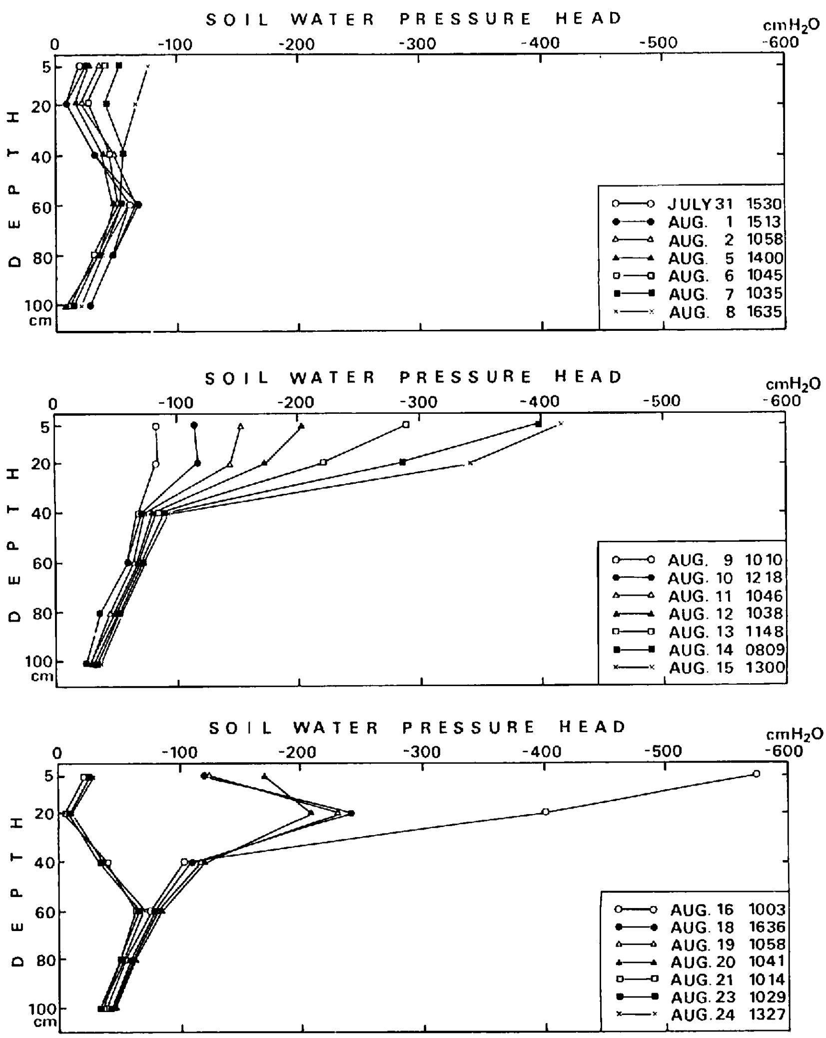

Fig. 3-2 Changes in soil water pressure head at different depths.

Figure 3-2 shows the variations of soil water pressure head profile. From July 30 to August 4, it was cloudy and occasionally rained. The evapotranspiration rate was low. Accordingly, there appeared little change in the soil water pressure head and the matric suction was small. The weather became clear from August 5 and hence the soil water pressure head began to decrease. The large changes in soil water pressure head occurred in the upper part of the soil profile, particularly above the 40-cm depth line from the surface. After reaching its driest condition on August 16, the soil was rewetted by rain on August 17, 18, and 20. On August 21, the soil water pressure head profile showed the same features as shown on July 31.

The smallest value of soil water pressure head at 5-cm depth was detected on August 16, at -574 cm H_2O, equivalent to pF 2.8, which was near the critical pF value at which the soil water controlling stage of evapotranspiration begins (Kaneki and Tomita, 1975). However, pF values at the 5-cm depth were less than 2.8 on other days. Consequently, it can be considered that evapotranspiration proceeded in the absence of soil water shortage and that actual evapotranspiration was equivalent to the potential evapotranspiration in accordance with the definition, based on a well-watered soil condition.

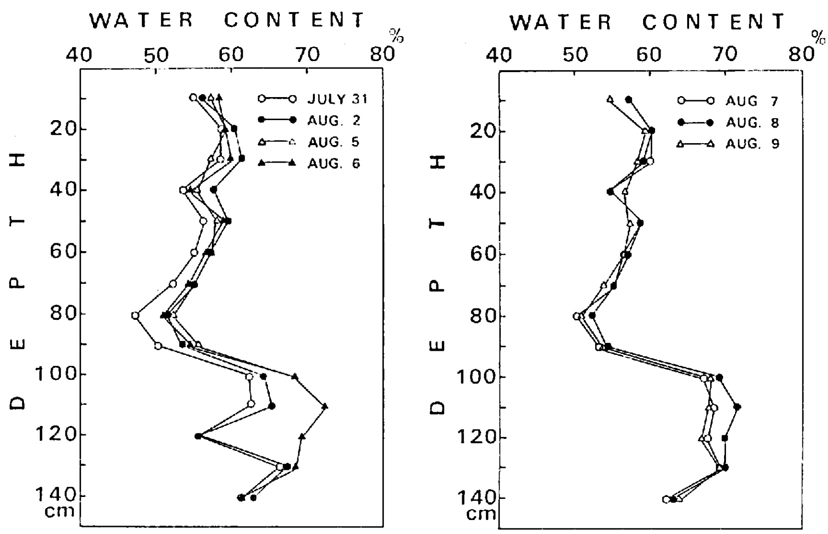

Fig. 3-3 Changes in soil water content profiles.

Figure 3-3 shows profiles of soil moisture content. Data after August 10 could not be obtained due to a breakdown in the soil moisture meter. It is shown in Fig. 3-3 that volumetric soil moisture content at the 10-cm depth was about 55% with slight fluctuation.

There are several formulae which are used to estimate evapotranspiration from an evaporating surface in a vapor-saturated condition. The most commonly used method is based on the Penman-Monteith equation with $r_c = 0$. In this study, the following formula is used as representative:

$$\lambda PE = \frac{\Delta}{\Delta + \gamma}(R_n - G) + \frac{\rho c_p}{\Delta + \gamma}\ \frac{e_a{}^* - e_a}{r_a} \qquad (3\text{-}13)$$

where PE is the potential evapotranspiration as defined by a vapor-saturated surface condition. The first term on the right-hand side of Eq. (3-13) is called the radiation term, which is equivalent to the equilibrium evaporation as proposed by Slatyer and McIlroy (1961), and the second term is the ventilation term. Van Bavel (1966) developed the following formula by excluding an empirical feature in Penman's ventilation term:

$$PE = \frac{\Delta}{\Delta + \gamma}\left(\frac{R_n - G}{\lambda}\right) + \frac{\gamma}{\Delta + \gamma}\ \frac{\rho \epsilon k^2}{p}\ \frac{u\,(e_a{}^* - e_a)}{[\ \ln\{\ (z - d)\ /\ z_0\}\ ]^2} \qquad (3\text{-}14)$$

where ϵ is the ratio of mole weights of water vapor to air (=0.622) and p the atmospheric pressure. This equation is slightly different from Eq. (3-13) but becomes equal after transformation.

As mentioned previously, the potential evapotranspiration as defined by an ample soil water condition is considered to be the same as the actual evapotranspiration in this study. Therefore, hereafter it is designated by E.

Examples of the diurnal variations of PE and E are shown in Fig. 3-4, in which (a) to (f) designate fine days and (g) and (h) cloudy days. It is noticeable on fine days that close agreement between PE and E is obtained in the morning, except on August 6. In addition, this agreement lasts until around noon in some cases. However, after this accordance, PE takes upon larger value than E and the discrepancies between them become large in the afternoon. The duration in which PE equals E differs from day to day. The condition of $PE \cong E$ last until 1100h, 1200h, and 1000h on July 22, July 23, and August 12, respectively. These were long duration cases. On the other hand, $PE \cong E$ last only until 0600h and 0700h on July 24 and August 9, respectively.

Table 3-1 shows the amount of dewfall during nighttime and the time when dew disappeared. The amount of dewfall was calculated by summing up the negative E values obtained by the EBWSP method. In this procedure, nighttime negative E values were considered as dewfall. The time at which dew disappeared was obtained as follows: first, in the cases where nighttime dewfall was detected, positive E values in the early morning were considered to occur due to all λE being generated by the evaporation of dew, i.e., transpiration from vegetation and evaporation from soil surface were assumed to be negligible. Secondly, hourly E values were accumulated until ΣE became larger than the amount of dewfall. The time when the accumulated E became larger than the amount of dewfall was considered to be the time when dew disappeared. There are two kinds of processes in dew formation, i.e., condensation (the turbulent transfer of water vapor from the atmosphere to leaves) and distillation (the turbulent transfer of water vapor from soil to leaves) (Monteith, 1957). However, in this study, a distinction between these processes was not made.

The amount of dewfall ranges from 0 to 0.38 mm/night. The maximum dewfall, observed on August 12, is consistent with other observation results (e.g., Lloyd, 1961; Baier, 1966; Hungerford, 1967; Burrage, 1972; Monteith, 1973). It is shown in Table 3-2 that the duration of dew evaporation coincides with the duration during which $PE \cong E$ is established in days when the relation $PE \cong E$ holds for a short time (July 24 and August 9). It is also clear that the relation $PE \cong E$ remains valid after the completion of dew evaporation in days when the relation $PE \cong E$

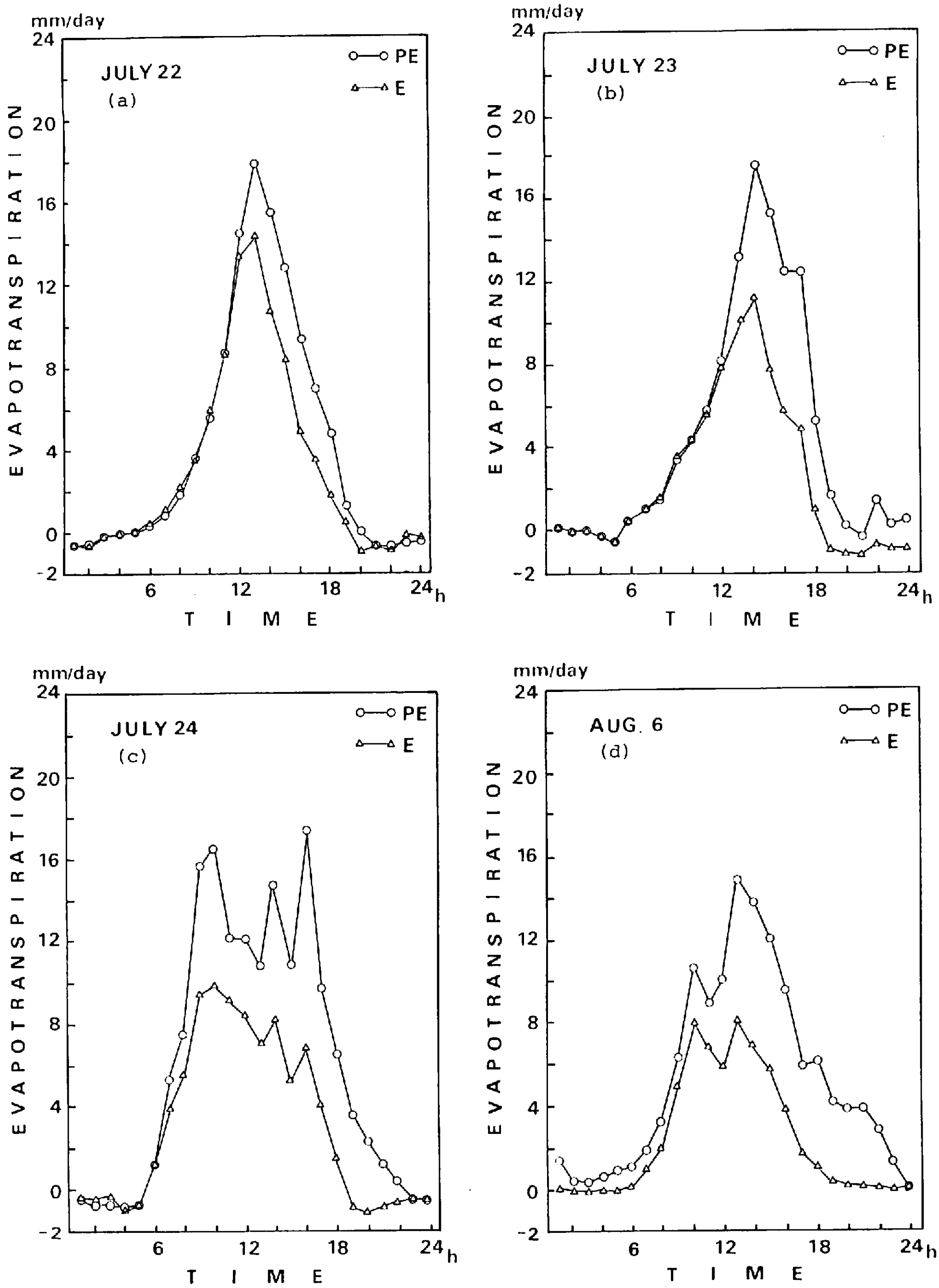

Fig. 3-4 Diurnal variations of actual evapotranspiration (E) and potential evapotranspiration (PE) defined by a vapor-saturated surface condition.

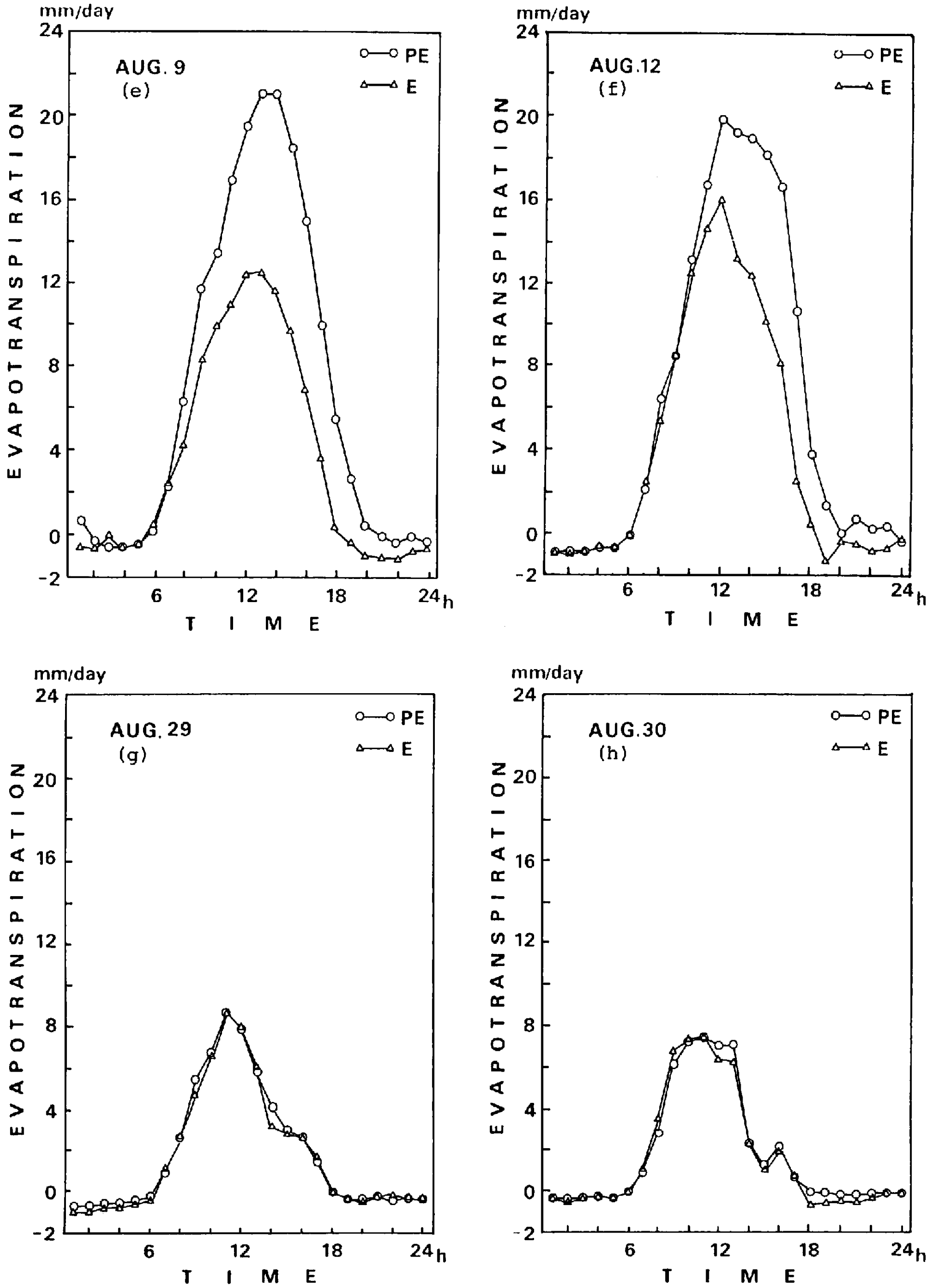

Fig. 3-4 (continued)

Table 3-1 Amount of dewfall and the time when dew disappeared.

date	amount of dewfall (mm/night)	dew diminishing time (hr)
July 22	0.12	0800
July 23	0.15	0800
July 24	0.18	0600
August 6	0.00	–
August 9	0.10	0700
August 12	0.38	0900

occurs for a long period (July 22, July 23, and August 12) and that the relation $PE > E$ proceeds throughout the daytime when there is no dewfall (August 6).

Figure 3-5 shows the diurnal variations of potential evapotranspiration (PE) in a vapor-saturated surface condition, available energy in terms of evaporation equivalence ($R_n - G^*$), radiation term of PE (E_{eq}), and ventilation term of PE (E_a'). It can be seen from Fig. 3-5 that the ventilation term in PE is relatively small compared to the radiation term in PE during evaporation of dew. This is also true in the case of the relation $PE \cong E$ lasting for a longer period. By contrast, it can be seen that the ratio of the ventilation term to the radiation term is larger during the period of $PE > E$ than during the period of $PE \leqq E$ and that PE becomes larger than $R_n - G^*$, with E_a'/E_{eq} nearly equal to 3.5 in some extreme cases (e.g., 1700h on July 24 and 1700h on August 6). The time when the condition of $PE > R_n - G^*$ was first established was 1300h on July 22 and 23, 0700h on July 24, 0800h on August 9, and 1200h on August 12.

Figure 3-6 shows the diurnal variations of heat balance components, vapor pressure deficit, and wind speed. Vapor pressure deficit and wind speed were measured at a height of 1.6 m. It is clear from Fig. 3-6 that vapor pressure deficit and wind speed are relatively low in the morning, but become larger in the afternoon. The larger values of vapor pressure deficit and wind speed in the afternoon cause E_a' to become larger than E_{eq}, resulting in the condition $PE > E$. When PE takes upon a value close to $R_n - G^*$, with an increase in E_a', the condition $PE \cong E$ no longer exists and PE becomes larger than E. A further increase in E_a' makes PE greater than $R_n - G^*$. This means that more energy than $R_n - G^*$ should be used as latent heat. However, sensible heat flux (H) takes a positive value even when PE is larger than E (see Fig. 3-6), so that the situation $PE > R_n - G^*$ shows a contradiction to the energy balance. The critical value of E_a'/E_{eq} for the transition from the condition $PE \leqq E$ to $PE > E$ proves to be approximately 0.33, i.e., if $E_a'/E_{eq} \leqq 0.33$, the condition $PE \cong E$ is established and if $E_a'/E_{eq} > 0.33$, $PE > E$ is established.

A summary of the above discussion is shown in Table 3-2. It is seen in Table 3-2 that the time lag between the terminations of dew evaporation and of the condition $PE \cong E$ is a few hours, with a maximum of 4 hours. During the evaporation of dew, the evaporating surface behaves as a completely wet surface with $r_c = 0$. Therefore, the evapotranspiration can be considered to be the same as the potential evapotranspiration when using the definition of a vapor-saturated surface condition. On the other hand, after the end of dew evaporation, water transfer through the stomata of vegetation leaves is prevalent and evapotranspiration no longer proceeds at a potential rate in a vapor-saturated surface condition. Therefore, evapotranspiration in this case occurs as a

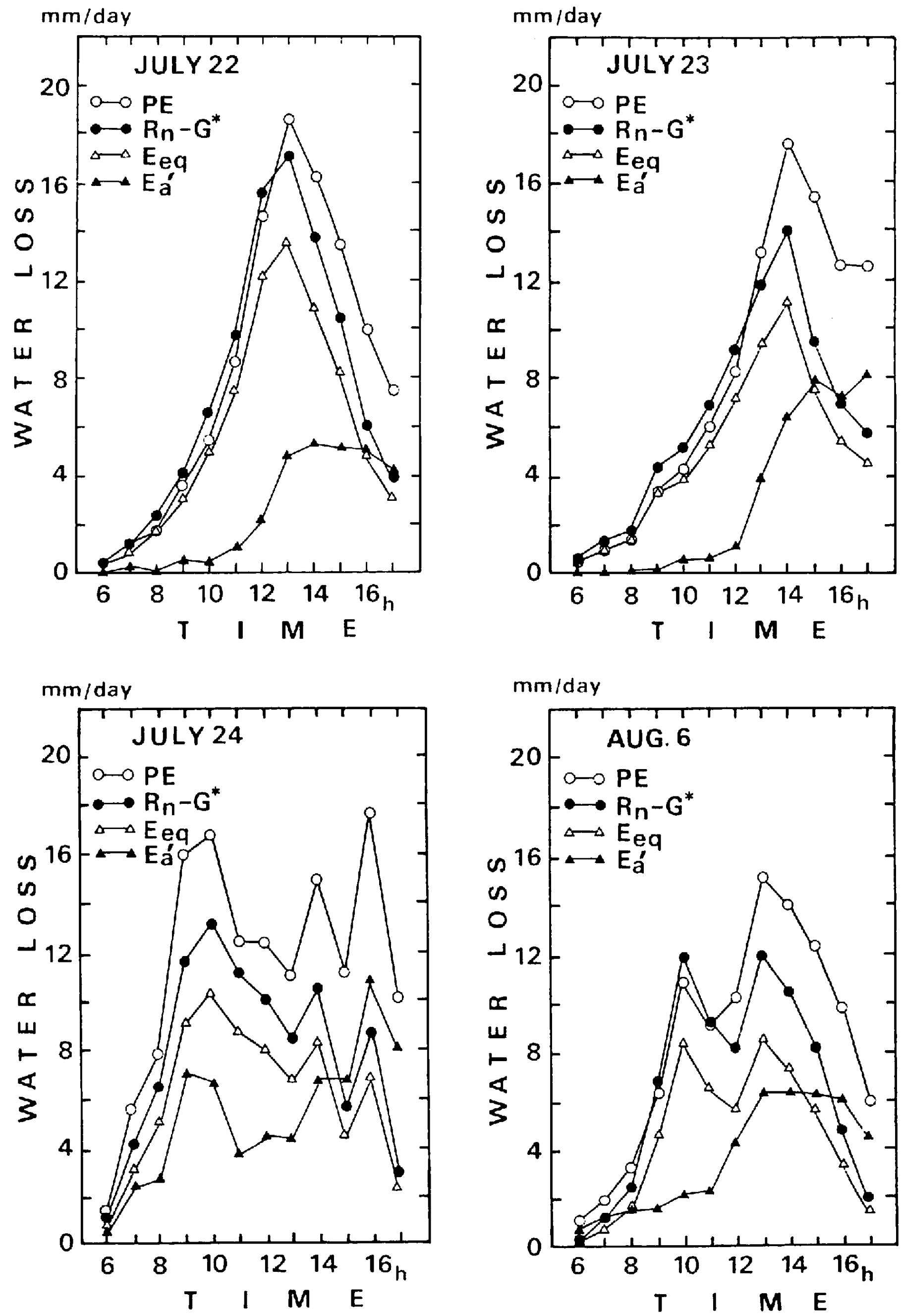

Fig. 3-5 Hourly variations of potential evapotranspiration (*PE*) defined by a vapor-saturated surface condition, available energy in evaporation equivalence ($R_n - G^*$), equilibrium evaporation (E_{eq}), and ventilation term in *PE* (E_a').

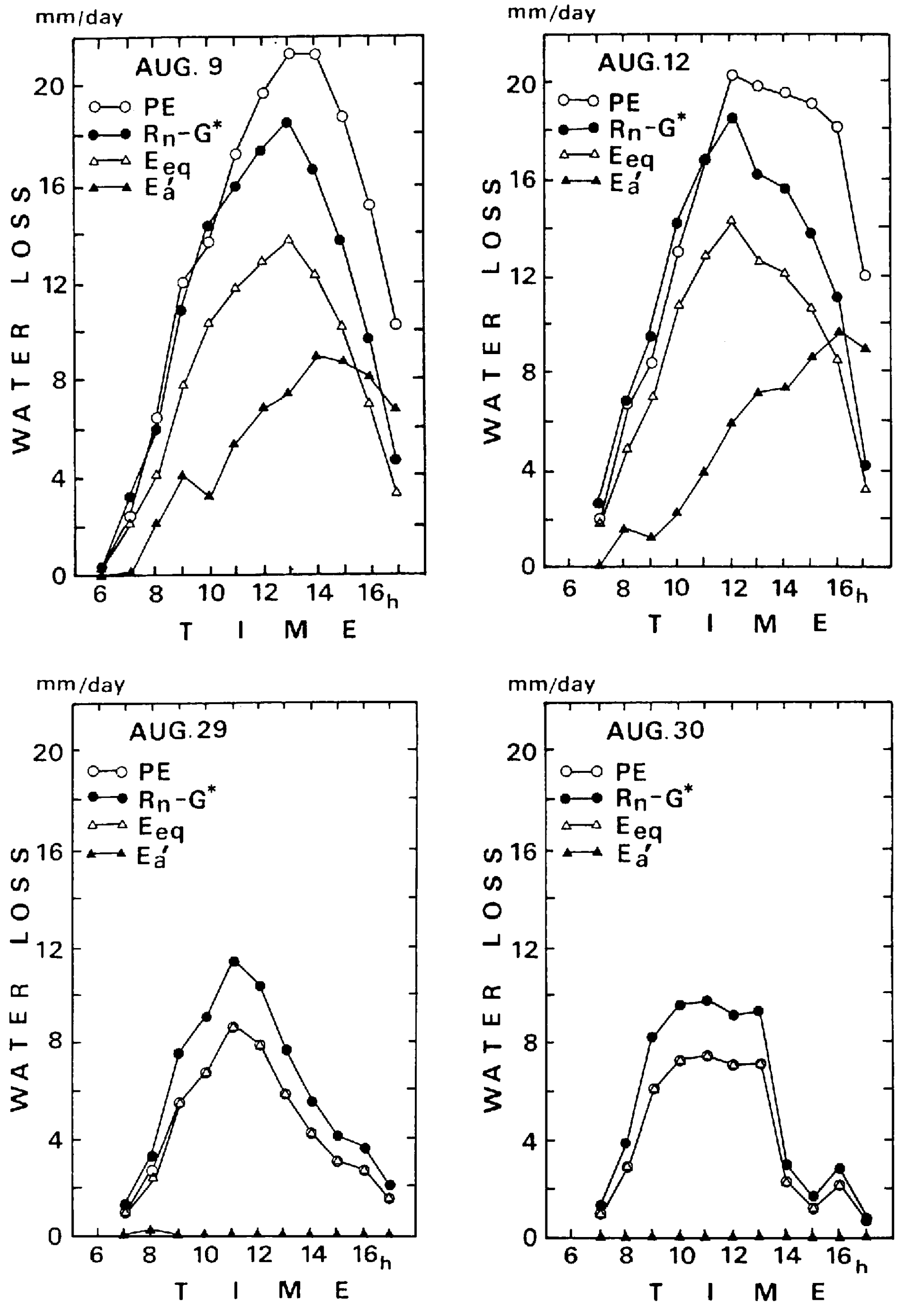

Fig. 3-5 (continued)

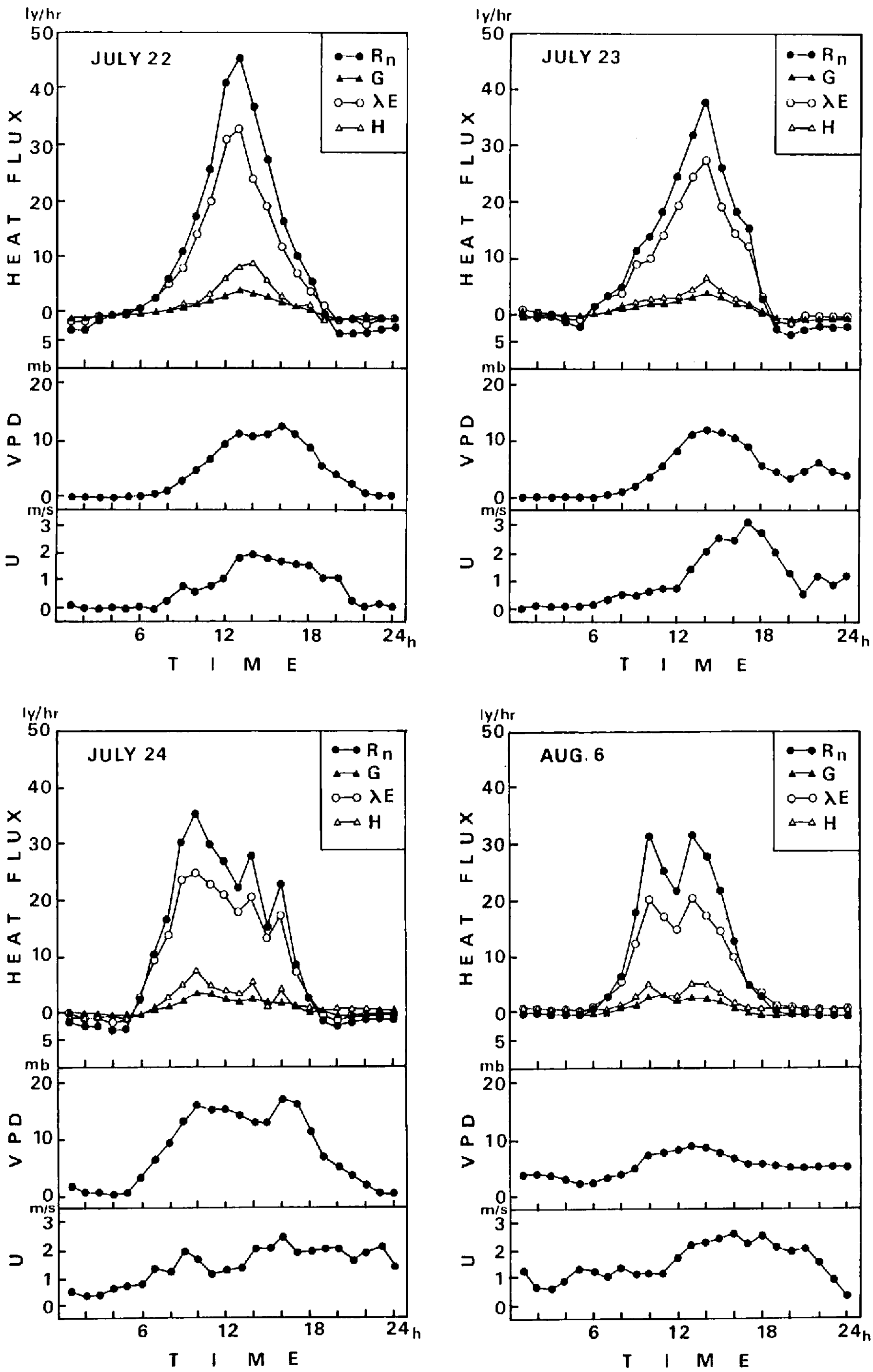

Fig. 3-6 Diurnal variations of net radiation (R_n), soil heat flux (G), latent heat flux (λE), sensible heat flux (H), vapor pressure deficit (VPD), and wind speed (u). Vapor pressure deficit and wind speed were measured at a height of 1.6 m.

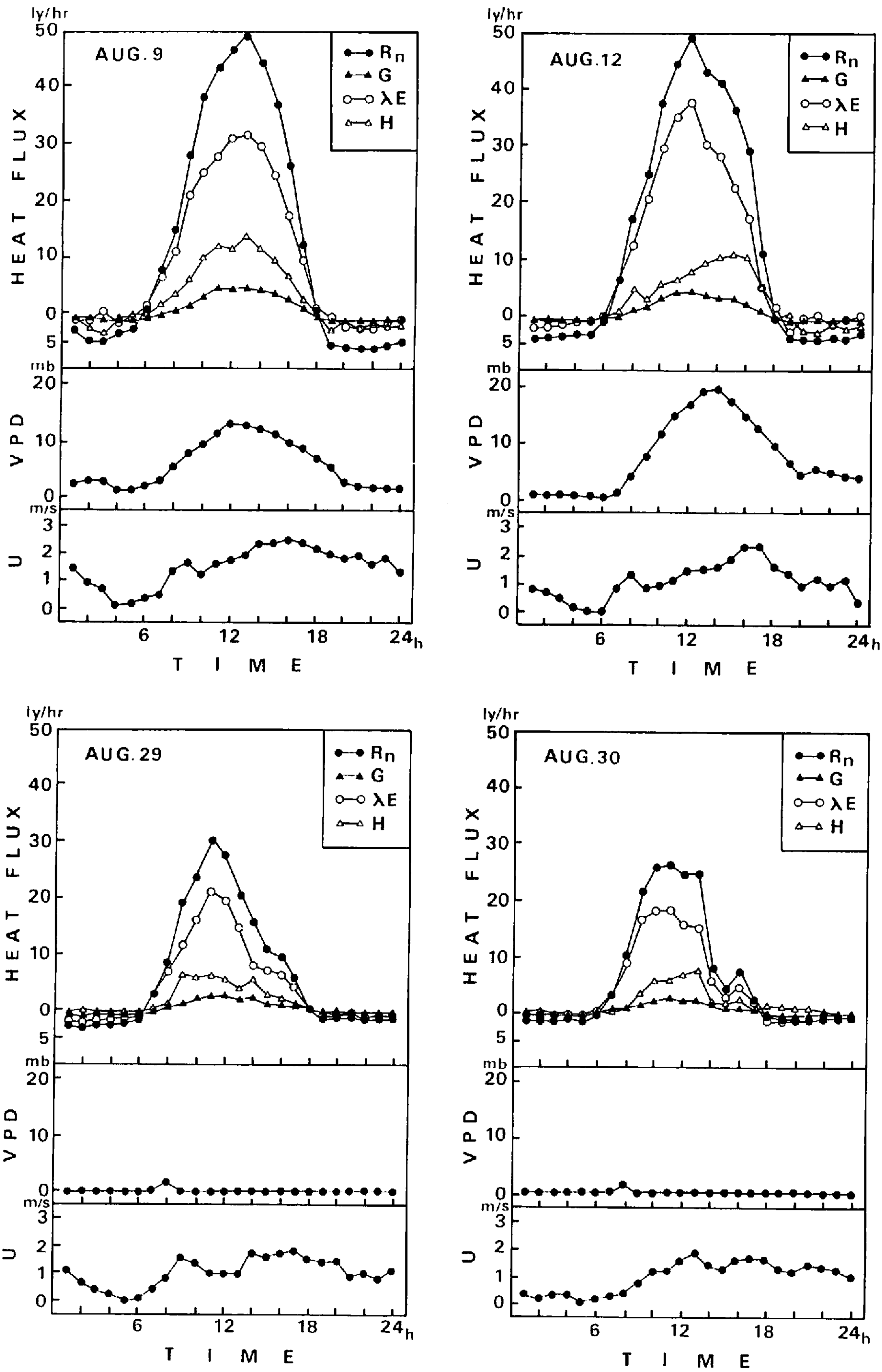

Fig. 3-6 (continued)

Table 3-2 The time when dew disappeared, when the relation $PE \leqq E$ terminated, and when the relation $PE > R_n - G^*$ began.

date	dew diminishing time (hr)	relation of $PE \leqq E$ terminating time (hr)	relation of $PE > R_n - G^*$ beginning time (hr)
July 22	0800	1100	1300
July 23	0800	1200	1300
July 24	0600	0600	0700
August 9	0700	0700	0800
August 12	0900	1000	1200

rate of the potential evapotranspiration in an ample soil water condition.

In contrast to that of fine days, PE is nearly equal to E all day long on cloudy and humid days (August 29 and August 30 in Fig. 3-4). The reason of this fact is that a low vapor pressure deficit makes the ventilation term (E_a') in Eq. (3-13) very small and consequently PE is put under control of the radiation term (E_{eq}) only (see Fig. 3-6). In such a case, $PE \cong E \cong E_{eq}$ proves to be true (see Fig. 3-5). In this case, $\beta = \gamma/\Delta$ can be deduced from Eqs. (3-6) and (3-13) and Bowen's ratio becomes a function of temperature only.

As stated before, there have been two types of definitions of potential evapotranspiration and confusion occurs in the application and the use of the idea of potential evapotranspiration. The following becomes clear from the analysis of evapotranspiration from actively growing pasture with no shortage of water in the root zone: the potential evapotranspiration as defined by a vapor-saturated surface condition is applicable only to the completely wet condition of an evaporating surface and to a vapor-saturated surface boundary layer. The former condition is established in the case where the leaf surface is completely wetted by dewfall or intercepted rainfall. However, such a condition does not last for a long period of time, because of the small interception capacity of pasture canopy.

Excluding the case mentioned above, the difference between the potential evapotranspiration as defined by a vapor-saturated surface condition and that defined by an ample soil water condition is found to be large. Sometimes the former becomes nearly twice as large as the latter. Federer (1975) stated that for a grass surface as initially postulated by Penman, the difference between values for the two kinds of definitions of potential evapotranspiration is small. This is because of a large r_a as compared to r_c. However, in the case of a forest, the difference becomes large as a result of a small r_a. The results obtained in this analysis show that even for a grass such as pasture, the difference between the two kinds of definitions of potential evapotranspiration cannot be considered small and that there are distinctively different conditions as to which the definitions can be applied.

3-3 Relationship between actual evapotranspiration and equilibrium evaporation

The concept and equation of equilibrium evaporation (E_{eq}) first proposed by Slatyer and McIlroy (1961) has been adapted widely because of its simple form. Later, Priestley and Taylor

(1972) took equilibrium evaporation as the basis for an empirical relationship for potential evaporation over a horizontally uniform saturated surface under conditions of minimal advection. They analyzed data obtained over ocean and saturated land surfaces to find the following equation for potential evaporation (E_p):

$$\lambda E_p = \alpha \frac{\Delta}{\Delta + \gamma} (R_n - G) = \alpha \lambda E_{eq} \tag{3-15}$$

where α is a parameter. They concluded that for a large saturated land surface and what they termed an 'advection-free' water surfaces the best estimate of λE_p was obtained with $\alpha = 1.26$. The method proposed by Priestley and Taylor (1972) has been applied to a variety of surfaces owing to its simple and reasonable form. As described in Chapter 1, $\alpha = 1.26$ was obtained for evaporation over open water surfaces. It has been reported, however, that the value of α is not necessarily equal to 1.26 over bare soil or vegetated surfaces and that it varies depending upon surface properties. For example, Barton (1979) for bare soil, McNaughton and Black (1973), Stewart and Thom (1973), Black (1979), and Spittlehouse and Black (1981) for forests, and Rouse and Stewart (1972), Yap and Oke (1974), and McNaughton et al. (1979) for crops have determined the value of α to be less than 1.26 from observations under ample soil water conditions.

Analyses were made to determine the value of α for pasture. As described in Section 3-2, soil water did not limit the evapotranspiration during the observation. Figure 3-7 shows the hourly variations of α as obtained from hourly values of actual evapotranspiration and equilibrium evaporation. Although the hourly variations of α differ from day to day, it can be seen as a general pattern that α takes maxima early in the morning and late in the afternoon, and that a minimum occurs near midday with few exceptions (e.g., August 12, 13, and 29). Except for these days, α values are between 0.95 and 1.4. The diurnal variations of α were mentioned for a grass by Yap and Oke (1974) and for a lake by de Bruin and Keijman (1979). Yap and Oke (1974) noticed from observations over an extensive grass surface at Ladner, British Columbia that α showed a midday minimum, followed by an increasingly sharp rise in mid-to-late afternoon. They explained the reason for this diurnal pattern of α, saying that sensible heat flux dropped off more rapidly than latent heat flux in the afternoon, and that latent heat flux commonly peaked 1~2 hr later than sensible heat flux. De Bruin and Keijman (1979) also found similar diurnal variations of α, with a minimum early in the day and a maximum late in the afternoon. Their observations were made over a large, shallow lake in the Netherlands. They concluded that diurnal variations of α were caused by the different diurnal variation patterns of air temperature and surface water temperature.

It is worth noting that the diurnal pattern of α obtained in this analysis differs from the previous observational results described earlier. That is, large α values occur not only in the afternoon but also early in the morning. In addition, the value of α early in the morning is larger than that in the late afternoon. The large α value observed in the early morning is considered to be caused by the fact that evaporating surface acts as a completely saturated surface due to the evaporation of nighttime dewfall. This occurs in the same way as discussed in the previous Section. The gradual drop of α observed late in the afternoon on August 12 and 13 may be attributed to an increase in the aridity of air caused by an increasing vapor pressure deficit and an increase in wind speed in the afternoon.

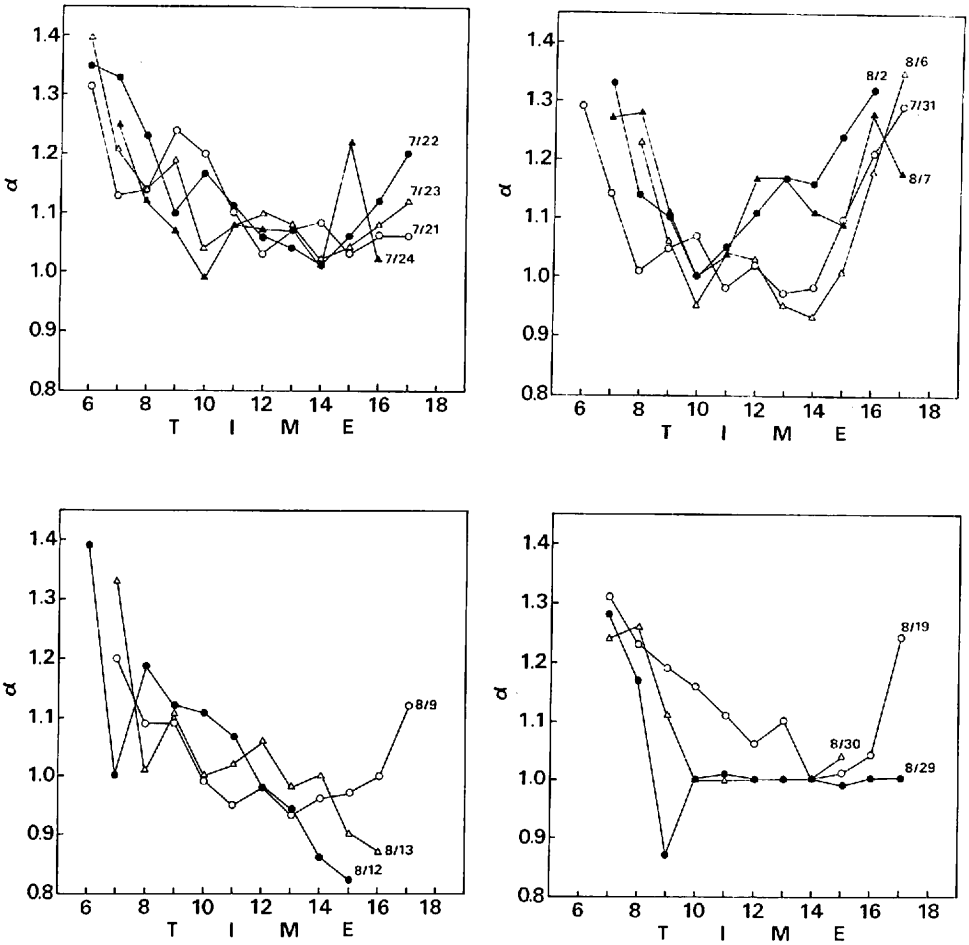

Fig. 3-7 Hourly variations of the proportional constant (α) in the Priestley and Taylor potential evaporation equation.

Figure 3-8 shows the relationship between actual evapotranspiration (E) and equilibrium evaporation (E_{eq}). The lines in Fig. 3-8 represent the values of α. It can be concluded from Fig. 3-8 that, on the whole, actual evapotranspiration falls in the range between the equilibrium evaporation ($\alpha=1$) defined by Slatyer and McIlroy (1961) and the potential evaporation ($\alpha=1.26$) by Priestley and Taylor (1972). From Figs. 3-7 and 3-8, the value of α early in the morning is usually near 1.26 but for the rest of the day α tends to be smaller than 1.26.

As discussed in the previous Section, the relationship of $PE \cong E$ is established during the evaporation of dew. The average value of α during dew evaporation proved to be 1.25 ± 0.02, which is very close to the value $\alpha = 1.26$ for saturated surfaces obtained by Priestley and Taylor (1972), Stewart and Rouse (1976, 1977), and de Bruin and Keijman (1979). The $\pm$ notation is used to denote the standard error of the mean. On the other hand, actual evapotranspiration is nearly equal to equilibrium evaporation on very humid days. The overall mean of α is taken as

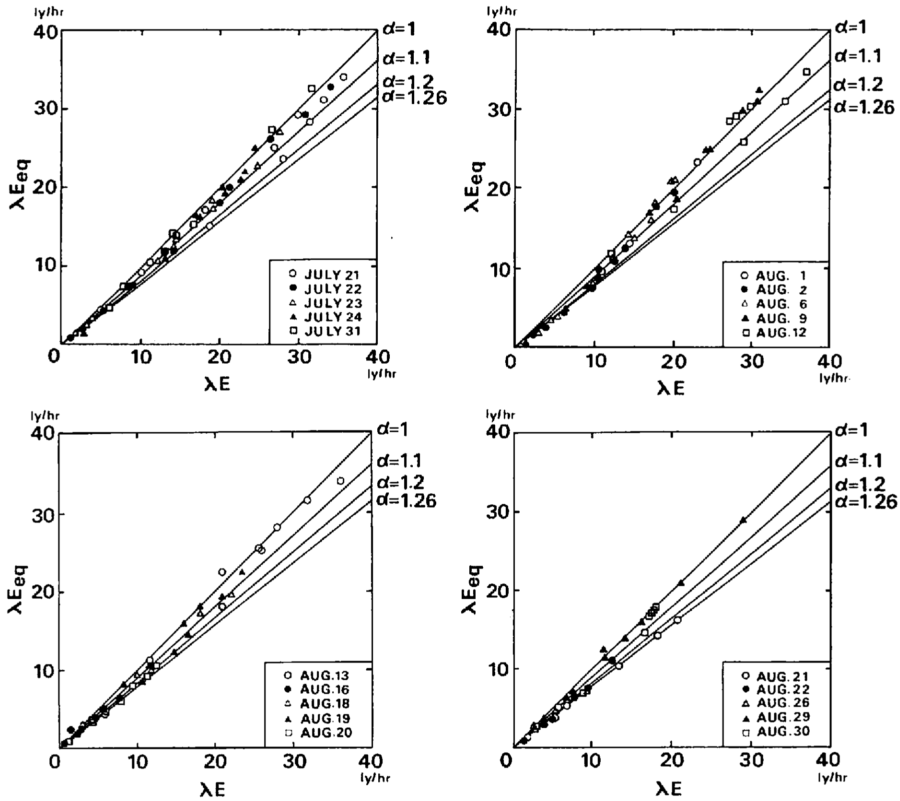

Fig. 3-8 Relationship between hourly actual latent heat flux (λE) and hourly equilibrium latent heat flux (λE_{eq}).

1.16 ± 0.01, which is smaller than the value of α for completely wet surfaces.

From the above discussion, it becomes clear that the upper limit of evapotranspiration is represented by the potential evaporation equation by Priestley and Taylor (1972) and the lower limit by the equilibrium evaporation equation by Slatyer and McIlroy (1961) for actively growing pasture with no shortage of water. Actual evapotranspiration, however, is usually smaller than potential evaporation even if there is no shortage of soil water for evapotranspiration. Actual evapotranspiration becomes equal to potential evaporation only when the evaporating surface acts as a completely saturated surface during the evaporation of dew.

3-4 Validity of the concepts on evapotranspiration

Potential evapotranspiration as first presented in 1944 and formulated independently by Thornthwaite and Penman has been widely used for studies on evapotranspiration due to its easiness of computation. However, these studies have been conducted without detailed discussion as to the definition of potential evapotranspiration. Progress in designing instruments for measuring and theories describing the transport phenomena in the surface boundary layer have promoted direct measurements of evapotranspiration over various surfaces. Thus, evaluations of evaporation formulae have been conducted. The use of the term "potential evapotranspiration", however, has been made without discrimination. Consequently, there is confusion as to the interpretation of and the use of the term "potential evapotranspiration". That is, principally two kinds of interpretation still exist. One is to consider potential evapotranspiration as the evapotranspiration that would occur if an evaporating surface is vapor-saturated. This interpretation may be deduced from the assumption that under a non-stressing soil water condition the vapor pressure deficit at the evaporating surface of actively growing vegetation becomes negligible. The other interpretation is to consider potential evapotranspiration as the evapotranspiration that would occur under an ample soil water condition. In this interpretation, the nature of the evaporating surface is not well specified.

The term "evaporation" is often used as a generalized term which describes all the processes of water transfer from the earth's surface into the atmosphere, and the term "potential evaporation" is frequently used as a synonym for "potential evapotranspiration". In addition, Priestley and Taylor (1972) proposed the term "potential evaporation" in the context of more closely defining the nature of the evaporating surface.

In this Chapter, the terms "potential evapotranspiration", "equilibrium evaporation", and "potential evaporation" were taken as the basic concepts on evapotranspiration and analyses were made to clarify the conditions to which they can be applied. Data obtained from actively growing pasture with no soil water control of evapotranspiration were used as the basic data for the analyses. The actual evapotranspiration during the observation could be considered to be the potential evapotranspiration under a sufficient soil water supply condition. The equation of potential evapotranspiration under a vapor-saturated surface condition proved to overestimate the potential evapotranspiration with an ample soil water condition. The potential evapotranspiration under a vapor-saturated surface condition and actual evapotranspiration are equal only when evaporation of dew occurs or when the air is very humid. This fact indicates that, in the case of vegetation, the assumption of a vapor-saturated condition of an evaporating surface can be established in the case of an externally wetted condition. This condition is a result of either dewfall or interception of rainfall. When an evaporating surface is not externally wetted, water is transferred through stomata of leaves. Under this condition, evapotranspiration is lower than the potential evapotranspiration of a vapor-saturated surface condition. In this case, a finite canopy resistance is exerted on the water transfer, even though there is sufficient soil water.

The canopy resistance (r_c) in the observation period was examined. The canopy resistance was determined by the following equation, which was obtained by combining Eqs.(1-6) and (3-13). That is,

$$r_c = r_a \left(1 + \frac{\Delta}{\gamma} \right) \left(\frac{PE}{E} - 1 \right) \tag{3-16}$$

Figure 3-9 shows the hourly variations of canopy resistance. The very low values of r_c in the morning are considered to be caused by the evaporation of dew. After taking upon low values in the morning, r_c increases rapidly in the afternoon. The increase in r_c in the afternoon is caused by the stomatal response to decreasing light intensity and increasing vapor pressure deficits. Most of the r_c values are below or nearly equal to 1.0 sec/m, except in the late afternoon on August 12 and 13, when the drying power of the air was large. The values of r_c in this study are similar to those reported by Szeicz and Long (1969) for irrigated grass.

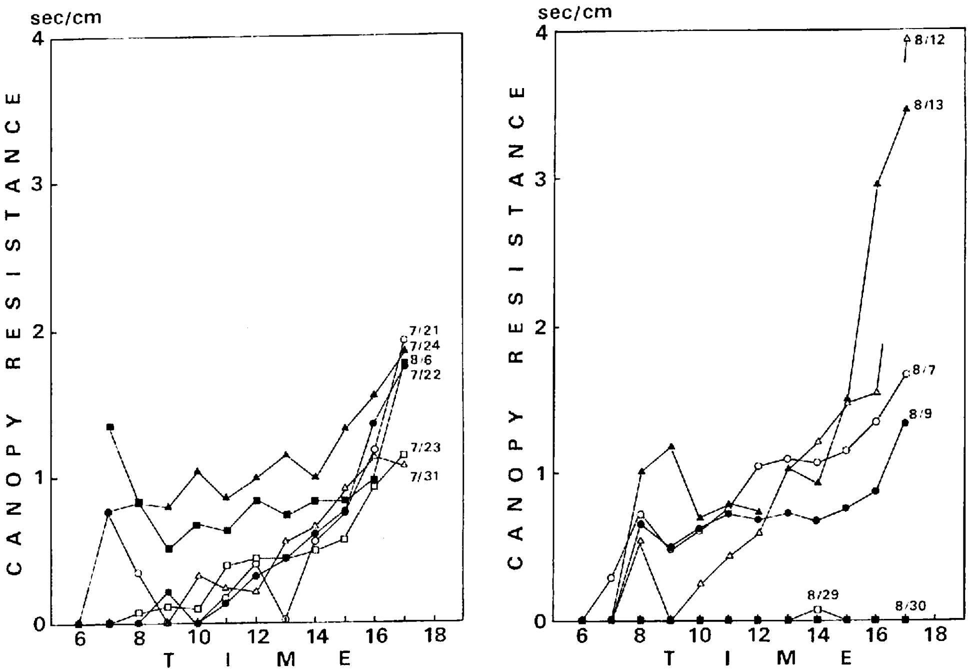

Fig. 3-9 Hourly variations of canopy resistance.

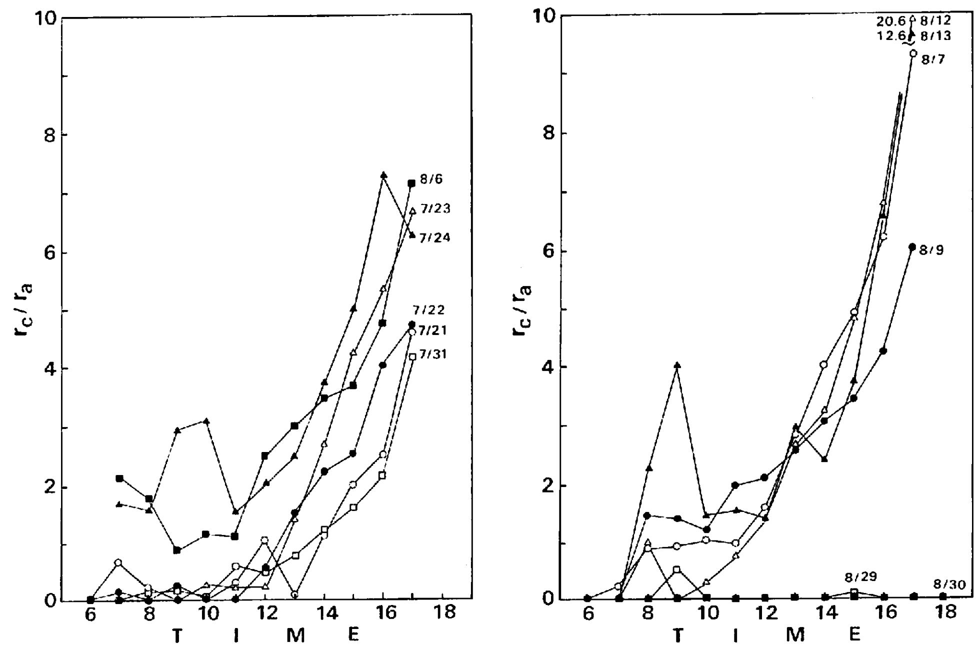

Fig. 3-10 Hourly variations of the ratio of canopy resistance (r_c) to aerodynamic resistance (r_a).

Figure 3-10 shows the hourly variations of r_c/r_a. From Eq. (3-16),

$$\frac{E}{PE} = \frac{\Delta + \gamma}{\Delta + \gamma\,(1 + r_c \,/\, r_a)} \tag{3-17}$$

is obtained. From Eq. (3-17), r_c/r_a is used as a basis for consideration of the relative evaporation ratio (E/PE). As shown in Fig. 3-10, a low value of r_c/r_a in the morning is followed by a sharp rise in the afternoon. It is considered that the low r_c/r_a in the morning is due to a low r_c and that the high r_c/r_a in the afternoon is due to a large vapor pressure deficit and strong wind. The value of r_c/r_a in the midday and in the afternoon exceeds unity and sometimes reaches close to 10.

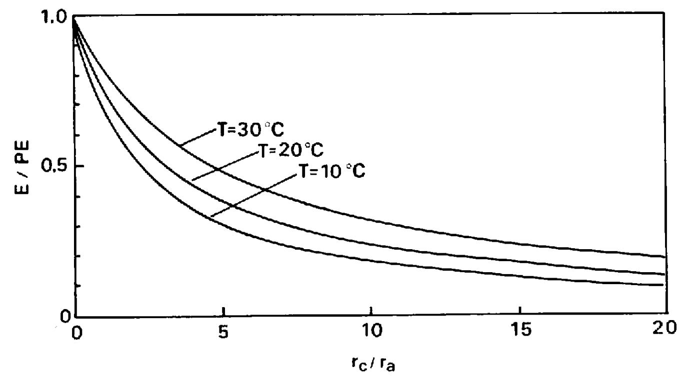

Fig. 3-11 Changes in evaporation ratio (E/PE) with r_c/r_a and temperature (T).

Figure 3-11 shows the relationship represented by Eq. (3-17). From Fig. 3-11, it can be seen that the increase in r_c/r_a corresponds to the decrease in E/PE. It is generally recognized that for field crops with the same order of magnitude of r_c and r_a, E/PE does not become much smaller than unity, but that for forests with a much larger r_c than r_a, E/PE becomes very small. This occurs even under ample soil water conditions (Rutter, 1975). If r_c/r_a is about 4, the typical value in the afternoon in this study, PE is shown to be nearly twice as large as E.

On the other hand, the analysis of evapotranspiration, with the equilibrium evaporation as a basis, clarifies that the "potential evaporation" as proposed by Priestley and Taylor (1972) can be applied only to the situation when the evaporation of dew takes place. In addition, the Priestley and Taylor equation overestimates the actual evapotranspiration in other situations, even under plentiful soil water content. Although the evaporating surface acts as a freely evaporating surface during dew evaporation, stomata regulate water transfer in the absence of dew. For equal weather conditions, the evapotranspiration rate under wetted surface conditions proves to be about 10% greater than that under dry surface conditions.

The fact that the potential evaporation is equal to the potential evapotranspiration under a vapor-saturated surface condition, during the evaporation of dew, indicates that the potential evaporation by Priestley and Taylor (1972) can be applied only to saturated, freely evaporating surfaces.

Even under an adequate soil water condition, evapotranspiration depends considerably on the wetness of an evaporating surface. Although this fact is generally taken into consideration in the studies on evaporation of intercepted rainfall over forests (e.g., McNaughton and Black, 1973; Stewart and Thom, 1973), little attention has been paid to low vegetation with a small interception capacity, resulting in the application of evaporation formulae indiscriminately. Therefore, the nature of an evaporating surface, especially its wetness, should be specified in the application of potential evapotranspiration or potential evaporation.

It is also clarified that the equilibrium evaporation represents the lower limit of evapotranspiration from actively growing pasture with no shortage of soil water, and that evapotranspiration falls in the range of 1.0 to 1.26 times as large as the equilibrium evaporation.

CHAPTER 4

ANALYSIS OF DAILY EVAPOTRANSPIRATION

The applicable conditions of several concepts of evapotranspiration were discussed in Chapter 3, using hourly data as a basis. These concepts were potential evapotranspiration, equilibrium evaporation, and potential evaporation. In water balance studies, daily values of evapotranspiration are often used as the basic data for investigations. The same procedure as that in Chapter 3 is used for the analysis of daily evapotranspiration rates. An evapotranspiration model, which is based on equilibrium evaporation, is presented and tested by the use of field data.

4-1 Relationship between actual evapotranspiration and potential evapotranspiration

As described in Chapter 1, there are two kinds of interpretation of the wetness conditions of an evaporating surface under a state of potential evapotranspiration; one assumes a vapor-saturated surface condition and the other assumes a well-watered soil condition. The Monteith equation (Eq. 1-6) with $r_c = 0$ was used as the equation which defines the potential evapotranspiration under a vapor-saturated surface condition. The equation is as follows:

$$\lambda E = \frac{\Delta}{\Delta + \gamma}(R_n - G) + \frac{\rho c_p}{\Delta + \gamma}(e_a{}^* - e_a)/r_a \qquad (4\text{-}1)$$

On the other hand, the Penman equation, using actual radiation over a vegetated surface instead of that over a hypothetical open water surface, was used as the equation for the potential evapotranspiration defined by a well-watered soil condition. The equation is

$$\lambda E = \frac{\Delta}{\Delta + \gamma}(R_n - G) + \frac{\lambda\gamma}{\Delta + \gamma} f(u)(e_a{}^* - e_a) \qquad (4\text{-}2)$$

As mentioned in Chapter 1, the following two kinds of wind functions were proposed:

$$f(u) = 0.26\,(1 + 0.54u) \qquad (4\text{-}3)$$

and

$$f(u) = 0.26\,(0.5 + 0.54u) \qquad (4\text{-}4)$$

The potential evapotranspiration under a well-watered soil condition was calculated by inserting Eqs. (4-3) and (4-4) into Eq. (4-2). Wind speed, air temperature, and vapor pressure measured at a height of 1.6 m were used in Eqs. (4-1) and (4-2). Actual evapotranspiration was measured by a weighing lysimeter. The data used in this analysis were obtained during the same period as in the analysis of hourly data. As discussed in Chapter 3, it could be considered that soil water did not limit evapotranspiration during the observation period.

In Fig. 4-1, the calculated daily evapotranspiration (E_{cal}) from Eqs. (4-1) or (4-2) is compared

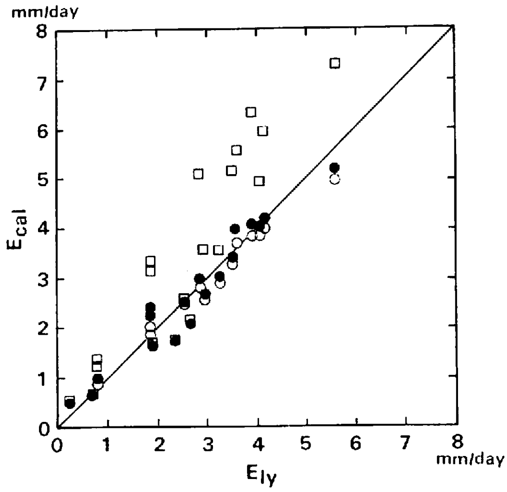

Fig. 4-1 Relationship between actual evapotranspiration (E_{ly}) by a weighing lysimeter and calculated evapotranspiration (E_{cal}) by evaporation formulae.

- • Penman method (with Eq. 4-3)
- o Penman method (with Eq. 4-4)
- □ potential evapotranspiration defined by a vapor-saturated surface condition

with the actual daily evapotranspiration (E_{ly}) measured by a weighing lysimeter. As can be seen in Fig. 4-1, the estimated potential evapotranspiration from Eq. (4-1) tends to overestimate the actual evapotranspiration. Overestimation reaches about 80% in extreme cases. The reason is that a completely wetted surface condition, i.e., $r_c = 0$, could be established only for a short period during the evaporation of dew, as stated in Chapter 3. As shown in Table 4-1, the potential evapotranspiration described by Eq. (4-1) overestimates the actual evapotranspiration by 35% in total. The canopy resistance (r_c) is shown to be larger than the aerodynamic resistance (r_a). Since the aerodynamic resistance of tall grass is effectively smaller than that of short vegetation, surface effects due to canopy resistance are at least as important as those of ventilation, caused by aerodynamic resistance, in determining the rate of evapotranspiration from pasture.

On the other hand, close agreement was obtained between the estimated potential evapotranspiration shown by Eq. (4-2) and the actual evapotranspiration (see Fig. 4-1 and Table 4-1). There exists little difference between the values estimated by the Penman method with different wind functions. The reason for this close agreement between E_{cal} from Eq. (4-2) and E_{ly} is considered to be that the effects of canopy resistance and of aerodynamic resistance are approximately self-cancelling as stated by Thom and Oliver (1977). This is a prominent feature of the Penman equation, which, in spite of its empiricism, has been remarkably successful in estimating potential evapotranspiration.

Table 4-1 Daily evapotranspiration rates and daily average resistances.

	E_{eq} mm/d	PE_{w1} mm/d	PE_{w2} mm/d	PE_{sv} mm/d	E_{ly} mm/d	r_c s/cm	r_a s/cm	r_c/r_a
July 20	4.45	4.96	5.19	7.32	5.57	0.47	0.39	1.21
21	3.58	3.87	4.06	4.96	4.06	0.54	0.61	0.89
22	2.72	2.92	3.05	3.56	3.26	0.28	0.75	0.37
23	2.29	2.56	2.70	3.60	2.95	0.47	0.51	0.92
*30	2.15	2.34	2.42	3.36	1.83	0.85	0.31	2.74
Aug. 2	1.62	1.90	2.02	3.18	1.83	0.87	0.38	2.29
6	2.34	2.81	3.02	5.10	2.87	0.89	0.37	2.41
9	3.30	3.85	4.08	6.34	3.90	0.73	0.37	1.97
12	3.51	4.01	4.27	5.97	4.14	0.83	0.51	1.63
13	3.32	3.76	4.01	5.62	3.50	1.08	0.51	2.12
15	2.50	2.49	2.50	2.63	2.55	0.05	0.34	0.15
16	0.63	0.62	0.62	0.63	0.24	2.59	0.49	5.24
*19	2.16	2.13	2.13	2.16	2.63	–	0.44	–
22	0.84	0.93	0.98	1.28	0.80	1.10	0.53	2.08
23	0.98	1.10	1.15	1.65	0.80	1.53	0.39	3.92
25	2.93	3.28	3.38	5.19	3.50	0.39	0.24	1.63
26	0.66	0.65	0.65	0.66	0.72	–	0.28	–
29	1.76	1.74	1.74	1.76	2.31	–	0.42	–
*30	1.67	1.68	1.68	1.67	1.91	–	0.57	–
total	43.41	47.60	49.65	66.64	49.37			

* day with rainfall

E_{eq} : equilibrium evaporation
PE_{w1} : potential evapotranspiration by Eqs. (4-2) and (4-4)
PE_{w2} : potential evapotranspiration by Eqs. (4-2) and (4-3)
PE_{sv} : potential evapotranspiration by Eq. (4-1)
E_{ly} : actual evapotranspiration by a weighing lysimeter
r_c : canopy resistance
r_a : aerodynamic resistance

4-2 Relationship between actual evapotranspiration and equilibrium evaporation

Figure 4-2 shows the relationship between the daily actual evapotranspiration (E_{ly}) and the daily equilibrium evaporation (E_{eq}). The proportional constant (α) of the Priestley and Taylor method (Eq. 3-15) is also shown in Fig. 4-2. The proportional constant (α) ranges from 1.0 to 1.26. This is the same range as the results obtained from the analysis of hourly data in Chapter 3. The average value of α is found to be 1.14 ± 0.03, which is almost the same average value of α (=1.16) obtained from the analysis of hourly data.

Yap and Oke (1974), Kanemasu et al. (1976), and Tanner and Jury (1976) reported that a smaller value of α will be obtained if daytime available energy is used rather than daily available energy. The reason for this difference in α, with averaging period, is considered to be a loss of available energy during the nighttime. Since the data used in this Chapter is not consistent with that in the previous Chapter, no discussion is made here for the difference in α with averaging period.

Daily actual evapotranspiration is found to be less than the potential evaporation (see Fig. 4-2). During the period listed in Table 4-1, the total amount of potential evaporation is 54.70 mm, which is about 10% larger than the actual evapotranspiration of 49.37 mm. This fact re-emphasizes the important regulating function of the surface wetness in evapotranspiration, even under a non-limiting soil water condition. It was found in Chapter 3 that the amount of dewfall over one night was 0.4 mm at a maximum and the latest dew evaporation terminated by 0900h. Hence, it is considered that dew evaporation does not influence the daily evapotranspiration amount.

The close relationship between daily equilibrium evaporation and actual evapotranspiration suggests that radiation is the dominant factor in determining evapotranspiration and that the effects of wind and humidity are relatively small.

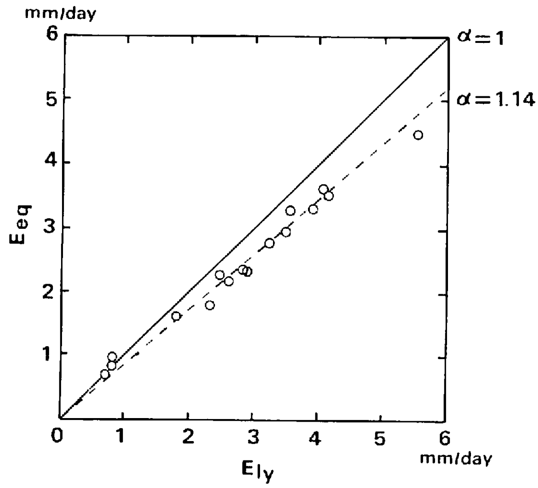

Fig. 4-2 Relationship between daily actual evapotranspiration (E_{ly}) and equilibrium evaporation (E_{eq}).

4-3 Development and test of equilibrium evaporation model

As stated in the previous Section, daily evapotranspiration from actively growing pasture under a non-limiting soil water condition can be expressed as

$$\lambda E = \alpha \lambda E_{eq} \tag{4-5}$$

where α is a constant equal to 1.14. Hereafter, the estimation method of evapotranspiration with equilibrium evaporation as a basis is referred to as "equilibrium evaporation model", which is expressed by Eq. (4-5). Since the equilibrium evaporation model does not require the data of wind speed and vapor pressure, it is simpler than the Penman method.

The test of $\alpha = 1.14$ in the equilibrium evaporation model is made for the whole observation period in the summer of 1980 (from July 20 to August 31). Figure 4-3 shows the daily variations of actual evapotranspiration (E_{ly}) measured by a weighing lysimeter and estimated evapotranspiration (E_{es}) from the equilibrium evaporation model with $\alpha = 1.14$. The plots for August 5 are missing because of lack of data. Evapotranspiration was zero on August 3 due to an all-day rain. As can be seen in Fig. 4-3, the daily variations of E_{es} and E_{ly} are closely related.

The summer of 1980 was a season of unusual weather because of low rainfall, low sunshine, and low temperature (Murakami, 1981). Hence, it is questionable whether the value of α obtained in 1980 is valid for other years. To investigate this question, the equilibrium evaporation model was tested again using summer data of 1978. The summer of 1978 was hot, in contrast to the summer of 1980. In 1978, net radiation and soil heat flux measurements were only available from August 4 to 24. Therefore, the relationship between actual evapotranspiration (E_{ly}) measured by a weighing lysimeter and estimated evapotranspiration (E_{es}) using the equilibrium evaporation model with $\alpha = 1.14$ was examined for the above period. Figure 4-4 represents the daily patterns of E_{ly} and E_{es}. Similar to the results for 1980, there exists a close relationship between E_{ly} and E_{es}, which confirms the validity of $\alpha = 1.14$ in the equilibrium evaporation model.

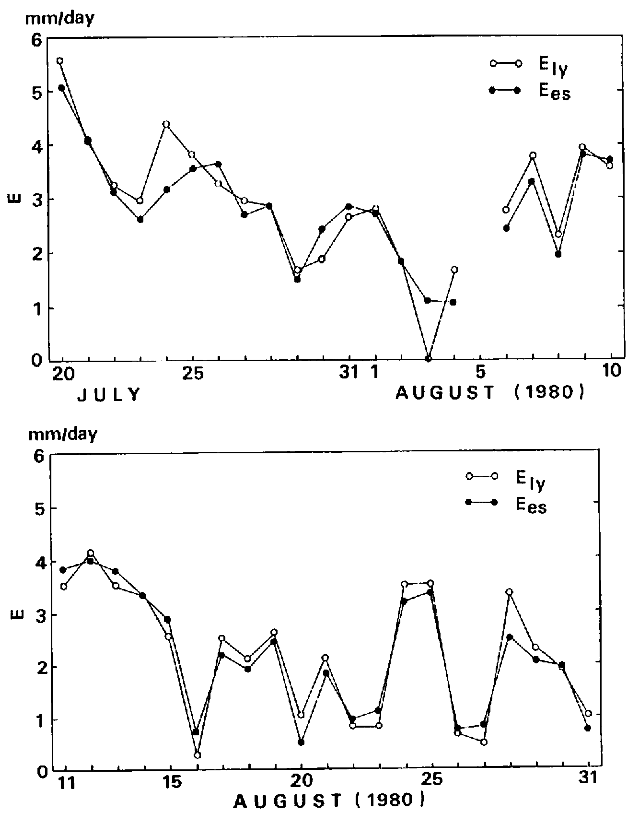

Fig. 4-3 Day-to-day variations of actual evapotranspiration (E_{ly}) and estimated evapotranspiration (E_{es}) by the equilibrium evaporation model with $\alpha = 1.14$ in 1980.

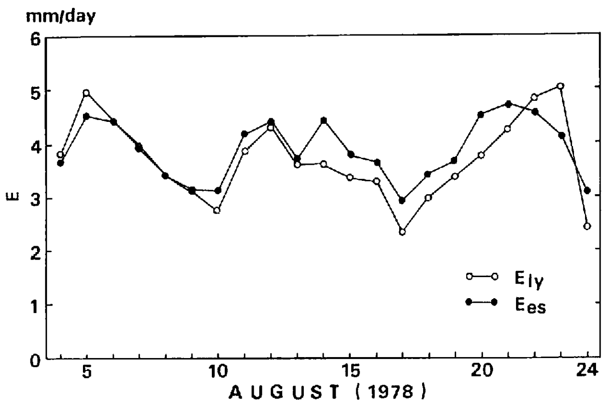

Fig. 4-4 Day-to-day variations of actual evapotranspiration (E_{ly}) and estimated evapotranspiration (E_{es}) by the equilibrium evaporation model with $\alpha = 1.14$ in 1978.

Table 4-2 Comparison of estimated evapotranspiration by the equilibrium evaporation model with actual evapotranspiration.

	E_{eq}	E_{es}	E_{ly}
1978 Aug. 4 ~ Aug. 24	71.4	81.4	77.0
1980 July 20 ~ Aug. 31	91.5	104.3	108.1

(unit: mm)

E_{eq} : equilibrium evaporation
E_{es} : estimated evapotranspiration by the equilibrium evaporation model with $\alpha = 1.14$
E_{ly} : actual evapotranspiration by a weighing lysimeter

Table 4-2 shows the totals of actual evapotranspiration (E_{ly}) measured by a weighing lysimeter, equilibrium evaporation (E_{eq}), and estimated evapotranspiration (E_{es}) using the equilibrium evaporation model with $\alpha = 1.14$ for both observation periods. It can be seen in Table 4-2 that the estimation of evapotranspiration can be made successfully with an accuracy of about 5% with the equilibrium evaporation model, using $\alpha = 1.14$.

It can be concluded from the above discussion that daily evapotranspiration from actively growing pasture, which completely shades the ground and has no soil water shortage, can be estimated successfully by the equilibrium evaporation model with $\alpha = 1.14$.

CHAPTER 5

ANALYSIS OF SEASONAL VARIATIONS OF EVAPOTRANSPIRATION

A simple "equilibrium evaporation model" was developed and tested in Chapter 4. The results show that the equilibrium evaporation model well represents evapotranspiration from actively growing pasture with a non-limiting soil water supply. The equilibrium evaporation model requires very few input parameters, which makes it a promising tool for the estimation of regional evapotranspiration.

In this Chapter, the equilibrium evaporation model is applied to the estimation of seasonal evapotranspiration and problems in the application of this model are discussed. Next, the applicabilities of the Thornthwaite and the Penman methods, which are most commonly used in Japan for the estimation of evapotranspiration, are discussed.

5-1 Application of equilibrium evaporation model to seasonal evapotranspiration

Net radiation, short-wave radiation, soil heat flux, air temperature, and evapotranspiration had been measured continuously until April 17, 1981 after intensive observations during the summer in 1980. The applicabilities of the equilibrium evaporation model, especially the variations of the proportional constant α in Eq. (4-5), were investigated using data obtained from September to April.

Figure 5-1 shows the day-to-day variations of daily values of evapotranspiration (E), equilibrium evaporation (E_{eq}), available energy in evaporation equivalence ($R_n - G^*$), and α. It can be detected as general patterns that the variations of E, E_{eq}, and $R_n - G^*$ coincide well. The value of α gradually increases from September to October, although there exist sharp changes in α due to rainfall. The value of α during this period, however, is greater than the value of 1.14 obtained from the summer data (in Chapter 4). Moreover, it is greater than the value of 1.26 obtained from completely saturated surfaces. It can be found from the comparison of E with $R_n - G^*$ that α becomes greater than 1.14 on days when E is nearly equal to or greater than $R_n - G^*$.

The value of α shows a decrease early in November and an increase late in November. The value of α in November is smaller than that in September or October. This may result from the smaller number of days with $E > R_n - G^*$ in November. However, the value of α in November is still greater than that in the summer.

Little coincidence is observed between the variations of E and E_{eq} from December to February. Very large fluctuations exist in the day-to-day variations in α and the value of α sometimes becomes extremely great. The large fluctuations in α might be caused by measurement errors in E and $R_n - G^*$, because E and $R_n - G^*$ are small in winter months. However, it can be seen in Fig. 5-1 that an extremely large α occurs when E_{eq} is very small. Therefore, measurement errors cannot be considered as the only reason for the fluctuations in α.

In March and April, the fluctuations in α are not large and the value of α is less than 1.0 in most cases.

Figure 5-2 shows the relationship between daily evapotranspiration (E) and equilibrium evaporation (E_{eq}) for each month. Monthly average value of α is shown in each figure by a broken line. The value of α varies widely, especially in winter months. Also, days with $E > R_n - G^*$ can be

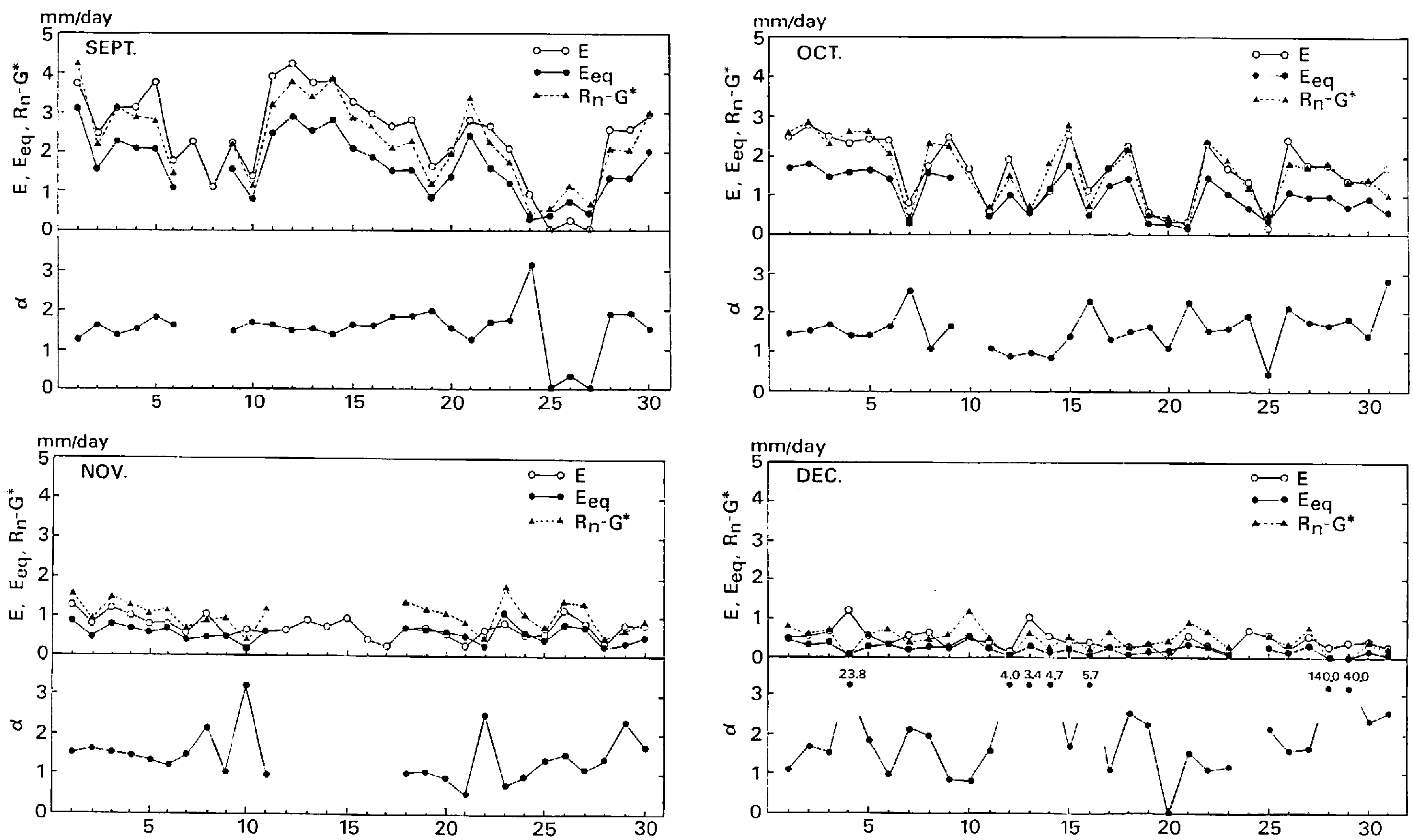

Fig. 5-1 Day-to-day variations of daily evapotranspiration (E), equilibrium evaporation (E_{eq}), available energy in evaporation equivalence ($R_n - G^*$), and the parameter α in the equilibrium evaporation model. Figures in α represent the values of α.

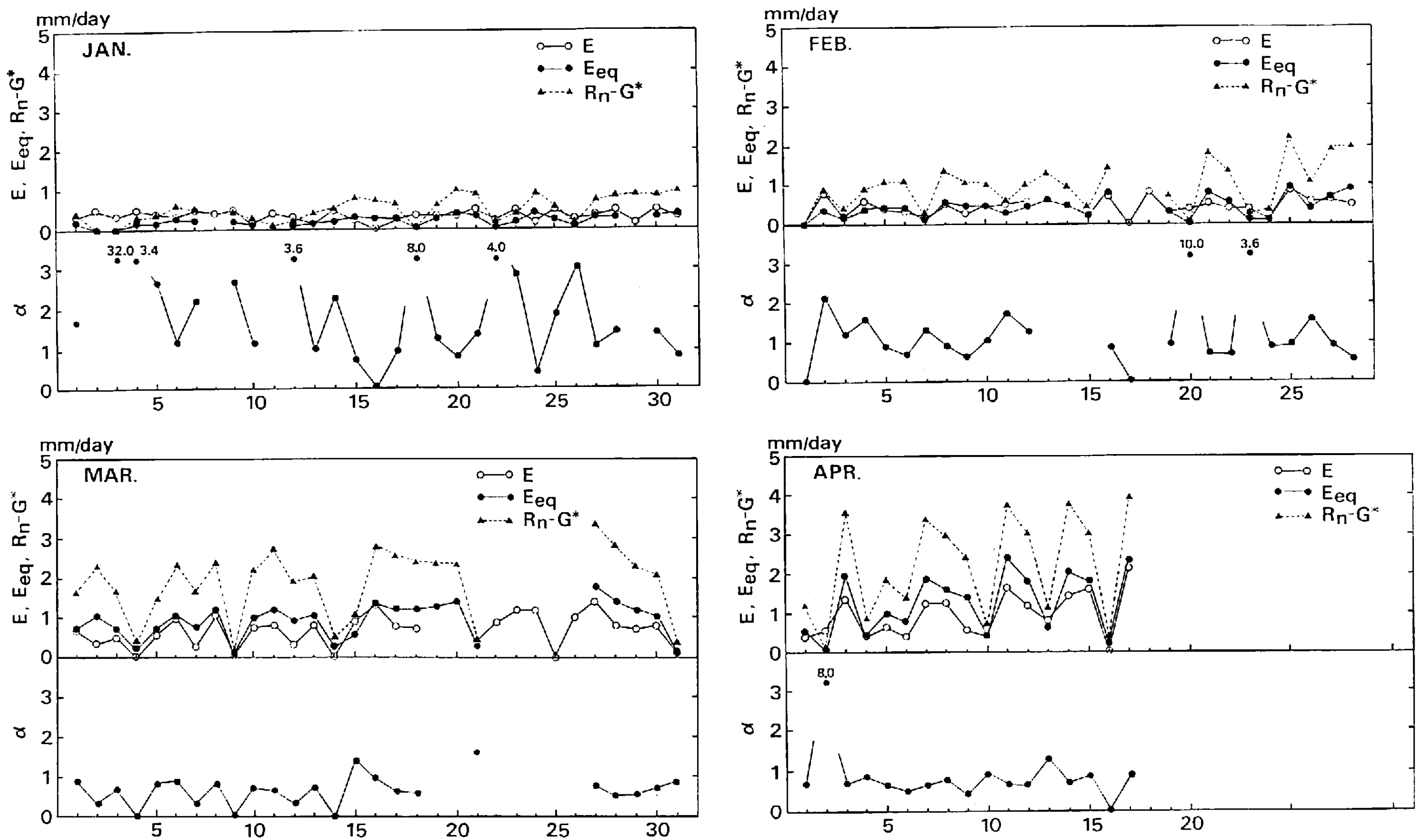

Fig. 5-1 (continued)

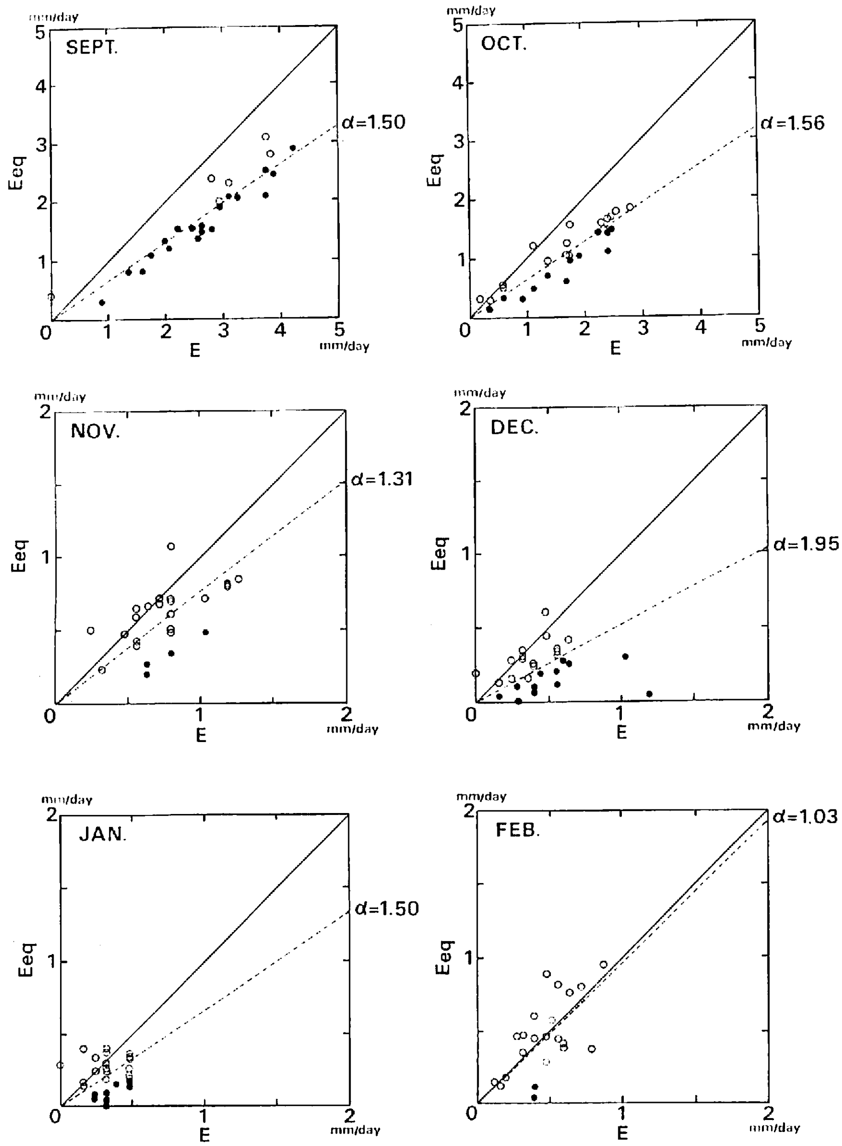

Fig. 5-2 Relationship between daily evapotranspiration (E) and equilibrium evaporation (E_{eq}). Values of α represent monthly mean values.

○ $E \leqq R_n - G^*$

● $E > R_n - G^*$

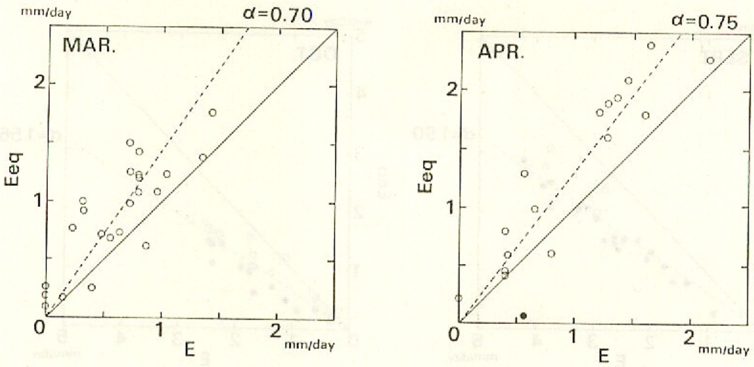

Fig. 5-2 (continued)

sometimes observed during cold spells.

As described above, the proportional constant α in the equilibrium evaporation model, with daily values of available energy and air temperature, shows very complex seasonal variation patterns. In addition, the value of α becomes greater than 1.26 for completely saturated surfaces, especially in winter months.

5-2 Factors affecting the proportional constant in equilibrium evaporation model

Seasonal variations of α have been reported by Jackson et al. (1976), McNaughton et al. (1979), de Bruin and Keijman (1979), and Nakayama and Nakamura (1982) for bare soil, pasture, shallow lake, and radish field, respectively. All of the above authors found an increase in α in cool season, but no one has discussed the reason for the great increase in α except Nakayama and Nakamura (1982). Nakayama and Nakamura (1982) analyzed data from September to December and found that the increase in α was due to the long-wave radiation exchange in the nighttime. However, seasonal variations of α from winter to spring have not been reported yet. The differences in α due to different averaging periods have been pointed out by Yap and Oke (1974), Kanemasu et al. (1976), and Tanner and Jury (1976). However, these authors did not discuss long term variations of α.

To evaluate the reasons for seasonal variations of α, an analysis based on daytime data was conducted. In this analysis, the daytime period was considered as the period during which available energy $(R_n - G)$ was greater than or equal to zero. The nighttime period was considered as the period during which $(R_n - G)$ was negative. The duration of daytime hours for each month is listed in Table 5-1. Figure 5-3 shows day-to-day variations of daytime evapotranspiration (E), equilibrium evaporation (E_{eq}), available energy in evaporation equivalence $(R_n - G^*)$, and the parameter α. Though abrupt changes in α are found on rainy days, day-to-day variations of daytime α are smaller than those of daily α. Values of α do not exceed the value of 1.14, which was obtained from the summer data, except in September where α is approximately 1.14. Furthermore, the march of α shows a distinctive seasonal trend.

The relationship between daytime values of evapotranspiration (E) and equilibrium evaporation (E_{eq}) is shown in Fig. 5-4 for each month, together with monthly mean values of α. The monthly mean value of α decreases gradually in the cold months with a minimum (=0.45) in January. From

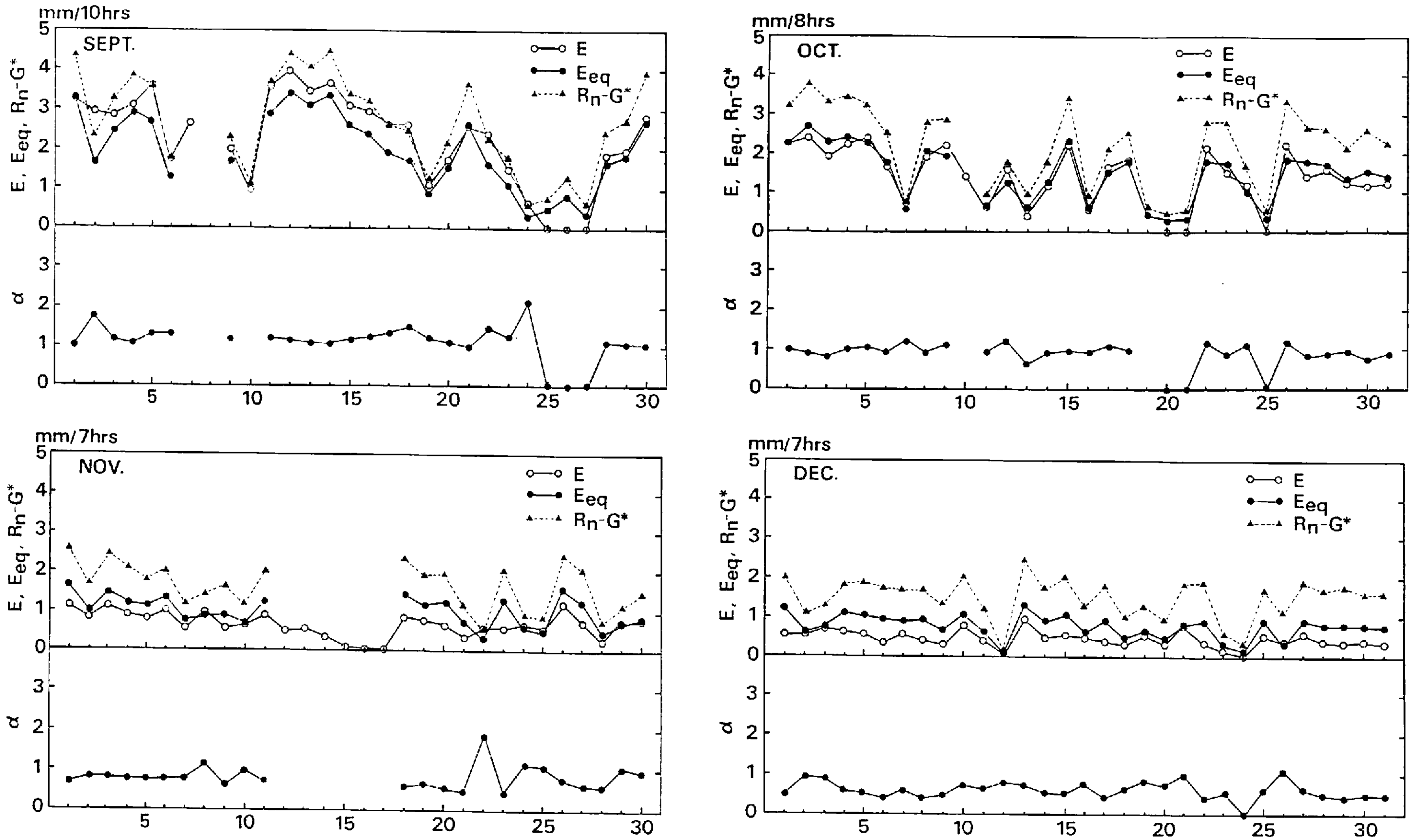

Fig. 5-3 Day-to-day variations of daytime evapotranspiration (E), equilibrium evaporation (E_{eq}), available energy in evaporation equivalence ($R_n - G^*$), and the parameter α in the equilibrium evaporation model.

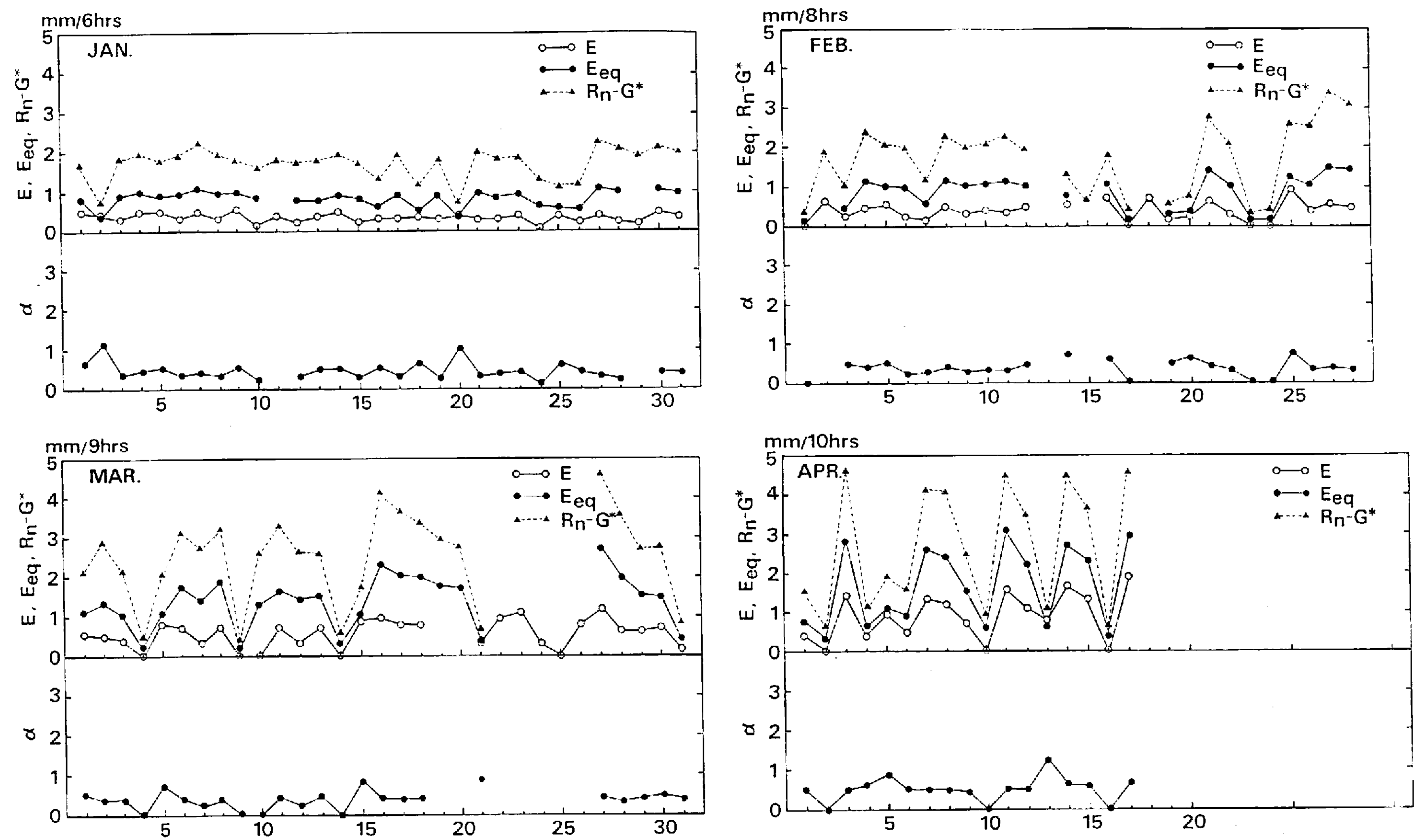

Fig. 5-3 (continued)

Table 5-1 Duration of daytime hours.

	duration of daytime		
	from	to	length
Sept.	0700	1700	10 hrs.
Oct.	0800	1600	8
Nov.	0800	1500	7
Dec.	0800	1500	7
Jan.	0900	1500	6
Feb.	0800	1600	8
Mar.	0800	1700	9
Apr.	0700	1700	10

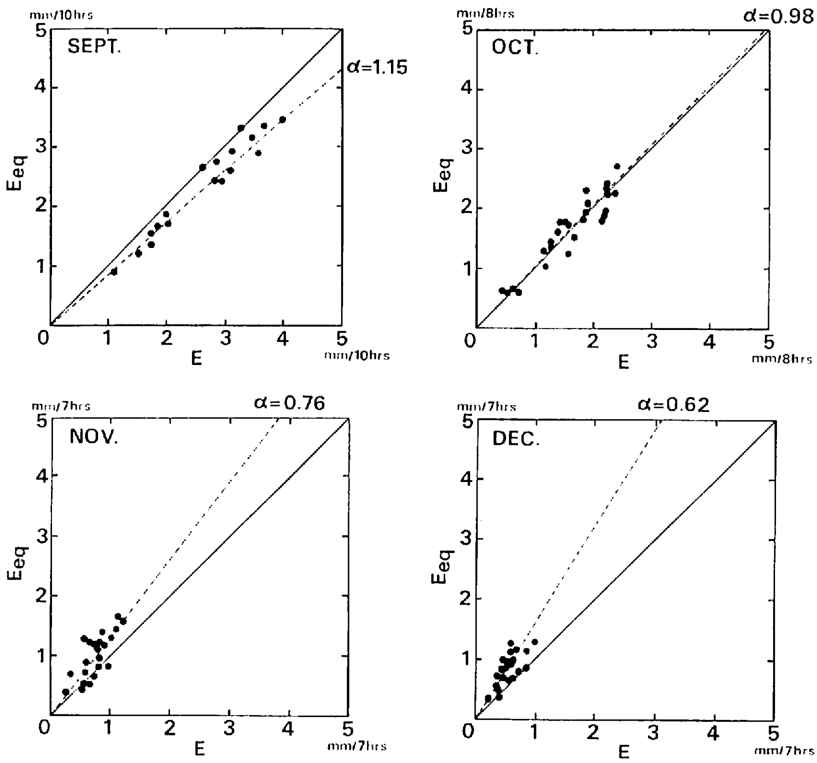

Fig. 5-4 Relationship between daytime evapotranspiration (E) and equilibrium evaporation (E_{eq}). Values of α represent monthly mean values.

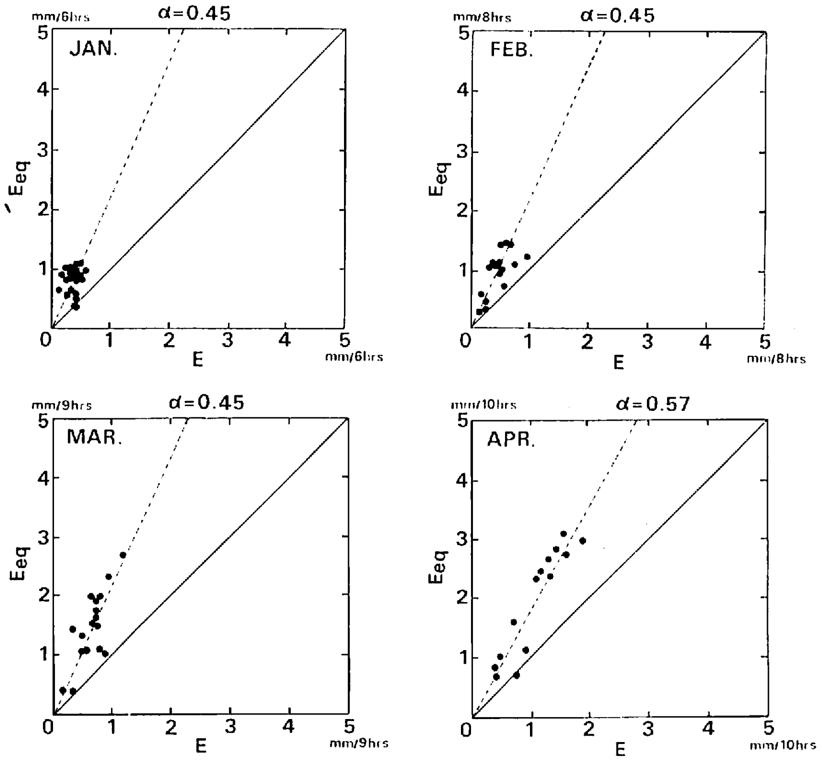

Fig. 5-4 (continued)

January to March, the monthly mean value of α remains constant at 0.45 and then begins to increase in April.

Maeno (1977) explained that temperate grass such as ryegrass gradually decreases in productivity from early autumn to early winter and ceases to grow in winter due to low temperature. It is generally considered that there are close relationships between the physiological and productive activities of plants and water consumption abilities. Therefore, it may be considered that seasonal trends of monthly mean values of α, obtained from daytime data, reflect the physiological and productive features of pasture. That is, the decrease in α from September to January may coincide with the decrease in productivity of pasture from early autumn to early winter, and the constant value of α from January to March may coincide with the stoppage of growth during the winter. The constant value of α from January to March implies that a large portion of evapotranspiration during these months is maintained by soil water evaporation and that the influence of soil water content on evaporation is undetectable. Soil water evaporation without the control of soil water content may result from the fact that the surface soil layer remains wet during the daytime hours due to the melting of needle ice. The formation of needle ice in winter is typical in this area (Ono, 1978), because of the high water content of Kanto loam soil.

Figure 5-5 shows seasonal variations of monthly mean values of daily α ($\overline{\alpha}$) and daytime α (α_d). Monthly mean values of daily $\alpha(\overline{\alpha}_*)$, which are derived from the restricted case $\lambda E < R_n - G$,

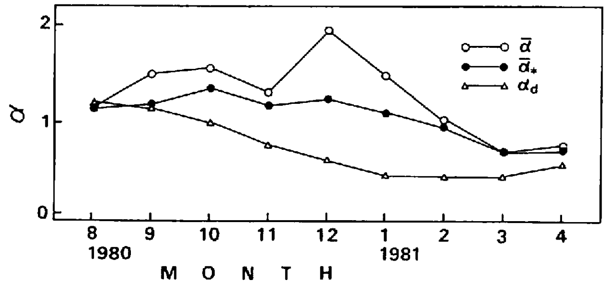

Fig. 5-5 Seasonal variations of monthly mean α.

$\bar{\alpha}$: daily value

$\bar{\alpha}_*$: daily value in the case $\lambda E < R_n - G$

α_d : daytime value

are also plotted in Fig. 5-5. The value of $\bar{\alpha}_*$ is the value of $\bar{\alpha}$ without advective energy supply. Even with this restriction, $\bar{\alpha}_*$ tends to be greater than α_d.

Values of $\bar{\alpha}$ and α_d are almost equal in August but $\bar{\alpha}$ is consistently greater than α_d after September. Month-to-month variations of $\bar{\alpha}$ show a complicated pattern. On the contrary, α_d shows a distinctive month-to-month variation pattern which may reflect the physiological nature of pasture. Although the variations of $\bar{\alpha}$ are more complicated than those of α_d, it can be detected as a general rule that the differences between $\bar{\alpha}$ and α_d show a gradual increase from autumn to winter and a gradual decrease from winter to spring.

Nakayama and Nakamura (1982) stated that the differences between $\bar{\alpha}$ and α_d were caused by the large amount of outgoing radiative energy in the nighttime period. They related the differences to the rate of decreasing daily net radiation due to nighttime outgoing radiation. However, they neglected soil heat flux and used net radiation as a basis for their calculation. Available energy, $R_n - G$, is considered to be preferable to net radiation in considering the energy apportionment at the ground surface. Therefore, available energy was used as a basis in this study.

Figure 5-6 shows examples of hourly variations of net radiation (R_n), soil heat flux (G), and available energy ($R_n - G$). It can be seen in Fig. 5-6 that outgoing radiation due to net long-wave radiation during the nighttime is very large in winter months. As a result, daily available energy becomes much smaller than daytime available energy, which makes daily equilibrium evaporation smaller than daytime equilibrium evaporation. Therefore, the daily value of α increases. If the ratio of the nighttime condensation rate to nighttime available energy was the same as that of the daytime evapotranspiration rate to daytime available energy, an increase in α should not occur. Such a relationship, however, may not be established.

To examine the degree of nighttime available energy, the nighttime decrease rate of available energy (R') is defined by the following equation:

$$R' = -\frac{(\overline{R_n - G}) - (R_n - G)_d}{(R_n - G)_d} \tag{5-1}$$

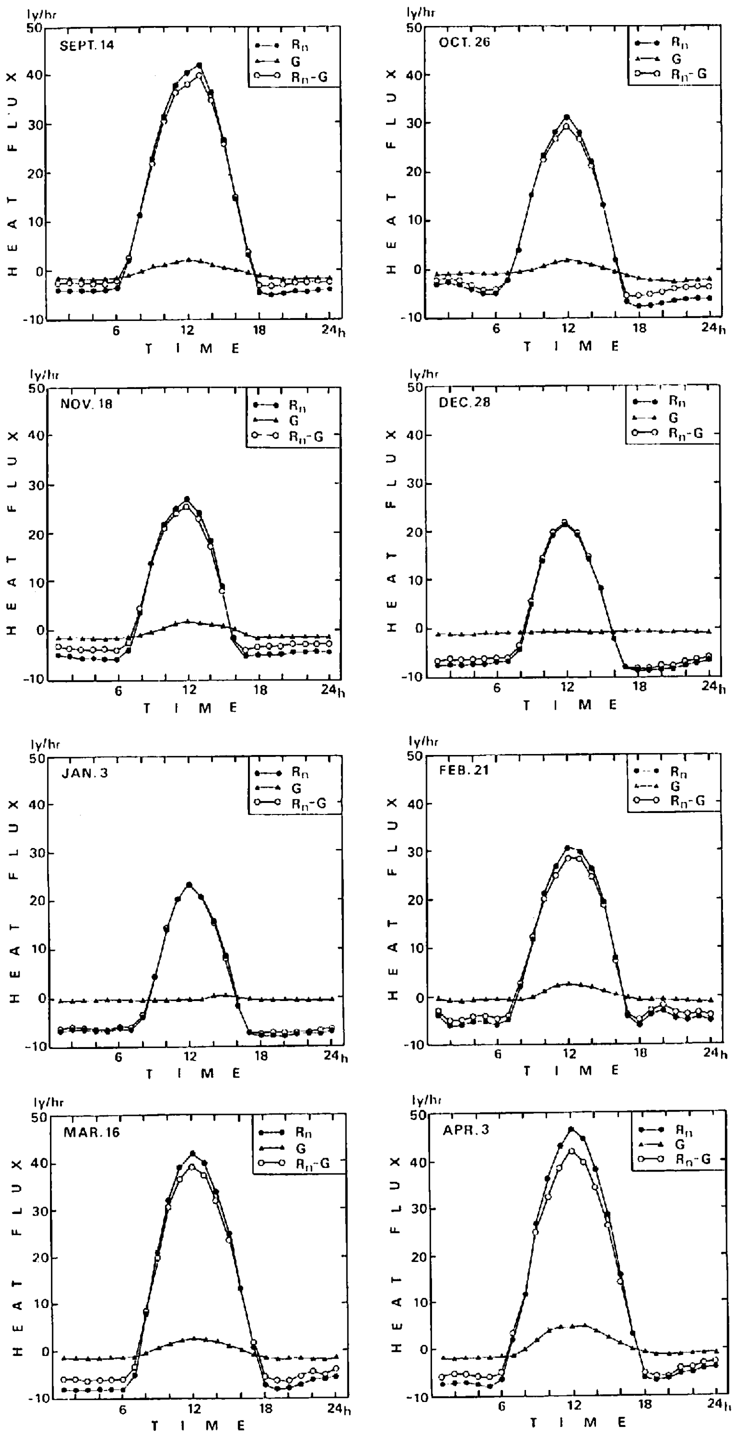

Fig. 5-6 Examples of diurnal variations of net radiation (R_n), soil heat flux (G), and available energy (R_n–G).

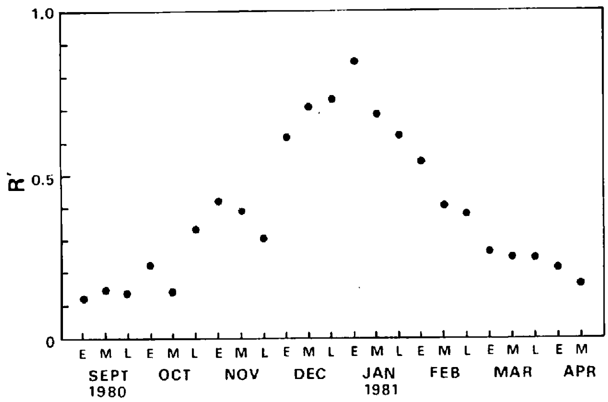

Fig. 5-7 Seasonal variations of 10-day mean values of the rate of decrease in daily available energy (R') due to nighttime radiative cooling.

where the bar and the subscript d represent daily values and daytime values, respectively. The large value of R' represents the large loss of available energy during the night as compared with daytime available energy.

Figure 5-7 shows the seasonal trend of R' as determined by Eq. (5-1). Ten-day averages of $\overline{(R_n-G)}$ and $(R_n-G)_d$ were used for calculation. Although there exist small fluctuations in R', depending on the weather conditions during the averaging periods, it can be detected as a general trend that R' increases up to early January from which it gradually decreases. The maximum value of R' is 0.72, which means that 72% of available energy received during the daytime is lost due to nighttime radiative cooling and that daily available energy is only 28% of daytime available energy.

The variations in the differences between $\overline{\alpha}$ and α_d and those in R' show very similar seasonal patterns. Hence, the differences in α may be considered to be caused by R'. To investigate the relationship between the differences in α and R', the rate of increase in α (α') is defined as

$$\alpha' = \frac{\overline{\alpha} - \alpha_d}{\alpha_d} \tag{5-2}$$

Figure 5-8 shows the relationship between α' and R' calculated from monthly mean values. It is noticeable in Fig. 5-8 that the abrupt increase in $\overline{\alpha}$ in December coincides with the sharp increase in R'. There exists an almost linear relationship between α' and R'. This fact indicates that the differences between $\overline{\alpha}$ and α_d are caused by the nighttime decrease in available energy. Nakayama and Nakamura (1982) obtained the same results from observations over a radish field from September to December (the growing season of radish). They found that the daytime value of α remained nearly constant throughout their observation period. The following becomes clear from the results in this study of pasture: the differences in α due to different averaging periods can be explained by the decrease in the daily available energy induced by nighttime radiative cooling; the above relations are established not only for cool spells, during which the activity of pasture decreases gradually, but also for the periods during winter and from winter to

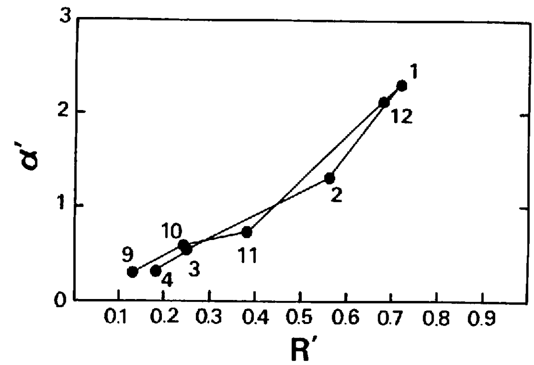

Fig. 5-8 Relationship between the rate of increase in α (α') and the rate of decrease in available energy (R'). Figures represent months.

spring. The condition of $\lambda E > R_n - G$, often observed in the analysis with daily data, may be caused by the decrease in daily available energy due to nighttime outgoing radiation.

As described above, the proportional constant α in the equilibrium evaporation model proves to be an effective parameter showing distinctive seasonal variations which may reflect the activity of pasture. Special attention, however, should be paid to the application of the equilibrium evaporation model to periods with strong nighttime radiative cooling.

5-3 Monthly evapotranspiration estimate with equilibrium evaporation model

The previous discussions have been made based on the observation data from July to April. The values of α for May and June must be included to apply the equilibrium evaporation model to the annual evapotranspiration estimate. The hourly integrated radiation data from the routine measuring system in ERC became available from August, 1981 with the introduction of a real time processor. The missing monthly mean values of α for May and June were calculated from the routinely observed data in 1982. Since data of soil heat flux were not observed in May and June in 1982, soil heat flux was estimated from net radiation by a regression equation. The daily values of net radiation and soil heat flux observed from July, 1980 to April, 1981 were used to obtain the regression equation. Figure 5-9 shows the relationship between net radiation (R_n) and soil heat flux (G). Although there appears a hysteresis in the relationship between R_n and G, the following regression equation was obtained:

$$G = 0.096\,R_n - 13.34 \quad (r = 0.888) \tag{5-3}$$

where G and R_n are in ly/day.

The values of soil heat flux for May and June were estimated from the measured net radiation by using Eq. (5-3). Then the values of α for May and June were calculated from equilibrium evaporation and evapotranspiration by a weighing lysimeter through Eq. (3-15). As a result, $\alpha = 1.15$ for May and $\alpha = 1.12$ for June were obtained. The monthly mean values of daily α are summarized in Table 5-2 together with values obtained in the previous Section. It is noteworthy that α takes nearly the same value from May to August, during which the pasture grew actively and the effects of nighttime radiative cooling on daily available energy value were considered slight.

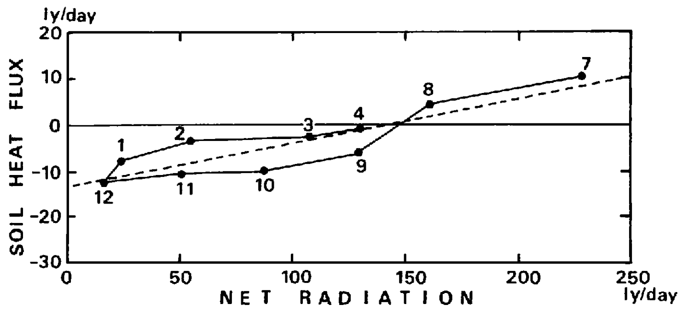

Fig. 5-9 Relationship between monthly mean net radiation and soil heat flux. Figures represent months.

Table 5-2 Monthly mean values of the proportional constant (α) in the equilibrium evaporation model.

	α
Jan.	1.50
Feb.	1.03
Mar.	0.70
Apr.	0.75
May	1.15
June	1.12
July	1.14
Aug.	1.14
Sept.	1.50
Oct.	1.56
Nov.	1.31
Dec.	1.95

The estimation of monthly evapotranspiration by the equilibrium evaporation model with α as listed in Table 5-2 was carried out for the periods from 1980 to 1982. However, the measured net radiation or soil heat flux values were sometimes not available at ERC. Therefore, the unavailable values of net radiation at ERC were estimated by those measured at the Aerological Observatory at Tateno, located 7.2 km southwest of ERC. The regression equation was determined by the monthly average data from July, 1980 to April, 1981. As a result, the following regression equation was obtained:

$$(R_n)_{ERC} = 0.8717\,(R_n)_{AO} + 0.7829 \qquad (r = 0.9837) \tag{5-4}$$

where $(R_n)_{ERC}$ is the net radiation at ERC and $(R_n)_{AO}$ the net radiation at the Aerological Observatory. The relationship between $(R_n)_{ERC}$ and $(R_n)_{AO}$ is shown in Fig. 5-10.

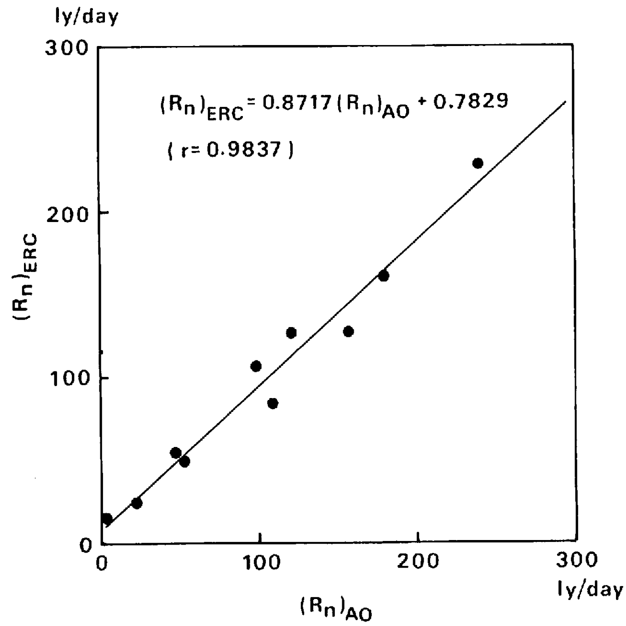

Fig. 5-10 Relationship between net radiation at ERC and that at the Aerological Observatory.

$(R_n)_{ERC}$: net radiation at ERC

$(R_n)_{AO}$: net radiation at the Aerological Observatory

Table 5-3 Monthly estimated evapotranspiration (E_{es}) by the equilibrium evaporation model and actual evapotranspiration (E_{ly}).

	1980		1981		1982	
	E_{es}	E_{ly}	E_{es}	E_{ly}	E_{es}	E_{ly}
Jan.	10.2	14.2	10.7	10.5	11.0	11.0
Feb.	14.3	16.7	12.4	12.5	10.9	12.6
Mar.	16.8	24.6	20.4	20.6	17.4	18.4
Apr.	29.5	37.4	29.2	35.2	27.5	30.0
May	65.0	71.4	64.4	72.7	77.8	77.8
June	78.6	87.2	52.9	55.0	69.7	69.7
July	75.6	83.7	95.5	100.7	68.5	67.6
Aug.	67.1	72.4	89.1	98.9	90.1	82.7
Sept.	73.1	72.4	74.9	64.5	66.0	59.5
Oct.	50.5	50.2	47.9	40.4	46.8	44.7
Nov.	22.4	21.8	16.0	17.1	17.3	21.7
Dec.	13.8	14.1	8.2	10.8	13.3	10.4
Total*	516.9	566.1	521.6	538.9	516.3	506.1

note: * mm/yr (unit: mm/month)

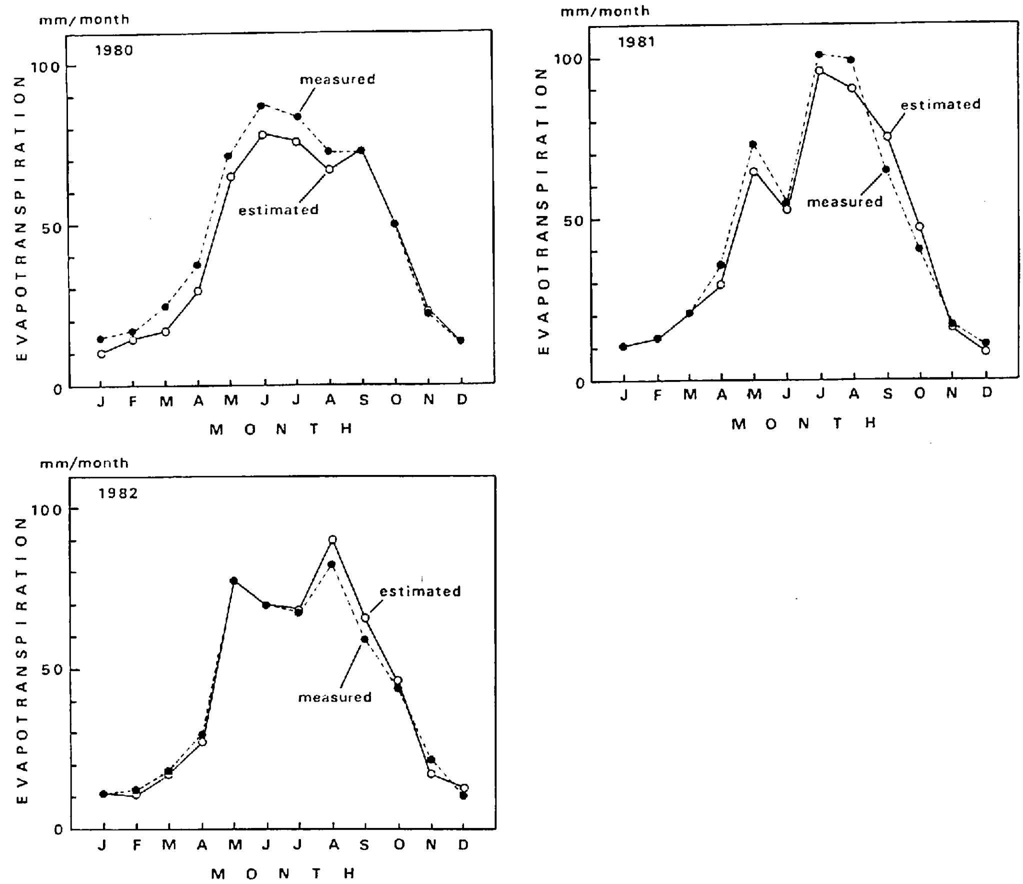

Fig. 5-11 Comparison of estimated monthly evapotranspiration by the equilibrium evaporation model with actual evapotranspiration by a weighing lysimeter.

Figure 5-11 shows month-to-month variations of actual evapotranspiration measured with a weighing lysimeter and estimated evapotranspiration by the equilibrium evaporation model with α as listed in Table 5-2. It is clear from Fig. 5-11 that there exists close agreement between the estimated and the measured monthly evapotranspiration. In addition, as shown in Table 5-3, annual evapotranspiration can be estimated within 10% accuracy by the equilibrium evaporation model. Therefore, if the parameter α is once determined by local calibration, the equilibrium evaporation model may produce a good estimate of evapotranspiration by use of a few input climatological parameters.

5-4 Evaporation formulae and their applicabilities

As described in Section 5-3, the estimation capability of the equilibrium evaporation model is excellent. However, local calibration is necessary to determine the parameter α in the model. In Japan, the Thornthwaite and the Penman methods have been used most frequently as the representative formulae for evapotranspiration in water balance studies. However, tests of these formulae have not been well conducted. Hence, applicabilities of these formulae were investigated by comparing with actual evapotranspiration measured by a weighing lysimeter. The routine data of vapor pressure were not available at ERC. Because of this, monthly averaged meteorological data observed at the Aerological Observatory were used as input data for calculation. Wind speed is measured at a height of 20.5 m at the Aerological Observatory. Since wind speed data observed at a height of 2 m is necessary in the Penman method, the following equation (Takeuchi and Kondo, 1981) was used to convert the wind speed at a height of 20.5 m to that at a height of 2 m:

$$\frac{u_2}{u_1} = \left(\frac{z_2}{z_1}\right)^{1/7} \tag{5-5}$$

where u is the wind speed and z the height at which wind speed is measured and subscripts, 1 and 2, represent the heights of the measurement.

It is necessary in the Penman method to calculate the net radiation received at a hypothetical open water surface exposed to the same weather as the reference area. The equation is

$$R_{no} = S\downarrow - S\uparrow + L\downarrow - L\uparrow = (1 - \alpha_a) S\downarrow + L^* \tag{5-6}$$

where R_{no} is the net radiation received by a hypothetical open water surface, S the short-wave radiation, L the long-wave radiation, and α_a the albedo. Arrows represent the directions of radiant energy. $S\downarrow$ and L^* are estimated from the following equations:

$$S\downarrow = (a + bn/N) R_A \tag{5-7}$$

$$L^* = -\sigma T_a^4 (a' - b'\sqrt{e_a}) (a'' + b'' n/N) \tag{5-8}$$

where n is the number of hours of bright sunshine, N the number of daylight hours, R_A the extra-terrestrial radiation, σ the Stefan-Boltzmann constant, T_a the air temperature in $^\circ$K, e_a the vapor pressure and a, b, a', b', a'', and b'' the constants. The constants in Eqs. (5-7) and (5-8) are different from place to place, which makes local calibrations necessary. Radiation terms ($S\downarrow$, $S\uparrow$, $S\downarrow + L\downarrow$, and R_n) are routinely observed at the Aerological Observatory. From the analyses of data observed at the Aerological Observatory, Kobayashi (1977) obtained the values $a = 0.22$, $b = 0.52$ and Nakagawa (1977) obtained the values $a' = 0.366$, $b' = 0.036$. These values are different from those presented by Penman (1948). However, the test of a'' and b'' in Eq. (5-8) is insufficient. Therefore, Eq. (5-6) with $a'' = 0.1$ and $b'' = 0.9$, which were proposed by Penman (1948), was tested using data obtained from 1980 to 1981 at the Aerological Observatory. Net radiation over a hypothetical open water surface was calculated from the measured radiation terms with $\alpha_a = 0.05$. The results are shown in Fig. 5-12.

As shown in Fig. 5-12, the estimated and the measured incoming short-wave radiations coincide well with each other. In contrast, the estimated net long-wave radiation underestimates the measured one. Table 5-4 shows the measured and the estimated values of $S\downarrow$, L^*, and R_{no}. The relative estimation errors defined by Eq. (5-9) are also listed in Table 5-4.

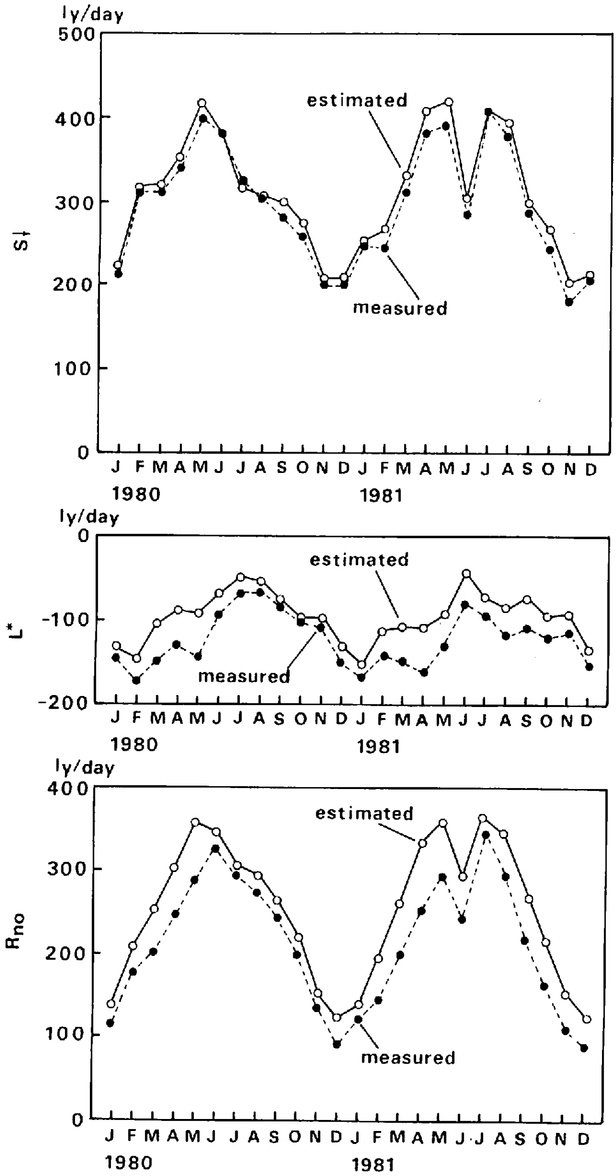

Fig. 5-12 Comparison of measured monthly incoming short-wave radiation ($S\downarrow$), net long-wave radiation (L^*), and net radiation (R_{no}) with measured values.

Table 5-4 Estimated and measured monthly radiation terms and relative estimation error percentages.

		measured			estimated			error		
		$S\downarrow$	L^*	R_{no}	$S\downarrow$	L^*	R_{no}	$\delta S\downarrow$	δL^*	δR_{no}
1980	Jan.	215.4	−147.0	57.6	226.5	−132.4	82.7	5	−10	43
	Feb.	314.6	−174.4	124.5	319.0	−148.8	154.3	1	−15	24
	Mar.	313.8	−148.6	149.5	323.6	−106.0	201.4	3	−29	35
	Apr.	344.2	−131.6	195.4	357.9	−89.8	250.2	4	−32	28
	May	401.1	−144.4	236.6	419.4	−93.5	305.0	5	−35	29
	June	385.4	−93.1	273.0	384.4	−69.7	295.4	0	−25	8
	July	327.1	−67.9	242.8	320.2	−50.6	253.7	−2	−25	4
	Aug.	305.7	−68.7	221.7	311.5	−57.4	238.5	2	−16	8
	Sept.	283.3	−78.8	190.3	302.4	−75.0	212.8	7	−5	12
	Oct.	260.2	−102.9	144.3	277.3	−99.5	164.0	7	−3	14
	Nov.	199.7	−111.1	78.6	207.7	−99.1	98.2	4	−11	25
	Dec.	199.5	−152.0	37.5	209.6	−132.0	67.2	5	−13	79
1981	Jan.	248.8	−168.9	67.5	254.3	−156.2	85.5	2	−8	27
	Feb.	245.1	−142.3	90.5	269.6	−114.6	141.6	10	−19	56
	Mar.	315.0	−151.7	147.6	336.2	−110.5	208.8	7	−27	41
	Apr.	383.6	−163.7	200.7	412.1	−112.6	278.9	7	−31	39
	May	395.3	−133.9	241.6	421.9	−95.2	305.6	7	−29	26
	June	286.7	−83.5	188.9	304.7	−45.6	243.9	6	−45	29
	July	408.6	−97.1	291.1	409.4	−75.7	313.2	0	−22	8
	Aug.	378.7	−119.1	240.7	395.9	−84.0	292.1	5	−29	21
	Sept.	290.0	−110.0	164.5	302.7	−74.2	213.4	4	−33	30
	Oct.	245.6	−122.2	111.1	270.0	−94.5	162.0	10	−23	46
	Nov.	181.5	−117.3	55.1	202.0	−94.5	97.4	11	−19	77
	Dec.	200.7	−154.7	36.0	213.8	−135.1	68.0	7	−13	89

$S\downarrow$: incoming short-wave radiation
R_{no} : net radiation at a hypothetical open water surface
L^* : net long-wave radiation
δ : relative estimation error percentage

$$\delta = \frac{X_{es} - X_{mea}}{X_{mea}} \times 100 \qquad (5\text{-}9)$$

where δ is the relative estimation error percentage, X_{es} the estimated quantity, and X_{mea} the measured quantity. As seen in Table 5-4, the relative estimation error in $S\downarrow$ is within 10% except for one extreme value of 11%. On the other hand, the relative estimation error in L^* is large and reaches into several tens percents, which makes the relative estimation error in R_{no} large. This effect is most significant in winter months with small R_{no}. Therefore, the use of a'' and b'' as suggested by Penman (1948) in Eq. (5-6) makes the estimated R_{no} too large. Since R_{no} is the net radiation received by a hypothetical open water surface, exposed to the same weather as the reference surface, the only different term between these surfaces in Eq. (5-6) is the albedo. In order to prevent the effects of estimation errors on radiation terms, R_{no} was calculated by the measured

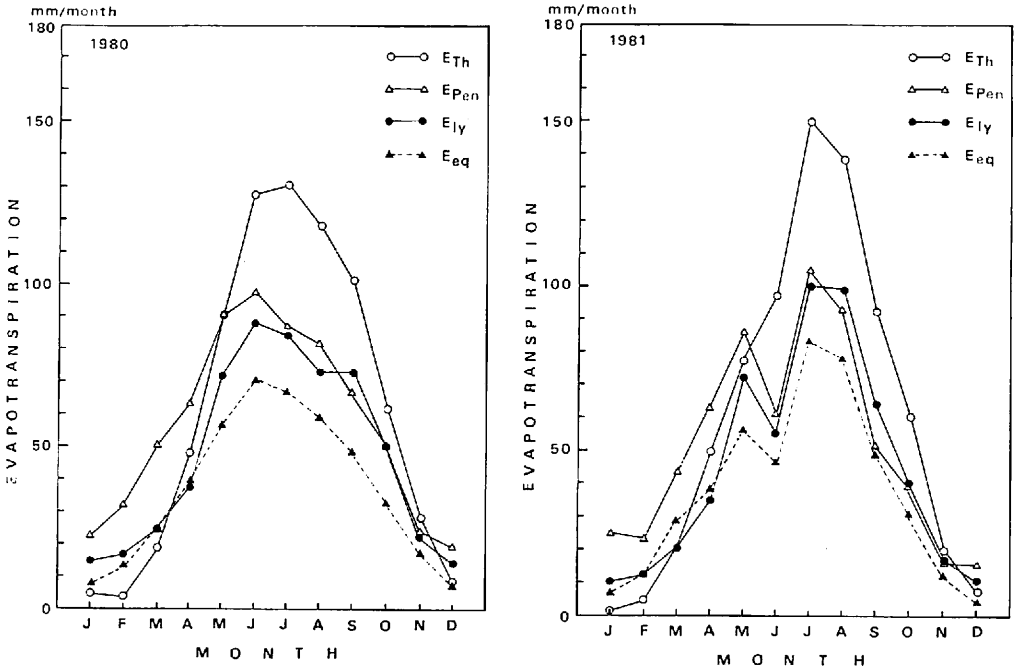

Fig. 5-13 Comparison of estimated monthly evapotranspiration rates by the Thornthwaite method (E_{Th}) and the Penman method (E_{Pen}) with measured evapotranspiration (E_{ly}) and equilibrium evaporation (E_{eq}).

$S\downarrow$, $S\uparrow$, $S\downarrow + L\downarrow$, and R_n at the Aerological Observatory by setting $\alpha_a = 0.05$. The hypothetical open water evaporation (E_o) was calculated by Eqs. (1-1) and (1-3). The potential evapotranspiration was obtained by multiplying the reduction factor (f), listed in Table 1-1.

Figure 5-13 shows the variations of monthly potential evapotranspiration as estimated by the Thornthwaite method (E_{Th}) and the Penman method (E_{Pen}), monthly equilibrium evaporation (E_{eq}), and monthly actual evapotranspiration (E_{ly}) measured by a weighing lysimeter. These monthly values are also listed in Table 5-5.

The potential evapotranspiration determined by the Thornthwaite method (E_{Th}) tends to overestimate summer evapotranspiration and to underestimate winter evapotranspiration. Furthermore, the variation pattern of E_{Th} does not always correspond to that of E_{ly}, which may be considered as a shortcoming of the Thornthwaite method based on air temperature only. The annual totals of E_{Th} are greater than those of E_{ly} by about 30%.

The soil in the observational field is Kanto loam with a high water content, which means that a shortage of soil water for evapotranspiration seldom occurs. Kayane and Kobayashi (1973) stated that the success of the Thornthwaite method may result from the high correlation between air temperature and net radiation. However, transpiration activities of vegetation may differ not only from species to species but also from season to season (Kayane, 1980). The available energy in summer is in the order of 1.33 times as large as equilibrium evaporation. However, the potential

Table 5-5 Comparison of estimated evapotranspiration rates by the Thornthwaite method (E_{Th}) and the Penman method (E_{Pen}) with measured evapotranspiration (E_{ly}) and equilibrium evaporation (E_{eq}).

		E_{ly}	E_{eq}	E_{Th}	E_{Pen}
1980	Jan.	14.2	6.8	4.4	23.0
	Feb.	16.7	13.9	3.2	32.5
	Mar.	24.6	24.0	18.2	50.2
	Apr.	37.4	39.3	47.9	63.4
	May	71.4	56.5	90.2	90.1
	June	87.2	70.2	127.5	97.0
	July	83.7	66.3	130.4	87.1
	Aug.	72.4	58.9	118.1	81.8
	Sept.	72.4	48.7	101.5	66.6
	Oct.	50.2	32.4	61.4	50.4
	Nov.	21.8	17.1	28.0	23.4
	Dec.	14.1	7.1	8.7	19.1
	Total*	566.1	441.2	739.5	684.6
1981	Jan.	10.5	7.1	1.3	25.1
	Feb.	12.5	12.0	4.5	23.5
	Mar.	20.6	29.1	20.8	44.5
	Apr.	35.2	38.9	49.7	63.4
	May	72.7	56.0	77.8	86.6
	June	55.0	47.2	97.1	60.7
	July	100.7	83.8	149.7	105.0
	Aug.	98.9	78.2	138.4	93.0
	Sept.	64.5	49.9	92.1	51.9
	Oct.	40.4	30.7	60.7	39.5
	Nov.	17.1	12.2	19.9	16.9
	Dec.	10.8	4.2	8.0	16.3
	Total*	538.9	449.3	720.0	626.4

note: * mm/yr (unit: mm/month)

evapotranspiration as determined by the Thornthwaite method reaches twice as large as equilibrium evaporation in summer (see Table 5-5). This fact indicates that the potential evapotranspiration as determined by the Thornthwaite method is larger than the available energy. The summer in Japan, especially in the region near the Pacific Ocean, is very hot and very humid due to the advection of humid air from the Pacific Ocean. Table 5-6 shows monthly mean meteorological data observed at the Aerological Observatory. It is considered that high humidity prevents the summer evapotranspiration of pasture from being as large as that which may be expected from air temperature. The empirical formula of Thornthwaite was derived from the data in the U.S.A., which is located in the middle latitudes. Kayane and Kobayashi (1973) stated that the Thornthwaite formula, derived in the middle latitudes, cannot be applied successfully to evapotranspi-

Table 5-6 Monthly mean meteorological data at the Aerological Observatory.

		T (°C)	u (m/s)	RH (%)	n/N
1980	Jan.	2.6	2.0	63	0.63
	Feb.	2.1	1.9	55	0.71
	Mar.	6.3	2.4	63	0.47
	Apr.	11.9	2.4	69	0.39
	May	17.4	2.1	74	0.43
	June	22.1	1.8	82	0.32
	July	22.2	1.8	86	0.21
	Aug.	21.7	1.8	84	0.25
	Sept.	21.2	1.9	78	0.34
	Oct.	15.5	1.7	76	0.46
	Nov.	9.7	1.4	77	0.45
	Dec.	4.3	1.9	69	0.63
1981	Jan.	1.0	1.9	57	0.76
	Feb.	2.5	1.9	66	0.53
	Mar.	6.6	2.0	70	0.51
	Apr.	11.9	2.0	67	0.51
	May	15.4	1.9	74	0.43
	June	18.0	1.6	88	0.17
	July	24.4	1.7	86	0.39
	Aug.	24.2	1.9	83	0.43
	Sept.	19.6	1.5	84	0.34
	Oct.	15.1	1.5	79	0.44
	Nov.	7.3	1.2	78	0.43
	Dec.	3.8	1.4	70	0.65

T : air temperature u : wind speed
RH : relative humidity n/N : fraction of sunshine hours

ration estimates in tropical regions, where cloud amount and humidity make a greater influence on evapotranspiration than does air temperature. The results in this study indicate that attention should be paid to the application of the Thornthwaite method even in Japan, which is located in the middle latitudes but has the different climatic conditions.

The potential evapotranspiration calculated with the Penman method (E_{Pen}) shows similar month-to-month variations as E_{ly}. Although the values of E_{Pen} are closely related to those of E_{ly} from summer to early winter, E_{Pen} is greater than E_{ly} from winter to spring. Annual totals of E_{Pen} are about 20% greater than those of E_{ly} because of the overestimation of E_{Pen} from winter to spring. Good correlation of E_{Pen} with E_{ly} results from the nature of the Penman equation which is based on the physical processes of evapotranspiration and contains more input meteorological parameters than does the Thornthwaite equation.

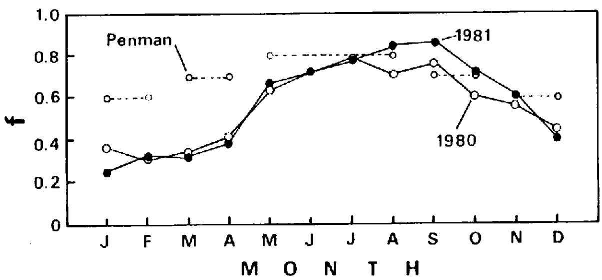

Fig. 5-14 Monthly variations of reduction factor (f) in the Penman method.

Figure 5-14 shows month-to-month variations of the reduction factor (f) in the Penman method, calculated from E_{Pen} and E_{ly}. The values of f proposed by Penman (1948) are also indicated. It is clear from Fig. 5-14 that differences in f are large from January to April and that Penman's original values of f are greater than those obtained at ERC. Evapotranspiration from pasture during winter and early spring is significantly small compared with daytime equilibrium evaporation as discussed in Section 5-2. The wilting of pasture during the above period is considered as the reason for small evapotranspiration. Therefore, it is considered that the large differences in f, recognized in Fig. 5-14, are caused by the fact that pasture withers and then ceases transpiration from winter to early spring. It is interesting that the monthly variations of f obtained at ERC show a very similar pattern to those of α_d obtained from the analysis of evapotranspiration using daytime data (see Fig. 5-5).

As described in Section 5-2, daily net radiation becomes very small in the winter months due to strong radiative cooling in the nighttime. Accordingly, the radiation term in the Penman equation becomes very small in winter. The monthly radiation term and ventilation term in the Penman equation are listed in Table 5-7. It is clear from Table 5-7 that the ventilation term becomes very large compared with the radiation term in winter months. Low humidity and strong wind in winter months are considered as the reason for the large ventilation term in this period. On the contrary, the radiation term is prevalent from late spring to autumn. In this case, E_{Pen} takes approximately the same value as E_{ly}. It is clear from the results in this analysis that the potential evapotranspiration calculated by the Penman method with Penman's original reduction factor cannot be applied to the periods in which pasture withers and strong radiative cooling occurs.

The corrected values of reduction factor for pasture at ERC were determined from Fig. 5-14 and are listed in Table 5-8. If the corrected reduction factor in Table 5-8 is used in place of the original one proposed by Penman, the estimated potential evapotranspiration by the Penman method is 580.7 mm/yr in 1980 and 533.3 mm/yr in 1981, both of which are very close to the measured evapotranspiration.

In this Section, applicabilities of evaporation formulae were discussed, based on evapotranspiration data obtained over pasture, which is a kind of grass surface often treated as a standard surface in the definition of potential evapotranspiration. It is proved that evaporation formulae

Table 5-7 Radiation term and ventilation term in the Penman equation.

		radiation term		ventilation term	
		$\frac{\Delta}{\Delta+\gamma} R_{no}^*$ (mm/month)	$\frac{\Delta}{\Delta+\gamma} \frac{R_{no}^*}{E_o}$	$\frac{\gamma}{\Delta+\gamma} E_a$ (mm/month)	$\frac{\gamma}{\Delta+\gamma} \frac{E_a}{E_o}$
1980	Jan.	13.0	(0.34)	25.4	(0.66)
	Feb.	26.4	(0.49)	27.8	(0.51)
	Mar.	39.1	(0.55)	32.6	(0.45)
	Apr.	57.9	(0.64)	32.7	(0.36)
	May	82.2	(0.73)	30.4	(0.27)
	June	99.9	(0.82)	21.3	(0.18)
	July	91.8	(0.84)	17.1	(0.16)
	Aug.	83.1	(0.81)	19.2	(0.19)
	Sept.	68.7	(0.72)	26.4	(0.28)
	Oct.	48.1	(0.67)	23.9	(0.33)
	Nov.	21.9	(0.56)	17.1	(0.44)
	Dec.	9.3	(0.29)	22.6	(0.71)
1981	Jan.	14.6	(0.35)	27.3	(0.65)
	Feb.	18.8	(0.48)	20.4	(0.52)
	Mar.	39.1	(0.61)	24.5	(0.39)
	Apr.	59.4	(0.66)	31.2	(0.34)
	May	80.3	(0.74)	27.9	(0.26)
	June	63.9	(0.84)	12.0	(0.16)
	July	113.8	(0.87)	17.4	(0.13)
	Aug.	93.9	(0.81)	22.3	(0.19)
	Sept.	57.6	(0.78)	16.5	(0.22)
	Oct.	36.6	(0.65)	19.8	(0.35)
	Nov.	14.4	(0.51)	13.8	(0.49)
	Dec.	8.6	(0.32)	18.6	(0.68)

derived in different climatic regions cannot be applied universally to Japan. The annual growing cycle of vegetation and the climatic conditions caused by air mass play important roles in the determination of evapotranspiration.

Table 5-8 Corrected reduction factor (f) in the Penman method.

	f
Jan.	0.3
Feb.	0.3
Mar.	0.3
Apr.	0.4
May	0.7
June	0.7
July	0.8
Aug.	0.8
Sept.	0.8
Oct.	0.7
Nov.	0.6
Dec.	0.4

CHAPTER 6

CONCLUSIONS

Evapotranspiration is one of the important processes in the hydrological cycle as well as one of the important factors in determining the climatic conditions of an area. The accurate measurement of evapotranspiration is essential not only for water resource management and irrigation planning, but also for the understanding of the mechanism of the hydrological cycle. However, routine observation of evapotranspiration is very difficult due to budgetary and technical limitations. Therefore, several estimation formulae and concepts for evapotranspiration have been proposed. The indiscriminate applications of these concepts and formulae have led to confusion in the interpretation of these concepts and in the use of formulae.

Firstly, potential evapotranspiration, equilibrium evaporation, and potential evaporation were taken as the representative concepts of evapotranspiration. Conditions to which they can be applied were investigated, based on evapotranspiration data obtained over a field of actively growing pasture with no soil water shortage. Pasture is a kind of grass, which is treated as a standard surface in the studies of potential evapotranspiration. The results are summarized as follows:

(1) The recent interpretation of the definitions and the formulae for potential evapotranspiration can be divided into two categories. These interpretations are caused by insufficient treatment of the wetness conditions of an evaporating surface and advective energy.

(2) The potential evapotranspiration, defined by a vapor-saturated surface condition, tends to overestimate the evapotranspiration from actively growing pasture with ample soil water. This type of definition of potential evapotranspiration can be applied only when the leaf surface behaves as a completely wetted surface during dew evaporation, or when the over-passing air is very humid.

(3) The potential evapotranspiration as defined by a vapor-saturated surface condition and that defined by an ample soil water condition cannot be used as the same meaning for pasture, because of the pasture's large aerodynamic resistance.

(4) Hourly evapotranspiration from pasture falls into the range of 1.0 to 1.26 times as large as the equilibrium evaporation, and on the average, the former is 1.16 times as large as the latter. The equilibrium evaporation coincides with the actual evapotranspiration under humid atmospheric conditions.

(5) The potential evaporation, which is defined as 1.26 times as large as the equilibrium evaporation, can be applied only to a pasture canopy, completely wetted by dewfall. Under other conditions, surface resistance acts to regulate the evapotranspiration and then potential evaporation becomes greater than actual evapotranspiration.

(6) The upper and the lower limits of evapotranspiration from actively growing pasture with no soil water shortage can be represented by the potential evaporation and by the equilibrium evaporation, respectively.

A simple equilibrium evaporation model, in which actual evapotranspiration is proportional to equilibrium evaporation, was developed and tested by using daily evapotranspiration data. The following results were obtained:

(7) The daily evapotranspiration from pasture, without a soil water shortage in the summer, can be estimated successfully by the equilibrium evaporation model with a proportional constant equal to 1.14.

(8) A distinctive seasonal trend, which may reflect the activity of pasture, exists in the relationship between daytime evapotranspiration and daytime equilibrium evaporation.

(9) The large amount of outgoing radiative energy due to radiative cooling in the nighttime causes large differences in the proportional constants of the equilibrium evaporation model which is based on the different averaging periods.

(10) The equilibrium evaporation model can estimate annual evapotranspiration within 10% accuracy if appropriate values of the proportional constant, which is determined from local calibration, are used in the model.

The Thornthwaite and the Penman methods for potential evapotranspiration, which are most commonly used in water balance studies in Japan, were tested by actual evapotranspiration measured by a weighing lysimeter. The results are as follows:

(11) The potential evapotranspiration calculated by the Thornthwaite method overestimates summer evapotranspiration and underestimates winter evapotranspiration. The annual evapotranspiration obtained by the Thornthwaite method is greater than the actual evapotranspiration by approximately 30%.

(12) The potential evapotranspiration calculated by the Penman method with the original reduction factor of Penman shows good accordance with the actual evapotranspiration in summer and autumn. However, the former is greater than the latter in winter and spring. The annual evapotranspiration determined by the Penman method exceeds the actual value by approximately 20%. The reasons for this overestimation are that pasture withers from winter to early spring and that a strong radiative cooling occurs in winter.

It is necessary in the application of evaporation formulae to consider climatic conditions and the annual growing cycle of vegetation in the area of interest.

ACKNOWLEDGEMENTS

The author wishes to express his gratitude to his adviser, Dr. I. Kayane, Professor of Institute of Geoscience, the University of Tsukuba, for his continuing guidance and encouragement during the graduate work and research. Special thanks are extended to Professor Dr. S. Takayama, Professor Dr. T. Kawamura, and Associate Professor Dr. K. Kotoda, Institute of Geoscience, the University of Tsukuba, for their suggestions and guidance. Thanks are also expressed to Professor Dr. M. Inokuchi, former Director of Environmental Research Center, the University of Tsukuba, and Professor Dr. M. M. Yoshino, present Director of Environmental Research Center, for making available the facilities for this research.

In addition, the author is grateful to Dr. N. Tase and Mr. Robert C. Backstrom for their helpful suggestions and critical reading of the manuscript. Thanks are due to Professor Dr. S. Shindo, Dr. T. Tanaka, Dr. Y. Suzuki, Institute of Geoscience, the University of Tsukuba, and Dr. Y. Sakura, Department of Earth Science, Chiba University, for their useful discussions.

Acknowledgement is also made to Mr. T. Kojima and Mr. K. Izumi, Environmental Research Center, for their helpful advice in making ventilated psychrometers.

The computations in the present research were performed with use of the ACOS-800 computer at the Science Information Processing Center, the University of Tsukuba, and with the MELCOM 70/25 computer at Environmental Research Center.

BIBLIOGRAPHY

Baier, W. (1966): Studies on dew formation under semi-arid conditions. *Agric. Meteorol.*, **3**, 103-112.

Bailey, W.G. and Davies, J.A. (1981): Bulk stomatal resistance control on evaporation. *Boundary-Layer Meteorol.*, **20**, 401-415.

Barry, P.J. and Robertson, E. (1975): The energy budget of Perch Lake. In: *Hydrological studies on a small basin on the Canadian Shield-evaporation studies*. ed. by Barry, P.J., *AECL Chalk River Nuclear Laboratories*, 375-416. cited in Stewart and Rouse (1977)

Barton, I.J. (1979): A parameterization of the evaporation from nonsaturated surfaces. *J. Appl. Meteorol.*, **18**, 43-47.

Black, T.A. (1979): Evapotranspiration from Douglas fir stands exposed to soil water deficits. *Water Resour. Res.*, **15**, 164-170.

Black, T.A., Tanner, C.B., and Gardner, W.R. (1970): Evapotranspiration from a snap bean crop. *Agron. J.*, **62**, 66-69.

Black, T.A., Thurtell, G.W., and Tanner, C.B. (1968): Hydraulic load-cell lysimeter, construction, calibration, and tests. *Soil Sci. Soc. Am. Proc.*, **32**, 623-629.

Blaney, H.F. (1954): Evapo-transpiration measurements in western United States. *General Assembly of Rome, IASH*, **3**, 150-160.

Brady, R.A., Goltz, S.M., Powers, W.L., and Kanemasu, E.T. (1975): Relation of soil water potential to stomatal resistance of soybean. *Agron. J.*, **67**, 97-99.

Brutsaert, W. (1982): *Evaporation into the atmosphere. Theory, history, and applications.* D. Reidel Publishing Company, Dordrecht, 299p.

Brutsaert, W. and Stricker, H. (1979): An advection-aridity approach to estimate actual regional evapotranspiration. *Water Resour. Res.*, **15**, 443-450.

Budyko, M.I. (1956): *Teplovoi balans zemnoi poverkhnosti (The heat balance of the earth's surface).* Gidrometeol. Izd., Leningrad. (Japanese translation by Uchijima, Z., 1959, 181p., Kasen Suion Chosakai, Tokyo, Japan)

Burrage, S.W. (1972): Dew on wheat. *Agric. Meteorol.*, **10**, 3-12.

Businger, J.A. (1956): Some remarks on Penman's equation for the evapotranspiration. *Neth. J. Agric. Sci.*, **4**, 77-80.

Chang, J.-F. (1965): On the study of evapotranspiration and water balance. *Erdkunde*, **19**, 141-150.

Clarke, R.H., Dyer, A.J., Brook, R.R., Reid, D.G., and Troup, A.J. (1971): The Wangara experiment: Boundary layer data. *Tech. Paper No. 19, Div. Meteorol. Physics, CSIRO*, Melbourne, Australia, 352p.

Davies, J.A. (1972): Actual, potential and equilibrium evaporation for a beanfield in southern Ontario. *Agric. Meteorol.*, **10**, 331-348.

Davies, J.A. and Allen, C.D. (1973): Equilibrium, potential and actual evaporation from cropped surfaces in southern Ontario. *J. Appl. Meteorol.*, **12**, 649-657.

Davies, J.A. and McCaughey, J.H. (1968): Potential evapotranspiration at Simcoe, southern Ontario. *Arch. Meteorol. Geophys. Biokl., Ser. B*, **16**, 391-417.

Deacon, E.L. and Stevenson, J. (1968): Radiation and associated observations made on Indian Ocean cruises: Part 1. Cruises during 1962. *Tech. Paper No. 16, Div. Meteorol. Physics, CSIRO*, Aspendale, Australia, 22p. cited in Priestley and Taylor (1972)

Deardorff, J.W. (1978): Efficient prediction of ground surface temperature and moisture, with inclusion of a layer of vegetation. *J. Geophys. Res.*, **83**, 1889-1903.

De Bruin, H.A.R. (1981): The determination of (reference crop) evapotranspiration from routine weather data. *Proceedings and Informations No. 28, Comm. Hydrol. Research TNO,* Hague, Netherlands, 25-37.

De Bruin, H.A.R. and Keijman, J.Q. (1979): The Priestley-Taylor evaporation model applied to a large, shallow lake in the Netherlands. *J. Appl. Meteorol.*, **18**, 898-903.

Denmead, O.T. and McIlroy, I.C. (1970): Measurements of non-potential evaporation from wheat. *Agric. Meteorol.*, **7**, 285-302.

Denmead, O.T. and Shaw, R.H. (1962): Availability of soil water to plants as affected by soil moisture content and meteorological conditions. *Agron. J.*, **54**, 385-390.

Dyer, A.J. and Hicks, B.B. (1970): Flux-gradient relationships in the constant flux layer. *Quart. J. Roy. Meteorol. Soc.*, **96**, 715-721.

Federer, C.A. (1975): Evapotranspiration. *Rev. Geophys. Space Phys.*, **13**, 442-445.

Federer, C.A. (1979): A soil-plant-atmosphere model for transpiration and availability of soil water. *Water Resour. Res.*, **15**, 555-562.

Ferguson, H.L. and den Hartog, G. (1975): Meteorological studies of evaporation at Perch Lake, Ontario. In: *Hydrological studies on a small basin on the Canadian Shield-evaporation studies.* ed by Barry, P.J., *AECL Chalk River Nuclear Laboratories,* 417-448. cited in Stewart and Rouse (1977)

Goldstein, R.A., Mankin, J.B., and Luxmoore, R.J. (1974): Documentation of PROSPER: A model of atmosphere-soil-plant water flow. *EDFB-IBP 73-9, Oak Ridge National Laboratory,* Tennessee, U.S.A., 75p.

Hansen, V. and Hegg, K. (1981): Potential evapotranspiration from barley and short cut grass in relation to the surface energy budget. *Arch. Meteorol. Geophys. Biokl., Ser. B,* **29**, 57-70.

Hayashi, Y. (1979): On zero-plane displacement and roughness length above a grass field. *Bull. Env. Res. Cen., Univ. of Tsukuba,* Japan. **3**, 25-31. (in Japanese)

Heilman, J.L., Kanemasu, E.T., Bagley, J.O., and Rasmussen, V.P. (1977): Evaluating soil moisture and yield of winter wheat in the Great Plains using Landsat data. *Remote Sens. Env.*, **6**, 315-326.

Hoeber, H. (1970): Measurements in the atmospheric surface layer above the tropical Atlantic. *Weather,* **25**, 419-424.

Holmes, R.M. and Robertson, G.W. (1959): A modulated soil moisture budget. *Monthly Weath. Rev.*, 87, 101-105.

Hungerford, K.E. (1967): An acetate dew gage. *J. Appl. Meteorol.*, **6**, 936-940.

Jackson, R.D., Idso, S.B., and Reginato, R.J. (1976): Calculation of evaporation rates during the transition from energy-limiting to soil-limiting phases using albedo data. *Water Resour. Res.*, **12**, 23-26.

Jones, H.G. (1976): Crop characteristics and the ratio between assimilation and transpiration. *J. Appl. Ecol.*, **13**, 605-622.

Jury, W.A. and Tanner, C.B. (1975): Advection modification of the Priestley and Taylor evapotranspiration formula. *Agron. J.*, **67**, 840-842.

Kaneki, R. and Tomita, M. (1975): Response of cucumber plants to water potential on the root hair surface – Basic studies on water movement in the soil-plant system –. *Trans. JSIDRE,* **57**, 26-32. (in Japanese with English summary)

Kanemasu, E.T., Stone, L.R., and Powers, W.L. (1976): Evapotranspiration model tested for soybean and sorghum. *Agron. J.*, **68**, 569-572.

Kayane, I. (1967): Recent problems in evapotranspiration studies. *Tenki,* **14**, 271-284. (in Japanese)

Kayane, I. (1980): *Hydrology*. Taimeido, Tokyo, 272p. (in Japanese)

Kayane, I. and Kobayashi, M. (1973): Evapotranspiration in Monsoon Asia – especially on the climatological estimation methods –. In: *Water resources in monsoon Asia.* ed. by Yoshino, M.M., Kokonshoin, Tokyo, 55-70. (in Japanese)

Kayane, I. and Takeuchi, A. (1971): On the annual runoff ratio of Japanese rivers. *Geogr. Rev. Japan,* 44, 347-355. (in Japanese with English summary)

Kobayashi, M. (1977): *Hydro-climatological study in monsoon Asia.* D.Sc. Thesis, Tokyo Univ. of Education, Japan, 153p. (in Japanese with English summary)

Kotoda, K. (1979): On wind profile parameters in and above a vegetative canopy. *Bull. Env. Res. Cen., Univ. of Tsukuba,* Japan, **3**, 17-24. (in Japanese)

Kotoda, K. and Hayashi, Y. (1980): On the wind profile above plant canopies and its aerodynamic properties. *J. Agric. Meteorol. Japan,* **35**, 221-228. (in Japanese with English summary)

Lloyd. M.G. (1961): The contribution of dew to the summer water budget of northern Idaho. *Bull. Am. Meteorol. Soc.,* **42**, 572-580.

Maeno, Y. (1977): Forage crop and meteorology. In: *Handbook of agricultural meteorology.* Yokendo, Tokyo, 347-354. (in Japanese)

Maki, T. (1969): On zero-plane displacement and roughness length in the wind velocity profile equation over a corn canopy. *J. Agric. Meteorol. Japan,* **25**, 13-18. (in Japanese with English summary)

Maki, T. (1975): Interrelationships between zero-plane displacement, aerodynamic roughness length and plant canopy height. *J. Agric. Meteorol. Japan,* **31**, 7-15.

Maki, T. (1976): Aerodynamic characteristics of wind within and above a plant canopy – Interrelationship between aerodynamic parameters and plant canopy height –. *Bull. Nat. Inst. Agric. Sci., Ser. A,* **23**, 1-67.

Marlatt, W.E., Havens, A.V., Willits, N.A., and Brill, G.D. (1961): A comparison of computed and measured soil moisture under snap beans. *J. Geophys. Res.,* **66**, 535-541.

McNaughton, K.G. (1976): Evaporation and advection I: Evaporation from extensive homogeneous surfaces. *Quart. J. Roy. Meteorol. Soc.,* **102**, 181-191.

McNaughton, K.G. and Black, T.A. (1973): A study of evapotranspiration from a Douglas fir forest using the energy balance approach. *Water Resour. Res.,* 9, 1579-1590.

McNaughton, K.G., Clothier, B.E., and Kerr, J.P. (1979): Evaporation from land surfaces. In: *Physical hydrology: New Zealand experience.* ed. by Murray, D.L. and Ackroyd, P., New Zealand Hydrol. Soc., Wellington North, 97-119.

Monteith, J.L. (1957): Dew. *Quart. J. Roy. Meteorol. Soc.,* **83**, 322-341.

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

Monteith, J.L. (1965): Evaporation and environment. In: *The state and movement of water in living organisms.* ed. by Fogg, G.E., *Sympos. Soc. Exper. Biol.,* **19**, Academic Press, New York, 205-234.

Monteith, J.L. (1973): *Principles of environmental physics.* Edward Arnold, London, 241p.

Monteith, J.L., Szeicz, G., and Waggoner, P.E. (1965): The measurement and control of stomatal resistance in the field. *J. Appl. Ecol.,* **2**, 345-355.

Mukammal, E.I. and Neumann, H.H. (1977): Application of the Priestley-Taylor evaporation model to assess the influence of soil moisture on the evaporation from a large weighing lysimeter and Class A pan. *Boundary-Layer Meteorol.,* **12**, 243-256.

Murakami, R. (1981): Distinctive feature and cause of the 1980 cool summer weather in Japan. *J. Agric. Meteorol. Japan,* **37**, 249-253. (in Japanese)

Nakagawa, K. (1977): Atmospheric radiation from cloudless sky. *Geogr. Rev. Japan*, **50**, 129-143. (in Japanese with English summary)

Nakayama, K. and Nakamura, A. (1982): Estimating potential evapotranspiration by the Priestley-Taylor model. *J. Agric. Meteorol. Japan*, **37**, 297-302. (in Japanese with English summary)

Nkemdirim, L.C. (1976): Crop development and water loss – A case study over a potato crop–.*Agric. Meteorol.*, **16**, 371-388.

Ono, Y. (1978): Pipkrakes as a periglacial process. *Bull. Env. Res. Cen., Univ. of Tsukuba*, Japan, **2**, 47-55. (in Japanese)

Paulson, C.A. (1967): Profiles of wind speed, temperature and humidity over the sea. *Scientific Rep., NSF Grant GP-2418*, Dept. Atmos. Sci., Univ. of Washington, Seattle, U.S.A., 128p. cited in Priestley and Taylor (1972)

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

Penman, H.L. (1949): The dependence of transpiration on weather and soil conditions. *J. Soil Sci.*, **1**, 74-89.

Penman, H.L. (1950): Evaporation over the British Isles. *Quart. J. Roy. Meteorol. Soc.*, **76**, 372-383.

Penman, H.L. (1956): Estimating evaporation. *Trans. Am. Geophys. Union*, **37**, 43-50.

Penman, H.L. (1963): *Vegetation and hydrology*. Tech. Commun. No. 53, Commonwealth Bureau of Soils, Harpenden, England, 124p.

Penman, H.L. and Schofield, R.K. (1951): Some physical aspects of assimilation and transpiration. *Sympos. Soc. Exper. Biol.*, **5**, 115-129.

Priestley, C.H.B. (1959): *Turbulent transfer in the lower atmosphere*. Univ. of Chicago Press, Chicago, 130p.

Priestley, C.H.B. and Taylor, R.J. (1972): On the assessment of surface heat flux and evaporation using large-scale parameters. *Monthly Weath. Rev.*, **100**, 81-92.

Rijtema, P.E. (1965): *An analysis of actual evapotranspiration*. Centrum voor landbouwpublikaties en landbouw-documentatie, Wageningen, Netherlands, 107p.

Ritchie, J.T. (1973): Influence of soil water status and meteorological conditions on evaporation from a corn canopy. *Agron. J.*, **65**, 893-897.

Ritchie, J.T., Burnett, E., and Henderson, R.C. (1972): Dryland evaporative flux in a subhumid climate: III. Soil water influence. *Agron. J.*, **64**, 168-173.

Rosenberg, N.J. (1974): *Microclimate: The biological environment*. Wiley-Interscience, New York, 315p.

Robinson, S.M. (1962): Computing wind profile parameters. *J. Atmos. Sci.*, **19**, 189-190.

Rouse, W.R. and Stewart, R.B. (1972): A simple model for determining evaporation from high-latitude upland sites. *J. Appl. Meteorol.*, **11**, 1063-1070.

Rutter, A.J. (1975): The hydrological cycle in vegetation. In: *Vegetation and the atmosphere. Vol. 1. Principles.* ed. by Monteith, J.L., Academic Press, London, 111-154.

Sellers, W.D. (1964): Potential evapotranspiration in arid regions. *J. Appl. Meteorol.*, **3**, 93-104.

Shepherd, W. (1972): Some evidence of stomatal restriction of evaporation from well-watered plant canopies. *Water Resour. Res.*, **8**, 1092-1095.

Shimshi, D. and Ephrat, J. (1975): Stomatal behavior of wheat cultivars in relation to their transpiration, photosynthesis, and yield. *Agron. J.*, **67**, 326-331.

Sinclair, T.R., Murphy, C.E., Jr., and Knoerr, K.R. (1976): Development and evaluation of simple models for simulating canopy photosynthesis and transpiration. *J. Appl. Ecol.*, **13**, 813-829.

Slatyer, R.O. (1967): *Plant-water relationships*. Academic Press, London, 366p.

Slatyer, R.O. and McIlroy, I.C. (1961): *Practical micrometeorology.* CSIRO, Melbourne, Australia, 310p. cited in Davies (1972)

Spittlehouse, D.L. and Black, T.A. (1981): A growing season water balance model applied to two Douglas fir stands. *Water Resour. Res.*, **17**, 1651-1656.

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

Stewart, R.B. and Rouse, W.R. (1976): A simple method for determining the evaporation from shallow lakes and ponds. *Water Resour. Res.*, **12**, 623-628.

Stewart, R.B. and Rouse, W.R. (1977): Substantiation of the Priestley and Taylor parameter $\alpha=1.26$ for potential evaporation in high latitudes. *J. Appl. Meteorol.*, **16**, 649-650.

Swift, L.W., Jr., Swank, W.T., Mankin, J.B., Luxmoore, R.J., and Goldstein, R.A. (1975): Simulation of evapotranspiration and drainage from mature and clear-cut deciduous forests and young pine plantation. *Water Resour. Res.*, **11**, 667-673.

Szeicz, G., Endrödi, G., and Tajchman, S. (1969): Aerodynamic and surface factors in evaporation. *Water Resour. Res.*, **5**, 380-394.

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

Szeicz, G., van Bavel, C.H.M., and Takami, S. (1973): Stomatal factor in the water use and dry matter production by sorghum. *Agric. Meteorol.*, **12**, 361-389.

Takeuchi, K. and Kondo, J. (1981): *Atmosphere near the ground.* Tokyo Univ. Press, Tokyo, 226p. (in Japanese)

Tan, C.S. and Black, T.A. (1976): Factors affecting the canopy resistance of a Douglas-fir forest. *Boundary-Layer Meteorol.*, **10**, 475-488.

Tani, N. (1960): The wind on the cultivated field. *J. Agric. Meteorol. Japan*, **16**, 89-93. (in Japanese with English summary)

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

Tanner, C.B. and Jury, W.A. (1976): Estimating evaporation and transpiration from a row crop during incomplete cover. *Agron. J.*, **68**, 239-243.

Tanner, C.B. and Pelton, W.L. (1960): Potential evapotranspiration estimates by the approximate energy balance method of Penman. *J. Geophys. Res.*, **65**, 3391-3413.

Thom, A.S. (1972): Momentum, mass and heat exchange of vegetation. *Quart. J. Roy. Meteorol. Soc.*, **98**, 124-134.

Thom, A.S. (1975): Momentum, mass and heat exchange. In: *Vegetation and the atmosphere. Vol. 1. Principles.* ed. by Monteith, J.L., Academic Press, London, 57-109.

Thom, A.S. and Oliver, H.R. (1977): On Penman's equation for estimating regional evaporation. *Quart. J. Roy. Meteorol. Soc.*, **103**, 345-357.

Thornthwaite, C.W. (1948): An approach toward a rational classification of climate. *Geogr. Rev.*, **38**, 55-94.

Thornthwaite, C.W. and Mather, J.R. (1955): The water balance. *Publications in Climatology*, **8**(1), 86p.

Udagawa, T. (1966): Variation of aerodynamical characteristics of a barley field with growth. *J. Agric. Meteorol. Japan*, **22**, 7-14. (in Japanese with English summary)

Van Bavel, C.H.M. (1966): Potential evaporation: The combination concept and its experimental verification. *Water Resour. Res.*, **2**, 455-467.

Van Bavel, C.H.M. (1967): Changes in canopy resistance to water loss from alfalfa induced by soil water depletion. *Agric. Meteorol.*, **4**, 165-176.

Veihmeyer, F.J. (1972): The availability of soil moisture to plants: Results of empirical experiments with fruit trees. *Soil Sci.*, **114**, 268-294.

Veihmeyer, F.J. and Hendrickson, A.H. (1955a): Does transpiration decrease as the soil moisture decreases? *Trans. Am. Geophys. Union*, **36**, 425-448.

Veihmeyer, F.J. and Hendrickson, A.H. (1955b): Rates of evaporation from wet and dry soils and their significance. *Soil Sci.*, 80, 61-67.

Webb, E.K. (1960): An investigation of the evaporation from Lake Eucumbene. *Tech. Paper No. 10, Div. Meteorol. Physics, CSIRO*, Aspendale, Australia, 75p. cited in Priestley and Taylor (1972)

Williams, R.J., Broersma, K., and van Ryswyk, A.L. (1978): Equilibrium and actual evapotranspiration from a very dry vegetated surface. *J. Appl. Meteorol.*, **17**, 1827-1832.

Wilson, R.G. and Rouse, W.R. (1972): Moisture and temperature limits of the equilibrium evapotranspiration model. *J. Appl. Meteorol.*, **11**, 436-442.

Yap, D. and Oke, T.R. (1974): Eddy-correlation measurements of sensible heat fluxes over a grass surface. *Boundary-Layer Meteorol.*, 7, 151-163.

Environmental Research Center Papers

No. 1 (1982) Kenji KAI: Statistical characteristics of turbulence and the budget of turbulent energy in the surface boundary layer. 54p.

No. 2 (1983) Hiroshi IKEDA: Experiments on bedload transport, bed forms, and sedimentary structures using fine gravel in the 4-meter-wide flume. 78p.

No. 3 (1983) Yousay HAYASHI: Aerodynamical properties of an air layer affected by vegetation. 54p.

No. 4 (1984) Shinji NAKAGAWA: Study on evapotranspiration from pasture. 87p.

No. 5 (1984) Fujiko ISEYA: An experimental study of dune development and its effect on sediment suspension. 56p.

発行　　　昭和59年3月8日

編集・発行者　筑波大学水理実験センター

〒305　茨城県新治郡桜村天王台1－1－1

TEL 0298 (53) 2532, 2534

印　刷　株式会社　三　協　社

〒164 東京都中野区中央4－6－16