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EVAPORATION FROM A PINE FOREST*

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ABSTRACT

Although intensive observations of forest evaporation for the past one or two decades revealed many interesting phenomena on forest evaporation, they were often based on the measurements on "typical" days, or clear summer days. In the present study observations of evaporation, interception, and transpiration activities were conducted at a pine forest over three years. Also observed was seasonal variation in amount of foliage, wetness of the canopy and soil water content. Analyses of these data with emphasis on seasonal variation revealed changes in wet- and dry-canopy evaporation rates and their mechanisms. The results are summarized as follows:

(1) The evaporation rate of wet canopy was usually within 0.3 mm/h during a daytime and 0.16–0.18 mm/h during a nighttime through the year. Evaporation rate of dry canopy, on the other hand, changed seasonally with highest rate of 0.1–0.8 mm/h in summer and lowest rate of 0.1–0.3 mm/h in winter. The integration of hourly evaporation rate of dry canopy over a daytime amounted to the smallest value of 0–1.5 mm in January and the biggest value of 5–5.5 mm in August.

(2) Observations with the heat pulse method showed that the origin of evaporation of wet canopy was intercepted rainfall and thus canopy resistance went to zero. Therefore evaporation rate can be determined solely by meteorological factors. Observations showed that both vapor pressure deficit and available energy stayed constant through the year in wet-canopy condition unlike in dry-canopy condition. This situation led to the lack of seasonal variation in wet-canopy evaporation rate.

(3) When canopy is dry, transpiration is controlled through canopy resistance. Canopy resistance stayed almost constant before early afternoon and then rapidly increased in the afternoon. This pattern of the diurnal variation was consistent through the year; magnitude of the resistance, however, differed seasonally. The seasonal variation can be explained with two mechanisms: changes in leaf area index and stomatal aperture. Leaf area index decreased with leaf fall from the maximum value of 4.0 in July to the minimum value of 1.68 in the middle of July and then increased with foliation. Stomatal aperture is usually caused by the two factors: soil water deficit and changes in meteorological elements. However, soil water deficit proved not to cause stomatal closure in the study area. As for the meteorological elements, vapor pressure deficit had the highest correlation coefficient for $R_n > 80 \text{ W/m}^2$. Regression analysis between stomatal resistance and vapor pressure deficit for four seasons showed that the minimum stomatal resistance was at 100–150 s/m and that slopes of the regression equation changed seasonally. The cause of the variations was suggested as changes in leaf age and air temperature.

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LIST OF SYMBOLS

Symbol	Definition	Unit
$C. I.$	clearness index	
C_p	specific heat of air at constant pressure	$J\ Kg^{-1}K^{-1}$
DBH	diameter breast high	cm
d	diameter of root	mm
E	evaporation	mm
E_{eq}	equilibrium evaporation	mm
E_f	forest floor evaporation	mm
E_i	evaporation of intercepted rainfall	mm
E_t	transpiration	mm
ET	total evaporation, i.e. $E_i + E_t + E_f$	mm
e	vapor pressure	hPa
e_C	vapor pressure at Tower C	hPa
e_E	vapor pressure at Tower E	hPa
e^*	saturation vapor pressure	hPa
G	soil heat flux	Wm^{-2}
H	sensible heat flux	Wm^{-2}
HPV	heat pulse velocity	$cm\ h^{-1}$
I	interception	mm
K_h	turbulent transfer coefficient for heat	m^2s^{-1}
K_m	turbulent transfer coefficient for momentum	m^2s^{-1}
K_v	turbulent transfer coefficient for water vapor	m^2s^{-1}
LAI	leaf area index	m^2m^{-2}
ℓE	latent heat flux	Wm^{-2}
ℓ	latent heat for vaporization	$J\ Kg^{-1}$
P_g	gross rainfall	mm
P_{gp}	precipitation measured by pot-type rain gauge	mm
P_{gt}	precipitation measured by tipping bucket rain gauge	mm
P_s	stemflow	mm
P_t	throughfall	mm
R_n	net radiation	Wm^{-2}
R_w	fractional extracted water in the root zone	
R_{wc}	critical value of R_w	
r_a	aerodynamic resistance	sm^{-1}
r_c	canopy resistance	sm^{-1}
r_s	stomatal resistance	sm^{-1}
T_a	air temperature	$^{\circ}C$
T_{aC}	air temperature at Tower C	$^{\circ}C$
T_{aE}	air temperature at Tower E	$^{\circ}C$
u	wind speed	ms^{-1}
u_C	wind speed at Tower C	ms^{-1}

u_E	wind speed at Tower E	ms^{-1}
u^*	friction velocity	ms^{-1}
u'	fluctuation of horizontal wind speed	ms^{-1}
VPD	vapor pressure deficit	hPa
W_b	weight of leaf biomass per unit area	Kg m^{-2}
W_f	field capacity	mm
W_s	water content in the root zone	mm
W_w	water storage at a wilting point	mm
w	vertical wind speed	ms^{-1}
w'	fluctuation of vertical wind speed	ms^{-1}
β	Bowen ratio	
γ	psychrometric constant	hPa K^{-1}
Δ	slope of saturation vapor pressure curve	hPa K^{-1}
ρ	density of air	Kg m^{-3}

LIST OF ABBREVIATIONS

Abbreviation	Definition
BREB	Bowen ratio/energy balance method
ECEB	eddy correlation/energy balance method
ERC	Environmental Research Center, University of Tsukuba

CHAPTER I

INTRODUCTION

1-1. Review of recent studies

General aspect

Evaporation from a land surface is one of the most important hydrological processes and has been studied from various view points. In water resources research, for example, changes in water yield from a catchment due to deforestation have been a major topic for forest hydrologists. Bosch and Hewlett (1982) demonstrated through the review of watershed experiments that deforestation caused an increase in the amount of discharge due to the decrease in evaporation. In meteorology and climatology, there has been increasing awareness that evaporation from a land surface may have an important role in global climatology. Shukla and Mintz (1982), for example, presented computations which indicate strong dependence of the rainfall and temperature distributions on land surface evaporation.

Evaporation from open water, bare soil or short vegetation has been studied extensively for a long time. On the contrary, difficulties in measurement delayed studies on forest evaporation. Especially studies based on micrometeorological observations have been conducted for only one or two decades. Intensive observations in humid regions, however, revealed many interesting phenomena during this period. Attempts to measure energy and radiation balance in a tropical rain forest also began recently (Shuttleworth *et al.*, 1984a, 1984b).

Methods for measuring forest evaporation

Conventional methods for measuring evaporation from a short vegetation are not necessarily applicable to the measurement of forest evaporation. This is mainly because a forest has aerodynamically rougher surface than a short vegetation. This results in great efficiency of gas and heat transfer from the canopy to air and thus creates small vertical gradients of temperature, water vapor, and other elements to be used in the conventional measurement methods.

Table 1-1 provides the list of methods used for measurements of forest evaporation together with information on their classification by space and time. Among these methods, vast observations of annual evaporation have been reported based on the water balance method (Hibbert, 1967; Bosch and Hewlett, 1982). Application to a small plot (Fedorov, 1965; Calder, 1976) or catchment (Linsley *et al.*, 1958; Hamon, 1961; Takase and Maruyama, 1978; Suzuki, 1980, 1985) for a short time period have also been made. Although there are difficulties in measuring drainage, changes in soil water content (Morton, 1984) and average precipitation (Kayane and Takeuchi, 1971), the water balance method is virtually the only method which determines an areal evaporation. The method, however, is not suitable for analytical studies, as diurnal changes in evaporation cannot be measured.

For measurements of diurnal variations in transpiration or evaporation of one tree or leaves, several methods have been used. They are a tracer method using deuterium oxide or tritiated water (Kline *et al.*, 1970, 1976; Jordan and Kline, 1977; Luvalla and Murphy, 1982; Calder *et al.*, 1986), a chamber method (Greenwood and Beresford, 1979, 1980; Greenwood *et al.*, 1981, 1985), a weighing lysimeter method (Fritschen and Doraiswamy, 1973; Fritschen *et al.*, 1973, 1977; Aston, 1984; Dunin and Aston, 1984), a cut-leaf method (Rutter, 1966; Morikawa, 1970) and a heat pulse method (Cross, 1958; Marshall, 1958; Sessler and Pitman, 1970; Morikawa, 1972; Morikawa and Sato, 1976; Lassolle *et al.*, 1977;

Table 1-1 List of method for evaporation measurement.

method	space	minimum time	object
water balance	catchment lysimeter	weeks-year weeks-month	$E_t + E_i + E_f$ $E_t + E_f, E_i + E_f$
eddy correlation	uniform area	1 hour	—
BREB	—	1 hour	—
ECEB	—	—	—
aerodynamic	—	1 hour	—
tracer	plant	weeks	E_t
tracer (heat pulse)	—	30 min.	E_t
chamber	plant or group of plants	1 hour	$E_t, E_t + E_f$
weighing lysimeter	plant or group of plants	1 hour	$E_t + E_f, E_i + E_f$
cut-leaf	shoots	minutes	E_t
interception	group of plants	1 rainfall event	E_i

BREB: Bowen ratio/energy balance method

ECEB: eddy correlation/energy balance method

E_t : transpiration E_f : forest floor evaporation

E_i : evaporation of intercepted rainfall

Balek and Pavlik, 1977; Tselentis, 1984; Sugita *et al.*, 1983; Yahata, 1984; Sugita and Kotoda, 1984b). These methods are superior to the water balance method in that they can measure transpiration or evaporation at shorter intervals; but in turn they lose the advantage of the water balance method to measure areal evaporation.

As a method to measure evaporation of uniform area with short intervals, an aerodynamic method, an energy balance method and an eddy correlation method have been applied to a forest. The aerodynamic method is based on the assumption of $K_v = K_m$, where K_v and K_m are the turbulent transfer coefficient for water vapor and for momentum. Relation between K_m and K_v is, however, affected strongly by the atmospheric stability: $K_v = K_m$ for neutral condition (Dyer, 1967; Webb, 1970; Oke, 1978); $K_v < K_m$ for stable condition (Kayane, 1980); $K_v > K_m$ for unstable condition (Dyer and Hicks, 1970; Paulson, 1970; Pruitt *et al.*, 1973; Oke, 1978). Therefore, an application to a forest has been limited only to the use for comparison with other methods (Grant, 1975; Thom *et al.*, 1975; McNeil and Shuttleworth, 1975; Milne *et al.*, 1985).

The eddy correlation method has the only provision that vertical wind speed should average out over the mean period in the analysis. A sonic anemometer can satisfy the requirement that an apparatus for measuring wind speed should have fast response and should be free from angle-of-attack response problem (McBean, 1972; Schotanus *et al.*, 1983). The application of this method was difficult because of lack of apparatus to measure fast fluctuation of a specific humidity. Recent technological development, however, enable us to conduct the direct measurement of forest evaporation through this method (Shuttleworth *et al.*, 1984a; Verma *et al.*, 1986) at least in a clear sky condition with a hygrometer using infrared ray (Hyson and Hicks, 1975; Raupach, 1978; Tsukamoto and Mitsuta, 1979; Ohtaki and

Matsui, 1982; Moore, 1983; Ohtaki *et al.*, 1985; Trevitt, 1986), one using ultraviolet ray (Fujitani, 1980), and one using Lyman- α ray (Buck, 1976; Redford *et al.*, 1980; Tsukamoto and Mitsuta, 1982, 1983).

The Bowen ratio/energy balance (BREB) method has been applied to a forest most successfully for the past 20 years. This method is based on the assumption of $K_h = K_v$, where K_h is the turbulent transfer coefficient for heat. Unlike the relation between K_m and K_v , K_h/K_v stays unity for a wide atmospheric stability range: K_h and K_v are nearly equal regardless of the atmospheric stability (Dyer, 1974; Kayane, 1980), although in the stable condition, which usually occurs at night, some researchers suggest $K_h > K_v$ (Campbell, 1973; Verma *et al.*, 1978; Motha *et al.*, 1979) and $K_h < K_v$ (Lang *et al.*, 1983). Therefore except for the nighttime, this method can be used satisfactorily.

Accuracy of this method strongly depends on the accuracy of Bowen ratio and thus data on air temperature and vapor pressure (Angus and Watts, 1984). As vertical gradients of temperature and vapor pressure above a forest canopy are usually small (Jarvis *et al.*, 1976; Lindroth, 1984), the equipment for measuring small difference of dry- and wet-bulb temperatures has been improved: new psychrometer using sensors with high resolution (Black and McNaughton, 1971; Revfeim and Jordan, 1976; Aston, 1985b) and the system that uses a rotating apparatus for periodically reversing two psychrometers to remove systematic errors (Black and McNaughton, 1971; Spittlehouse and Black, 1980, 1981a; McCaughey, 1981, 1985b; Aston, 1985b; Ritari and Strömmer, 1985). The spatial variation of net radiation also affect the result (Droppo and Hamilton, 1973). The Bowen ratio/energy balance method has been applied to many forests for measuring hourly or daily evaporation with great success (Denmead, 1969; Storr *et al.*, 1970; Black and McNaughton, 1971, 1972; Tajchman, 1972; Stewart and Thom, 1973; McNaughton and Black, 1973; Revfeim and Jordan, 1976; Gay and Fritschen, 1979; Thompson, 1979; Munro, 1979; Hattori *et al.*, 1981; Ogushi *et al.*, 1981; Kotoda, 1982; Sugita and Kotoda, 1984a; Sugita, 1985). Long term measurements, however, have been scarcely made because the method has practical difficulties in maintenance like care of a wet-bulb wick.

In addition to the Bowen ratio/energy balance method, eddy correlation/energy balance (ECEB) method has come to be used in measurement of forest evaporation (Moore, 1976a; Hicks *et al.*, 1975; Spittlehouse and Black, 1979; Kotoda, 1982; Kotoda and Sugita, 1984; Sugita, 1985). In this method sensible heat flux is determined with the eddy correlation method, and net radiation (R_n) and soil heat flux (G) are measured directly, then evaporation is calculated as a residual in the energy balance equation. Thus provision in this method is the same as in the eddy correlation method. This method has readiness in maintenance if an on-line operation (Mitsuta *et al.*, 1971, 1973; Kotoda *et al.*, 1983; Lloyd *et al.*, 1984) is employed for calculation of eddy correlation and thus has an advantage in long term measurements.

In the BREB and ECEB methods, storage term and energy used for assimilation have been neglected. Some reports are available on these terms. The flux to storage is usually as small as 1–10% of R_n except at sunrise and sunset hours (Hicks *et al.*, 1975; Tajchman, 1981; McCaughey, 1985a; Sugita, 1985), though Aston (1985a, 1985b) reported that it accounted for more than 10% of R_n in a eucalypt forest. The energy used for plant assimilation was estimated as 1.9–4.7% of R_n at maximum (Hattori, 1985a).

Foregoing review on the methods for measurement of forest evaporation reveals that the eddy correlation/energy balance method is most suitable for a long term observations at present. This method, however, cannot distinguish evaporation of intercepted rainfall from transpiration. Therefore an additional method to measure only transpiration or evaporation of intercepted rainfall should be used for analytical studies.

Factors affecting forest evaporation

Studies on evaporation of temperate-forest have revealed the relation between evaporation and meteorological or physiological elements. The results may be summarized as follows:

1) Available energy

Available energy usually limits the upper side of evaporation rate. The available energy above a forest is larger than that above grassland. Sugita (1984), for example, reported that summertime available energy above a forest was 38% larger than that above a pasture field. Tajchman (1971) observed net radiations of a forest, of an alfalfa field, and of a tomato field. The net radiation of the forest was 20% and 16% larger than that of the alfalfa and that of the tomato fields, respectively. These differences may be attributed to the peculiarity of a forest canopy: smaller albedo (Tajchman, 1971; Oke, 1978; Baumgartner, 1984) and smaller upward long wave radiation, i.e. lower canopy temperature (Tajchman, 1971; Oikawa, 1983; Kawashima, 1986). Moore (1976b) and Van Josef (1984), however, observed lower canopy temperature in a forest in summer and higher in winter. Therefore observation of seasonal variations in radiation balance may be necessary.

2) Canopy wetness

When leaves transpire, water evaporates from cell walls into substomatal cavities, then into the leaf boundary layer through stomatal pores, and finally into the atmosphere. This process can be described as an analogy to electric circuits: evaporating water is current, inverse of transfer coefficient for vapor being the resistance, and vapor pressure deficit being the voltage. Transpiring water receives two resistances: stomatal and aerodynamic resistance. On the other hand, evaporation of intercepted rainfall occurs at leaf surface and thus evaporating water short circuits the stomatal resistance effectively. Accordingly evaporation rate from a wet canopy becomes faster than that from a dry canopy under the same weather condition. This phenomenon has been confirmed both by theoretical models (Murphy and Knoerr, 1975; Lockwood and Sellers, 1982; Wronski, 1984) and by observations (Rutter, 1967; McNaughton and Black, 1973; Moore, 1976a; Stewart, 1977; Singh and Szeicz, 1979; Sugita, 1985; Lindroth, 1985; Hattori, 1985b). Monteith (1965) argued with his "relative transpiration ratio" that evaporation rate from a wet canopy was particularly fast at a forest because of its height and rough surface.

Because available energy for wet-canopy evaporation is usually small as a result of rainy or cloudy sky condition, the additional energy other than the net radiation to maintain this high evaporation rate may be required. Stewart (1977) reported that the inversion of temperature gradient was found in two thirds of 245 meteorological data obtained above a pine forest when the canopy was wet. Therefore he argued that the additional energy came from downward sensible heat flux originating in large scale advection. Pearce *et al.* (1980) obtained wet-canopy evaporation at night using the interception model proposed by Gash (1979). The nighttime evaporation had mean rate of 0.37 mm/h, which was as high as daytime transpiration. They pointed out that the only energy source to cause that high evaporation rate at night was downward sensible heat flux. Moore (1976a), however, presented the observation showing that wet-canopy evaporation was usually limited by net radiation.

The downward sensible heat flux to maintain high evaporation rate has been discussed in terms of an accuracy of observation method (Morton, 1984; Calder, 1985; Morton, 1985). For a thorough understanding of this phenomenon, observations which covers the height of whole atmospheric boundary layer may be necessary.

3) Crown density

A crown density affects the ratio of canopy transpiration to forest floor evaporation. Roberts (1978), who reviewed forest evaporation data in UK, suggested that changes in crown density did not cause total evaporation (ET) variations because forest floor evaporation (E_f) acted as a buffer. At small crown density, total evaporation mainly consisted of forest floor evaporation, while canopy transpiration was dominant in a forest with closed canopy.

The data on the forest floor evaporation, however, are insufficient in number because of difficulties in measurement. Accordingly the amount of E_f or ratio of E_f/ET is far from settled. Some of the recent studies show large variations of these values. Federov (1965) measured forest floor evaporation using evaporimeter as 27–33% of total forest evaporation. Tan *et al.* (1977) reported that up to 70% of total evaporation was contributed by the understory vegetation (salal) below Douglas fir. Roberts *et al.* (1980) presented data which showed that understory bracken evaporation accounted for 60% of total evaporation in Scots pine forest when soil was dry and that even under normal conditions it was still around 25% of the total. Rutter (1966) estimated forest floor evaporation to be 10–20% of total forest evaporation in various forests. Sugita (1984) measured pine forest evaporation and showed that forest floor evaporation accounted for only 6% of total evaporation. Hattori (1983) reported evaporation from forest floor as 8.9% of annual rainfall at a Japanese cypress forest.

In a deciduous forest, both foliation and defoliation cause significant changes in energy and water balance. Kotoda and Sugita (1984) found that the rate of latent heat to net radiation increased from 56% to 82% after foliation in a Japanese oak forest. Rauner (1972) reported that forest floor evaporation of a deciduous tree decreased from 0.16–0.24 mm/h to 0.04–0.07 mm/h following the flush of new leaves. Baldocchi *et al.* (1984) observed difference of radiation penetration into the forest floor between leafless and leafy seasons.

4) Decrease in soil water

When soil water decreased to a certain critical amount, a plant root cannot extract soil water. This causes decrease in turgor pressure in guard cells, closure of stomata, and consequently suppression of transpiration.

The relationship between soil water deficit and transpiration has been studied by many researchers, particularly on crop field (e.g., Brun *et al.* (1985) on spring wheat, Denmead and Shaw (1962) on corn, Mukammal and Neumann (1977) on grass, and Stanhill (1957) on various crops). This relationship in a forest, however, is poorly understood because of difficulties in use of a experimental methods to control soil water content (Sugita and Kotoda, 1985).

Some researchers considered the effects of soil water deficit on forest transpiration in their model (e.g., Spittlehouse and Black, 1981b; Halldin *et al.*, 1984), while Roberts (1983) argued that soil water deficits rarely limit transpiration. This discrepancy may be results of the differences in environment and tree species in the observed area. Therefore in addition to the observation for various tree species, information on seasonal variation in soil water should be collected. Though some data in grassland have been reported (e.g., Ratliff *et al.*, 1983, who compiled soil water data in the U.S.), little is known on seasonal changes in soil water in a forested area.

5) Physiological control

Stomatal aperture controls transpiration rate physiologically. This can be described with analogy to electric circuit as stated previously. The level of stomatal aperture has been expressed as a stomatal resistance (r_s) or a canopy resistance (r_c), which is a resistance of stomata in an imaginary leaf when a

whole canopy is considered as one "big leaf".

The data on these resistances have been accumulated through progress in observation techniques (Stewart and Thom, 1973; Gash and Stewart, 1975; Tan and Black, 1976; Milne, 1979; Leverenz *et al.*, 1982; Shuttleworth *et al.*, 1984a, 1984b). Relations between these resistances and the other meteorological and hydrological elements have also been analyzed (Tan *et al.*, 1977; Attiwill *et al.*, 1982; Bringfelt, 1982; Whitehead *et al.*, 1984; Simpson *et al.*, 1985; Körner, 1985). These field studies suggested that stomatal resistance was a function of vapor pressure deficit together with plant water potential, air temperature, solar radiation, and other elements. The results agree with those by plant physiologists who observed relations between stomatal aperture and the other elements in an environment-controlled laboratory (Lösch and Tenhunen, 1981).

Though stomatal resistance is obviously a function of vapor pressure deficit, the functional form has been controversial, which is attributable, in part, to the fact that reports on seasonal variation in r_s or r_c were scanty. Calder (1977, 1978) proposed a model which expressed annual cyclic changes in stomatal resistance with a cosine curve. Gash and Stewart (1977) obtained an equation in terms of vapor pressure deficit and an annual average of diurnal variations in stomatal resistance. Although these models were useful at places where they were developed, further studies may be necessary to apply them to the other forests.

Lack of information on seasonal variation

The foregoing review has presented some problems concerning measurement and mechanism of forest evaporation. Allowing for these problems, probably the most important topics to be studied now are to measure and to analyze seasonal variation in evaporation and its mechanism.

Although many observational results have been accumulated for forest evaporation, most of the results were unfortunately based on observation on a "typical day", which usually means the day under clear sky condition in summer. This is because observational error is expected small due to the large fluxes on these days. Therefore we have only a little information on seasonal variation in forest evaporation and related factors. There are possibilities, however, that different factors are dominant in determining forest evaporation in different seasons and weather conditions. As a result, analyses of only the typical days may prevent us from proper understanding of forest evaporation. Therefore we should conduct the long term observations including various conditions. Although the Bowen ratio/energy balance method has practical difficulties in maintenance as mentioned previously, we can now use the eddy correlation method over a long time.

1-2. Objectives of the study

In the present study, seasonal variations in the forest evaporation and its mechanism were investigated. Three methods for measurements of evaporation were employed: the energy balance/eddy correlation method for evaporation, the heat pulse method for transpiration, and interception measurement for evaporation of intercepted rainfall. The obtained data were analyzed for the following objectives:

- 1) To make clear seasonal variation in energy balance and evaporation rate of a forest.
- 2) To clarify the mechanism of wet-canopy evaporation.
- 3) To study the mechanism of forest transpiration, with emphasis on physiological control.

CHAPTER II

METHOD

2-1. Study area

The observations were conducted at a 1.7-ha pine forest in University of Tsukuba, Japan ($36^{\circ}05'N$, $140^{\circ}06'E$) on an upland with an altitude of 30 m (Figs. 2-1, 2-2). The forest consists of 10–13-m red pine trees (*Pinus Densiflora*) with an average diameter breast high (DBH) of 12.4 cm and a stand

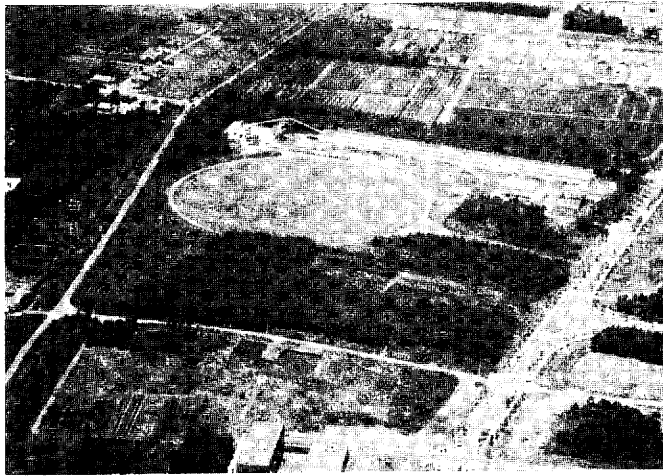


Fig. 2-1 Areal view of study area.

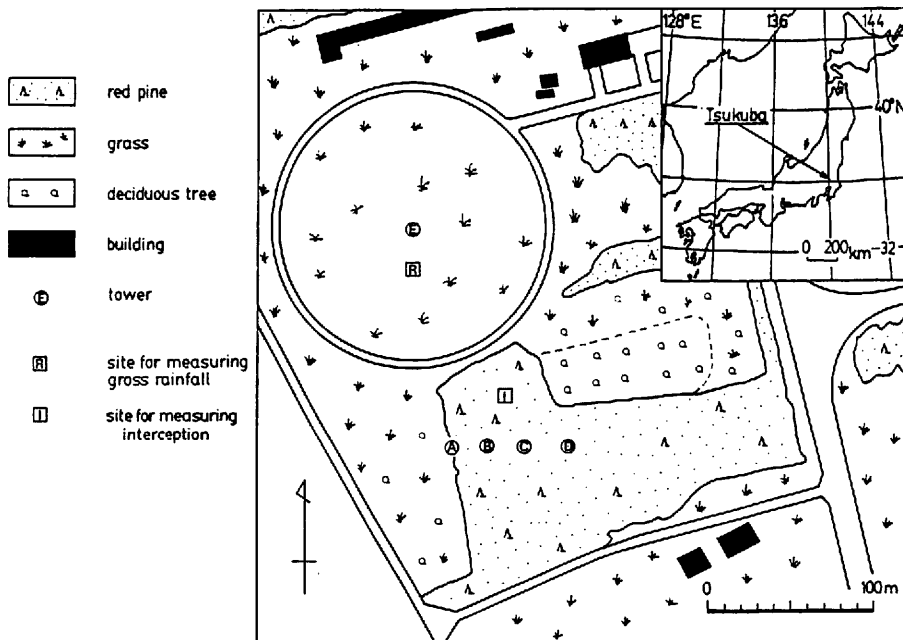


Fig. 2-2 Map of study area.

density of 27 trees per 100 m² (Sugita, 1985). Forest floor is partly covered with an understory vegetation like a lacquer tree, Japanese oak, etc. Figure 2-3 shows the vertical distribution of cross sectional area of a pine tree with DBH of 13.1 cm in the forest. The leaves concentrated at a height of 9–12 m and the leaf area index measured 4.0. Figure 2-4 provides a photograph of a root system of the same tree. The root extended to a depth of 1.4 m. Table 2-1 gives the description of soil profile in the study area. The soil is Kanto loam (volcanic ash soil), 1.7 m in depth, with its top being composed of humus and loam, underlain by clay.

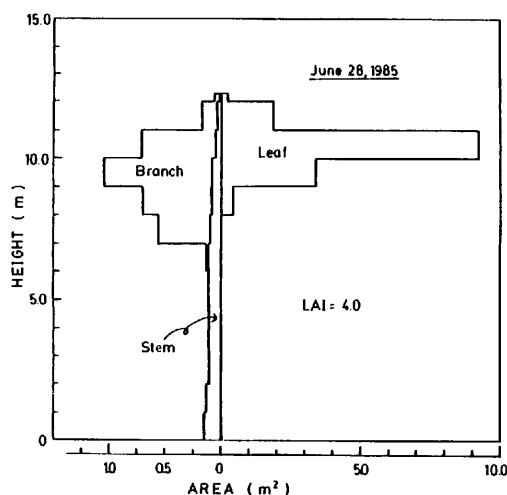


Fig. 2-3 Vertical distribution of cross-sectional area of a pine tree.

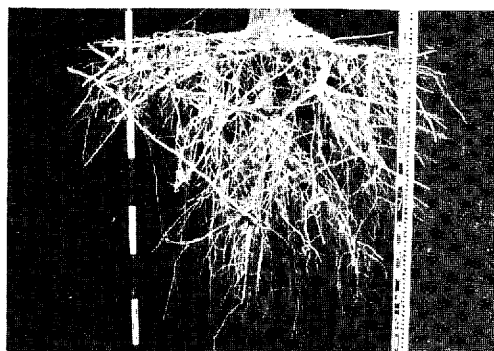


Fig. 2-4 Photograph of root system of a pine tree.

Table 2-1 Description of soil profile.

horizon		description
A ₀	+3–0 cm	litter
A ₁₁	0–7 cm	dark brown (7.5YR/3/2), SiL, fine granular, weak, boundary: plane and clear.
A ₁₂	7–25 cm	dark reddish brown (5YR/3/4), LiC, angular blocky, weak, boundary: plane and clear
B ₁	25–45 cm	dark reddish brown (5YR/3/4), CL, angular blocky, firm, boundary: gradual.
B ₂	45–79 cm	dark reddish brown (5YR/3/4), CL, angular blocky, firm, boundary: gradual
B ₃	79–98 cm	dark reddish brown (5YR/3/4), CL, angular blocky, firm, boundary: plane and clear.

observation: July 7, 1986

2-2. Data collection

Four observation towers with a height of 13.5 m were constructed in the forest, which are the Towers A, B, C, and D in Fig. 2-2. Meteorological and hydrological measurements were made with apparatus on these towers (Fig. 2-5). Some data used in this study have been measured routinely at the

Tower E and Point R (Fig. 2-2) in Environmental Research Center, University of Tsukuba (ERC). All elements measured in the present study are listed in Table 2-2 together with information on observation period and method.

Net radiation was measured with a net radiometer (Eko Instruments Co., Ltd., type CN-11) placed at a height of 13.5 m on the Tower C (Fig. 2-6). Output of the sensor was integrated hourly by an analog integrator (Eko Instruments Co., Ltd., type, MP-20) and recorded with a dotted line recorder.

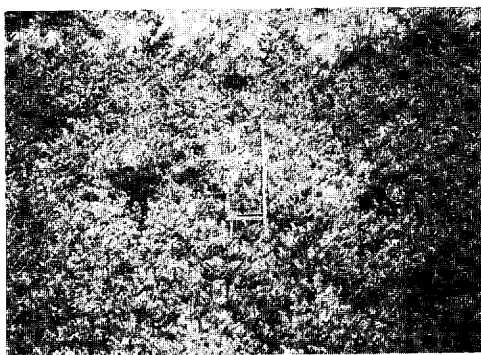


Fig. 2-5 Photograph of Tower C with meteorological instruments.



Fig. 2-6 Photograph of net radiometer.

Table 2-2 Observed elements and observed period.

	item	place	height of sensor	averaging periods or duration of observation	period
Energy Balance	net radiation	C	13.5 m	1 hr.	July, 1983-June, 1986
	soil heat flux	C	- 0.01	1 hr.	-
	sensible heat flux	C	13.5	1 hr.	-
	air temperature	C, E	11.0, 29.5	1 hr.	-
	vapor pressure	C, E	11.0, 29.5	1 hr.	-
	wind speed	C, E	11.0, 29.5	1 hr.	-
	friction velocity	C	13.8	10 min.	July-Sep., 1986
Transpiration	heat pulse velocity	C	6.0	15 min.	July-Aug., 1983
Interception	gross rainfall	R	0	after rainfall event	Apr., 1983-June, 1986
	stemflow	I	1.2	after rainfall event	-
	throughfall	I	0	after rainfall event	-
	wetness of canopy	C	6-12	continuously	July 1983-June, 1986
Subsurface Water	water content	I		1 month	Apr., 1985-June, 1986
	water table	I		1 month	-
Leaf Biomass		I			June, 28, 1985
	amount of litter fall	A-D		1 month	July, 1985-Sep., 1986
	amount of foliation	C		1 month	-

Symbols of place are same as Fig. 2-2

Soil heat flux was measured with a soil heat flux plate (Eko Instruments Co., Ltd., type CN-9) buried at a depth of 1 cm and recorded by the dotted line recorder.

Sensible heat flux was obtained with the eddy correlation method given by eq. (2-1).

$$H = \rho C_p \overline{w'Ta'} \quad (2-1)$$

where H is the sensible heat flux, ρ the density of the air, C_p the specific heat of air at constant pressure, w the vertical wind speed, and Ta the air temperature. The prime denotes a departure from the mean value and the bar a time average. The fluctuations of vertical wind speed and air temperature are measured with a one-dimensional sonic anemometer-thermometer (Kaijo Denki Co., Ltd., type DAT-100). The sensor was mounted at a height of 13.5 m on the Tower C (Fig. 2-7). Output of the sensor was analyzed by eq. (2-1) with an analog flux meter (Kaijo Denki Co., Ltd., type UDF-03) and integrated hourly with the integrator (Eko Instruments Co., Ltd., type MP-20).

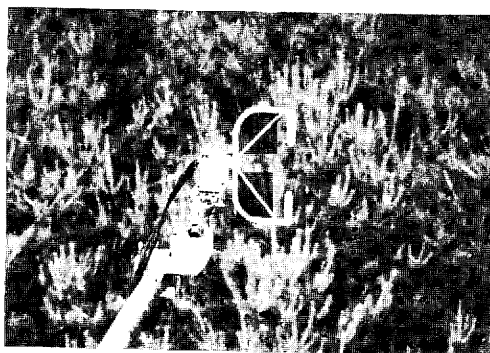


Fig.2-7 Photograph of sonic anemometer-thermometer.

Dry- and wet-bulb temperatures were measured with a ventilated psychrometer employing platinum resistance temperature sensors (Iio Electric Co., Ltd., type SH-20) mounted at a height of 11.0 m on the Tower C. To compensate the missing of data, regression equations for air temperature and vapor pressure between at the Tower C and at the Tower E were calculated (Figs. 2-8 and 2-9). Air and dew-point temperatures at the Tower E have been measured with a ventilated psychrometer using Pt sensor and a dew-point hygrometer, respectively, at a height of 29.5 m (Kotoda *et al.*, 1983). Regression equations were obtained with correlation coefficient of 0.998 for air temperature and for vapor pressure as follows:

$$Ta_C = 0.988 \times Ta_E + 1.044 \quad (2-2)$$

$$e_C = 0.998 \times e_E - 0.310 \quad (2-3)$$

where Ta is the air temperature and e the vapor pressure. Suffixes C and E denote the name of the towers.

Wind speed was measured with a 3-cup anemometer installed at a height of 11.0 m on the Tower C. A correlation equation with the data at the Tower E was also calculated (Fig. 2-10). The wind speed at the Tower E has been measured with a sonic anemometer (Kaijo Denki Co., Ltd., type SA-200) at a height of 30.5 m (Kotoda *et al.*, 1983). The regression equation was obtained with a correlation coefficient of 0.962 as:

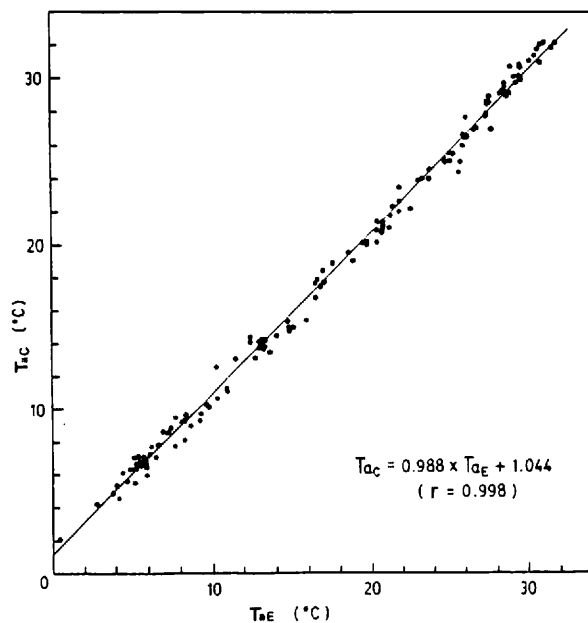


Fig. 2-8 Correlation of air temperature between Tower C (T_{aC}) and Tower E (T_{aE}).

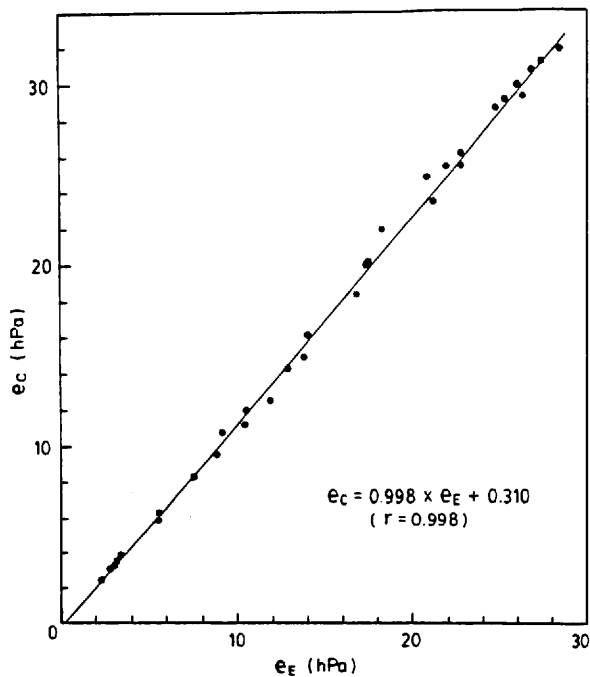


Fig. 2-9 Correlation of vapor pressure between Tower C (e_C) and Tower E (e_E).

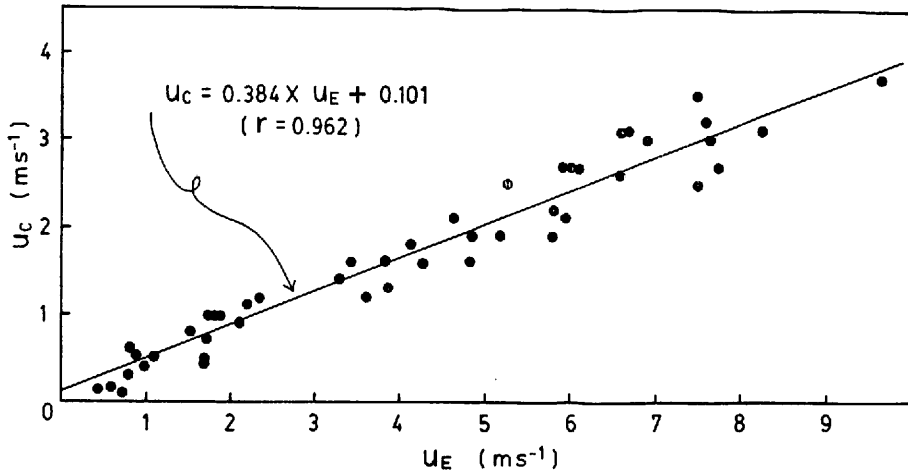


Fig. 2-10 Correlation of wind speed between Tower C (u_C) and Tower E (u_E).

$$u_C = 0.384 \times u_E + 0.101 \quad (2-4)$$

where u is the wind speed and suffixes denote the names of the tower.

Friction velocity (u^*) is given as:

$$u^* = \sqrt{-\overline{u'w'}} \quad (2-5)$$

where u and w are the horizontal and vertical wind speed, respectively. Overbar and prime denote mean and fluctuation therefrom, respectively. The fluctuations of both wind speeds were measured with a three-dimensional sonic anemometer-thermometer (Kaijo Denki Co., Ltd., type DAT-311) placed at a height of 13.8 m on the Tower C (Fig. 2-11) and analyzed by eq. (2-5) with a flux meter (Kaijo Denki Co., Ltd., type UDF-03). Output of the flux meter was recorded with a data logger every 30 minutes.

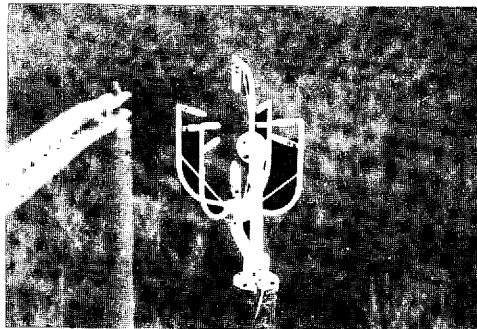


Fig. 2-11 Photograph of sonic anemometer-thermometer for measuring the friction velocity.

The heat pulse method is used to measure the transpiration activity. In this method, a speed of transfer of a heat pulse given to a tree trunk is calculated from measurement of temperature variations at two points: upstream and downstream of the heat source. One pine tree with *DBH* of 11.8 cm was chosen and a set of sensor and heater was inserted into the tree trunk to a depth of 2.0 cm at a height of 5.7 m (Fig. 2-12). Heat pulse velocity (*HPV*) was measured automatically at 15-min. intervals with the apparatus constructed by Sugita *et al.* (1983) and Sugita and Kotoda (1984b). The heat pulse method cannot measure small value of *HPV* so that the threshold value when transpiration begins with sunrise should be determined by other methods. Sugita and Kotoda (1984b) showed from *HPV* and energy balance measurements in sunny days that transpiration came to be active when *HPV* just below a canopy of a pine tree exceeded 1.2 cm/h. Thus this *HPV* value was adopted as the threshold in the present study.



Fig. 2-12 Photograph of the sensor used for heat pulse method.

Gross rainfall has been measured with a tipping bucket rain recorder by ERC at Point R shown in Fig. 2-2 (Kotoda *et al.*, 1983). Throughfall was obtained as an average of 50 pot-type rain gauges (Fig. 2-13) distributed at 2-m intervals in a rectangular plot at Point I (Fig. 2-2). Trap efficiency of this rain gauge was compared by putting the pot-type and the tipping bucket rain gauge of ERC side by side. The



Fig. 2-13 Pot-type rain gauge for measuring throughfall.

comparison of measurements revealed no serious difference between two rainfall data (Fig. 2-14). Accordingly no correction was made to the throughfall data obtained with the pot-type rain gauges. Stemflow was measured with seven stemflow samplers (Fig. 2-15) mounted on stems of pine trees at a height of 1.2 m at Point I. Throughfall and stemflow were measured after each rainfall event.

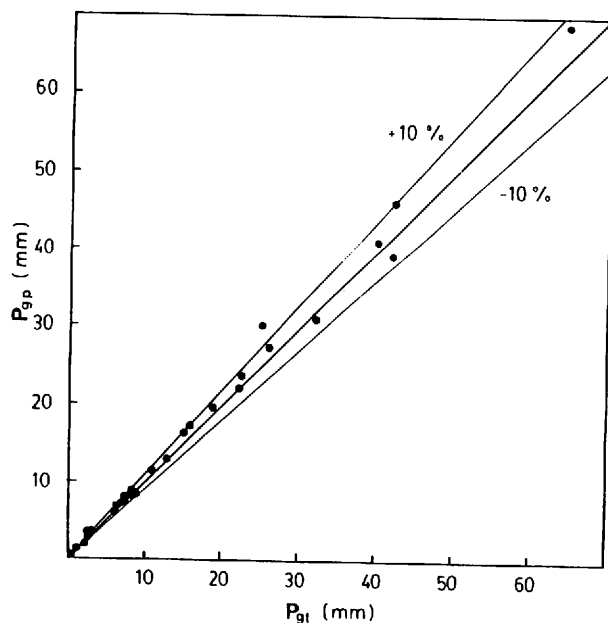


Fig. 2-14 Correlation between rainfall measured by tipping bucket rain gauge (P_{gt}) and pot-type rain gauge (P_{tp}).



Fig. 2-15 Photograph of stemflow sampler.

Wetness of the forest canopy was observed with three wetness indicators (Murata Mfg. Co., Ltd., type HOS003, Fig. 2-16) in the canopy. The sensor has dimensions of $1.5 \times 1.2 \times 0.15$ cm and changes its electric resistance remarkably with ambient relative humidity (Sugita, 1985). As a preliminary test of this sensor, twelve indicators were installed in the canopy and the outputs were compared with visual observation of canopy wetness. The test showed good agreement between them. It was also shown that the canopy may be classified into three parts in terms of speeds of wetting and drying (Sugita and Kotoda, 1984b). Therefore in the present study three wetness indicators were installed in each part of the canopy. The outputs were converted into a voltage and then recorded by a dotted line recorder.

Soil water content was determined by a gravimetric method. Soil samples were taken every month at 10-cm intervals to a depth of 1.4 m. Groundwater level was measured in an observation well with a depth of 2.3 m at Point I (Fig. 2-2).

Amount of leaf was measured by cutting a pine tree with *DBH* of 13.1 cm on June 28, 1985 (Sugita *et al.*, 1986). Amount of leaf fall was measured as an average of litter accumulated in four litter traps (Fig. 2-17). The area of 3 m² is necessary for litter trap to measure amount of litter fall in a forest (Yoda, 1971), so that in the present study four traps each of which has an area of 1 m² were placed on the forest floor. Amount of foliage was obtained by measuring the growth of foliage in the crown of the pine tree that is selected as a representative.

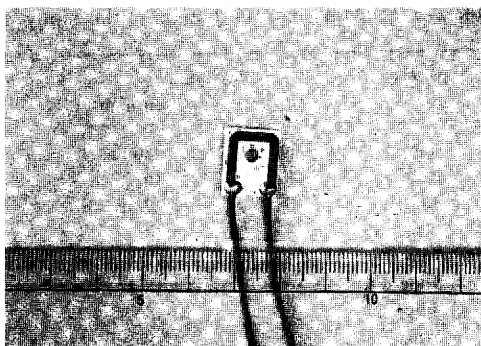


Fig. 2-16 Photograph of wetness indicator.

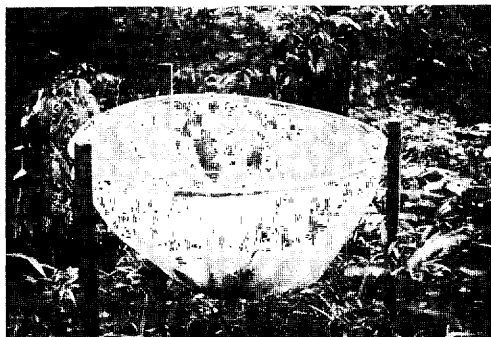


Fig. 2-17 Photograph of litter trap.

2-3. Terminology and calculation of evaporation

The terms concerning evaporation are defined as follows:

Evaporation (E) is the physical process in which liquid water is converted into water vapor. *Transpiration* is the evaporation through plant stomata. *Dry-canopy evaporation* (E_d) is the evaporation from a dry canopy. This is a synonym of transpiration of a canopy. *Wet-canopy evaporation* (E_w) is the evaporation from a wet canopy. This is a synonym of evaporation of intercepted rainfall or interception (I). *Forest floor evaporation* (E_f) is the evaporation from a forest floor or transpiration of understory vegetation. *Total evaporation* (ET) is the sum of dry- and wet-canopy evaporation and forest floor evaporation. *Equilibrium evaporation* (E_{eq}) is defined in eq. (2-6) (Slatyer and McIlroy, 1961) and represent the minimum evaporation of vegetation without water shortage (Nakagawa, 1984).

$$\ell E_{eq} = \Delta (R_n - G) / (\Delta + \gamma) \quad (2-6)$$

where ℓ is the latent heat for vaporization, Δ the slope of saturation vapor pressure curve, R_n the net radiation, G the soil heat flux and γ the psychrometric constant.

As an averaging time, terms daytime and hourly are used to denote daytime mean and hourly mean, respectively. Daytime is determined for each month as shown in Table 2-3.

Table 2-3 Daytime period defined in this study.

Month	daytime	
	from	through
Jan.	0900 h	1500 h
Feb.	8	16
Mar.	7	17
Apr.	7	17
May	6	17
June	6	18
July	6	18
Aug.	6	18
Sep.	7	17
Oct.	7	16
Nov.	8	15
Dec.	9	15

Evaporation from a canopy is obtained with the eddy correlation/energy balance method:

$$\ell E = R_n - H - G \quad (2-7)$$

where H is the sensible heat flux. Although evaporation obtained by eq. (2-7) includes forest floor evaporation, the preliminary observations revealed that it accounted for only several percent of total evaporation. Therefore evaporation calculated by eq. (2-7) is treated as evaporation from a canopy in the present study.

Evaporation of intercepted rainfall (E_i) or amount of interception (I) can be obtained as follows:

$$I = E_i = P_g - P_t - P_s \quad (2-8)$$

where P_g is the gross rainfall, P_s the stemflow, and P_t the throughfall.

CHAPTER III

ENERGY BALANCE AND EVAPORATION RATE

3-1. Energy balance

Figures 3-1 through 3-3 show seasonal variations in energy balance elements in various weather conditions. Weather conditions are classified by canopy wetness and clearness index, which is defined as the ratio of global solar radiation to extra-terrestrial solar radiation (Ikeda and Ichiki, 1977): (1) dry canopy and $C.I. \geq 0.5$ (Fig. 3-1), (2) dry canopy and $C.I. < 0.5$ (Fig. 3-2), and (3) wet canopy and $C.I. < 0.5$ (Fig. 3-3). Figure 3-4 illustrates seasonal variations in overall mean of energy balance components.

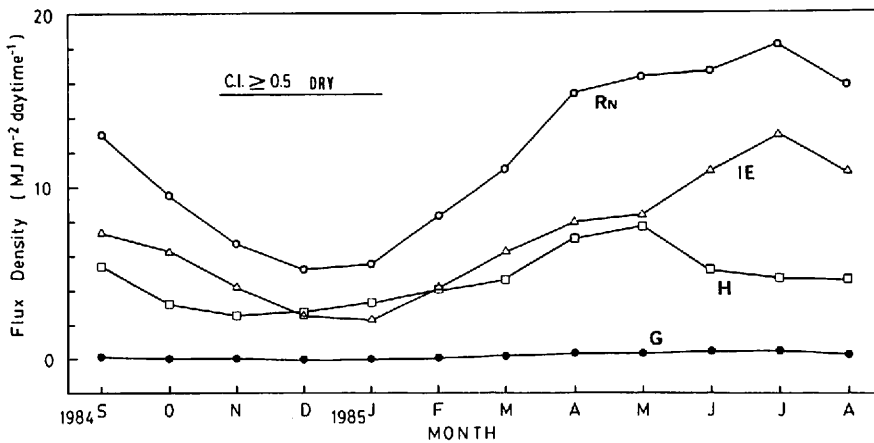


Fig. 3-1 Seasonal variation in energy balance with $C.I. \geq 0.5$ and dry canopy.

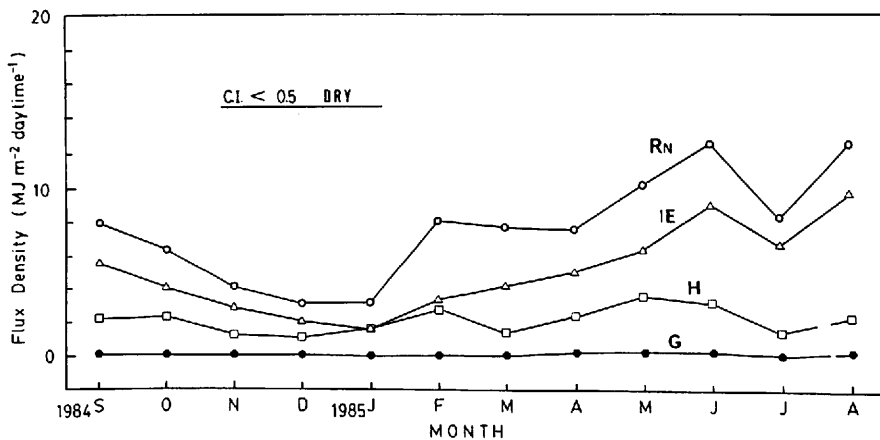


Fig. 3-2 Seasonal variation in energy balance with $C.I. < 0.5$ and dry canopy.

These figures show that the way of division of net radiation into soil, sensible and latent heat flux differed seasonally and with weather conditions. To clarify this, Figs. 3-5, 3-6, and 3-7 illustrate seasonal variations in the ratio of soil heat flux to net radiation (G/R_n), of latent heat flux to net radiation ($\ell E/R_n$) and of sensible to latent heat flux, or Bowen ratio ($H/\ell E$), respectively.

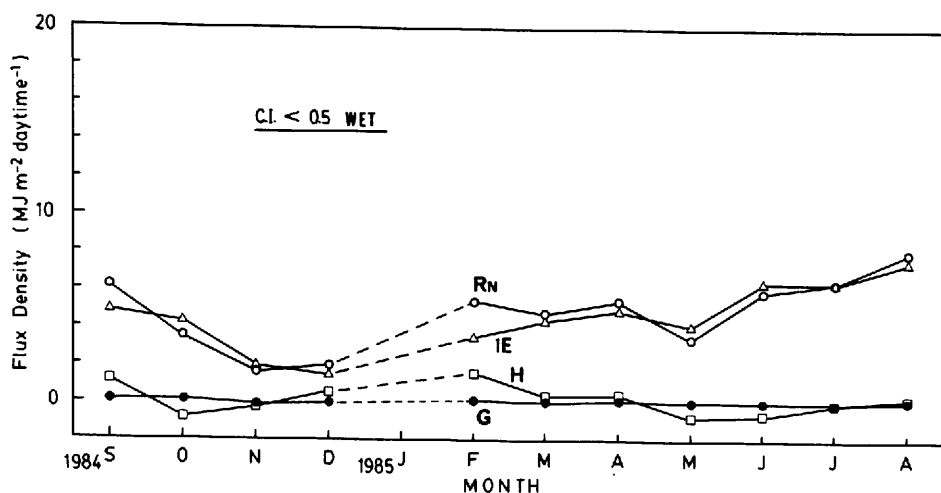


Fig. 3-3 Seasonal variation in energy balance with $C. I. < 0.5$ and wet canopy.

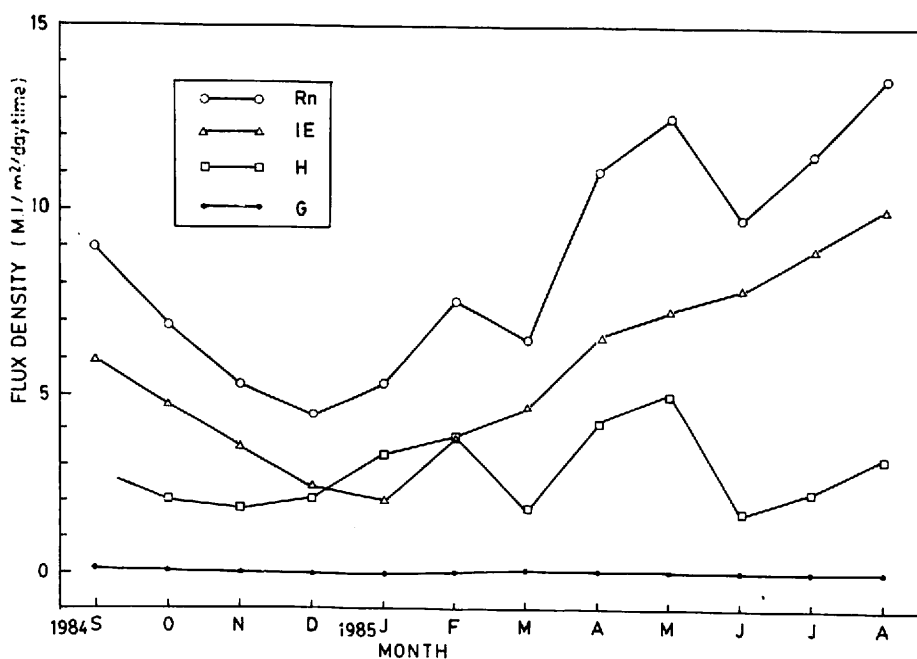


Fig. 3-4 Seasonal variation in energy balance.

Figure 3-5 shows that G was usually several percent of R_n except in wet-canopy condition. This result is same as those observed in various forest (Hicks *et al.*, 1975; Milne, 1979; Kotoda, 1982). Fluctuation of G/R_n ratio in the wet-canopy condition may be caused by measurement failure of a heat flux plate due to infiltrated water.

Figure 3-6 reveals that the ratio of latent heat to net radiation was lowest on sunny day and became as high as more than unity when canopy was wet. The ratio $\ell E/R_n$ also changed seasonally, for example, from 0.42 to 0.74 for $C. I. \geq 0.5$. Bowen ratio shows similar seasonal variation to the ratio $\ell E/R_n$ (Fig. 3-7). Hence factors affecting energy partition are likely to vary from one season to another.

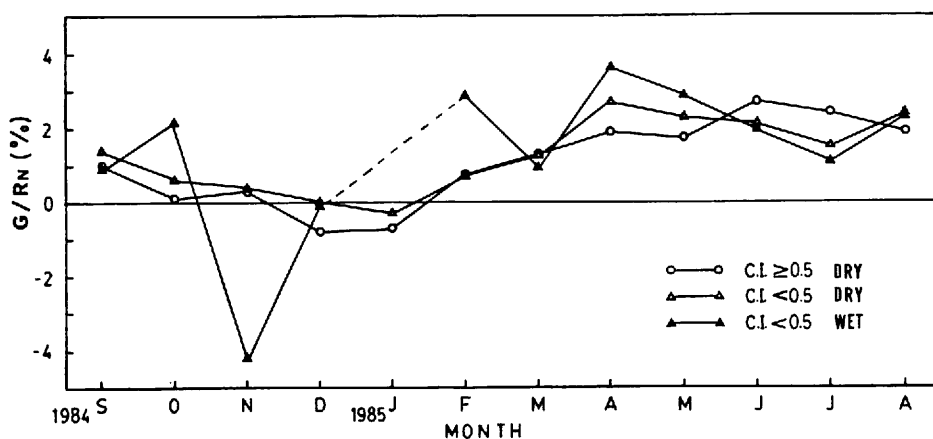


Fig. 3-5 Seasonal variation in the ratio of G/R_n .

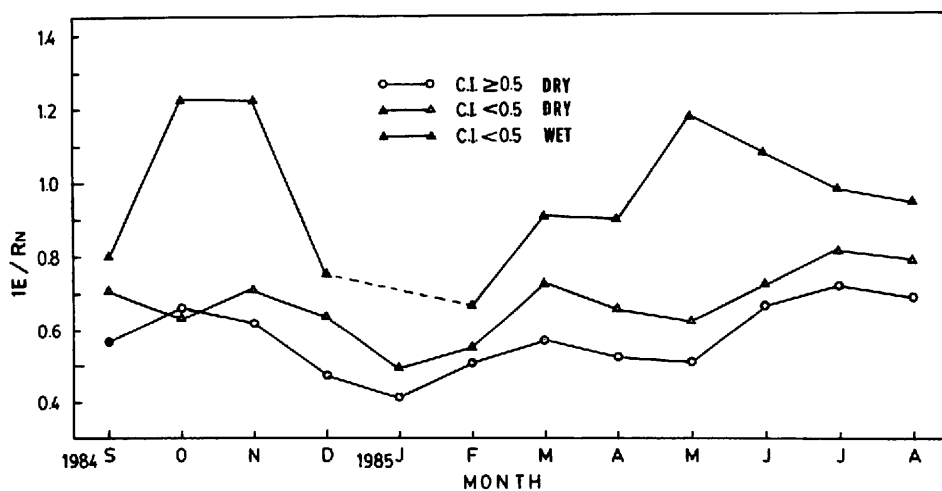


Fig. 3-6 Seasonal variation in the ratio of $\ell E/R_n$.

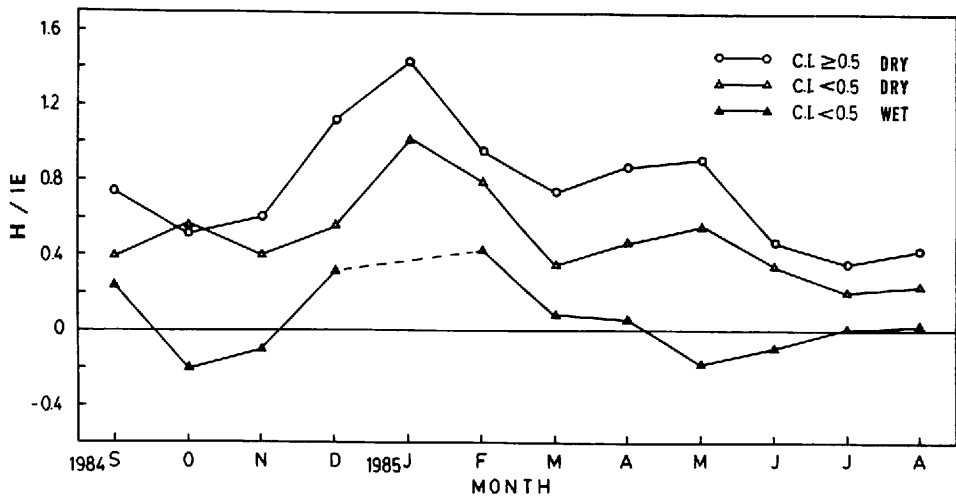


Fig. 3-7 Seasonal variation in the ratio of $H/\ell E$.

3-2. Seasonal variation in dry-canopy evaporation

Figure 3-8 shows seasonal variation in frequency distribution of daytime transpiration. Information on the day numbers for each month used in the analysis is given in Table 3-1. Some transpiration data in January and February were excluded from the analysis. This is because sensible heat flux sometimes exceeded the net radiation during this period and thus resulted in "negative transpiration" in energy balance equation. Possible explanations of this are: (i) inadequate averaging time in the eddy correlation

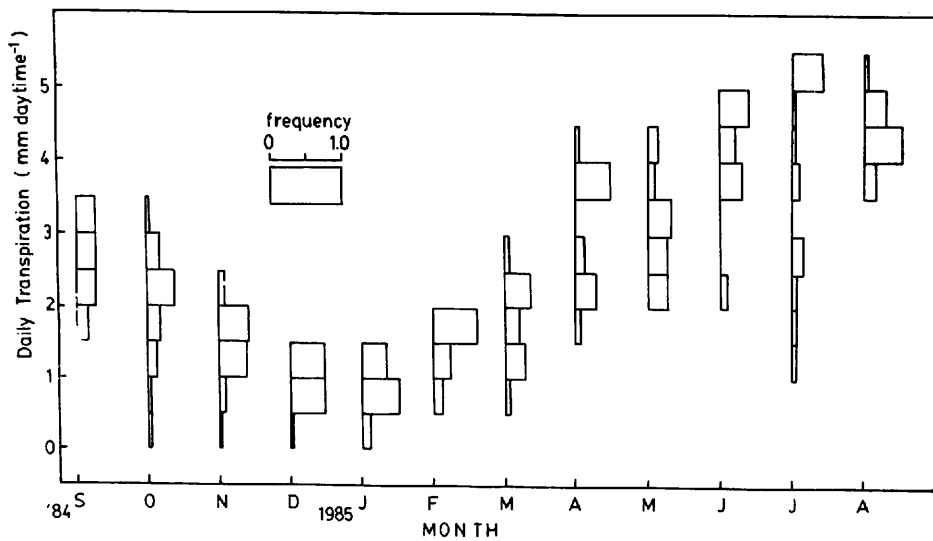


Fig. 3-8 Seasonal variation in frequency distribution of daily transpiration.

Table 3-1 Numbers of data used for analysis of dry-canopy evaporation.

Month	Number of data
Sep., 1984	18 days
Oct.	23
Nov.	24
Dec.	25
Jan., 1985	17
Feb.	8
Mar.	14
Apr.	15
May	21
June	10
July	18
Aug.	24

method and (ii) local advection in a strong northwest wind by monsoon prevailing during this period. Since both causes associate with strong wind speed, the criterion for data exclusion was determined so as to except data on days with pressure pattern of monsoon; the criterion is that daytime wind speed at a height of 30.5 m on the Tower E was more than 2 m/s.

Figure 3-8 shows that amount of transpiration was smallest in December and January with 0-1.5 mm/daytime and largest in August with 5-5.5 mm/daytime. Transpiration in June and July showed wide range of variety from the minimum of 1-2.5 mm/daytime to the maximum of 5-5.5 mm/daytime. These variations were a result of frequent weather changes during a rainy season of this period, or Bai-u.

3-3. Seasonal variation in wet-canopy evaporation

Daytime evaporation

Figure 3-9 shows seasonal variations in relative frequency distribution of evaporation rate from wet canopy. Information on the number of data is provided in Table 3-2. Evaporation occurred at a rate of 0-0.5 mm/h and mostly at 0-0.3 mm/h. These rates are nearly equal to those observed in various temperate-forests (McNaughton and Jarvis, 1983). The evaporation rate of wet canopy can be compared with that of dry canopy (Fig. 3-10), which is in the range of 0.1-0.3 mm/h in winter and 0.1-0.8 mm/h

Table 3-2 Numbers of data used for analysis of wet-canopy evaporation.

Month	Number of data
Sep., 1984	34 hours
Oct.	34
Nov.	21
Dec.	27
Jan., 1985	3
Feb.	17
Mar.	50
Apr.	61
May	69
June	93
July	37
Aug.	26

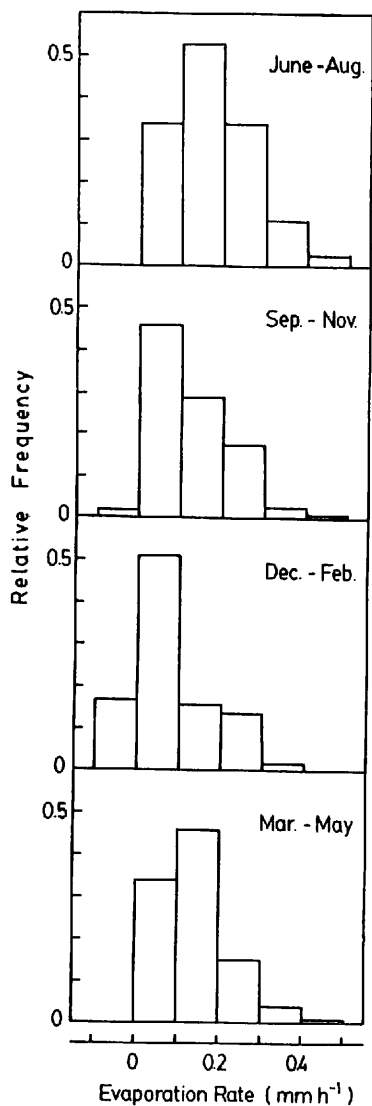


Fig. 3-9 Seasonal variation in frequency of wet-canopy evaporation rate.

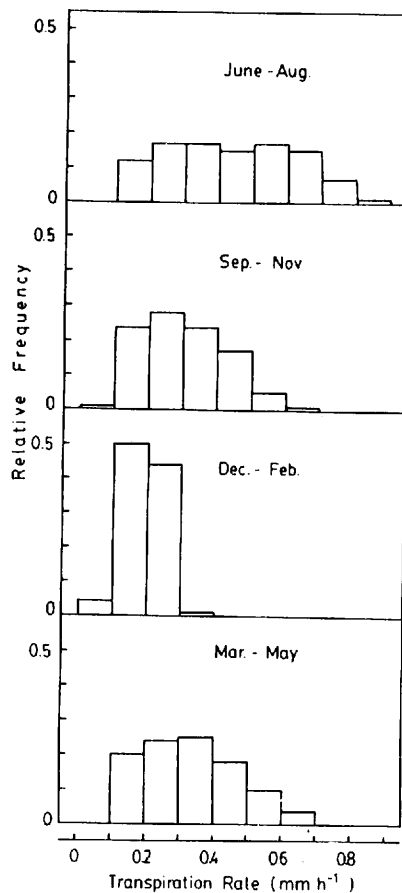


Fig. 3-10 Seasonal variation in frequency of dry-canopy evaporation rate.

in summer. Summertime evaporation rate of dry canopy is also comparable with those reviewed by McNaughton and Jarvis (1983): transpiration rates of 0.15–0.45 mm/h were given as a typical value for various temperate-forests.

Comparison between Fig. 3-9 and Fig. 3-10 reveals that evaporation rate of wet canopy was smaller than that of dry canopy and that dry-canopy evaporation rate changed seasonally, while wet-canopy

evaporation had virtually constant range of rate through the year. Although evaporation rate of wet canopy was usually smaller than that of dry canopy, a wet canopy evaporated faster than a dry canopy under the same weather condition as shown in Fig. 3–6. Under the weather condition of $C. I. < 0.5$, the ratio of latent heat to net radiation ($\ell E/R_n$) was 0.67–1.23 for a wet canopy and 0.41–0.72 for a dry canopy. Therefore the rate of evaporation of wet canopy proved to be 1.6 to 1.7 times faster than that of transpiration in the same weather condition.

Nighttime evaporation

Pearce *et al.* (1980) obtained mean nighttime evaporation rate of intercepted rainfall as 0.37 mm/h for a hardwood forest through Gash's interception model (Gash, 1979). This rate exceeds even daytime evaporation rate in the pine forest. Nighttime evaporation has significant effect in the water balance calculation, as argued by Pearce *et al.* (1980), because amount of nighttime evaporation is a net water loss. On the contrary, daytime evaporation of intercepted rain occurs concurrently with suppression of transpiration and thus net water loss turns out to be the amount as interception less suppressed transpiration. Despite this importance in forest water balance, further studies have not been conducted because of the difficulties in direct measurement of nighttime evaporation.

Interception data were analyzed to estimate wet-canopy evaporation at night in the present study. First, rainfall events that started after a sunset and ceased before the next sunrise were selected. These data were further sorted out so as to fulfill the condition: the canopy dried up before sunrise. Then, these data together with duration of canopy wetness enable us to calculate mean evaporation rate of the wet canopy at night.

The results are summarized in Table 3–3 using the data of ten rainfall events selected from the period between April, 1983 and July, 1986. Amount of rainfall ranged from less than 0.5 mm to 10.5 mm and corresponding interception ranged from less than 0.5 mm to 1.7 mm. The mean evaporation rate was calculated through division of these interception data by the duration of canopy wetness. It was

Table 3–3 Nighttime evaporation of intercepted rainfall.

	year	duration of canopy wetness from	canopy wetness through	P_g mm	P_i mm	P_s mm	I mm	E_i mm/h
1	1983	1745 h, July 8	0615 h, July 9	<0.5	0	0	<0.5	<0.04
2	1984	2015 h, May 9	2345 h, May 9					
		0300 h, May 10	0600 h, May 10	3.5	3.1	0	0.4	0.06
3		1715 h, June 1	2400 h, June 1	8.5	7.7	0.1	0.7	0.10
4		1645 h, Oct. 3	2330 h, Oct. 3	<0.5	0	0	<0.5	<0.07
5		1630 h, Oct. 4	2030 h, Oct. 4	<0.5	0	0	<0.5	<0.13
6		2200 h, Oct. 5	0430 h, Oct. 6	3.0	1.8	0	1.2	0.18
7	1985	1845 h, Feb. 13	0330 h, Feb. 14	10.5	9.3	0.1	1.1	0.13
8		1545 h, Mar. 31	0215 h, Apr. 1	6.5	4.8	0	1.7	0.16
9		1430 h, Oct. 17	2400 h, Oct. 17	2.5	1.0	0	1.5	0.16
10	1986	1900 h, Mar. 10	0200 h, Mar. 11	<0.5	0	0	<0.5	<0.07

P_g : gross rainfall P_i : throughfall P_s : stemflow I : amount of interception
 E_i : evaporation rate <: less than

from less than 0.04 mm/h to 0.18 mm/h. The evaporation rate seems to change according to the degree of canopy saturation. As terms concerning interception, water depth on a canopy is called canopy storage and maximum canopy storage is called canopy capacity. Calder and Wright (1986) measured wet-canopy evaporation rate and canopy storage of Sitka spruce forest directly using Gamma ray attenuation. The results showed rapid decrease in evaporation rate when canopy storage reduced to less than canopy capacity.

To consider this variation in evaporation rate, mean evaporation rate was compared with amount of interception (Fig. 3-11). The figure shows constant evaporation rate when amount of interception was approximately more than 1.2–1.4 mm. Below this value evaporation rate fell gradually with amount of interception. Accordingly nighttime evaporation rate of intercepted rainfall turned out to be at around 0.16–0.18 mm/h for a saturated canopy. Although these evaporation rates are small compared with daytime evaporation rate (Fig. 3-9) or transpiration rate (Fig. 3-10), integration over night results in 3 mm in winter and 2 mm in summer at maximum. These amounts are comparable to daytime transpiration in spring or fall (Fig. 3-8).

In this chapter, energy balance and evaporation rates in dry- and wet-canopy condition were presented. The results showed significant difference in both conditions. This may be attributed to the differences in mechanism of evaporation and environmental condition. In the following chapters, therefore, these problems will be discussed based on the data observed over a year.

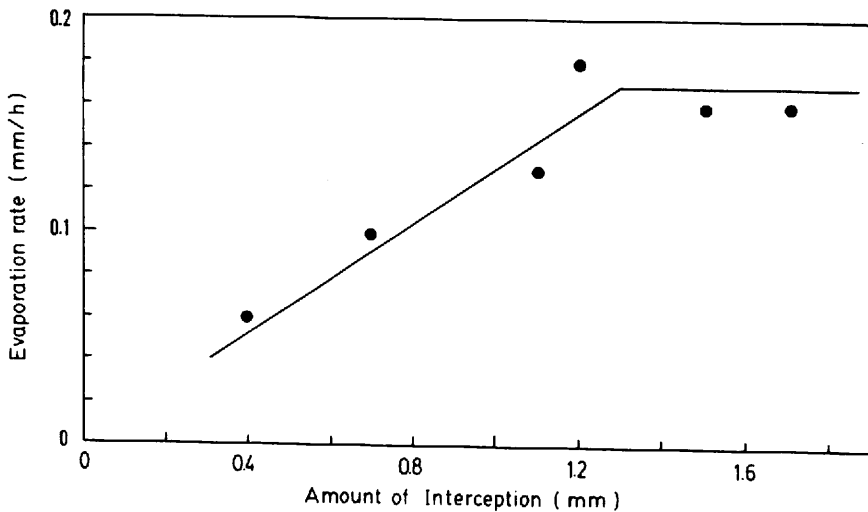


Fig. 3-11 Relation between evaporation rate and amount of interception.
Lines were drawn by eye.

CHAPTER IV

MECHANISM OF WET-CANOPY EVAPORATION

4-1. Theory of wet-canopy evaporation

If canopy resistance goes to zero for a wet canopy, evaporation from a canopy can be expressed with the following equation (Monteith, 1965):

$$\ell E = \frac{(R_n - G) + \rho C_p (e^* - e)/r_a}{\Delta + \gamma} \quad (4-1)$$

where ℓE is the latent heat flux, R_n the net radiation, G the soil heat flux, ρ the density of the air, C_p the specific heat of air at constant pressure, Δ the slope of the saturation vapor pressure curve, e the vapor pressure, e^* the saturation vapor pressure, r_a the aerodynamic resistance, and γ the psychrometric constant. The right side of eq. (4-1) includes two terms: radiation term and ventilation term. The radiation term represents energy necessary for vaporization of water, while the second term denotes energy used for diffusion of vapor into the atmosphere. Using the daytime average of the present study, we can take $r_a = 21$ s/m, $R_n - G = 50$ W/m², $e^* - e = 2$ hPa, and $\Delta = 1.45$. Then the ventilation term accounts for 61% of the latent heat flux; evaporation is determined almost evenly by both terms.

Radiation term comprises available energy and air temperature (T_a) as variables; but air temperature is of little consequence in determining radiation term because changes in Δ with regard to T_a are small. Ventilation term comprises vapor pressure deficit and aerodynamic resistance as variables. Aerodynamic resistance was almost constant over 0-4 m/s wind speed range as described in chapter V. Thus evaporation is a function of environmental factors, or available energy and vapor pressure deficit, if canopy resistance approaches zero.

4-2. Evidence of zero-stomatal resistance for wet canopy

Stewart (1977) found canopy resistance of completely wet canopy around zero through observations in a pine forest. The calculation of canopy resistance for a wet canopy using eq. (4-1) is, however, still open to question because aerodynamic resistance in the equation may be affected by the inversion of air temperature on a wet canopy (Morton, 1984). Therefore, in the present study, the heat pulse method was employed to assess canopy resistance indirectly through observations of transpiration activity with canopy wetness.

Figure 4-1 illustrates diurnal variations of evaporation, canopy wetness, heat pulse velocity (HPV), and other elements on July 28-29, 1983. Degree of canopy wetness is classified into three categories: completely wet, partly wet and dry. Completely wet is the case when all three wetness indicators show wetness of canopy. Partly wet is the case when at least one of wetness indicator indicates wet. Dry is the case when no wetness indicators registers wet.

The canopy became saturated by the 9.5-mm rainfall and stayed completely wet until 0600 h and partly wet until 0830 h. HPV became greater than 1.2 cm/h at 0815 h; transpiration became active under the partly wet condition. Accordingly all of latent heat flux under the completely wet condition and some of that under the partly wet condition turned out to be consist of evaporation of intercepted rainfall.

Similar results were obtained as shown in Fig. 4-2. A peak of latent heat flux appeared at around

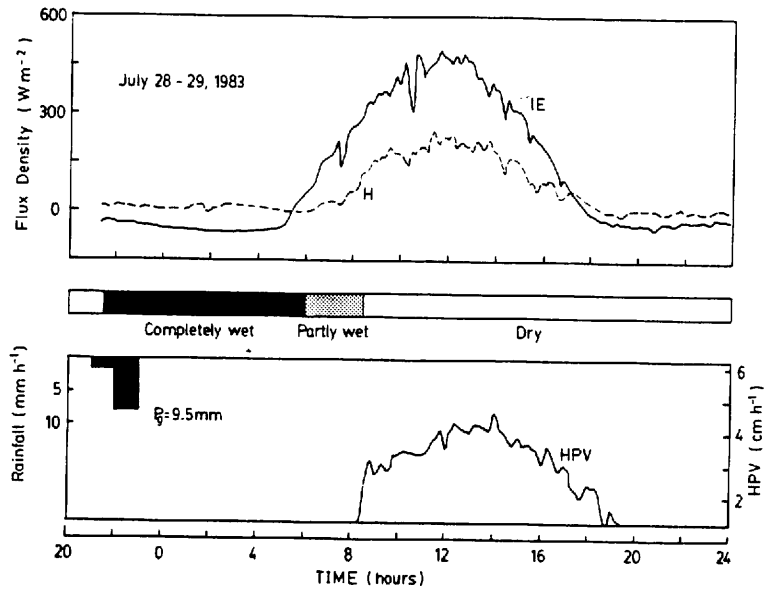


Fig. 4-1 Diurnal variation in heat pulse velocity (HPV), latent heat and sensible heat fluxes (ℓE and H), and canopy wetness on July 28-29, 1983. P_g : gross rainfall

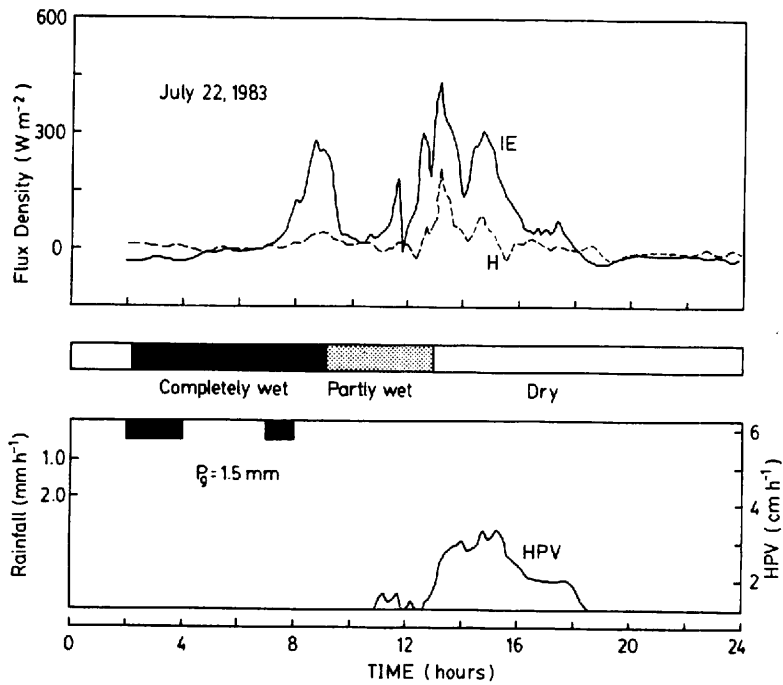


Fig. 4-2 Diurnal variation in heat pulse velocity (HPV), latent heat and sensible heat fluxes (ℓE and H), and canopy wetness on July 22, 1983. P_g : gross rainfall

0840 h on July 22 when canopy was completely wet. On the other hand, there were no corresponding peaks of *HPV*. Therefore evaporation from a completely wet canopy did not include transpiration. Above results confirm that wet-canopy evaporation consists of evaporation of intercepted rainfall and thus canopy resistance goes to zero in wet-canopy condition.

4-3. Environmental conditions for wet canopy

The foregoing sections have revealed that evaporation rate of a wet canopy is controlled solely by the environmental factors, especially vapor pressure deficit and available energy. Therefore to assess the mechanism of wet-canopy evaporation, variations in these factors should be analyzed.

Figures 4-3 and 4-4 show seasonal variations in relative frequency distribution of vapor pressure deficit during a daytime and nighttime both in wet-canopy and dry-canopy conditions. Vapor pressure deficit was almost within 4 hPa and only a slight seasonal change can be seen in wet-canopy condition.

Figure 4-5 shows relative frequency distribution of available energy in both dry-canopy and wet-canopy conditions. Under the wet-canopy condition, almost 50% of available energy was less than 50 W/m^2 through the year; and differences of distribution among seasons were small. Available energy for a dry canopy, on the contrary, scattered in the range as wide as 600 W/m^2 and showed significant seasonal variations.

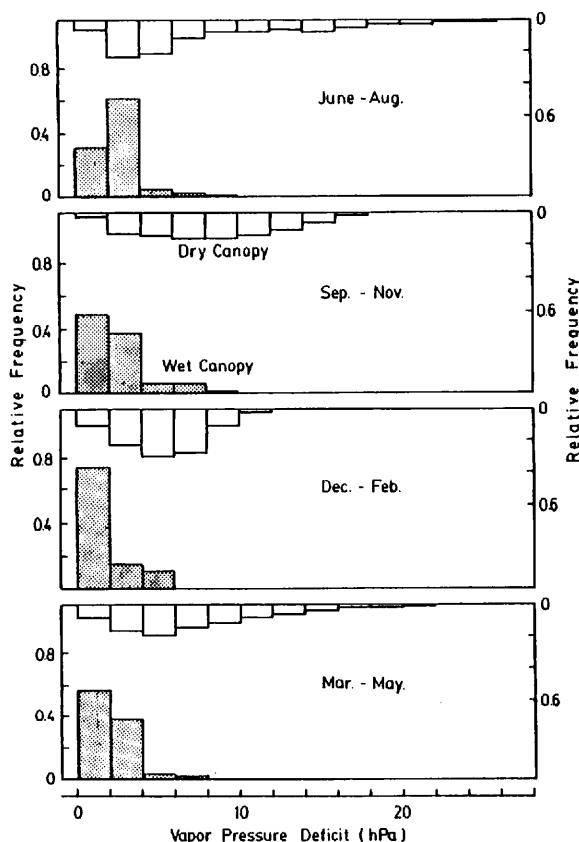


Fig. 4-3 Seasonal variation in frequency of daytime vapor pressure deficit

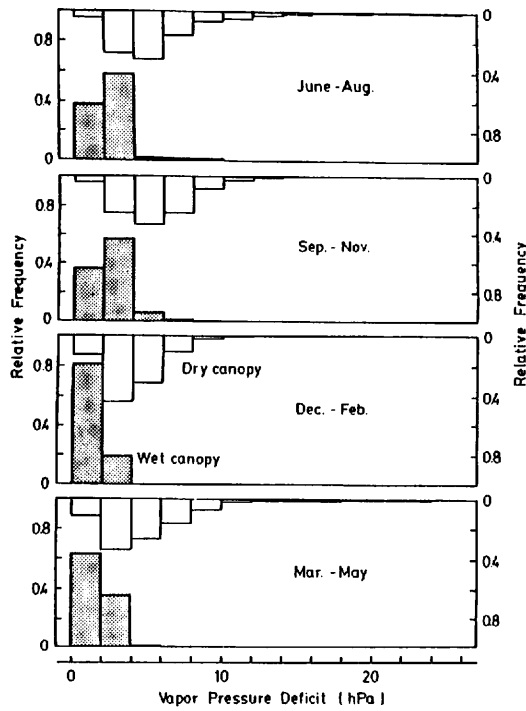


Fig. 4-4 Seasonal variation in frequency of nighttime vapor pressure deficit.

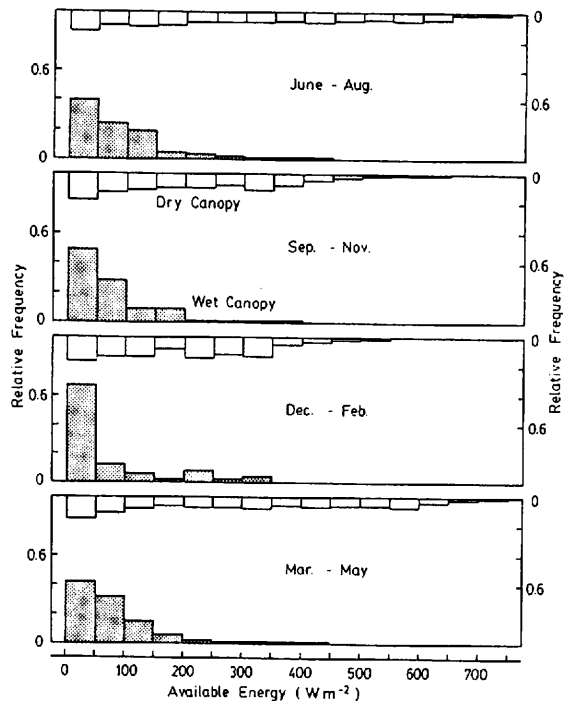


Fig. 4-5 Seasonal variation in frequency of daytime available energy.

These results lead to the evaporation rate of a wet canopy shown in chapter III, which was almost constant as 0–0.3 mm/h through the year. On the other hand, latent heat flux sometimes exceeded net radiation for wet-canopy condition (Fig. 3–6). As a result, additional energy is necessary for evaporation in this case. This energy source has been controversial in recent years. Some researchers argued it came from large scale advection (Stewart, 1977; Calder, 1985), while Morton (1984, 1985) questioned that idea.

Figure 4–6 illustrates relationship between hourly available energy and latent heat flux in the pine forest when canopy was wet. Also shown in the figure is the isoline of sensible heat flux. Latent heat flux exceeded available energy especially when available energy was less than 100 W/m². As a result, downward sensible heat flux compensated the shortage of energy required for evaporation. Therefore additional energy source for evaporation was identified as downward sensible heat flux.

The mechanism of occurrence of this flux, however, remains vague, because we have no information on the condition of upper atmosphere. For a thorough understanding, therefore, observations which cover the height of atmospheric boundary layer may be necessary.

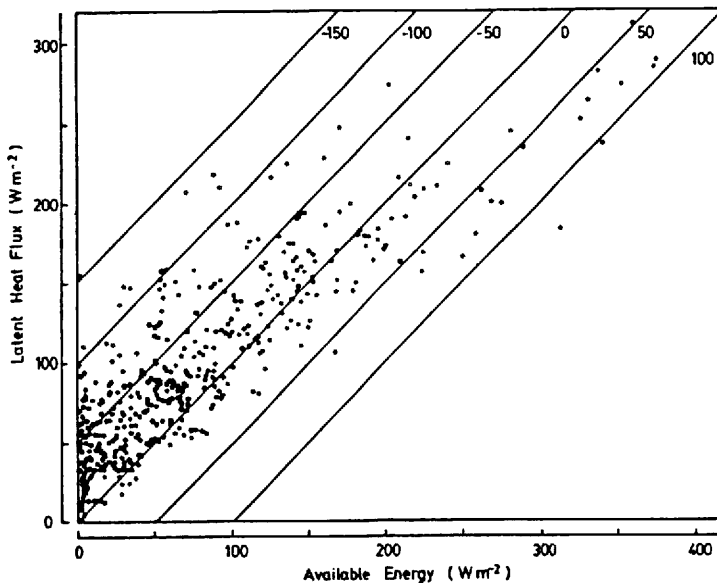


Fig. 4–6 Relation between available energy and latent heat flux.

CHAPTER V

MECHANISM OF DRY-CANOPY EVAPORATION

5-1. Theory of dry-canopy evaporation

The evaporation from a vegetation can be expressed with the Penman-Monteith equation (Monteith, 1965) with an assumption that gas exchange occurs at a height of imaginary one "big leaf" as stated in chapter I:

$$\ell E = \frac{(R_n - G) \Delta + \rho C_p (e^* - e)/r_a}{\Delta + \gamma (1 + r_c/r_a)} \quad (5-1)$$

where ℓE is the latent heat flux, R_n the net radiation, G the soil heat flux, ρ the density of the air, C_p the specific heat of air at constant pressure, Δ the slope of the saturation vapor pressure curve, e the vapor pressure, e^* the saturation vapor pressure, r_a the aerodynamic resistance, r_c the canopy resistance, and γ the psychrometric constant.

The most different aspect of dry-canopy evaporation from wet-canopy evaporation is that transpiration is controlled not only by the environmental factors but also by a plant itself through stomatal aperture. Accordingly to assess the mechanism of dry-canopy evaporation, understanding of the stomatal behavior is essential. The behavior is expressed in eq. (5-1) as a canopy resistance. Therefore in the following sections, behavior of canopy resistance will be discussed.

5-2. Diurnal and seasonal variations in canopy resistance

Canopy resistance (r_c) may be calculated by rearranging eq. (5-1):

$$r_c = \left(\frac{\Delta}{\gamma} \beta - 1 \right) r_a + \frac{\rho C_p (e^* - e)}{\gamma \ell E} \quad (5-2)$$

where ℓE is the latent heat flux, ρ the density of the air, C_p the specific heat of air at constant pressure, Δ the slope of the saturation vapor pressure curve, e the vapor pressure, e^* the saturation vapor pressure, r_a the aerodynamic resistance, β the Bowen ratio, and γ the psychrometric constant. Aerodynamic resistance (r_a) can be obtained with friction velocity (u^*) as follows:

$$r_a = u/u^{*2} = u/(\overline{u'w'}) \quad (5-3)$$

where u and w is the horizontal and vertical wind speed, respectively, and u^* the friction velocity. Prime and overbar denote mean and fluctuation therefrom. Figure 5-1 shows relation between 10-min. average u and u^* observed in the summer of 1986 and Fig. 5-2 illustrates relation between u and r_a calculated by eq. (5-3). Aerodynamic resistance was almost constant over a wind speed range of 0-4 m/s. Accordingly overall mean of 21 s/m was adopted as r_a .

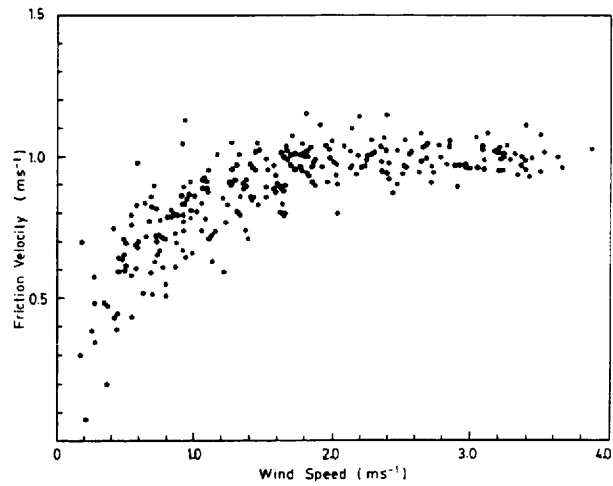


Fig. 5-1 Relation between wind speed and friction velocity.

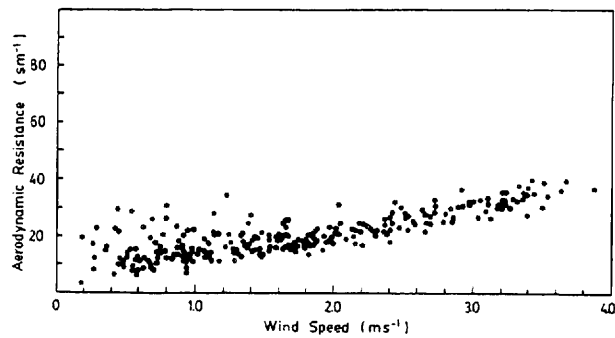


Fig. 5-2 Relation between wind speed and aerodynamic resistance.

Figure 5-3 gives diurnal variations in canopy resistance in August, 1985. The small resistance in the early morning stayed fairly constant until early afternoon, after which time it rose progressively, increasing almost by a factor of two by early evening.

Figures 5-4a and 5-4b illustrate diurnal variation in r_c averaged for each month. An average value was obtained as a harmonic mean of hourly data calculated by eq. (5-2). Although general trend of diurnal variations in r_c was similar to each other, difference in magnitude of r_c still existed among them

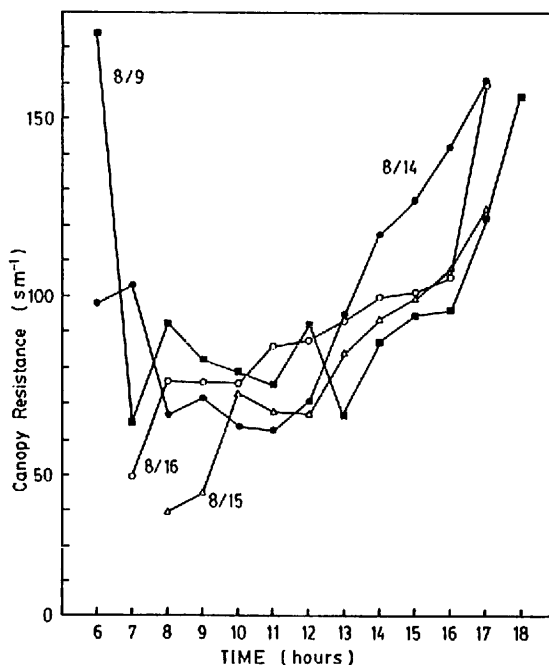


Fig. 5-3 Diurnal variations in canopy resistance in August, 1985.

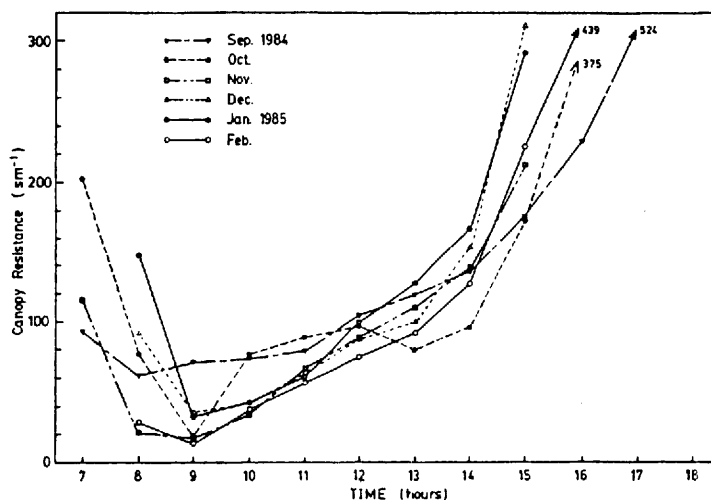


Fig. 5-4a Diurnal variations in canopy resistance obtained as a monthly mean.

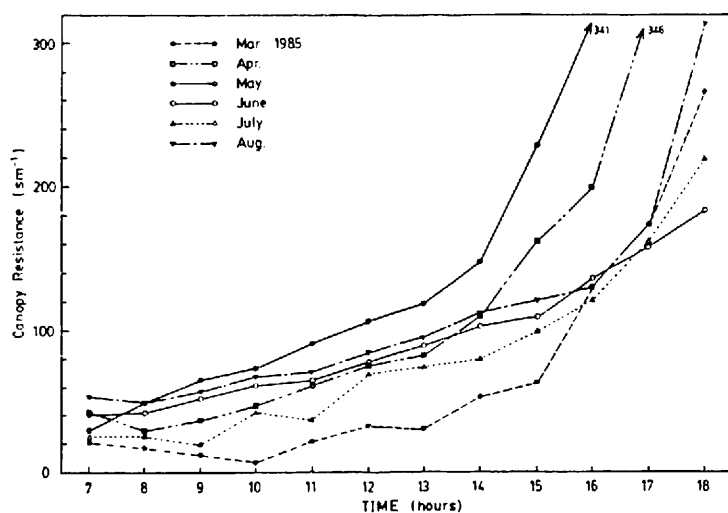


Fig. 5-4b Diurnal variations in canopy resistance obtained as a monthly mean.

unlike the results in Gash and Stewart (1975). This is obvious in Fig. 5-5 which gives seasonal variation in daytime canopy resistance. Canopy resistance changed seasonally and the curve was not likely to take a simple functional form. This suggests that many factors with different seasonal variations make up canopy resistance in a complex way.

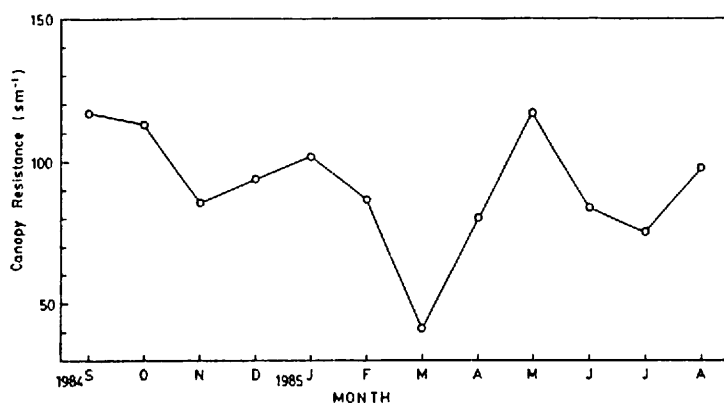


Fig. 5-5 Seasonal variation in canopy resistance.

5-3. Mechanism of variations in canopy resistance

Canopy resistance can be given as stomatal resistance divided by leaf area index (Federer, 1975; Tan and Black, 1976; Federer, 1979; Choudhury and Idso, 1985) as:

$$r_c = r_s / LAI \quad (5-4)$$

where r_s and r_c are stomatal and canopy resistance. Accordingly variations in canopy resistance are caused by two different variables.

Seasonal variations in leaf area index

Leaf area index (LAI) was obtained using the data of weight of leaf biomass per unit area (W_b). To convert W_b to LAI , the regression equation was obtained based on the data of biomass measurement on June 28, 1985:

$$LAI = 5.784 \times W_b - 0.086 \quad (r = 0.998) \quad (5-5)$$

Seasonal variation in LAI calculated from this equation is illustrated in Fig. 5-6 together with weights of leaf fall and foliation. After shedding of leaves in October to November, LAI approached 63% of that in July by the early December. Leaf area index took the smallest value of 1.68 in the middle of May and it increased rapidly to 2.98 by the middle of June with foliation. Therefore LAI changed almost twice through the year, which resulted in corresponding changes in canopy resistance.

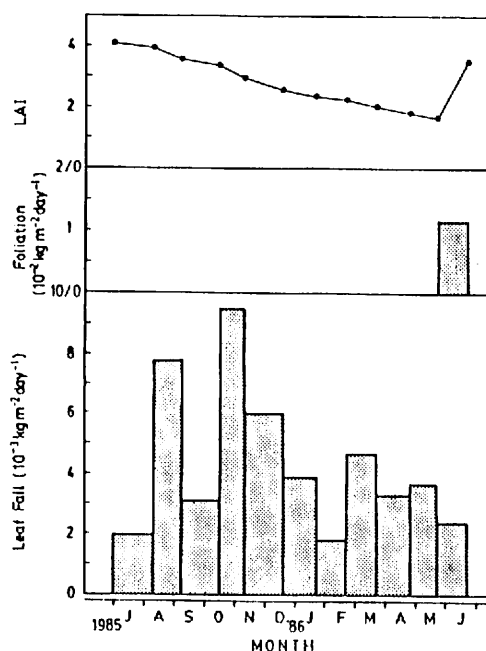


Fig. 5-6 Seasonal variation in leaf area index (LAI), amount of leaf fall and foliation.

Stomatal closure due to soil water deficit

Stomatal closure of a plant to prevent excess water loss may have two mechanisms (Lösh and Tenhunen, 1981). One is the feedback mechanism which begins functioning after a threshold water potential of leaves is surpassed. Another is the feedforward mechanism which controls stomatal aperture with changes in ambient humidity independent of water potential of the leaf. Confirmation of the feedforward mechanism in a plant has been still in a limited number. The well known feedback mechanism usually has origin in water deficit in soil water. When soil water decreases and the rate of water loss from a plant exceeds the rate of water uptake from the soil, the plant responds by closing the stomata as a result of turgor loss in the leaf (Davies *et al.*, 1981).

The stomatal closure causes suppression of transpiration. Many researches for crop and pasture have reported the decrease in transpiration due to soil water deficit (for example, Stanhill, 1957). The influences of soil water on forest transpiration, however, have been studied only a little so far as mentioned in chapter I. These results are compiled and given in Table 5-1. In this table, soil water contents when transpiration begins to decrease were expressed as critical value of fractional extracted water in the root zone (R_{wc}). Fractional extracted water is defined as:

$$R_w = (W_f - W_s) / (W_f - W_w) \quad (5-6)$$

where W_f is the field capacity, W_s the water content in the root zone, and W_w the water storage at a wilting point. The denominator in eq. (5-6) is called available water and it represents water content available to the plant in the root zone. The R_{wc} values by Rutter (1968) were determined from Figure 5 in the original literature, although some obscure data were excluded from analysis. The R_{wc} value by Halldin *et al.* (1984) was not obtained from direct field observations; they calculated transpiration with their model and optimized the R_{wc} value by comparing it with measured transpiration data. Other R_{wc} values were calculated or excerpted from each original literature.

Table 5-1 Effect of soil water deficits on transpiration.

vegetation	soil	depth of root system	available water	R_{wc}	reference
<i>Pinus sylvestris</i>	sandy loam	1.8 m	370 mm	60%	Rutter (1967)
Mixed chaparral species	sandy clay loam	1.0	155	70	Rutter (1968)
Mixed chaparral species	sandy clay loam	1.5	250	50	Rutter (1968)
<i>Populus tremuloides</i>	?	1.8	300	50-70	Rutter (1968)
?	sandy clay loam	1.8	190	60-80	Rutter (1968)
Mixed chaparral species	clay loam	1.0	210	40-60	Rutter (1968)
<i>Pseudotsuga menziesii</i>	gravelly sandy loam	0.5	88	65	Black (1979)
<i>Pseudotsuga menziesii</i>	gravelly sandy loam	0.7	82	63	Black (1979)
<i>Eucalyptus maculata</i>	?	?	?	60	Dunin and Aston (1984)
<i>Quercus sessiliflora</i>	sand	1.0	133	74	Halldin <i>et al.</i> (1984)

available water: $W_f - W_w$, R_{wc} : critical value of R_w , R_w : fractional extracted water

Table 5-1 reveals that the R_{wc} values stayed in the range of 50-70% although vegetation and available water differed largely among literatures. This means that extraction of 50-70% of available water causes decrease in transpiration even in a forested area. Tanner and Ritchie (1974) compiled 24 crop transpiration data in a similar way to the present study and obtained the value of R_{wc} as 65-85%. This value is the same order of magnitude with the R_{wc} value in a forest. This suggests that many plants have the critical point at similar level regardless of vegetation type.

To compare with the critical point value of R_{wc} in Table 5-1, seasonal variations in the R_w value in the root zone were obtained in the study area. First, depth of root zone was determined through excavation of a whole root system of the pine tree used for the biomass measurement. The roots were classified in terms of their diameter and then the surface area was measured. Figure 5-7 illustrates

distribution of the root system. The fine root, which plays principal part in soil water extraction, distributed from soil surface down to a depth of 140 cm. Thus the layer with a depth of 0–140 cm was defined as a root zone.

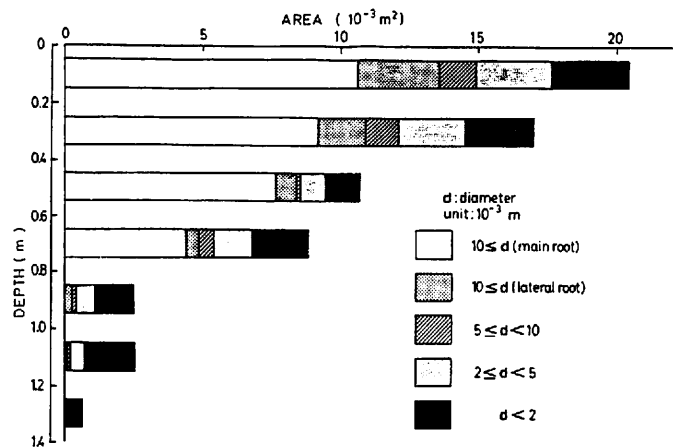


Fig. 5-7 Vertical distribution of surface area of root system.

Secondly, field capacity and water content at a wilting point were determined through pF test. Soil samples were taken at 10-cm intervals with a 100-cm³ soil sampler. Figures 5-8a, 5-8b, and 5-8c illustrate soil-water characteristic curves for these soil samples. Table 5-2 provides water contents at each

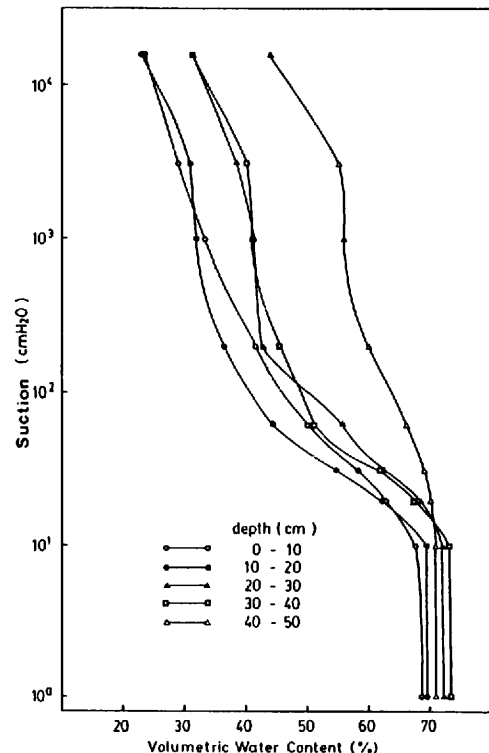


Fig. 5-8a Moisture characteristic curves.

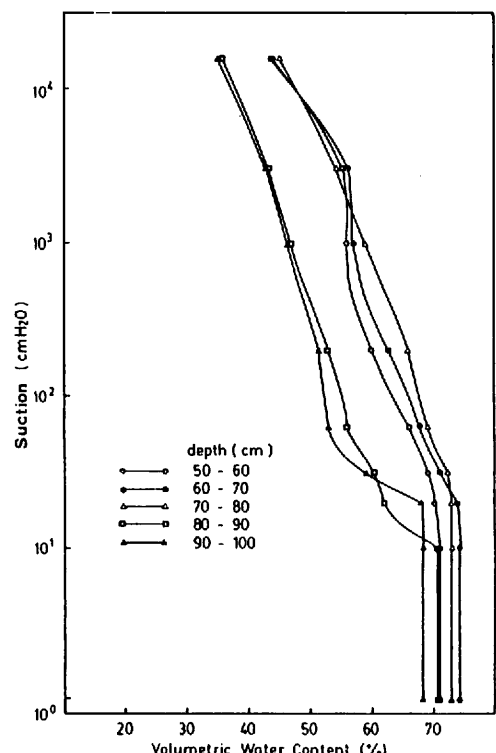


Fig. 5-8b Moisture characteristic curves.

water constant. These data enable us to calculate available water in the root zone as 295.8 mm (21.1% in volumetric water content).

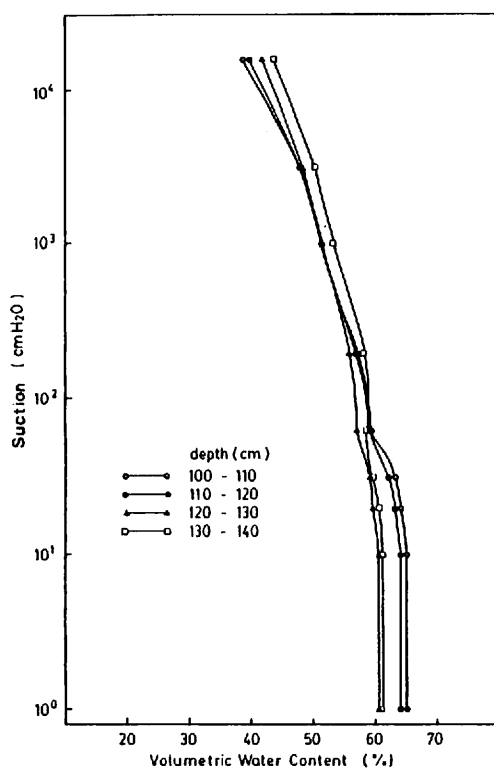


Fig. 5-8c Moisture characteristic curves.

Table 5-2 Soil-water characteristics.

W_f (at pF 1.8)	813.0
W_w (at pF 4.2)	517.2
$W_f - W_w$	295.8

$W_f - W_w$: available water

W_f : field capacity

W_w : water storage at a wilting point

unit: mm/1.4 m

Thirdly, seasonal variations in water content were measured (Fig. 5-9). The changes in water content do not correspond clearly to individual rainfall events, since soil water content was measured at one-month intervals. Soil water took constant values as an order of 50% for a depth of 20–40 cm and 60% for 80–140 cm layer in spite of uneven distribution of rainfall events through the year. The conversion of these water contents into the R_w values following eq. (5-6) revealed that fractional extracted water accounted for 40.3% at maximum and usually took values of 10–25%. These R_w values are much smaller than the values of R_{wc} provided in Table 5-1, which suggests that decrease in transpiration due to soil water deficit will not occur in the study area.

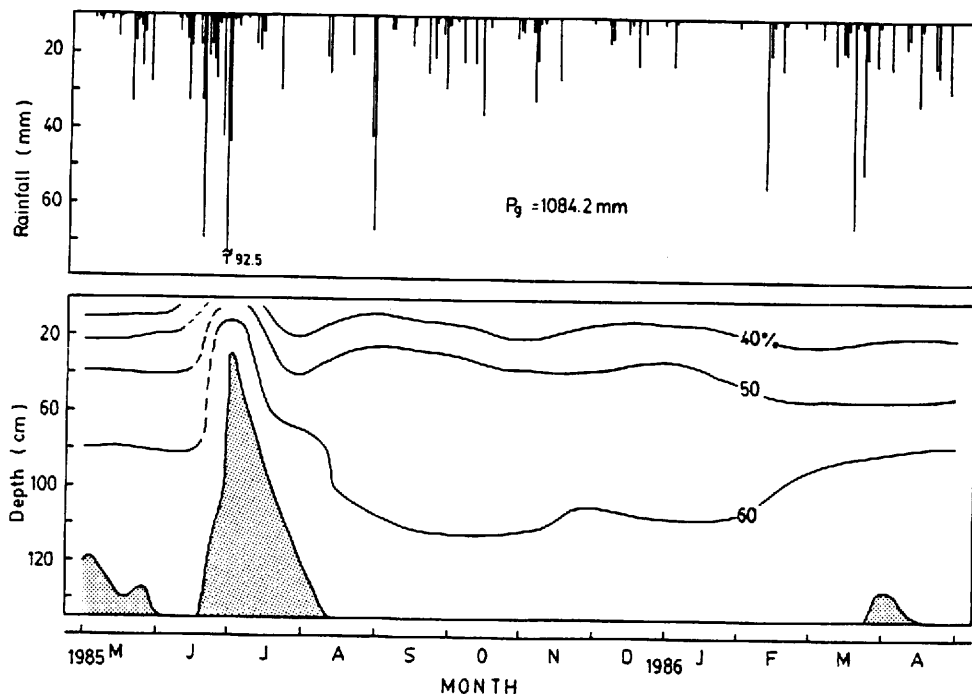


Fig. 5-9 Seasonal variation in soil water content (volumetric, %) with rainfall. Shaded parts represent water body. P_g : gross rainfall

To confirm this, day to day changes in transpiration were measured concurrently with suction variations at a depth of 40 cm from October 10 to December 20, 1984 (Fig. 5-10). In 1984, amount of rainfall was extremely small with only 56% of an average rainfall for the past 20 years (Yamashita, 1985) so that soil was expected to be unusually in dry condition. No correlation was found between suction and normalized transpiration in terms of equilibrium evaporation (E_t/E_{eq}). This means that even a smallest rainfall in 20 years did not cause suppression of transpiration. Consequently transpiration proved not to decrease usually due to soil water deficit in this study area.

Stomatal aperture with meteorological variables

As mentioned previously, recent studies have revealed that stomatal aperture of some species was controlled with ambient humidity together with other environmental elements.

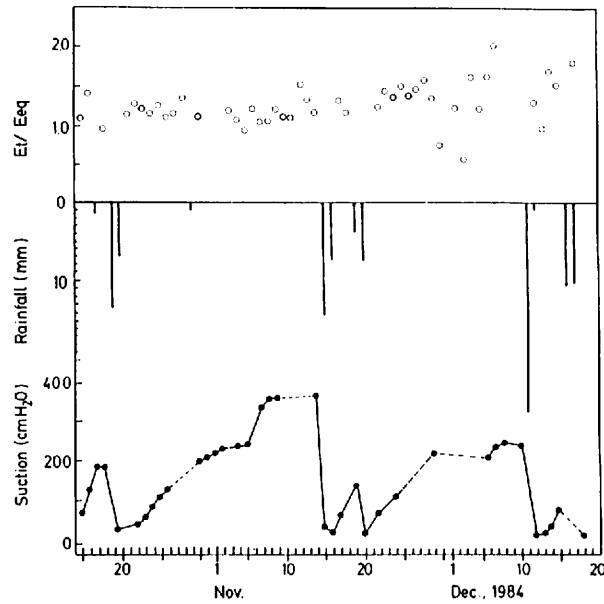


Fig. 5-10 Day to day variations in rainfall, E_t / E_{eq} , and suction at a depth of 40cm.

Figure 5-11 shows relation between hourly-averaged net radiation and natural logarithm of stomatal resistance. The lines in the figure are drawn by eye and they reveal that the stomatal resistance became large remarkably below the net radiation of 80 W/m^2 . Therefore the net radiation of 80 W/m^2 is thought to denote the threshold value, above which stomatal response was light saturated and independent of

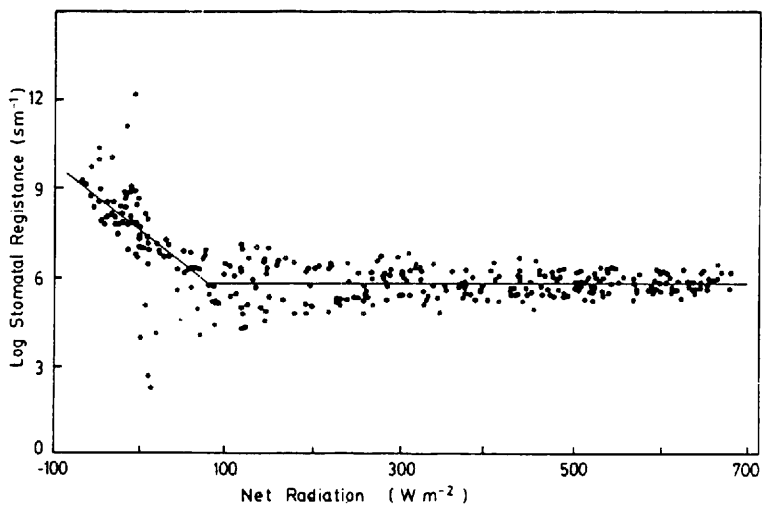


Fig. 5-11 Relation between net radiation and stomatal resistance. Lines were drawn by eye.

irradiance. This value is in the same order of magnitude as those reported for various forest tree species. Simpson *et al.* (1985) showed that stomata of a Douglas fir had a radiation threshold at 150 W/m^2 in global radiation. Other studies also provided data that suggest rather low levels of radiation threshold for various trees: 150 W/m^2 for Douglas fir (Tan and Black, 1976), $100\text{--}200 \text{ W/m}^2$ for white oak (Hinckley *et al.*, 1975), and less than 100 W/m^2 for various woody angiosperms (Hinckley *et al.*, 1978). Accordingly, for a forest tree, there seems to be little effect of solar radiation on r_s for the most of daytime, i.e. the period except for early morning and late afternoon.

To exclude the radiation effect from the following analysis, stomatal resistance and other meteorological data when R_n was greater than 80 W/m^2 were selected. Table 5-3 provides results of regression analysis based on the selected data between stomatal resistance and other meteorological variables. The correlation coefficient between stomatal resistance and vapor pressure deficit shows that 61% of changes

Table 5-3 Regression analysis between stomatal resistance and meteorological elements.

dependent	regression equation	correlation coefficient
VPD	$r_s = 133.73 + 13.27 \times VPD$	0.779
R_n	$r_s = 153.87 + 0.29 \times R_n$	0.592
T_a	$r_s = -326.65 + 20.83 \times T_a$	0.569
u	$r_s = 198.53 + 57.69 \times u$	0.383
VPD, R_n, T_a	$r_s = 239.50 + 16.90 \times VPD$ $+ 0.07 \times R_n - 4.02 \times T_a$	0.788

VPD : vapor pressure deficit, R_n : net radiation,
 T_a : air temperature, u : wind speed,
 r_s : stomatal resistance.

in stomatal resistance can be explained with vapor pressure deficit. The relation between stomatal resistance and vapor pressure deficit is shown in Fig. 5-12. Following vapor pressure deficit, net radiation and air temperature have correlation coefficients of 0.569-0.592. Wind speed is poorly related to variations in stomatal resistance. The high correlations of air temperature and net radiation with stomatal

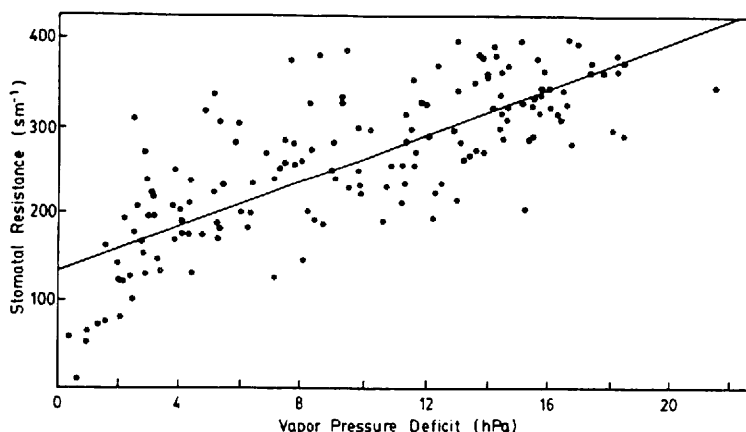


Fig. 5-12 Relation between vapor pressure deficit and stomatal resistance.

resistance are, however, a result of the fact that vapor pressure deficit is closely related to net radiation and air temperature. This can be seen in the multiple regression analysis using three meteorological variables as dependents in Table 5-3, which shows only a slight increase in the correlation coefficient from single regression analysis using vapor pressure deficit as a dependent. Consequently stomatal resistance turned out to be determined by vapor pressure deficit for $R_n > 80 \text{ W/m}^2$.

Table 5-4 provides similar results of regression analysis between stomatal resistance and vapor pressure deficit for different seasons. The regression equations revealed that the stomatal resistance went to similar minimum values around 100–150 s/m when vapor pressure deficit approached zero. This value may represent the stomatal resistance when stomata fully opens. The slope of the regression equations shows seasonal variations: it was smallest in spring and biggest in winter.

Table 5-4 Regression analysis between stomatal resistance and vapor pressure deficit.

season	regression equation	correlation coefficient
spring	$r_s = 94.47 + 9.68 \times VPD$	0.470
summer	$r_s = 145.62 + 11.87 \times VPD$	0.679
fall	$r_s = 131.72 + 14.81 \times VPD$	0.590
winter	$r_s = 111.63 + 18.55 \times VPD$	0.473

r_s : stomatal resistance VPD : vapor pressure deficit

Tan *et al.* (1977) pointed out that twig xylem pressure potential influenced the stomatal behavior. Simpson *et al.* (1985) obtained relation between r_s and the product of vapor pressure deficit and xylem pressure potential and found good correlation between them. These results have shown that increase in xylem pressure caused increase in stomatal resistance. Although no water potential data are available in the present study, Sato (1977) observed seasonal variation in xylem pressure potential of a young Japanese cypress tree. The result showed that the pressure was smallest in winter and biggest in summer, which suggests the smallest stomatal resistance in winter and biggest in summer. This conflicts with the observed variation in stomatal resistance. Accordingly if the seasonal variation in xylem pressure potential is applicable to the present study, it cannot explain changes in stomatal resistance of the pine trees.

Other possible explanations for the changes in the slope of the regression equations are changes in carbon dioxide concentration, air temperature and age of leaves of the pine tree. Changes in carbon dioxide are reported to cause relatively small variations in stomatal aperture (Lösh and Tenhunen, 1981). When air temperature decreases, on the other hand, rate of changes in stomatal resistance with vapor pressure deficit increases (Lösh and Tenhunen, 1981). This can explain the biggest stomatal resistance in winter. The foliage of pine trees just prior to leafing is oldest through the year. This period is spring for a pine forest as shown in Fig. 5-6. The old leaf may lose physiological control on stomatal aperture. This is likely to lead to insensitivity of stomatal resistance to vapor pressure deficit and thus results in the small slope of the regression equation in spring.

The foregoing consideration has revealed that stomata of a pine tree opened with vapor pressure deficit for a light saturated condition. The sensitivity of stomatal resistance with the changes in vapor pressure deficit changed seasonally and the cause is suggested as changes in air temperature and leaf

age. The factors affecting stomatal aperture are, however, difficult to identify, because these factors depend each other. Feedback and feedforward mechanism also make analysis difficult. For example, decrease in leaf water potential results in decrease in transpiration through stomatal closure, which in turn causes increase in leaf water potential. To study the mechanism of stomatal aperture, physiological studies like experiments in a controlled condition may be necessary together with field observation.

CHAPTER VI

CONCLUSIONS

To assess the mechanism of forest evaporation, observations over three years were conducted at a pine forest. Analyses of these data with emphasis on seasonal variation revealed changes in wet- and dry-canopy evaporation rates and their mechanisms. The results are summarized as follows:

Seasonal variations in evaporation

Ratio of sensible heat flux to net radiation changed seasonally and with weather condition. It took the largest value in summer when canopy was wet, while lowest in winter in clear sky condition. The evaporation rate of wet canopy was usually within 0.3 mm/h during a daytime and 0.16–18 mm/h during a nighttime through the year. Evaporation rate of dry-canopy, on the other hand, changed seasonally with highest rate of 0.1–0.8 mm/h in summer and lowest rate of 0.1–0.3 mm/h in winter. The integration of hourly evaporation rate of dry canopy over a daytime amounted to the smallest value of 0–1.5 mm in January and the biggest value of 5–5.5 mm in August.

Mechanism of wet-canopy evaporation

Observations with the heat pulse method showed that evaporation comprised of intercepted rainfall when canopy was wet and thus canopy resistance went to zero. This means that evaporation rate can be expressed with available energy and vapor pressure deficit. Both vapor pressure deficit and available energy showed little seasonal variations in the wet-canopy condition unlike in the dry-canopy condition. This environment can explain the lack of seasonal variation in wet-canopy evaporation rate.

Mechanism of dry-canopy evaporation

When canopy was dry, transpiration is controlled through canopy resistance. Therefore behavior of this resistance was analyzed.

(1) Canopy resistance stayed almost constant before early afternoon and then rapidly increased in the afternoon. This pattern of the diurnal variation was consistent through the year. Magnitude of resistance, however, differed seasonally. The seasonal variation had two origins: changes in leaf area index and stomatal aperture.

(2) Leaf area index decreased through leaf fall from the maximum value of 4.0 in July to the minimum value of 1.68 in the middle of July and then increased with foliation.

(3) Stomatal aperture was studied on the two factors: soil water deficit and meteorological elements. Analyses of recent studies in terms of fractional extracted water (R_w) revealed that transpiration began to decrease when R_w value exceeded 50–70%. In the pine forest, however, transpiration proved not to be suppressed due to soil water deficit because soil water scarcely reached R_w value of 50%.

(4) Vapor pressure deficit had the highest correlation coefficient among the meteorological elements for $R_n > 80 \text{ W/m}^2$. Regression analysis between stomatal resistance and vapor pressure deficit for four seasons showed that the minimum stomatal resistance was at around 100 s/m and that slopes of the regression equation changed seasonally. The causes of this variation were suggested as changes in leaf age and air temperature.

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