AN EVAPORATION MODEL FOR

HIGH LATITUDE LICHEN SURFACES

AN EVAPORATION MODEL FOR

HIGH LATITUDE UPLAND LICHEN SURFACES

by

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SCOPE AND CONTENTS:

Energy-budget calculations and equilibrium model estimates of
evaporation from a lichen-dominated upland site in the Hudson Bay low-
lands are presented.   The energy budget calculations reveal that the
lichen surface is relatively resistant to evaporation with an average of
only 54 percent of the daily net radiation being utilized in the evap-
orative process.   Equilibrium estimates of evaporation consistently over-
estimate actual evaporation by 5 and 8 percent for hourly values and daily
totals respectively.   A simple model, a function of the equilibrium model,
is derived from a comparison of actual and equilibrium evaporation.   The
only inputs required for the model are net radiation, soil heat flow and
screen temperatures.   Tests of the model indicate that it will predict
actual evaporation within 5 percent and that it can probably be applied
to any high latitude surface which exhibits a relatively large resistance
to evaporation.

ACKNOWLEDGEMENTS

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CHAPTER I

INTRODUCTION AND PURPOSE

Evaporation from land surfaces is a function of energy input and

variable interactions between the atmosphere, soil and vegetation. A

significant amount of research has been devoted to defining mathematically

the relation of the evaporative flux to these environmental parameters.

One of the more simple models for estimating evaporation is the

equilibrium model presented by Slatyer and McIlroy (1961). The model

describes the evaporative flux in terms of temperature-dependant terms

and the radiant energy supply.

The equilibrium model as originally presented by Slatyer and

McIlroy, and subsequently tested by Monteith (1965) and Pruitt and Lourence

(1968), was considered only applicable in conditions of a saturated

atmosphere. Further tests, however, have revealed that the model is also

applicable in non-saturated conditions (Denmead and McIlroy, 1970; Wilson,

1971). Wilson found that the equilibrium model performed exceptionally

well over a moderately dry surface exhibiting a relatively high resistance

to evaporation.

Rouse and Kershaw (1971), using the energy balance method to

evaluate evaporation, found that subarctic lichen dominated surfaces were

relatively resistant to evaporation although the soil moisture was at, or

near field capacity. A comparison of the Bowen ratio values observed by

Wilson, with those of Rouse and Kershaw revealed that the lichen surfaces could be equated to the moderately dry surfaces studied by Wilson. This implies, that the equilibrium model or some function of the equilibrium model might be used to estimate the evaporation for subarctic lichen-covered surfaces.

This study presents results of actual evaporation determined from the energy budget approach for a subarctic site. In addition, equilibrium model estimates of evaporation are compared to the energy balance measurements in order to test the hypothesis that the equilibrium model or some function of the equilibrium model can be used to accurately estimate the evaporation for a moderately dry subarctic surface.

The research was conducted on a lichen-covered raised beach system in the Hudson Bay lowlands of northern Ontario during July and August, 1971.

CHAPTER II

THEORY OF ENERGY BALANCE

AND EQUILIBRIUM EVAPOTRANSPIRATION

1.   The Energy Balance

The energy balance approach is based on the principle of the
conservation of energy which equates the gains and losses of energy at
the earth's surface. The energy balance which applies in the absence of
advective sensible and latent heat and which neglects photosynthesis is
expressed as

$$Rn - G \quad = \quad H + LE \tag{1}$$

where Rn is the net radiation, G is the soil heat flux, H is the sensible
heat flux and LE the latent heat flux (L representing the latent heat of
vapourization and E the amount of evapotranspiration). Photosynthesis
can be neglected on a daily basis since measured values for the most part
have been less than 5 percent of Rn (Lemon, 1960; Yocum, Allen, and Lemon,
1964). Similarly, storage of heat energy in the plant biomass is
considered negligible since it tends to approach zero on a daily basis
(King, 1961). These latter two omissions are certainly valid for a
slow-growing lichen canopy in the subarctic.

LE can be determined from (1) if the remaining components in the

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equation are known. Rn and G can be measured or calculated directly whereas the techniques to measure H and LE directly are still in the experimental stage. One indirect solution is to partition the available radiant energy represented by Rn - G into individual values for H and LE. This is done by expressing the sensible and latent heat fluxes in a ratio form H/LE known as the Bowen Ratio (Bowen, 1926). Dividing (1) by LE gives

$$Rn - G = LE(H/LE + 1) \tag{2}$$

where H/LE = β the Bowen ratio. Solving for LE produces

$$LE = \frac{Rn - G}{1 + \beta} \tag{3}$$

With the use of mass transfer theory H and LE can be expressed over a finite gradient as

$$H = -\rho Cp K_H \frac{\Delta T}{\Delta z} \tag{4}$$

and

$$LE = \frac{-\rho L\epsilon}{P} K_W \frac{\Delta e}{\Delta z} , \tag{5}$$

where ρ is the air density, Cp is the specific heat of air at constant pressure, P is the atmospheric pressure, ϵ = 0.622 the ratio of molecular weights of water to air, K_H and K_W are the eddy diffusivities of heat and water vapour, and ΔT, Δe are the gradients of dry bulb temperature and

vapour pressure over a height increment Δz. Evaluation of ΔT and Δe implies measurement of T and e at two or more levels.

Swinbank and Dyer (1967) and Dyer (1967) have shown that $K_W = K_H$ over a wide range of atmospheric stability. Assuming this equivalence and dividing (4) by (5) gives the Bowen Ratio as

$$\beta \quad = \quad H/LE \quad = \quad \gamma \frac{\Delta T}{\Delta e} \tag{6}$$

where $\gamma = PCp/L\epsilon = Cp/AL = 0.66$ mb C^{-1} the pyschrometric constant ($\epsilon = PA$ with A representing the accurately known psychrometric constant). Although there is a small error with the use of γ for practical purposes it is considered insignificant (Slatyer and McIlroy, 1961).

Substituting (6) in (3) produces

$$LE \quad = \quad \frac{Rn - G}{1 + \gamma \frac{\Delta T}{\Delta e}} \tag{7}$$

Following Dilley (1968) the vapour pressure can be calculated from

$$e \quad = \quad e_s - AP(1 + 0.00115 \; Tw)(T - Tw) \tag{8}$$

where e_s is the saturation vapour pressure at the wet bulb temperature Tw, and T is the dry bulb temperature. With the aid of the saturation vapour pressure-temperature curve the vapour pressure over a finite gradient can be expressed as

$$\Delta e = (S + \gamma) \Delta Tw - \gamma \Delta T \tag{9}$$

where S is the slope of the saturation vapour pressure-temperature curve
at the mean wet bulb temperature between any two levels of measurement,
$\gamma = AP(1 + 0.00115\ Tw)$ where $(1 + 0.00115\ Tw)$ normally approximates
unity, and $D = T - Tw$ the wet bulb depression.

Substituting (9) into (7) gives

$$LE = (Rn - G)\left[1 - \left(\frac{\gamma}{S + \gamma}\right)\frac{\Delta T}{\Delta Tw}\right] \tag{10}$$

Dilley has shown that calculated values for S are accurate to within 0.1
percent of the values observed by the formulae developed by Goff and
Gratch (1946) over a temperature range of 0 - 50°C. This is accomplished
by differentiating Tetens (1930) equation

$$e_s = 6.1078\ \exp\frac{17.269\ Tw}{Tw + 237.3} \tag{11}$$

with respect to temperature

$$S = \frac{de_s}{dTw} = \frac{25.029}{(Tw + 237.3)^2}\ \exp\frac{17.269\ Tw}{Tw + 237.3} \tag{12}$$

The error in estimating S by the above equations decreases in magnitude
as ΔTw decreases.

The determination of LE using (10) requires the measurement of
Rn and G at the surface, and T and Tw for at least two levels. No

knowledge of the surface characteristics, such as wetness, roughness or any other parameter that has to be calculated or estimated, is required. It is essential, however, to provide an adequate fetch to avoid advection affects. This insures that temperature and humidity profile measurements are maintained within the boundary layer thereby yielding representative values for the surface fluxes.

2. The Equilibrium Model

The equilibrium model represents a particular case of the combination energy balance-aerodynamic approach to estimating the evaporative heat flux. A general expression defining the combination method was first presented by Penman (1948). McIlroy's more recent derivation (Slatyer and McIlroy, 1961) will be presented in this discussion.

Slatyer and McIlroy express the combination model as

$$LE = \frac{S}{S + \gamma} (Rn - G) + h(D_z - D_o) \tag{13}$$

where D_z and D_o are the wet bulb depressions at a height z above the surface and at the height z_o at the surface, and h is a transfer co-efficient dependant on windspeed which operates between z_o and z, and varies somewhat with roughness and atmospheric stability.

Since the combination method combines the energy balance and aerodynamic approaches, it is subject to the assumptions of each. The mass transfer equations (4) and (5) can be rewritten as

$$H = h \Delta T \tag{14}$$

$$\text{and} \qquad LE \quad = \quad \frac{L\varepsilon}{PCp} \, h \, \Delta e \qquad\qquad (15)$$

where Δ is the gradient with respect to height, and

$$h \quad = \quad \frac{Cp\rho K_A}{z} \quad = \quad \frac{\rho Cp}{r_a} \qquad\qquad (16)$$

where $K_A = K_W = K_H$, and r_a is the aerodynamic resistance which is defined as the time taken for 1 cm^3 of air to exchange heat with 1 cm^2 of surface.

Equations (14) and (15) can be equated in terms of Tw and D using (9) to give

$$LE \quad = \quad \frac{h}{\gamma} \, (S\Delta Tw - \gamma \Delta D) \quad = \quad \frac{hS}{\gamma} \, \Delta Tw - h \, \Delta D \qquad\qquad (17)$$

$$\text{and} \qquad H \quad = \quad h(\Delta Tw + \Delta D) \qquad\qquad (18)$$

Substituting (17) and (18) into (1) gives

$$h \, \Delta Tw \quad = \quad (Rn - G) \, (1 + \frac{S}{\gamma}) \qquad\qquad (19)$$

By substituting (19) into (17) the latent heat flux becomes

$$LE \quad = \quad \frac{S}{S + \gamma} \, (Rn - G) + h \, \Delta D \qquad\qquad (20)$$

where $\Delta D = D_z - D_o$.

The practical use of (20) is limited by the difficulty of measuring D_0. The model is, however, informative, since it separates the basic energy sources utilized in evapotranspiration. The terms represent the contributions to the evaporative process made by the available radiant energy and atmospheric influences respectively. The atmospheric term, h ΔD, is the main factor contributing to differences in LE between surfaces of differing wetness. In the case where the air in proximity to an effectively moist surface becomes saturated $D_0 = 0$, and LE proceeds at the potential rate. When the water supply is restricted, however, D_0 attains a finite value and LE is less than the potential rate. Slatyer and McIlroy (1961) introduced the concept of equilibrium evapotranspiration by taking the limited case $D_z = D_0$. In this instance (20) reduces to

$$LE = LE_{EQ} = \frac{S}{S + \gamma} (Rn - G) \qquad (21)$$

where LE_{EQ}, the equilibrium evapotranspiration, can be calculated as a function of temperature and available radiant energy.

Various interpretations of the physical meaning of (21) have been made. Monteith (1965) and Tanner and Fuchs (1968) describe the simplest case where $D_z = D_0 = 0$. This is the evaporative flux which would occur in a saturated atmosphere. Slatyer and McIlroy (1961) and Denmead and McIlroy (1970) interpret LE_{EQ} as the lower limit of potential LE, with the former authors considering that $D_z = D_0$ points to a state of mutual adjustment between the surface and overlying air. The physical meaning for the purposes of this paper is similar to that put forth by Wilson

and Rouse (1972) who postulate that the depressions (D_z and D_o) may have finite values which can be equal or nearly equal. In either case the use of (21) can be considered as a valid indicator of actual LE.

CHAPTER III

SITE AND EXPERIMENTAL METHOD

1. <u>Site</u>

The research was conducted during July and August 1971 near the

Hudson Bay coastline adjacent to East Pen Island in northern Ontario

(lat. 57° 45' N; long. 88° 45' W). The exact location is shown in Fig. 1.

Observations were made on the predominantly flat raised beach ridge

illustrated in Plate 1, located 10-15 mi. north of the treeline and

approximately 1 mi. inland from the coast.

The vegetation in the vicinity of the site was typical of tundra

vegetation, comprising lichens, flowering plants, mosses and a few stunted

shrubs on the beach ridges, with intermittent lakes and muskeg between

the ridges. The beach ridge illustrated in Plate 1 was covered by a

lichen and flowering plant vegetation varying from 3-7 cm in mat thick-

ness. The main species of lichen were <u>Cetraria islandica</u>, <u>Cetraria</u>

<u>nivalis</u>, <u>Cetraria cuculata</u>, <u>Alectoria ochroleuca</u>; <u>Pryas integrifolia</u>, and

<u>Rhododenron lapponicium</u> were the main flowering plants. Although vegeta-

tion species and mat thickness varied from ridge to ridge lichen vegeta-

tion was the typical cover of the ridges.

The vegetation cover was underlain by a sandy soil varying from

fine to coarse sand. The sands were very porous to rainwater and no

surface drainage was evident.

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FIGURE 1

LOCATION OF RESEARCH
AND MODEL TESTING SITES
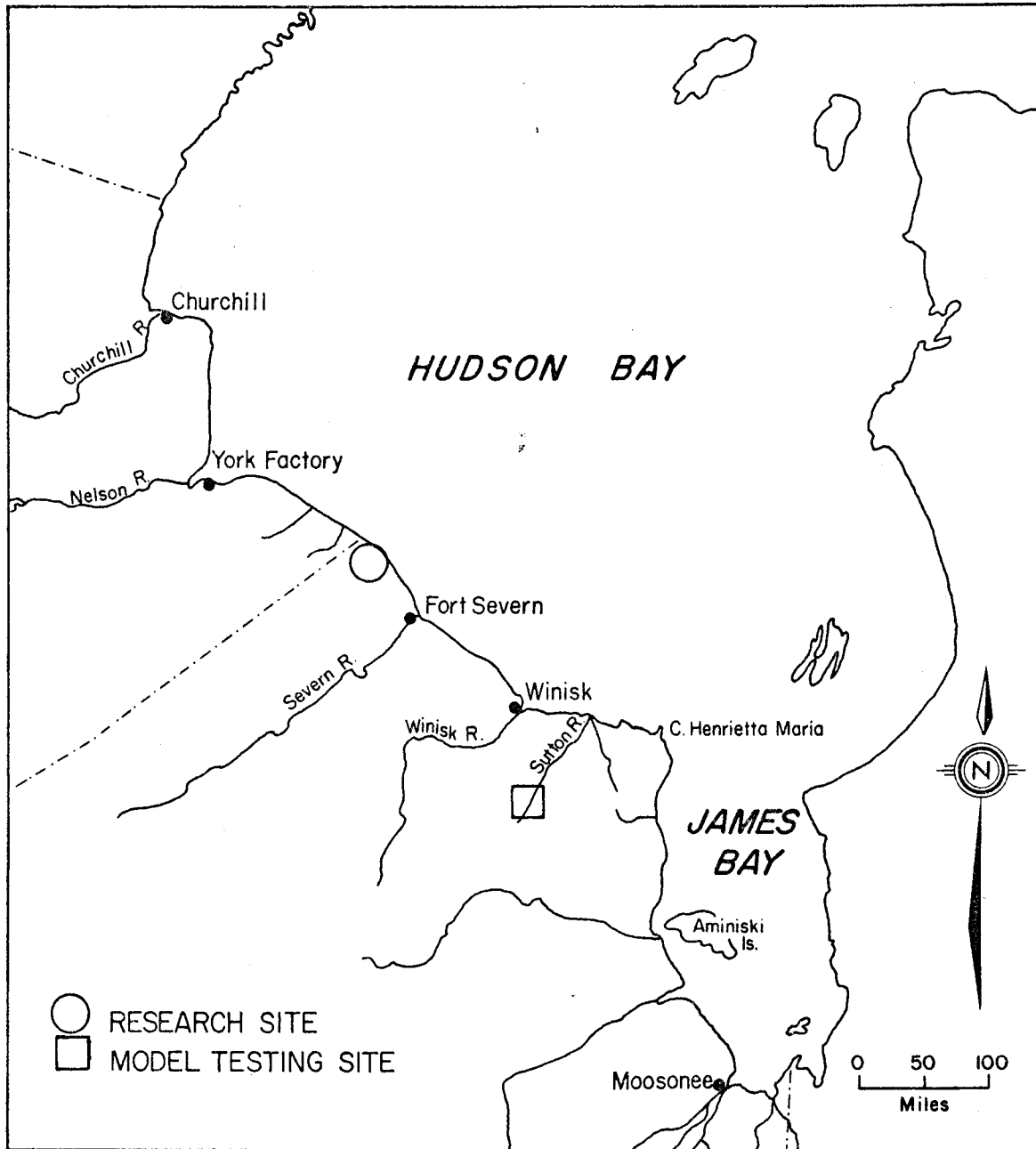
2. Energy Balance Measurements

Evaluation of the energy balance using (10) requires the measurement of net radiation, soil heat flux and, dry and wet bulb vertical temperature gradients above the surface.

a) Net Radiation

Net radiation was measured using a net radiometer (Swissteco, Type S-1) mounted at 0.5 m above the surface. Polyethylene domes, which protected the thermopile surfaces, were inflated by the flow of dessicated air from an aquarium pump. Net radiation signals were continuously recorded on a millivolt strip chart recorder, and integrated hourly values determined by planimetering the pen trace.

b) Soil Heat Flux

The soil heat flux, G, at the surface was calculated from the equation

$$G = G_5 + C \frac{\overline{\Delta T_s}}{\Delta t} \Delta z \qquad (22)$$

where G_5 is the soil heat flux measured at 5 cm below the surface, C is the soil heat capacity between the surface and 5 cm, $\overline{T_s}$ is the mean soil temperature between the surface and 5 cm, t is the time and Δz is the soil layer from 0-5 cm depth.

The soil heat flux at 5 cm (G_5) was measured with three soil heat flux plates (Middleton and Pty. Ltd.) connected in series giving a signal that was continuously recorded on a strip chart recorder. Hourly totals were calculated by planimetering the pen trace.

PLATE 1

View of the Site at Pen Island

Changes in the mean soil temperature $(\overline{\Delta T_s})$ in the 0-5 cm layer were monitored bi-hourly, at 2 sites about a meter apart, with ice referenced thermocouple units. Thermocouple junctions were located at 0 cm (the soil surface), 2 and 5 cm. The temperatures of the two sites were averaged to determine the mean temperature of the 0-5 cm layer.

The soil heat capacity was estimated from the equation by De Vries (1963)

$$C = 0.46 \; X_M + 0.60 \; X_0 + \theta \qquad (23)$$

where X_M, X_0 and θ are the volumetric fractions of mineral matter, organic matter and water respectively. X_M and X_0 were found to be 0.98 and 0.02 respectively from "loss on ignition" treatments of surface samples. These results allowed (23) to be reduced to

$$C = 0.463 + \theta \qquad (24)$$

The average soil moisture content was estimated on a daily basis with the use of a surface neutron probe (Nuclear Chicago 5901).

c) <u>Temperature and Wet Bulb Depression Gradients</u>

Wet and dry bulb air temperatures were measured with 5-junction thermopiles similar to those described by Rouse and Kershaw (1971) and Wilson (1971). The junctions, enclosed in an aluminum sleeve, were constructed from Standard 36 gauge copper constantan wire. Protection and rigidity was enhanced by enclosure of the wires in a stainless steel shaft attached to the aluminum sleeve. When calibrated against platinum

resistors, two thermopile units gave an output of 203 uV C^{-1}. It was assumed that the remaining sensors performed similarly, since side by side testing of all the thermopiles gave perfect agreement.

The 3-level Bowen ratio mast illustrated in Plate 2 provided for the measurement of wet and dry bulb temperatures at each level. The sensors, located 75 cm apart at each level, were contained within a double walled, aspirated radiation shield in order to eliminate radiation and other local heating errors. Each sensor was referenced to an ice bath.

The wet-bulb sensors were enclosed in a tight-fitting muslin wick which extended from the base of the sensor through plastic tubing to a water reservoir. The rate of water feed to the thermopile could be readily adjusted by varying the height of the reservoir relative to the sensor head.

There were no fetch limitations at the site as illustrated by the near infinite homogeneous flat surface shown in Plate 1. Temperatures monitored at heights of 25, 50 and 75 cm above the surface were well within the atmospheric boundary layer.

All temperature signals were stepped through a double 3-channel stepping switch and recorded on a 2 pen millivolt recorder. This allowed the wet and dry bulb temperatures at a given level to be recorded simultaneously. The speed of the stepping switch was adjusted to give 30 signals per hour for each sensor. These were averaged to give the hourly mean temperature.

The hourly values of the Bowen ratio and evaporative and sensible heat fluxes, were computed by a CDC 6400 computer for the levels: 25-50 cm, 50-75 cm and, 25-75 cm. Three level evaluation ensured that values
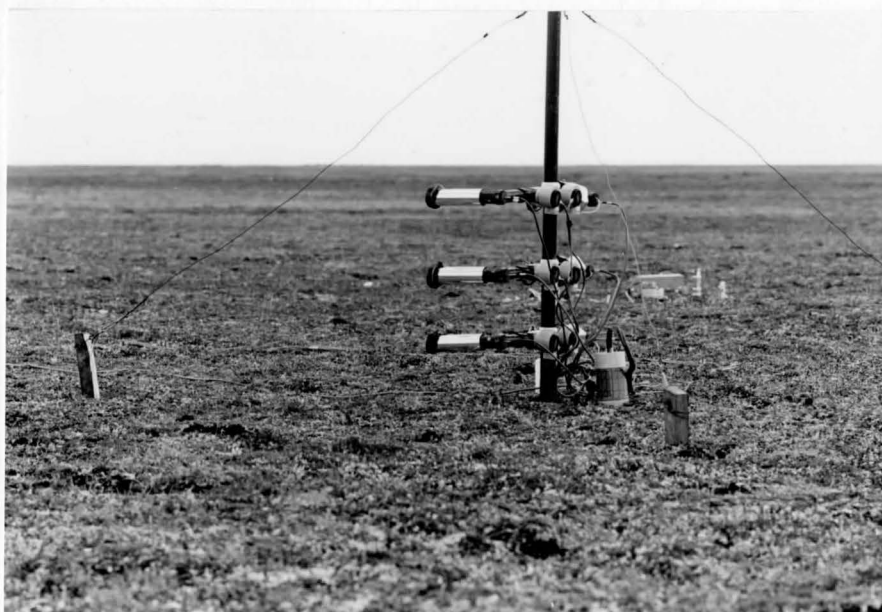
PLATE 2

3-Level Bowen Ratio Mast

could still be determined in the event of a sensor failure at one level.

In addition to measurements required for evaluation of the energy balance, wind speed and direction, precipitation, and maximum temperatures were recorded.

The energy balance estimates of LE were utilized as control data for evaluation of the equilibrium model.

3.   Measurements and Calculations for the Equilibrium Model

The same measurements made for the energy balance were utilized in evaluating the equilibrium evaporation as developed in (12) and (21). The hourly equilibrium evaporation was computed by the same computer program used to calculate the energy balance.

CHAPTER IV

ENERGY BALANCE COMPONENTS

AND EQUILIBRIUM EVAPORATION

1. Energy Balance

Daily variations of the energy balance components, each presented

as a percentage of Rn during clear sky conditions, are shown in Fig. 2a.

With a few exceptions, it can be seen that the components remain fairly

constant throughout the study. An average of 54, 39 and 7 percent of Rn

was utilized in the fluxes of LE, H and G respectively.

Average daily Bowen ratios are plotted in Fig. 2b. The mean value

for β was 0.731 for the 19 days of measurement with a maximum and minimum

value ranging between 1.778 on Aug. 22 and 0.389 on Aug. 3. Comparing

Fig's. 2a and 2b it can be seen that any significant deviation from the

mean value of the components in Fig. 2a show up as marked fluctuations

of the Bowen ratio from its mean. In particular, this is exemplified on

Aug. 3, 4, 22 and 23. The generally high β values indicate that the lichen

surface is highly resistant to evaporation with the sensible heat compon-

ent comprising a significant proportion of the available radiant energy.

The low β value for Aug. 3 and 4 are a result of a heavy rainfall

of 2.75 cm during two of the three preceding days (July 31 and Aug. 1).

From Table 1 it can be seen that the surface moisture content increased

significantly which implies that the available moisture for evaporation

TABLE 1

DAILY METEOROLOGICAL AND SURFACE SOIL MOISTURE DATA

July

Date	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
Maximum Temperature (°C)						10.1	19.0	17.2	4.4	12.2	20.7	15.0	13.3	5.6	4.4	4.4
Wind Direction													NW	N	NW	NW
Precipitation (cm)								0.1	1.0	0.2			0.2			0.8
Surface Soil Moisture (% Vol.)					0.19	0.19	0.16	0.17	0.19	0.20	0.16		0.16		0.14	

Date	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	
Maximum Temperature (°C)	4.6	6.7	7.8	11.2	7.3	6.1	10.6	15.0	4.4	10.0	13.4	7.3	13.9	19.0	14.0	
Wind Direction	NW	NNW	NNW	N	NNE	NNE	W	NW	N	NNW	W	N	NNW	SW	NNW	
Precipitation (cm)		0.1		0.1	0.9	0.3	0.5				0.3			T	2.2	
Surface Soil Moisture (% Vol.)	0.17	0.15	0.14		0.15			0.14	0.14	0.14			0.13			

(cont'd)

TABLE 1

DAILY METEOROLOGICAL AND SURFACE SOIL MOISTURE DATA

August

Date	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
Maximum Temperature (°C)	5.4	8.4	22.9	26.7	21.2	13.9	16.7	6.5	6.7	6.2	10.1	12.3	9.5	19.5	12.3	6.7
Wind Direction	N	N	SW	SW	WNW	NNW	NNW	NNE	NNE	E	NNW	NNW	NW	W	NW	NE
Precipitation (cm)	0.6						0.4	0.7		1.4						
Surface Soil Moisture (% Vol.)	0.14	0.14	0.14	0.13	0.13	0.13	0.13				0.14		0.13	0.12		

Date	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	
Maximum Temperature (°C)	8.9	8.4	10.6	7.3	6.1	7.8	11.2	18.0	18.4	21.2						
Wind Direction	ESE	ESE	NW	NNW	NNW	NNE	S	S	S	SW						
Precipitation (cm)	0.1	0.3	0.1	0.2	0.2											
Surface Soil Moisture (% Vol.)	0.13															

FIGURE 2

VARIATION OF THE ENERGY BALANCE COMPONENTS
AND BOWEN RATIO ON SAMPLE DAYS
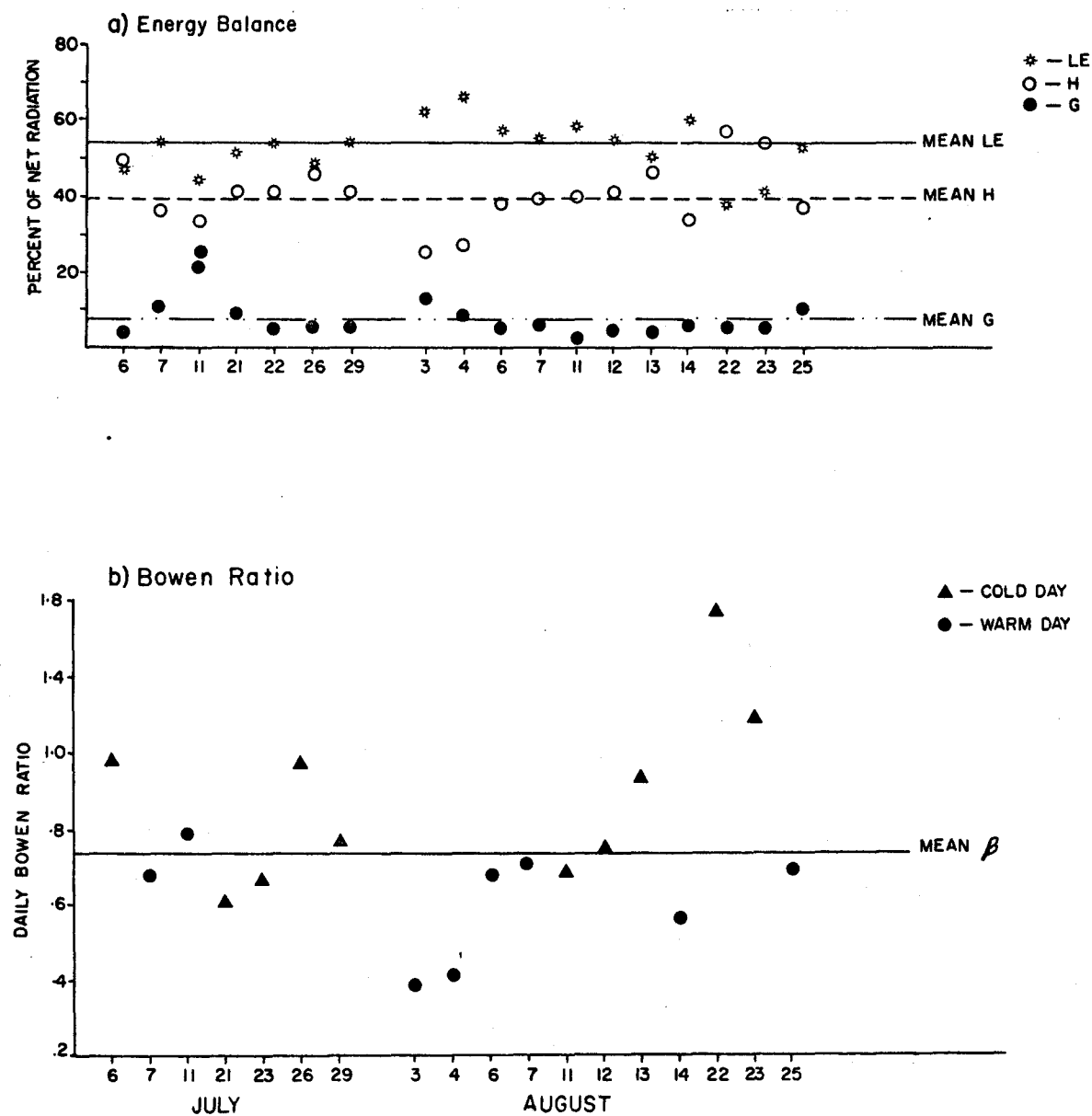
increased. This, in combination with the influence of high temperatures (discussed later) accounts for the low Bowen ratio values. It is apparent from Table 1 that the high Bowen ratio values of Aug. 22 and 23, were not due to a restriction on available water. Although soil moisture measurements are not available a rainfall of 0.93 cm in the period Aug. 16-21, insured moist surface conditions. It would appear, therefore, that cold atmospheric temperatures were responsible for the high β values. The greater sensible heat flux is promoted by the utilization of more radiant energy in the sensible heating of the cooler atmosphere.

The variations in the energy balance components noted above imply a temperature control of evaporation. To test this hypothesis the data were subdivided according to the temperature conditions observed over the study period.

2.   The Energy Balance of Warm and Cold Days

Table 1 shows that the maximum temperature varied significantly with changing wind direction. The winds can be divided into two primary directional groups: northerly winds, which include all winds from west to north to east; southerly winds which include those winds from west to south to east. A marked difference in the maximum temperature range is observed, varying from 4.4-19.5 C for northerly and 10.6-26.7 C for southerly winds. The data were divided, therefore, into warm and cold days. Warm days are defined as any day that has a maximum temperature greater than 15 C, whereas, cold days are those with a maximum less than or equal to 15 C.

Figure 2b shows the daily variations in the Bowen ratio for warm

FIGURE 3

ENERGY BALANCE COMPONENTS
FOR WARM AND COLD DAYS
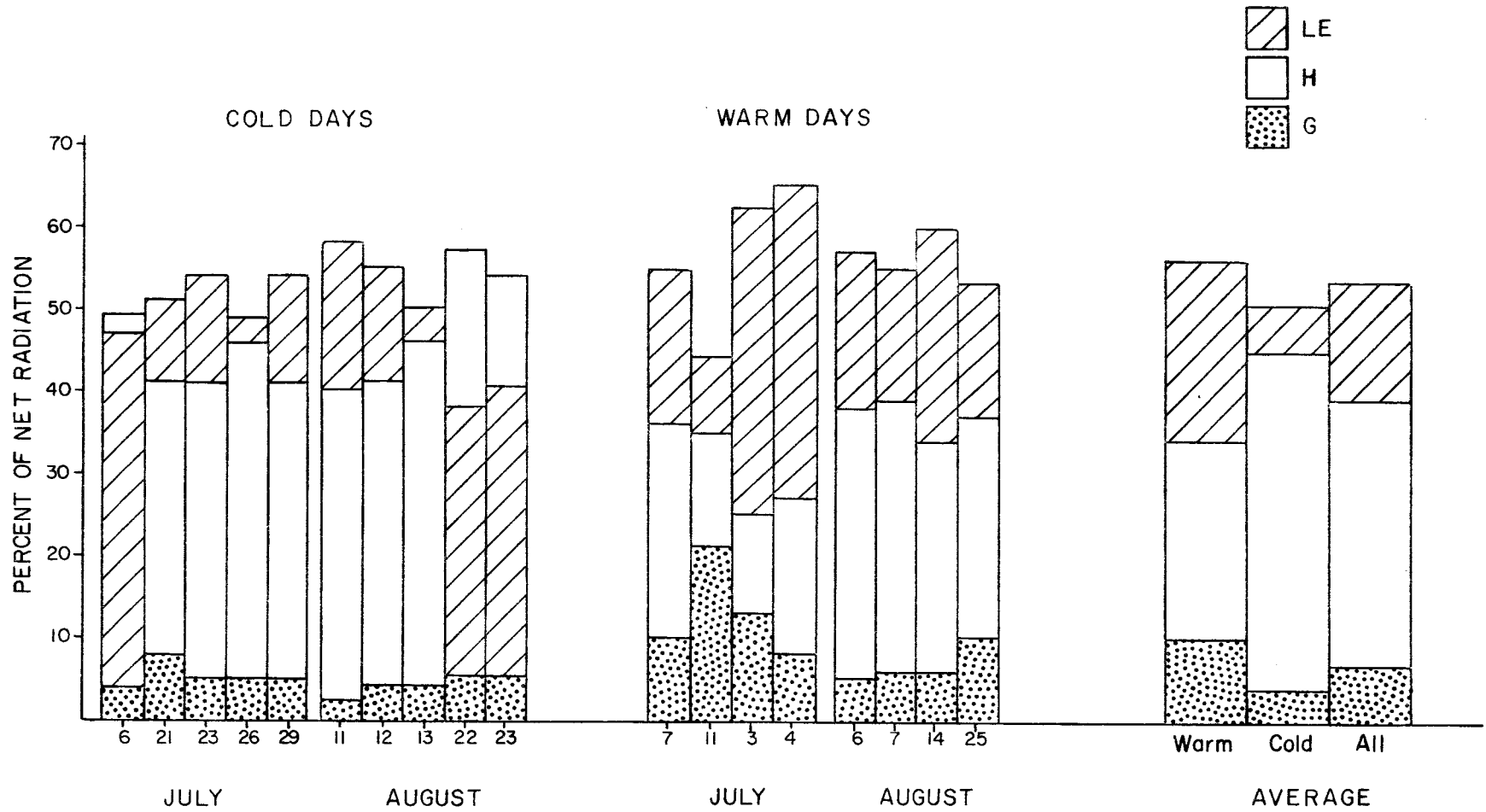
and cold days. The β values on cold days are higher averaging 0.886 as compared to 0.610 for the warm days. This shows that the energy balance components H and LE are temperature dependant.

Fig. 3 shows the energy balance components for warm and cold days. For individual days LE does not show much difference between warm and cold days. The sensible heating, however, is obviously larger on cold days. This is readily apparent when the daily values are averaged for the season. LE on warm days exceeds the cold day value by only 5 percent, whereas, H is 11 percent lower on the warm days. As H grows larger on the cold days, G grows small. The opposite effect is observed for warm days. With a warm atmosphere more energy is utilized in G, since the soil temperature gradient, $\Delta T_s/\Delta z$, is larger relative to the atmospheric temperature gradient, $\Delta T/\Delta z$, than for a colder atmosphere.

3. Equilibrium Evaporation

Tests were applied to both hourly and daily data in order to compare equilibrium estimates of evaporation to those calculated for the energy budget.

The 148 hourly values examined pertain exclusively to a non-saturated atmosphere. Fig. 4 shows the energy balance calculations of LE plotted as a function of equilibrium estimates. The equation

$$LE = -0.008 + 0.955\ LE_{EQ} \tag{25}$$

$$r = 0.953 \text{ and } S_y = 0.034 \text{ cal cm}^{-2}\text{min}^{-1}$$

shows that LE_{EQ} consistantly overestimates LE by about 5 percent. S_y is

FIGURE 4

SEASONAL COMPARISON OF AVERAGE HOURLY
LE TO LE_{EQ}

$LE = -0.008 + 0.955\ LE_{EQ}$

$r = 0.953 \quad Sy = 0.034\ (cal\ min^{-1})$

LE (cal cm^{-2}min^{-1})

LE_{EQ} (cal cm^{-2}min^{-1})
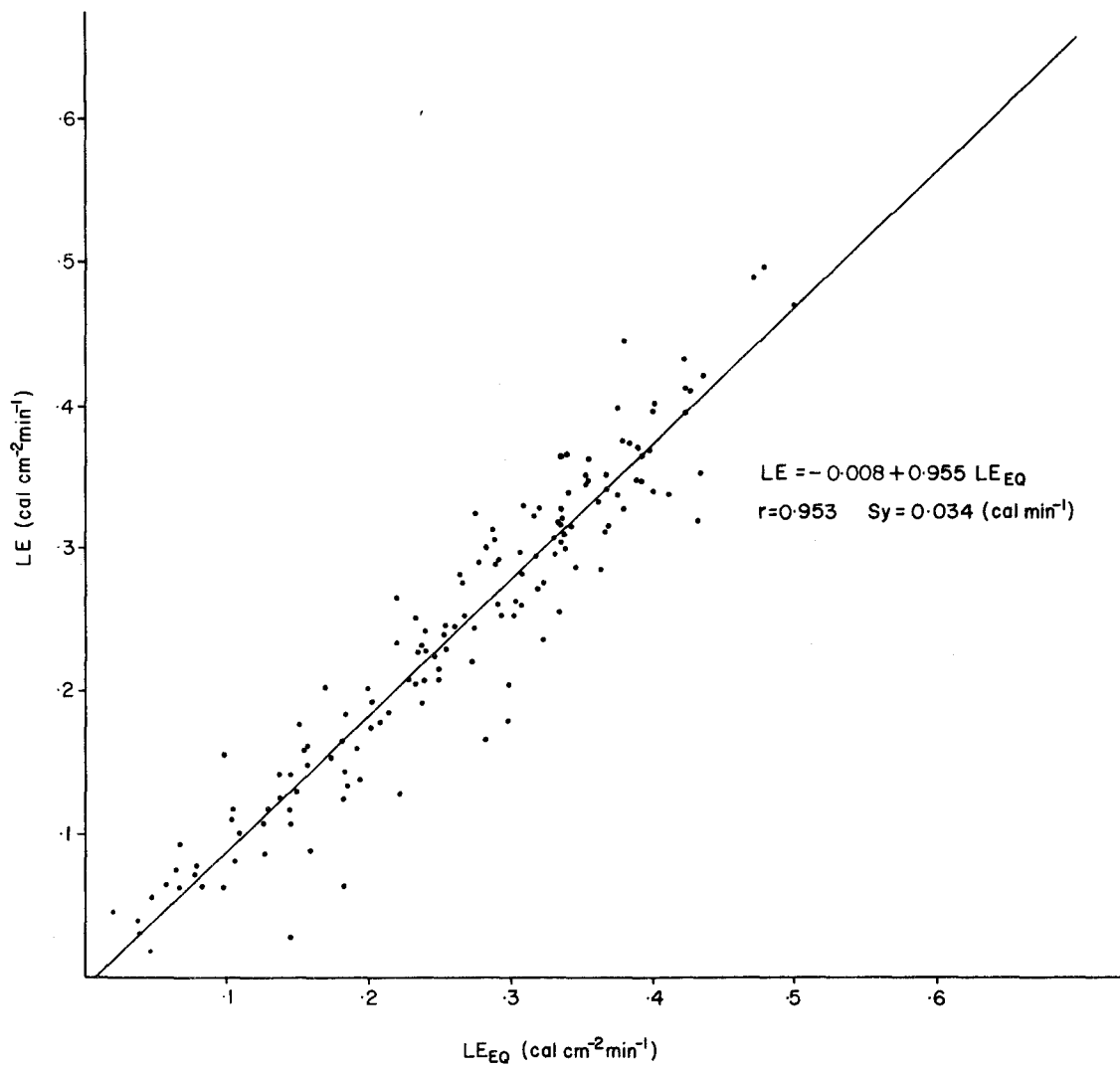
FIGURE 5
COMPARISON OF DAILY LE TO LE$_{EQ}$

$LE = -1.172 + 0.929$ LE_{EQ}

$r = 0.977$ $Sy = 10.565$ cal. cm^{-2} day^{-1}

LE (cal. cm^{-2} day^{-1})

LE$_{EQ}$ (cal. cm^{-2} day^{-1})
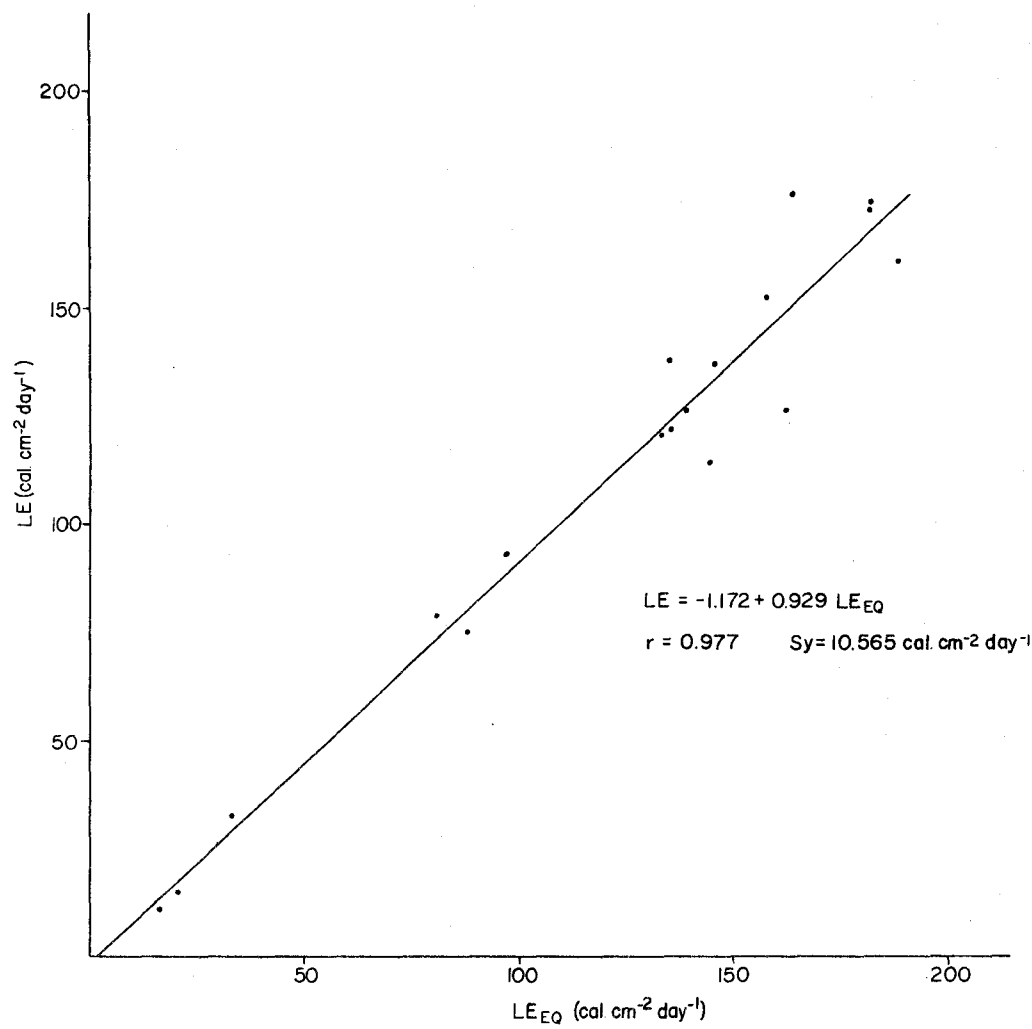
the standard error.

Daily totals of LE versus LE_{EQ} are shown in Fig. 5. The daily totals of LE are consistently overestimated by approximately 8 percent. The high correlation coefficient and low standard error indicate a very reliable relationship.

The consistency with which LE_{EQ} overestimates LE implies that actual evaporation can be determined as some function of the equilibrium model.

4.   Conditions in which the Equilibrium Model Applied

In theory the equilibrium model should apply when the second term on the right hand side of (13) tends toward zero. Three conditions which can promote this are when $D_z = D_o$ or is nearly equal; when r_a is large; when both of the above conditions occur together. In this study the good agreement between LE and LE_{EQ} is attributed to the second condition, a high diffusive resistance to evaporation. One reason for this can be attributed to negative differences in the wet-bulb depressions monitored between the 25 and 75 cm heights averaging 0.6 C and varying between 0.2 and 1.0 C. On the assumption that ΔD_{25-75} indicates ΔD between the surface and overlying atmosphere, the above observed conditions imply the existence of substantial wet-bulb depression gradients. From (13) and (16) it is evident that with a large ΔD the resistance (r_a) must be large in order to reduce the second term on the right of the equation toward zero. In addition, the high Bowen Ratios indicate the existence of a substantial diffusive resistance to evaporation.

The aerodynamic resistance by definition refers to the resistance

to diffusion of water vapour from the effective evaporating surface into the air. In the case of a non-transpiring lichen vegetation, the effective surface is not at the normally dry canopy top, but at the moisture source near the canopy base. Since r_a is proportional to the inverse of the wind speed (Monteith, 1965), r_a becomes large at the base of the lichen canopy where the windspeed approaches zero. When this occurs $(\rho c_p/r_a)(D_z - D_o)$ tends toward zero.

The reason for LE_{EQ} exceeding LE consistently can be explained by the fact that the wet-bulb depressions observed over the lichen canopy in all cases were found to decrease with height between the 25 and 75 cm heights. If by analogy D_{25-75} can be assumed indicative of D_{0-25}, then $D_o > D_z$. In this case combined with a large r_a the value of LE_{EQ} will exceed LE as determined by (20).

The fact that $D_o > D_z$ does not necessarily mean that the specific humidity of air at height z is more than at height 0 because as long as evaporation from the lichen surface is taking place it must be less. It does mean, however, that although the latent heat flux can still be large (> H for example), the dry bulb temperature decreases more rapidly with height than the wet-bulb temperature. This indicates the existence of a high diffusive resistance to the water loss from the lichen mat.

5. Significance of Results

The excellent relationship between LE_{EQ} and LE implies that a simple model based on the concept of the equilibrium model can be utilized to estimate the actual evaporation as a function of temperature and radiant energy.

CHAPTER V

DERIVATION AND TEST

OF A SIMPLE EVAPORATION MODEL

1. Derivation of the Evaporation Model

The LE_{EQ} computation of evaporation as expressed in (25) and

Fig. 4 explains 91 percent of the actual LE for hourly values. This

gives exceptionally good agreement when one considers the wide range of

temperature and moisture conditions over which LE_{EQ} was estimated. By

substituting (21) into (25), the acutal evaporation can be expressed as

$$LE = -0.008 + 0.955 \frac{S}{S + \gamma} (Rn - G) \qquad (26)$$

where LE is now expressed in terms of radiant energy and a temperature

dependant term.

Further simplification of (26) was shown in the work of Wilson

and Rouse (1972). They showed that $S/(S + \gamma)$ can be calculated as a

linear function of temperature with a high degree of accuracy over the

temperature range of 17-32 C, nearly equivalent to the "warm day"

classification of this study. In similar fashion, and with inclusion of

the "cold day" temperature range, the determination of $S/(S + \gamma)$ as a

linear function of hourly mean temperatures gave

$$\frac{S}{S + \gamma} = 0.434 + 0.012 \, T \tag{27}$$

for the temperature range 6.6–27.7 C. The correlation coefficient of 0.998 and standard error of 0.003 C^{-1} show the reliability of this relationship.

Equation (27) is very similar in magnitude to that calculated by Wilson and Rouse, the difference in this case being attributed to the incorporation of a larger temperature range in the evaluation of the $S/(S + \gamma)$ function.

Substituting (27) into (26) gives

$$LE = LE_M = (0.406 + 0.011 \, T)(Rn - G) \tag{28}$$

where LE_M is the prediction model by which actual LE can be determined (cal $cm^{-2}min^{-1}$). Equation (28) can readily be converted for use in terms of hourly and daily totals.

2. A Test of the Evaporation Model

To test the validity of (28) as a general model for estimating evaporation for resistant subarctic surfaces, values of measured LE were obtained from another site at a similar latitude for comparison to corresponding LE_M values.

Data were obtained from Rouse and Kershaw (1971), who made energy budget measurements at Hawley Lake, a site approximately 200 miles ESE of Pen Island. The two surface types which they considered were a dense

FIGURE 6

HOURLY RELATIONSHIP BETWEEN
LE AND LE$_M$ FOR HAWLEY LAKE

$LE = 0.004 + 1.015 LE_M$

$r = 0.967 \quad Sy = 0.033$ cal cm^{-2}min^{-1}

LE (cal cm^{-2}min^{-1})

LE$_M$ (cal cm^{-2}min^{-1})
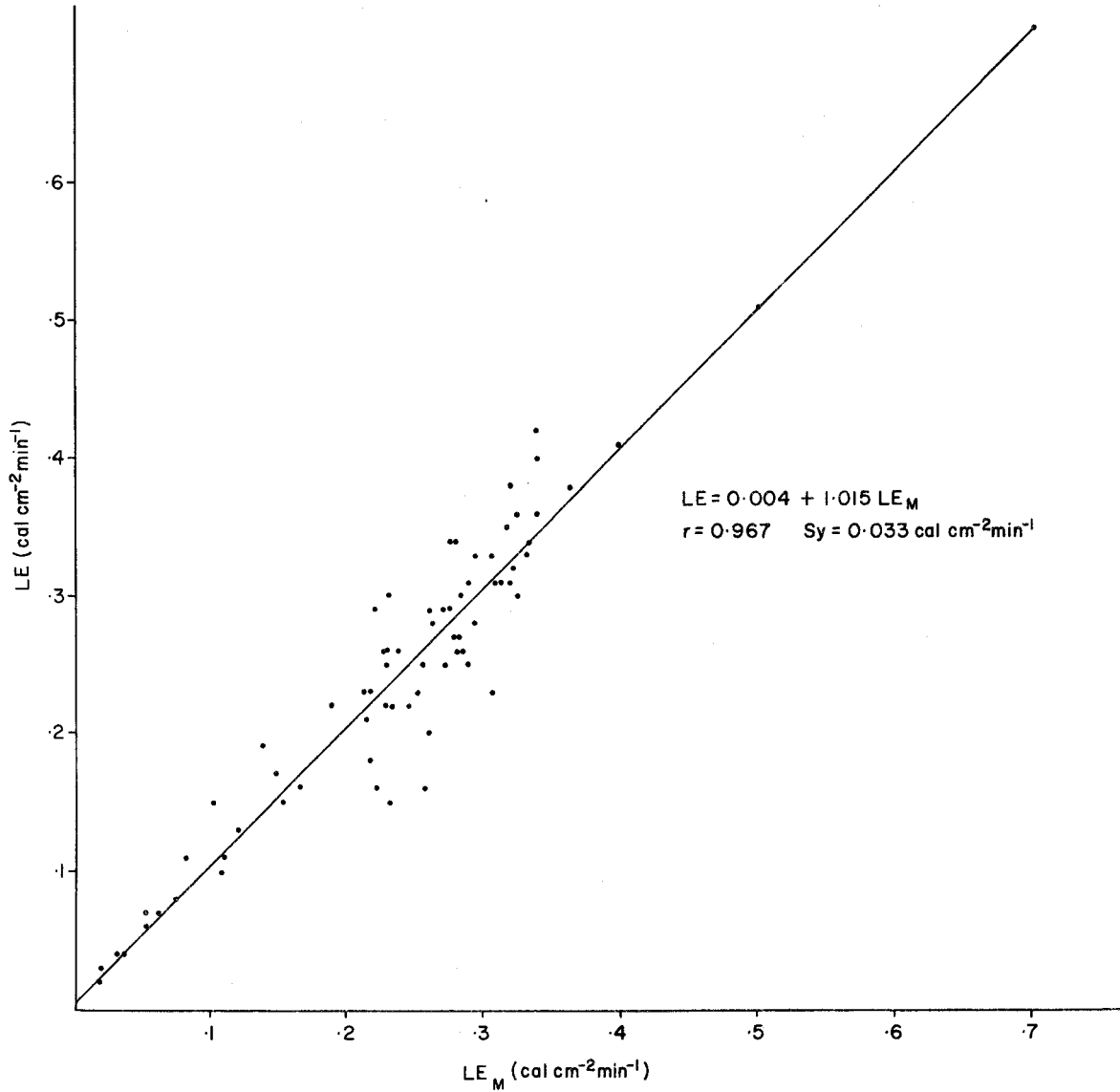
FIGURE 7

DAILY TOTALS RELATIONSHIP BETWEEN
LE AND LE_M FOR HAWLEY LAKE

$LE = 4.748 + 0.986 LE_M$

$r = 0.978$ $Sy = 10.043 \, cal \, cm^{-2} day^{-1}$
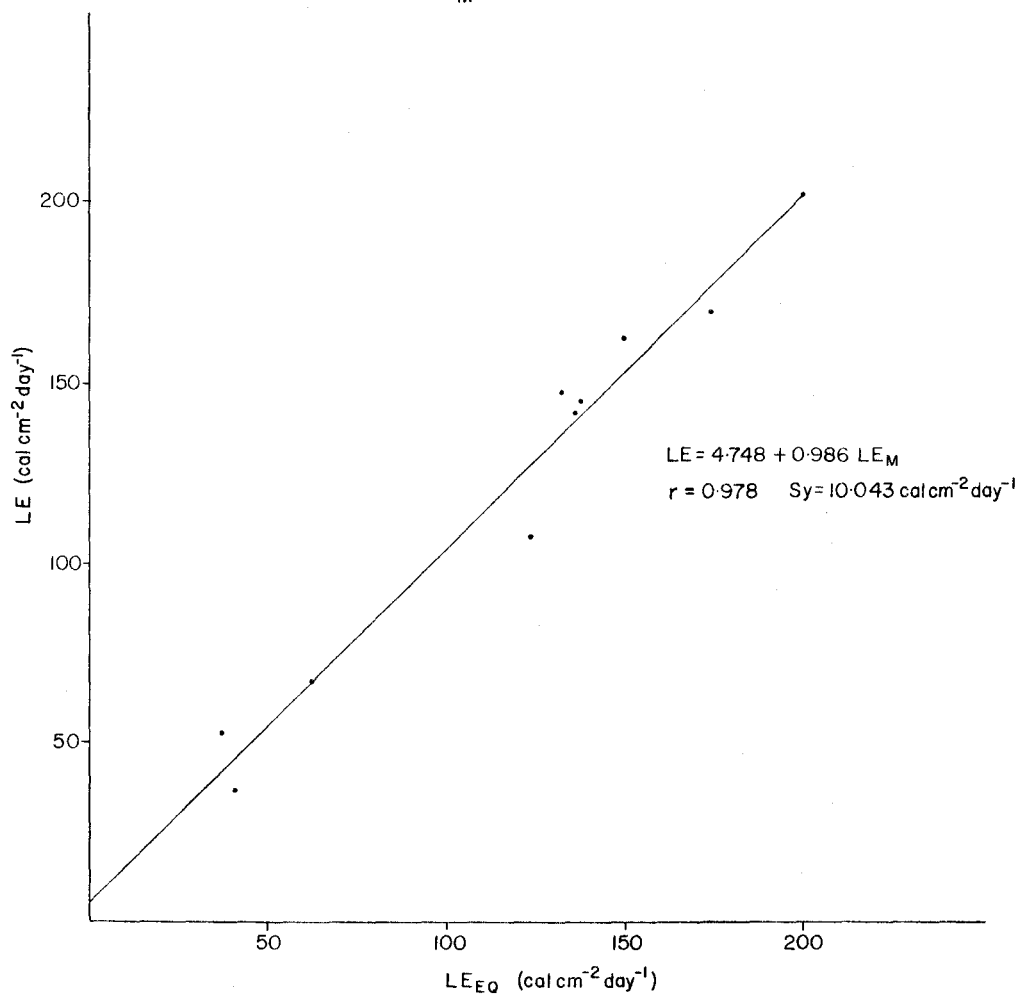
natural lichen surface composed almost entirely of <u>Cladonia alpestris</u>

averaging 11 cm in thickness, and a previously burned lichen surface that

was just beginning to be revegetated but was still covered with the ash

of the burned lichen. Temperature and available radiant energy conditions

varied from -0.3 to 24.0 C and -0.012 to 0.690 cal $cm^{-2}min^{-1}$ respectively.

Fig. 6 shows the relationship between hourly predicted values of

LE_M and measured values for both surfaces. The excellent agreement of

the relationship yielded a correlation coefficient of 0.967 with a

standard error of 0.033 cal $cm^{-2}min^{-1}$ for the 82 hours of measurement.

For daily totals the agreement was even more significant as seen in Fig.

7. A correlation coefficient of 0.978 and standard error of 10.043 cal

$cm^{-2}day^{-1}$ was obtained for the 9 days of comparison.

These results indicate that the generalized evaporation model

expressed in (28) is very reliable in estimating the hourly and daily

totals of evaporation from subarctic surfaces which are fairly resistant

to evaporation.

3. <u>Significance of Results</u>

The excellent agreement of LE estimates for the lichen-covered

and burned surfaces as shown in Fig's. 4, 6 and 7 is of significant

importance. Previous discussion has revealed that the diffusive resist-

ance to evaporation was the prime reason for the equilibrium model apply-

ing in the subarctic. The successful testing of the model on the various

lichen and burned sites implies that the resistances are of comparable

magnitudes, further suggesting that all subarctic upland sites exhibit

similar diffusive resistances to evaporation. It is apparent from the

above that LE_M should be applicable over any subarctic upland land surface in estimating evaporation.

CHAPTER VI

SUMMARY AND CONCLUSION

The energy balance method was used to estimate the hourly and daily totals of evaporation for clear sky conditions. All observations were made on a lichen covered raised beach ridge adjacent to East Pen Island in northern Ontario. The lichen surface was found to be relatively resistant to evaporation with only 54 percent of the daily net radiation being utilized in the evaporative process. The sensible heat component comprised a significant proportion of the net radiation, with daily Bowen ratios averaging 0.731. Fluctuations in β with temperature had little effect on evaporation over the temperature range 6.6-26.7 C. The warm and cold day fluctuations in Bowen Ratios was the result of changing proportions of energy being used in the sensible and soil heat fluxes.

Hourly and daily evaporation estimates of the equilibrium model were compared to corresponding energy balance values in order to test the hypothesis that the equilibrium model or some function of the equilibrium model could be used to estimate accurately the evaporation from an upland lichen surface. Hourly and daily values of evaporation determined by the equilibrium model were found to consistently overestimate the actual evaporation by 5 and 8 percent respectively. An evaporation model, a function of the equilibrium model, was derived from the relation-

ship to estimate the actual evaporation. The derived model explains 91 and 95 percent of the actual evaporation at the Pen Island site for hourly and daily totals respectively. Further tests of the model for a natural lichen and for a burned surface at Hawley Lake gave excellent results when compared to actual evaporation, predicting both hourly and daily totals within 5 percent.

In conclusion, the results of this study show that hourly and daily totals of evaporation can be estimated with a 5 percent accuracy for an upland lichen surface in the subarctic and can probably be estimated with excellent reliability for any subarctic upland surface exhibiting a relatively high resistance to evaporation. The model now needs to be tested on a variety of high latitude surfaces in order to define its overall applicability.

APPENDIX A

ENERGY BALANCE AND

EQUILIBRIUM DATA FOR PEN ISLAND

The following symbols are used:

ID $=$ time period (month-day-hour)

T $=$ air temperature used in determining $S/(S + \gamma)$ (°C)

LE_{EQ} $=$ evaporation from the Equilibrium model (cal $cm^{-2}min^{-1}$)

LE $=$ evaporation from the Bowen ratio method (cal $cm^{-2}min^{-1}$)

H $=$ sensible heat flux (cal $cm^{-2}min^{-1}$)

ΔT_D $=$ mean hourly dry-bulb temperature gradient (°C)

Δe $=$ mean hourly vapour pressure gradient (mb)

β $=$ mean hourly Bowen ratio (dimensionless)

Hours are listed on the 24-hour clock and represents the period of measurement i.e., ID 071612 represents measurements from 1100-1200 on July 16.

The flux of any parameter is considered positive if directed towards the surface.

ID	T(°C)	Rn	LE_{EQ}	LE	H	G	ΔT_D	Δe	β
070614	10.2	0.740	0.398	0.370	0.343	0.027	0.684	0.488	0.925
070615	10.9	0.732	0.400	0.340	0.401	0.027	0.908	0.453	1.322
070616	11.1	0.661	0.363	0.285	0.351	0.025	0.743	0.398	1.232
070617	10.8	0.559	0.304	0.263	0.274	0.022	0.648	0.411	1.041
070708	16.8	0.256	0.146	0.028	0.197	0.031	1.809	0.169	7.049
070709	18.9	0.329	0.193	0.138	0.148	0.043	0.995	0.613	1.071
070710	20.4	0.440	0.266	0.276	0.109	0.055	0.774	1.298	0.394
070711	22.4	0.563	0.353	0.351	0.144	0.068	1.119	1.799	0.411
070712	25.1	0.627	0.412	0.338	0.218	0.071	2.867	2.937	0.644
070713	23.5	0.657	0.434	0.353	0.245	0.059	2.719	2.578	0.696
070714	22.7	0.650	0.432	0.319	0.294	0.047	2.684	1.987	0.892
070715	21.6	0.532	0.342	0.316	0.170	0.046	1.119	1.379	0.536
070716	20.7	0.456	0.291	0.292	0.126	0.038	0.773	1.186	0.430
070717	17.4	0.327	0.199	0.200	0.104	0.023	0.505	0.644	0.518
070718	15.1	0.148	0.083	0.064	0.069	0.015	0.174	0.107	1.072
071108	14.0	0.161	0.074	0.001	0.121	0.039	3.585	0.017	5.312
071109	17.8	0.348	0.183	0.064	0.213	0.071	4.206	0.834	3.330
071110	21.3	0.514	0.298	0.179	0.246	0.089	4.466	2.140	1.377
071111	24.4	0.559	0.334	0.255	0.199	0.105	4.238	3.572	0.783
071112	25.2	0.447	0.249	0.207	0.129	0.111	3.238	3.418	0.625
071113	23.9	0.442	0.246	0.224	0.113	0.105	1.311	1.719	0.503
071114	24.2	0.451	0.268	0.253	0.112	0.086	1.294	1.926	0.443
071115	22.2	0.276	0.149	0.132	0.078	0.066	1.512	1.690	0.591
071116	23.0	0.403	0.237	0.191	0.138	0.074	2.337	2.134	0.723

ID	T(°C)	Rn	LE$_{EQ}$	LE	H	G	ΔT_D	Δe	β
071117	21.7	0.472	0.274	0.244	0.144	0.084	1.724	1.920	0.593
071118	21.4	0.361	0.202	0.192	0.096	0.073	0.962	1.262	0.503
071119	20.4	0.252	0.138	0.125	0.074	0.053	0.431	0.477	0.596
071120	18.8	0.124	0.057	0.065	0.020	0.039	0.172	0.370	0.306
072117	12.7	0.493	0.276	0.285	0.176	0.028	4.790	5.208	0.607
072118	10.7	0.278	0.144	0.142	0.112	0.024	2.518	2.104	0.790
072119	9.1	0.215	0.104	0.118	0.074	0.023	0.506	0.533	0.627
072120	8.7	0.083	0.036	0.040	0.028	0.015	0.247	0.234	0.697
072314	11.0	0.689	0.379	0.376	0.290	0.023	0.784	0.669	0.774
072315	13.1	0.696	0.400	0.397	0.271	0.028	0.953	0.923	0.682
072316	12.8	0.507	0.286	0.313	0.168	0.026	0.736	0.906	0.536
072317	13.1	0.359	0.202	0.192	0.145	0.022	0.996	0.872	0.754
072318	13.0	0.327	0.185	0.135	0.174	0.018	1.214	0.619	1.295
072319	12.2	0.226	0.126	0.108	0.106	0.012	0.834	0.565	0.974
072320	10.9	0.076	0.038	0.031	0.036	0.009	0.613	0.348	1.164
072614	11.9	0.431	0.237	0.230	0.177	0.024	0.577	0.492	0.773
072615	13.3	0.641	9.366	0.312	0.296	0.033	1.007	0.702	0.946
072616	12.4	0.570	0.318	0.271	0.268	0.031	0.989	0.662	0.986
072617	12.4	0.463	0.254	0.229	0.202	0.032	0.936	0.698	0.885
072618	12.0	0.338	0.182	0.126	0.185	0.027	0.888	0.397	1.476
072619	11.3	0.203	0.106	0.082	0.102	0.019	0.614	0.329	1.234
072909	11.4	0.317	0.172	0.155	0.145	0.017	0.360	0.254	0.935
072910	12.7	0.480	0.272	0.220	0.239	0.021	0.407	0.248	1.085
072911		0.569				0.027			

ID	T(°C)	Rn	LE$_{EQ}$	LE	H	G	ΔT_D	Δe	β
072912	15.0	0.711	0.426	0.411	0.271	0.029	0.756	0.757	0.660
072913		0.703				0.031			
072914	14.0	0.575	0.334	0.316	0.230	0.029	0.589	0.533	0.729
072915	12.9	0.589	0.337	0.310	0.254	0.025	0.627	0.505	0.819
072916	12.6	0.593	0.335	0.321	0.246	0.026	0.593	0.510	0.767
072917	12.4	0.472	0.264	0.282	0.166	0.024	0.544	0.608	0.591
072918	11.5	0.334	0.183	0.183	0.135	0.016	0.500	0.447	0.738
072919	9.7	0.206	0.108	0.101	0.096	0.009	0.262	0.180	0.959
080308	9.1	0.104	0.047	0.055	0.032	0.017	0.146	0.167	0.576
080309	12.1	0.137	0.064	0.075	0.034	0.028	0.254	0.369	0.454
080310	14.1	0.261	0.137	0.142	0.082	0.037	0.244	0.280	0.575
080311	18.4	0.377	0.220	0.233	0.096	0.048	0.663	1.059	0.413
080312	20.3	0.400	0.235	0.227	0.113	0.060	0.654	0.867	0.497
080313	23.2	0.538	0.340	0.339	0.132	0.067	0.909	1.544	0.388
080314	24.5	0.715	0.472	0.490	0.152	0.073	0.758	1.615	0.310
080315	24.9	0.721	0.479	0.497	0.151	0.073	0.545	1.181	0.305
080316	25.0	0.401	0.253	0.239	0.102	0.060	0.973	1.499	0.429
080409	17.8	0.216	0.129	0.118	0.078	0.020	0.230	0.228	0.665
080410	20.5	0.328	0.207	0.178	0.121	0.029	0.464	0.449	0.681
080411	22.2	0.441	0.288	0.288	0.116	0.037	0.567	0.928	0.403
080412	24.6	0.538	0.367	0.352	0.146	0.040	0.698	1.111	0.415
080413	26.2	0.605	0.423	0.412	0.150	0.043	0.756	1.369	0.365
080414	27.0	0.619	0.436	0.422	0.152	0.045	0.685	1.256	0.360
080415	27.7	0.595	0.424	0.396	0.157	0.042	0.666	1.112	0.396

ID	T(°C)	Rn	LE_{EQ}	LE	H	G	ΔT_D	Δe	β
080416	27.6	0.519	0.367	0.342	0.137	0.040	0.579	0.949	0.403
080417	27.5	0.416	0.291	0.261	0.119	0.036	0.434	0.628	0.457
080418	26.7								
080419	25.4	0.163	0.103	0.111	0.027	0.025	0.153	0.411	0.246
080518	14.8	0.246	0.144	0.117	0.114	0.015	0.367	0.250	0.971
080519	15.5	0.115	0.067	0.064	0.043	0.008	0.213	0.205	0.684
080520	14.0	-0.003	-0.002	-0.006	0.002	0.001	0.015	-0.028	-0.361
080608	11.2	0.124	0.066	0.093	0.022	0.009	0.032	0.089	0.236
080609	12.4	0.278	0.156	0.161	0.104	0.013	0.237	0.244	0.641
080610	14.0	0.409	0.239	0.229	0.162	0.018	0.509	0.478	0.704
080611	15.0	0.515	0.308	0.331	0.163	0.021	0.588	0.790	0.492
080612	15.8	0.601	0.368	0.366	0.213	0.022	0.694	0.710	0.582
080613	16.0	0.640	0.392	0.348	0.267	0.025	0.768	0.662	0.765
080614	15.6	0.646	0.389	0.371	0.244	0.031	0.744	0.740	0.659
080615	14.6	0.572	0.338	0.300	0.246	0.026	0.642	0.515	0.822
080616	13.4	0.534	0.307	0.281	0.228	0.025	0.445	0.361	0.814
080617	14.4	0.436	0.260	0.245	0.176	0.015	0.677	0.625	0.715
080618	13.11	0.368	0.214	0.185	0.173	0.010	0.406	0.286	0.935
080712	15.2	0.608	0.361	0.332	0.244	0.032	0.911	0.821	0.732
080713	15.9	0.647	0.392	0.366	0.250	0.031	0.910	0.880	0.682
080714	16.1	0.638	0.389	0.349	0.260	0.029	0.789	0.699	0.745
080715	17.2	0.608	0.375	0.338	0.237	0.033	0.861	0.809	0.703
080716	18.2	0.530	0.331	0.296	0.201	0.033	0.846	0.821	0.680
080717	16.4	0.400	0.239	0.207	0.165	0.028	0.783	0.647	0.799

ID	T(°C)	Rn	LE_{EQ}	LE	H	G	ΔT_D	Δe	β
080718	15.8	0.268	0.156	0.149	0.097	0.022	0.496	0.499	0.656
080719	06.1	0.136	0.079	0.079	0.044	0.013	0.202	0.239	0.557
081110	8.6	0.419	0.219	0.264	0.147	0.008	0.462	0.549	0.556
081111	8.9	0.544	0.288	0.305	0.230	0.009	0.693	0.606	0.754
081112	9.4	0.631	0.339	0.366	0.255	0.010	0.783	0.741	0.696
081113	9.9	0.688	0.375	0.398	0.279	0.011	0.863	0.811	0.702
081114	10.5	0.684	0.379	0.445	0.229	0.010	0.650	0.332	0.515
081115	10.6	0.640	0.354	0.363	0.264	0.013	0.783	0.709	0.728
081116	10.5	0.577	0.316	0.323	0.239	0.015	0.773	0.690	0.740
081117	9.8	0.443	0.239	0.241	0.191	0.011	0.600	0.501	0.790
081118	9.2	0.292	0.154	0.159	0.125	0.008	0.474	0.400	0.781
081119	8.2	0.145	0.077	0.072	0.072	0.001	0.315	0.210	0.987
081213	10.4	0.651	0.353	0.346	0.283	0.022	1.002	0.807	0.820
081214	11.4	0.639	0.354	0.348	0.268	0.023	1.016	0.871	0.770
081215	12.6	0.592	0.335	0.328	0.239	0.025	0.935	0.849	0.727
081216	13.9	0.527	0.306	0.297	0.205	0.025	0.738	0.752	0.688
081309	7.7	0.287	0.145	0.107	0.174	0.006	0.190	0.077	1.632
081310	8.5	0.445	0.233	0.205	0.232	0.008	0.340	0.198	1.131
081311	9.0	0.534	0.282	0.300	0.221	0.013	0.797	0.715	0.736
081312	9.4	0.601	0.319	0.329	0.255	0.017	1.018	0.864	0.777
081313	9.7	0.623	0.333	0.319	0.284	0.020	1.173	0.872	0.889
081314	10.4	0.614	0.330	0.308	0.280	0.026	1.177	0.852	0.912
081315	11.2	0.582	0.317	0.294	0.260	0.028	1.252	0.934	0.884
081316	10.5	0.381	0.201	0.174	0.184	0.023	0.904	0.565	1.056

ID	T(°C)	Rn	LE$_{EQ}$	LE	H	G	ΔT_D	Δe	β
081409	11.7	0.277	0.151	0.177	0.085	0.015	0.262	0.359	0.482
081410	14.1	0.407	0.233	0.251	0.134	0.022	0.524	0.650	0.533
081411	16.1	0.534	0.322	0.275	0.229	0.030	0.489	0.389	0.830
081412	17.5	0.618	0.383	0.374	0.208	0.036	0.526	0.623	0.557
081413	19.0	0.632	0.400	0.400	0.192	0.040	0.485	0.665	0.481
081414	20.7	6.653	0.422	0.432	0.176	0.045	0.579	0.936	0.408
081415	20.8	0.588	0.379	0.328	0.216	0.044	0.546	0.547	0.658
081416	20.5	0.520	0.335	0.304	0.180	0.036	0.477	0.532	0.591
082217	7.3	0.372	0.183	0.145	0.210	0.017	0.728	0.331	1.453
082218	7.3	0.256	0.126	0.087	0.158	0.011	0.056	0.208	1.807
082219	6.6	0.096	0.046	0.018	0.073	0.005	0.393	0.065	3.980
082310	7.8	0.391	0.191	0.160	0.204	0.027	0.407	0.211	1.274
082311	10.1	0.476	0.249	0.215	0.233	0.028	0.586	0.358	1.081
082312	11.9	0.557	0.307	0.260	0.268	0.029	0.579	0.372	1.028
082313	12.9	0.593	0.335	0.364	0.199	0.030	0.363	0.438	0.546
082314	13.9	0.593	0.345	0.287	0.279	0.027	0.593	0.403	0.972
082315	12.3	0.569	0.322	0.236	0.313	0.020	0.743	0.368	1.331
082316	11.2	0.508	0.282	0.167	0.328	0.015	0.556	0.187	1.960
082317	10.4	0.407	0.221	0.130	0.264	0.013	0.560	0.182	2.036
082318	9.7	0.295	0.158	0.089	0.197	0.009	0.341	0.101	2.225
082508	6.4	0.053	0.020	0.046	-0.006	0.013	-0.024	0.116	-0.135
082509	9.7	0.198	0.098	0.156	0.021	0.021	0.075	0.369	0.135
082510	12.6	0.317	0.169	0.202	0.083	0.032	0.312	0.505	0.408
082511	14.6	0.410	0.228	0.208	0.160	0.042	0.597	0.512	0.770

ID	T(°C)	Rn	LE_{EQ}	LE	H	G	ΔT_D	Δe	β
082512	16.4	0.477	0.276	0.235	0.194	0.048	0.625	0.499	0.826
082513	19.1	0.489	0.298	0.204	0.236	0.049	0.816	0.468	1.152
082514	20.7	0.480	0.301	0.253	0.180	0.047	0.857	0.798	0.709
082515	19.7	0.469	0.292	0.252	0.176	0.041	0.936	0.885	0.698
082516	18.6	0.412	0.253	0.246	0.132	0.034	0.816	1.004	0.536
082517	17.9	0.299	0.181	0.165	0.109	0.025	0.706	0.704	0.662
082518	17.2	0.166	0.097	0.063	0.086	0.017	0.804	0.384	1.384
082519	15.0	0.042	0.020	0.003	0.028	0.011	0.637	0.045	9.249

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