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FACTORS AFFECTING EVAPORATION FROM A SUBARCTIC TUNDRA, CHURCHILL, MANITOBA

FACTORS AFFECTING EVAPORATION FROM A SUBARCTIC TUNDRA, CHURCHILL, MANITOBA

MSc

by

MONIKA M. DOBSON, B.A.

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AUTHOR: Monika M. Dobson, B.A. (McMaster University)

SUPERVISOR: Professor W.R. Rouse

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ABSTRACT

Evaporation was calculated for a subarctic beach ridge, near Churchill, Manitoba, using the energy balance approach. Energy balance calculations for the measurement season revealed an average Bowen ratio, β, of 0.68, with a value of 1.00 representing α' (the evaporability parameter). Fifty-seven percent of the net radiation was utilized by the evaporative heat flux over this tundra surface. Regressions were used to determine the most likely combination of environmental variables responsible for the behaviour of evaporation. Surface soil moisture remained relatively constant throughout the summer measurement period and soil temperatures appeared to be unrelated to evaporation. Air temperature proved to be insignificant to the evaporation flux, and net radiation alone could only account for 54% of the variability. The combination of the net radiation and the wet and dry bulb temperature depression at 1 m accounted for 88% of the variability of the evaporative heat flux. The mean α' for a site is assumed to be controlled by the surface type in simplified variations of the combination model. The conclusion has been drawn from this study that the variability of α' can be accounted for by variable atmospheric humidities as well as net radiation. The importance of this atmospheric control on the rate of evaporation is emphasized.

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Chapter 1

INTRODUCTION

In an Arctic environment, the radiation and energy balances respond to the long winters, short snow-free summers, long daylight periods, permafrost, xerophytic and semi-xerophytic tundra vegetation and the cold air masses present throughout most of the year. Plant species are strongly regulated by the presence of permafrost and the cold air. The vegetation, in part, influences the soil moisture, soil temperature and evaporation, and exerts a strong influence on the net radiant energy available at the surface. Interrelationships between climatic, soil and vegetative variables in a northern environment are often very complex and difficult to determine.

The aim of this study is to examine the environmental factors influencing the evaporative heat flux LE. LE was determined using the energy balance approach for select periods in June, July, August and September, 1978, at an upland tundra site at Churchill, Manitoba. Results of past studies have indicated that a relationship between soil moisture and evaporation exists in a variety of different environments (Barton, 1979; Marsh et al., 1979 unpublished; Williams et al., 1978; Rouse et al., 1977; Davies and Allen, 1973). As well, Rouse and Stewart (1972) showed that a simplified form of the equilibrium model (utilizing only net radiation and air temperature) could accurately estimate evaporation from low arctic upland tundra by using a constant evaporability factor, α'.

The present study examines these relationships and particularly considers the role of local atmospheric conditions in exerting a control over evaporation.

Chapter 2

PHYSICAL BASIS

The energy balance of a given surface can be expressed as

$$Q^* = LE + H + G \tag{1}$$

where

Q^* = net radiation
LE = latent heat flux
H = sensible heat flux
G = soil heat flux

Components such as snowmelt and photosynthesis have been neglected due to negligible local importance during the measurement period. Q^* and G are generally measured directly, H can be determined by residual from (1) provided LE has been determined. Following Bowen (1926), evaporation may be expressed as

$$LE = \frac{Q^* - G}{1 + (H/LE)} = \frac{Q^* - G}{1 + \beta} \tag{2}$$

where the Bowen ratio, β, can be defined in terms of the ratio of the vertical dry bulb temperature between two levels, ΔT, to the vertical vapour pressure between two levels, Δe, in the form

$$\beta = \frac{\gamma \Delta T}{\Delta e} \tag{3}$$

in which γ is the psychrometer constant. Substitution of (3) into (2) results in

$$LE = \frac{Q^* - G}{1 + \left(\frac{\gamma \Delta T}{\Delta e}\right)} \tag{4}$$

Δe can be calculated in terms of dry and wet bulb temperature differences (Dilley, 1968) where

$$\Delta e = (s + \gamma)\ \Delta T_w - \gamma \Delta T \tag{5}$$

s = slope of saturated vapour pressure versus temperature curve at the mean of the wet bulb temperature at two levels
T_w = wet bulb temperature
ΔT_w = vertical wet bulb temperature differences

Substituting (5) into (4) results in the energy balance equation,

$$LE = (Q^* - G)\ \left(1 - \frac{\gamma}{s + \gamma} \quad \frac{\Delta T}{\Delta T_w}\right) \tag{6}$$

which allows the calculation of LE from direct measurement of ΔT and ΔT_w, providing temperature measurements approaching an accuracy of .01C are made, and that all measurements take place within the surface boundary layer. The depth of the surface boundary layer is a function of the distance downwind from a terrain type which has different surface characteristics from the one where measurements are made. To be within the boundary layer, the height to fetch ratio should be at least 1:100.

In subarctic and tundra regions of high latitudes, a variation of the equilibrium model developed by Priestley and Taylor (1972) has been found to be applicable in the form

$$LE = \alpha' \frac{S}{S + \gamma} (Q^* - G) \quad (7)$$

where

α' = evaporability parameter relating actual to equilibrium evaporation.

α' values have been found to increase with available soil moisture (Rouse et al., 1977), with $\alpha' = 1.26$ representing potential evaporation for wet, freely evaporating surfaces (Priestley and Taylor, 1972; Rouse et al., 1977; Stewart and Rouse, 1976; Stewart and Rouse, 1977).

The combination model was expressed by Slatyer and McIlroy (1961) in the form

$$LE = \frac{S}{S + \gamma} (Q^* - G) + \frac{\rho\, Cp}{r_a} (D_z - D_o) \quad (8)$$

with

ρ = air density (kg m^{-3} X 10^3)
Cp = specific heat of air at a constant pressure (J kg^{-1} K^{-1} X 10^3)
r_a = aerodynamic resistance to the diffusion of water vapour (s m^{-1})
D_z = wet bulb depression at height z (K)
D_o = wet bulb depression at the surface (K)

From (7) and (8)

$$(\alpha' - 1)\ (Q^* - G)\ \frac{S}{S + \gamma} = \frac{\rho\ Cp}{r_a}\ (D_z - D_o) \tag{9}$$

and;

$$\alpha' - 1 = \frac{\rho\ Cp}{r_a}\ \frac{D_z - D_o}{Q^* - G}\ \frac{S + \gamma}{S} \tag{10}$$

expressing α' as a function of $D_z - D_o$, $Q^* - G$, air temperature, and r_a. r_a can be defined as

$$r_a = \frac{\left(\ln \left(\frac{z-d}{z_o} \right) \right)^2 \phi}{k^2\ u(z)} \tag{11}$$

where

z	=	height (m)
d	=	zero plane displacement (m)
z_o	=	roughness length (m)
ϕ	=	stability function
k	=	von Karman's constant (.41)
$u(z)$	=	windspeed at height z ($m\ s^{-1}$)

Equilibrium evaporation, as presented by Slatyer and McIlroy (1961) was considered to be valid only under the conditions of a saturated atmosphere, where $D_z = D_o$. Under these conditions, α' (from (10)) equates to 1.00, defining equilibrium evaporation, LE_{eq}, in the form of equation (7), with

$$LE_{eq} = \frac{S}{S + \gamma} (Q^* - G) \tag{12}$$

Later, equilibrium evaporation was revealed as applicable under non-saturated conditions over moderately dry surfaces where a constant resistance to evaporation was in effect (Denmead and McIlroy, 1970; Wilson and Rouse, 1972; and, Rouse and Stewart, 1972). It was considered that the wet bulb depressions would be equal if a mutual adjustment between the air and the surface had been reached with regard to moisture (Monteith, 1965, and; Tanner and Fuchs, 1968). Wilson and Rouse (1972) felt that the depressions may have finite values and may still be equal or nearly equal. In the subarctic, resistances due to the xerophytic vegetation create a substantial canopy resistance.

Chapter 3

SITE AND METHODS

Site

Research was undertaken during the period between May 1 and mid-September 1978, at a site situated in an open tundra environment on a raised beach system, approximately 2 km south of the Hudson Bay coastline and 10 km east of Churchill, Manitoba (58^{o} $45^{'}$N 94^{o} $04^{'}$W) (Figure 1). Due to recent glaciation, immature drainage networks combine with the existence of continuous permafrost to form an abundant number of small shallow lakes (Figure 2). The dominant vegetation on this subarctic ridge consists of *Dryas integrifolia* and *Carex rupestris*, which are highly adapted to dry Arctic conditions. Sparse occurences of lichen are encountered, but the plant environment is not lichen dominated, and in this respect differs from the study site of Rouse and Stewart (1972). Figure 3 illustrates that the soils are comprised of sand intermixed with thin gravel layers at varying depths throughout the tundra site. The organic layer immediately below the surface is 10 cm deep, on average. Figure 4 provides a detailed map of instrument locations. A small lake east of the instrument site would result in inadequate height: fetch requirements if winds were from the east. As well, a spruce lichen

FIGURE 1: SITE LOCATION

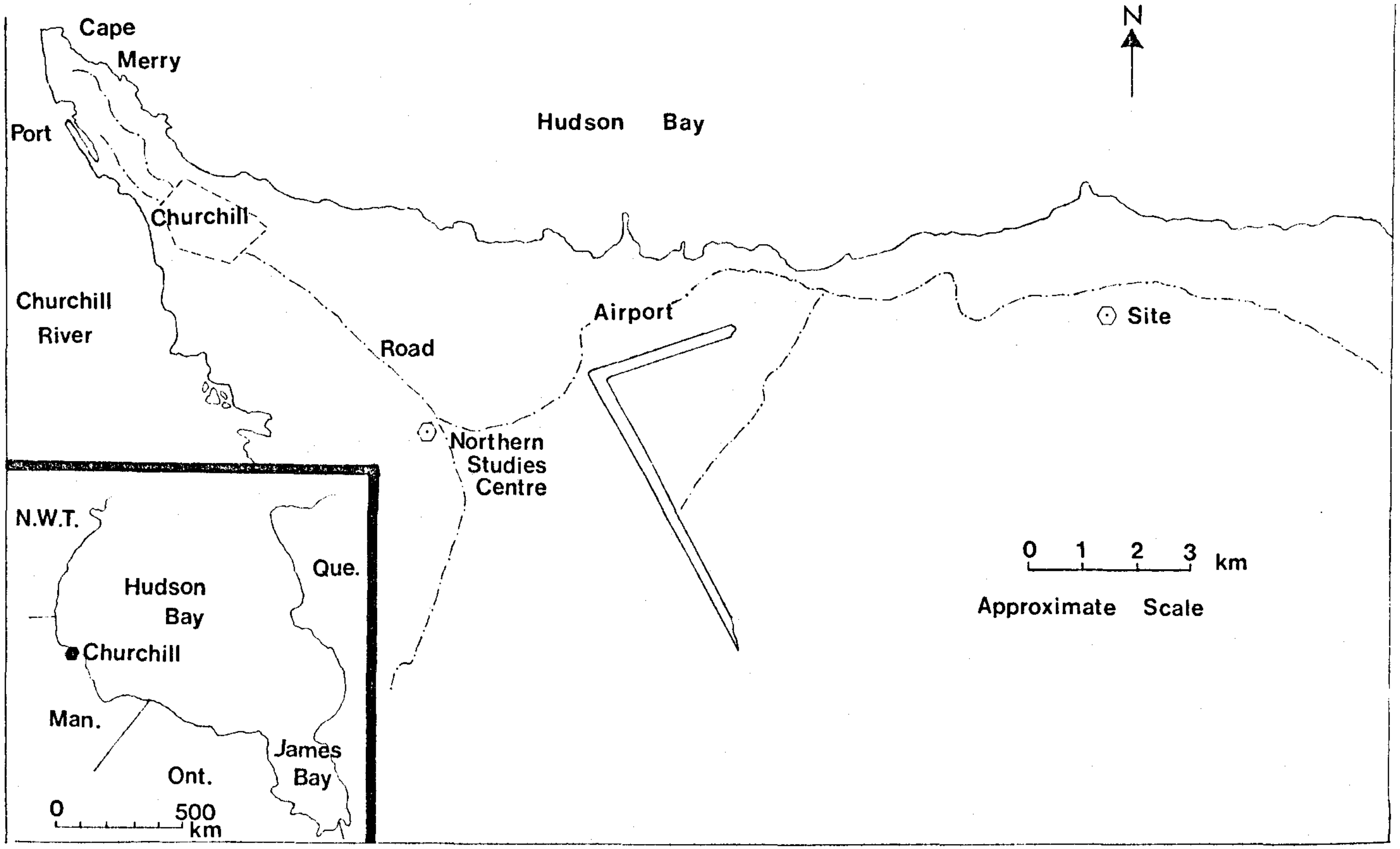
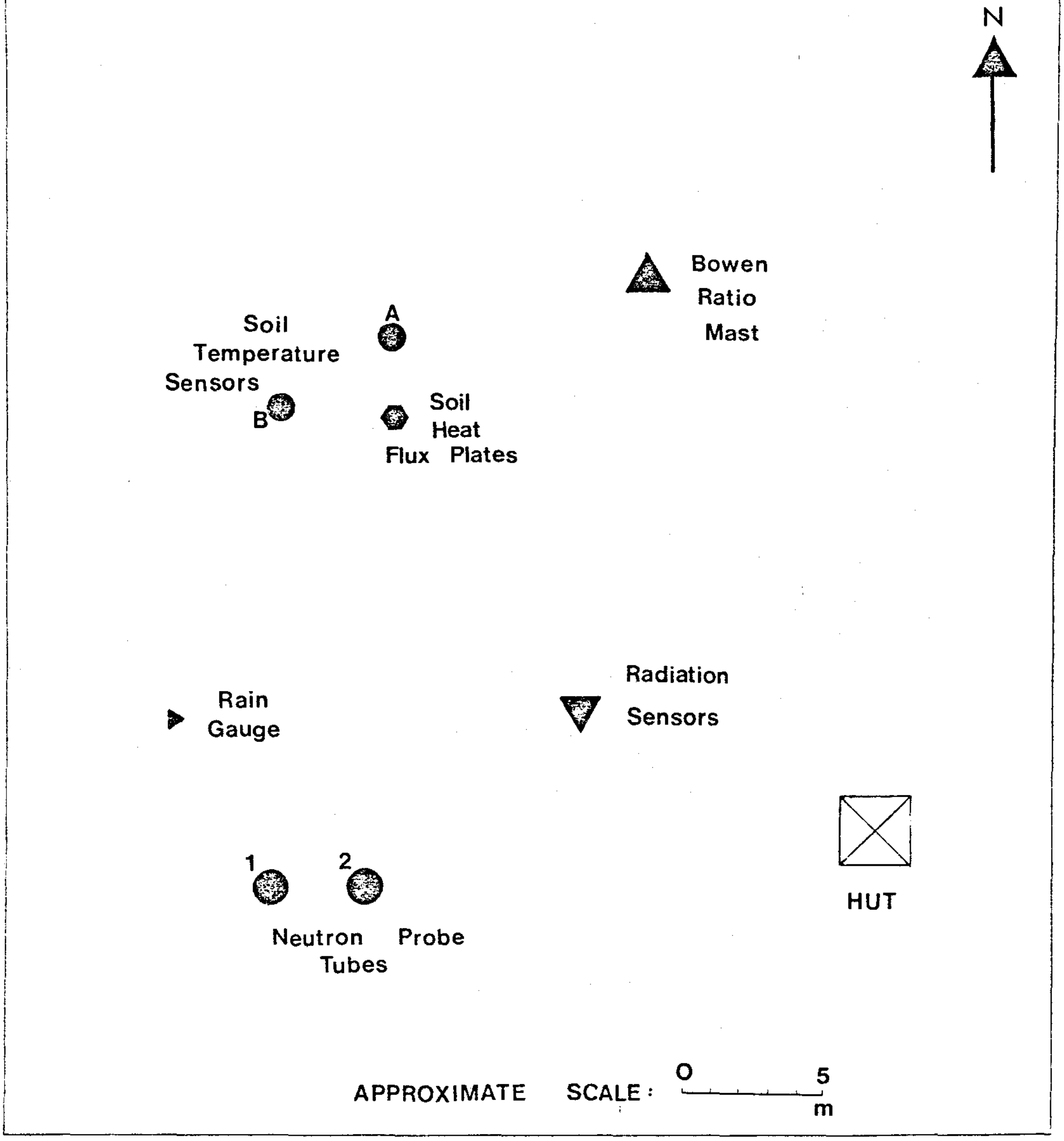


Figure 2: Research Site From Air

Figure 3: Sample Soil Pit

FIGURE 4: INSTRUMENT LOCATION



forested site SSE of the site would produce a discontinuous fetch. Winds at the site were exclusively from the N, NE, NW and W during the data collection periods, giving an adequate height: fetch ratio of at least 1:100. This allowed for energy balance determination of LE within an adjusted set of temperature and humidity profiles.

Methods

In order to evaluate LE from (6) and to identify any relationships between individual variables in this environment, basic energy and radiation balance components were monitored. For the daily analyses, only daylight totals were used, which were calculated from a variable number of hours.

Net radiation was measured with a Swissteco net radiometer (Figure 5). This instrument consists of a blackened thermopile transducer enclosed within aspirated polyethylene domes, on the top and bottom, which are transparent to both longwave and shortwave radiation. The mV output is proportional to temperature differences between the top and bottom surfaces of the thermopile and hence the differences in radiation receipt. This instrument is accurate to within $\pm$ 10% (Latimer, 1972; Sinclair et al., 1975; Fuchs and Tanner, 1970). Radiation measurements were derived from half-hourly integrated mV signals recorded on a Campbell CR5 Digital Recorder.

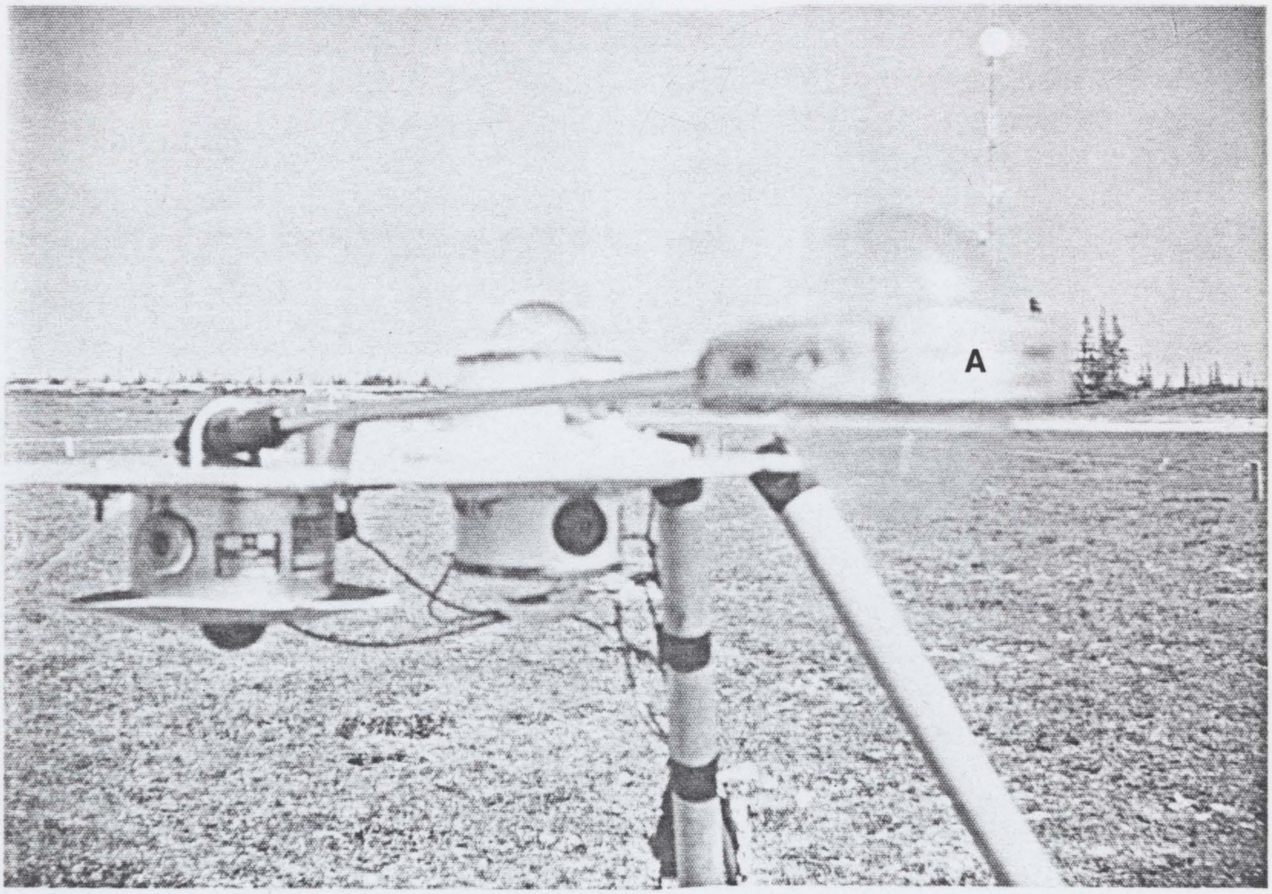


Figure 5: Radiation Instruments
(A) Net radiometer

As well, the Campbell recorder monitored the soil heat flux at 5 cm below the surface with three Middleton Heat Flux Plates, connected in series. Possible error, using this method for measuring G can be as high as $\pm$ 15% (Fuchs and Tanner, 1970).

Soil temperatures were manually recorded from two profiles of YSI thermistors set at depths of 5, 10, 20, 40, 80 and 160 cm below the surface. The accuracy of the thermistor reading is approximately $\pm$2%.

Wet and dry bulb temperatures were measured at 25, 50, 75 and 100 cm above the surface, in shielded and aspirated housings (Figure 6). Copper constantan thermopiles (accurate to $\pm$ .01C), consisting of five junctions, were referenced to an electrically maintained zero C temperature reference bath. The mV signal was monitored at two minute intervals with an Esterline-Angus D2020 recorder, and stored on magnetic tape for computer analysis. The wet bulb sensor was covered with a muslin wick, wetted from an individually regulated water feed. As evaporation proceeds and latent heat is lost, the wet bulb temperature becomes less than the dry bulb temperature. ΔT's and ΔT_w's were calculated between six levels (25 cm to 50 cm, 25 cm to 75 cm, 25 cm to 100 cm, 50 cm to 75 cm, 50 cm to 100 cm, and 75 cm to 100 cm). The values of ΔT and ΔT_w **for each pair of levels were used to calculate an LE value each time, with the average of the six LE results comprising the** final LE value. This was the situation 35% of the time. The remainder of the calculations used only the average of the 25 to 50, 25 to 75, and 25 to 100 cm differences. Highly irregular LE results (greater

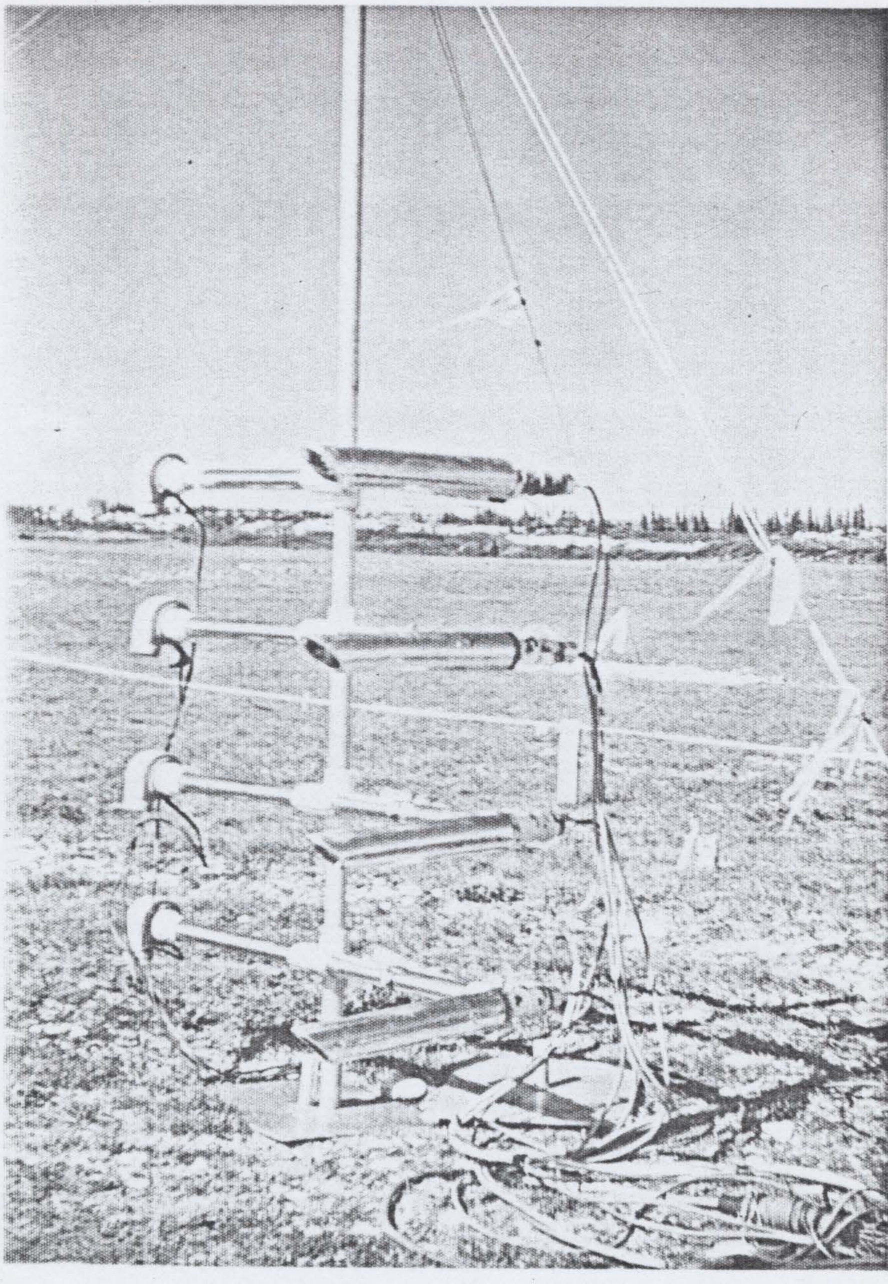


Figure 6: Bowen Ratio System

than $\pm$ 200% of LE_{eq}) emerged when the other three measurement levels were used. Kinks in the vertical distribution of T and T_w values may have been responsible for these erratic results. Other climatological occurances common to this 65% group of days could not be found.

A 'typical' error value for the calculation of the latent heat flux, using the combination model or the energy balance approach is $\pm$ 15% (Bailey, 1977). When humidity gradients are large (Δe > 25 Pa), the error in the calculation of LE is approximately $\pm$ 9%. Smaller Δe values leads to errors of 19% or greater. Therefore, the calculation of LE is more reliable when wet conditions prevail.

Soil moisture was measured routinely every three days, as well as before and after each rainfall. Measurements were made at 10 cm intervals between 5 cm and 95 cm depth, at two separate neutron probe aluminum access tubes. A Nuclear Chicago Subsurface Neutron Probe was used. This instrument emits neutrons from a beryllium source embedded in a probe which is lowered into an access tube. Neutrons encountering hydrogen atoms lose energy through molecular collision and some diffuse back to the boron trifluoride detector. To prevent loss of neutrons to the air for measurements at the 5 cm depth, a 30 X 30 X 20 cm soil basket, maintained in the same environment as the soil adjacent to the access tube, was placed over the tube. The wetter the soils, the greater the rate of neutron return. Neutron count rate is a linear function of volumetric soil water. An error margin of $\pm$ 5% should be allowed for depths of 15 cm and more, with the possibility of a slightly higher error resulting from the use of the soil basket for the 5 cm measurement. The surface organic layer may slightly alter the rate of neutron return but the effects can be considered negligible in this study due to the relatively low count rates encountered near the surface.

Wind, precipitation, and cloud data were obtained from the local Atmospheric Environment Station, at Churchill Airport.

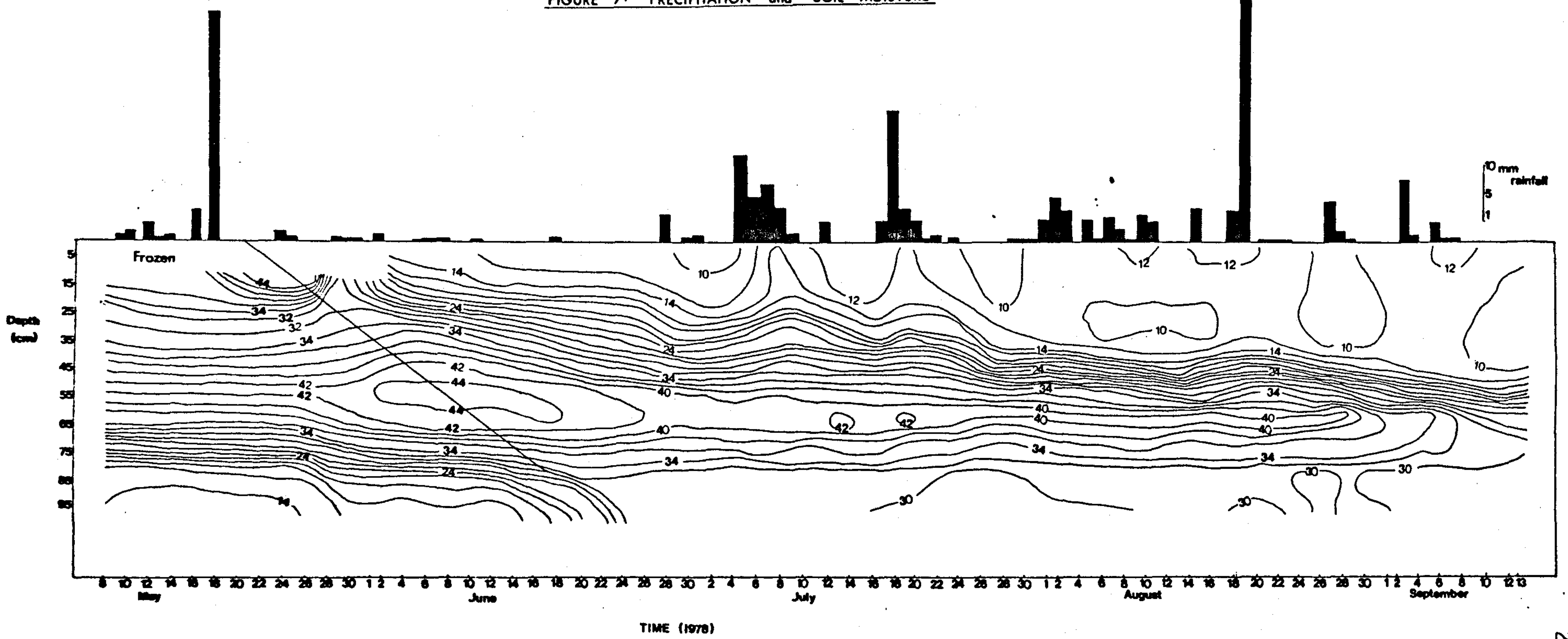
Chapter 4

RESULTS AND DISCUSSION

Results

Soil moisture patterns in Figure 7 show very wet surface soils immediately after snowmelt. Percolation and evaporation have reduced volumetric moisture levels to 10% by the end of June, after which time values at 5 cm depth remain between 8% and 12% throughout the remaining measurement period. Figure 7 shows precipitation over the entire measurement period. May, July and August received above average rainfall for the Churchill region, to contrast with the month of June which was relatively dry. An important feature emerges from the soil moisture plot. The rain water appears to be percolating into the soil very quickly after a rainfall. These soils, as illustrated in Figure 3, can be considered to be saturated in the range of 40 to 44% soil moisture by volume. This was derived from the fact that when soil pits were dug on the tundra site, standing water was reached at the same depth as the 40 to 44% soil moisture reading with the neutron probe. The thawing of the frost table does not appear to have affected the volume of soil moisture.

FIGURE 7: PRECIPITATION and SOIL MOISTURE



The soil temperature regime illustrated in Figure 8 shows several interesting patterns. The surface started to thaw on May 14, but a major snowstorm on May 17 delayed final thaw for a week. The frost table started to recede from the surface while there was still snow on the tundra, and followed a fairly constant and rapid rate of retreat until the end of July after which the rate slowed substantially (Figure 9). The fairly constant retreat rate was probably maintained by heat input from percolating water.

If one arbitrarily defined the summer period as one when mid-afternoon surface soil temperatures generally exceed 10C, the following pattern emerges. Mid-afternoon temperatures exceeded 10C only after the frost table fell below a depth of 70 cm, an event that coincided with the start of a clear dry period on June 12. On extremely cold days, such as June 17 when midday soil temperatures dropped to 4C, the rapid response of the surface soil temperatures to air temperature change is emphasized. High soil temperatures, at 5 cm, occurred during the mid-summer evaporation period, reaching values larger than 20C on warm days in June and July. Surface soils started to get cold at the end of August, caused by the onset of a very cold, overcast period for a week and a half. This really marked the end of summer since subsequent mid-September values during a warm spell did not rise above 10C at 5 cm depth.

Figure 10 illustrates the diurnal trends of Q^*, LE, G, α' and β, measured only on clear days. On overcast days, energy levels were too low to distinguish any significant trends, and, on rainy days,

FIGURE 8 : 5 cm SOIL TEMPERATURES at 1400 (SOLAR TIME)

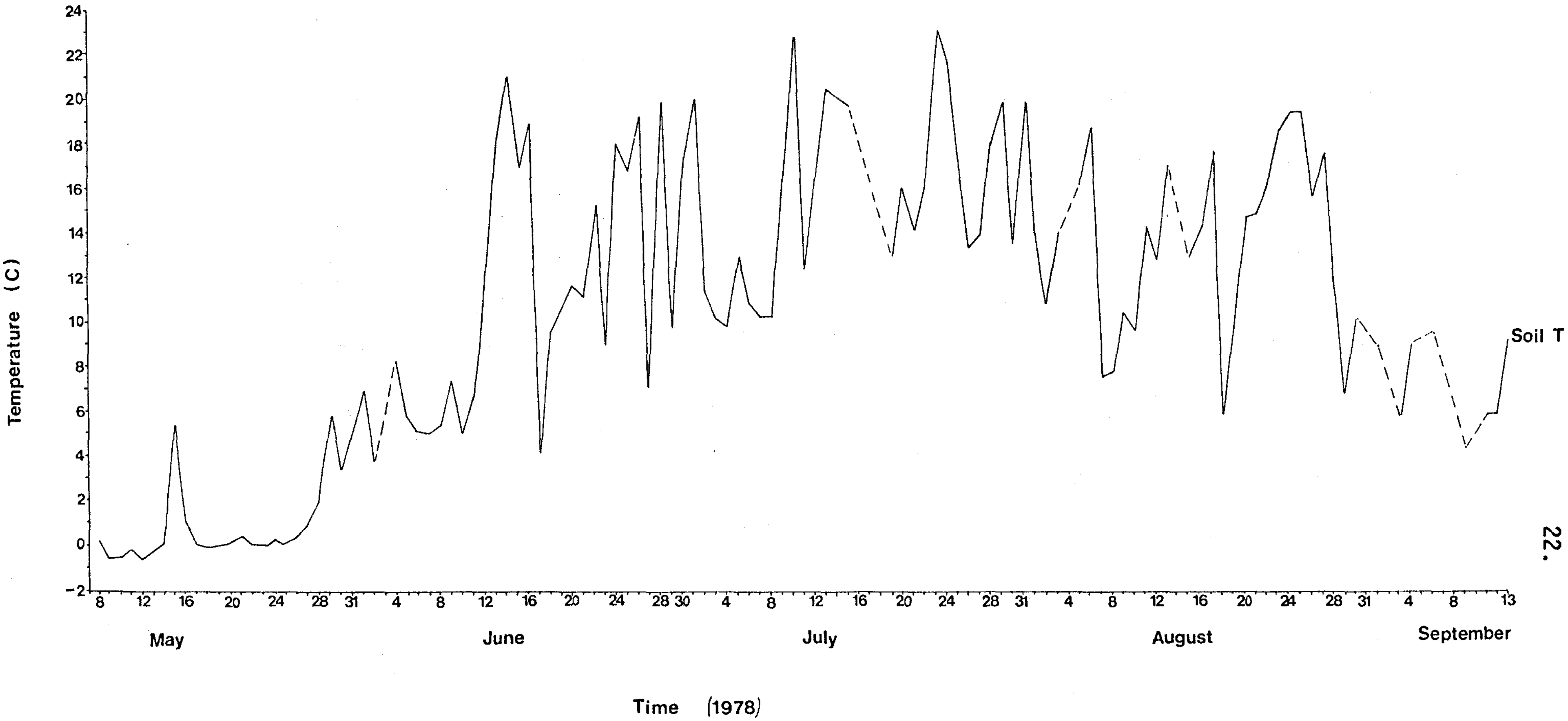


FIGURE 9 : $^G/_{Q*}$ and ACTIVE LAYER DEVELOPMENT during the MEASUREMENT SEASON

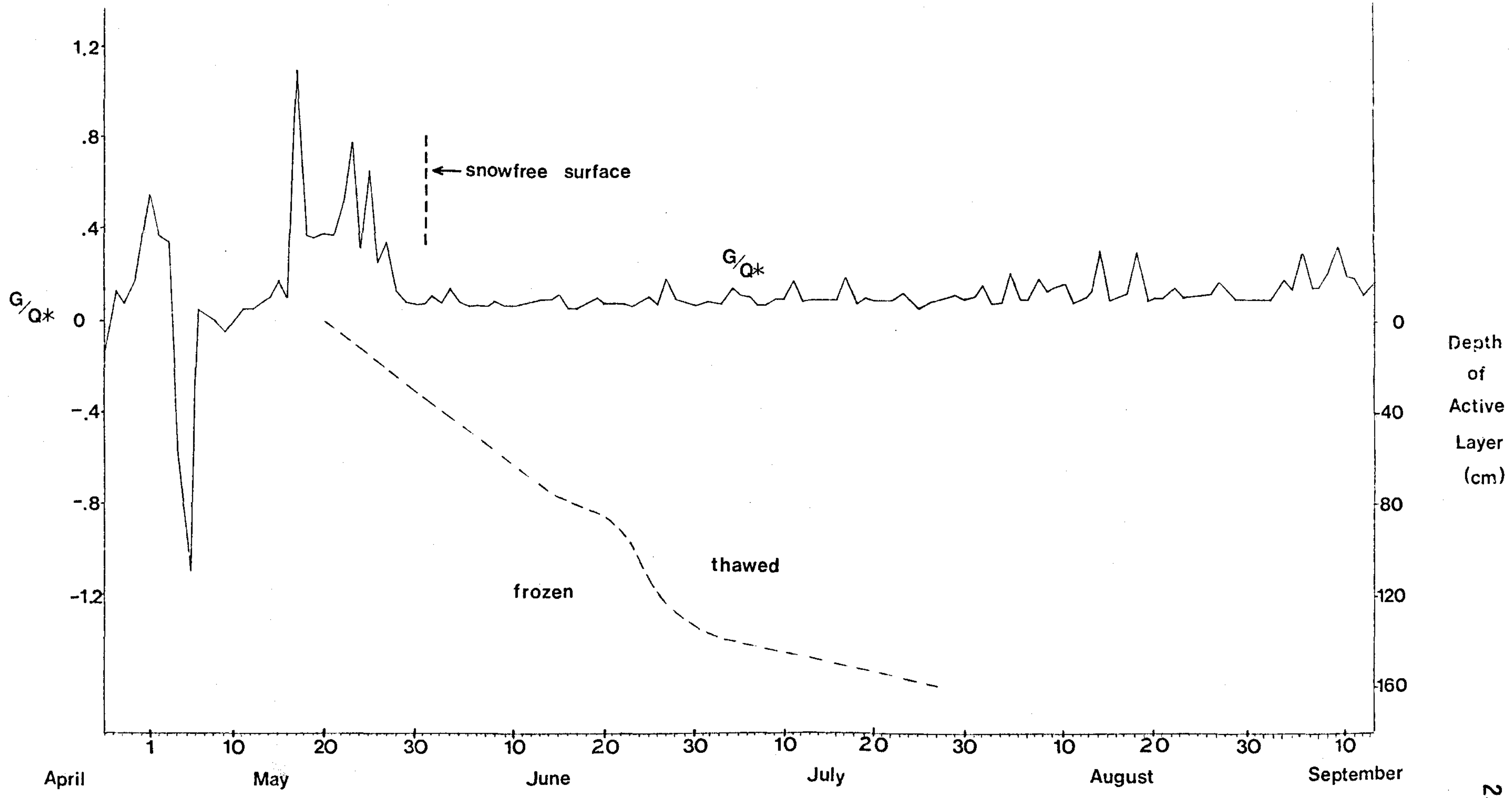


FIGURE 10 ENERGY BALANCE DIURNAL TRENDS

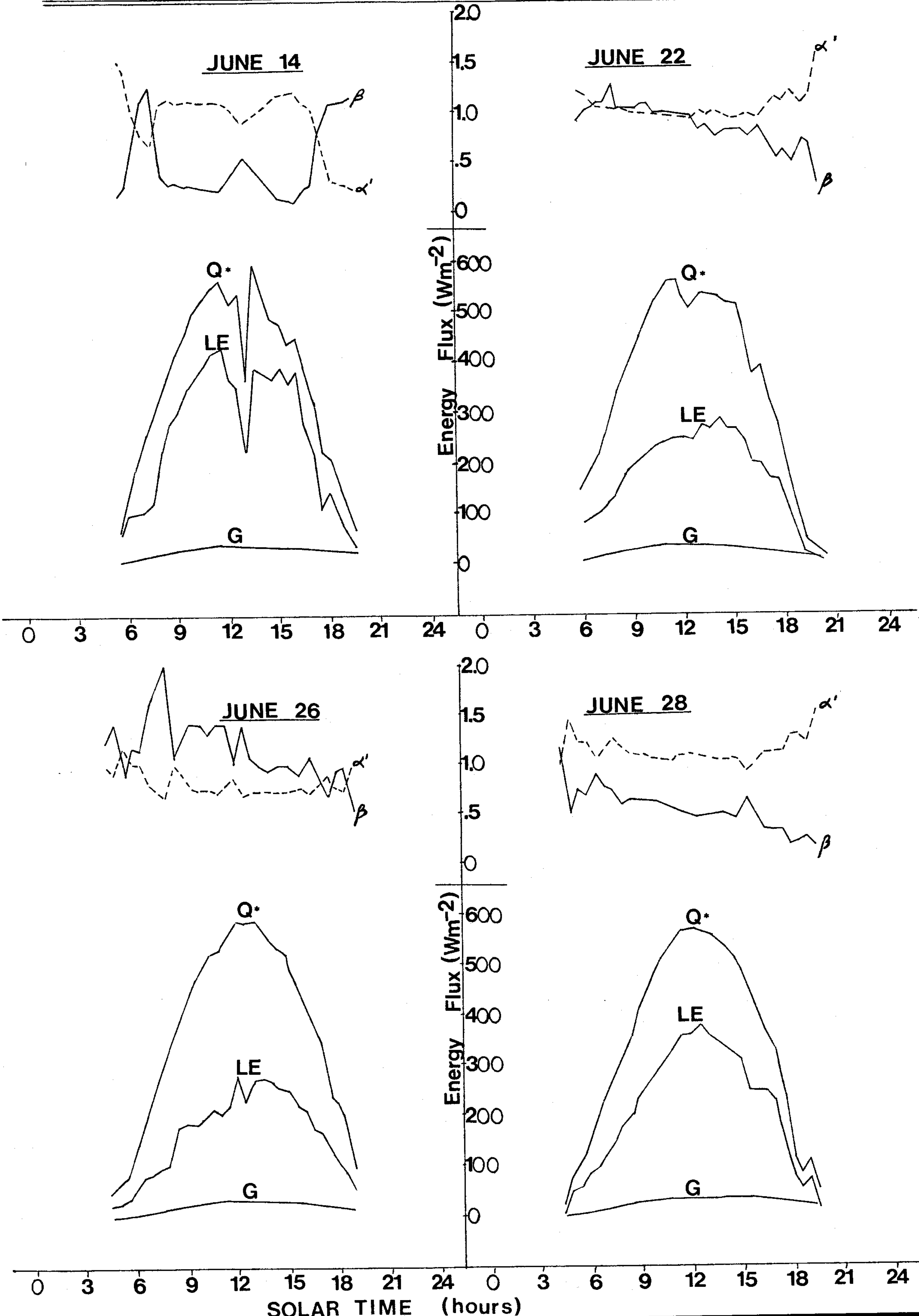


FIGURE 10 cont.

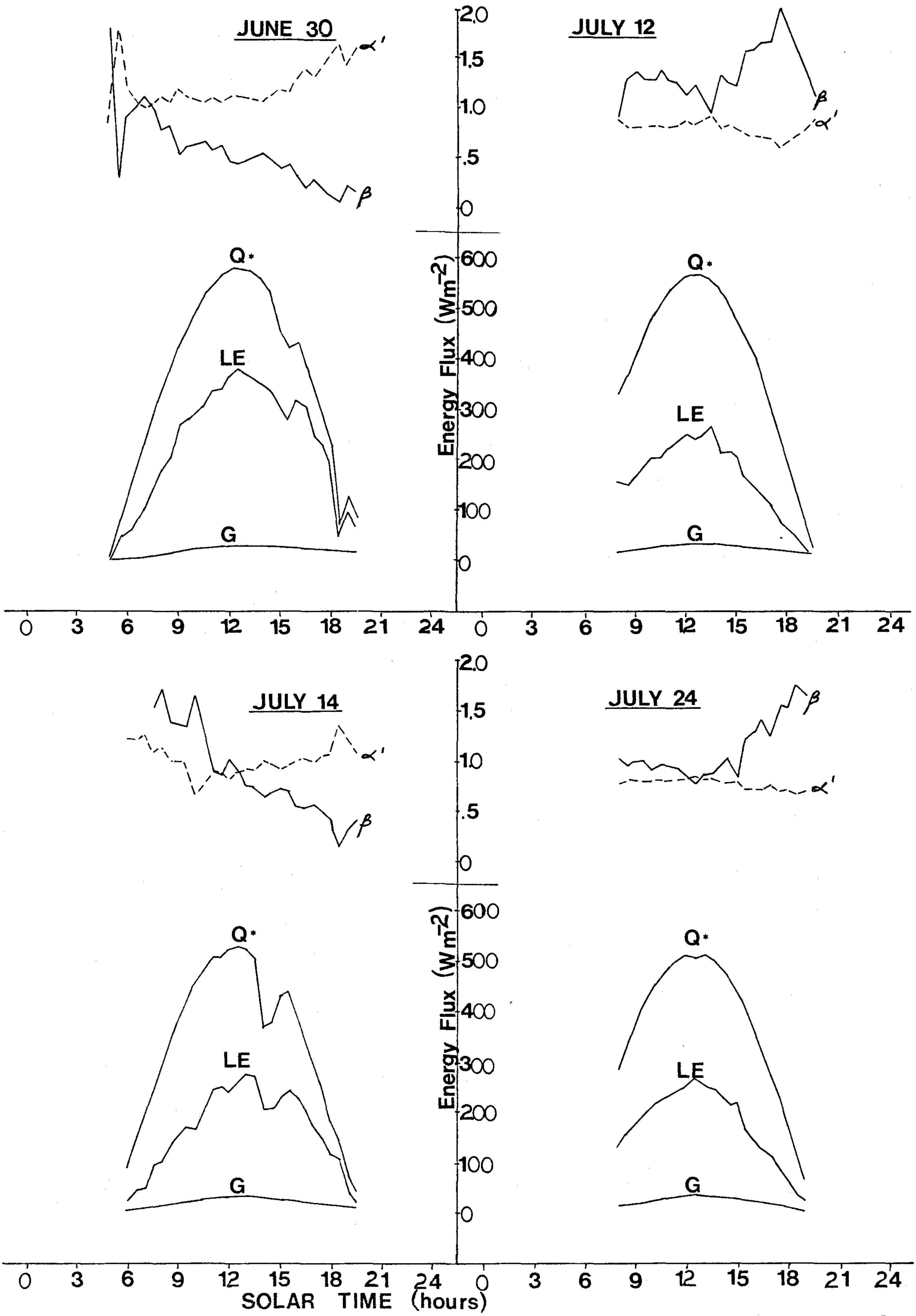


FIGURE 10 cont.

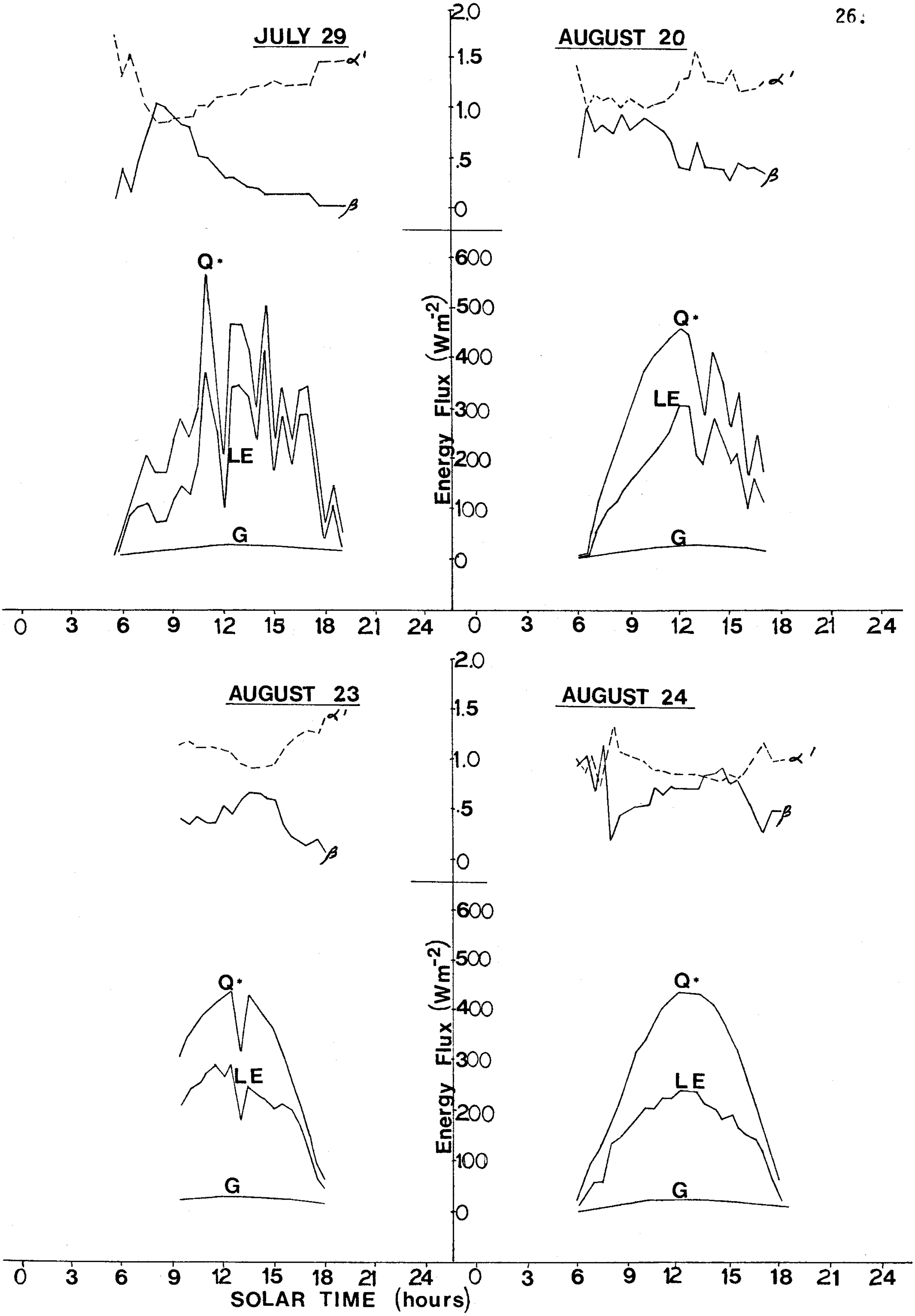
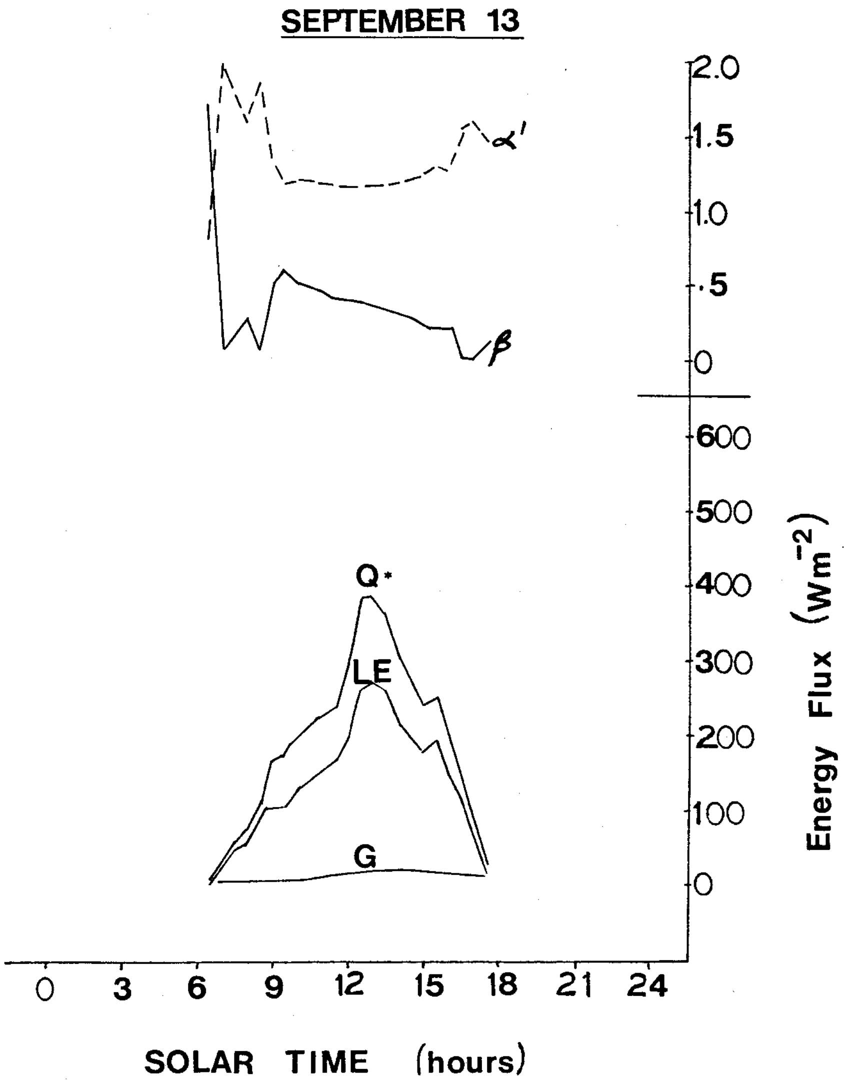


FIGURE 10 cont.



evaporation cannot be accurately monitored. Net radiation values at solar noon, around the time of the summer solstice, were close to 600 Wm^{-2}, after which, values steadily decreased. By September, the net radiation at solar noon was less than 400 Wm^{-2} on a clear day.

Table 1 lists the daily values of all the measured and calculated components. Q* and LE are daily totals. D_z, soil temperature, air temperature, windspeed and r_a each represent the mean of the daily measurement period. The soil moisture measurement was taken at 0900 each day. α' and β are weighted means for the daily measurement periods, and are weighted with 1/2 hourly measurements of Q*.

Diurnal trends of α' and β (Figure 10) for each of the 13 days of measurement are not evident. The average for β over the entire measurement period was 0.68, indicating that moderately wet conditions prevailed. This agrees with the soil moisture measurements. Daily averages of β, however, varied between the values of 0.34 and 1.30, with no apparent relationship to the relatively constant surface soil moisture. The mean α' value during the measurement season was 1.00, with a large scatter. The value of 1.26 was reached once. If the daily averages of α' in Table 1 are studied, and compared to the surface soil moisture results in Figure 7, a relationship is not evident. For example, there was not any significant precipitation during the latter part of June, yet daily α' values increased during this time. This sample would indicate that the surface resistance to evaporation is decreasing as the soils dry, which is contrary to normal physical behaviour.

Table 1

Pertinent Environmental Variables on a Daily Basis

Date		(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)	(9)	(10)	(11)
June	14	0530-1930	19.06	13.34	7.88	1.06	.35	11.8	16.4	18.4	5.07	55.23
	22	0600-2000	18.88	9.24	3.51	.97	.93	11.3	10.4	8.1	3.12	89.74
	26	0430-1900	19.94	8.75	2.89	.78	1.17	10.3	13.8	11.6	3.75	74.67
	28	0430-1930	19.06	11.59	4.25	1.07	.53	10.3	15.0	12.5	2.19	127.85
	30	0500-1930	19.48	12.32	4.07	1.15	.51	9.6	11.6	10.0	4.58	61.14
July	12	0800-1930	17.28	7.03	1.55	.80	1.30	11.0	14.4	8.8	3.56	78.65
	14	0600-1930	17.16	8.36	3.73	.96	1.00	10.4	16.7	12.0	3.41	82.11
	24	0800-1900	15.80	7.32	2.14	.79	1.03	10.6	18.4	13.0	3.17	88.33
	29	0530-1900	12.67	9.00	5.95	1.12	.35	8.1	15.1	16.7	5.22	53.64
Aug.	20	0600-1700	11.72	6.98	2.97	1.18	.59	11.5	10.2	8.8	3.52	79.55
	23	0930-1830	10.30	6.63	5.11	1.07	.44	10.7	17.5	16.1	1.72	162.79
	24	0600-1800	12.45	6.95	3.89	.91	.68	10.4	15.9	15.6	2.77	101.08
Sept.	13	0630-1730	7.93	5.60	4.06	1.26	.34	9.7	6.7	10.8	3.65	76.71

(1) Time Interval
(2) Q* (MJ m^{-2} daily $period^{-1}$)
(3) LE (MJ m^{-2} daily $period^{-1}$)
(4) Daily Mean D_z (K)
(5) α' (weighted average)
(6) β (weighted average)
(7) Soil Moisture at 5 cm Depth (% moisture by volume)
(8) Daily Mean Soil Temperature at 5 cm Depth (C)
(9) Daily Mean Air Temperature (C)
(10) Daily Mean Windspeed (m s^{-1})
(11) Daily Mean r_a (s m^{-1})

Rouse et al. (1977) suggest that α' is not controlled by soil moisture, except in the instance when the surface is very wet, and that a constant value of α' may be appropriate for a surface type. Table 2 compares the α' values obtained at various different subarctic sites (Rouse et al., 1977) to the Churchill values. The data for the old and new burns includes only non-potential values. For the new burn and the lichen heath sites, the variance of the α' values about the mean is not different from the Churchill results. The largest standard deviation occurs in the Churchill data set.

Table 3 lists the daily ratios of LE/Q*, H/Q*, G/Q* for measurement days. Clear cut trends for the partitioning of LE/Q* and H/Q* during any one of the drying periods is not apparent. A seasonal average value for LE:H:G of 57:36:7 can be calculated from the data.

Daily ratios of G to Q* are shown in Figure 9. Inconsistent daily values, in the range of $-1.1 < G/Q^* < 1.1$, are evident while snow remained on the ground. From the period immediately following snowmelt, until mid-August, G/Q* values were more constant, within the range of 0.0 to 0.2 most of the time. On clear days, throughout the entire measurement period, the G/Q* ratio was smaller than on cloudy days, indicating that more of the available energy was being used by LE and H. Between mid-August and September, slightly higher ratios than 0.2 appeared on a few occasions when significant (>3C) air temperature changes occurred from one day to the next. Soil temperatures, while responding to sudden air temperature changes, do so on a modified basis, by creating a temporarily larger differential

Table 2

α' Values Compared Statistically From Five Sites

	Mean α'	Standard Dev.	Sample Size
Old Burn (Rouse et al., 1977)	.97	.07	16
New Burn (Rouse et al., 1977)	.91	.11	15
Lichen Heath (Rouse et al., 1977)	.95	.12	30
Beach Ridge Churchill, 1978	1.00	.14	13

Churchill, 1978 vs.	F-Statistic	Degrees of Freedom_1	Degrees of Freedom_2	$F_{.05}$
Old Burn	4.00	12	15	2.48
New Burn	1.62	12	14	2.53
Lichen Heath	1.36	12	29	2.10

The variance of the data about the mean is significantly different from the Churchill data for the Old Burn site, and not significantly different for the New Burn and Lichen Heath sites.

Table 3

Daily Proportions of LE, H and G to Q*

Over the Measurement Period

Date	LE/Q*	H/Q*	G/Q*
June 14	.70	.22	.076
22	.49	.45	.063
26	.44	.50	.061
28	.61	.32	.073
30	.63	.31	.056
July 12	.41	.53	.063
14	.49	.44	.069
24	.46	.47	.068
29	.71	.22	.074
Aug 20	.60	.33	.066
23	.64	.28	.076
24	.56	.37	.071
Sept. 13	.71	.23	.062
Mean	.57	.36	.068
Standard Deviation	.106	.109	.0063

between the surface and the ground at 5 cm than would normally occur of air temperatures remained constant over a period of days.

Table 4 lists the results of a series of regressions, which attempt to isolate the variable, or combination of variables, which most effected the magnitude of LE. As can be seen from Table 4, LE is related to Q*, such that 54% of the variation in LE can be accounted for by variation in Q*. Regressing LE against D_z resulted in a relatively poor but nevertheless significant correlation coefficent. However, a multiple regression of Q* and D_z help explain 88% of the variability in LE for the measurement season. A test of the main terms in the combination model is presented when LE is regressed with $\frac{S}{S + \gamma}$ (Q* - G) and D_z, resulting in a correlation coefficient of .91.

The regressions in Table 4 indicate that as well as the energy driving force, in the form of Q*, effecting evaporation, the ability of the atmosphere to accept evaporated water is important. In this subarctic environment, a strong atmospheric control was in effect; an effect not dependant on air temperature alone, but also involving atmospheric humidity.

During the relatively dry period experienced between June 14 and June 30, Q* values for the five relatively clear measurement days varied only slightly in magnitude, whereas the LE flux experienced differences as large as 4 MJ m^{-2} daily $period^{-1}$ from day to day (Table 1). This latter variability paralleled the pattern of change in D_z. The largest daily LE occurred on June 14, coinciding with the

Table 4

LE Regressed With Environmental Variables

	r	r^2	Standard Deviation of LE About the Regression Line
$LE = 1.95 + .435\, Q^*$	.73	.54	1.69
$LE = 5.54 + .790\, D_z$	.55	.30	2.09
$LE = -1.77 + .455\, Q^* + .853\, D_z$	.94	.88	.89
$LE = -0.35 + .785\, ^{**} + .565\, D_z$	.91	.83	1.09

$$^{**}\ \frac{S}{S + \gamma}\ (Q^* - G)$$

All regressions were significant

largest daily mean D_z. In similar fashion, subsequent smaller daily mean D_z values were mirrored with smaller LE values. The strong influence of D_z on LE diminished toward the end of summer when smaller amounts of net energy were available, at which time Q* exerted substantially more influence over LE than did D_z. For example, on September 13, the smallest Q* and LE values were encountered, yet D_z values were the same as the average for all measurement periods.

DISCUSSION

As noted earlier, Q* was quite large at the solstice, after which it steadily decreased. This did not appear to influence the ratio of LE/Q* as the season progressed, as is seen in Table 3. Despite the large changes in zenith angles and daylengths encountered between June 14 and September 13, the proportion of the available energy utilized by evaporation at the start of the measurement period is almost the same as at the end, in mid-September. Therefore, the evaporation regime, in this study, appears to be time independant when air temperatures are greater than OC. The seasonal value for LE/Q*, at 0.57, was slightly lower than the range of $0.6 < LE/Q^* < .74$ acquired in mid-latitudes by Davies and Allen (1973), suggesting the existance of a resistance to evapotranspiration in this northern environment. The average LE/Q* is similar to the Rouse and Stewart (1972) value of 0.54 achieved in a similar study at Penn Is., but under slightly drier conditions (11.1 cm rainfall in July and August compared to 18.1 cm in Churchill during this time). The fact that the Rouse and Stewart value is smaller may be due to the difference in soil moisture between the two sites. However, results from Table 1 indicate that LE/Q* in this study is not related to soil moisture. An example of this is the drying period started on July 24, with LE/Q* increasing from 0.46 to 0.71 by July 29, with no rainfall during the period. Measurement error cannot be held responsible for such a large fluctuation. This evidence emphasizes that in this particular environment, on a daily basis, evaporation responds to controls other than soil moisture.

The α' term has been described as a partial indicator of the control exerted on the yield of water to the atmosphere by the surface, and has been assumed to be a function of the surface type and the surface soil moisture. The comparison of the α' values at the different sites (Table 2) states that the daily Churchill α' values were as constant as the values of the new burn and the lichen heath, where a constancy of α' was claimed for each site. From these results, therefore, an α' value of 1.00 can be claimed as representative of this surface.

Equation 10 defines α' to be a function of several atmospheric controls, as well as the surface and radiation dominated ones. Therefore, one cannot rule out the influence of the atmosphere on α'. D_o is a surface control and cannot be measured. ρ and Cp are constant. The r_a term represents the aerodynamic resistance, dependant on and related inversely to windspeed. As the windspeed increases (actual windspeeds varied from 1.7 to 5.2 m s^{-1}), the first part of the second term, $\frac{\rho\, Cp}{r_a}$, would tend to increase α' if the value of $D_z - D_o$ was positive, or, decrease the α' value if $D_z - D_o$ was negative. $\frac{\rho\, Cp}{r_a}$ and α' are totally uncorrelated. The magnitude of $\frac{\rho\, Cp}{r_a}$ and D_z are also uncorrelated. However, α' does show a positive correlation with D_z although the relationship is statistically non-significant because of the variability of the other parameters which influence α'. Therefore, it is felt that atmospheric humidity did exert an important control over evaporation during the 1978 summer season at Churchill. The very high statistical correlation of the regression of Q^* and D_z with LE (Table 4) provides the use of a simple regression-based model for Churchill which only requires the measurement input of Q^* and wet and dry bulb temperature at 1 m.

The fact that 12% of the behaviour of LE is unexplained by the combined factors of Q^* and D_z may be due in part to measurement error. The calculation of LE requires the measurement inputs stated, each variable possessing individual measurement errors. The surface vegetation may have been behaving in some controlling but inconsistent manner. The influence of soil moisture cannot totally be ruled out. Probably, this 12% is a response to the combination of all these factors.

A major portion of the growing season was spanned during this study. With the longer measurement period, inconsistencies had a better chance to emerge. The close proximity of Hudson Bay may have acted as an unsettling factor for the atmosphere temperature and humidity as compared to other studies. Soil temperatures near the surface spanned a temperature range of 20C, allowing measurements to be taken under varying conditions. It is felt, therefore, that the explanations regarding the behaviour of LE are valid, and it is concluded that surface type and site characteristics appear to be controlling the mean α' value at this site, with the range of daily values about the mean controlled by the atmospheric humidity. The suggestion is that earlier works did not explore the possibility of atmospheric influences to improve their simplified models developed to calculate LE, and that it should be investigated in the future.

Chapter 5

SUMMARY AND CONCLUSIONS

The evaporative heat flux was determined for clear days during the summer months of 1978 using the energy balance approach. Soil moisture conditions remained moderately wet throughout the measurement period, with moisture values varying between an average of 10% at the surface, to saturation at 60 cm depth (in mid-September.) The active layer increased rapidly, starting on May 22 and reaching 1 m depth by June 25. Soil temperatures were found to respond to surface temperatures once the active layer had reached 70 cm. High energy days often resulted in soil temperatures reaching values greater than 20 C. Mean β and α values were .68 and 1.00 for the season, with daily results showing a spread of values about the mean. A seasonal value of .57 was obtained for LE/Q*, which corresponded with the .54 found by Rouse and Stewart (1972).

When LE was regressed on a set of environmental variables, the highest and most significant correlation emerged when LE was regressed on the combination of Q* and D_z. This multiple regression explained 88% of the variability in LE. Air temperature and Q* in combination could not be significantly regressed with LE. Soil moisture and soil temperatures appeared to be unrelated to daily LE.

In conclusion, atmospheric controls exert a strong influence on rates of evapotranspiration and cannot be ignored. Simplified models derived for high latitudes based on only T, (Q^* - G) and a constant α' as inputs are not applicable at this site. A regression model was derived for Churchill to show that LE at a subarctic beach ridge can be accurately estimated as a function of Q^* and D_z. It is suggested that if previous studies had accounted for atmospheric humidity, better LE results could have been calculated. As well, the constancy of α' may be found in the subarctic to generally depend, to some extent, on the humidity. Further studies are warranted to examine this phenomena in more detail.

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