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Annual Variability in Summer Evapotranspiration and Water Balance at a Subarctic Forest Site

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Evapotranspiration (*ET*) and precipitation were measured during five summers (1989-1993 inclusive) at a subarctic forest site near Churchill, Manitoba, Canada. Mean daily *ET* varied from 2.14-3.18 mm d^{-1} during the five summers, while mean daily precipitation (*P*) ranged from 1.46-3.15 mm d^{-1}. Yearly variability in summer *ET* was most influenced by availability of surface moisture, then by atmospheric conditions (i.e. temperature), and least of all by net radiation. In four of the five years total summer *ET* exceeded *P* resulting in significant soil water deficits and in the other year summer *ET* and *P* were similar in magnitude.

The use of equilibrium evaporation (*EE*) as a predictor of *ET* was explored. Separate relationships between *ET* and *EE* were computed for all five years. Three statistically dissimilar groups of equations were found: 1989/1990, 1991/1992, and 1993. A single regression equation describing all years is presented.

Introduction

Evapotranspiration, *ET*, the combined process of evaporation and transpiration, is an important component of the energy and water balances of northern watersheds. The bulk of annual *ET* losses in northern latitudes occur during the short snow-free period (Weller and Holmgren 1972; Ohmura 1982). Since *ET* is controlled by meteorological and surface factors, it varies both spatially and temporally. Thus, for a given watershed, *ET* and its relative importance in the water balance can vary significantly from year to year.

The present study reports on summertime *ET* and water balance at a forested site near Churchill, Manitoba, Canada (58°45'N, 94°04'W) during five years: 1989-1993 inclusive. The objective of the study was to examine the inter-annual variability in *ET* and precipitation and their effects on the soil-water balance at the site. Since there has been little previous *ET* research on subarctic forests (Rouse 1990), this work also compares the evapotranspiration from this site with other northern locations. Previous studies on *ET* processes at this site and comparison with a nearby tundra site were conducted by Lafleur (1992) and Lafleur *et al.* (1992).

Study Site

Terrain in the Churchill region is generally flat with many small, shallow lakes dotting the landscape. There is a transition of landscape from the coast inland, in which treeless tundra turns to a forest-tundra transition zone, then to continuous forest (Scott *et al.* 1987). Although sediments of marine origin (clay-silt) underlie most of the landscape, gravel materials are found in numerous low-lying features such as remnant beach ridges and glacial features such as kames and eskers. Peat topsoil varies from a few cm to up to 0.5 m depending on local topography and vegetation; it is generally thicker in wet and forested areas.

The study site was located in the continuous open forest approximately 40 km east of the town of Churchill, Manitoba and 15 km south of the Hudson Bay coast. In 1989 a central 18 m instrument tower was constructed 1,000 m south of the forest edge and 300 m to the west of an access road. The terrain around the tower was generally flat except for a gentle slope (≈1.2%) to the south where there were two small rain-fed lakes situated atop a glacial kame (locally known as the Twin Lakes Ecological Reserve). A survey of the forest in 1989 revealed that black spruce and tamarack were the dominant species, comprising 40% and 48% of the trees, respectively, with a small amount (12%) of white spruce. The mean tree height was 5.4 m, mean distance between trees was 3.3 m, the areal crown cover was 25%, and leaf area index of the forest was estimated at 1.5. There was a sparse layer of shrubs beneath the tree canopy consisting of birch, willow, and Labrador Tea.

The forest floor was very irregular, composed of a hummock-hollow arrangement. Local relief of the hummocks was 0.5-0.75 m. Vegetation on the hummock tops varied from moss only, to mixture of vascular plants, moss and lichen, to a thick (100-150 mm) mat of lichen, depending upon local factors such as moisture, light and exposure. Standing water or mosses covered the depressions between the hummocks, depending upon local relief. The topsoil was approximately 0.2-0.3 m of fibrous peat, overlaying marine clay-silt. The soils were usually thawed by mid-June and permanent ground ice was found only in the larger hummocks.

Climate of Churchill

It is well known that Hudson Bay has a significant impact on the climate of the surrounding terrestrial zone (Rouse 1991). The study site is located along a section of the Hudson Bay coast which has an east-west orientation into the Bay. As the dominant regional wind direction is from the north, the coastal orientation enchances the influence of the Bay on the local terrestrial climate and is the reason that despite its subarctic location it experiences low arctic climatic conditions (Rouse 1991). The annual mean maximum and minimum temperatures for Churchill are -3.2°C and -11.3°C, respectively, giving a mean annual temperature of -7.2°C. The annual total precipitation is 411 mm of which 235 mm is rain occurring mostly in the months of May to September.

This study is concerned with the summer period which is defined here as the months of June, July and August. The mean temperatures for June, July, and August are 6.2°C, 11.8°C, and 11.3°C, respectively, giving an average summer temperature of 9.9°C. Analysis of a 50-year period for Churchill indicated that for 95% of the years the summer means ranged from 7.5-12.3°C. The average summer precipitation total is 159 mm, which represents 39% of the annual precipitation. The long-term mean daily precipitation derived from the summer total is 1.73 mm d^{-1}. Precipitation tends to be more variable than temperature, with 95% of the summer totals falling between 55 mm and 263 mm.

Instrumentation and Methods

The simple water balance of a flat site can be given as

$$S = P - ET \quad \text{(mm)} \tag{1}$$

where S is storage and P is precipitation. In this study, P and ET were measured directly and S was calculated as residual relative to the start of the measurement period. Although the net horizontal flux (surface or subsurface) of water at this site was not measured here, site conditions suggested that it may not have been zero. Some subsurface flow probably occurred as water moved down off the kame slopes to the south of the site.

P was derived from three sources. Precipitation records for all years were obtained from the Atmospheric Environment Services (AES) meteorological station in Churchill (≈40 km to the west). P was measured during all years with a tipping-bucket rain gauge located at a site on the open tundra 2 km north of the forest tower. Finally, daily precipitation was measured in a small clearing near the forest tower in 1991 and 1992 with two manual gauges. The two manual gauges compared closely with each other and with the tipping bucket located outside the forest. In general, the tipping-bucket gauge compared well with the AES record, but there were days when the two measurements disagreed. These were usually associated with convective storm activity.

ET was calculated via the Bowen ratio energy balance method (Thom 1975; Halliwell and Rouse 1989), where profiles of temperature and humidity are used to compute the ratio of sensible to latent heat flux, known as the Bowen ratio, β. β is then combined in the surface energy balance to give ET as follows

$$ET = \frac{R_n - G}{(1+\beta) L_v} \quad (2)$$

where R_n is net radiation (J s^{-1} m^{-2}), G is the soil heat flux in (J s^{-1} m^{-2}), and L_v is the latent heat of vaporization (J kg^{-1}). The procedure for applying Eq. (2) involves taking 30 min averages of R_n, G, and β to calculate ET, summing these over a 24-hour period and multiplying by the number of seconds in 30 min to obtain daily ET in mm d^{-1}.

Similar instrumentation was maintained during each year of the study. R_n was measured with a net radiometer (CN1, Middleton Instruments, Melbourne, Australia) mounted at the top (18 m) of the tower. G was measured with four soil heat flux transducers (CN3, Middleton Instruments) placed 3 cm beneath the surface in representative terrain types. The signals from the four instruments were averaged. Since these instruments often underestimate the heat flux in peat soils (Halliwell and Rouse 1987), an attempt was made to correct the heat flux plate measurement of G via calorimetric computations of total heat storage obtained from soil temperature profiles and soil heat capacity (for details on the method see Halliwell and Rouse 1987). Wet- and dry-bulb temperatures were measured with ventilated psychrometers at heights of 9 m, 10.5 m, 12 m, 13.5 m and 15 m above the forest floor. The minimum fetch to height ratio for the highest sensor was 100:1 to the north of the tower. In all other directions the fetch ratio exceeded 300:1. Wind speeds were measured with cup anemometers (R.M. Young Co., Michigan, U.S.A.) at the same heights as the psychrometers and wind direction was measured at the top of the tower. All instruments were recorded on an electronic data logger (CR7, Campbell Scientific, Logan, Utah) once every 5 s and averaged every 30 min.

Bowen ratios were calculated using a computer program which plots vapour pressure, e_a, against potential temperature, θ, for all sensor levels. The half-hourly plots were visually inspected in order to eliminate suspect points in the profile (such as when a wet-bulb sensor had dried out) from the analysis. Following Thom (1975), β was computed from the regression slope for all remaining levels as $\Delta\gamma\theta/\Delta e_a$, where γ is the psychrometric constant (≈66.2 Pa K^{-1}). ET was then calculated as in Eq. (2). The error in ET was assessed as follows. Although the R_n measurement is usually assigned an error of ±5%, it is recognised that an additional error in ET can arise because of spatial variability in R_n in the up-wind regions from the tower. Unfortunately, there were no spatial measurements of R_n available with which to assess this error. Because of the complexity of the forest floor G was the most uncertain measurement in this study. It was assigned an error of ±20%. Combining the errors in R_n, G, and the typical error in β (±11%), the mean error in ET was found to be ±12%.

This study considers the concept of equilibrium evaporation, EE (Slatyer and McIlroy 1961), given as

$$EE = \frac{s(R_n - G)}{L_v(s+\gamma)} \tag{3}$$

where s is the slope of saturation vapour pressure *versus* temperature (Pa K^{-1}). Daily values of EE were computed as described above for ET. Since its inception, EE has been the subject of much discussion. It was originally defined as the rate of evaporation from a wet surface into a saturated atmosphere or, more loosely, the lower limit to potential evaporation (Slatyer and McIlroy 1961). EE was later recognized as the rate of evaporation rate which would occur whenever the vapour deficit at the surface was equal to that in the atmosphere (Wilson and Rouse 1972), as well as the evaporation rate achieved downwind of an infinitely large surface with constant radiation and surface resistance (McNaughton 1976). More recently, McNaughton and Jarvis (1983) suggested that it is the evaporation occurring when the surface is completely decoupled from the regional air stream, which is typical of very smooth surfaces with small surface resistance. A popular variant of the equilibrium concept was proposed by Priestley and Taylor (1972), where the potential evaporation from a well-watered surface could be described by EE modified by some constant parameter α (empirically defined as 1.26). Despite the confusion over its definition, EE has been shown to closely approximate actual ET in many environments, including a number of arctic and subarctic sites (Kane *et al.* 1990). The present study only attempts to show that EE is a useful approximation of ET for this forest site, whereby EE can be used to estimate missing data from the ET record.

Data Collection Periods

Data were collected for periods of variable length during the summers of 1989, 1990, 1991, 1992, and 1993 (Table 1). The earliest date of measurement was Julian day (*jd*) 161 (June 9) in 1992, the latest date was *jd*243 (August 31) in 1989, and the longest record was 76 days from *jd*168 (June 17) to *jd*243 (August 31) in 1989.

Meteorological conditions during these five summers varied considerably (Fig. 1). Comparing the mean temperature for the period of instrumental record to the normal summer temperature (9.9°C), two years, 1989 and 1991, were significantly warmer than normal, 1990 was only slightly warmer than the normal, and 1992 was significantly cooler than normal (Table 2). Mean daily precipitation was significant-

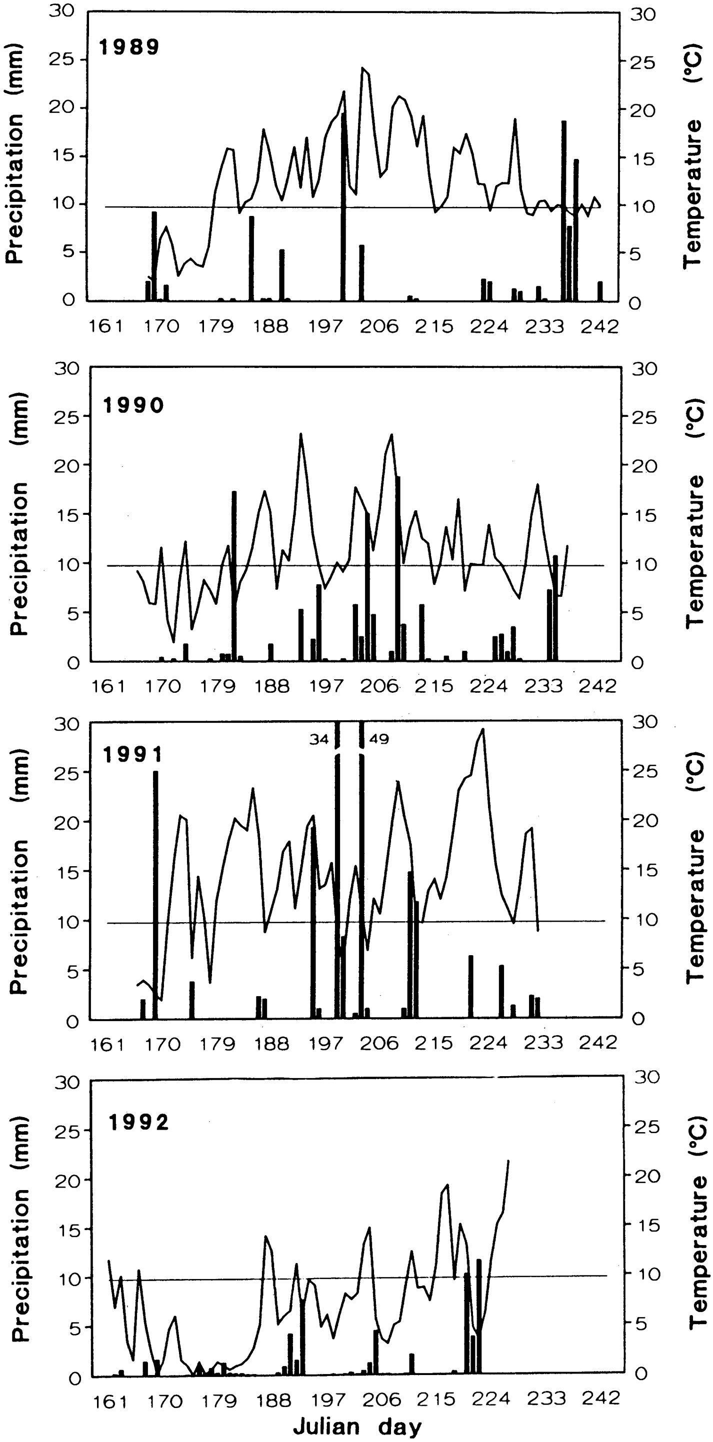
1989
1990
1991
1992
34
49
Precipitation (mm)
Temperature (°C)
0
5
10
15
20
25
30
161
170
179
188
197
206
215
224
233
242
Julian day

Fig. 1.

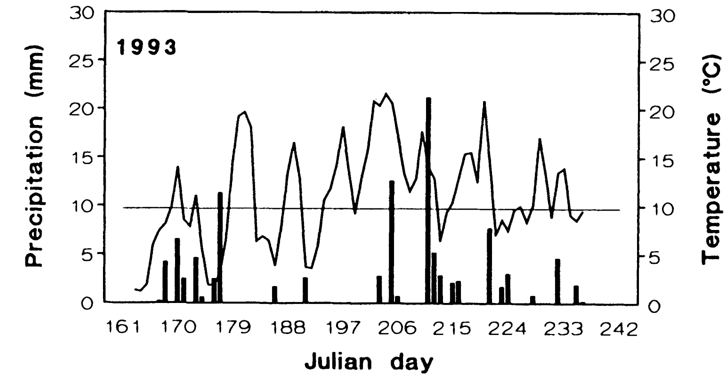

Fig. 1. Meteorological records for study periods. Bars indicate daily precipitation and thick solid line is mean daily air temperature. The thin solid line is mean summer (June, July, August) temperature (9.9°C).

Table 1 – Dates of study periods. *jd* represents Julian days and *n* is total number of days

year	dates	*jd*	*n*
1989	June 17 – August 30	168-242	76
1990	June 15 – August 25	166-237	72
1991	June 15 – August 18	166-230	65
1992	June 9 – August 14	161-227	67
1993	June 12 – August 24	163-236	74

Table 2 – Comparison of meteorological conditions during the study periods. T is mean daily air temperature, P_T is total precipitation, P_d is mean daily precipitation (*i.e.*., total precipitation during the period divided by number of days in the study period), and #rd is the number of days with measurable precipitation ($\geq$0.2 mm)

year	T (°C)	P_T (mm)	P_d (mm d^{-1})	#rd
1989	12.2	106	1.4	20
1990	11.0	116	1.6	30
1991	14.4	189	2.9	21
1992	7.0	57	0.9	26
1993	11.0	106	1.4	23

ly less than the normal (1.64 mm d^{-1}) during 1992, similar to normal in 1990 and 1993, and significantly greater than normal in 1991. Although mean daily precipitation during 1989 was near normal, the distribution of rain was such that 80% of the total fell in two events, the largest of which occurred at the end of the measurement period (Fig. 1). Therefore, compared with other years the 1989 experimental period was drier than normal. On this basis temperature and precipitation characteristics for the four years were qualitatively described as follows: 1989 was hot and dry, 1990 was normal, 1991 was hot and wet, 1992 was cool and dry, and 1993 was near normal. The two years with the greatest contrast were 1991 and 1992, where 1991 was the warmest summer in the previous 50 years and 1992 was the coolest summer during the same period. 1991 was a very wet summer, while 1992 was a moderately dry summer.

Results

Evapotranspiration and Equilibrium Evaporation

ET for the five summer periods varied moderately between years (Table 3). Statistical analysis of the mean daily values indicated the following: the 1989 and 1993 means were similar, the 1990 and 1992 means were similar and significantly greater than for 1989 and 1993, and the 1991 mean was significantly larger than for all other years. The smallest mean daily *ET* was 2.1 mm d^{-1} which occurred in 1989, a warm and dry summer. The largest mean daily *ET* (3.2 mm d^{-1}) occurred in 1991 and was 49% larger than in 1989. 1991, which was the warmest and one of the wettest years on record at Churchill, also experienced the single largest daily *ET* (5.99 mm d^{-1}) during the study. The evaporation efficiency of the site (ET/R_n, where R_n is expressed as an equivalent depth of water) varied from 49% in 1989 to 65% in 1991 (Table 3).

Table 3 – Summary of mean daily evapotranspiration, *ET*, minimum observed daily evapotranspiration, ET_{min}, and maximum observed daily evapotranspiration, ET_{max}, for the five study periods. Mean daily *ET* is given as a per cent of net radiation, R_n, where R_n is expressed as an equivalent depth of water by dividing the daily R_n by the latent heat of vaporization. *n* is number of days

year	*n*	*ET* (s.d.) mm d^{-1}	ET_{min} mm d^{-1}	ET_{max} mm d^{-1}	ET/R_n (s.d.) %
1989	76	2.1 (0.97)	0.22	3.49	49 (11)
1990	63	2.6 (0.91)	0.63	4.34	53 (14)
1991	50	3.2 (1.36)	0.41	5.99	65 (13)
1992	67	2.6 (0.94)	0.66	4.28	53 (09)
1993	67	2.3 (1.00)	0.59	4.56	52 (11)

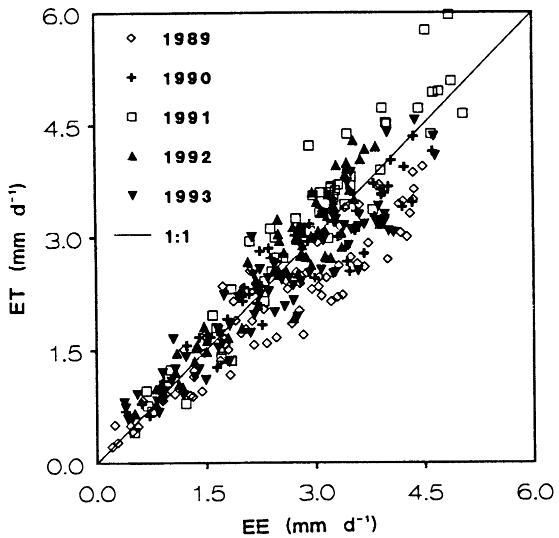

Fig. 2.
Relationship between daily evapotranspiration, *ET*, and equilibrium evaporation, *EE*.

Fig. 2 shows the relationship between daily *ET* and *EE* for all years. While there appears to be an overall clustering of points about the one-to-one line, the linear relationships between *ET* and *EE* varied between years (Table 4). Slopes for these relationships ranged from 0.77 (1989) to 1.09 (1991) and the intercepts ranged from 0.02 (1991) to 0.48 (1990). The goodness of fit (r^2) was similar for all data sets, ranging between 0.85-0.90. A statistical comparison of the regression slopes revealed that the slopes for the 1989 and 1990 regressions were not significantly different (at α=0.05 level) and that there was no significant difference between the slopes for the 1991 and 1992 regressions. However, these two sets of equations (1989/90 and 1991/92) were statistically different, and both sets were different from the 1993 slope. A single relationship constructed using data for all years resulted in only a slightly poorer fit (r^2=0.83) and lower accuracy (s_y=0.45 mm d^{-1}).

Table 4 – Regression equations for actual evapotranspiration, *ET, versus* equilibrium evaporation, *EE* for the five study periods as shown in Fig. 2. *n* is number of days, r^2 is the coefficient of determination, and s_y is the standard error of the *y* estimate

year	regression equation
1989	ET=0.21+0.77EE (mm d^{-1}) n=76, r^2=0.87, s_y=0.36
1990	ET=0.48+0.78EE (mm d^{-1}) n=63, r^2=0.85, s_y=0.29
1991	ET=0.02+1.09EE (mm d^{-1}) n=50, r^2=0.92, s_y=0.40
1992	ET=0.10+1.01EE (mm d^{-1}) n=68, r^2=0.88, s_y=0.33
1993	ET=0.31+0.81EE (mm d^{-1}) n=74, r^2=0.88, s_y=0.35
All years	ET=0.20+0.89EE (mm d^{-1}) n=319, r^2=0.84, s_y=0.45

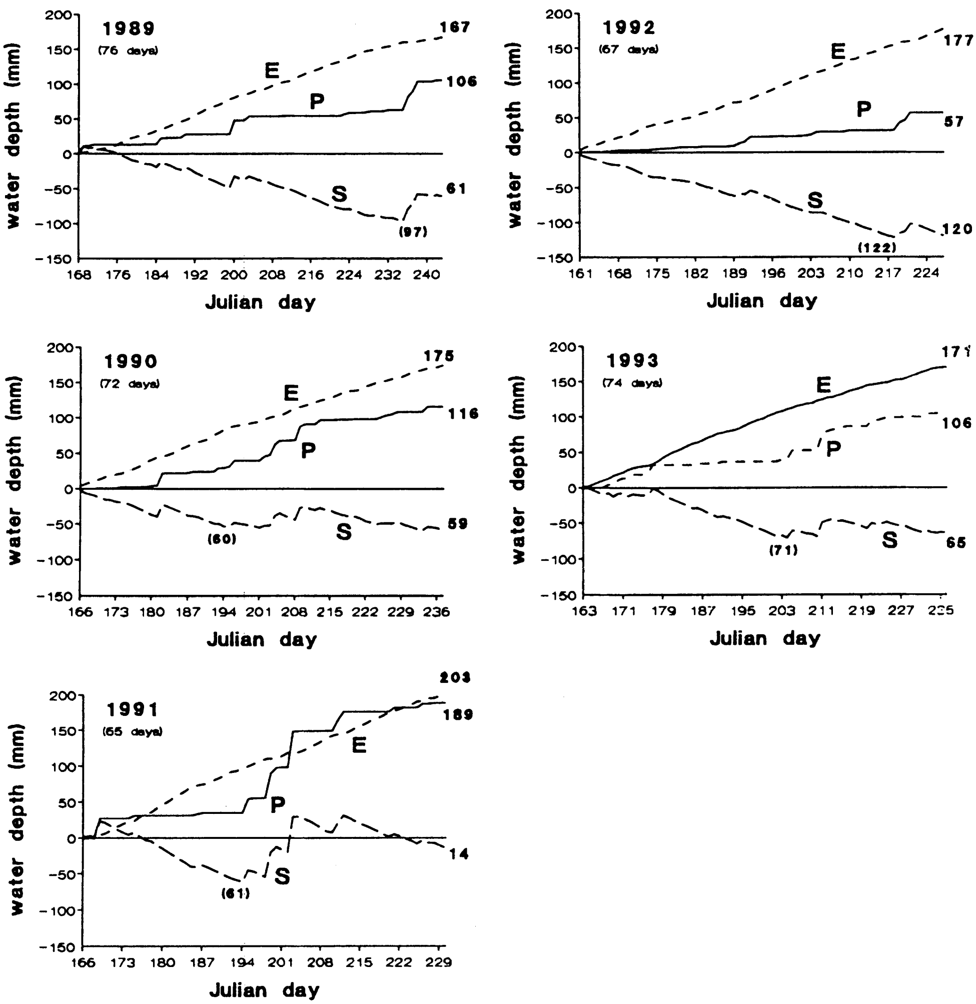

Fig. 3. Cumulative water balances for the study periods. *ET* is evapotranspiration, *P* is precipitation, and *S* is the soil water deficit ignoring horizontal water flux. Values to the right are the accumulated totals and values in brackets are the maximum water balance deficit.

Water Balances

Water balances for the site were constructed for the five study periods (Fig. 3). During two years (1990 and 1991) evapotranspiration data were missing on some days because of instrument failure. The number of missing days was 9 in 1990 and 15 in 1991. In order to complete these records, *ET* for the missing days was calculated using the relationship between *ET* and *EE* for the appropriate year (Table 4). *EE* was computed with R_n and air temperature obtained from a nearby

(≈2 km) tundra site. It has been shown previously that R_n is similar between these two sites (Lafleur *et al.* 1993). G for these days was calculated as a per cent of daily R_n based upon the ratio G/R_n for ten-day periods on either side of the missing days. Precipitation for the water balances was taken from the on-site gauges when they were available, when there was no precipitation record form the site the tipping-bucket record at the tundra site was used, and finally, if neither of these were available, the Churchill weather office record was used.

Moderate summertime water deficits (60-70 mm) developed during three of the study years (1990, 1991, and 1993) and substantial water deficits (>95 mm) were recorded in two years (1989 and 1992) (Fig. 3). Only during the very wet year (1991) was a positive water balance accumulated and then only for a short period of time from mid to late July. In all years except 1991 there was a significant water deficit (>50 mm) accumulated by the end of the study period. It is clear in Fig. 3 that the pattern of the accumulated water balance was strongly affected by the precipitation regime. As well, yearly differences in the summertime accumulated water deficit were largely a function of variability in total summer precipitation and less dependent upon variability in ET. Mean daily precipitation for the five study periods varied by a factor of 3 (Table 2). The distribution of rain was also important, as evident in 1989 which had normal rainfall totals but experienced significant drying in mid-July and early August. ET proceeded at a more constant rate during summer months and the difference between the smallest and largest mean daily ET for the five summer periods was only about 1.5 times (Table 3). For the five study years, the ratio of ET/P expressed as a percentage was 156%, 138%, 108%, 310% and 161% for 1989, 1990, 1991, 1992 and 1993, respectively.

Discussion and Conclusions

The magnitude and range of mean daily summer ET from this subarctic forest site is comparable with that from many arctic environments. Kane *et al.* (1990) cite several studies of evaporation from high latitude sites with values generally ranging between 1.0 and 4.5 mm d^{-1}. ET/R_n ranged between 49% and 65% for this site which is also comparable to previous work on tundra surfaces (Weller and Holmgren 1974, Ohmura 1982, Kane *et al.* 1990), but it is lower than typical values for arctic lakes (Marsh and Bigras 1988, Bello and Smith 1990).

The maximum difference in mean daily ET between the five summers of study was only 49% despite a considerable difference in meteorological conditions. Compared with the works surveyed in Kane *et al.* (1990), this yearly variability in ET is similar to arctic tundra sites and it is larger than for arctic lakes. The inter-annual variability in ET is a function of changes in several factors and although the available data set was not large enough to provide conclusive results, some generalizations can be made. R_n was remarkably similar in the five study years. Average daily net radiation varied from 11.18 MJ m^{-2} in 1993 to 12.38 MJ m^{-2} in 1992. Thus, changes in surface and atmospheric factors were probably more important in determing the yearly differences in summer *ET*. The highest mean *ET* occurred in the wettest and warmest year, 1991 (Fig. 1). *ET* was lowest during the hot and dry summer of 1989, when surface moisture became limiting during the mid summer. Although 1992 had a lower summer precipitation, it was distributed evenly throughout the measurement period and surface moisture did not become as limiting as in 1989.

The variation in the relationship between *ET* and *EE* between the five summers is somewhat puzzling. For example, the slope coefficients for the 1989 and 1990 data were statistically similar. However, 1989 was warmer and drier than 1990. In these years the slope of the relationships were strongly influenced by the large ($\geq$3.0 mm d^{-1}) *EE* values which were consistently greater than *ET* values (Fig. 2). The wettest and warmest year, 1991, had the largest slope coefficient (1.09), but it was not statistically different from that for 1992 which was cool and dry by comparison. No consistent trend was found between summer meteorological conditions and the nature of the relationship between *ET* and *EE*. This result is most likely related to the empirical nature of this type of exercise which cannot account for the complexities of the *ET* process. It is clear that a model of evaporation based upon *EE* from only one year of data has limited value for application to the prediction of *ET* for other summers. An expression based upon several years of data would be preferable.

Variability in the summer water balance at this site is strongly controlled by the stochastic nature of the precipitation regime (Fig. 3). Summer *P* varies considerably more than does *ET*. For this site summer *ET* usually exceeds summer *P*; only during the wettest year (1991) were *ET* and *P* totals similar. *ET*/*P* ratios varied from 108%-310% at this site. Similar variability in the *ET*/*P* ratio has been reported for other northern locations. For example, Roulet and Woo (1986) reported that *ET*/*P* for a watershed at Baker Lake, N.W.T. was 171% and 607% in two successive summers. Kane *et al.* (1990) measured the summer water balance of a tundra watershed in nothern Alaska and found that *ET* was smaller than *P* in two of three years, with *ET*/*P* ranging from 77%-124%. In high arctic watershed studies, Rydén (1977) found that *ET* was only marginally greater that *P* in three summers on Devon Island and data from Woo (1983) for Cornwallis Island showed that summer *ET*/*P* varied from 61%-135% over a six-year period. The general conclusion one might draw from this brief review is that, while it is generally accepted that both *ET* and *P* decrease northward of the treeline (Prowse 1990), the relative importance of *ET* in the summer water balance tends to decrease moving northward of the treeline. The same conclusions are not true of the annual water budget. In all studies, annual *ET* is always less than annual *P*.

In conclusion, the major points of this paper can be summarised as follows:

1) Summer ET at this subarctic forest site is usually greater than P, resulting in a soil moisture deficit. This behaviour is most likely a function of the regional climate and site characteristics which control ET.
2) The annual variations in the summer water balance are a function of both ET and P, but P typically has a larger influence because of its stochastic nature.
3) The EE model can be used to predict daily ET. However, given the variation in the relationships found here, a model developed from data obtained in one year could not be applied with confidence in other years. Several years of data are needed to develop a more general relationship. The EE model is probably best suited, as used in this study, for estimating missing data from a record where there is adequate measured data to construct the model.

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