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Evapotranspiration from a Wetland Complex on the Arctic Coastal Plain of Alaska

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Evapotranspiration (*ET*) from an arctic coastal wetland near Prudhoe Bay, Alaska, was studied during the summers between 1994 and 1996. The purpose of the study was to compare different *ET* models and to gain a better understanding of evapotranspiration from arctic wetlands. The models used to obtain *ET* from the watershed were the Bowen ratio energy balance (BREB), Priestley-Taylor (PT), Penman-Monteith (PM), Penman Combination (PC), energy balance (EB), water balance (WB), and WB based on Time Domain Reflectometry (TDR). For one of the ponds, evaporation determined by the EB, PT, PC, BREB, WB, and the aerodynamic (AD) methods were also compared. *ET* during the summer snow-free period for the watershed averaged 1.45 mm/day obtained via the BREB model. Evaporation from all ponds after spring snowmelt averaged 3.11 mm/day (obtained via the WB). Evaporation rate from ponds was on average twice that of the tundra as a whole. Latent heat flux was the dominant energy sink in wetlands and ponds, whereas sensible heat flux dominated in the drier upland area. The PT and PM models compared well to the BREB (used as the standard of comparison for *ET*) for 1994 and 1995, once parameters were properly calibrated using 1996 data. The BREB compared well with independent values of *ET* from the water balance and eddy correlation methods. For the pond, the EB, BREB, WB, PT, and AD methods gave very similar evaporation results for the summer.

Introduction

In the Arctic, hydrology plays an important role in the understanding of numerous physical, chemical, and biological processes from climate to nutrient dynamics to carbon and methane fluxes. A major component of the hydrology of wetlands, including those in the Arctic, is evapotranspiration, *ET* (Lafleur 1990; Souch *et al.* 1996). On the Alaskan Arctic coastal plain, *ET* can be the dominant water loss mechanism, the magnitude of which usually exceeds summer precipitation (Rovansek *et al.* 1996). In addition to its important role in the hydrologic cycle, *ET* also provides the bridge that connects the water and energy budgets (Bello and Smith 1990). Therefore, improving our understanding of evapotranspiration will improve our knowledge of the hydrologic and thermal processes of the Arctic. Many studies have been conducted on evaporative processes from arctic ecosystems (*e.g.* Mather and Thornthwaite 1958; Addison 1972; Wight 1973; Ohmura 1982; Marsh and Bigras 1988; Kane *et al.* 1990; Lafleur 1990; Rouse *et al.* 1992; Wessel and Rouse 1994; Rouse 1998). The majority of these have taken place in the Canadian Arctic, but some studies have been performed in the Alaskan Arctic.

A small wetland complex located in the Arctic coastal plain of Alaska has been the subject of ongoing hydrologic studies since 1992 (Rovansek *et al.* 1996; Robinson 1995; Rovansek 1994). The present study, which extends from 1994 to 1996, is aimed at testing and comparing several techniques of calculating evapotranspiration as well as at advancing our understanding of evapotranspirative processes from this region. In addition, this study increases the baseline data relevant to *ET* and advances our basic understanding of the hydrology of arctic wetlands.

The methods for obtaining *ET* from the small watershed that were investigated are the following: Bowen ratio energy balance (BREB), Priestley-Taylor (PT), Penman-Monteith (PM), Penman Combination (PC), energy balance (EB), water balance (WB), and WB based on Time Domain Reflectometry (TDR). The BREB was chosen as the standard of comparison for the models. Two reasons support this choice. First, it was not possible to use the more rigorous EB model as the standard of comparison since it could not be applied during 1994 and 1995. Second, the BREB model has been used extensively as the standard of comparison and calibration model for *ET* estimates from varied surface types (*e.g.* Abbaspour 1991; Tomlinson 1996; Wessel and Rouse 1994; Moore *et al.* 1994; Lafleur 1990; Rouse 1982). It should be kept in mind, however, that the BREB does not provide an actual measurement of evapotranspiration; it just provides an estimate of *ET*. However, results from many other studies (*e.g.* Assouline and Mahrer 1993; Tomlinson 1996; Malek and Bingham 1993) have compared the BREB results to other independent methods of obtaining *ET* such as eddy correlation, lysimeter, and water balance methods. From results of such studies it can be determined that the BREB is capable of providing *ET* values that are in the range of actual measurements of *ET*.

The evaporation from one of the ponds in the area was also investigated during

1996. For this pond, evaporation obtained from the PC, WB, PT, BREB, and the aerodynamic (AD) models was compared to results from the energy balance method.

Study Site Description

The study site is a small wetland complex (22.4 hectares) on the Arctic Coastal Plain of Alaska, near the Prudhoe Bay oil field, named Betty Pingo (Fig. 1). Within the Betty Pingo site there is a small watershed of 8.15 ha, which is the focus of this study (Fig. 2). The center of the watershed is situated at 148^o 53' 44.5" west longitude and 70^o 16' 46.9" north latitude. Approximately 78% of the watershed is characterized as a wetland (Rovansek 1994). The dominant landforms in the wetland area are low centered polygons, strangmoor ridges, and small thaw ponds less than 1 m in depth. The remaining 22% of the watershed is slightly higher (about 1 m) and is a drier, better-drained upland tundra dominated by high-centered polygons and ice wedges. The vegetation in the watershed is dominated by sedges, mainly *Carex* and *Eriophorum* species with an average height of 10 to 15 cm. In the drier upland areas the sedges are intermixed with lichen, mosses, dwarf *Salix* and *Betula* species, and a variety of flowering herbaceous species (Walker *et al.* 1980; Robinson 1995; Rovansek 1994).

The Betty Pingo site has a very flat surface topography, with slopes of approximately 0.2%. The active layer is on average 40 to 50 cm in depth. The soil within the active layer is mainly organic peat. Hydraulic conductivity is on the order of 10^{-4} cm/sec, and the specific yield is approximately 0.1 (Rovansek 1994). Beneath the thin active layer, Osterkamp *et al.* (1985) found the permafrost to be 680 m in thickness.

Peak runoff occurs during snowmelt; afterwards the flow gradually decreases, ceasing by mid-June. The melt water from the watershed drains towards a thaw lake basin through a notch in the scarp surrounding the basin. Roughly half of the snowpack exits the watershed as runoff (Robinson 1995). The remainder either sublimates or goes into surface storage in ponds or subsurface storage to make up for water deficits from the previous summer's evapotranspiration.

The study site is characterized by nearly continuous winds. Summer wind speeds during the three years of study reached up to 16 metres/second. The predominant wind direction is usually from the east-northeast. The fetch over wetland surfaces in the study site is more than a kilometre from this direction. There is a road and a 1 m diameter pipeline, elevated about 2 m above the surface, about 0.5 km away in the downwind direction. Beyond that is a relatively flat uniform surface for many kilometres.

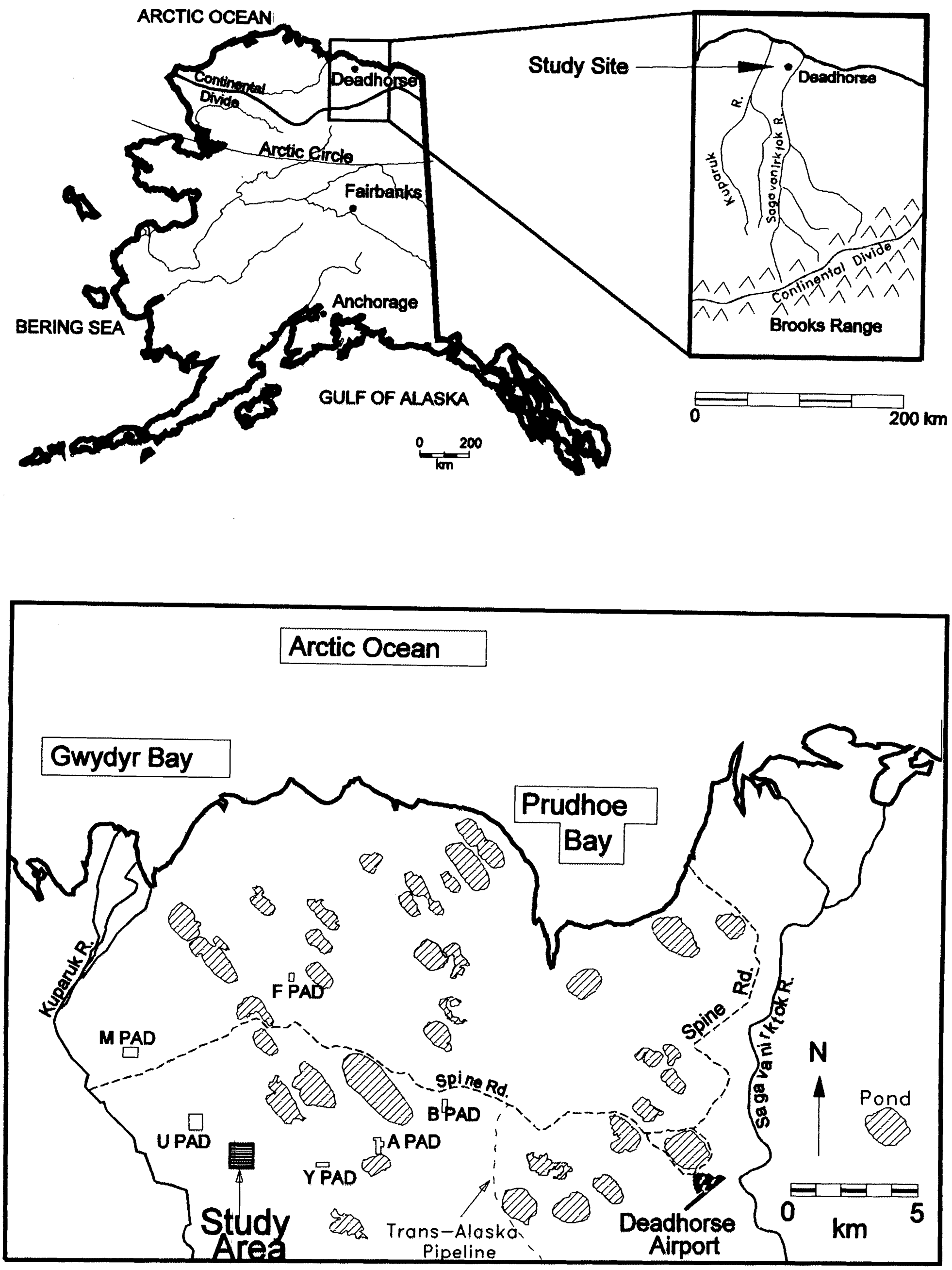

Fig. 1. Maps with location of study site at Prudhoe Bay, Alaska.

Instrumentation and Data Collection Methods

Data collection for the present study began in 1994 and extended to 1996. Meteorological data was collected from a 10 m tower located at the center of the site. The data collected at the 10 m tower were 1 m and 10 m air temperature and relative humidity (RH), using a Campbell Scientific Model 207 probe. The air temperature measurement accuracy of this probe is typically $\pm 0.1^{\circ}$ C. The RH sensor of the probe is a Phys-Chem Scientific PCRC-11 which has a reported accuracy typically better than 5% over the range of 12 to 100% RH. Wind speed is measured with Met-One 014A cup anemometers at 1 m and 10 m, and wind direction is obtained with a Met-One 024A sensor. The accuracy for wind speed and direction are ±1.5% (or 0.125 m/s) and $\pm 5^{\circ}$, respectively.

Net radiation at the site was measured at 1.5 m over wetland and upland areas using a REBS Q6 net radiometer with a reported accuracy of ±5% over vegetated surfaces (worst case scenario). The radiometers were mounted on 1.5 m towers separate from the main 10 m meteorological tower to minimize the interference of the tower on the sensors' field of view. In addition, during 1996, a net radiometer and an air temperature/RH probe were placed over one of the larger ponds in the area (Pond S11) at heights of 1.1 m and 0.9 m, respectively.

Surface and subsurface temperatures were measured using YSI 44034 precision thermistors (or equivalent) with an accuracy and interchangeability of $\pm 0.1^{\circ}$ C. These thermistors were placed at different depths in the active layer. Thermistors were also placed at different depths in pond S11 during 1996: water surface, 6 cm below water surface, pond bottom, and 4 cm below pond bottom in the sediment. Surface thermistors were shaded to avoid direct sun heating. All thermistors were calibrated in the lab using an ice-water bath at 0° C. The calibration was also checked in the field during phase change in spring and summer.

Data from all the above-mentioned instruments were measured at one-minute intervals and averaged or totaled hourly and stored on Campbell Scientific CR10 data loggers. Evapotranspiration was calculated using this hourly data and then summed for the day.

Soil moisture data was collected using Time Domain Reflectometry (TDR). Data was collected with a Campbell Scientific automated TDR soil moisture measurement system. The system uses a Tektronix 1502B TDR Cable Tester (the reflectometer), which is connected to a Campbell Scientific CR10 data logger. The 30 cm probes were installed horizontally in four sites with ranging moisture regimes (slightly dry to very wet). The probes were primarily installed near the surface (5 cm, 10 cm, 15 cm) because in the wetter areas the deeper soil remained saturated throughout our experience in this study area. In the dry site, deeper probes were also installed to make the profile measurements at 5, 10, 15, 20, 25, 30, 35, and 40 cm. TDR data was collected every two hours, except for several days in the beginning of the season when it had to be collected from every 12 to every 4 hours. The internal

Campbell Scientific algorithm for obtaining soil moisture was not used in this study since it was not properly calibrated for organic soils; instead the entire TDR wave forms were stored in the CR10 data logger. These were later reduced to soil moisture values following the procedure outlined in Stein and Kane (1983) using a computer program developed by the author. The program uses a calibration curve of soil moisture *versus* apparent dielectric constant developed by Meade (personal communication) for peat soils from the Betty Pingo study site. The equation has the following form

$$SM = (7\times10^{-5}\, KK^3 - 0.018\, KK^2 + 2.1778\, KK + 5.0982)100 \tag{1}$$

where

SM – Soil moisture (% by volume)
KK – Apparent dielectric constant (obtained from the TDR trace)

The water table at the site was monitored using a series of shallow wells on land surfaces and staff gages in all ponds of the watershed. The well and staff gage levels were read every three days. The staff gages had 6.1 mm (0.02 ft) divisions, but readings were obtained to the nearest half division (3.05 mm). Well level readings were taken with an accuracy of 2 mm.

Precipitation was measured at the site with a standard 20 cm (8-inch) diameter tipping bucket rain gauge with an alter shield. During 1996, potential evaporation was measured by means of a standard class A evaporation pan.

Theory

The following is a brief description of the models used for calculating evapotranspiration in this study. All the terms and units are defined in Table 1. For all energy-based approaches, the latent heat flux can be converted to evapotranspiration (in units of depth of water over time) through the relationship

$$ET = -\frac{Qe}{\rho_w \lambda} \tag{2}$$

For further detail refer to Mendez (1997).

Aerodynamic Approach (AD)

This approach takes into account turbulent transfer mechanisms and vertical gradients of temperature and vapor pressure in order to obtain sensible, Qh, and latent, Qe, heat fluxes. In this context, the gradient (a) and bulk (b) forms of the fluxes are (Moore 1983)

$$Q.h = \rho\, C_{pa}\, Kh \frac{dT}{dz} = \rho\, C_{pa}\, Dh (T_a - T_s) \tag{3.1}$$

$$Qe = \frac{\rho}{\gamma} C_{pa} Ke \frac{de}{dz} = \frac{\rho}{\gamma} C_{pa} De (e_a - e_s) \tag{3.2}$$

(a) (b)

The bulk exchange coefficients, *Dh, De*, can be obtained as functions of wind speed, roughness lengths, and the Monin-Obukhov stability parameter (Launiainen 1995; Lo 1966).

Energy Balance Method (EB)

In this model, latent heat flux is obtained as the residual term of the energy balance. Taking all fluxes to be positive towards the surface, and assuming negligible advection, we can write the surface energy balance as (Kane *et al.* 1990)

$$-Qe = Q_{net} + Qh + Qg \tag{4}$$

In Eq. (4), net radiation is obtained from net radiometer measurements at 1.5 m above the surface. The ground heat flux is obtained from soil temperature gradients as

$$Qg = Ks \frac{T_D - T_S}{D} \tag{5}$$

The convective heat flux term can be obtained from the aerodynamic approach Eq. (3.1).

When applying the energy balance for the pond, a term that accounts for the change in water heat storage, Q_w, needs to be added to Eq. (4) (Bello and Smith 1990)

$$Q_w = C_w \rho_w z_p \left(\frac{\Delta T}{\Delta t}\right) \tag{6}$$

Bowen Ratio Energy Balance (BREB)

This approach, originally developed by Bowen (1926), provides a slightly simpler way of obtaining *ET* from the energy balance since it eliminates the need to calculate the *Qh* component of the energy budget. Thus, complicated atmospheric stability corrections, which are related to the computation of the bulk transfer coefficient for sensible heat, need not be considered. The BREB method combines the energy balance equation Eq. (4) with the flux-gradient equations Eq. (3) for heat and vapor to obtain an expression for the latent heat flux, which is as follows

$$-Qe = \frac{Q_{net} + Qg}{1 + \beta} \tag{7}$$

Where the Bowen ratio, β, is defined as the ratio of *Qh*/*Qe* (Bowen 1926)

$$\beta = \gamma \left(\frac{Kh}{Ke}\right) \left(\frac{dT}{de}\right) \tag{8}$$

Table 1 – Definition of terms and units

Variable	Definition	Units (SI)
C_{pa}	Specific heat of air	J/kg/°C
ρ	Air density	kg/m^{-3}
Kh	Eddy diffusivity for heat	m^2/s
Ke	Eddy diffusivity for water vapor	m^2/s
γ	Psychometric constant = $(P_a\ C_{pa})/(\lambda\ 0.622)$	Pa/°C
P_a	Air pressure	Pa
λ	Latent heat of vaporization, $\lambda = (2.501 \times 10^6 - 2370 \times T_s)$	J/kg
Dh	Bulk exchange coefficient for heat	m/s
De	Bulk exchange coefficient for water vapor	m/s
T_a	Air temperature at height z above the ground	°C
T_s	Surface temperature	°C
e_s	Surface vapor pressure	Pa
k	von Karman's constant	0.41
z	Height of meteorological measurements	m
z_{om}	Roughness length for momentum	m
z_{oh}	Roughness length for heat ≈ $(z_{om}/10)$	m
u_z	Wind speed at height z	m/s
e_a	Air vapor pressure at height z	Pa
e_{as}	Saturation air vapor pressure	Pa
Qnet	Net radiation	W/m^2
Qe	Latent heat flux	W/m^2
Qh	Convective/sensible heat flux	W/m^2
Qg	Ground heat flux	W/m^2
Q_w	Change in water heat storage in time period Δt	W/m^2
ET	Evapotranspiration	m/s
ρ_w	Water density	kg/m^3
Ks	Thermal conductivity of the soil	W/m °C
T_D	Soil temperature at depth D	°C
D	Depth of soil temperature measurement	m
C_w	Heat capacity of water	kJ/kg °C
z_p	Average depth of the pond	m
$\Delta T/\Delta t$	Average hourly temperature change in the lake	°C/3600s
β	Bowen ratio	-
Δ	Slope of the saturated vapor pressure curve	Pa/°C
r_c	Canopy resistance	s/m
r_a	Aerodynamic resistance	s/m
d	Zero plane displacement height	m
α	Priestley-Taylor parameter relating actual and equilibrium evaporation	-
P	Precipitation	m/s
ΔS	Change in storage in soil or ponds (positive when storage decreases)	m/s
dT/dz	Vertical temperature gradient	°C/m
de/dz	Vertical vapor pressure gradient	Pa/m

In the absence of net horizontal advection, the bulk transfer coefficients, *Dh, De*, or eddy diffusivities, *Kh, Ke*, for heat and water vapor can be assumed to be equal (Tomlinson 1996), and Eq. (8) simplifies to

$$\beta = \gamma \left(\frac{dT}{de} \right) \tag{9}$$

The temperature and vapor pressure gradients in Eq. (9) can be approximated from measurements at two heights above the surface, provided that the instruments are within the internal boundary layer (Tomlinson 1996; Heilman and Brittin 1989).

Penman-Monteith Model (PM)

The Penman-Monteith model (Monteith 1965) is a refinement of the original Penman Combination equation. It incorporates aerodynamic and energy balance components as does the former, but the Penman-Monteith model also takes into account the physiological control of plants over the transpiration process through a canopy resistance term, r_c. The Penman-Monteith equation has the following form (Monteith 1965)

$$-Qe = \frac{\{\Delta(Q_{net}+Qg)+\left[\rho\, C_{pa}(e_{as}-e_a)\right]/r_a\}}{\{\Delta+\gamma\left[(r_c+r_a)/r_a\right]\}} \tag{10}$$

The fluxes are defined the same as in Eq. (4), and the aerodynamic resistance term is defined as (Campbell 1977)

$$r_a = \frac{\{\ln\left[(z-d+z_{oh})/z_{oh}\right]\ \ln\left[(z-d+z_{om})/z_{om}\right]\}}{u_z\, k^2} \tag{11}$$

This equation assumes neutral stability conditions, which can be common in windy areas like Prudhoe Bay during the summer (Weller and Holmgren 1974).

Penman Combination Model

This model was first developed by Penman (1948) to describe evaporation from open water or well-watered short green vegetation. It has the form of the Penman-Monteith equation with the term $r_c = 0$. Thus from Eq. (10) we can write the Penman Combination equation as

$$-Qe = \frac{\{\Delta(Q_{net}+Qg)+\left[\rho\, C_{pa}(e_{as}-e_a)\right]/r_a\}}{\Delta+\gamma} \tag{12}$$

This equation combines an energy balance component (first term in the numerator on the right-hand side) and a mass transfer (aerodynamic) component (second term in the numerator on the right-hand side). The aerodynamic resistance, r_a, is defined as in Eq. (11).

Priestley-Taylor Model

This model, developed by Priestley and Taylor (1972), is commonly used to estimate evapotranspiration where meteorological data is not abundant. It requires less input data than either of the Penman models and is given by the following equation

$$-Qe = \alpha \left(\frac{\Delta}{\Delta+\gamma} \right) (Q_{net}+Qg) \tag{13}$$

Alpha, α, is an empirical parameter that relates actual to equilibrium evaporation (α=1). Alpha is reported to be, on average, 1.26 for open water and saturated surfaces (Priestley and Taylor 1972) but it can vary considerably from site to site, depending on soil moisture, atmospheric conditions, vegetation, and other factors (Rovansek *et al.* 1996; Eichinger *et al.* 1996; Bello and Smith 1990). Thus, for best results it should be calibrated to a particular surface type.

Water Balance Method

Due to the flat terrain in the watershed, the runoff component of the water balance can be ignored after mid June when spring snowmelt runoff ends. In addition, the presence of continuous permafrost eliminates the possibility of deep groundwater percolation. Therefore, a water mass balance can be greatly simplified and *ET* can be calculated, after snowmelt runoff is completed, as follows (Rovansek *et al.* 1996)

$$ET = P + \Delta S \tag{14}$$

The change in storage, ΔS, is obtained by performing periodic readings (every three days) of staff gages in ponds and shallow wells on the tundra.

The water balance for the entire watershed is obtained by an area-weighted average of *ET* from upland (22%) and wetland (78%) areas. The evaporation of the ponds is incorporated as part of *ET* from the wetland area.

TDR-Based Water Balance

This method is similar to that described in the previous section, except that change in storage in the soil is measured by using Time Domain Reflectometry (TDR) (Stein and Kane 1983). TDR soil moisture data was only available for the wetland area. *ET* obtained from TDR data was substituted for *ET* obtained from wells in the wetland area. These values were then combined with pond *ET* and upland *ET* from the water balance as explained previously in order to obtain *ET* from the entire watershed. It should be noted, however, that the TDR probes were located in a small area at the center of the watershed (Fig. 2) and were not distributed throughout the watershed like the wells.

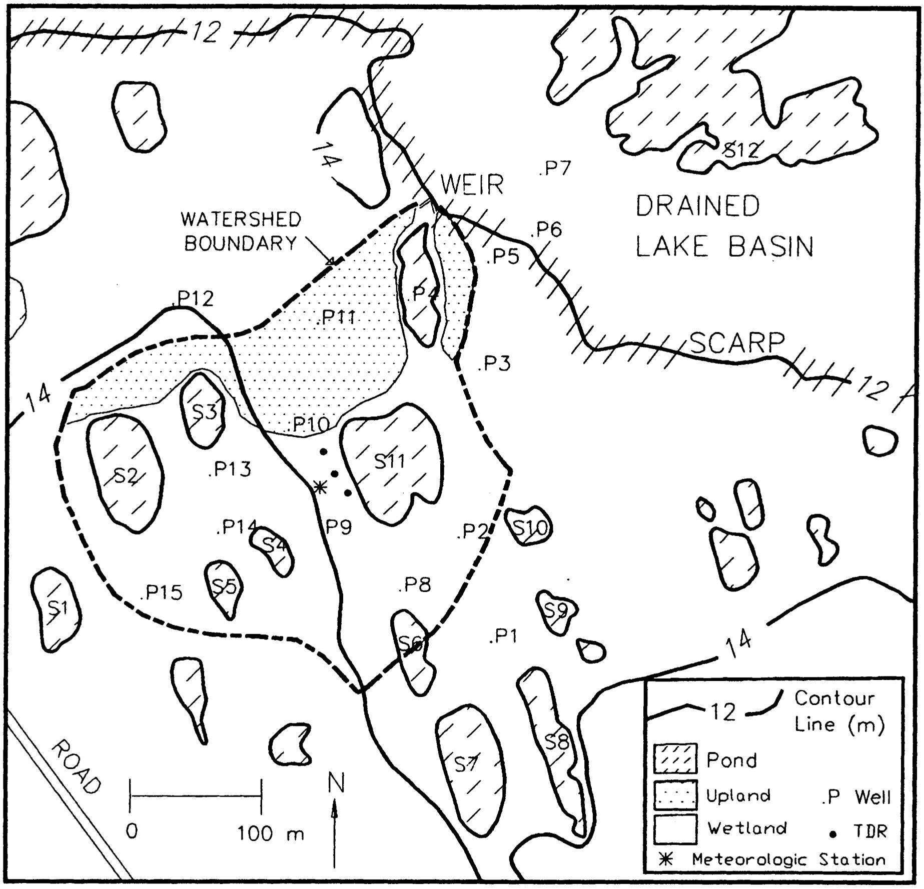

Fig. 2. Detailed map of Betty Pingo study site with 8.15-ha watershed delineated.

Results and Discussion

For the three summers of 1994, 1995, and 1996, evapotranspiration for the entire watershed obtained using the different *ET* models was compared to the BREB results. The EB model was only used during 1996 because of lack of sufficiently accurate surface temperatures in 1994 and 1995. A sensitivity analysis performed by Mendez (1997) indicated that the energy balance model was nine times more sensitive to surface temperatures than the BREB and PT models, and about eleven times more sensitive than the PC and PM models. Thus, small errors in surface temperatures can have a big impact on errors in *ET* values obtained via the energy balance. The TDR-based water balance model was also only used during 1996, when TDR data collection began.

Table 2 – Parameter values for *ET* models

Parameter	Value	Comments
Z_{om} for watershed	1×10^{-3} m	
Z_{oh} for watershed	$Z_{oh} = Z_{om}/10$	
$Z_{om}=Z_{oh}$ for pond	0.1×10^{-4}m	if 1 m wind speed < 2 m/s
	0.5×10^{-4}m	if 1 m wind speed is between 2 and 5 m/s
	1.0×10^{-4m}	if 1 m wind speed > 5 m/s
r_c wetland	25 s/m	before July 19
	30 s/m	after July 19
r_c upland	58 s/m	before July 19
	70 s/m	after July 19
α wetland	1.15	before July 19
	1.1	after July 19
α upland	0.95	before July 19
	0.91	after July 19
α pond	1.5	

For 1996, it was possible to obtain a complete energy balance for one of the ponds (pond S11) in the watershed (Fig. 2). For this pond, evaporation obtained using the Priestley-Taylor, Penman Combination, BREB, and aerodynamic methods was compared to evaporation obtained using the energy balance method.

The aerodynamic method was only used at the pond because the pond surface vapor pressure could be determined as the saturation vapor pressure at the temperature of the water surface. This assumption would not be very accurate for the land surfaces, since they are not always saturated. In addition, even when transpiring vegetation grows on saturated soil, its surface cannot be considered saturated except after rain or dew deposition (Brutsaert 1982). Therefore, the aerodynamic method was not employed for measuring land *ET*.

Two types of comparison between the models were performed. One comparison was based on the cumulative *ET* curves and values for the entire summer. The second comparison was based on a statistical analysis of daily *ET* values throughout the summer. These two types of comparisons are presented here for the entire watershed and for pond S11.

The parameters used for the models are shown in Table 2. The canopy resistance term in the PM model, as well as the alpha parameter in the PT model, were optimized against the BREB data during the summer of 1996 until the best index of agreement between the models was found (Mendez 1997). Two sets of values were obtained for r_c and the α parameter, one for early summer and one for late summer in order to account for different moisture conditions and plant growth stages. These sets were further divided into parameters for the upland and wetland areas. The models were then tested using these parameters unchanged in the 1994 and 1995 data sets.

The α parameters obtained for wetlands in this study are very similar to the 1.13 value obtained by Rovansek *et al.* (1996) for the month of August from eddy correlation measurements at this same site. It also is comparable to the 1.19 value obtained by Rouse *et al.* (1987) during onshore winds at a wetland site near Hudson Bay. Wight (1973) used an alpha of 1.0 for a sedge site near Yellowknife, Northwest Territories, Canada.

The α values for uplands found in this study are also similar to the 0.95 value used by Rovansek *et al.* (1996) for the upland areas in this same watershed. Comparable values were also reported by Kane *et al.* (1990) ($\alpha = 0.95$) and Rouse (1982) ($\alpha = 0.94$) for tundra with similar "drier" conditions.

The canopy resistance values obtained in this study (Table 2) are within the range of values obtained by Lafleur and Rouse (1988) for onshore winds at a wet marsh site and drier backshore site. They found that for the marsh site, the values ranged between 13 and 100 s/m with a median of 48 s/m. The range for the backshore site was 10 to 132 s/m with a median of 50 s/m. Rouse *et al.* (1992) obtained an r_c of 64.4 s/m for a normal wet year at a wetland tundra site, which compares favourably to the r_c for uplands in this study. The roughness length for momentum, z_{om}, presented in Table 2 was obtained from the median of a large number ($n = 3645$) of logarithmic wind profiles obtained during neutral atmospheric conditions using 1 and 10 m wind speed data from all three summers. The roughness and alpha parameters for the pond were obtained from values reported in the literature for open water surfaces.

Cumulative *ET* Comparisons

The purpose for comparing the cumulative *ET* curves is twofold. First, it gives a visual means of comparing the seasonal *ET* trends predicted by the models. Second, it provides a means for comparing models that employ different time steps such as the water balance (three-day time step) and the PT (one-day time step) models.

The cumulative *ET* comparisons begin after runoff at the weir ceases. This usually occurs between June 13 and June 17. The reason for starting at this time is that the water balance method (as is used in this study) assumes negligible runoff.

Figs. 3a, b, c, and d show the cumulative *ET* plots for 1996, 1995, and 1994. It can be observed that with the 1996 parameter calibration, the Penman-Monteith and Priestly-Taylor methods approximate best the cumulative *ET* values obtained from the BREB during 1994 and 1995. The Penman Combination model overestimated *ET* for all three summers. This is probably due to the fact that this method was developed for vegetation with unlimited water supply (*i.e.* no resistance to *ET* by plants) and open water surfaces. Unlike the PM model, the PC model does not take into account the biological control that the vegetation exerts on transpiration.

The water balance for the entire watershed slightly underestimates *ET* during all three years as compared to the BREB. Most of the under-estimation seems to occur during the later part of the summer. Under-estimation was less in 1994, which had

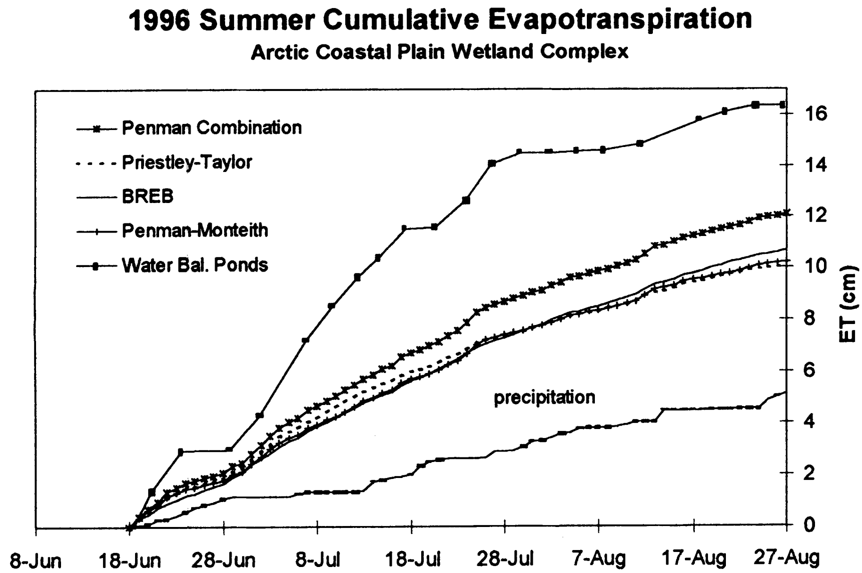

Fig. 3a. Cumulative evapotranspiration calculated by Priestley-Taylor, Penman-Monteith, Penman Combination, and BREB and precipitation; summer 1996.

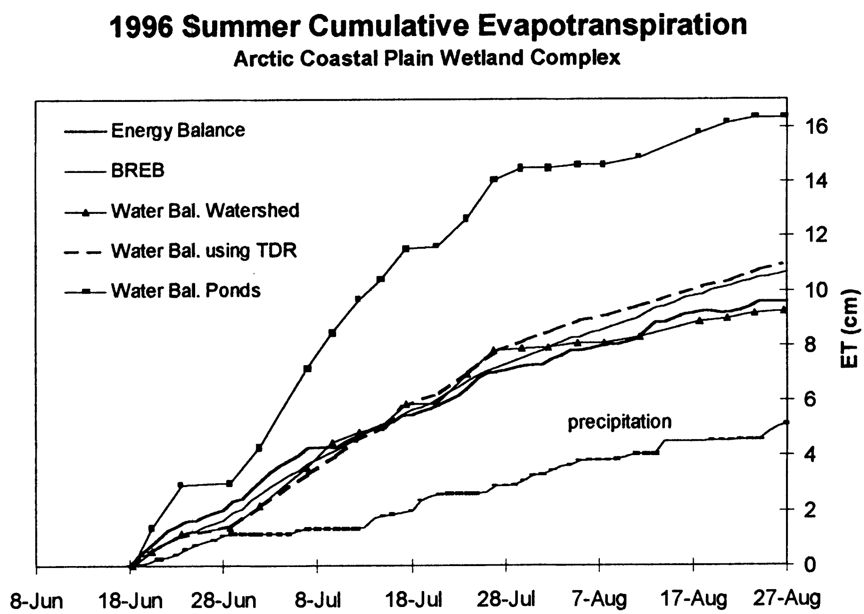

Fig. 3b. Cumulative evapotranspiration calculated by the energy balance, water balance, TDR-based water balance, and BREB and precipitation; summer 1996.

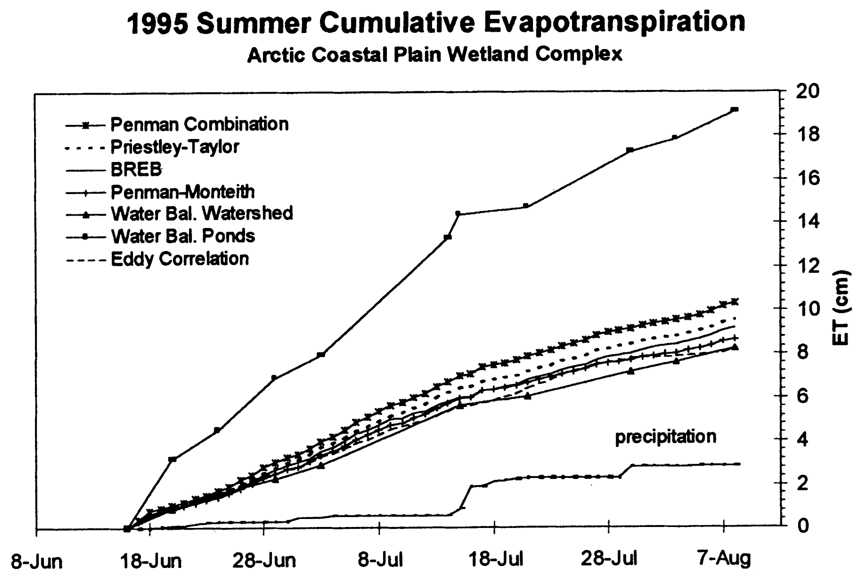

Fig. 3c. Cumulative evapotranspiration and precipitation; summer 1995.

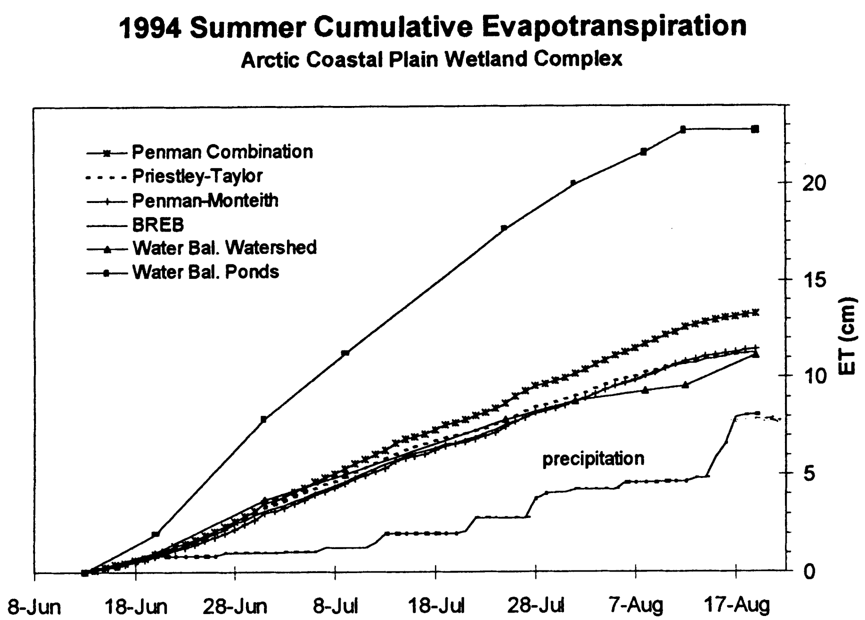

Fig. 3d. Cumulative evapotranspiration and precipitation; summer 1994.

the most August rainfall. Human error in reading the staff gages and wells could be a factor in the discrepancy between the water balance and BREB. Another factor could be precipitation undercatch, since a great portion of the precipitation in arctic coastal areas is in the form of light drizzle (Clebsch and Shanks 1968). In addition, trace events and condensation are probably not measured by the tipping-bucket rain gage used in this study. Yang *et al.* (1995) reported that on average, trace recordings constitute 45 to 50% of total annual precipitation days in Barrow, Alaska. With these values, they estimated that trace precipitation made up 12 to 13% of annual precipitation. But probably the most important factor is accuracy in determination of change in storage, ΔS, since this term is often larger than precipitation in the water balance equation (sometimes it accounts for 70 to 90% of ET).

The TDR-based water balance model employed in 1996 (Fig. 3b) gave ET results that were more comparable to the BREB than the regular water balance. This could be due to more accurate estimates of change in storage than those provided by wells. It is possible that the single specific yield value used for all the wells is not representative of the entire watershed. Also, as summer progresses and the water table drops, the specific yield of the deeper sections of soil might be different than those closer to the surface. This might also explain why the water balance cumulative curve diverges from the BREB towards the end of the summer.

Included in Figs. 3c and 4 are the results from eddy correlation measurements of ET, which were provided by Vourlitis and Oechel (1997). These measurements were taken from a tower located approximately 300 metres from our main 10 m tower. Eddy correlation provides rapid measurements of vertical wind speeds and vapor pressures, from which the evaporative flux is calculated (Kizer and Elliott 1991). This method obtains water vapor fluxes from the covariance of vertical wind and absolute humidity over a suitable averaging period (Brutsaert 1982; Kizer and Elliott 1991). As can be seen in Figs. 3c and 4, the BREB gives results that are in the range of values of eddy correlation measurements on a cumulative basis; on a daily basis there are several significant over and under-estimations of the comparisons. The fact that the BREB gives results that are comparable to completely independent methods such as the eddy correlation method as well as the water balance method serves as a verification for the model results.

The average evaporation from all ponds in the watershed obtained by the WB model was also included in Figs. 3a, b, c, and d. The evaporation of these ponds is comparable to that obtained from an evaporation pan (Fig. 5) at the site; thus ponds are evaporating at a rate near the maximum potential. Looking at the cumulative ET results for the three years, cumulative ET obtained via the BREB for the whole watershed is on average 0.54 times the cumulative evaporation from the ponds alone. This compares well with the evaporation pan coefficient of 0.49 obtained by Kane *et al.* (1990) for a watershed in the northern foothills of the Brooks Range (approximately 200 km south of our site). Figs. 3a, b, c, and d also show that cumulative ET for the watershed is higher than cumulative precipitation for each of the three sum-

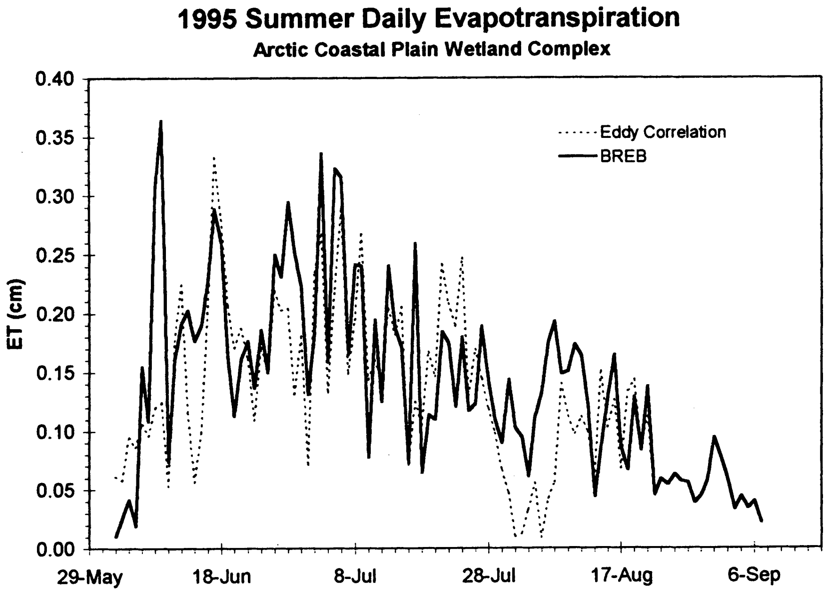

Fig. 4. Comparison of BREB and eddy correlation *ET*; summer 1995.

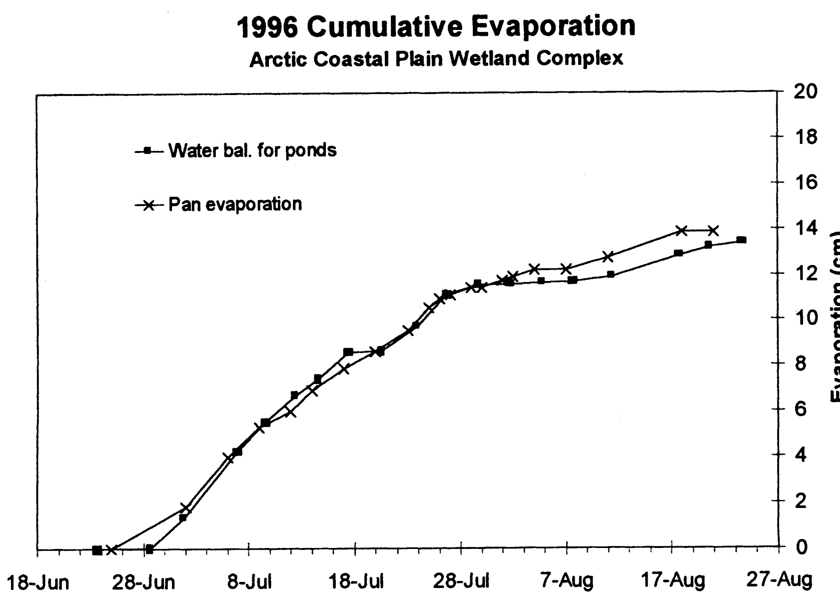

Fig. 5. Comparison of cumulative evaporation from all ponds (obtained via the water balance) to cumulative pan evaporation.

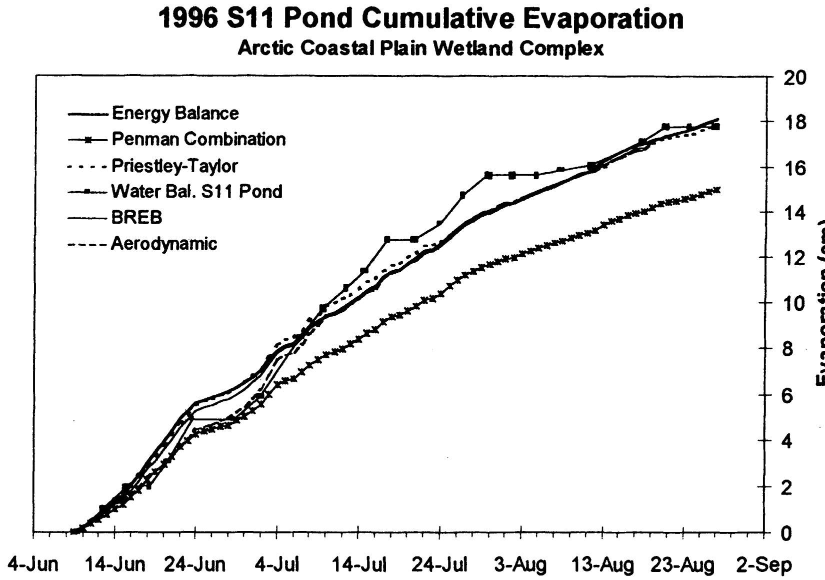

Fig. 6. Cumulative evaporation for pond S11; summer 1996.

mers. Thus, by the end of the summer, there is a water deficit in the tundra and ponds, which is replenished by the next year's snowmelt. Similar processes have been observed in the lakes of the Mackenzie Delta (Marsh and Bigras 1988).

The cumulative evaporation for pond S11 is presented in Fig. 6. The Priestley - Taylor, aerodynamic, BREB, and WB cumulative evaporation curves all match the energy balance cumulative evaporation very well. The Penman Combination method shows the least favorable results, underestimating evaporation during the entire period. A possible explanation is that the model was not originally developed to handle the very small roughness lengths that were used for the pond.

Statistical Comparison of Models

A statistical comparison provides a quantitative means of comparing the daily performance of the *ET* models. Table 3(a) shows how some of the models compared to the BREB on a daily basis for the entire period of record for the three summers. The water balance model is not included in this comparison since it had a different time step (three-day). The main statistical tool used to compare the performance of the models is a relative index of agreement (Ia) defined in Wilmott and Wicks 1980. This index has values ranging from 0 (worst performance) to 1 (best possible performance), thus providing a simple normalized way of comparing the models. The root mean square error (RMSE), which indicates non-systematic error, and the mean bias error (MBE), which gives a measure of systematic error (Halliwell and Rouse 1989), are also provided in Table 3(a).

Table 3(a) – Statistical Analysis of ET Models for Watershed

Time period	Model:	Penman Combination	Priestley-Taylor	Penman-Monteith	Energy Balance	BREB
	Ia	**0.89**	**0.95**	**0.91**	**0.83**	**1.00**
	RMSE (mm/d)	0.57	0.35	0.46	0.75	0.00
1996	MBE (mm/d)	0.19	-0.05	-0.05	-0.08	0.00
Jun 7-Sep 13	mean (mm/d)	1.57	1.34	1.32	1.31	1.38
	N = 99					
	Ia	**0.95**	**0.98**	**0.96**	**n/a**	**1.00**
	RMSE (mm/d)	0.37	0.24	0.30	n/a	0.00
1995	MBE (mm/d)	0.22	0.08	-0.04	n/a	0.00
Jun 3-Sep 8	mean (mm/d)	1.64	1.49	1.37	n/a	1.41
	N = 98					
	Ia	**0.91**	**0.96**	**0.93**	**n/a**	**1.00**
	RMSE (mm/d)	0.44	0.28	0.35	n/a	0.00
1994	MBE (mm/d)	0.24	-0.01	-0.02	n/a	0.00
Jun 1-Aug 28	mean (mm/d)	1.81	1.56	1.56	n/a	1.57
	N = 89					

The high index of agreement values in Table 3(a) indicate that the PT, PM, PC, and EB models compared well with the BREB on a daily basis. With the 1996 parameter values, the Priestley-Taylor and Penman-Monteith models compared best to the BREB during 1994 and 1995. The Penman Combination model was the third best. The same can be observed in the RMSE and MBE values. As indicated by the MBE, both the PT and PM models slightly under-predict *ET* values as compared to the BREB. The exception was during 1995 when the PT model slightly over-estimated *ET*. On the other hand, the Penman Combination model consistently over-predicts *ET* by as much as 13.5 to 15.2%.

The RMSE values in Table 3(a) indicate that there is a fair amount of non-systematic error between the models and the BREB results on a daily basis. These errors could arise from different sensitivities of the models to changes in the input variables. For instance, the EB model is about twice as sensitive to net radiation as the other radiation-based models (Mendez 1997).

During 1996 it was also possible to compare the energy balance to the BREB results. The energy balance presented the least favorable comparison to the BREB (Ia=0.83). As with the PM and PT models, it too slightly under-predicted *ET* rates (MBE of -0.08 mm/d). The energy balance method calculates *ET* as the residual term. Therefore, it is very susceptible to accumulation of errors. In addition, the energy balance requires very accurate surface temperature measurements for calculation of the convective heat flux term, Q_h, which is a major component of the energy

Table 3(b) – Statistical Analysis of ET Models for S11 Pond

Time period	Model:	Penman Combination	Priestley-Taylor	BREB	Aerodynamic	Energy Balance
	Ia	**0.94**	**0.95**	**0.97**	**0.87**	**1.00**
1996	RMSE (mm/d)	0.52	0.49	0.38	0.80	0.00
Jun 9-Aug 27	MBE (mm/d)	-0.40	-0.04	-0.01	-0.01	0.00
	mean (mm/d)	1.89	2.25	2.28	2.28	2.29
	N= 80					

budget in the tundra. Small errors in temperature measurements could have big impacts on the determination of Qh and thus Qe. The Qh term is not used in the other models, thus they have one less source of error. The other models, as well as the energy balance, do require surface temperatures for the calculation of the conductive heat flux, Qg. However, Qg is the smallest term in the energy budget, thus its impact on ET is not as significant.

Table 3(b) shows the statistical comparisons of the evaporation models for pond S11. The standard of comparison for the pond was the EB model. There was more confidence in using the more rigorous EB model as the standard of comparison over the pond than over land. This is because the thorough mixing of the shallow pond's water caused by steady winds makes the water temperature isothermal throughout its depth. This, combined with the availability of thermistors at different depths in the pond, increased our confidence in the accuracy of the surface temperature measurements. In addition, over water surfaces, there is not the complication found over land where it is necessary to find a representative temperature of the vegetation canopy. Consequently, we had more confidence in the evaporation rates obtained from the energy balance over the pond than we did over land surfaces. The index of agreement values in Table 3(b) indicate that the best performance was that of the BREB, followed by the Priestley-Taylor, Penman Combination, and lastly the aerodynamic method. A similar outcome is observed when looking at the RMSE values.

The MBE results for the pond indicate that the Penman Combination model under-predicts evaporation by the greatest amount (MBE of -0.4 mm/d or -17.5%) as compared to the other models. This result was also observed in the cumulative plots in Fig. 6.

Additional Findings

From the energy balance results presented in Tables 4(a) and 4(b), several observations can be made. First, latent heat flux, Qe, over wetlands is the largest energy sink, dissipating on average 45.8% of net radiation during the summer. Similar results were found by other studies in the Arctic (*e.g.* Addison 1972; Moore *et al.* 1994) as well as in temperate regions (*e.g.* Souch *et al.* 1996). Over the upland areas, sensible heat flux, Qh, is a much greater energy sink than Qe (51.7 *versus* 23% of

Table 4(a) – Energy Partitioning as Percentage of Qnet for Watershed; summer 1996

	wetland			upland		
Month	*Qh*	*Qg*	*Qe*	*Qh*	*Qg*	*Qe*
June	27.9	20.5	51.6	50.1	34.2	15.7
July	41.0	18.4	40.6	43.5	24.0	32.4
August	44.6	10.3	45.1	61.4	17.6	21.0
Average	**37.8**	**16.4**	**45.8**	**51.7**	**25.3**	**23.0**

Table 4(b) – Energy Partitioning as Percentage of *Q*net for S11 pond; summer 1996

Month	Qh	Qg	Qw	Qe
June	31.1	9.2	2.0	57.7
July	29.1	16.4	-1.1	55.5
August	37.8	9.4	-3.7	56.4
Average	**32.7**	**11.7**	**-0.9**	**56.5**

*Q*net). Rott and Obleitner (1992) and Lewis and Callaghan (1976) found similar results for dry tundra sites. This is possibly related to the drier and warmer surface conditions found in the upland area. Such energy partitioning explains why *ET* rates are higher over wetland areas than upland areas (Fig. 7). Pond S11 displays an energy partitioning similar to that of the tundra wetland area. Latent heat flux dissipates

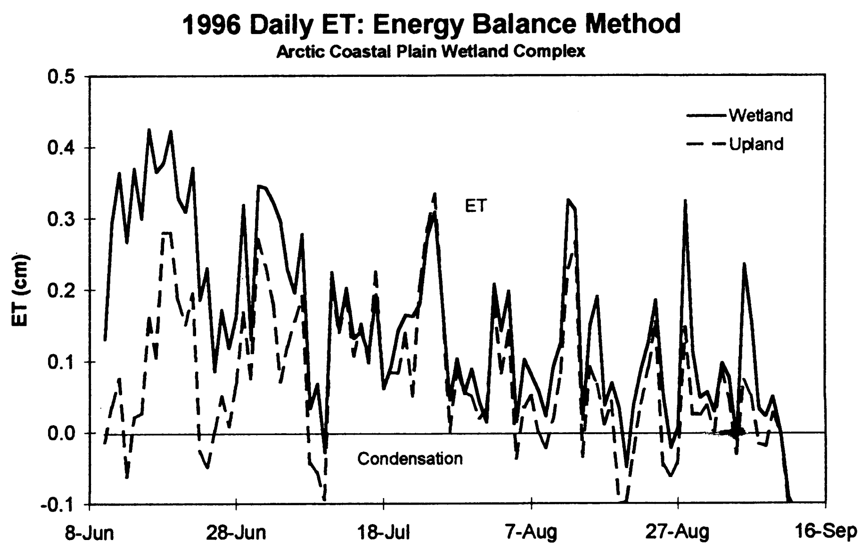

Fig. 7. Comparison of upland and wetland daily *ET* obtained from the energy balance model; summer 1996.

Table 5 – Monthly Average ET Rates (mm/day)

Month	1996 (S11 pond)	1996	1995 (watershed)	1994	Average (watershed)
June	3.06	1.81	1.76	1.63	1.73
July	2.48	1.81	1.73	1.85	1.80
August	1.40	1.06	1.02	1.20	1.09

the largest percentage of the incoming net radiation (56.5%), followed by *Qh* (32.7%). Conductive heat flux at the bottom of the pond is mainly negative during the summer and constitutes a small percentage of *Q*net (11.7%). The change in heat storage term, *Qw*, can be a source or sink of energy.

From Table 3(a), the mean watershed *ET* rate (using the BREB) for the entire length of the data record was 1.57, 1.41, and 1.38 mm/d for 1994, 1995, and 1996, respectively. The average for the three years was 1.45 mm/d. The average post-runoff (*i.e.* after mid June) summer evaporation rate (obtained via the WB) from all the ponds in the watershed averaged 3.40, 3.60, and 2.33 mm/d for 1994, 1995, and 1996 respectively. The average for the three years is 3.11 mm/d. These values were obtained from the water balance approach. These *ET* and evaporation rates are comparable to those obtained by other studies in the Arctic for similar surfaces (*e.g.* Mather and Thornthwaite 1958; Clebsch and Shanks 1968; Kane and Carlson 1973; Rouse 1982; Ohmura 1982; Marsh and Bigras 1988; Rovansek *et al.* 1996).

ET and evaporation rates in general were higher during the month of June and July and decreased in August (Table 5). Over the three-year period from 1994 to 1996, the average *ET* rate for June and July was 1.73 and 1.80 mm/d respectively, and for August it was 1.09 mm/d. For pond S11, the evaporation rate for June and July was 3.06 and 2.48 mm/d respectively, and for August it was 1.40 mm/d. These results follow the pattern of net radiation, which is larger in June and July than in August (Fig. 8). Similar findings were observed by Hinzman *et al.* (1996) and Rouse (1982).

Table 5 also shows that during August of 1994 the *ET* rate was higher than during August of 1995 and 1996. Precipitation data indicate that August of 1994 was about three times wetter than August of 1995 and 1996. Temperature data also indicates that August of 1994 was 4.5 and 6.4°C warmer than August of 1995 and 1996, respectively. This indicates that the combination of wetter and warmer conditions can increase the rate of *ET* from this tundra wetland. Kane (1997) also observed a similar behavior in a watershed at the northern foothills of the Brooks Range, where wetter years displayed higher *ET* rates. Such findings could have an important implication for studies of global climate change. In a global warming scenario, the ice-free period in the Beaufort Sea would be prolonged (Kane 1997), leading to a possible increase in precipitation. This combined with higher temperatures and possibly greater vapor pressure deficits could increase the rate of evapotranspiration from the

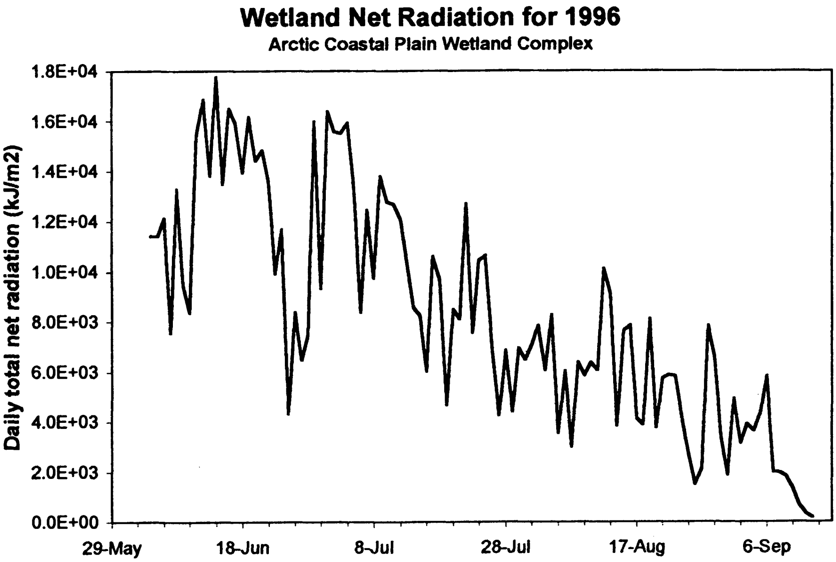

Fig. 8. Typical pattern of net radiation during the summer.

tundra. It should be kept in mind, however, that there are other feedback mechanisms, such as the response of vegetation to warmer climates and precipitation distribution, that should also be taken into account if such predictions are undertaken. Such considerations are beyond the scope of this study and the reader is referred to other works (*e.g.* Kane 1997; Lafleur and Rouse 1988; Rouse *et al.* 1992) for further information on this topic.

Conclusions

Evapotranspiration is a very important component of the water balance in arctic coastal wetlands, with a magnitude usually exceeding that of precipitation during the summer. As determined from the BREB model, evapotranspiration rates for the summer snow-free period in this arctic wetland complex were 1.38, 1.41, and 1.57 mm/day for 1996, 1995, and 1994, respectively. The average for the three years was 1.45 mm/d. *ET* rates were higher during wetter, warmer conditions (August 1994) than during drier, cooler ones (August 1996 and August 1995). In addition, *ET* was higher in June and July than in August.

Average evaporation rates from all the ponds in the watershed obtained via the water balance method for the post-runoff summer period were 2.33, 3.60, and 3.40 mm/d for 1996, 1995, and 1994, respectively. Comparison of these evaporation results to *ET* values obtained for the entire watershed by the BREB for the same time periods indicate that the *ET* rate from the tundra as a whole is approximately one-half the evaporation rate from the ponds (per unit area). If pond evaporation is taken as a measure of the potential for *ET* in this wetland complex, then it can be concluded that actual *ET* rates from the tundra are approximately half of the maximum potential rate that can be sustained by current atmospheric and energy conditions.

Energy balance results showed that over wetland and ponds, latent heat flux, *Qe*, is the most important energy sink, but over the drier and warmer upland areas *Qe* plays a secondary role to sensible heat flux, *Qh*, as the most important energy sink.

Cumulative and daily model comparisons indicated that, in general, all the models gave fairly similar results. The energy-based models such as the BREB, PT, PM, and EB compared well with the water balance, eddy-correlation, and aerodynamic methods, which are independent methods of obtaining *ET* and evaporation. The PC model seemed to over-estimate *ET* for all three years as compared to the BREB. The Priestly-Taylor and Penman-Monteith models showed consistently good comparisons to the BREB results in 1994 and 1995, with the α and r_c parameters obtained through calibration against the BREB in 1996. The simplicity of the PT model, which requires fewer input variables than most of the other models, together with its performance being similar to the more complicated models, makes it a very attractive method to use to estimate *ET* from wetland tundra.

One disadvantage of the PT model is that it lumps several physical processes into the α parameter, thus making it less desirable for investigations on how different variables (*i.e.* *Q*net, wind speed, and vapor pressures) affect *ET* and what their feedback relationships are. In this respect, the PM model is a better choice for modeling *ET*.

Acknowledgments

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