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Evapotranspiration from a Small Alaskan Arctic Watershed

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D. L. Kane, R. E. Gieck and L. D. Hinzman

University of Alaska, Fairbanks, AK-99775, U.S.A.

Evapotranspiration (*ET*) vies with runoff as the primary mechanism for water loss from a watershed underlain by permafrost, yet past attempts to predict *ET* have proven to be less than completely successful in the Arctic. Imnavait Creek, a small 2.2 km^2 watershed underlain by continuous permafrost has been studied for 4 years. Evapotranspiration on a watershed scale has been calculated from water balance studies. These results are compared with point measurements of pan evaporation and daily estimates of *ET* by the energy balance and Priestley-Taylor methods. Since it is difficult to determine the daily change in soil moisture, the energy balance approach appears to be the best method to determine daily *ET*. The water balance approach is the best method to determine total *ET* over the course of the summer because it is possible to delete the soil moisture term due to an insignificant change annually in this watershed. Priestley-Taylor gave adequate estimates of *ET* with only limited data. After a pan coefficient is determined, the evaporation pan functions well over extended time periods but is less accurate for shorter periods. Evapotranspiration is greatest in early summer, immediately following the spring snowmelt, during the period of maximum incoming radiation but not necessarily maximum air or soil temperatures. The cumulative potential evaporation is greater than the cumulative summer precipitation. The source of moisture for *ET* in early summer is from snowmelt or moisture stored in the active layer.

Introduction

The Arctic functions as a sink to the excess energy radiated on the more temperate regions of the Earth. It is important to understand the energy transformations taking place in the Arctic in order to determine the role of the Arctic on atmospheric circulation and in turn, global weather processes. Evaporation (E) and evapotranspiration (ET) are among the primary mechanisms of summer energy dissipation in the Arctic.

This paper presents the results of an investigation into the evaporative process using several techniques to determine the daily ET. Although this analysis is based upon a comprehensive data collection effort, emphasis was placed on evaluating methods of accurately determining ET based upon a minimum amount of data because extensive meteorologic data are normally not available for most areas of the Arctic. The accuracy of the simplified approaches are checked by comparison with the results of the energy and water balance measurements.

The purpose of this investigation was to develop a better understanding of the hydrologic, meteorologic, and thermal processes occurring in the arctic environment. Imnavait Watershed is a small headwater watershed located in the foothills of the Philip Smith Mountains (latitude 68°37′ N, longitude 149°17′ W, elevation ≈900 m) on the Arctic North Slope of Alaska, U.S.A. The basin is located between the headwaters of the Toolik and Kuparuk Rivers, approximately 150 km south of the Arctic Ocean. Imnavait Watershed is 2.2 km^2 in area and is completely underlain by permafrost to a depth of 250 to 300 m (Osterkamp *et al.* 1985). The active layer, or surficial soil which progresses through freeze and thaw each year, is between 25 and 60 cm thick. Imnavait Creek is a north flowing stream in which discharge completely ceases during the winter months and often approaches zero flow in summer after extended dry periods. The soils are *Histic Pergelic Cryaquepts* and consist of 10 to 15 cm of highly organic soil overlying a fine-grained till. The vegetation is primarily water tolerant plants such as sedge tussocks and mosses, but they are accompanied by lichens and shrubs such as willow, alder and dwarf birch. A few small ponds exist in the valley bottom, but these occupy less than 0.5 % of the basin area.

Relevant Research

Research related to evaporation in the Arctic is very limited for several reasons. Problems common to all research in the Arctic are difficulties associated with the stark remoteness of the research site; *i.e.* sites are frequently inaccessible and expensive to reach, continuous measurements throughout the year are formidable and even throughout the summer are quite rare, and instrumentation must operate from batteries or small generators. Measurement of precipitation is more difficult than in more temperate regions due to high winds and a higher percentage of precipitation falling as snow. The challenges of measuring runoff are greater due to the presence of ice and snow and the changing configuration of channels due to thawing. In addition to the above adversities, evaporation studies are normally quite labor intensive requiring frequent personal observation and hands-on data collection by the researcher. By nature, evapotranspiration is quite variable depending upon meteorologic and soil moisture conditions as well as plant status, therefore studies usually must be of relatively long duration to be meaningful.

Evaporation represents a coupling of the meteorologic and hydrologic regimes. Although eddy correlation and aerodynamic methods representing mass transfer have been developed (Brutsaert 1982), many experiments designed to determine evaporation measure or calculate all the other components of either the water or energy balance and determine the evaporative vapor flux as the residual term. In the more temperate environments, calculations of evaporation are usually based upon energy balance or vapor transfer considerations rather than a water balance approach. This is due to the large uncertainty in determining the amount of moisture infiltrating to groundwater. In arctic regions underlain by permafrost, percolation into groundwater and soil water storage is severely limited and can be eliminated from the water balance for seasonal estimates (Kane *et al.* 1989), thus making this approach more viable. Evapotranspiration from a watershed isolated from groundwater flow is easily calculated from a water balance as

$$ET = P - Q - dS \tag{1}$$

where

ET – evapotranspiration, mm
P – precipitation, mm
Q – surface runoff, mm
dS – change in surface and/or soil water storage, mm

By selecting a time frame where the change in soil storage dS is negligible, it is possible to neglect this term.

The energy balance approach is a commonly applied method in determining evaporation. This concept is easily understood; however it is not always easy to accurately determine the magnitude of each of the components. The data requirements for a complete energy budget are extensive and the instrumentation is expensive but the primary problem with this method is the same as the water balance; the error in measurement of each component is collected in the residual term, in this case the evaporation and/or transpiration. Each of the components of the energy balance involves individual but related processes. Due to the ease of measuring the radiative components with accurate, well-calibrated instruments, net radiation, Q_{net}, is perhaps the easiest component to quantify.

The energy advected in water can often be neglected under most conditions if there is not significant movement of soil water. Precipitation in the Arctic often occurs as a slow drizzle and is usually within a few degrees of the soil temperature. This addition of energy may be significant over a summer, but can be neglected for most days. So the energy balance can be expressed as

$$Q_e = Q_{net} + Q_h + Q_c \tag{2}$$

where

Q_e – the energy utilized for evaporation and/or transpiration of water from the surface, W/m^2
Q_{net} – energy transferred at the surface through net radiation, W/m^2
Q_h – sensible heat flux between surface and air, W/m^2
Q_c – energy flux via conduction between surface and subsurface, W/m^2

As mentioned previously, some components of the energy balance are difficult to accurately quantify; consequently, several alternate techniques have been derived from the energy balance. Use of the Bowen ratio (Bowen 1926) is one such technique. This approach considers the ratio of sensible heat flux, Q_h, to the latent heat flux, Q_e to avoid the requirement of obtaining all the data necessary for a complete energy balance. The Priestley-Taylor method (Priestley and Taylor 1972) to estimate evapotranspiration is a technique similar to the Bowen Ratio method but requires less data. In a similar derivation as the Bowen ratio, the following formula is achieved

$$Q_e = \alpha\left(\frac{s}{s+\gamma}\right)(Q_{net} - Q_c) \tag{3}$$

where

α – an evaporability parameter relating actual to equilibrium evaporation
s – slope of the specific humidity and temperature curve, °C^{-1}
γ – psychrometric constant in terms of specific humidity, °C^{-1}

In subarctic studies, Rouse and Stewart (1972) and Stewart and Rouse (1976) found in a well drained, upland lichen heath an average $\alpha = 0.95$ was applicable. They also found they could relate $s/(s + \gamma)$ to a linear function of the screen air temperature (T_a) as

$$\frac{s}{s+\gamma} = 0.406 + 0.011\,(T_a) \tag{4}$$

Rouse *et al.* (1977) later tested the above relationships in tundra and forested sites in northern Canada using the linearization of screen air temperature and omitting the soil conduction term. They found the evaporability parameter to vary depending upon vegetation type and soil moisture regime, but the linearization of the $s/(s + \gamma)$ equation appeared to function well. In a study in the High Arctic, Marsh *et al.* (1981) tested the reliance of the evaporability parameter, α, on soil moisture and soil type and found that there is no unique relationship between surface soil moisture and evaporation rates.

Direct measurement of evaporation or evapotranspiration usually involves placing a lysimeter within the snow or soil matrix and periodically determining the change in weight. Although actual measurements using lysimeters are almost always preferred for comparison with calculated values, this technique presents many problems and errors associated with the thin active layer, saturated soils and snow bridging, all of which occur commonly in the Arctic. When used correctly, lysimeters are capable of measuring true evapotranspiration as compared to evaporation pans which give an estimate of potential evaporation given an unlimited water source. Evaporation pans do offer a relatively inexpensive and easy way to measure water loss due to evaporative processes.

Most research into evaporation in the Arctic concerns determining the amount of loss from lakes (Kane and Carlson 1973; Marsh 1986; Marsh and Bigras 1988). Although a high percentage of the Arctic is covered by open water, estimations of evaporation and/or transpiration from tundra surfaces are also needed for water balance studies and hydrologic and ecologic modeling. One study dedicated primarily to the determination of the amount of evaporation in the High Arctic was conducted by Ohmura (1982). Evaporation was calculated using the Bowen ratio method and measured using a weighing lysimeter. Although he found total evaporation to be small compared to more temperate environments, it was the primary mechanism of water loss on Axel Heiberg Island in Northern Canada. Evaporation increased abruptly as snow free patches appeared during the snowmelt period, with the year's highest evaporation rates occurring soon after the completion of the melt (Table 1).

Concurrent measurements of all components of the energy balance have been collected only a few times in the Arctic. Weller and Holmgren (1974) used this technique near Barrow to elucidate the magnitude of the components of the energy balance throughout the year. They also found the highest rates of evapotranspiration immediately following the snowmelt period corresponding to the period of maximum radiation but not necessarily maximum air temperatures. These results are summarized and presented with other relevant research in Table 1.

Table 1 – Evaporation data calculated and/or collected from arctic watersheds

Location	Time period	Evaporation (mm/day)	(mm)	Method		Reference
McMaster	Oct 1-Sept 30, 1976	-	31	PT	*	Woo, 1983
River Basin,	Oct 1-Sept 30, 1977	-	31	PT	*	
Cornwallis	Oct 1-Sept 30, 1978	-	38	PT	*	
Island,	Oct 1-Sept 30, 1979	-	30	PT	*	
N.W.T.	Oct 1-Sept 30, 1980	-	51	PT	*	
	Oct 1-Sept 30, 1981	-	47	PT	*	
Baker Lake,	May 29-Jul 31, 1982	2.9	183	PT,BR	*	Roulet and
N.W.T.	Jun 9-Aug 5, 1983	3.7	215	PT,BR	*	Woo, 1986
	Jun 4-Aug 5, 1984	3.5	223	WB	***	

cont.

Table 1 – cont.

Location	Time period	Evaporation (mm/day)	Evaporation (mm)	Method		Reference
Barrow, Alaska	Jul 11-31, 1965	2.2	47	PAN	*	Brown, et al., 1968
	Aug 1-27, 1965	1.6	44	PAN	*	
	Jun 22-30, 1966	2.7	24	PAN	*	
	Jul 1-30, 1966	2.1	63	PAN	*	
	Aug 1-28, 1966	1.8	49	PAN	*	
Barrow, Alaska	Jul & Aug, 1957	1.6	102	EB	*	Mather and Thornthwaite, 1958
	Jul & Aug, 1957	1.3	81	LY	*	
	Jul & Aug, 1958	0.8	52	EB	*	
	Jul & Aug, 1958	1.3	81	LY	*	
Putuligayuk Basin, Alaska	Jun 9-Aug 27, 1971	2.4	193	PAN	*	Kane and Carlson, 1973
	Jun 24-Aug 27, 1971	2.0	127	KE	**	
Resolute, Cornwallis Is., N.W.T.	Jul 1-Aug 31, 1978	1.0	61	PT	*	Woo, Heron and Steer, 1981
	Jul 1-Aug 26, 1979	0.9	52	PT	*	
Axel Heiberg Island, N.W.T.	Jun 20-Aug 31, 1969	1.3	84	BR,EB	*	Ohmura, 1982
	Jun 1-Aug 24, 1970	1.6	86	BR,EB	*	
	Jun 28-Aug 22, 1972	1.5	82	BR,EB	*	
Barrow, Alaska	Jun 14-17, 1971	4.5	18	EB	*	Weller and Holmgren, 1974
	Jul 15-28, 1971	2.6	36	EB	*	
	Aug 25-Sept 1, 1971	1.1	9	EB	*	
	Jun 14-17, 1971	4.2	17	PAN	*	
	Jul 15-28, 1971	3.0	42	PAN	*	
	Aug 25-Sept 1, 1971	0.7	6	PAN	*	
Imnavait Watershed, Toolik Lake, Alaska	Jun 1-Aug 31, 1986	3.4	310	PAN	*	Hinzman, 1990
	Jun 1-Aug 31, 1986	1.7	153	WB	***	
	Jun 1-Aug 31, 1987	3.5	320	PAN	*	
	Jun 1-Aug 31, 1987	1.4	130	WB	***	
	May 15-Aug 31, 1988	3.0	330	PAN	*	
	May 15-Aug 31, 1988	1.6	180	WB	***	
	May 25-Aug 31, 1989	4.2	420	PAN	*	
	May 25-Aug 31, 1989	2.4	240	WB	***	
Resolute, Cornwallis Is., N.W.T.	May-Sept, 1976	-	46	PT	*	Marsh and Woo, 1979
	May-Sept, 1976	-	40	PT	*	
	May-Sept, 1976	-	39	PT	*	
	May-Sept, 1976	-	41	PT	*	
	May-Sept, 1978	-	40	PT	*	

cont.

Table 1 – cont.

Location	Time period	Evaporation (mm/day)	(mm)	Method		Reference
NRC Lake, MacKenzie River Delta, N.W.T.	Jun 7-Aug 31, 1984	2.9	247	PT	*	Marsh and Bigras, 1988
	Jun 9-Sept 1, 1985	2.9	243	PT	*	
	Jun 16-Sept 2, 1986	2.5	200	PT	*	
	Jul 13-Sept 2, 1984	2.2	114	PT	*	
	Jul 13-Sept 2, 1984	2.0	103	WB	**	
	Aug 5-Sept 1, 1985	2.1	56	PT	*	
	Aug 5-Sept 1, 1985	2.3	63	WB	**	
	Aug 6-Sept 2, 1986	1.8	50	PT	*	
	Aug 6-Sept 2, 1986	1.5	40	WB	**	
Dishwater Lake, MacKenzie Delta River, N.W.T.	Jun 15-Sept 11, 1982	3.9	349	WB	**	Marsh and Bigras, 1988
	Jun 15-Sept 1, 1983	4.1	322	WB	**	
	Jun 7-Aug 29, 1984	4.6	387	WB	**	
	Jun 9-Aug 27, 1985	3.9	310	WB	**	
Meadow Watershed, N.W.T.	Open water period 1978	-	107	MT	**	Newbury, et al., 1979
Barrow, Alaska	snow free period, long term average data	-	140	EB	*	Dingman, et al., 1980
			60	WB	***	
Ellesmere Is., N.W.T.	Jul 6-Aug 17, 1975	-	27	CM	*	Marsh and Woo, 1977
Devon Island N.W.T.	Summer 1972	-	69	WB	***	Rydén, 1977
	Summer 1972	-	70	EB	*	
	Summer 1973	-	110	WB	***	
	Summer 1973	-	120	EB	*	
	Summer 1974	-	65	WB	***	

EB – Energy balance
PT – Priestley-Taylor
WB – Water balance
PAN – Potential pan evaporation
LY – Lysimeter
MT – Mass transfer for lake evaporation
CM – Combination model – Priestley-Taylor and mass transfer
KE – Kohler's equation for lake evaporation
BR – Bowen ratio

* – Point
** – Lake
*** – Watershed

Measurements at Imnavait Watershed

Two comprehensive meteorological sites were established within the basin. These sites were instrumented to record data from all radiation components including incident, reflected and emitted radiation using annually calibrated longwave and shortwave radiometers. Air temperature, relative humidity, and wind speed were measured at several heights, as well as snow and soil temperature at various depths. Temperatures were measured using calibrated thermistors and thermocouples. Relative humidity was measured with a chemical film resistor chip. Wind speeds were measured by cup type anemometers. The preceding data were measured every 60 seconds, averaged hourly and recorded on dataloggers. Precipitation was measured using shielded tipping bucket raingages (rainfall) and Wyoming (snow and rainfall) gages mounted with a water level recorder. The Wyoming snowgage data were recorded on a punch tape and reduced as daily total precipitation. The tipping bucket raingages were also recorded on dataloggers and totaled hourly.

Measurements of the snow on the ground were made using an Adirondack snow sampler. Snow surveys, comprised of at least 40 data points, were made periodically prior to the initiation of melt and each morning during melt. The liquid water content of the soil at various depths was measured by time domain reflectometry (Stein and Kane 1983). Soil moisture was measured several times each week during snowmelt and at least weekly throughout the summer. Soil thermal conductivities were determined in the laboratory for various soil moisture and temperature conditions for both organic and mineral soils using a guarded hot plate (Hinzman *et al.* 1991). Stream stage was recorded continuously using a water level recorder attached to a stilling well in a 1 meter H-type flume. Streamflow measurements were periodically made using cup type and electromagnetic current meters. Pan evaporation was measured using a standard Class A four foot (1.22 m) diameter by ten inch (25 cm) deep evaporation pan.

Evapotranspiration Estimation

Mass Balance

Evapotranspiration on a watershed scale was estimated using Eq. (1). Runoff, precipitation and change in soil moisture were measured and ET was determined as the residual term. Problems with the mass balance estimates of ET arise mainly from measurement error. The estimation of ET using water balance calculations again collects the error associated with each measurement in the residual term. Inaccuracies in streamflow and precipitation measurements and difficulties in measuring groundwater and soil moisture storage changes, especially during short time periods, can compound to result in errors greater than the term being estimated (Woo 1983).

Clagett (1988) compared precipitation (P) measurements taken with unshielded

Table 2 – Water balance over the spring and water year

Year	Snowpack Water (mm)	Summer Precip (mm)	Total Precip (mm)	Snowmelt Runoff (mm)	Summer Runoff (mm)	Total Runoff (mm)	Evapo-trans (mm)
1986	109	163	272	57	62	119	153
1987	108	272	380	71	179	250	130
1988	78	252	330	39	72	111	219
1989	155	257	412	94	78	172	240

weather service precipitation gages and the Wyoming snowgage at several sites in Arctic and Subarctic Alaska. He found that the Wyoming gage is considerably more accurate, although it still under-reports precipitation amounts especially for snowfall during high wind conditions. Point measurements of stream discharge (Q) are generally assumed to be accurate to within 10 to 15%. However, since our measurements utilized a calibrated flume with stage-discharge relationship that was checked by many independent observations, our accuracy should be considerably better. It was possible to calculate the annual water balance by assuming no change in soil moisture storage (Kane and Hinzman 1988). This is due to near saturation conditions each fall caused by limited soil moisture storage capacity in the active layer and relatively high precipitation in August and September. In this basin, there is no loss of water to subpermafrost groundwater due to the ice-rich conditions of the permafrost and no source of water from subpermafrost springs. Therefore an accurate determination of ET can be obtained through accurate measurements of the snowpack, summer precipitation, and runoff (Table 2). The total length of time between ablation in the spring and freezing in the fall varies by several weeks from year to year; consequently, estimates of total ET from the water balance may not directly compare with those of other years. In order to facilitate comparisons between years, the cumulative summer ET was normalized by the length of time included in the calculation and reported as the amount of evaporation and/or transpiration per day (Table 3).

Pan Evaporation

Pan measurements were taken manually as often as possible by our field technicians, usually at least once each day and often several times daily. Pan evaporation alone is not necessarily a good estimate of evapotranspiration for several reasons. The evaporation pan is slightly elevated and has a reflectance that differs from the soil or leaf surfaces; consequently, it is representative of a different energy environment than the soil or plants. Much of the evapotranspiration occurs on the leaf surface where conditions can be radically different from those occurring at the pan water surface. Therefore, determining the relationship between the pan evaporation and potential evaporation or actual evapotranspiration is difficult.

Table 3 – Comparison of methods used to predict evapotranspiration

Method			Calculated Evapotranspiration (1)	(2)	(3)	(4)	Average ET	Period Begin	End
Year	Pan Evap (mm/day)	Pan Coeff.	Pan ET (mm/day)	Energy Balance (mm/day)	Priestley Taylor (mm/day)	Water Balance (mm/day)	Method (2,3,4) (mm/day)		
1986	4.1	0.39	2.0	*	1.9	1.3	1.6	6 Jun	20 Aug
1987	3.1	0.45	1.5	1.4	1.7	1.1	1.4	23 May	26 Aug
1988	3.2	0.59	1.6	1.7	1.9	2.0	1.9	15 May	26 Aug
1989	4.1	0.51	2.0	2.3	1.7	2.2	2.1	29 May	4 Sep
Average	3.6	0.49	1.8	1.8	1.8	1.7	1.8		

* – data unavailable

Pan coefficients may be used to compensate for differences such as the water availability in the soil, soil temperature characteristics of the sites and different climatic characteristics between sites. Patric and Black (1968) compared pan evaporation to Penman and Thornthwaite estimates of actual and potential evapotranspiration at Fairbanks, Alaska. Even though they demonstrate that pan coefficients do vary seasonally, these results from Subarctic Alaska suggest that pan evaporation may be used as a rough estimate of potential evapotranspiration using a pan coefficient of 0.98 for interior Alaskan sites. A pan coefficient of 0.59 for actual evapotranspiration was derived from their data for Fairbanks, Alaska. There are many factors which affect the accuracy and comparability of pan evaporation data. Hickox (1944) compared pans of various sizes, albedo or color, rim heights and water depths under controlled laboratory conditions. He found that pans should be well maintained to insure consistent albedo and a consistent rim height above the water surface. Evaporation can increase by as much as 10% in dirty pans with lower albedos and an increase in rim height of 2.5 cm also increased evaporation by as much as 10%.

A comparison of four years of measured pan evaporation to the estimates of actual evapotranspiration yield a pan coefficient ranging between 0.39 and 0.59, with an average of 0.49 (Table 3). The pan coefficient was determined by dividing the total pan evaporation by the average *ET* determined from the energy balance, water balance and Priestley-Taylor methods. The pan *ET* in Table 3 was calculated by multiplying the pan evaporation by 0.49.

Energy Balance

An energy balance was based on radiation, air temperature, soil temperature and wind speed data collected within the basin. The evaporative energy flux was then used to determine the mass of water evaporated and/or transpired. The individual processes are described below through their theoretical development and via equations which easily lend themselves to computer analysis.

The energy used in conduction warms the soil and thaws the ice in the active layer. This is a significant percentage of the energy balance in the summer. It is not necessary to know how much ice is melted if one considers only the amount of energy which is conducted across the surface

$$Q_c \equiv K_s \frac{T_d - T_{ss}}{L} \tag{5}$$

where

K_s – thermal conductivity of the soil, W/m° C
T_d – soil temperature at depth d, °C
T_{ss} – soil surface temperature, °C
L – thickness of soil zone between surface and depth d, m

Calculation of the convective heat transfer term (Q_h) is more complex. Following the formulation of Moore (1983), the integrated form is developed using the bulk heat exchange coefficient D_h

$$Q_h = \rho_a C_{pa} D_{h[n,u \text{ or } s]} (T_a - T_s) \tag{6}$$

The stability of the air just above the ground surface must be considered and this relationship must be corrected for non-neutral (isothermal) conditions. This is possible using either the Monin-Obuchov factor or, as in this paper, the Richardson number (Price and Dunne 1976).

$$D_{hn} \equiv \text{neutral heat exchange coefficient} = k^2 (u_z) / (\ln(\frac{z}{z_o}))^2 \tag{7}$$

$$D_{hs} = \text{stable heat exchange coefficient} \equiv D_{hn} / (1 + (\sigma R_i)) \tag{8}$$

$$D_{hu} = \text{unstable heat exchange coefficient} = D_{hn} (1 - (\sigma R_i)) \tag{9}$$

$$R_i \equiv \text{Richardson number} = (g z (T_a - T_z)) / (u_z^2 (T_a + 273.15)) \tag{10}$$

where

ϱ_a – density of air, kg/m^3
C_{pa} – specific heat of air, J/kg°C
k – von Karman's constant, 0.41
u_z – wind speed at height z, m/s
z – height of the wind speed measurement, m
z_o – roughness length, m
σ – empirical constant = 10
g – gravitational constant, m/s^2
T_z – temperature at height z, °C
T_s – effective surface temperature, °C
T_a – air temperature, °C

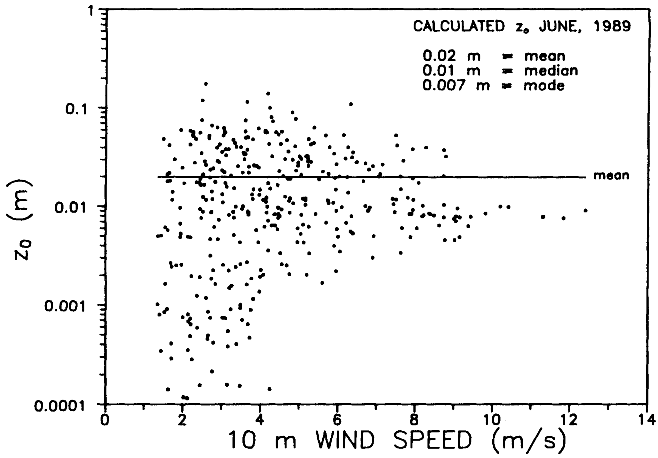

Fig. 1. Roughness lengths (Z_0) calculated from June 1989 wind profile data.

Daily heat exchange coefficients were adjusted to compensate for air stability based on the air temperature profile between the surface and 10 m using D_{hn} for neutral, D_{hs} for stable and D_{hu} for unstable conditions. The average surface roughness length, z_0, was determined from wind speed profiles between 1.5 and 10 m. A roughness length of 0.02 m was estimated by averaging calculations made from several hundred wind profile measurements during near neutral conditions (Fig. 1). This value falls well within the reported values for short grasses (Szeicz *et al.* 1969) and yields reasonable results. Weller and Holmgren (1974) calculated z_0 values between 0.02 and 0.04 m for fully developed tundra vegetation. Effective surface temperatures for the Q_h calculation were obtained by back calculation using the Stefan-Boltzman law and the emitted terrestrial longwave radiation.

The relative magnitudes of the components of the surface energy balance are presented for 1987, 1988 and 1989 on a weekly basis (Figs: 2 a, b and c). The meteorological measurements and subsurface soil temperatures, when made with calibrated sensors, produce very small errors. The most difficult measurement to make accurately was surface temperature. These energy balance calculations were very sensitive to slight variations in the surface temperature. Surface temperature was used in the calculation of both the convective heat flux (Q_h) and the conductive heat flux (Q_c). Surface temperature thermistors, even when placed very near the soil surface, recorded temperatures which were much too cool during the summer months when surface heating occurred. The temperatures measured by soil thermistors were probably low because of the slight insulation of the thermistor and because much of the evapotranspiration occurred at the leaf surface which is not in direct contact with the soil surface. Geiger (1965) suggests that the temperature of

Fig. 2. Relative magnitudes of components of surface energy balance calculated weekly during the summer of a) 1987, b) 1988, c) 1989.

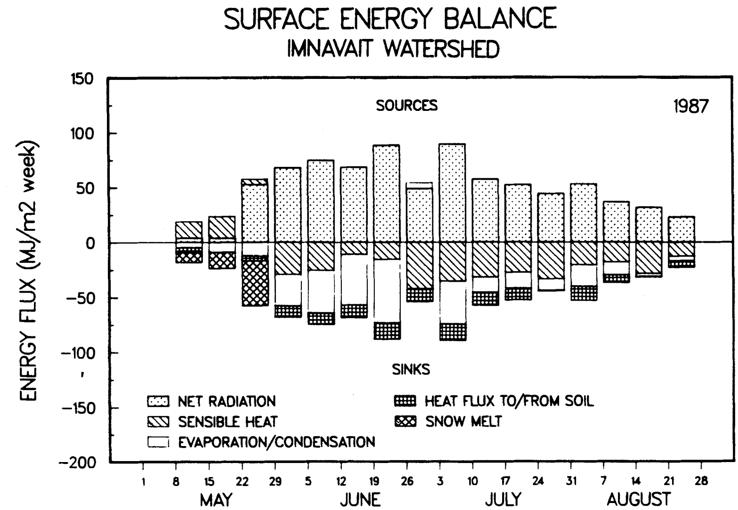

Fig. 2a.

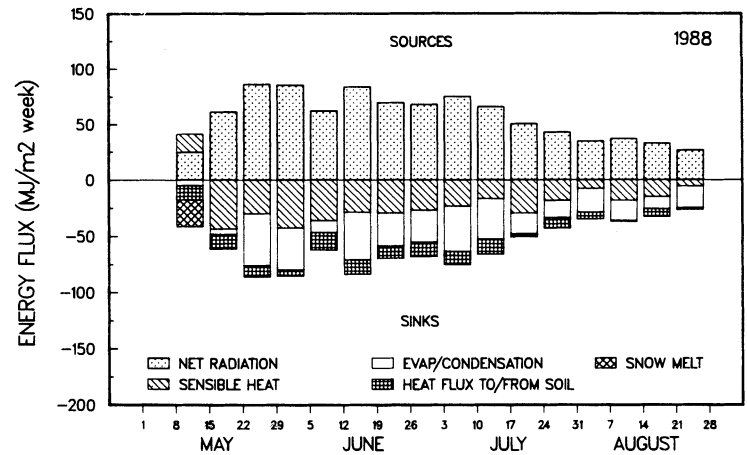

Fig. 2b.

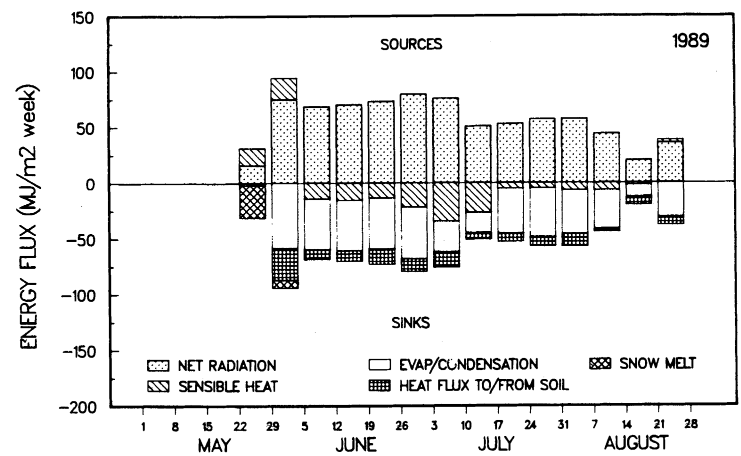

Fig. 2c.

the "outer effective surface" (T_S), which represents the energy exchange between the atmosphere and the soil, vegetation and boundary layer air, be used to calculate the convective heat exchange. The terrestrial longwave radiation was used to estimate this "outer effective surface" temperature. The radiometer was calibrated to surface emissivity by assuming a surface emissivity that yielded temperatures close to surface thermistor measurements in periods without intense sunlight just after the snowpack disappeared as the surface was going through phase change at 0°C. Rouse (1984b) used this method to determine the radiative surface temperature in energy balance calculations for a subarctic site in Canada.

Radiation is consistently the dominant source of energy during the snowfree period, but sensible heat transfer is quite important during snowmelt. During snowmelt, warm air is advected over the mountains from the south, warming the cold snowpack. Prior to snowpack ablation, the albedo is quite high and about 80 % of the incoming shortwave radiation is reflected, substantially reducing the net radiation absorbed. Theoretical and experimental work in Finland and Sweden (Kuusisto 1984; Lemmelä and Kuusisto 1974; Bengtsson 1980; Lemmelä 1972) has shown that evaporation from the snowpack is very small during the winter, increasing during the spring melt. Evaporation during the spring snowmelt in the Arctic is a significant component of both the hydrologic and energy balances.

Priestley-Taylor

The adaption of the Priestley and Taylor (1972) model derived by Rouse *et al.* (1977) was used to estimate evapotranspiration within the watershed. The equivalent depth of evaporated and transpired water was obtained by dividing the energy utilized for ET by the density of water and the latent heat of vaporization. The Priestley-Taylor method uses a control factor or coefficient to correct for surface moisture, temperature and vegetation conditions. Rouse *et al.* (1977) and Rouse (1984a) determined surface control factors for several types of tundra and woodland surfaces in the Canadian Subarctic. A control factor for a drier upland setting (upland lichen heath, $\alpha = 0.95$) was used for this comparison.

This method ignores heat conduction into the ground. Rouse (1984b) found conduction in a permafrost region to be greater than 15 % of net radiation during soil thaw and freeze back. This indicates that, at least in permafrost areas, conduction may be significant. From measurements of the energy balance, conduction was $\approx$17 % of the total net radiation over the course of the 1987 summer (Fig. 2a), $\approx$15 % in the 1988 summer (Fig. 2b) and $\approx$16 % in the 1989 summer (Fig. 2c).

The Priestley-Taylor method (Eq. (3)) relies on the Bowen ratio to apportion the net radiation between convective and evaporative losses. Weisner (1970) states that if the air temperature is measured outside of the boundary layer, under intense heating, this ratio varies considerably due to convection. Therefore, it may not accurately estimate evapotranspiration under the conditions that are common at this site.

Discussion of Results

Summarized in Table 1 are most of the evapotranspiration data from the North American Arctic. This data is useful for general comparisons, but one should avoid making detailed comparisons for several reasons. First on a spatial scale, this data has been derived for point, watershed, and lake areas. On the terrestrial side there can be considerable variation in vegetation, soils, and soil moisture availability. Second, on a temporal scale, measurements have been made at various times and durations and therefore are not directly comparable. Third, numerous methods have been used to calculate evapotranspiration. This is probably not as important as the quality of the data used in these methods or how it is reported. For example, some data is reported as actual versus potential, as in the case of pan evaporation. Finally, in many of these hydrologic studies *ET* was not the prime concern and was only a by-product of the study.

The average daily value of *ET* was determined for the period of record in each case. This average in some cases is for the entire summer period and in other studies only for a few days. Even with the restrictions mentioned several generalizations are evident from the data:

1) evapotranspiration decreases as the latitude increases,
2) evaporation losses from lakes are greater than evapotranspiration losses from an equivalent terrestrial area, and
3) evapotranspiration is greatest following snowmelt and decreases throughout the summer.

A comparison of calculated and measured *ET* for 1987, 1988 and 1989 are presented in Figs. 3 a, b and c. These are plots of cumulative precipitation, calculated *ET* by the Priestley-Taylor method (Eq. (3)) and the energy balance approach (Eq. (2)) and pan evaporation. The cumulative precipitation appears to be less than the estimates of evapotranspiration most of the time in Figs. 3 a, b and c. The cumulative precipitation includes only the rain and snow which fell after the spring melt. Some of the water from snowmelt is stored in the active layer and is the source of moisture for evapotranspiration in early summer when rainfall is low. Evaporation rates are usually highest during the early summer, as shown in Figs. 3 a, b and c. This is reasonable when one considers the pattern of incoming radiation, air temperatures and relative humidity. This appears to be a general trend in the Arctic supported by the results compiled in Table 1.

The total amount of evapotranspiration as determined from the mass balance technique ranged between 130 and 240 mm, *i.e.* 1.4 and 2.0 mm/day. The estimates from the energy balance method ranged between 1.4 and 2.3 mm/day and those of the Priestley-Taylor method ranged between 1.7 and 1.9 mm/day. The Priestley-Taylor, energy balance and pan evaporation methods are developed from point measurements. The results from a water balance (Table 2) are developed on a

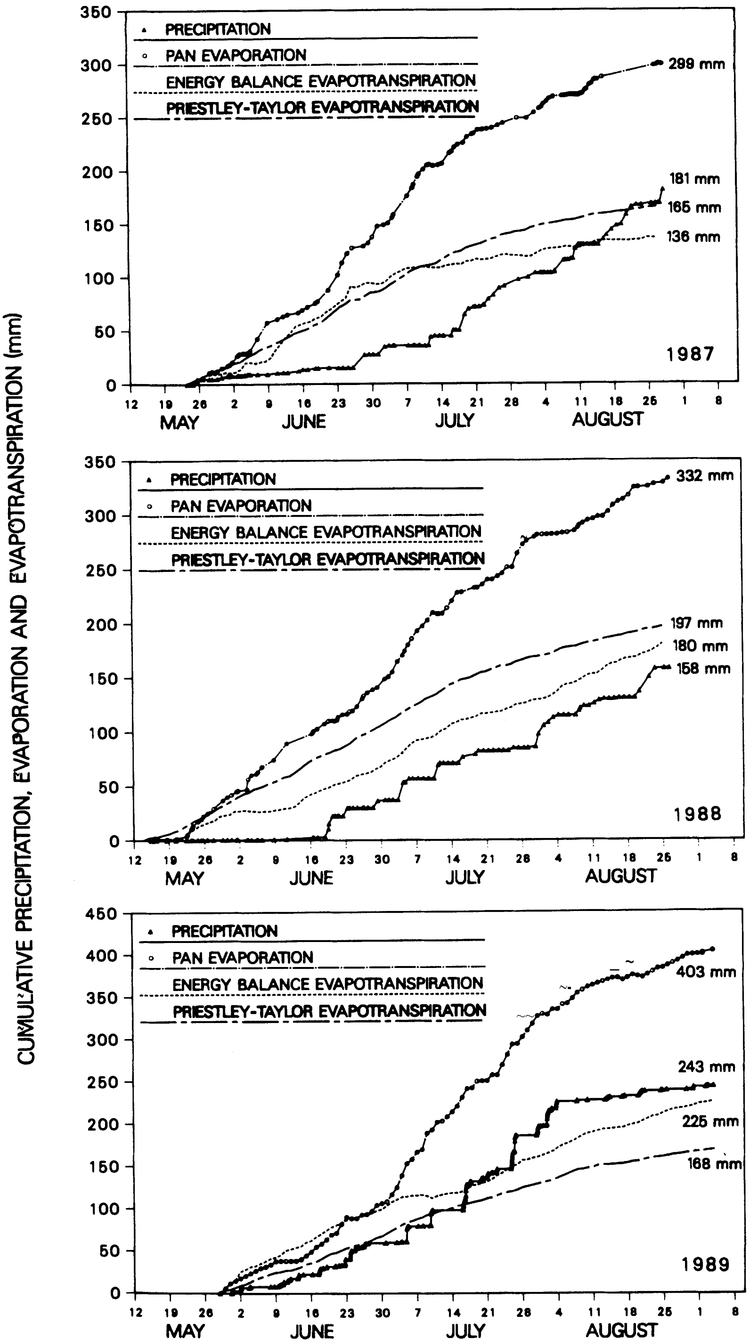

Fig. 3a.

Fig. 3b.

Fig. 3c.

Fig. 3. Comparison of precipitation, pan evaporation and calculations of evapotranspiration during the summer of a) 1987, b) 1988, c) 1989.

watershed scale. The amount of evapotranspiration could not be related with certainty to the season length, the amount of precipitation or the summation of the net radiation. It is undoubtedly related to the complex interaction of many of these features.

Conclusions

Evapotranspiration is a major component of the summer energy and water balances in the Arctic. In the summers of 1987, 1988, and 1989, *ET* consumed the equivalent of about 39 %, 46 % and 65 % of the net radiation respectively. From the calculation of the annual water balance in 1986, 1987, 1988 and 1989, *ET* represented 56 %, 34 %, 66 % and 58 % of the annual precipitation respectively. These two theoretically complete calculations of evapotranspiration were compared with the Priestley-Taylor method and pan evaporation. The techniques investigated all performed adequately to estimate *ET*. The estimates from the modified Priestley-Taylor method required average daily air temperature and net radiation. The estimates from pan evaporation required measurements from an evaporation pan and precipitation. Although quite variable throughout the summer and between years, the average *ET* of each method was quite close. The average *ET* from the energy balance approach was 1.8 mm/day, the Priestley-Taylor method yielded 1.8 mm/day, and the water balance method had an average of 1.7 mm/day. From these data the average pan coefficient to estimate actual evapotranspiration is 0.49.

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Address:
University of Alaska Fairbanks
Water Research Center,
Inst. of Northern Eng.,
460 Duckering Building,
Fairbanks, Alaska 99775-1760, U.S.A.