Surface Energy Balance of the Western and Central Canadian Subarctic: Variations in the Energy Balance among Five Major Terrain Types

ANDREA K. EATON AND WAYNE R. ROUSE

School of Geography and Geology, McMaster University, Hamilton, Ontario, Canada

PETER M. LAFLEUR

Department of Geography, Trent University, Peterborough, Ontario, Canada

PHILIP MARSH

National Water Research Institute, Environment Canada, Saskatoon, Saskatchewan, Canada

PETER D. BLANKEN

Department of Geography and Environmental Studies, University of Colorado, Boulder, Colorado

(Manuscript received 9 April 2000, in final form 30 January 2001)

ABSTRACT

In this study, the surface energy balance of 10 sites in the western and central Canadian subarctic is examined. Each research site is classified into one of five terrain types (lake, wetland, shrub tundra, upland tundra, and coniferous forest) using dominant vegetation type as an indicator of surface cover. Variations in the mean summertime values (15 June–25 August) of the energy balance partitioning, Bowen ratio (β), Priestley–Taylor alpha (α), and surface saturation deficit (D_o) are compared within and among terrain types. A clear correspondence between the energy balance characteristics and terrain type is found. In addition, an evaporative continuum from relatively wet to relatively dry is observed among terrain types. The shallow lake and wetland sites are relatively wet with high Q_E/Q^* (latent heat flux/net radiation), high α, low β, and low D_o values. In contrast, the upland tundra and forest sites are relatively dry with low Q_E/Q^*, low α, high β, and high D_o values.

1. Introduction

In Canada, the subarctic is an important ecoclimatic region accounting for approximately 20% of the total land area (Fig. 1). However, evaluating the physical characteristics of this ecoregion is difficult, because of its considerable size and its heterogeneous terrain. The heterogeneity is evident in the substantial site-to-site variation seen in physical properties such as vegetation, topography, and soil moisture. These variations contribute to significant differences in the surface energy balance among sites in this ecoregion.

In this study, the seasonal energy balance dynamics of 10 sites in the western and central Canadian subarctic are examined. The research sites are grouped into five distinct terrain types based on their vegetative, topographic, and hydrologic properties. The surface energy balance partitioning, Bowen ratios, Priestley–Taylor alphas, and surface saturation vapor pressure deficits are compared both within and among these five terrain types. It is hypothesized that sites belonging to a given terrain type will display similar energy balance characteristics.

Several prior studies have examined the summertime energy balance at one or more sites in the Canadian subarctic (Rouse et al. 1977; Rouse and Bello 1985; Lafleur and Rouse 1988; Bello and Smith 1990; Lafleur et al. 1992; Blanken and Rouse 1994; Lafleur 1994; Boudreau and Rouse 1995; Lafleur and Rouse 1995; Blanken et al. 2000; Petrone et al. 2000). In addition, short-term comparisons of the surface energy balance have been made among multiple sites on the Alaskan North Slope (Eugster et al. 1997; McFadden et al. 1998). Lynch et al. (1999) synthesized summertime energy balance data collected at numerous sites in the Alaskan North Shore as part of the Land–Atmosphere–Ice Interactions Flux Study. To date, however, there have been no comprehensive studies that explore the summertime

---

Corresponding author address: Andrea K. Eaton, School of Geography and Geology, McMaster University BSB 311, Hamilton, ON L8S 4K1, Canada.
E-mail: eatonak@mcmaster.ca

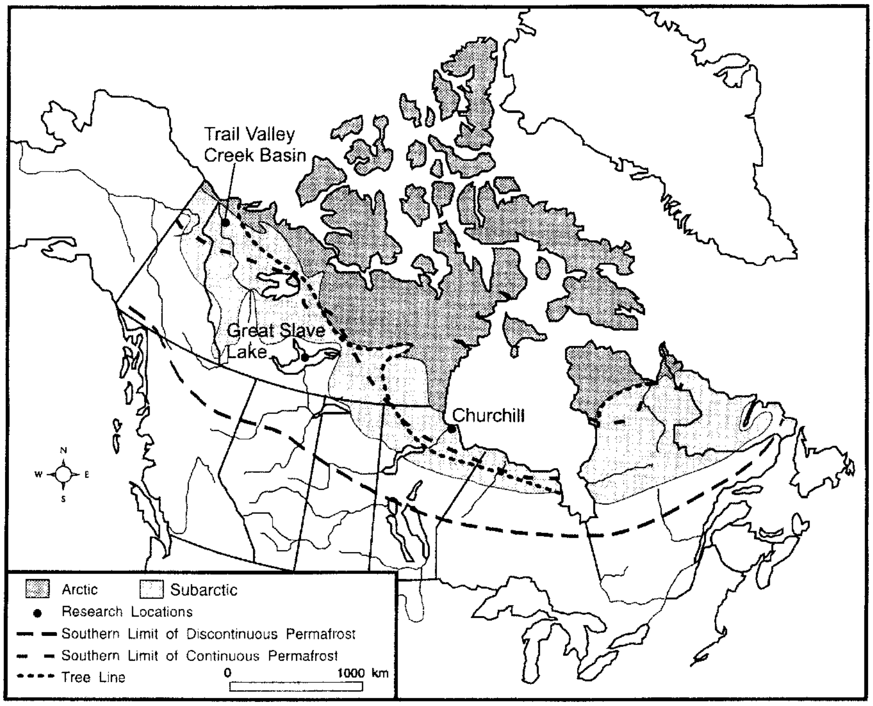

FIG. 1. The arctic and subarctic ecoclimatic regions of Canada (modified from Ecoregions Working Group 1989).

energy balance across multiple sites in the Canadian subarctic.

This study is part of the Global Energy and Water Balance Experiment (GEWEX), an international initiative aimed at improving our understanding of the connectivity among the energy balance, the water balance, and the climate system. The Mackenzie GEWEX Study (MAGS) is a component of GEWEX initiated in order to understand and model energy and water cycles in the Mackenzie River basin and to assess changes to these cycles that may arise from natural climate variability and anthropogenic climate change (Stewart et al. 1998; Rouse 2000).

2. Study sites

Data used in this study have been collected from 10 sites representing many of the terrain types found in the western and central Canadian subarctic (Fig. 2). At each site, data were collected from approximately 15 June to 25 August, which roughly corresponds to the summer growing season. At most sites, data from multiple (usually successive) years were available. Many of the datasets obtained for this study have been previously published. A brief description of each site is provided below, and a summary of the site measurements (including references to prior studies that have produced datasets used in this study) is given in Table 1.

a. Trail Valley Creek lowland (Tw1)

This wetland tundra site is located in the Trail Valley Creek basin (TVC) approximately 150 km south of the Beaufort Sea. The terrain comprises mineral soil tussocks, with peat in the intertussock hollows. It is underlain by continuous permafrost with an active layer depth of 0.5 m in early September (Petrone et al. 2000). Mackenzie sedge (*Carex mackenzii*), which grows in tussocks, is the dominant vascular vegetation. A 20–25-cm-deep *Sphagnum* mat is found in the intertussock hollows. Other vascular vegetation includes cotton grass (*Eriophorum* spp.), Labrador tea (*Ledum groenlandicum*), scrub birch (*Betula glandulosa*), trailing willow (*Salix arctophila*), and numerous berry species (Petrone et al. 2000).

b. Churchill sedge fen (Tw2)

This sedge wetland is located 18 km southeast of Churchill, Manitoba (CH). The substrate is porous peat (0.3–0.5 m thick) overlying a thick layer of glaciomarine silty clay (Lafleur and Rouse 1995). Carbonate cobbles form a thin layer at the top of the marine clays (Burton et al. 1996). The terrain is hummocky, and both the hummocks and hollows are comprised of peat (Petrone et al. 2000). The site is underlain by continuous permafrost with an active layer depth of 1.0 m in late August (Lafleur et al. 1992). The primary vascular vegetation is water sedge (*Carex aquatilis*), but rocky-

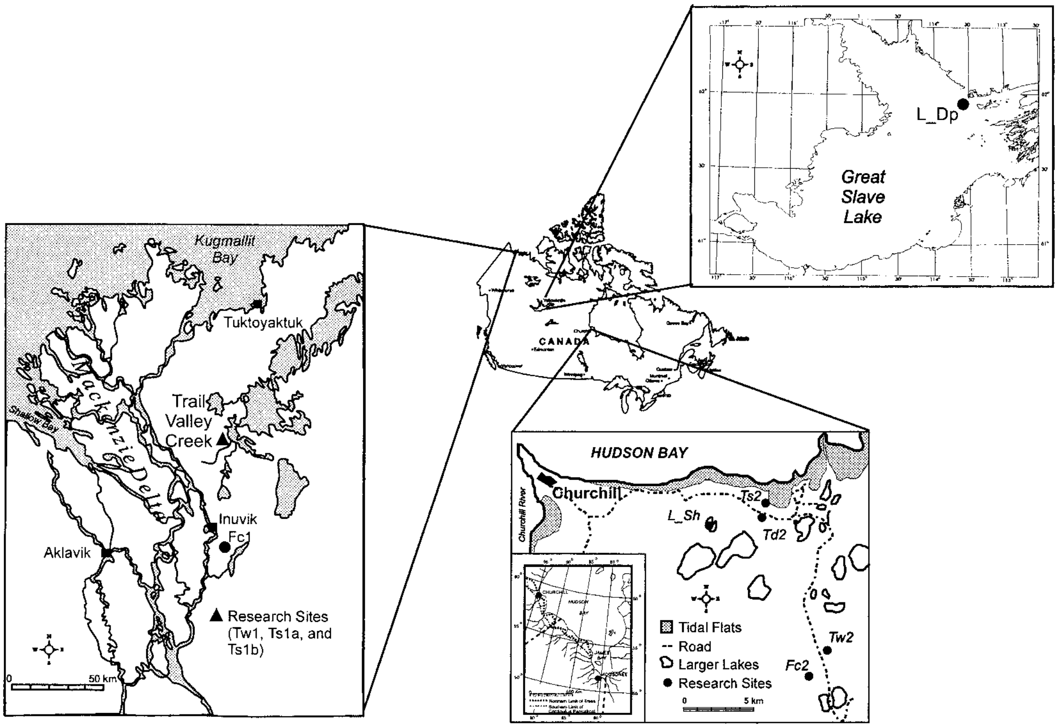

FIG. 2. Research site locations.

ground sedge (*Carex saxatilis*), northern bog sedge (*Carex gynocrates*), and mud sedge (*Carex limosa*) are also present. Other vascular vegetation includes dwarf Labrador tea (*Ledum decumbens*), scrub birch, trailing willow, deer grass (*Scirpus caespitosus*), and purple saxifrage (*Saxifraga oppositifolia*). Nonvascular vegetation consists primarily of lichens from the *Cetraria* and *Cladina* genera. In addition, a thin layer of moss (*Scorpidium turgescens*) covers approximately 15% of the surface (Burton et al. 1996; Petrone et al. 2000).

c. Trail Valley Creek shrub tundra 1 (Ts1a)

This shrub tundra site is located on a hilltop in TVC. The soils are primarily mineral, with an organic content of $< 5\%$ in the top 0.4 m. The active layer in this continuous permafrost region reaches a depth > 0.80 m by early September (Petrone et al. 2000). The microtopography includes small hummocks that affect the distribution of vegetation. Lichens, mosses, various berry species, and low deciduous shrubs [e.g., alders (*Alnus* sp.), birches, and willows] dominate the hummocks, while sedges [Mackenzie sedge and golden sedge (*Carex aurea*)] and trailing willow dominate the interhummock areas (Marsh and Pomeroy 1996).

d. Trail Valley Creek shrub tundra 2 (Ts1b)

At this shrub tundra site, scrub birch covers approximately 70% of the surface area. The birch canopy is underlain by a lichen–heath mat that comprises Labrador tea, crowberry (*Empetrum nigrum*), cranberry (*Vaccinium vitis-idaea*), cloudberry (*Rubus chamaemorus*), and numerous lichens (e.g., *Cetraria* spp., *Cladina* spp., and *Alectoria nigricans*) (R. M. Petrone 1998, personal communication).

e. Churchill shrub tundra (Ts2)

At the CH shrub tundra site, soils consist of a 0.20-m organic layer covering a thick layer of sand (Blanken and Rouse 1994). Unlike the two TVC shrub tundra sites, the surface at this location is always damp. Standing water occupies, on average, 6% of the surface area. Continuous permafrost occurs beneath this site, and the active layer depth is 1.0 m in early September (Blanken and Rouse 1995). Vegetation includes silver willow (*Salix candida*), flat-leaved willow (*Salix planifolia*), snow willow (*Salix reticulata*), scrub birch, and water sedge (Blanken and Rouse 1994).

f. Churchill upland lichen–heath tundra (Td2)

Soils at this lichen–heath tundra site are poorly developed and well drained, consisting of a thin organic layer (<0.01 m) over sand. Vegetation is distributed evenly between lichen and heath. Common lichens include members of the *Cladina, Cetraria,* and *Alectoria* genera (Boudreau and Rouse 1995). The heath consists of low-growing vascular vegetation such as white moun-

TABLE 1. Research site measurements [*for terrestrial sites, "1" indicates a western subarctic site, and "2" indicates a central subarctic site; **this column contains references to prior studies that have produced datasets used in this study (n/a = not available)].

Site	Site I.D.*	Lat	Lon	Energy balance methodology	Data year(s)	References**
Trail Valley Creek lowland	Tw1	68°45′N	133°32′W	VPEC	1996, 1997	Petrone et al. (2000)
Churchill sedge fen	Tw2	58°40′N	94°40′W	BREB	1990–95	Boudreau and Rouse (1995) Lafleur and Rouse (1995) Schreader et al. (1998)
Trail Valley Creek shrub tundra 1	Ts1*a*	68°45′N	133°32′W	SEC	1996, 1997	Petrone et al. (2000)
Trail Valley Creek shrub tundra 2	Ts1*b*	68°45′N	133°32′W	VPEC	1997	n/a
Churchill shrub tundra	Ts2	58°45′N	94°37′W	BREB	1990, 1991	Blanken and Rouse (1994) Boudreau and Rouse (1995)
Churchill upland tundra	Td2	58°45′N	94°37′W	BREB	1991, 1996	Boudreau and Rouse (1995)
Havikpak Creek forest	Fc1	68°32′N	133°32′W	VPEC	1996	n/a
Churchill spruce–tamarack forest	Fc2	58°39′N	94°39′W	BREB	1993, 1994	Lafleur (1994) Lafleur and Rouse (1995)
Golf Lake	L_Sh	58°46′N	94°33′W	BREB	1991, 1995	Boudreau and Rouse (1995)
Great Slave Lake	L_Dp	61°55′N	113°43′W	SEC	1997, 1998	Blanken et al. (2000)

SEC = eddy correlation using a sonic anemometer.
VPEC = eddy correlation using a vertical propeller anemometer.
BREB = Bowen ratio–energy balance method.

tain-avens (*Dryas integrifolia*), dwarf Labrador tea, Lapland rose-bay (*Rhododendron lapponicum*), crowberry, red bearberry (*Arctostaphylos rubra*), and bog rosemary (*Andromeda polifolia*) (Boudreau and Rouse 1995).

g. Havikpak forest (Fc1)

Havikpak forest is located in the Havikpak Creek basin (15 km^2), which is 15 km south of Inuvik, Northwest Territories, in the Mackenzie Delta. The terrain is sloped and well drained, resulting in a dry surface. This open-canopy forest comprises primarily black spruce (*Picea mariana*) with some poplars (*Populus* spp.) growing along the ridge tops. The understory consists of a lichen mat with protruding vascular vegetation (M. Russell 1998, personal communication).

h. Churchill spruce–tamarack forest (Fc2)

At the CH coniferous forest site, soils are glaciomarine silts with a 0.25-m-thick layer of fibrous peat at the surface (Lafleur 1992). There is no permafrost beneath this site and soils have typically thawed completely by mid-July (Lafleur et al. 1992). The forest canopy consists of 48% tamarack (*Larix laricina*), 40% black spruce, and 12% white spruce (*Picea glauca*). The trees are widely spaced with a leaf area index of approximately 1.6 (Lafleur 1992). The understory consists of small shrubs (e.g., scrub birch, trailing willow, and Labrador tea), mosses, lichens, and some standing water (Lafleur and Schreader 1994).

i. Golf Lake (L_Sh):

Golf Lake is relatively small (0.2 km^2 in area) and shallow ($<$2 m deep). The lake is oval in shape with its long axis oriented northeast–southwest. It is surrounded by sedge wetlands to the south, north, and west, and upland lichen–heath tundra to the east. The lake level is strongly dependent on precipitation input and varies seasonally as precipitation patterns change (Boudreau 1993).

j. Great Slave Lake (L_Dp):

Great Slave Lake is the fourth largest lake in North America, covering an area of roughly 28 570 km^2. It is irregular in shape, with several long arms extending from the main body. Lake depths can exceed 600 m in the lake's east arm, but the average depth of the main body is 41 m (Rawson 1950). The research site is located near the main body of the lake on a rock islet known as Inner Whaleback Island.

3. Theory and methods

Based on their vegetative, topographic, and hydrologic characteristics, the research sites are grouped into five distinct terrain types. These are wetlands (Tw1, Tw2), shrub tundra (Ts1*a*, Ts1*b*, Ts2), upland lichen–heath tundra (Td2), coniferous forests (Fc1, Fc2), and lakes (L_Sh, L_Dp). To examine the variability among terrain types with respect to their energy and water balances, mean values of the energy balance partitioning, Bowen ratio (β), Priestley–Taylor alpha (α), and surface saturation vapor pressure deficit (D_o) are compared. All variables are calculated over 30-min intervals, which are averaged to produce summertime mean values.

a. Surface energy balance

The surface energy balance is expressed differently for terrestrial and lake sites. For a homogenous terrestrial surface, the energy balance is represented by

$$Q^* = Q_E + Q_H + Q_G, \tag{1}$$

where Q^* is net radiation (W m^{-2}), Q_E is the latent heat flux (W m^{-2}), Q_H is the sensible heat flux (W m^{-2}), and Q_G is the ground heat flux (W m^{-2}). For a lake, the energy balance is determined using the equation:

$$Q^* = Q_E + Q_H + Q_W + Q_B, \tag{2}$$

where Q_W is the lake heat storage (W m^{-2}) and Q_B is the heat flux through the bottom of the lake (W m^{-2}). Heat flux Q_B is negligible for deep lakes and has been shown to account for $<$ 4% of Q^* for shallow subarctic lakes (Marsh and Bigras 1988; Bello and Smith 1990). It is not included in this study.

Energy balance partitioning is used to make direct comparisons among the surface energy balance characteristics at each site (and each terrain type). Energy balance partitioning distributes the total available energy at the surface among the energy balance components by calculating the ratios Q_E/Q^*, Q_H/Q^*, and Q_G/Q^* (or Q_W/Q^*). These ratios indicate the relative magnitudes of Q_E, Q_H, and Q_G (or Q_W) in the surface energy balance.

Three different approaches are used to evaluate the convective heat fluxes: the Bowen ratio–energy balance method (BREB), eddy correlation using a sonic anemometer (SEC), and eddy correlation using a vertical propeller anemometer (VPEC) (Table 1). For sites at which data have been previously published, details regarding instrumentation and data processing (i.e., instrument heights, fetches, missing data handling, etc.) are given in the references cited in Table 1. For site Ts1*b*, instrumentation and data processing techniques are identical to those at site Tw1. The reader is referred to Petrone et al. (2000) for details.

At site Fc1, a VPEC system is used to determine Q_H. The vertical wind sensor is mounted 8.5 m above the surface, adjacent to a fine wire chromel–constantan thermocouple (lateral separation = 0.04 m). Ground heat flux Q_G is measured with a soil heat flux plate buried 0.025 m below the surface. Calorimetric corrections (Rouse 1984a; Halliwell and Rouse 1987) are applied to the Q_G data to adjust for permafrost soils. Latent heat flux Q_E is solved as a residual of the energy balance equation [Eq. (1)]. Radiation sensors (net radiometer and pyranometer) are mounted 8.5 m above the surface. Air temperature and relative humidity are measured at a height of 1.5 m. Precipitation is measured with a tipping-bucket rain gauge located near the meteorological tower. The fetch exceeds 800 m to the north, east, and west. To the south, a change in topography restricts the fetch to 250 m; however, vegetation cover remains uniform beyond this boundary. Data processing techniques are identical to those described for site Tw1 by Petrone et al. (2000).

b. Bowen ratio

The Bowen ratio (β) is the ratio of the sensible and latent heat fluxes above a surface (Bowen 1926):

$$\beta = \frac{Q_H}{Q_E} = \gamma\left(\frac{\Delta T_a}{\Delta e_a}\right), \tag{3}$$

where γ is the psychrometric constant (65 Pa K^{-1}), T_a is air temperature (K), and e_a is atmospheric vapor pressure (Pa). When $\beta > 1$, more energy is directed from the surface as sensible heat than as latent heat. In contrast, when $\beta < 1$, the latent flux is the primary pathway for convective heat loss. Wet surfaces have a lower β than dry surfaces.

c. Priestley–Taylor alpha

The Priestley–Taylor alpha coefficient (α) is determined using the following equation (Priestley and Taylor 1972):

$$Q_E = \alpha\left[\frac{S}{(S + \gamma)}\right](Q^* - Q_G), \tag{4}$$

where S is the slope of the saturation vapor pressure–temperature curve (Pa K^{-1}). By incorporating the Bowen ratio into Eq. (4), α can be calculated as follows:

$$\alpha = \frac{(S + \gamma)}{S(1 + \beta)}. \tag{5}$$

Coefficient α varies widely according to surface type and micrometeorological conditions. Wet surfaces have a higher α than dry surfaces.

d. Surface saturation vapor pressure deficit

Another way to evaluate surface–atmosphere interactions is to measure the surface saturation vapor pressure deficit (D_o). Deficit D_o represents how close the thin layer of air (~1 cm) immediately above the ground surface is to saturation. It is calculated as the difference between the surface saturation vapor pressure (e_{so}) and the ambient surface vapor pressure (e_o):

$$D_o = e_{so} - e_o. \tag{6}$$

In this study, D_o is calculated only for the terrestrial sites because it is assumed that the surface air over the lakes is saturated at all times ($D_o = 0$).

The quantity e_{so} is a function of surface temperature (T_o) and is calculated using (Bolton 1980) the following equation:

$$e_{so} = 0.6112\ \exp\left(\frac{17.67T_o}{T_o + 243.5}\right). \tag{7}$$

Traditionally, e_o is determined by extrapolating from an atmospheric vapor pressure profile (Sellers 1965; Thom 1975; Halliwell and Rouse 1989; Oke 1992). However, this method could not be used in this study because vapor pressure profiles were not measured at all sites. Instead, e_o is derived from Eq. (3) as follows:

$$e_o = e_a - \left[\frac{\gamma(T_a - T_o)}{\beta}\right], \qquad (8)$$

where atmospheric vapor pressure (e_a) and air temperature (T_a) are measured at the same height.

It should be noted that vapor pressure profiles are measured at three of the ten research sites used in this study (Tw2, Ts2, and Td2). As a result, the accuracy of e_o values calculated using Eq. (8) can be verified. At these three sites, daily mean e_o calculated using Eq. (8) are compared with daily mean e_o extrapolated from a vapor pressure profile (L. D. Boudreau, unpublished data) for the summer period. At all sites, the mean percent differences between the two sets of e_o values is $<$ 10%. This suggests that Eq. (8) is a reliable tool for calculating the ambient surface vapor pressure.

4. Data compatibility

Prior to analysis, steps were taken to ensure that comparisons within and among terrain types were valid. Each dataset was restricted to the same period (15 June–25 August). Daily means were adjusted to represent the 24-h period centered on solar noon. These measures ensure that the same seasonal and diurnal characteristics are captured in each dataset.

A potential limitation to this study exists because comparisons are made among data obtained using three different approaches for determining the surface energy balance: BREB, SEC, and VPEC (Table 1).

To evaluate how BREB and SEC systems compare, daytime Q_H data determined using both methods are examined for two sites (Tw2 and Fc2; Fig. 3). Data were collected 23–25 August 1997 at site Tw2, and from 2 to 14 July 1992 at site Fc2. At both sites, similar Q_H fluxes are measured by each system. This suggests that, at least for these sites and periods, comparisons among convective fluxes determined using these two methodologies are valid.

Because of lack of availability, it is not possible to make comparisons between VPEC data and BREB or SEC data at any of the sites in this study. VPEC systems, however, are known to consistently underestimate Q_H because vertically oriented propeller anemometers tend to stall at low wind speeds and to be unresponsive to high-frequency eddies (Brutsaert 1982; Blanford and Gay 1992). To compensate for this, a frequency response correction factor (Moore 1986; Blanford and Gay 1992) was applied to Q_H at the three sites employing VPEC systems. At Tw1 and Ts1*b* a correction factor of 1.29 was used, while at Fc1 a correction factor of 1.32 was used. These values fall in the midrange of correction factors calculated for VPEC systems in other studies (Tsvang et al. 1973; McNeil and Shuttleworth 1975; Moore 1976; Spittlehouse and Black 1979; Blanford and Gay 1992). Blanford and Gay (1992) found that VPEC data corrected in this manner closely approximate SEC measurements. Thus, we believe that the comparisons among corrected VPEC Q_H data and BREB or SEC Q_H data made in this study are valid, and they have been verified using the best of our resources.

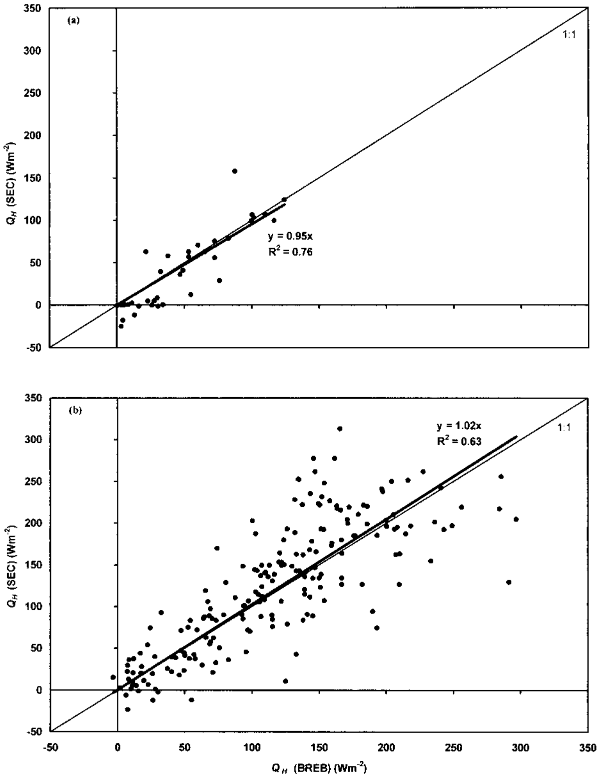

FIG. 3. Daytime values (averaged over 30-min intervals) of the sensible heat flux (Q_H) measured using the Bowen ratio–energy balance method (BREB) and eddy correlation using a sonic anemometer (SEC) at (a) site Tw2 and (b) site Fc2.

5. Results and discussion

At each research site, mean summertime values (15 June–25 August) of the energy balance partitioning, β, and α are calculated for each data year. Generalized mean values, which average each variable over all years, are also calculated where possible (Table 2). The data show that terrestrial sites belonging to the same terrain type classification have similar values for all descriptor variables. In addition, an evaporative continuum, from wet (shallow lakes) to dry (coniferous forests), is evident across terrain types (Fig. 4).

The shallow lake (L_Sh) and wetland sites (Tw1, Tw2) have the highest Q_E/Q^*. This is due to high moisture availability and low surface resistance to evaporation. The three shrub tundra sites (Ts1*a,* Ts1*b,* Ts2) exhibit midrange Q_E/Q^*. This is consistent with their vegetative cover; there is a large amount of available surface moisture, but there is also a higher vegetative resistance to evapotranspiration. The upland lichen–heath tundra (Td2) has a small Q_E/Q^*. This is due to high surface

TABLE 2. Mean summertime values of energy balance partitioning [Q_E/Q^*, Q_H/Q^*, Q_G/Q^* (or Q_W/Q^*)], Bowen ratio (β), Priestly–Taylor alpha (α), and surface saturation deficit (D_o) for individual data years at each research site. Generalized mean values (across all data years) are also included where possible.

Site	Terrain type	Year	Q_E/Q^*	$Q_H Q^*$	Q_G/Q^* [a](Q_W/Q^*)	β	α	D_o (kPa)
L_Dp	Deep lake	1997	0.50	0.04	0.45[a]	0.09	1.51	—
		1998	0.18	−0.06	0.90[a]	−0.31	2.32	—
		Average	0.34	−0.01	0.68	−0.11	1.92	—
L_Sh	Shallow lake	1991	0.85	0.08	0.05[a]	0.09	1.45	—
		1995	0.64	0.38	0.01[a]	0.60	1.17	—
		Average	0.75	0.23	0.03	0.34	1.31	—
Tw1	Wetland	1996	0.64	0.30	0.05	0.46	1.07	0.30
		1997	0.64	0.27	0.07	0.42	1.10	0.25
		Average	0.64	0.28	0.06	0.44	1.08	0.28
Tw2	Wetland	1990	0.65	0.25	0.11	0.39	1.20	0.46
		1991	0.85	0.17	0.09	0.20	1.42	0.55
		1992	0.55	0.35	0.10	0.64	1.08	0.31
		1993	0.58	0.30	0.13	0.51	1.08	0.42
		1994	0.49	0.40	0.11	0.81	0.83	0.78
		1995	0.73	0.22	0.10	0.30	1.46	0.33
		Average	0.64	0.28	0.11	0.48	1.18	0.48
Ts1*a*	Shrub tundra	1996	0.50	0.30	0.20	0.59	1.00	0.39
		1997	0.53	0.26	0.22	0.50	1.08	0.51
		Average	0.52	0.28	0.21	0.54	1.04	0.45
Ts1*b*	Shrub tundra	1997	0.57	0.28	0.15	0.49	1.07	0.37
Ts2	Shrub tundra	1990	0.49	0.36	0.15	0.74	0.95	0.25
		1991	0.63	0.23	0.14	0.36	1.20	0.38
		Average	0.56	0.30	0.14	0.55	1.08	0.32
Td2	Upland tundra	1991	0.45	0.47	0.09	1.03	1.00	0.98
		1996	0.54	0.30	0.18	0.55	0.81	0.91
		Average	0.50	0.38	0.14	0.79	0.90	0.94
Fc1	Coniferous forest	1996	0.46	0.45	0.09	0.98	0.77	0.70
Fc2	Coniferous forest	1993	0.53	0.37	0.10	0.71	0.95	0.46
		1994	0.56	0.40	0.04	0.72	0.94	0.63
		Average	0.54	0.39	0.07	0.72	0.94	0.54

resistance to evaporation from the nontranspiring lichen vegetation and the small water-holding capacity of the sandy substrate.

The smallest terrestrial Q_E/Q^* occurs at the spruce forest site (Fc1). Evapotranspiration from this site is limited by the stunted physiognomy of the trees, the abundant lichen in the understory, and the well-drained sloping terrain. The spruce–tamarack forest site (Fc2) experiences a higher Q_E/Q^* than Fc1. This is attributed to more freely available water due to ponding on the forest floor.

The deep lake (L_Dp) exhibits quite different behavior from the other research sites. It has a Q_E/Q^* of only 0.34. This is due to the large percentage of the total energy budget used to heat the water body (resulting in a very large Q_W/Q^*). Since so much energy is stored in the lake, little is left over for evaporation. It is important to note, however, that this condition is only representative of the summer season. In the fall and early winter, when the lake is warmer than the atmosphere, the stored energy promotes a large evaporative heat flux (Blanken et al. 2000).

It should be noted that Q_E/Q^* at all of the above sites (except L_Dp) exceeds 0.46. We believe that a full range of subarctic terrain types (from very wet to very dry) are represented in this study, and that the relatively large Q_E/Q^* indicates that, on average, evapotranspiration is the principle component of the summertime energy budget in this region. The study site locations are such, however, that airmass control can alter the energy balance partitioning on a day-to-day basis (i.e., lower Q_E/Q^* values occur on days with cold, onshore winds). This effect has been demonstrated over several sites in the North American subarctic (Rouse 1984b; Rouse and Bello 1985; Rouse et al. 1987; Weick and Rouse 1991; Harazono et al. 1998). It is also evident in a study that examines the surface energy balance characteristics at our study sites during warm and cold temperature quartiles (Eaton 1999).

The shallow lake, wetland, and shrub tundra sites have a Q_H/Q^* near 0.28. At the upland and coniferous forest sites this ratio is slightly higher—with a maximum of 0.45 over the spruce forest (Fc1). In all cases, Q_G/Q^* ranges from 0.06 to 0.21. The magnitude of the ground heat flux is large as compared with more temperate regions; however, a large Q_G is typical of subarctic landscapes. This is due to the steep temperature gradient that exists in the substrate between the surface and the permafrost layer, combined with a high thermal conductivity in the melting frost zone (Rouse et al. 1997).

Sites with high Q_E/Q^* (shallow lake and wetlands)

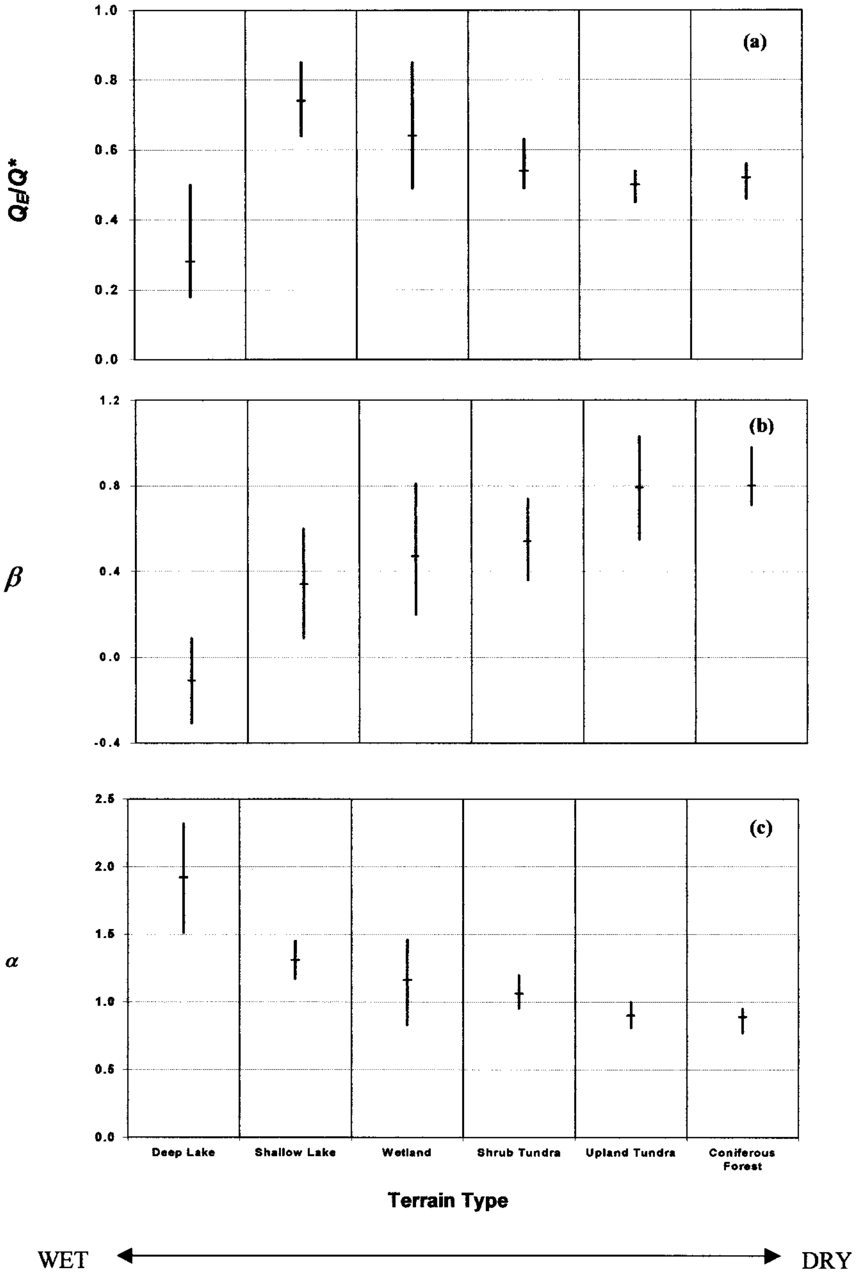

FIG. 4. Range of mean summertime (a) Q_E/Q^*, (b) Bowen ratios (β), and (c) Priestley–Taylor alphas (α) experienced at each terrain type. Ranges include all annual mean summertime values from sites belonging to a given terrain type (e.g., "wetland" includes data from sites Tw1 and Tw2).

have a relatively low β, while sites with low Q_E/Q^* (upland tundra and coniferous forests) have a high β. At the deep lake site (L_Dp), there is a negative mean β. This is a result of warm air advection and persistent downward-directed sensible heating during the summer period. In all cases mean βs are less than 1.

High α values are associated with a high Q_E/Q^*, while low values are associated with a low Q_E/Q^*. In this study, the shallow lake and wetland sites have higher α values than the upland tundra and forest sites. The extremely high α over the deep lake is a result of strong advective conditions at that site during the summer period. It is, however, meaningless in terms of actual evaporation rates (Blanken et al. 2000).

Energy balance studies conducted at other locations in the North American subarctic have produced energy partitioning data similar to the values in the present study (Table 3). For shallow subarctic lakes, Q_E/Q^* was found to exceed 0.74 by both Stewart and Rouse (1976) and Bello and Smith (1990). For subarctic wetlands,

TABLE 3. Mean summertime values of the energy balance partitioning [Q_E/Q^*, Q_H/Q^*, Q_G/Q^* (or Q_W/Q^*)], Bowen ratio (β), and Priestley–Taylor alpha (α) from other sites in the North American subarctic (— data not available).

Terrain type	Site location	Q_E/Q^*	Q_H/Q^*	Q_G/Q^* [a](Q_W/Q^*)	β	α	Energy balance methodology	Source
Shallow lake	Hudson Bay Coast, Ontario	0.74	0.26	0.00[a]	0.35	1.26	BREB	Stewart and Rouse (1976)
Shallow lake	Northern Manitoba	1.05	−0.14	0.04[a]	−0.13	1.35	BREB	Bello and Smith (1990)
Wetland	Hudson Bay Coast, Ontario	0.68	0.30	0.06	0.44	1.26	BREB	Rouse et al. (1977)
Wetland	Lake Melville, Newfoundland	—	—	—	—	1.20	BREB	Price et al. (1991)
Wetland	Schefferville, Quebec	0.63	0.25	0.10	0.50	—	SEC	Moore et al. (1994)
Wetland	Happy Valley, Alaska	0.57	0.29	0.09	0.51	—	SEC	Harazono et al. (1998)
Upland tundra	Hudson Bay Coast, Ontario	0.57	0.41	0.04	0.72	0.95	BREB	Rouse et al. (1977)
Upland tundra	Bethel, Alaska	0.40	0.40	0.20	1.1	—	SEC	Fitzjarrald and Moore (1992)
Upland tundra	Happy Valley, Alaska	0.49	0.40	0.16	0.81	—	SEC	Harazono et al. (1998)
Coniferous forest	Lake Athabasca, Northwest Territories	0.63	0.32	0.05	0.51	1.13	BREB	Rouse et al. (1977)
Coniferous forest	Schefferville, Quebec	0.29	0.64	0.07	2.5–3.0	—	SEC	Fitzjarrald and Moore (1994)

SEC = eddy correlation using a sonic anemometer.
BREB = Bowen ratio–energy balance method.

Q_E/Q^* was reported as approximately 0.63 by Rouse et al. (1977), Moore et al. (1994), and Harazono et al. (1998). These results are consistent with values from this study, and support the claim that, on average, evapotranspiration from these terrain types is the principle component of Q^*. Values for Q_H/Q^* and Q_G/Q^* (or Q_W/Q^*) reported in the above studies are also similar to those from this study—with one exception. Bello and Smith (1990) report a downward-directed sensible heat flux over a shallow lake in the Hudson Bay Lowland. This was attributed to advection of warm air over the lake from the adjacent dry lichen–heath tundra.

Measurements over upland lichen–heath tundra have found that Q_E/Q^* ranges from 0.40 to 0.57 (Rouse et al. 1977; Fitzjarrald and Moore 1992; Harazono et al. 1998), while Q_H/Q^* is always near 0.40. Again, these ratios correspond to the data presented in this study. To the authors' knowledge, there is no work that examines the summertime energy balance over North American shrub tundra at sites other than those included in this study.

The literature values for energy balance partitioning at open canopy coniferous forests show the widest variation. Rouse et al. (1977) report a Q_E/Q^* of 0.63 from a black spruce forest near Lake Athabasca, Northwest Territories, while Fitzjarrald and Moore (1994) found the ratio to be only 0.29 from a black spruce forest near Schefferville, Quebec. The coniferous forest sites in the present study behave similar to the Rouse et al. (1977) site, with Q_E as the dominant component of the surface energy balance. In contrast, the Schefferville data show that Q_H is the dominant convective flux. This latter result is similar to those from energy balance studies at coniferous forest sites in the northern boreal ecoregion.

McCaughey et al. (1997), Joiner et al. (1999), and Moore et al. (2000) report that Q_H dominates the energy balance partitioning in boreal Jack Pine (*Pinus banksiana*) forests. However, a low Q_E/Q^* is expected at Jack Pine forest sites because they are ecologically adapted dry environments. Betts et al. (1999) found Q_H to be the dominant convective flux at a boreal spruce forest site. The low Q_E flux at this site can be attributed to the forest's closed canopy. In boreal spruce forests, the closed canopy suppresses evaporation from the understory, resulting in a lower Q_E than at the open-canopy spruce forests of the subarctic. Northern tree species are adapted for water conservation; thus, direct evaporation from the forest floor is the primary component of stand evapotranspiration at high latitudes. In fact, Lafleur and Schreader (1994) have shown that water loss from the understory is the largest component of stand evapotranspiration for a subarctic forest (site Fc2).

Literature values of β and α from other sites in the North American subarctic correspond well to the data in this study, with two exceptions. In the first case, Fitzjarrald and Moore (1994) report a higher β for their Schefferville forest site. This corresponds to the low Q_E/Q^* noted above. In the second case, Bello and Smith (1990) found a negative β over their shallow tundra lake (and a higher α than found at shallow lakes in the present study) due to the very strong advectional effects also noted above.

Variations in energy balance characteristics among research sites in this study of the subarctic environment are clearly associated with terrain type. Terrestrial sites belonging to the same terrain type classification have similar Q_E/Q^*, β, and α values. The two lake sites, however, behave quite differently from each other. This difference indicates that shallow and deep lakes do not use Q^* in the same manner because of the large dif-

TABLE 4. Mean summertime temperature and total summertime precipitation in each data year at the meteorological station nearest to each research site relative to the 30-yr climate normals (1961–90) from the same meteorological station. ''Normal'' represents temperatures within ±2°C and precipitation within ±20% of the 30-yr normal. (*Data from Yellowknife A are not representative of conditions over Great Slave Lake.)

Meteorological station	Corresponding research sites	Data year	Temperature	Precipitation
Yellowknife A	L_Dp*	1997	Normal	Dry
		1998	Hot	Dry
Churchill A	L_Sh, Tw2, Ts2, Td2, Fc2	1990	Normal	Normal
		1991	Hot	Wet
		1992	Cold	Normal
		1993	Normal	Dry
		1994	Normal	Dry
		1995	Normal	Wet
		1996	Normal	Normal
Inuvik A	Tw1, Ts1*a*, Ts1*b*, Fc1	1996	Normal	Normal
		1997	Normal	Normal

ferences in their thermal masses and in their heating and cooling characteristics. A similar disparity between shallow and deep lakes was identified in Eugster et al. (2000), who compiled largely short-term data from circumpolar arctic and subarctic locations.

An evaporative continuum, from high to low, associated with terrain type is evident across the western and central Canadian subarctic. This trend is observed in Fig. 4, which plots the range of mean Q_E/Q^*, β, and α values experienced at all sites within each terrain type classification (each data year is considered individually). The two lake sites are plotted as separate terrain types due to their differing characteristics. Decreasing Q_E/Q^*, increasing β, and decreasing α are clearly associated with increasing surface dryness. There is, however, a substantial overlap in the range bars among terrain types. This is attributed to interannual variability at each site resulting from changing meteorological and soil moisture conditions (see below).

This evaporative continuum is also evident in the literature values presented in Table 3. A further example of this is a study by McFadden et al. (1998) conducted in the Alaskan tundra. In this study, the energy partitioning and β from a range of terrain types were compared over a 5-day period in 1994. A twofold difference in Q_E/Q^* was observed between the wettest and driest terrain types ($Q_E/Q^* = 0.42$ at a wetland as compared with 0.21 at a dry heath). Significant variation was also observed in β. The authors attributed these variations to changes in topography and surface hydrology among sites.

To further examine differences among terrain types in the western and central Canadian subarctic, the mean summertime surface saturation vapor pressure deficit (D_o) was calculated at each terrestrial research site using all available data years (Table 2). Deficit D_o represents how close the surface atmospheric boundary layer is to saturation. A surface with a small D_o is more saturated than a surface with a large D_o. A completely saturated surface would have a D_o of zero. This condition is assumed to occur at both lake sites, where surface water is not limited.

Mean summertime surface saturation deficits range from 0.28 (Tw1) to 0.94 kPa (Td2). There is no well-defined relationship between D_o and terrain type. Furthermore, the relationship between D_o and Q_E/Q^* is not as strong as the relationship between β or α and Q_E/Q^*. This is because the principal evaporating surface represented by D_o is at ground level; therefore, the zero plane displacement within a vegetated canopy is not accurately represented. D_o can be used, however, to identify ''wet'' (high evapotranspiration) and ''dry'' (low evapotranspiration) sites. Wet sites (wetland and shrub tundra) have a small D_o, while dry sites (upland tundra and forest) have a large D_o.

Because of limited data availability, it was not possible to ensure that the energy balance data used for comparison in the present study were collected under similar meteorological conditions. Hence, it is possible that differences observed among terrain types are due to variations in the meteorological conditions among individual years rather than to changes in terrain. To evaluate whether the meteorological conditions in a given data year differ from ''normal,'' mean summertime temperature and total summertime precipitation values were obtained for each data year from the meteorological station nearest to each site and were compared with the summertime 30-yr climate normals (1961–90) at the same station (Table 4). It should be noted that the meteorologic conditions experienced over Great Slave Lake (L_Dp) are generally significantly different from those recorded at the nearby Yellowknife A station.

Although ''normal'' temperature conditions are experienced in $>$ 72% of the data years (Table 4), there is significant variability in total summertime precipitation among years, making meteorological variation among data years a substantial concern. We believe, however, that although individual data years are affected by meteorological conditions, the overall differences among terrain types are large and consistent enough to be definitive. This is supported by the work from McFadden et al. (1998) discussed earlier and by Boudreau and Rouse (1995). The latter study details significant differences in the energy balance among a shallow lake, a wetland, an upland tundra, and an open canopy forest in a small subarctic basin during a single growing season.

6. Summary and conclusions

This study shows that the surface energy balance of the western and central Canadian subarctic is more strongly influenced by its heterogeneous terrain than by climatic differences resulting from geographical position. Five distinct terrain types are identified in this region (lakes, wetlands, shrub tundra, upland tundra, and coniferous forests), each with unique energy balance characteristics. Terrestrial sites belonging to the same terrain type have similar energy balance partitioning, Bowen ratios, and Priestley–Taylor alphas. The energy balance characteristics of subarctic lakes, however, depend on their physical properties. Shallow and deep lakes have very distinct energy balances due to their differing depths and thermal masses.

An evaporative continuum, from high to low, is observed among subarctic terrain types. Shallow lakes and wetlands have the highest Q_E/Q^*, while upland tundra and coniferous forests have the lowest Q_E/Q^*. It should be noted that sites with high Q_E/Q^* also have a low β and a high α, and vice versa. These relationships are expected because of the algebraic relationships among these variables shown in Eqs. (1), (3), (4) and (5). Surface saturation vapor pressure deficit is useful for characterizing the evapotranspiration conditions at terrestrial sites. In general, sites with a high Q_E/Q^* have a small D_o, and sites with a low Q_E/Q^* have a large D_o.

Acknowledgments. The authors would like to thank Dale Boudreau, Bob Kochtubajda, Natasha Neumann, Richard Petrone, and Mark Russell for their assistance in helping prepare and assemble the large dataset required for this project. Maps were prepared by Ric Hamilton. Financial support was provided by a Natural Science and Engineering Research Council of Canada (NSERC) Collaborative Partnership Research Grant to the Mackenzie GEWEX Study, NSERC research grants to P. M. Lafleur and W. R. Rouse, and Northern Student Training Grants awarded by the Department of Indian and Northern Affairs. Substantial logistical support during data collection has been provided by Aurora College, Inuvik, Northwest Territories; The Churchill Northern Studies Centre, Churchill, Manitoba; Environment Canada, Yellowknife, Northwest Territories; and the Polar Continental Shelf Project, Ottawa, Ontario.

REFERENCES

Bello, R. L., and J. D. Smith, 1990: The effect of weather variability on the energy balance of a lake in the Hudson Bay Lowlands, Canada. *Arct. Alp. Res.,* **22,** 98–107.

Betts, A. K., M. Goulden, and S. Wofsy, 1999: Controls on evaporation in a boreal spruce forest. *J. Climate,* **12,** 1601–1618.

Blanford, J. H., and L. W. Gay, 1992: Test of a robust eddy correlation system for sensible heat flux. *Theor. Appl. Climatol.,* **40,** 53–60.

Blanken, P. D., and W. R. Rouse, 1994: The role of birch-willow forest in the energy balance at treeline. *Arct. Alp. Res.,* **26,** 403–411.

——, and ——, 1995: Modelling evaporation from a high subarctic willow-birch forest. *Int. J. Climatol.,* **15,** 97–106.

——, and Coauthors, 2000: Eddy covariance measurements of evaporation from Great Slave Lake, Northwest Territories, Canada. *Water Resour. Res.,* **36,** 1069–1077.

Bolton, D., 1980: The computation of equivalent potential temperature. *Mon. Wea. Rev.,* **108,** 1046–1053.

Boudreau, L. D., 1993: The energy and water balance of a high subarctic wetland underlain by permafrost. M.S. thesis, Department of Geography, McMaster University, 205 pp.

——, and W. R. Rouse, 1995: The role of individual terrain units in the water balance of wetland tundra. *Climate Res.,* **5,** 31–47.

Bowen, I. S., 1926: The ratio of heat losses by conduction and evaporation from any water surface. *Phys. Rev.,* **27,** 779–787.

Brutsaert, W., 1982: *Evaporation into the Atmosphere.* D. Reidel, 299 pp.

Burton, K. L., W. R. Rouse, and L. D. Boudreau, 1996: Factors affecting the summer carbon dioxide budget of subarctic wetland tundra. *Climate Res.,* **6,** 203–213.

Eaton A. K., 1999: Energy balance and climate change in the western and central Canadian subarctic. M.S. thesis, School of Geography and Geology, McMaster University, 84 pp.

Ecoregions Working Group, 1989: *Ecoclimatic Regions of Canada, First Approximations.* Minister of Supply and Services Canada, 118 pp.

Eugster, W., J. P. McFadden, and F. S. Chapin III, 1997: A comparative approach to regional variation in surface fluxes using mobile eddy correlation towers. *Bound.-Layer Meteor.,* **85,** 293–307.

——, and Coauthors, 2000: Land–atmosphere energy exchange in Arctic tundra and boreal forest: Available data and feedbacks to climate. *Global Change Biol.,* **6** (Suppl. I), 85–115.

Fitzjarrald, D. R., and K. E. Moore, 1992: Turbulent transport over tundra. *J. Geophys. Res.,* **97** (D15), 16 717–16 729.

——, and ——, 1994: Growing season boundary layer climate and surface exchanges in a subarctic lichen woodland. *J. Geophys. Res.,* **99** (D1), 1899–1917.

Halliwell, D. H., and W. R. Rouse, 1987: Soil heat flux in permafrost: Characteristics and accuracy of measurement. *J. Climatol.,* **7,** 571–584.

——, and ——, 1989: A comparison of sensible and latent heat flux calculations using the Bowen ratio and aerodynamic methods. *J. Atmos. Oceanic Technol.,* **6,** 563–574.

Harazono, Y., M. Yoshimoto, M. Mano, G. L. Vourlitis, and W. C. Oechel, 1998: Characteristics of energy and water budgets over wet sedge and tussock tundra ecosystems at North Slope in Alaska. *Hydrol. Processes,* **12,** 2163–2183.

Joiner, D. W., J. H. McCaughey, P. M. Lafleur, and P. A. Bartlett, 1999: Water and carbon dioxide exchange at a boreal young jack pine forest in the BOREAS northern study area. *J. Geophys. Res.,* **104** (D22), 27 641–27 652.

Lafleur, P. M., 1992: Energy balance and evapotranspiration from a subarctic forest. *Agric. For. Meteor.,* **58,** 163–175.

——, 1994: Annual variability in summer evapotranspiration and water balance at a subarctic forest site. *Nord. Hydrol.,* **25,** 331–344.

——, and W. R. Rouse, 1988: The influence of surface cover and climate on energy partitioning and evaporation in a subarctic wetland. *Bound.-Layer Meteor.,* **44,** 327–347.

——, and C. P. Schreader, 1994: Water loss from the floor of a subarctic forest. *Arct. Alp. Res.,* **26,** 152–158.

——, and W. R. Rouse, 1995: Energy partitioning at treeline forest and tundra sites and its sensitivity to climate change. *Atmos.–Ocean,* **33,** 121–133.

——, ——, and D. W. Carlson, 1992: Energy balance differences and hydrologic impacts across the northern treeline. *Int. J. Climatol.,* **12,** 193–203.

Lynch, A. H., F. S. Chapin III, L. D. Hinzman, W. Wu, E. Lilly, G. Vourlitis, and E. Kim, 1999: Surface energy balance on the arctic tundra: Measurements and models. *J. Climate,* **12,** 2585–2606.

Marsh, P., and S. C. Bigras, 1988: Evaporation from Mackenzie Delta lakes, N. W. T., Canada. *Arct. Alp. Res.,* **20,** 220–229.

——, and J. W. Pomeroy, 1996: Meltwater fluxes at an Arctic forest–tundra site. *Hydrol. Processes,* **10,** 1383–1400.

McCaughey, J. H., P. M. Lafleur, D. W. Joiner, P. A. Bartlett, A. M. Costello, D. E. Jelinski, and M. G. Ryan, 1997: Magnitudes and seasonal patterns of energy, water, and carbon exchanges at a boreal young jack pine forest in the BOREAS northern study area. *J. Geophys. Res.,* **102** (D4), 28 997–29 007.

McFadden, J. P., F. S. Chapin III, and D. Y. Hollinger, 1998: Subgrid-scale variability in the surface energy balance of arctic tundra. *J. Geophys. Res.,* **103** (D2), 28 947–28 961.

McNeil, D. D., and W. J. Shuttleworth, 1975: Comparative measurements of the energy fluxes over a pine forest. *Bound.-Layer Meteor.,* **9,** 297–313.

Moore, C. J., 1976: Eddy flux measurements above a pine forest. *Quart. J. Roy. Meteor. Soc.,* **102,** 913–918.

——, 1986: Frequency response corrections for eddy correlation systems. *Bound.-Layer Meteor.,* **37,** 17–35.

Moore, K. E., D. R. Fitzjarrald, S. C. Wofsy, B. C. Daube, J. W. Munger, P. S. Bakwin, and P. Crill, 1994: A season of heat, water vapor, total hydrocarbon, and ozone fluxes at a subarctic fen. *J. Geophys. Res.,* **99** (D1), 1937–1952.

——, ——, R. K. Sakai, and J. M. Freedman, 2000: Growing season water balance at a boreal jack pine forest. *Water Resour. Res.,* **36,** 483–493.

Oke, T. R., 1992: *Boundary Layer Climates.* 2d ed. Routledge, 435 pp.

Petrone, R. M., W. R. Rouse, and P. Marsh, 2000: Comparative surface energy budgets in western and central subarctic regions of Canada. *Int. J. Climatol.,* **20,** 1131–1148.

Price, J. S., D. A. Maloney, and F. G. Downey, 1991: Peatlands of the Lake Melville coastal plain, Labrador. *Northern Hydrology Selected Perspectives: Proc. of the Northern Hydrology Symp.,* Saskatoon, SK, Canada, National Hydrology Research Institute, 293–302.

Priestley, C. H. B., and R. J. Taylor, 1972: On the assessment of surface heat flux and evaporation using large-scale parameters. *Mon. Wea. Rev.,* **100,** 81–92.

Rawson, D. S., 1950: The physical limnology of Great Slave Lake. *J. Fish. Res. Board Can.,* **8,** 3–66.

Rouse, W. R., 1984a: Microclimate at Arctic tree line 2. Soil microclimate of tundra and forest. *Water Resour. Res.,* **20,** 67–73.

——, 1984b: Microclimate at Arctic tree line 3. The effects of regional advection on the surface energy balance of upland tundra. *Water Resour. Res.,* **20,** 74–78.

——, 2000: Progress in hydrological research in the Mackenzie GEWEX study. *Hydrol. Processes,* **14,** 1667–1685.

——, and R. L. Bello, 1985: Impact of Hudson Bay on the energy balance in the Hudson Bay Lowlands and the potential for climate modification. *Atmos.–Ocean,* **23,** 375–392.

——, P. F. Mills, and R. B. Stewart, 1977: Evaporation in high latitudes. *Water Resour. Res.,* **13,** 909–914.

——, S. G. Hardill, and P. M. Lafleur, 1987: The energy balance in the coastal environment of James Bay and Hudson Bay during the growing season. *J. Climatol.,* **7,** 165–179.

——, R. L. Bello, and P. M. Lafleur, 1997: The low arctic and subarctic. *The Surface Climates of Canada,* W. G. Bailey, T. R. Oke, and W. R. Rouse, Eds., McGill–Queen's University Press, 198–221.

Schreader, C. P., W. R. Rouse, T. J. Griffis, L. D. Boudreau, and P. D. Blanken, 1998: Carbon dioxide fluxes in a northern fen during a hot, dry summer. *Global Biogeochem. Cycles,* **12,** 729–740.

Sellers, W. D., 1965: *Physical Climatology.* University of Chicago Press, 272 pp.

Spittlehouse, D. L., and T. A. Black, 1979: Determination of forest evapotranspiration using Bowen ratio and eddy-correlation measurements. *J. Appl. Meteor.,* **18,** 647–653.

Stewart, R. B., and W. R. Rouse, 1976: A simple method for determining the evaporation for shallow lakes and ponds. *Water Resour. Res.,* **12,** 623–628.

Stewart, R. E., and Coauthors, 1998: The Mackenzie GEWEX study: The water and energy cycles of a major North American river basin. *Bull. Amer. Meteor. Soc.,* **79,** 2665–2683.

Thom, A. S., 1975: Momentum, mass and heat exchange of plant communities. *Vegetation and the Atmosphere.* Vol. 1, *Principles,* J. L. Monteith, Ed., Academic Press, 57–109.

Tsvang, L. R., B. M. Koprov, S. L. Zubkovskii, A. J. Dyer, B. Hicks, M. Miyake, R. W. Stewart, and J. W. McDonald, 1973: A comparison of turbulence measurements by different instruments: Tsimlyansk field experiment 1970. *Bound.-Layer Meteor.,* **3,** 499–521.

Weick, E. J., and W. R. Rouse, 1991: Advection in the coastal Hudson Bay Lowlands, Canada. I. The terrestrial surface energy balance. *Arct. Alp. Res.,* **23,** 328–337.