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Nonlinear controls on evapotranspiration in arctic coastal wetlands

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Abstract. Projected increases in air temperature and precipitation due to climate change in Arctic wetlands could dramatically affect ecosystem function. As a consequence, it is important to define controls on evapotranspiration, the major pathway of water loss from these systems. We quantified the multi-year controls on midday Arctic coastal wetland evapotranspiration, measured with the eddy covariance method at two vegetated, drained thaw lake basins near Barrow, Alaska. Variations in near-surface soil moisture and atmospheric vapor pressure deficits were found to have nonlinear effects on midday evapotranspiration rates. Vapor pressure deficits (VPD) near 0.3 kPa appeared to be an important hydrological threshold, allowing latent heat flux to persistently exceed sensible heat flux. Dry (compared to wet) soils increased bulk surface resistance (water-limited). Wet soils favored ground heat flux and therefore limited the energy available to sensible and latent heat flux (energy-limited). Thus, midday evapotranspiration was suppressed from both dry and wet soils but through different mechanisms. We also found that wet soils (ponding excluded) combined with large VPD, resulted in an increased bulk surface resistance and therefore suppressing evapotranspiration below its potential rate (Priestley-Taylor $\alpha < 1.26$). This was likely caused by the limited ability of mosses to transfer moisture during large atmospheric demands. Ultimately, in addition to net radiation, the various controlling factors on midday evapotranspiration (i.e., near-surface soil moisture, atmospheric vapor pressure, and the limited ability of saturated mosses to transfer water during high VPD) resulted in an average evapotranspiration rate of up to 75 % of the potential evapotranspiration rate. These multiple limitations on midday evapotranspiration rates have the potential to moderate interannual variation of total evapotranspiration and reduce excessive water loss in a warmer climate. Combined with the prevailing maritime winds and projected increases in precipitation, these curbing mechanisms will likely prevent extensive future soil drying and hence maintain the presence of coastal wetlands.

1 Introduction

The response of Arctic wetland hydrology to projected climate warming is uncertain. Evapotranspiration is the least understood component in the hydrologic cycle (Kane et al., 1989, 1992; Vörösmarty et al., 2001; Woo et al., 2008), though regional studies have proposed both increased (Lafleur, 1993) and unchanged (Rouse et al., 1992) future evapotranspiration rates in Arctic coastal wetlands. Evapotranspiration is the major pathway for water loss from the flat tundra landscape (Rovansek et al., 1996; Mendez et al., 1998; Bowling et al., 2003). A region that has for a long time sequestered carbon will shift to a carbon source if soil drying occurs (Oechel et al., 1998; Olivas et al., 2010) and thus, causing a positive feedback to global climate warming.

A vast majority of extremely low-gradient tundra is located within 135 km of the Arctic Ocean (Walker et al., 2005; Minke et al., 2007). The summer climate of the coastal zone is controlled by a steady mesoscale phenomenon – a nearly

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24-h sea breeze (Moritz, 1977; Walsh, 1977; Kozo, 1979, 1982), resulting in low diurnal temperature fluctuations and low atmospheric vapor pressure deficits (VPD). All components of the coastal wetland energy balance, except net radiation, depend on wind direction, with cold, moist maritime air suppressing evapotranspiration losses (Rouse et al., 1987). One may expect the sea breeze to continue in a warmer climate, though the fate of future evapotranspiration rates from coastal wetlands is uncertain.

Measurement of energy fluxes in arctic environments is challenging due to climatic and logistical constraints. Hence, most field studies are of relatively short duration. There are several field studies of arctic surface energy exchange (see Eugster et al., 2000) and water balance (see Kane and Yang, 2004), but few studies have conducted multi-year analyses of evapotranspiration measured using the eddy covariance technique.

Here we present results from one of the longest time series of flux measurements available for any arctic ecosystem, represented by two vegetated, drained thaw lake basins (five summers at one site; three summers in an adjacent site) on the Arctic Coastal Plain of Alaska. Our objective is to define mechanisms controlling midday evapotranspiration rates from seasonally inundated arctic coastal wetlands. We hypothesize that the evapotranspiration experiences multiple controls apart from surface net radiation. Defining these controls is important in order to refine our understanding of the hydrologic regime and the role of tundra ecosystems in global climate feedbacks.

2 Background

Extremely low-relief wetlands represent a significant portion (>400 000 km^2) of the pan-Arctic landscape (Walker et al., 2005) and are unique in that they exist in an environment with a desert-like annual precipitation (∼250 mm yr^{-1}). Sparse summer runoff (Brown et al., 1968; Kane et al., 2008) limits the summer net water balance to summer precipitation minus evapotranspiration. A negative summer net water balance is common (Mendez et al., 1998), but this is offset by an annual replenishment of water from snowmelt (Rovansek et al., 1996). The abundance of snowmelt water results in extensive surface inundation during the first week following snowmelt (Bowling et al., 2003; Woo et al., 2006). Spring runoff is not generated until surface stores are replenished (Rovansek et al., 1996; Bowling and Lettenmaier, 2010). Accordingly, evapotranspiration is the major pathway for water loss in summer and also affects the lateral exports of water.

Evapotranspiration from wet and moist tundra ecosystems of the North Slope of Alaska is estimated to be 0.8–4.2 mm day^{-1}, resulting in estimated annual totals ranging from 70 to 190 mm (see summary by Vourlitis and Oechel, 1997). A majority (55 to 85 %) of this evapotranspiration is represented by evaporation from moss and open water (see review by Engstrom et al., 2006), even though bryophytes receive only 10–20 % of direct solar radiation during a clear summer day (Miller and Tieszen, 1972). Upward migration of water in *Sphagnum* moss, attributed to capillary flow, has been shown to occur from water tables located at 0.2 m depth (Hayward and Clymo, 1982; Price et al., 2009). Capillary water flow in moss, and hence moss evaporation, is negligible at water potentials below −0.1 MPa (Hayward and Clymo, 1982). In comparison, stomatal closure due to water stress by the typical vascular plants occurs at soil water potentials of −0.4 MPa (*Arctophila fulva*) to −1.2 MPa (*Carex aquatilis*) (Stoner and Miller, 1975; Johnson and Caldwell, 1975). Total transpiration is closely related to Leaf Area Index (LAI), as stomatal closure is rare at wet, coastal Arctic sites (Miller and Tieszen, 1972). However, plant-scale studies have also shown that the conductance of tundra plants can be reduced by leaf cell water stress, induced by vapor pressure gradients between the leaf and the ambient air that range from 0.7 to 2 kPa (Johnson and Caldwell, 1975). Tundra bryophytes are extremely sensitive to air vapor pressure deficits, due to the direct changes in tissue water content (Oechel and Sveinbjörnsson, 1978).

The effect of maritime air mass on surface energy partitioning has been observed in tundra 135 km inland from the Arctic coast (Harazono et al., 1998). Cold, moisture-laden air along the coast boosts the partitioning of surface energy into sensible heat flux (H), caused by (a) a steep temperature gradient between the ground surface and air, favoring H; and (b) a nearly saturated air mass that reduces latent heat flux (*LE*) (Rouse et al., 1987; Lafleur and Rouse, 1988, 1995; Price, 1991; Harazono et al., 1998; Boike et al., 2008). This dynamic explains, at least partly, the observation that despite wet soils, evapotranspiration is generally below its potential rate in coastal arctic wetlands (Rouse et al., 1987; Mendez et al., 1998). However, it is unclear what values in vapor pressure deficits lead to significant changes in surface energy partitioning and evapotranspiration rates.

Soil moisture may also play a major role in tundra surface energy balance partitioning. Surface energy partitioning shifted from being dominated by latent heat in the early season, when the water table was near the ground surface, to being dominated by sensible heat in late summer, when the water table was at 30 cm depth at a coastal wet and moist herbaceous tundra site (Vourlitis and Oechel, 1997). Further, wet organic soils transfer heat more efficiently than dry organic soils (Farouki, 1981; Hinzman et al., 1991), which, in theory, would leave less net radiation available to sensible and latent heat fluxes. This prompts questions about whether the Arctic wetlands display important controlling mechanisms on the local hydrological system, constraining evapotranspiration rates not only when near-surface conditions are dry, but also when they are wet.

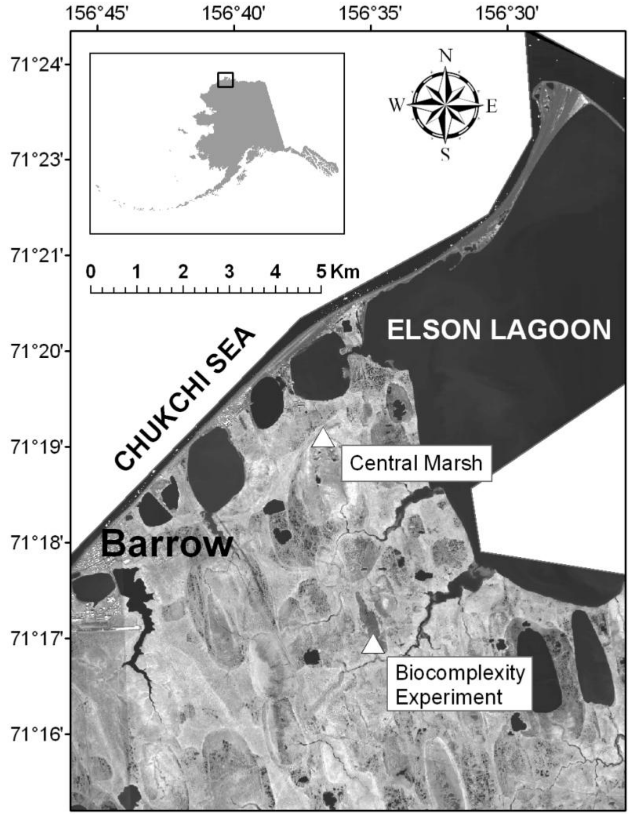

Fig. 1. The Central Marsh (CM) and the Biocomplexity Experiment (BE) sites are located at separate vegetated, drained thaw lake basins within 3 km of the ocean, outside the town of Barrow, Northern Alaska.

3 Site description

The two sites, hereafter referred to as Central Marsh (CM, at 71°19′12.5″ N, 156°37′20.211″ W, elevation 1 m) and the Biocomplexity Experiment (BE, at 71°16′51.17″ N, 156°35’47.28″ W, elevation 4.5 m), are located 4.5 km apart, and both are only a few kilometers from the ocean near Barrow, Alaska, on the Arctic Coastal Plain (Fig. 1). Mean annual air temperature at Barrow Airport is −12 °C (1977–2009), with a summer (June through August) average of 3.3 °C. A large amount of the annual adjusted precipitation (173 mm, 1977–2009) falls during June through August (72 mm). Fog and drizzle are common during the summer because the area receives a steady cool, moist wind (mean 5 m s^{-1}) off the ocean from the east-northeast (Shulski and Wendler, 2007). The BE site is located in the control treatment of a large-scale hydrologic manipulation experiment that began in 2007 (identified as the South site in the work by Zona et al., 2009). Unlike the other treatments, this site did not have manipulated water tables.

The BE and CM sites are representative of vegetated, drained thaw lake basins that appear to have drained between 50 and 300 yr ago (Hinkel et al., 2003). The sites are poorly drained and are characterized by wet meadow tundra with Typic Aquiturbels soils (Bockheim et al., 1999) underlain by 600 m thick permafrost (Brown and Johnson, 1965). Low-centered ice-wedge polygons are found at the vegetated, drained thaw lake basin while high-centered ice-wedge polygons cover the upland areas of the watershed. Vegetated, drained thaw lake basins (Mackay, 1963) occupy approximately 26 % of the Arctic Coastal Plain (Hinkel et al., 2005) and 50 % of the Barrow Peninsula north of ∼71° latitude (Hinkel et al., 2003). Longer-term (>2 yr) energy balance measurements of vegetated, drained thaw lakes are limited, constraining our understanding of interannual controls of evapotranspiration rates from this vast region.

Non-vascular vegetation contributes significantly to biomass and land cover (Webber, 1974, 1978; Oechel and Sveinbjörnsson, 1978; Rastorfer, 1978). Bryophytes represent between 60 and 95 % of the overall live biomass in similar wet meadow communities (Tieszen, 1978), with much of the variation due to small-scale heterogeneity associated with micro-topography (Tieszen, 1978; Hollister and Flaherty, 2010). Across the BE drained lake bed, mosses represent most of the live above-ground biomass (Zona et al., 2009, 2010, 2011). Up to 60 % of the ecosystem’s net daytime CO_2 uptake at the end of the growing season at BE is represented by *Sphagnum* (Zona et al., 2011). Accordingly, controls on evapotranspiration rates from this landscape are likely dominated by moss evaporation processes.

The sites differ somewhat in vascular plant composition, LAI (green biomass, unless otherwise stated), and the amount of standing dead biomass (which is defined as attached or upright dead plant matter). *Arctophila fulva* is the dominant vascular plant species at the CM site, where vegetation is also represented by sedges, mosses, and lichens. LAI at the CM site reached 1.4 in mid-August of 2001 (Mano, 2003). Mid-August LAI reached 0.58 in 2006 at the BE vegetated, drained lake, where the vascular plant coverage is dominated by *Carex aquatilis* (Zona et al., 2011). Sedges at the BE site did not experience water stress in mid-July of 2008 (P. Olivas, unpublished data). Standing dead leaf biomass in the Barrow area reaches 1.23 m^2 m^{-2} (Dennis et al., 1978). The CM site has a larger abundance of standing dead biomass than the BE site (personal observation). End of growing season plant senescence extends from the end of August to late September (Myers and Pitelka, 1979).

The sparseness of live subsurface material at depths greater than 25 cm at Barrow suggests that the cold temperatures near the bottom of the active layer limit vascular root growth (Dennis and Johnson, 1970). Moss may reach a thickness of 20 cm at wet sites, but the bulk of their living biomass is usually within ∼1 cm of the soil surface (Engstrom et al., 2005). The rate of thaw is at its highest in early summer and the maximum thaw depth (active layer depth) is reached in late August/September. The active layer depth at a nearby drained lake basin varied from 19 to 62 cm (36 cm mean) from 1995–2009, while the mean active layer depth at the BE site was 30 cm in 2006 and 26 cm in 2007 and 2008 (Shiklomanov et al., 2010).

4 Methods

The controls on midday evapotranspiration rates were assessed through surface energy balance partitioning, the McNaughton and Jarvis Ω-factor, and by solving for parameters in the Penman-Monteith and the Priestley-Taylor equations. The results were then analyzed in the context of soil moisture and meteorological conditions.

4.1 Measurements

We collected summer (June through August) measurements for five years (1999–2003) at the CM site and for three years (2006–2008) at the BE site. Energy flux measurements were taken at a 10 Hz sampling interval using an eddy covariance system. The path length of our anemometer and gas analyzer sensor at CM was 10 cm, and the separation distance between the center of the sonic anemometer and the open-path IRGA sensors was 16 cm. The three components of wind speed, air temperature, and water vapor concentration from these sensors were recorded on a magneto-optical disc by a digital recorder (Teac, DRM3). At BE, the sensor separation of the Li-7500 and WindMasterPro was 10 cm. The Li-7500 was calibrated every two to four weeks, using ultra high purity nitrogen as zero and a dew point generator (Li-610, Li-COR) that produced an air stream with a known water vapor dew point. Micrometeorological variables were sampled on a data logger every 5 s (CM) or 10 s (BE), and then averaged every 30 min. Additional descriptions of the measurements and data analysis are presented in the work of Harazono et al. (2003) and Zona et al. (2009).

Measurements of volumetric water content (VWC) at two locations within the CM drained lake basin were taken in 2000–2003 by inserting a 7 cm Vitel probe (Hydra soil moisture probe, Vitel Inc.) vertically into the ground. The instrument was calibrated through comparison to multiple oven-dried soil samples (Engstrom et al., 2005). The CM site was often inundated in early summer. Such events are presented here as 100 % VWC to indicate ponding.

Hourly atmospheric air pressure for the years 1999–2003 were obtained from the NCDC web archive (http://cdo.ncdc.noaa.gov/cgi-bin/cdo/cdostnsearch.pl) and used in the calculations of the psychrometric constant. Long-term records of daily precipitation and air temperature were retrieved from the National Climatic Data Center (NCDC) web archive for the Barrow Airport (STN 700260, WBAN 27502, http://www.ncdc.noaa.gov/cgi-bin/res40.pl?page=gsod.html). The characteristic increase in net radiation (R_n) defined the start date of the summer. We adjusted the daily precipitation to account for gage undercatch, according to Yang et al. (1998). Snow accumulation was retrieved from the Circumpolar Active Layer Monitoring program (http://www.udel.edu/Geography/calm/).

4.2 Eddy covariance calculations

We calculated flux of heat and momentum at 30 min. intervals, according to typical covariance calculation procedures. The following corrections were applied (Harazono et al., 2003; Zona et al., 2009): humidity effect on the sonic thermometry (Kaimal and Gaynor, 1991); effects of path length and sensor separation on the spectrum for high-frequency flux ranges (Moore, 1986); air density effects (Webb et al., 1980; Leuning et al., 1982); and coordinate rotation (Tanner and Thurtell, 1969). We removed calculated flux during rain, fog, and low wind, which may have caused a bias (i.e., reduced representation of low evapotranspiration rates). Extreme amplitudes in the flux data (greater than three times the average) were removed. At the BE site, fluxes of latent heat, sensible heat and momentum were calculated using the EdiRe program and software (version 1.4.3.1169, Robert Clement, University of Edinburgh). No gap-filling was performed when analyzing the bulk parameters and energy flux ratios. Midday represented half-hourly values around solar noon (defined as ±2 h from local solar noon, ∼14:00 Alaska Standard Time). Extreme (greater than six times the average) amplitudes in bulk parameters were removed. All analyses, except the total daily evapotranspiration, represent non-gap-filled midday values. For daily evapotranspiration rates, gap filling was performed for missing data (<3.5 consecutive hours), using linear interpolation. The ending date of the study periods was 31 August and the start date in each year was, at the earliest, the first day after the snowmelt completion, although the effective dates depended upon the available data.

4.3 Soil moisture analysis

Unfrozen soil moisture, in percent saturation, was estimated from volumetric water content measurements at 10 cm depth at the BE site. The spring peak in soil moisture was assumed to represent saturated conditions (100 %, all micro and macro pore spaces filled with liquid water). In winter, the organic soil was assumed to have 6 % saturation (Hinzman et al., 1991). Soil water potential (ψ) was calculated by fitting a curve, after van Genuchten (1980), to a measured water potential sequence (WP4-T, Decagon Devices) from a surface organic moss layer sampled at the BE site.

4.4 Analysis of resistances and equilibrium evaporation

The Penman-Monteith equation (Monteith, 1973) is expressed in terms of aerodynamic (r_a) and bulk surface resistance (r_c):

$$LE = \frac{sQ_\mathrm{a} + \frac{\rho C_p[e_\mathrm{s}(T_\mathrm{a})-e_\mathrm{a}]}{r_\mathrm{a}}}{s+\gamma[1+r_\mathrm{c}/r_\mathrm{a}]} \quad (1)$$

where s is the slope of the saturation vapor pressure curve versus temperature modified from Brutsaert (1982); Q_a is available energy ($\mathrm{W\,m^{-2}}$); γ is the psychrometric constant ($\mathrm{kPa\,K^{-1}}$); ρ the air density ($\mathrm{kg\,m^{-3}}$); C_p is the specific heat capacity of air ($\mathrm{kJ\,kg^{-1}\,K^{-1}}$) at constant pressure; e_s is the saturation vapor pressure (kPa) at T_a, which is the ambient air temperature (K); and e_a is the air vapor pressure (kPa). Shallow ponded water can represent a significant portion (<50 %) of the net radiation partitioning (Harazono et al., 1998). Therefore, we defined Q_a as the sum of sensible (H) and latent heat (LE) fluxes, since no water temperature measurements were obtained. The aerodynamic resistance, r_a ($\mathrm{s\,m^{-1}}$), is calculated from Eq. (2) following Monteith (1973), with an additional term on the right side representing the laminar boundary layer resistance from Thom (1975) and Lafleur and Rouse (1988):

$$r_\mathrm{a} = \frac{u}{u^{*2}} + \frac{4}{u^*} \quad (2)$$

where u^* is friction velocity ($\mathrm{m\,s^{-1}}$) obtained by eddy covariance measurements and u is wind speed. From here forward, the sum of the aerodynamic and laminar boundary layer resistance in Eq. (2) is referred to as aerodynamic resistance (r_a). The aerodynamic resistance is the bulk meteorological descriptor of the role of atmospheric turbulence in evaporation.

Isothermal resistance, r_i, ($\mathrm{m\,s^{-1}}$) was originally defined by Monteith (1965) and is sometimes referred to as the climatological resistance. It is the ratio of the water vapor deficit to available energy at the canopy:

$$r_\mathrm{i} = \frac{\rho \cdot C_p}{\gamma}\frac{[e_\mathrm{s}-e_\mathrm{a}]}{Q_\mathrm{a}} \qquad \text{(Stewart and Thom, 1973)} \quad (3)$$

Equations (1), (3), and the Bowen ratio, β, which is the ratio of sensible over latent heat, can be combined to solve for the bulk surface resistance, r_c ($\mathrm{m\,s^{-1}}$):

$$r_\mathrm{c} = (1+\beta)r_\mathrm{i} + \left(\beta\frac{s}{\gamma}-1\right)r_\mathrm{a} \quad (4)$$

The bulk surface resistance characterizes the control of water loss by vascular plants, non-vascular vegetation, and bare ground.

The bulk surface resistance approaches zero either because the surface boundary layer becomes saturated and VPD = 0 or the air travels over an unsaturated surface with constant r_c and the moisture deficit in the air becomes equal to the value of the surface. A r_c close to 0 results in the Penman-Monteith Equation (Monteith, 1973) collapsing into:

$$LE = \alpha\left(\frac{s}{s+\gamma}\right)Q_\mathrm{a} \qquad \text{(Priestley and Taylor, 1972)} \quad (5)$$

which is known as the Priestley-Taylor equation. The evapotranspiration is referred to as "equilibrium" when α equals one, which is most commonly achieved when VPD = 0 (note that equilibrium rates can also be measured over unsaturated surfaces and VPD > 0). This method assumes that latent heat flux depends only upon the absolute temperature and available energy. Results from a variety of arctic sites, both wet and dry, indicate that latent heat flux is often above the equilibrium rate (see Engstrom et al., 2002), as originally suggested by Priestley and Taylor (1972) at a non-water-limited grassland. Large-scale mixing of the planetary boundary layer and the entrainment of drier air from above the mixed layer results in evaporation over saturated surfaces greater than the "equilibrium" rate (McNaughton and Jarvis, 1983; DeBruin, 1983). DeBruin (1983) indicates that α is a function of wind speed, surface roughness, and bulk surface resistance. Here we defined the potential evapotranspiration by setting the α-value to 1.26 (Priestley and Taylor, 1972).

The McNaughton and Jarvis Ω-factor sets the relative importance of r_c and r_a:

$$\Omega = \left(1+\frac{s}{s+\gamma}\frac{r_\mathrm{c}}{r_\mathrm{a}}\right)^{-1} \quad (6)$$

A vigorous turbulent mixing of the air mass suppresses Ω by promoting increased VPD at the surface. Limited atmospheric mixing results in Ω approaching unity (McNaughton and Jarvis, 1983). However, Ω will approach 0 as long as $r_\mathrm{c} \gg r_\mathrm{a}$. In general, VPD is the main driver of evapotranspiration when Ω is low, while net radiation has the dominant control when Ω is near 1.

5 Results

5.1 Meteorological and hydrologic conditions

Analyzed measurements represented the thawed season through August (1999–2003 and 2006–2008). Mean air temperature (June–August, 3.2 °C) and precipitation (June–September 86 mm) were near the long-term means (3.4 °C and 99 mm, respectively, from 1979–2008), though large interannual variations occurred (Table 1). Summer 2007 brought unusually high air temperatures (5.4 °C) and low precipitation (24 mm). Most of the 2007 summer precipitation occurred in a single event in mid-August. During all study periods, 77 % of daily precipitation rates were less than 2 $\mathrm{mm\,day^{-1}}$. Trace observations (<0.13 mm) represented 33 % of all recorded events. Accumulated winter precipitation ranged from 93 to 158 mm of snow water equivalent (SWE).

The maritime nature of both sites led to low variability in VPD and air temperature. Mean daily VPD was 0.08 kPa, with a typical diurnal min and max of 0.02 and 0.17 kPa, respectively. Mean midday VPD was similar during all years (0.10–0.13 kPa) except during summer of 2007, when VPD was higher (0.17 kPa) (Table 1). The maximum VPD

Table 1. Meteorological conditions during the study period (1999–2008), including Snow Water Equivalent (SWE) prior to snowmelt; total precipitation from June through September; mean air temperature June through August; mean midday (12:00 through 16:00) air vapor pressure deficit (VPD); and the number of days that experienced a VPD above 0.3 kPa. Mean precipitation and air temperature, 1999–2008 (86 mm and 3.2 °C, respectively), were near the long-term (1979–2008) conditions of 99 mm and 3.4 °C, respectively. The end of the study period was 31 August (Julian day 242).

	1999	2000	2001	2002	2003	2006	2007	2008	Mean
SWE (mm)	122	113	123	93	95	137	98	158	117 ± 23
Precipitation, June–September (mm)	82	128	124	114	72	72	24	68	86 ± 35
Mean air temperature, June–August (°C)	4.2	3.1	2.1	2.3	2.5	2.9	5.4	3.3	3.2 ± 3.2
Mean midday VPD, June–August (kPa)	0.12	0.13	0.1	0.12	0.12	0.11	0.17	0.12	0.12 ± 0.02
VPD > 0.3 kPa, June–August (days)	12	10	14	9	8	10	13	10	11 ± 2
Start study period (Julian-Day)	163	163	164	156	158	162	162	169	162 ± 4

Table 2. Differences in midday (12:00 through 16:00) energy balance partitioning, bulk parameters, and air conditions during offshore (from land to ocean) and onshore (from ocean to land) winds at the CM site during the thawed season, 1999–2003. The temperature gradient (ΔT) represents the air minus the ground surface temperature (sensor located at 1 cm depth).

	Offshore	Onshore
LE/R_n	0.33 ± 0.17	0.28 ± 0.14
H/R_n	0.22 ± 0.12	0.35 ± 0.14
G/R_n	0.22 ± 0.08	0.15 ± 0.07
β	0.87 ± 0.82	1.37 ± 0.64
α	1.08 ± 0.24	0.95 ± 0.21
ΔT (°C)	2.3 ± 3.5	-1.3 ± 3.8
VPD (kPa)	0.21 ± 0.22	0.12 ± 0.13

recorded was 1.76 kPa, but days exceeding VPDs of 0.3 kPa were few (8 to 14 days per summer).

Onshore summer winds, defined as winds originating from between 1–135 and 225–360 degrees, occurred 89 % of the time (1999–2008). Air during onshore winds was colder than the ground surface (Table 2). Offshore winds (from land to sea) typically produced higher VPDs and air temperatures than onshore winds, reversing the typical midday temperature gradient between the air and the ground surface.

The moss surface and organic soils remained close to saturation throughout the study periods, with the exception of 2007 (Fig. 2). The soils within the vegetated, drained lake basins were unusually dry in late July 2007 (when the water table dropped below 15 cm depth). Snowmelt water recharged the drained lake soil water storage in spring. Water table measurements at the BE drained lake basin show a multi-week long ponding period following the snowmelt (note that the water table measurements do not capture the start of the inundation). About 10 cm of water accumulated above the ground surface following snowmelt. The drained lake basins also experienced inundation in late summer (2001 and 2008), resulting in an inundation for at least half of the summer's duration. No soil water measurements were made in summer 1999, but the near-normal precipitation (82 mm) suggests wet soil conditions.

5.2 Surface energy exchange

Energy balance closure was not complete (CM 80 % and BE 95 %, Table 3), though it was comparable to other tundra and grassland ecosystems reported by Eugster et al. (2000), Wilson et al. (2002), Cava et al. (2008), and Ryu et al. (2008). A major portion of the midday surface energy balance was partitioned into sensible heat flux (CM 35 % and BE 48 %), resulting in a mean midday Bowen ratio above unity at both sites (1.40 at CM and BE). Latent heat flux represented 29 % and 35 %, and ground heat flux averaged 16 % and 12 %, at CM and BE, respectively. A plot of the seasonal and interannual variations in midday energy partitioning across the thawed season suggests a somewhat consistent partitioning into *LE* (Fig. 3a, b) despite large interannual differences in soil moisture (Fig. 2).

Mean midday evapotranspiration (ET) rates showed large day-to-day variations, which were also found in total daily ET. Measured total evapotranspiration ranged from 0 to 4.7 mm day^{-1} (mean 1.5 mm day^{-1}), and 0.2 to 3.4 mm day^{-1} (mean 1.9 mm day^{-1}), at the CM (311 days) and BE (46 days) sites, respectively. The high McNaughton and Jarvis Ω-factor (CM 0.74 and BE 0.75) suggests that net radiation was the main control on evapotranspiration rates. Overall mean midday evapotranspiration was near or below the equilibrium rate (CM 0.94 and BE 0.88) (Table 3). Overall average evapotranspiration rates of 75 % (CM) and 70 % (BE) of the potential rate (α1.26) indicate additional controls than net radiation alone.

The effect of soil moisture on Priestley-Taylor α was gradual. Approximately 64 % ($P < 0.01$, probability that correlation is zero) of the variance in the Priestley-Taylor α was correlated to the near-surface soil moisture (Fig. 4a). Additional comparisons were possible by focusing on a time period (20 July through 12 August) that displayed unusually dry soils ($\Psi < -0.13$ MPa) in 2007 but that was wet (although not inundated) in the other years (Table 4). In

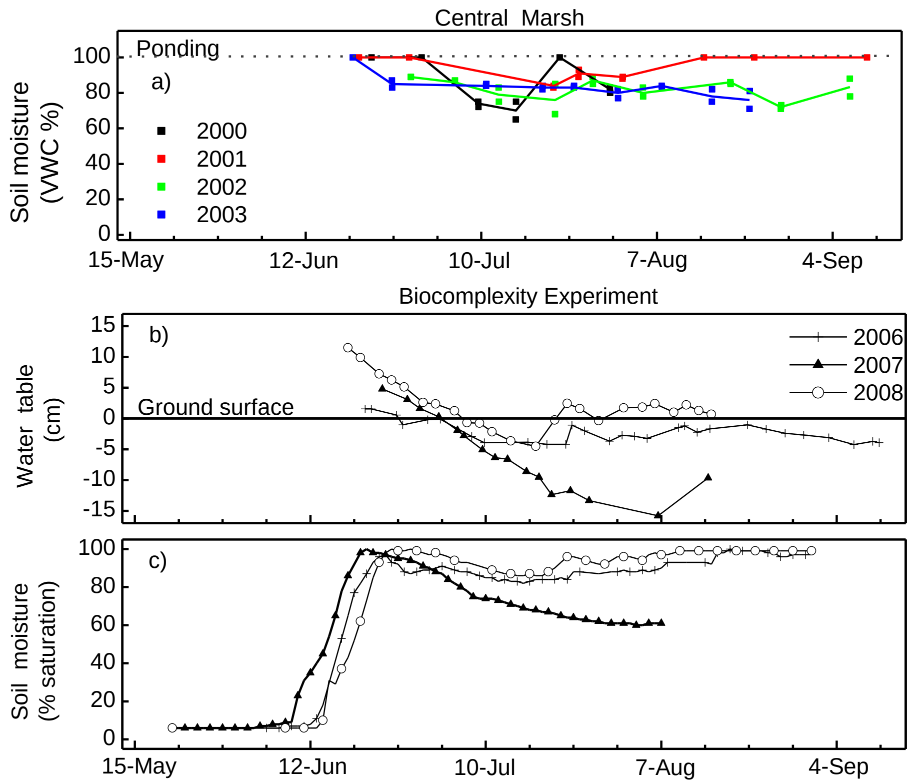

Fig. 2. The soil water status during the study period (no measurements from 1999) at the CM site **(a)** and the BE site (**b** and **c**). Panels **(a)** and **(b)** represent multiple locations across the vegetated, drained lake basins, while panel **(c)** is a continuous record of volumetric soil water content measurements at 10 cm depth near the BE eddy covariance tower, converted into % saturation.

Table 3. Mean midday (12:00 through 16:00) energy balance partitioning and bulk parameters at the CM (1999–2003) and BE (2006–2008) sites during the thawed season (through August).

	Central Marsh	Biocomplexity Experiment
LE/R_n	0.29 ± 0.15	0.35 ± 0.07
H/R_n	0.35 ± 0.14	0.48 ± 0.12
G/R_n	0.16 ± 0.07	0.12 ± 0.05
Closure	0.80 ± 0.23	0.95 ± 0.13
β	1.40 ± 0.67	1.40 ± 0.39
Ω	0.74 ± 0.22	0.75 ± 0.17
α	0.94 ± 0.22	0.88 ± 0.15
r_c (s m^{-1})	46 ± 40	46 ± 38
r_a (s m^{-1})	62 ± 38	63 ± 27
r_i (s m^{-1})	14 ± 14	12 ± 12

Closure = $(LE + H + G)/R_n$

addition to decreased Priestley-Taylor α-value, the dry soil conditions showed reduced Ω. Bulk surface resistance also responded to dry (r_c 57 s m^{-1}) and wet (41 s m^{-1}) soils, and displayed a statistically significant trend in summer 2003 and 2007. Accordingly, reduced soil moisture had a suppressing effect on ET.

Despite the differences in Priestley-Taylor α and r_c, late-summer partitioning of net radiation into LE was strikingly similar between dry and wet soils (37 and 34 %, respectively). This suggests controls additional to bulk surface resistance for ET. Wet soils increased the partitioning to ground heat flux from ~6 % (dry) to 14 % (saturated conditions) (Fig. 4b, Table 4), which resulted in less energy available to midday LE and H.

VPD affected the energy balance partitioning and the Priestley-Taylor α. Latent heat fluxes from a wet surface were always slightly larger than sensible heat fluxes (Bowen ratios below unity) if VPDs were above 0.25 (2006), 0.31 (2007), and 0.28 kPa (2008) (Fig. 5). A VPD above these thresholds during wet soils (including ponding) resulted in a Priestley-Taylor α near one or higher. On the other hand, unusually dry soil ($\Psi < -0.13$ MPa, 20 July–12 August 2007) resulted in evapotranspiration below the equilibrium rate, despite VPD reaching 1.7 kPa. Nevertheless, a VPD > 1.2 kPa resulted in latent heat flux that exceeded the sensible heat flux at dry soils.

A VPD below and above 0.3 kPa resulted in significantly different bulk parameters during wet soils (no ponding) (Table 5). A VPD > 0.3 kPa resulted in slightly increased Priestley-Taylor α and a doubled r_c, while β and Ω was

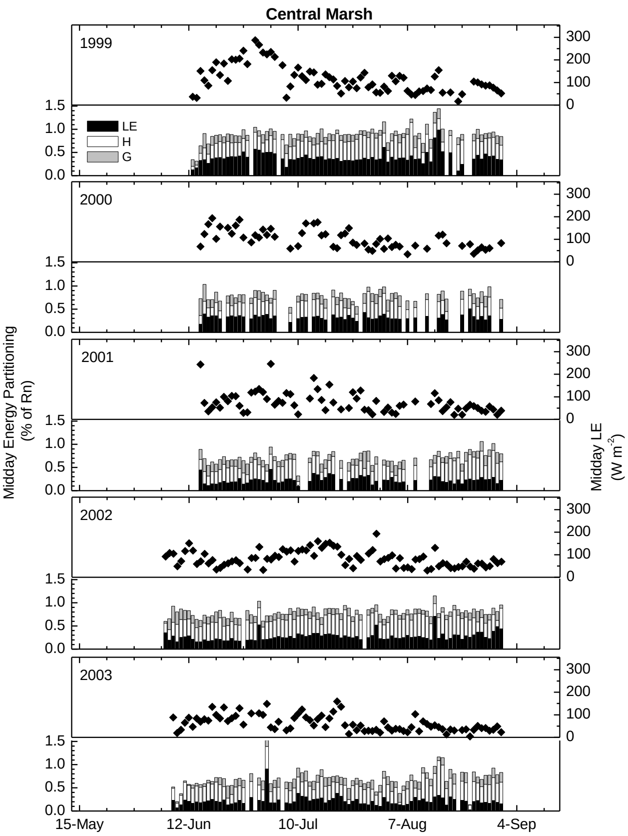

Fig. 3a. The variation in the mean midday energy balance partitioning and evapotranspiration rates during summer 1999–2003 at the Central Marsh site.

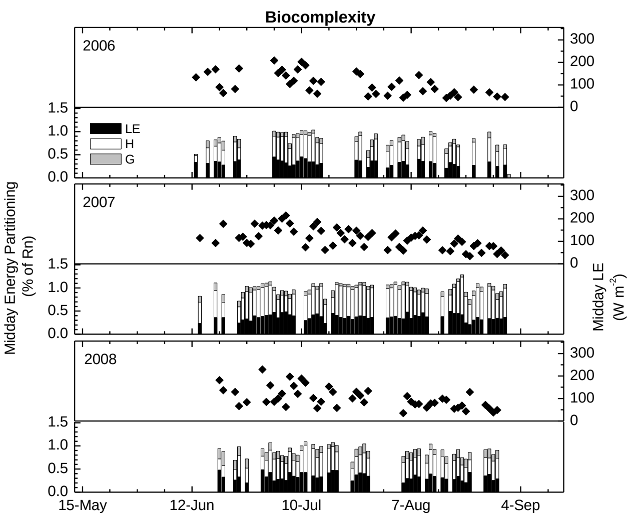

Fig. 3b. The variation in the mean midday energy balance partitioning and evapotranspiration rates during summer 2006–2008 at the Biocomplexity Experiment site.

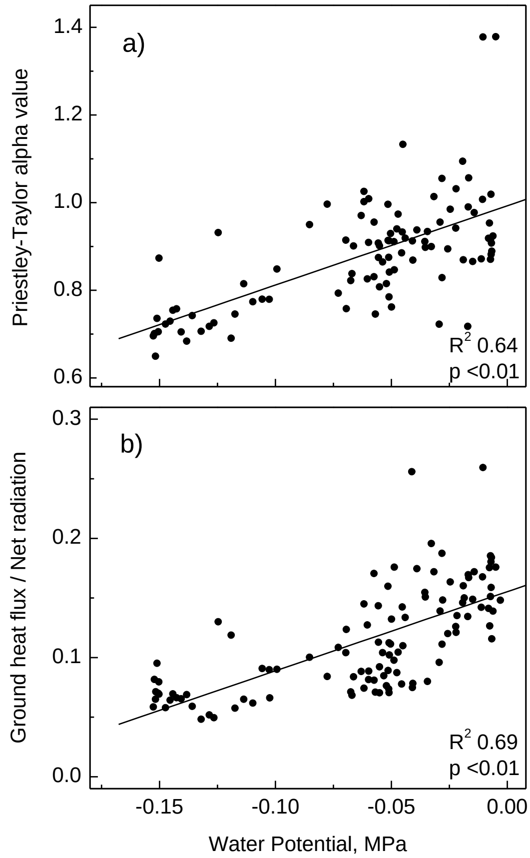

Fig. 4. The rate of evapotranspiration in relation to **(a)** the equilibrium rate (Priestley-Taylor α) and **(b)** near-surface soil moisture at the BE site. The partitioning of net radiation, R_n, into ground heat flux is linearly correlated to near-surface soil moisture (10 cm depth). The results represents mean midday values at the BE site.

reduced (15 July–15 August 1999–2003 and 2006). High bulk surface resistance ($\sim$100 s m^{-1}) often occurred with elevated VPD throughout the study period (Fig. 6). The results suggest that (a) net radiation was the primary control on ET in wet soils when VPD < 0.3 kPa and (b) increased bulk surface resistance suppressed the evapotranspiration during large atmospheric demand even if the soils were wet.

Surface energy partitioning depended on wind direction (Table 2). Onshore winds favored energy partitioning into sensible heat flux (β1.37), while the Bowen ratio was slightly below unity during offshore conditions (0.87) at the CM site. Partitioning into both ground and latent heat increased with offshore winds, while the sensible heat flux portion decreased. No onshore-offshore analysis was performed at the BE site, as onshore and offshore wind directions represent differing landscape features (drained thaw lake and uplands, respectively).

Two days in late July 2000 showed the cascading effects on meteorological conditions and surface energy balance that were induced by altered wind directions (Fig. 7). The first

Table 4. Mean midday energy partitioning and bulk parameters at the BE site between 20 July and 12 August during wet, but not inundated (2006 and 2008), and dry (2007) soil moisture conditions. Only days with VPD below 0.3 kPa are included. Dry soil moisture conditions represent $\Psi < -0.13$ MPa at 10 cm depth, which equals a water table at ∼15 cm depth, and no prior precipitation.

	Wet	Dry
LE/R_n	0.34 ± 0.08	0.37 ± 0.05
H/R_n	0.43 ± 0.10	0.65 ± 0.06
G/R_n	0.14 ± 0.04	0.06 ± 0.02
Closure	0.91 ± 0.14	1.08 ± 0.08
β	1.33 ± 0.32	1.79 ± 0.25
Ω	0.76 ± 0.08	0.63 ± 0.04
α	0.89 ± 0.09	0.72 ± 0.06
r_c (s m^{-1})	41 ± 22	57 ± 14
r_a (s m^{-1})	63 ± 19	48 ± 10
r_i (s m^{-1})	10 ± 6	7 ± 6

Closure = $(LE + H + G)/R_n$, Dry = $\Psi < -0.13$ MPa

Table 5. Average midday energy balance partitioning and bulk parameters during VPD below and above 0.3 kPa at the CM (1999–2003) and BE (2006) sites, during wet, but not inundated, near-surface soil moisture conditions (15 July–15 August).

	Wet Soils	
	VPD < 0.3 kPa	VPD > 0.3 kPa
LE/R_n	0.30 ± 0.14	0.36 ± 0.14
H/R_n	0.38 ± 0.16	0.35 ± 0.14
G/R_n	0.15 ± 0.08	0.14 ± 0.07
Closure	0.83 ± 0.21	0.83 ± 0.19
β	1.45 ± 0.61	1.04 ± 0.53
Ω	0.72 ± 0.22	0.59 ± 0.17
α	0.89 ± 0.20	0.91 ± 0.25
r_c (s m^{-1})	51 ± 47	114 ± 61
r_a (s m^{-1})	63 ± 45	64 ± 42
r_i (s m^{-1})	13 ± 12	39 ± 24

Wet soils = $\Psi \gg -0.13$ MPa

day represented near-normal meteorological conditions, with onshore winds resulting in an equal partitioning of LE and H. Offshore winds occurred during the following day, which resulted in high VPD (1.3 kPa), with LE dominating H. The LE exceeded H when the VPD passed 0.37 kPa (see Fig. 7, vertical dashed lines). Conversely, LE and H became equal later in the afternoon when the VPD returned to 0.37 kPa. Bulk surface resistance and Priestley-Taylor α responded accordingly, with increasing mean midday bulk surface resistance (from 75 to 128 s m^{-1}) and increasing Priestley-Taylor α (from 0.84 to 1.03).

6 Discussion

Our analyses confirm earlier landscape-scale work from coastal Arctic wetlands that relied on Bowen ratio and energy balance techniques as well as plant-scale ecohydrological studies from the Arctic Coastal Plain. The high McNaughton and Jarvis Ω-factor suggests that net radiation was the main control on evapotranspiration rates, but our results show that midday evapotranspiration rates are additionally constrained during both wet and dry near-surface conditions. We concur with previous studies that state the importance of maritime air mass favoring sensible heat (large temperature gradients) and suppressing latent heat flux (low VPD) (Rouse et al., 1987; Lafleur and Rouse, 1988; Price, 1991; Harazono et al., 1998). We also show that near-surface soil moisture conditions and VPD express nonlinear effects on midday evapotranspiration. Ultimately, the various controlling factors (i.e., net radiation; soil moisture; VPD; and, despite wet near-surface soils, bulk surface resistance during high VPD) reduced the evapotranspiration under a range of meteorological and hydrologic conditions, which have the potential to buffer interannual variation of total evapotranspiration. Midday evapotranspiration rates were, on average, 70 % (BE) and 75 % (CM) of the potential rate as defined by a Priestley-Taylor α-value of 1.26.

The generally low vapor pressure deficits (mean midday 0.12 kPa) play an important role in suppressing the evapotranspiration from arctic coastal wetlands. A VPD near 0.3 kPa appears to represent a threshold during wet near-surface soils (Fig. 5). Above 0.3 kPa, latent heat fluxes always dominated sensible heat fluxes, and the evapotranspiration rates always remained near or above the equilibrium rate (Fig. 5, Table 5).

Despite large interannual variations in mean summer air temperature, the number of days exceeding a VPD of 0.3 kPa varied only between 8 (2003) and 14 days (2001). In addition, it was the coldest summer (2001) that had the most days above 0.3 kPa, although the two warmest summers (1999 and 2007) trailed closely behind (12 and 13 days, respectively). Hence, warmer mean summer air temperatures did not necessarily mean an increased number of days with VPDs above 0.3 kPa.

An increased atmospheric demand favored the partitioning of net radiation into LE, but an increased bulk surface resistance – despite wet soils – prevented evapotranspiration from reaching its potential rate ($\alpha \sim 1.26$) (Table 5). The reduction in Ω suggests that VPD increased its role in driving evapotranspiration when VPD reached above 0.3 kPa. Simultaneously, a VPD > 0.3 kPa more than doubled the bulk surface resistance, which limited any increase in the Priestley-Taylor α. The rate of water movement through moss (capillary forces upwards from the water table) is likely not able to support potential evaporation rates. Our landscape-scale findings agree with earlier plot-scale studies of tundra vascular and non-vascular conductance (inverse of resistance) (Johnson and Caldwell, 1975; Oechel and Sveinbjörnsson,

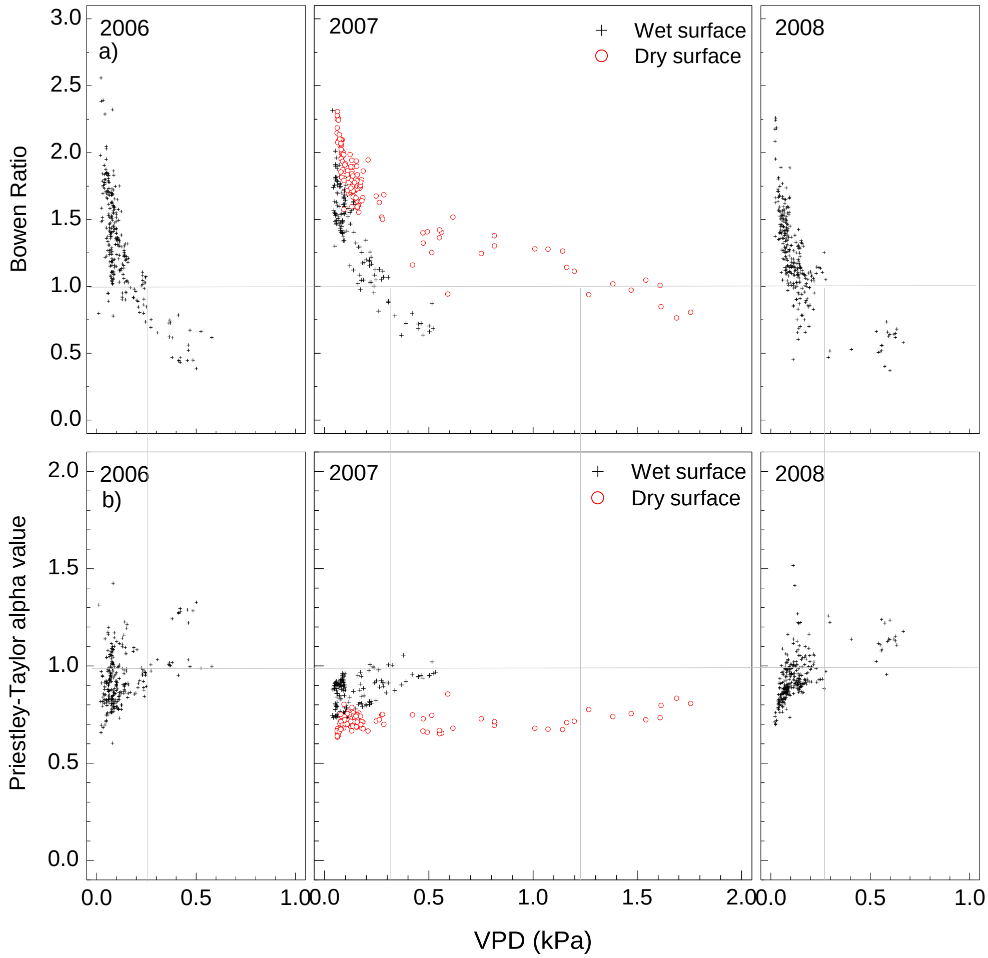

Fig. 5. The relationship between mean hourly air vapor pressure deficit (VPD) and **(a)** Bowen ratio (β) or **(b)** Priestley-Taylor α during differing soil moisture conditions at the BE site, 2006–2008. Dry soils represent a soil water potential < -0.13 MPa at 10 cm depth. The vertical dashed lines represent the identified critical value of VPD. VPDs above this threshold resulted in a $\beta < 1$ and a Priestley-Taylor α near or above 1. The identified VPD thresholds were 0.25 (2006), 0.31 (2007), and 0.28 kPa (2008) for wet soils and 1.19 kPa for dry soils (2007).

1978), in which surface cover (despite wet soils) was unable to deliver enough moisture when atmospheric demands were high.

Hence, the evapotranspiration from the two studied vegetated, drained thaw lakes was suppressed during both low and high VPDs, but through differing mechanisms. The lower VPDs present a direct atmospheric constraint as the air is unable to hold much additional moisture. High VPD results in an indirect constraint on evapotranspiration rates through an insufficient transfer rate of water through the moss layer, expressed through an increased bulk surface resistance.

Near-surface soil moisture plays an important role in controlling energy balance in vegetated, drained thaw lake basins. The higher Priestley-Taylor α and the lower bulk surface resistances during high near-surface soil moisture presented reduced constraints on evapotranspiration (Fig. 4a, Table 4). However, the linkage between evapotranspiration and soil moisture appears to be more complex, since the ratio of latent heat flux to net radiation was similar between dry and wet soils during VPDs < 0.3 kPa. The increased partitioning into ground heat flux during wet (compared to dry) soils reduced the energy available for midday sensible and latent heat flux – a phenomenon which has also been discussed by McFadden et al. (1998). Not unlike the discussion about the nonlinear controls of VPD on evapotranspiration, we suggest that midday evapotranspiration was suppressed during both dry and wet soils but through differing mechanisms: (a) energy limitations (wet soils) through an increased partitioning of net radiation into ground heat flux and (b) water limitations (dry soils).

The multiple nonlinear controls may moderate the spatial variability in the energy partitioning from different vegetation types. Short-term mid-summer measurements of LE/R_n at tussock, tussock-shrub, shrub, and wet sedge tundra

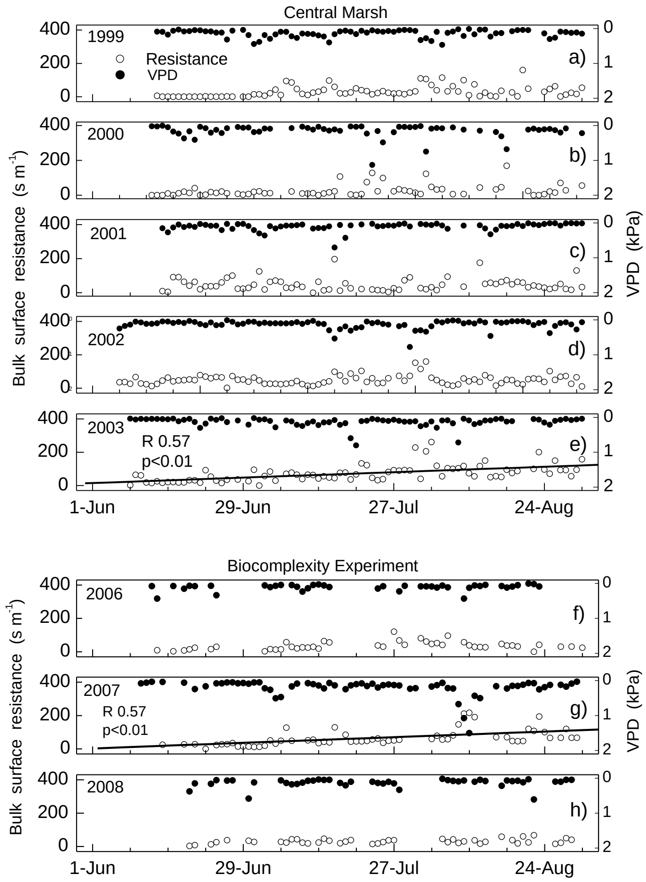

Fig. 6. Mean midday values of bulk surface resistance (r_C) and VPD at the CM **(a–e)** and BE **(f–h)** sites.

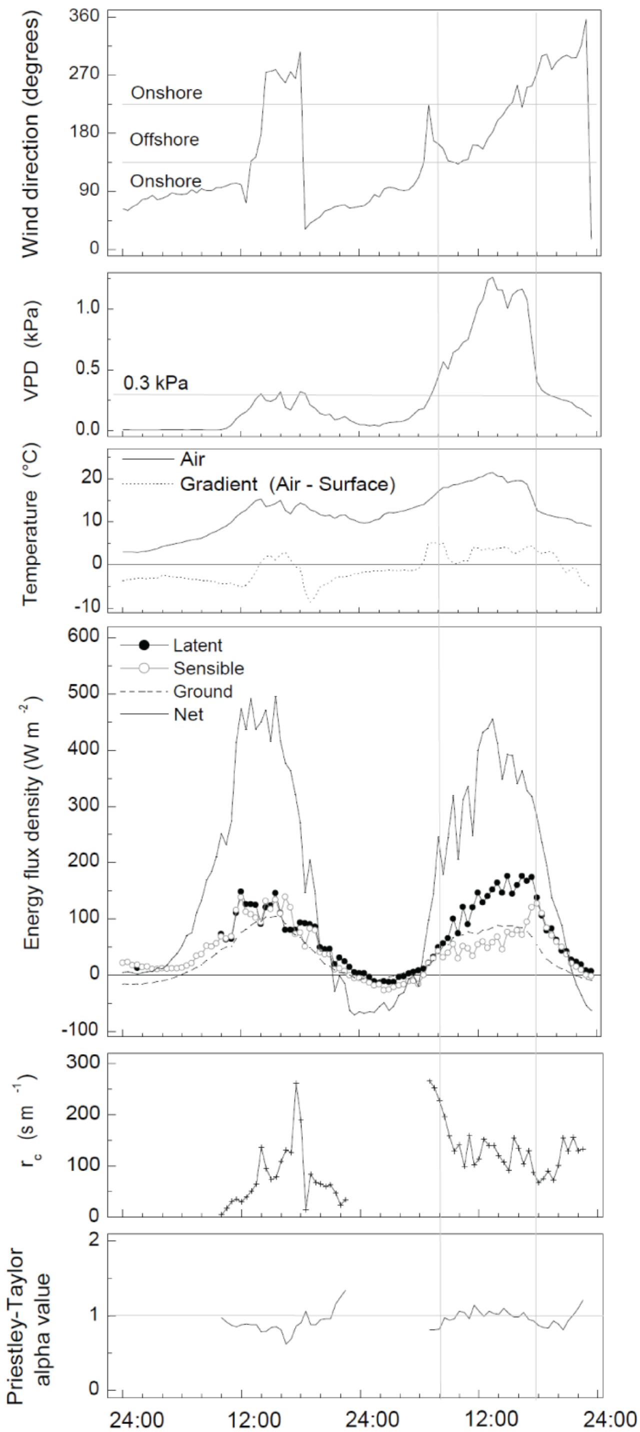

Fig. 7. Meteorological conditions, energy balance, and bulk parameters during a two-day time period (22 and 23 July 2000), when the wind shifted from onshore to offshore. The high VPD on 23 July coincides with offshore winds. Latent heat became the dominant heat sink when air vapor pressure deficit reached above 0.3 kPa.

ranged from 35 to 42 % (mean 38 %) (McFadden et al., 1998), which is close to our wet and dry mid-summer conditions (34–37 %). It is apparent, however, that there is quite a large variability in LE/R_n between sites and time periods from a multitude of short-term eddy covariance measurements across the North Slope of Alaska (Eugster et al., 2000). Nevertheless, values in LE/R_n and other measures presented in this study, agree well with those reported by Harazono et al. (1998), Eugster et al. (2000), McFadden et al. (1998), and McFadden and Chapin III (2003) of non-shrub coastal sites in Arctic Alaska. Accordingly, the details presented in this study are representative of the Arctic Coastal Plain (<135 km from the ocean) even though the energy partitioning (and controls on ET) may show similarities to other locations.

The measurements employed in this study cannot distinguish transpiration from evaporation, but our results can be compared to past findings. Firstly, measured mean ET rates (1.5 and 1.9 mm day^{-1}) were more than twice the maximum vascular transpiration (0.2 mm day^{-1}) estimated by Miller and Tieszen (1972) during peak LAI. Secondly, the measured near-surface soil water potentials never reached the soil water potentials for stomatal closure typical for tundra vascular vegetation, a value established by Stoner and Miller (1975) and Johnson and Caldwell (1975). However, soil water potentials surpassed the limit for effective water transport through moss (−0.1 MPa) in July 2007, which simultaneously saw an increased canopy resistance. Had the vascular vegetation played a dominant role, the observed increase in bulk surface resistance in late summer of 2003 and 2007 would have been less likely. This is because the soil water potential to initiate stomata closure was never reached. In addition, as mosses represent the majority of the live biomass (Zona et al., 2011), one could argue that they represent a key hydrologic pathway between land and atmospheric systems. And, in fact, boreal mosses are known to act as a heat and moisture "rectifier", allowing heat and moisture fluxes to proceed when they are moist, and reducing heat and moisture fluxes under hot and dry conditions when the uppermost moss surfaces are dry (Oechel and Van Cleve, 1986). However, a determination of the amount of transpiration and evaporation on total evapotranspiration would require hydrologic model simulations or isotopic analyses, which are beyond the scope of this study.

Overall, the two vegetated, drained thaw lake basins experienced similar distribution in the energy balance partitioning and bulk parameters, despite differences in weather between the years (Tables 1, 3). It should be noted that at least a third of the data from the BE site represents the unusually low precipitation and warm summer of 2007, which may explain differences in site averages. Still, the general partitioning of the energy balance components were similar, even under a relatively wide range of soil moisture conditions. The limited differences between the two sites and the agreement with previous studies suggest that our findings can represent the larger arctic coastal wetland domain.

Future projections

According to global climate model projections for the mid-21st century, summer air temperature and precipitation will generally increase in the Arctic (Walsh, 2008). Some parts of the land-ocean-atmosphere system are projected to change, but we may also hypothesize resistance in some components of the system. For example, future summer air-mass conditions at the Arctic Coastal Plain are likely to continue to be dominated by a 24-h sea breeze, which brings in moist, cool air that suppresses the evapotranspiration in addition to the nonlinear controls by soil moisture. The future evapotranspiration rates may therefore remain dampened, which is in agreement with Rouse et al. (1992) and in contradiction to the hypothesis of Lafleur (1993).

However, it is challenging to predict long-term effects of climate warming on Arctic wetland hydrology. The integrated response of the coupled permafrost-vegetation-hydrology system to a warmer climate could drastically affect the surface energy exchange by curbing or accelerating hydrologic fluxes. The presented results offer an example of this complex system, as evapotranspiration rates integrated across a common landscape type respond nonlinearly to a multitude of controlling factors.

7 Conclusions

Evapotranspiration from low-relief, vegetated, drained thaw lake basins experience multiple limitations through nonlinear relationships to atmospheric vapor demand and near-surface soil moisture. We estimated that current midday evapotranspiration rates represent, on average, <75 % of the potential rates despite the typical saturated near-surface conditions. Midday evapotranspiration was suppressed through different mechanisms: (a) vapor pressure deficits near 0.3 kPa appeared to be an important hydrological threshold, allowing latent heat flux to persistently exceed sensible heat flux; (b) dry compared to wet soils increased the bulk surface resistance (water-limited); (c) wet soils favored ground heat flux and therefore limited the energy available to sensible and latent heat flux (energy-limited); and (d) wet soils (ponding excluded) combined with large atmospheric demands resulted in an increased bulk surface resistance and therefore suppressing the evapotranspiration to below its potential rate (Priestley-Taylor $\alpha < 1.26$). The latter was likely caused by the limited ability of mosses to transfer moisture during large atmospheric demands. Our landscape-scale analyses agree well with plant-scale ecohydrological studies from the Arctic Coastal Plain. To conclude, there is a resistance in the hydrologic system that curbs soil drying in coastal Arctic wetlands. We propose that the wetness of the arctic coastal wetlands will persist – despite a warming climate – due to the prevailing maritime winds, increased precipitation, and multiple controls on evapotranspiration. Refined projections of future evapotranspiration should also include linkages to geomorphology and vegetation dynamics, which were beyond the scope of this study.

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