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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 functioning. As a consequence, it is important to define the controls on evapotranspiration, which is 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 near and above 0.3 kPa appeared to be an important hydrological threshold, allowing latent heat fluxes to persistently exceed sensible heat fluxes. Dry soils increased the bulk surface resistance (water-limited). Wet soils favored ground heat flux and therefore limited the energy available to sensible and latent heat fluxes (energy-limited). Thus, midday evapotranspiration was suppressed on both dry and wet soils through different mechanisms. We also found that 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 $\alpha < 1.26$). This is 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 (near-surface soil moisture, atmospheric vapor pressure, and the limited ability of mosses that are saturated at depth to transfer water during high atmospheric vapor demands) resulted in an average evapotranspiration rate of up to 75 % of the potential evapotranspiration rate. The multiple limitations on midday evapotranspiration rates have the potential to moderate interannual variation of total evapotranspiration and dampen excessive water loss during a warmer climate. Combined with the prevailing maritime winds and the projected increase in precipitation, these dampening mechanisms will likely prevent extensive future soil drying and hence maintain the presence of coastal wetlands.

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1 Introduction

The response of arctic wetland hydrology to projected climate warming is uncertain. Evapotranspiration is the least understood component in the Arctic hydrologic cycle (Kane et al., 1989, 1992; Vörösmarty et al., 2001; Woo et al., 2008). Regional studies have proposed increased (Lafleur, 1993) to unchanged (Rouse et al., 1992) future evapotranspiration rates from arctic coastal wetlands. As evapotranspiration is the major pathway of water loss from the flat tundra landscape (Rovansek et al., 1996; Mendez et al., 1998; Bowling et al., 2003), an increase in evapotranspiration could reduce the extent of arctic wetlands (Barnett et al., 2005). If soil drying occurs the region that for a long time has sequestered carbon will shift to a carbon source causing a positive feedback to global climate warming (Oechel et al., 1998; Olivas et al., 2010).

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

Measurement of energy fluxes in arctic environments are 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 arctic water balance (see Kane and Yang, 2004) but few studies have conducted multi-year analyses of evapotranspiration measured by 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

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basins (5 yr at one site; 3 yr in an adjacent site) on the Arctic Coastal Plain, Alaska. We hypothesize that the evapotranspiration from these wetlands experience multiple controls apart from surface net radiation. Defining these controls is important in refining our understanding of the future hydrologic regime, Arctic ecosystem changes, and their global implications.

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}). The negative summer net water balance, which is limited to the summer precipitation minus evapotranspiration (P–E) (Mendez et al., 1998) due to sparse summer runoff (Brown et al., 1968; Kane et al., 2008), is offset by the 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) and spring runoff is not generated until the surface stores are replenished (Rovansek et al., 1996; Bowling and Lettenmaier, 2010). Accordingly, evapotranspiration, through P-ET, is not only the major pathway of water loss in summer but it 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 of the evapotranspiration is represented by evaporation from moss and open water (55 to 85 %) (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, attributed to capillary flow, has shown to occur with 0.2 m deep water tables in *Sphagnum* moss (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). 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 ranging from 0.7 to 2 kPa between the leaf and the ambient air (Johnson and Caldwell, 1975). Arctic 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 affect tundra 135 km inland from the Arctic Coast (Harazono et al., 1998). Cold moisture-laden air along the coast increases the partitioning of surface energy into sensible heat flux (*H*) due to (a) a steep temperature gradient between the ground surface and air, which favors *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, at least partly, explains the finding that, despite the wet soils, evapotranspiration is in general below its potential rate in coastal arctic wetlands (Rouse et al., 1987; Mendez et al., 1998). However, it is unclear what values in air vapor pressure deficits result in significant changes to surface energy partitioning and evapotranspiration rates.

Soil moisture may play a major role on tundra surface energy balance partitioning. Coinciding with a seasonal reduction in the water table depth (from surface ponding to 30 cm below the ground surface), a larger portion was partitioned into latent heat rather than sensible heat in early season (Bowen ratio marginally <1), while sensible heat dominated the late season energy balance 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. That evokes the question whether the arctic wetlands display important controlling mechanisms on

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the local hydrological system that constrain evapotranspiration rates not only during dry near-surface conditions but also when wet.

3 Site description

The two sites, hereafter referred to as Central Marsh, CM, (71°19′12.5”N, 156°37′20.211”W, elevation 1 m) and the Biocomplexity Experiment, BE, (71°16′51.17”N 156°35′47.28”W, elevation 4.5 m) are located 4.5 km apart, and both are only 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 and moist wind (mean 5 m s^{-1}) off the ocean from 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., 2009a).

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 polygons are found at the vegetated drained thaw lake basin while high-centered polygons cover the upland areas of the watersheds. 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 studies of drained thaw lakes are limited, which constrains our understanding of interannual controls of evapotranspiration rates from this vast region.

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Non-vascular vegetation contributes significantly to biomass and land cover (Webber 1974, 1978; Oechel and Sveinbjornsson, 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., 2009a, b; 2011). Up to 60 % of the ecosystem net day time CO_2 uptake at the end of the growing season at BE is represented by *Sphagnum* (Zona et al., 2011).

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 2001 (Mano, 2003). Mid-August LAI at the BE vegetated drained lake reached 0.58 in 2006 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 2008 (P. Olivas, unpublished data). Standing dead leaf biomass in the Barrow area reaches LAI of 1.23 (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 are limiting to vascular root growth (Dennis and Johnson, 1970). Moss may reach thicknesses 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). While the rate of thaw is higher in early summer, 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 (mean 36 cm) in 1995–2009 while the mean active layer depth at the BE site was 30 cm (2006), and 26 cm (2007 and 2008) (Shiklomanov et al., 2010).

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4 Methods

4.1 Measurements

We collected summer measurements for five years at the CM site (1999–2003) and three years at the BE site (2006–2008). Energy flux measurements were made at a 10 Hz sampling interval using an eddy covariance system. The path length of anemometer and gas analyzer sensor at CM was 10 cm and the separation distance between the center of sonic anemometer and open-path IRGA sensors was 16 cm. The three components of wind speed, air temperature and water vapor concentration from the above 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. (2009a).

Measurements of the volumetric water content (VWC) at two locations within the CM drained lake basin were obtained 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 here presented as 100 % VWC to indicate ponding.

Hourly atmospheric air pressure for years 1999–2003 were obtained from the National Climatic Data Center 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 (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 defined

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the start date of the summer in order to evaluate the total summer solar radiation received by the snow-free ground surface.

4.2 Eddy covariance calculations

We calculated fluxes of heat and momentum in 30 min. intervals according to typical covariance calculation procedures. The following corrections were applied (Harazono et al., 2003; Zona et al., 2009a): the 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 fluxes 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, which represented half hourly values around solar noon (defined as ±2 h from local solar noon ∼ 14:00 Alaska Standard Time). Extreme amplitudes in the bulk parameters (greater than six times the average) were removed. For daily evapotranspiration rates, gap filling was performed for missing data (<2 consecutive hours) using linear interpolation. Negative latent heat fluxes were given a value of zero.

4.3 Soil moisture analysis

Unfrozen soil moisture as 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), and during winter, the organic soil was assumed to have 6 % saturation with unfrozen water content (Hinzman et al., 1991). Soil water potential

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(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. We adjusted the daily precipitation to account for gage undercatch according to Yang et al. (1998).

4.4 Analysis of resistances and equilibrium evaporation

We quantified parameters affecting evapotranspiration by applying measured heat fluxes to the Penman-Monteith equation (Monteith, 1973) expressed in terms of aerodynamic, r_a, and bulk surface resistance, r_c:

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

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 the saturation vapor pressure (kPa) at T_a, which is the ambient air temperature (K), and e_a 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):

$$r_\mathrm{a} = \frac{u}{u_*^2} + \frac{4}{u_*} \tag{2}$$

where u_* is friction velocity ($\mathrm{m\,s^{-1}}$) obtained by eddy covariance measurements and u is wind speed. From here and onwards, the sum of the aerodynamic and laminar

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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.

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

$$r_i = \frac{\rho \times Cp}{\gamma} \frac{[e_s - e_a]}{Q_a} \quad \text{(Stewart and Thom, 1973)} \tag{3}$$

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(m s^{-1}):

$$r_c = (1+\beta) r_i + \left(\beta \frac{s}{\gamma} - 1 \right) r_a \tag{4}$$

The bulk surface resistance characterizes the control of water loss by a 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 if 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 Penman-Monteith Equation (Monteith, 1973) collapsing into:

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

which is known as the Priestley-Taylor equation. When α equals one, the evapotranspiration is referred to as "equilibrium", which is most commonly achieved when VPD = 0 (note that equilibrium rates can also be measured over unsaturated surfaces and VPD > 0). The method assumes that the latent heat flux depends only upon the absolute temperature and the 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. Larger 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 α is a function of wind speed, surface roughness, and bulk surface resistance. Here we define the potential evapotranspiration by setting the α-value to 1.26.

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

$$\Omega = \left(1 + \frac{s}{s+\gamma}\frac{r_c}{r_a}\right)^{-1} \tag{6}$$

Also, a vigorous turbulent mixing of the air mass suppress Ω by promoting increased VPD at the surface while Ω approaches unity during limited atmospheric mixing (McNaughton and Jarvis, 1983). However, as long as $r_c \gg r_a$, Ω will approach 0. In general, VPD is the main driver of evapotranspiration when Ω is low, while net radiation has the dominant control during Ω near 1.

5 Results

5.1 Meteorological and hydrologic conditions

During the study periods (1999–2003 and 2006–2008), the mean air temperature (3.2 °C) and precipitation (86 mm) were near the long-term means (3.4 °C and 99 mm, respectively year 1979–2008) but large interannual variations occurred (Table 1). Summer 2007 experienced unusually high air temperatures (5.4 °C) and low precipitation (24 mm) with most of the precipitation occurring in a single event in mid-August. During the study period, 77 % of the daily precipitation rates were less than 2 mm day^{-1} and

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33 % of all recorded observations were trace events. Accumulated winter precipitation ranged from 93 to 158 mm of snow water equivalent (SWE).

The maritime air mass reduced variation 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 amongst all years (0.10–0.13 kPa) except summer of 2007 which was higher (0.17 kPa) (Table 1). The maximum VPD recorded was 1.76 kPa, but days exceeding VPD's of 0.3 kPa were few (8 to 14 days per summer).

Onshore summer winds (defined as 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 VPD's and air temperatures than onshore winds, reversing the typical midday temperature gradient between the air and the ground surface.

The porous organic soils remained close to saturation near the surface throughout the study period with the exception of 2007 (Fig. 2). Even though the soils within the vegetated drained lake basins were unusually dry in late summer 2007 (water table dropped below 15 cm depth), the typical extensive ponding of snowmelt water in spring recharged the soil water storage in spring 2008. Water table measurements at the BE basin show a multi-week long ponding period following the snowmelt with ∼10 cm of water accumulating above the ground surface (note that the water table measurements did not capture the start of the inundation). In two of the eight summers (2001 and 2008), the drained lake basins also experienced inundation in late summer, resulting in an inundation for at least half of the summer's duration. Unfortunately, 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

A major portion of the midday surface energy balance was partitioned into sensible heat flux CM: 35 %, BE: 46 %) resulting in a mean midday Bowen ratio above unity at both sites (CM:1.40, BE:1.38) (Table 3). Latent heat flux represented 29 % and

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36 %, and ground heat flux averaged 16 % and 12 % at CM and BE, respectively. Measured evapotranspiration ranged from 0.2 to 5.7 mm day^{-1} (mean 1.8 mm day^{-1}) and 1.3 to 7.2 mm day^{-1} (mean 4.2 mm day^{-1}) at the CM and BE sites, respectively. The high McNaughton and Jarvis Ω-factor (CM:0.74, BE:0.79) suggests that net radiation was the main control on evapotranspiration rates. The energy balance closure was not complete (BE 97 % and CM 80 %) but comparable to other tundra and grassland ecosystems reported by Cava et al. (2008), Ryu et al. (2008), and Wilson et al. (2002).

The large variation in near-surface soil moisture at the BE site in 2007 allowed an analysis of its affect on the surface energy balance. Approximately 71 % ($P < 0.01$, probability that correlation is zero) of the variance in the Priestley-Taylor α was correlated to the near-surface soil moisture (Fig. 3a). Accordingly, bulk surface resistance was higher during dry (Ψ <–0.13 MPa) near-surface soil conditions (Table 4). Wetter soils also increased the partitioning to ground heat flux from ~5 % (dry soil, i.e. soil water potential of <–0.13 MPa) to 15 % (saturated conditions) (Fig. 3b, Table 4). An individual day in early summer reached up to 40 % (G/net radiation). Late-summer (20 July through 12 August) midday partitioning of net radiation into LE was strikingly similar between dry and wet soils (38 and 35 %, respectively), suggesting a suppression of midday LE also under wet near-surface conditions.

The surface energy partitioning depended on wind direction (Table 2). Onshore winds favored energy partitioning into sensible heat flux (β 1.37), while the Bowen ratio was slightly below unity during offshore conditions (0.87) at the CM site. Both the partitioning into 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 the two wind directions represents differing landscape features (drained thaw lake and uplands, respectively).

High bulk surface resistance (~100 s m^{-1}) often occurred with elevated VPD throughout the study period (Fig. 4, Table 5). Focusing on a period with wet soils but with no ponding (15 July–15 August 1999–2003 and 2006) midday VPD's above 0.3 kPa resulted in a higher mean bulk surface resistance (120 s m^{-1}) than those below 0.3 kPa

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($50\,\mathrm{s\,m^{-1}}$). The increased bulk surface resistance suppressed the evapotranspiration during large atmospheric demand.

Increased 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) at the BE site (Fig. 5). The same threshold for dry soils ($\Psi < -0.13$ MPa) was at 1.19 kPa (2007). 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, such in late July through early August 2007, resulted in evapotranspiration below the equilibrium rate despite VPD reaching 1.7 kPa. Separating the midday bulk parameters from both the BE and CM site based upon the ~0.3 kPa threshold found in Fig. 5 resulted in a similar trend during wet soils (no ponding) (Table 5), where a high VPD resulted in an increased Priestley-Taylor α and reduced β. Overall mean midday evapotranspiration was near or slightly below the equilibrium rate (CM 0.94, BE 0.88) (Table 3) resulting in an average evapotranspiration rate of up to 75 % (CM) and 70 % (BE) of the potential rate.

Two days in late July 2000 show the cascading effects on meteorological conditions and surface energy balance that were induced by altered wind directions (Fig. 6). The first day represent 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) and LE dominating H. The LE exceeded H when the VPD passed 0.37 kPa (see 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 $128\,\mathrm{s\,m^{-1}}$) and Priestley-Taylor α (from 0.84 to 1.03).

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6 Discussion

Our analyses confirm earlier landscape-scale work from other locations 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 (CM: 0.74, BE: 0.79) 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 non-linear effects on midday evapotranspiration. Ultimately, the various controlling factors (net radiation, soils 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 has 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. However, this percentage should be considered an upper value as it does not include the affect of increased partitioning of net radiation into ground heat flux during wet soils.

The generally low vapor pressure deficits (mean midday 0.12 kPa) play an important role in suppressing the evapotranspiration from the 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 the sensible heat fluxes, and the evapotranspiration rates always remained near or above the equilibrium rate (Fig. 5, Table 5).

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 (α ~1.26) (Table 5). The 19 % reduction in

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Ω suggests that VPD increased its role in controlling evapotranspiration when VPD reached above 0.3 kPa. Simultaneously, a VPD >0.3 kPa more than doubled the bulk surface resistance, which limited the increase in the Priestley-Taylor α to +5 %. The rate of water movement through moss (capillary forces upwards from the water table) are 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, 1978) where the 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 lake was suppressed during both low and high VPD's, but through differing mechanisms. The lower VPDs present a direct atmospheric constraint as the air is unable to hold much additional moisture, while the high VPD results in an indirect constraint on evapotranspiration rates through an insufficient transfer rate of water through the moss layer that is expressed through an increased bulk surface resistance.

Despite large interannual variations in mean summer air temperatures, 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 do not necessarily mean an increased number of days with VPD's above 0.3 kPa.

Near-surface soil moisture plays an important, possibly non-linear, role in controlling energy balance in vegetated drained thaw lake basins. Firstly, the higher Priestley-Taylor α and the lower bulk surface resistances during high near-surface soil moisture presented reduced constraints on evapotranspiration (Fig. 3, 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 amongst dry and wet soils during VPD's below 0.3 kPa (Table 4). The partitioning into ground heat flux was larger for wet soils, which reduced the energy available for midday sensible and latent

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heat flux. 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 through but through differing mechanisms; (a) water-limitations and (b) energy-limitations through an increased partitioning of net radiation into ground heat flux.

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 (1.8 and 4.2 mm day^{-1}) were more than twice as high as 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, which value was 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 because the soil water potential initiating 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 dry conditions when the uppermost moss surfaces dry (Oechel and Van Cleve, 1986). However, a determination of the contribution and 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 amongst the years (Tables 1 and 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 the site-to-site difference in mean evapotranspiration rate, Priestley-Taylor α, and energy closure in Table 3. Also, the VPD analysis includes (Fig. 5) and excludes (Table 5) ponded conditions. Still, the general partitioning of the energy balance components were similar. The limited differences between the two sites suggest that our findings are representative of other coastal vegetated drained thaw lake basins.

6.1 Future projections

According to global climate model projections for the mid-21st century, 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-hour sea breeze, which brings in moist cool air that suppresses the evapotranspiration in addition to the non-linear controls by soil moisture. The future evapotranspiration rates may therefore remain dampened, which is in agreement with Rouse et al. (1992), while in contradiction to the hypothesis of Lafleur (1993).

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

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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, during both high and low VPD as well as both wet and dry near-surface soils. Our landscape-scale analyses agree well with plant-scale ecohydrological studies from the Arctic Coastal Plain. In other words, there is a resistance in the hydrologic system that dampens soil drying of coastal arctic wetlands. Assuming no changes in vegetation and microtopography, 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.

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Table 1. Meteorological conditions during the study period (1999–2008) such as Snow Water Equivalent (SWE) prior to snowmelt, total precipitation from June through September, mean air temperature June through August, mean midday 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.

	1999	2000	2001	2002	2003	2006	2007	2008	Mean
SWE (mm)	122	113	123	93	95	137	98	158	117
Precipitation, Jun–Sep (mm)	82	128	124	114	72	72	24	68	86
Mean air temperature, Jun–Aug (°C)	4.2	3.1	2.1	2.3	2.5	2.9	5.4	3.3	3.2
Mean midday VPD, Jun–Aug (kPa)	0.12	0.13	0.1	0.12	0.12	0.11	0.17	0.12	0.12
VPD > 0.3 kPa, Jun–Aug (days)	12	10	14	9	8	10	13	10	11

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Table 2. Differences in 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. The temperature gradient (ΔT) represents the air minus the ground surface temperature.

	Offshore	Onshore
LE/Q	0.41	0.29
H/Q	0.22	0.35
G/Q	0.22	0.15
β	0.87	1.37
α	1.08	0.95
ΔT (°C)	2.33	−1.30
VPD (kPa)	0.21	0.12

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Table 3. Energy balance partitioning and bulk parameters at the CM (1999–2003) and BE (2006–2008) sites.

	Central marsh	Biocomplexity experiment
LE/Q	0.29	0.36
H/Q	0.35	0.46
G/Q	0.16	0.12
Closure	0.80	0.97
β	1.40	1.38
Ω	0.74	0.79
α	0.94	0.88
r_c (s m^{-1})	46	40
r_a (s m^{-1})	62	62
r_i (s m^{-1})	14	6

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

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Table 4. Midday (12:00 through 16:00) energy partitioning and bulk parameters at the BE site between 20 July and 12 August during wet and dry near-surface conditions. Only days with VPD below 0.3 kPa are included.

	Wet	Dry
LE/Q	0.35	0.38
H/Q	0.43	0.65
G/Q	0.14	0.06
Closure	0.92	1.10
β	1.30	1.75
Ω	0.84	0.61
α	0.90	0.73
r_c (s m^{-1})	29	63
r_a (s m^{-1})	62	48
r_i (s m^{-1})	6	10

Closure = (LE + H + G)/Q
Dry = Ψ < –0.13 MPa

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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) site 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/Q	0.30	0.37
H/Q	0.38	0.36
G/Q	0.15	0.14
Closure	0.83	0.82
B	1.45	0.94
Ω	0.72	0.58
α	0.89	0.93
r_c (sm^{-1})	50	120
r_a (sm^{-1})	62	64
r_i (sm^{-1})	12	14

Wet soils = $\Psi > -0.13$ MPa but no ponding
Represents 15 July – 15 August in 1999–2003 and 2006

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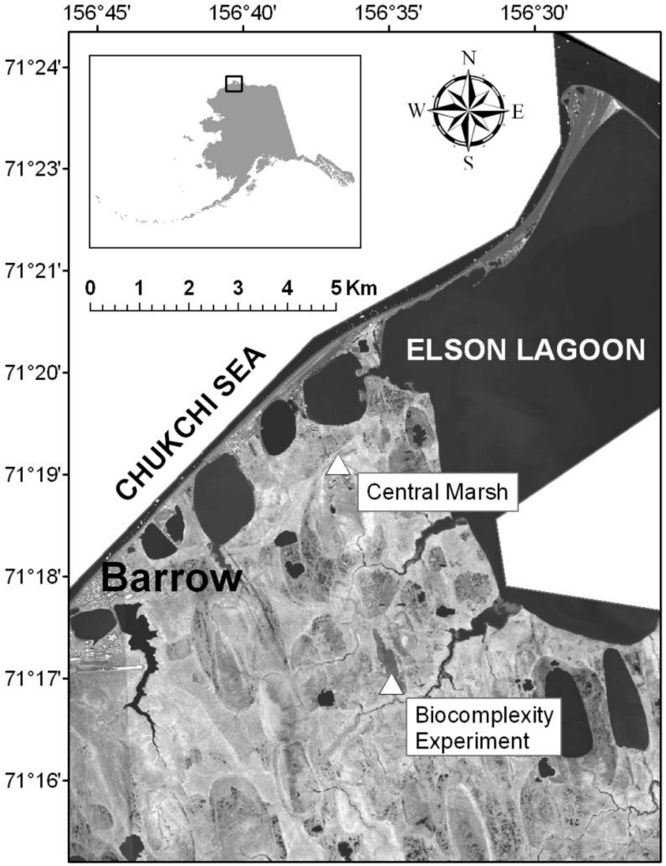

Fig. 1. The Central Marsh (CM) and the Biocomplexity Experiment (BE) sites are located at separate vegetated drained thaw lake basins within 3 km from the ocean outside the town of Barrow, Northern Alaska.

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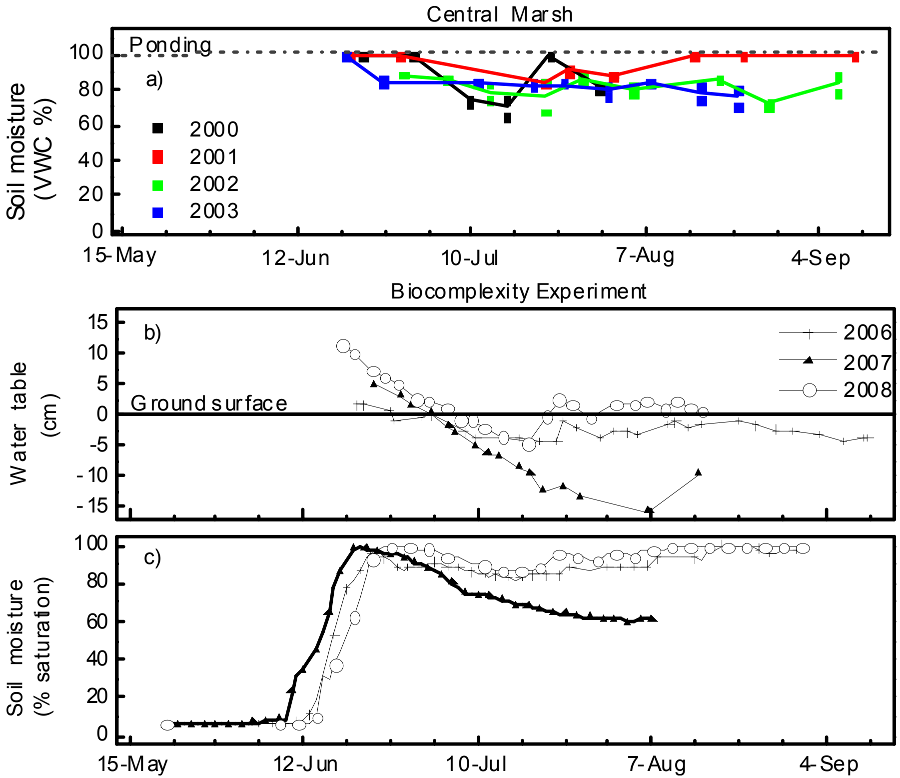

Fig. 2. The soil water status during the study period (no measurements from 1999) at the CM site **(a)** and the BE site **(b and c)**. Figure 2a and b represents multiple locations across the vegetated drained lake basins, while Fig. 2c is a continuous record of volumetric soil water content measurements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

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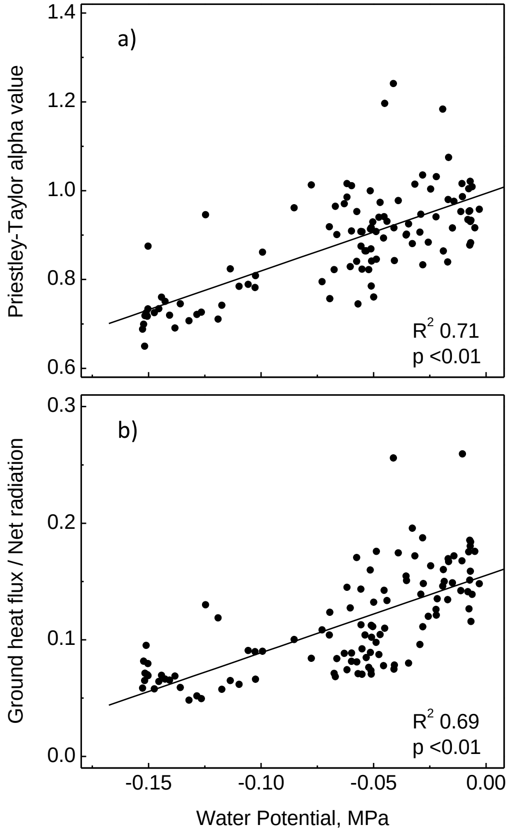

Fig. 3. The rate of evapotranspiration in relation to the equilibrium rate (Priestley-Taylor α) **(a)** and near-surface soil moisture **(b)** at the BE site. The partitioning of net radiation into ground heat flux is linearly correlation to near-surface (10 cm depth). The results represents mean midday values at the BE site.

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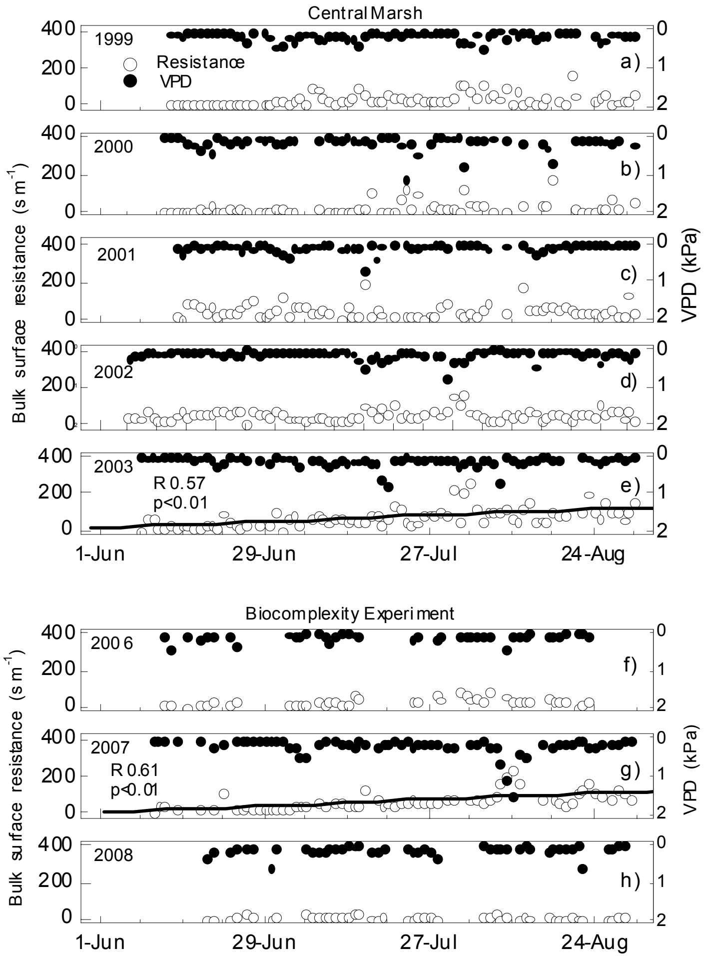

Fig. 4. Mean midday values of bulk surface resistance (r_c) and air vapor pressure deficit (VPD) at the CM **(a–e)** and BE **(f–h)** sites.

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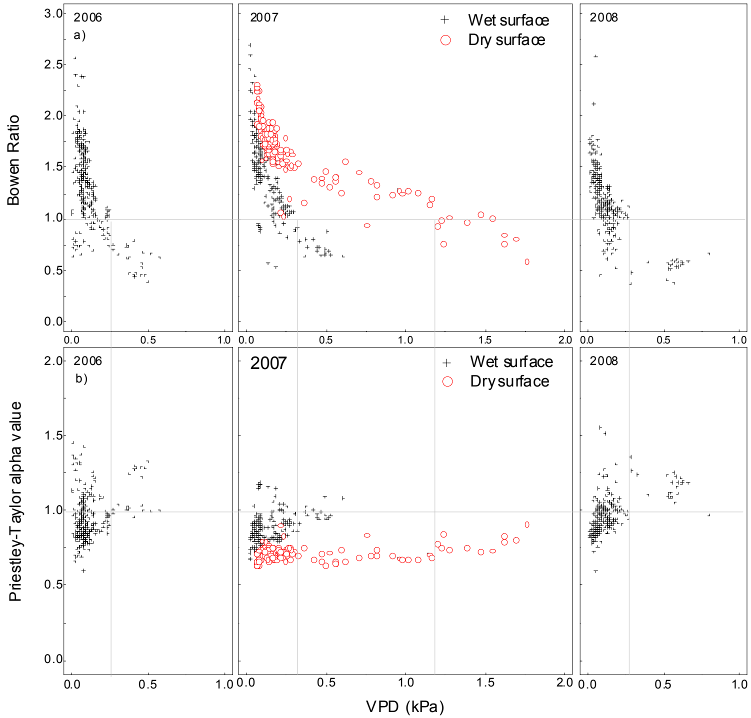

Fig. 5. The relationship between hourly air vapor pressure deficit (VPD) and Bowen ratio (β) **(a)** or Priestley-Taylor α **(b)** during differing soil moisture conditions at the BE site 2006–2008. Dry soils represent a soil water potential below –0.13 MPa. The vertical dashed lines represent a VPD-threshold higher VPD's resulted in a β below 1 and the subsequent response in the Priestley-Taylor α with an 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).

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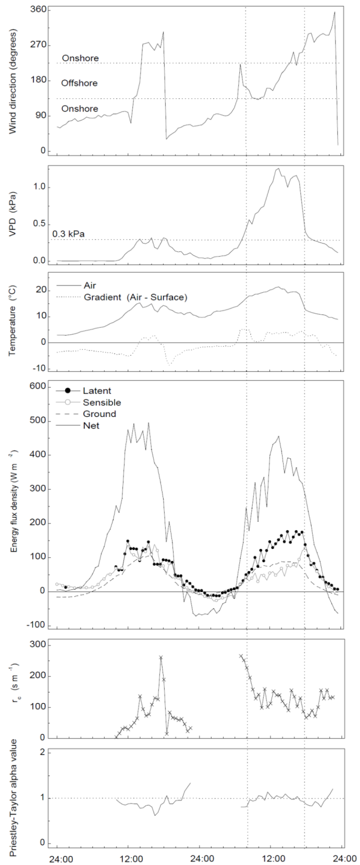

Fig. 6. Meteorological conditions, energy balance, and bulk parameters during a two-day time period (22 and 23 July 2000) when the wind shifted from onshore to offshore. The high VPD on 23 July coincides with offshore winds. Latent heat became the dominant heat sink when air vapor pressure deficit reached above 0.3 kPa.

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