Agricultural and Forest Meteorology 178–179 (2013) 106–119

Contents lists available at SciVerse ScienceDirect

Agricultural and Forest Meteorology

journal homepage: www.elsevier.com/locate/agrformet

Effect of long-term water table manipulation on peatland evapotranspiration

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ARTICLE INFO

Article history:
Received 14 August 2012
Received in revised form 18 April 2013
Accepted 20 April 2013

Keywords:
Eddy covariance
Peatland
Evapotranspiration
Surface resistance
Water table

ABSTRACT

Continuous measurements of ecosystem scale evapotranspiration (ET) were obtained using the eddy covariance method over the 2010 and 2011 growing seasons (May–September) at three adjacent peatlands that have undergone long-term water table manipulation. The three (wet, dry and intermediate) sites represent peatlands along a hydrological gradient, with different average depths to water table (WTD) and different resulting vegetation and microform assemblages. The 2010 growing season was warmer and wetter than normal, while 2011 conditions were near normal. The difference in maximum daily ET values (95th percentiles) between sites were greater in 2010 (3.14 mm d^{-1}–4.17 mm d^{-1}) compared to 2011 (3.68 mm d^{-1}–3.95 mm d^{-1}), yielding cumulative growing season ET that followed the wet to dry gradient in both 2010 and 2011. Synoptic weather conditions (i.e. air temperature, vapour pressure deficit, and incoming solar radiation, etc.) could not explain differences in ET between sites due to their proximity to one another. Peat surface wetness was more spatially homogeneous at the wet site due to small average microtopographic variations (0.15 m) compared to the intermediate (0.30 m) and dry (0.41 m) sites. Although average Bowen ratios were less than one at all three sites, greater surface wetness and heating at the wettest site lead to differences in energy partitioning, with higher average Bowen ratios at the sites with a shallow average WTD. No significant relation between normalized ET and WTD was found at any of the sites that were consistent across both study years. In addition, the lack of a relation between ET and near-surface moisture suggests that the unsaturated hydraulic conductivity and the boundary layer resistance created by the vascular canopy combined with low surface roughness limits evaporative losses from the peat surface. This study suggests that the low ET of a dry site compared to a wet site may be due to the impact of a long-term change in WTD on leaf area and the relative distribution of plant functional groups.

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1. Introduction

While peatland ecosystems cover only approximately 3% of the world's total land surface (Gorham, 1991), peatlands can cover in excess of 50% of the landscape in regions such as Northern Alberta, the Hudson/James Bay Lowland (e.g. Tarnocai, 1984), Fenno-Scandinavia, and the West Siberian Plain (e.g. Paavalainen and Paivanen, 1995). In areas where peatlands are abundant, these ecosystems can have a significant impact on regional climate through surface–atmosphere exchanges of mass and energy, where evapotranspiration (ET) is a large component of the water balance across peatland types (Price and Maloney, 1994; Kellner and Halldin, 2002).

Peatland ecosystems also represent a potential long-term impact on future climate (Frolking. et al., 2011) through their role in the global carbon cycle, where the oxidation and/or combustion of the large carbon pool stored in peatlands (e.g. Gorham, 1991; Vasander and Kettunen, 2006; Turetsky et al., 2011) would represent a positive feedback to current climate forcings. Rapid and/or long-term persistent oxidation of peatland carbon stores has been inferred from future climate scenario modelling studies (Ise et al., 2008) and from experimental manipulation where sites are impacted by multiple disturbances, such as temperature change and drainage (Strack and Waddington, 2007; Chivers et al., 2009; Huemmrich et al., 2010). However, to improve our understanding of the role of peatland surface–atmosphere interactions on regional climate and the potential of peatlands to feedback on global climate change, a greater level understanding on peatland surface energy partitioning and the controls on ET is required (Den Hartog et al., 1994; Kellner, 2001; Kettridge and Baird, 2006).

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http://dx.doi.org/10.1016/j.agrformet.2013.04.013

Due to the wet conditions prevalent in many peatlands, past research has sought to characterize simple metrics relating peatland ET to open-water or equilibrium evaporation and water table depth (WTD). Equilibrium evaporation is defined as the evaporation rate from an extensive, well watered surface where the overlying air mass has no saturation deficit (Monteith, 1981), and differs from open-water evaporation since the effect of entrainment of air from above the convective boundary layer (Priestley and Taylor, 1972) is not taken into account. In general, peatlands evaporate at or near the equilibrium level, similar to high productivity broadleaf forests in the boreal region, but less than open-water evaporation (Baldocchi et al., 2000; Humphreys et al., 2006), suggesting a weak but important control of vapour pressure deficit (D) or water limitation. The relation between ET, and WTD for peatlands in general is not straight forward as studies have shown a range of responses including a strong WTD relation (Lafleur and Roulet, 1992), no significant effect of WTD (Paramentier et al., 2009 and Wu et al., 2010), and a moderate or threshold type response (Lafleur et al., 2005). The control of ET by D in peatlands is similarly varied. In peatlands dominated by vascular vegetation having a relatively high LAI, there is evidence that ET is strongly controlled by surface resistance (Comer et al., 1999), and thus by D. In contrast, clear-day ET was shown to be more strongly controlled by net radiation in a moss-dominated peatland (Price, 1991). In assessing the controls on ET, surface resistance has been shown to be better correlated with D in comparison to water availability (Kellner, 2001). However, the strength of the relation between ET and D can deteriorate under drying conditions (Admiral et al., 2006), where a change in the ET regime may only be detected at relatively large WTD (Lafleur et al., 2005).

These findings suggest that the range of responses in peatland ET are, in part, due to the disparate controls on moss evaporation and vascular plant ET, and their respective links to the spatio-temporal variability in moisture availability. Moreover, the relative abundance of vascular vegetation within a given peatland will dictate whether a peatland can be modelled using a big-leaf model with ET partitioned into soil, vegetation, and interception loss components such as in the ORCHIDEE land surface scheme (Ducoudre et al., 1993), or if a more detailed dual source model is required (Kim and Verma, 1996; Lafleur et al., 2005). The Canadian land surface scheme (CLASS) and Community Land Model have recognized the importance of both organic soils (e.g. Letts et al., 2000; Lawrence and Slater, 2008) as well as the need for a dual source model to better characterize surface evaporation (e.g. Verseghy et al., 1993; Lawrence et al., 2011). Nevertheless, despite having dual source ET and organic soil parameterization, Comer et al. (1999) show that CLASS does not simulate ET well for peatlands with a sparse vascular canopy, suggesting a potential need for better parameterization of moss evaporation. Furthermore, potentially important dynamics which are missing from peatland classifications in land surface schemes include the role of microtopography on the spatial variability in moisture (e.g. Wu et al., 2011) and plant functional groups in peatlands (e.g. Korner et al., 1979), as well as the ability for microtopography and plant functional groups to change through time as a result of changing environmental conditions.

The response of peatland ET to changes in moisture availability has been characterized using interannual variability at a site (Lafleur et al., 2005; Liljedahl et al., 2011), climatic gradients between sites (Humphreys et al., 2006), and through field or laboratory manipulations (Whittington and Price, 2006). Measurements in years with different synoptic weather conditions can be used to quantify short-term, direct impacts on surface energy partitioning that result from warmer/colder or wetter/drier conditions (Teklemariam et al., 2010), but often both moisture availability, temperature, and other synoptic weather variables vary simultaneously so specific impact attribution can be difficult to elicit. Comparisons between sites along climatic gradients are often used to represent long-term impacts of climate (e.g. Chapin et al., 2000; Eugster et al., 2000), where plant community composition is assumed to be in quasi-equilibrium with the local climate. However, attribution of change can be confounded by other overlapping gradients (e.g. nitrogen deposition, Lund et al., 2009). Manipulation studies, which are predominantly focused on carbon fluxes (e.g. Oechel et al., 1998; Weltzin et al., 2003; Strack and Waddington, 2007; Chivers et al., 2009; Huemmrich et al., 2010), use warming or draining to control for different anticipated climate change effects, but are often limited to a few years and thus are potentially measuring the response of a system in disequilibrium. As such, there is a need to characterize peatland ET following a persistent change in moisture availability, where sufficient time has elapsed for both peat hydrophysical properties and the vegetation community to come into a quasi-equilibrium with the new hydrological conditions. In order to be able to constrain the degrees of freedom in a comparative study, the ideal research design would be to examine impacted and control sites in close enough proximity to one another that they can be considered to be subjected to the same synoptic weather conditions. To address this research need we took advantage of a large scale 'natural experiment' resulting from the construction of berms perpendicular to water flow in a large peatland complex in northern Michigan ca. 75 years ago by measuring growing season ET at three adjacent peatlands along a post-disturbance hydrosequence using the eddy covariance (EC) method. The objectives of this study were to: (i) quantify ET along a hydrosequence under the same synoptic weather conditions; (ii) determine if growing season ET was different between sites; and (iii) determine whether energy partitioning and/or controls on ET differed between sites.

2. Methods

2.1. Study site

The study area is located in the Seney National Wildlife Refuge (SNWR) in the Upper Peninsula of Michigan (46.20° N, 86.02° W, elev. ~205 m.a.s.l.). The region is characterized by relatively flat topography which slopes towards the south-east at 1.1–2.3 m km^{-1} (Heinselman, 1965). Land cover consists primarily of upland forests (67%), and open peatlands (20%), while the remainder of the region is made up of mostly forested swamps and open water (Casselman, 2009). Wet sand of glacial origin overlies the regional bedrock up to a thickness of 60 m (Albert, 1995). Surface soils throughout SNWR are mostly a complex of poorly drained muck and sand, while the study area itself contains poorly drained peats (Casselman, 2009).

The peatland complex within SNWR is subdivided by a combination of upland sand ridges of lacustrine origin, creeks, old drainage ditches, as well as a network of roads and berms constructed mostly during the late 1930s and early 1940s by the Civilian Conservation Corps (Kowalski and Wilcox, 2003; Wilcox et al., 2006) for the purpose of creating open-water habitat for migratory birds. As a result of the construction of berms perpendicular to water flow in the study area ca. 75 years ago, a natural experiment was created where an upslope peatland site (hereafter referred to as WET) likely became wetter and a downstream peatland site (hereafter referred to as DRY) likely became drier. Aerial photography of the study area shows a clear increase in ponding at the WET site and a rapid terrestrialization at the DRY site following berm construction. Moreover, contemporary WTD data shows a large difference between the WET and DRY sites, with the latter being up to 50–60 cm deeper. A site with an intermediate WTD variation (hereafter referred to as INT) is located adjacent to the WET site. All three study sites are located in the southern portion of SNWR and have greater than 200 m of

continuous fetch in the dominant wind sector, with only 100 m along the lateral wind sectors. All three sites are located directly adjacent to one another along a roughly north–south transect, and separated by sandy upland ridges between 25 and 100 m wide. The maximum distance between tower-based measurements (Section 2.2) is approximately 750 m.

Dominant vegetation at all three peatland sites consists of a ground cover of *Sphagnum* sp. (*S. angustifolium, S. capillifolium, S. magelanicum*), with an overstory of vascular species. The dominant vascular vegetation consists of *Carex oligosperma*, *Eriophorum vaginatum*, *Ericaceae* (e.g. *Chamaedaphne calyculata*, *Ledum groenlandicum*, *Kalmia polifolia*, and *Vaccinium oxycoccus*), with trees becoming more abundant at sites with increasing WTD (e.g. *Picea mariana*, *Pinus banksiana*, and *Larix laricina*), particularly towards the margins. The berm and upland area surrounding the research sites is dominated by coniferous species such as red pine (*Pinus resinosa*), eastern white pine (*Pinus strobus*), jack pine (*Pinus banksiana*), and black spruce (*Picea mariana*). Although each peatland site has almost identical species richness, the relative proportion of given species differs between sites (J. Hribljan, personal communication). Generally, shrubs dominate hummocks, and sedges dominate hollows across sites, where the proportion of hummocks, height of hummocks relative to adjacent hollows, and acidity of near-surface water increases from the WET, to INT and DRY sites (Table 1). Moreover, both leaf area index (LAI) and plant area index (PAI) are highest at the WET site (Table 1).

Based on the 1981–2010 climate normals from the Newberry 3S weather station (46.32° N, 85.50° W, elev. ~260 m.a.s.l.) ~45 km from the study sites, the area has an average annual temperature of 5.6 °C and receives 784 mm of precipitation, where greater monthly totals occur during May through October (NOAA NCDC, 2012).

2.2. Instrumentation

Standard EC equipment was used at all three sites to measure surface energy and mass exchanges based on the method described by Baldocchi et al. (1996). Three component wind speed and sonic air temperature were measured using a CSAT3 3D sonic anemometer-thermometer [Campbell Scientific Inc. (CS), Logan, UT, USA]. Water vapour and CO_2 concentrations were measured using an LI-7500A open-path infrared gas analyzer (IRGA) (LI-COR Biosciences, Lincoln, Nebraska). All EC sensors were mounted between 1.7 and 2.1 m above the average hummock surface, had a vertical and lateral separation less than 0.15 and 0.3 m, respectively, and were oriented upwind of the tower based on the dominant summertime wind direction. EC data was sampled at 10 Hz, with mean values and fluxes calculated every 30 min.

Radiation measurements were made using a combination of CMP3 and LP02 pyranometers (CS), NR-LITE net radiometers (CS), and CNR2 net short- and long-wave radiometers (CS). A CNR2 was used at the INT and DRY site mounted at a height of 1.78 and 3.2 m, respectively, while a NR-LITE was used at the WET and INT site mounted at a height of 1.7 and 2.3 m, respectively. Both types of net radiometers were used at the INT site to provide a continuous cross-comparison of instrument type in order to correct for any systematic bias. In order to calculate albedo for all three sites, a single CMP3 was mounted 1.7 m above the sedge canopy at the WET site to measure incoming short-wave radiation. Outgoing short-wave radiation was measured at the WET site using a LP02 mounted at 1.7 m above the sedge canopy. Outgoing short-wave radiation was obtained by difference from net shortwave radiation at each of the INT and DRY sites using a CNR2.

Supplementary air temperature and relative humidity (RH) were measured and recorded at 10 Hz using a HMP45C (Vaisala Oyj, Helsinki, Finland) temperature and RH probe mounted in a radiation shield at a height of 1.35 m on all EC towers. Peat temperature profiles were measured in a representative hummock and hollow at each site using T-type thermocouple (Omega Engineering, CT, USA) wire inserted at depths of 0.01, 0.05, 0.1, 0.2, and 0.5 m relative to the local surface.

Measured hydrometric data included rainfall, soil volumetric water content (θ), and WTD. Rainfall was measured using a TE525 tipping bucket rain gauge (Texas Electronics, Dallas, TX, USA) mounted 0.7 m above the surface. Soil moisture profiles were measured using CS-616 water content reflectometers (CS). θ sensors were installed at all three sites in a representative hummock and hollow at depths of 0.05, 0.15, and 0.25 m relative to the surface, with an additional sensor at 0.5 m in hummocks. WTD was measured hourly in 1.5 m deep wells using self-logging Levellogger Junior pressure transducers [Solinst, Georgetown, ON (Solinst)]. WTD measurements were corrected for changes in atmospheric pressure using a Barologger Gold barometric logger (Solinst). WTD was recorded at all three sites in two separate groundwater wells with the average local topography used as a datum, where the height of the average local surface was determined from 50 m transects with the groundwater wells at the centre.

All high-frequency data was recorded on a CR5000 datalogger (CS) at the WET and DRY site, while a CR3000 datalogger (CS) was used at the INT site. Meteorological, peat temperature and θ data were measured every minute and averaged and recorded at 30 min intervals using a CR1000 (CS) or CR10X (CS) datalogger.

2.3. High-frequency data processing and corrections

Prior to calculating half-hour covariances, high-frequency EC measurements were subjected to a spike detection algorithm analogous to that presented in Vickers and Mahrt (1997), where spikes were identified when a measurement exceeded the recursive mean by a standard deviation of 2.6 (also derived recursively). The mean was constructed as a recursive digital filter (Kaimal and Finnigan, 1994):

$$\tilde{c}_t = \left(1 - \frac{\Delta t}{\tau_f}\right)\tilde{c}_{t-1} + \frac{\Delta t}{\tau_f} c_t \quad (1)$$

where c_t is the measured value at time t, Δt is the incremental time step between measurements (0.1 s), and τ_f is the RC filter time constant (60 s). The cut-off for spike detection is lower than the typical 3–5 SD range reported by others (e.g. Baldocchi et al., 1997; Vickers and Mahrt, 1997; Humphreys et al., 2006) because, given the above time constant, the recursive mean is more responsive to coherent transient departures from the long-term mean compared to block averaging.

Sonic anemometer wind vectors were mathematically rotated based on the tilt correction algorithms presented by Wilczak et al. (2001), also known as the planar fit method. The planar fit method helps to address the problem of over-rotation in sloping terrain associated with the more commonly used method of Tanner and Thurtell (1969) as outlined by Lee et al. (2004).

Before calculating energy and mass fluxes, a time lag was introduced into the appropriate mass or energy time series in order to maximize the average covariance with the rotated vertical wind speed. In addition to flux loss that results from the asynchrony in the measured time series due to finite instrument processing times, spectral transfer functions were used to correct for high-frequency spectral losses that result from sensor separation, line and volume averaging, and digital filtering. Frequency response corrections were calculated according to the analytical solutions presented in Massman (2000) and applied to despiked, rotated, and lagged covariances.

Errors in EC flux measurements associated with variations in air density due to changing temperature and humidity were

Table 1
Summary of microtopography and vegetation characteristics across study sites.

Variable	WET	INT	DRY
Avg. height (m) of hummocks (±SE)	0.15 (0.01)	0.30 (0.02)	0.41 (0.03)
Proportion of hummocks and hollow	0.41/0.59	0.52/0.48	0.63/0.37
pH (±SE)	4.10 (0.06)	3.77 (0.02)	3.69 (0.01)
LAI[a] ($m^2\,m^{-2}$) – *hummocks* (±SE)	2.7 (0.18)	1.5 (0.06)	1.9 (0.19)
LAI[a] ($m^2\,m^{-2}$) – *hollows* (±SE)	2.0 (0.32)	1.9 (0.45)	1.3 (0.04)
Plant Area Index[b] ($m^2\,m^{-2}$)	2.8 (0.52)	1.9 (0.18)	2.6 (0.27)

[a] Leaf area index based on destructive sampling of 0.25 m^2 hummock and hollow plots ($n = 4$).
[b] Plant area index based on gap light photography of hummock and hollow plots ($n = 4$).

corrected based on the method outlined by Webb et al. (1980). The sensible heat flux is used in the WPL correction, and is itself dependent on air density fluctuations when measured using a sonic-anemometer. Sonic temperatures were thus corrected using the method of Kaimal and Finnigan (1994).

2.4. Quality assurance and gap filling

Half-hour flux measurements were rejected based on a number of statistical and physical environmental conditions. Basic statistical criteria for rejection of fluxes were based on second-, third-, and fourth-moment statistics. The thresholds for skewness and kurtosis were based both on those presented by Vickers and Mahrt (1997) and measured empirical probability distributions. A site-dependent friction velocity (u_*) threshold of ~0.08 m s^{-1} was used as a basic rejection criterion for removing measurements made under conditions without well-developed turbulence. Although data was not explicitly rejected during periods of rainfall, data was rejected for approximately 99% of all half-hours with rainfall as a result of the aforementioned statistical criteria.

A comprehensive data quality flagging system was also used to identify half-hours with high-quality, questionable, and bad data (Foken et al., 2004). Data quality was assessed based on integral turbulence characteristics, stationarity (Foken and Wichura, 1996), and wind direction, where wind sectors down-wind of the tower were considered inappropriate.

In order to estimate growing season total ET, missing data was filled using an artificial neural network (ANN). ANNs have been shown to be suitable for gap-filling EC flux data (Moffat et al., 2007) where standard meteorological and soil variables were used as driving variables. Day and nighttime gaps were modelled separately, where the disparity in available data for these two time periods would result in a bias towards daytime conditions during ANN calibration and validation. However, due to the relatively small contribution of nighttime latent heat fluxes, LE (W m^{-2}), the effect of nighttime bias was assumed to be negligible.

2.5. Calculations

A common method to investigate the controls on ET is by using the Penman–Monteith (Monteith, 1973) equation which takes the following form:

$$LE = \frac{s(R_n - G) + \rho_a C_p D r_a^{-1}}{s + \gamma\left(1 + {}^{r_s}\!/_{r_a}\right)} \tag{2}$$

where s (kPa °C^{-1}) is the slope of the saturation vapour pressure versus temperature curve, γ (kPa °C^{-1}) is the psychrometric constant, R_n (W m^{-2}) is the net radiation flux, G (W m^{-2}) is the ground heat flux, ρ_a (kg m^{-3}) is the air density, C_p is the heat capacity of air (J kg^{-1} °C^{-1}), D (kPa) is the vapour pressure of the atmosphere, r_a (s m^{-1}) is the aerodynamic resistance, and r_s (s m^{-1}) is the bulk surface resistance.

Aerodynamic resistance to heat and vapour transport was calculated as the sum of the resistance to momentum transfer, r_{am}, and the boundary layer resistance, $r_b - r_a$ was calculated according to Stewart and Thom (1973):

$$r_a = \frac{u_z}{u_*^2} + r_b \tag{3}$$

where u_z is the wind speed (m s^{-1}), u_* is the friction velocity (m s^{-1}), and r_b is calculated according to the empirical formula of Kellner (2001):

$$r_b = \frac{a\left(u_* z_0/\upsilon\right)^{0.25} - b}{k u_*} \tag{4}$$

where z_0 is the roughness length (m), υ is the kinematic viscosity of air ($m^2\,s^{-1}$), k is the von Karman constant, and a and b are empirical parameters. Estimates of 1.58 and 3.4 for a and b from a Swedish bog were used to calculate r_b (Molder and Kellner, 2001). Stability correction factors were omitted from Eqs. (3) and (4) since u_* was calculated based on turbulent measurements of all three Cartesian wind speed components rather than from wind speed profiles.

Surface roughness was estimated using a direct search algorithm (The MathWorks, R2010a) to minimize a least square difference function based on the log wind profile equation for near-neutral stability:

$$f(d, z_0) = \left(\frac{u_*}{k}\ln\left(\frac{z-d}{z_0}\right) - u_z\right)^2 \tag{5}$$

where d is the zero-plane displacement (m), subject to the constraint that $d > z_0$.

Bulk surface resistance, r_s (s m^{-1}), was calculated by rearranging the Penman–Monteith equation (Monteith, 1973) and substituting the Bowen ratio, $\beta = H/LE$, which yields:

$$r_s = r_a\left(\frac{s}{\gamma}\beta - 1\right) + \frac{\rho C_p D}{R_n - G}\left(1 + \beta\right) \tag{6}$$

In order to determine whether cumulative differences in ET between sites exceeded measurement uncertainty, a simple error analysis was used where the error in high quality data was assumed to be the result of random error. The calculation of random error was adapted from the method of Moncrieff et al. (1996):

$$Err_i = pDV_{\mathrm{ET},0.95}n^{-0.5} \tag{7}$$

where p is the magnitude of random error proportional to the measured flux, $DV_{\mathrm{ET},0.95}$ is the half-hourly diurnal variation of the 95th percentile of ET using a two week moving window, and n is the ith measurement. In order to ensure that non-overlapping error bounds represented meaningful differences in cumulative ET, p was set equal to one. Assuming the true value of p cannot exceed one for high quality data, then non-overlapping error bounds also represent a statistically significant difference in cumulative ET.

Unless otherwise stated, the period of analysis corresponds with the 2010 and 2011 growing seasons. Given that there is no universal definition for the climatological growing season (Walther and

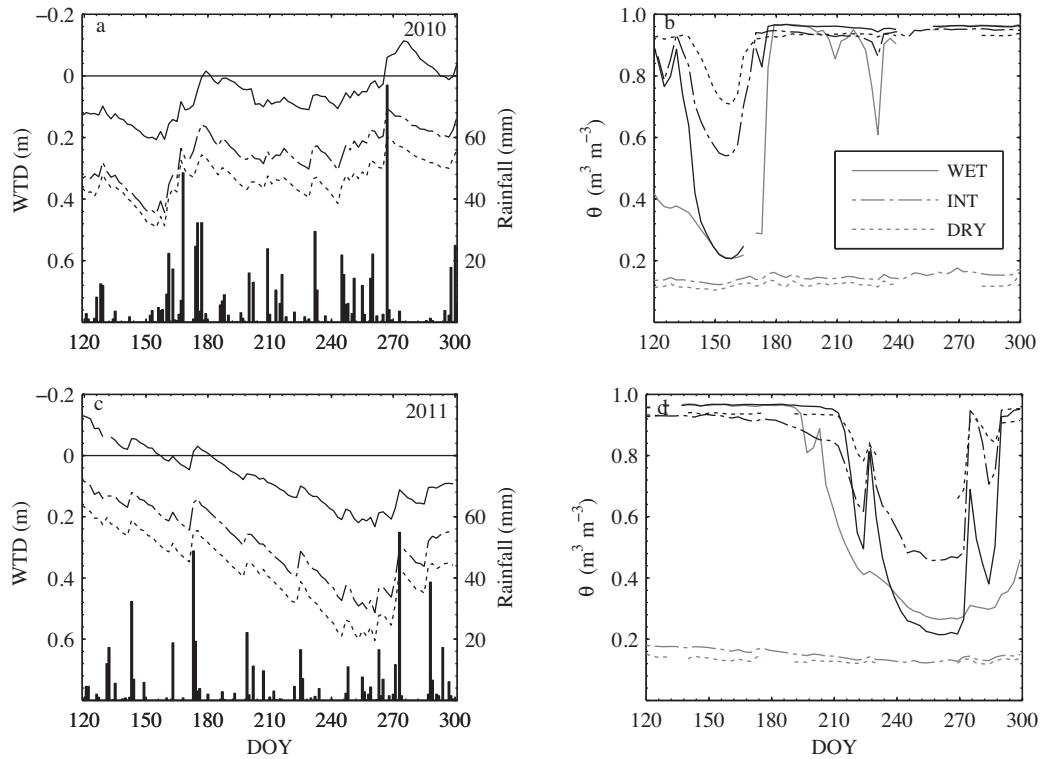

Fig. 1. Daily total rainfall (bars) and water table levels (a and c) for the WET (solid), INT (dash-dot), and DRY (dot) sites. Volumetric water content (three day average shown for clarity) at all three sites (b and d) measured at 0.05 m depth in a representative hummock (grey) and hollow (black) over the 2010 (top) and 2011 (bottom) growing seasons.

Linderholm, 2006), the growing season herein corresponds to the time of year when air temperature (T_a) was above 5 °C for over five consecutive days (Frich et al., 2002).

Finally, all data processing, quality assurance, and statistical tests were done using MATLAB 7.8 (The MathWorks Inc., 2009). Statistical tests were considered significant when the p-value was less than 0.01.

3. Results

3.1. Meteorological and hydrological conditions

The length and timing of the growing season was similar in both years (Table 2). Over the course of the two study years, 2011 was slightly warmer based on a comparison of cumulative GDD (Table 2). GDD for 2010 and 2011 were similar to the historical average of 1629 °C based on climate normals from the Newberry weather station (1981–2010 US Normals Data – National Climatic Data Center – National Oceanic and Atmospheric Administration). Considering GDD during the period defined by the climatological growing season alone, the difference between years becomes negligible. Moreover, during this period, there were no strong seasonal departures in $\underline{T}_a$ from the normal in either study year.

Rainfall during the 2010 and 2011 growing season was higher than normal (430 mm) over the same period (Table 2). The greater seasonal rainfall total in 2010 was the result of both fewer dry days and greater average rainfall on wet days (Table 2). Despite a greater growing season rainfall total, there was an extended dry period in 2010 where there was a 22 day period with only 6.4 mm of rainfall (day of year (DOY) 130–151). The maximum period with no rainfall in 2010 and 2011 was 10 and 12 days, respectively.

The difference in the average WTD (Fig. 1) between sites was fairly consistent over 2010 (median: $WTD_{WET} - WTD_{INT} = 0.20$ m; and $WTD_{WET} - WTD_{DRY} = 0.35$ m) with the exception of some anomalous values where WTD_{WET} was increasing while WTD_{INT} and WTD_{DRY} were decreasing (DOY 266–273). The difference in average WTD was smaller in 2011 (median: $WTD_{WET} - WTD_{INT} = 0.15$ m; and $WTD_{WET} - WTD_{DRY} = 0.28$ m) where the magnitude of the difference between WTD_{WET} and WTD_{DRY} increased as WTD_{DRY} moved into deeper peat late in the growing season. Overall, the average WTD was greater in 2011, which corresponds with a lower seasonal rainfall total. In terms of seasonality, WT was relatively stable in 2010 except early in the growing season during the extended dry period (DOY 130–151). Conversely, 2011 was characterized by a shallow WTD early in the growing season with progressive drying due to fewer and smaller rainfall events.

The effect of WTD variation was not apparent in the near-surface (0.05 m) θ of hummocks at two of the study sites (Fig. 1). Near-surface $\theta_{INT,hum}$ and $\theta_{DRY,hum}$ remained relatively stable throughout the growing season in both years, varying between 0.10 and 0.20 $m^3\ m^{-3}$. Both $\theta_{WET,hum}$ and all θ_{hol} had much greater ranges of variability in both years, loosely corresponding to periods of large WTD variability. The magnitude of variability in θ_{hol} differs between sites. During the drying phase of both study years, θ_{hol} reached a minimum daily value of 0.21 and 0.22 $m^3\ m^{-3}$ at the WET site compared to 0.56 and 0.47 $m^3\ m^{-3}$ for the INT site, and 0.73 $m^3\ m^{-3}$ (2011 data missing) for the DRY site despite a greater WTD at the INT and DRY sites.

3.2. Surface energy exchange

3.2.1. Evapotranspiration

Daily ET at the WET site had a maximum of 5.0 and 4.8 mm for 2010 and 2011, with typical (25th–75th percentile) values ranging from 1.9 to 3.4 mm d^{-1} and from 1.6 to 3.1 mm d^{-1} in each of the

Table 2
Summary of average climatological and flux variables for the study area.

Variable	2010	2011
Climatological growing season	DOY: 136–306	DOY: 130–303
GDD[a] (total/growing season)	1721 °C/1567 °C	1601 °C/1558 °C
Rainfall (total/dry days/avg. depth)	660 mm/96 days/8.2 mm	470 mm/112 days/7.1 mm
Total ET	WET: 1.03 (417)	WET: 0.93 (378)
10^3 MJ m^{-2} (and mm)	INT: 0.80 (323)	INT: 0.84 (340)
	DRY: 0.77 (314)	DRY: 0.82 (336)
Total R_n (±std. dev.[b])	WET: 1.70 (0.04)	WET: 1.82 (0.05)
10^3 MJ m^{-2}	INT: 1.86 (0.04)	INT: 1.92 (0.04)
	DRY: 1.86 (0.04)	DRY: 1.92 (0.05)
Total G (±std. dev.[b])	WET: 0.076 (0.002)	WET: 0.065 (0.008)
10^3 MJ m^{-2}	INT: 0.026 (0.002)	INT: 0.021 (0.006)
	DRY: 0.007 (0.001)	DRY: 0.010 (0.005)

[a] Cumulative growing degree days (GDD) were calculated using the method established by the Atmospheric Environment Service, Environment Canada, using a base temperature of 5 °C.
[b] The sum of square errors is calculated by assuming each half-hourly measurement has no systematic error and a maximum random error equal to the midday flux value.

two growing seasons, respectively (Fig. 2). In addition to having small differences in typical ET values, seasonality was more pronounced in 2011 compared to 2010, particularly due to lower ET values early in the 2011 growing season. Fig. 2 shows that throughout the 2010 growing season daily ET at both the INT and DRY sites was often lower compared to the WET site with a median difference of −0.41 mm d^{-1} for INT and −0.56 mm d^{-1} for DRY. In 2011, the differences between sites were not as pronounced, particularly early and late in the growing season, with values of −0.13 mm d^{-1} and −0.15 mm d^{-1} for INT and DRY, respectively. Total ET over the growing season was substantially higher at the WET site in 2010 (Table 2), with total ET corresponding to 0.63, 0.49, and 0.48 of the total rainfall over the same period. In 2011, ET accounted for a greater proportion of total rainfall at all sites, where total ET corresponded to 0.80, 0.72, and 0.72 of total rainfall.

When only periods of overlapping, high quality, non-gap-filled data are considered, the same general trends exist (Fig. 3). When simple error analysis was applied to the data set [Eq. (7)], we determined that, in 2010, ET at the WET site was significantly greater than the other two sites, whereas ET at the DRY site was less than at the INT site (Fig. 3). While ET at both WET and DRY sites was different from INT, the magnitude of the difference was on the order of five times greater at the WET site. In 2011, the relatively large differences in mid growing season ET between the WET and other two sites (Fig. 2) resulted in significantly different cumulative totals over that period. However, later in the growing season, where daily ET values were similar, the accumulation of random error resulted in there being no net significant difference in the cumulative total ET. Because the fundamental atmospheric controls on ET (i.e., incoming solar radiation and D) were the same between sites, the differences in surface radiative, thermal, and biophysical properties were examined to explain the differences in ET between sites, where significant.

3.2.2. Net radiation

Despite the potential for differences in surface radiative properties between sites, the difference in half-hourly average net radiation at each site was small. In order to investigate the significance of potential differences, a simple residual analysis was carried out for each unique site combination. Based on the Kolmogorov–Smirnov test, all three residual R_n-pairs were non-normally distributed ($p \ll 0.01$; $D = \{0.190, 0.140, 0.215\}$; $n = \{15{,}131, 16{,}320, 16{,}273\}$), where there tended to be a strong clustering of residuals about zero, with relatively infrequent yet large residuals (Fig. 4). Due to the non-normality of the residuals, the Wilcoxon signed rank test was used to determine whether the

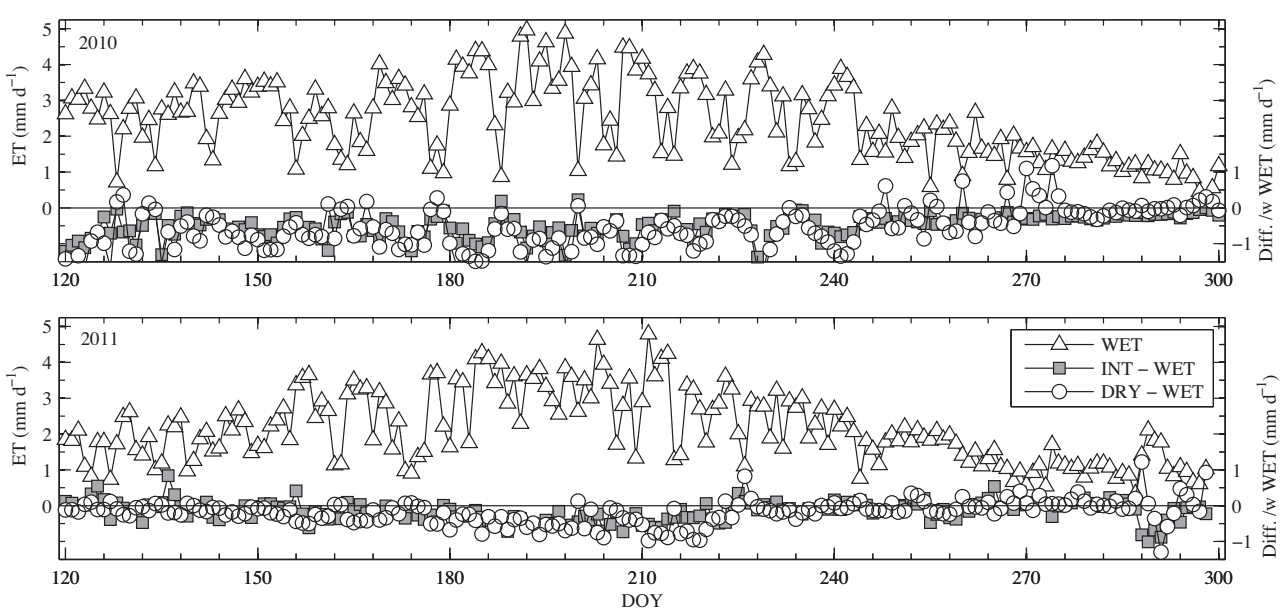

Fig. 2. Daily total gap-filled evapotranspiration at all three study sites over the 2010 and 2011 growing seasons. INT and DRY traces represent the difference in daily evapotranspiration relative to the WET site.

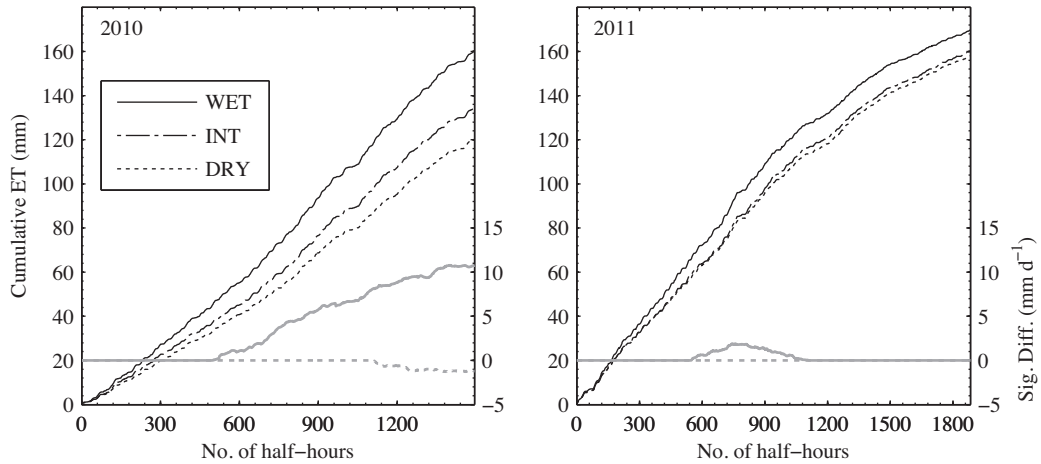

Fig. 3. Cumulative sum of all common half-hourly evapotranspiration measurements (thick black lines) for WET (solid), INT (dash-dot), and WET (dashed) sites where quality control flagging system indicated high quality data based on integral turbulence characteristics, wind direction, stationarity, and a u_* threshold of 0.1 m s^{-1}. Significant differences between sites (thin grey lines) are relative to the INT site and are based on an assumption of purely random error.
Source: Moncrieff et al. (1996).

distributions had a zero median. As a result of a lack of symmetry in the residual distributions, all residual R_n-pairs had medians significantly different from zero ($p \ll 0.01$). However, in absolute terms, the median residuals were relatively small (−5.0, −3.4, 0.49), with corresponding mean absolute deviations of 30.6 W m^{-2}, 28.9 W m^{-2}, and 29.1 W m^{-2} and root mean square errors of 57.4 W m^{-2}, 44.1 W m^{-2}, and 57.7 W m^{-2} (WET–INT; WET–DRY; and INT–DRY). The results indicate that the median half-hourly $R_{n,\,\mathrm{wet}}$ was less than the other two sites.

It is important to note that these differences are based on half-hourly measured values which are susceptible to small variations in local apparent cloud cover. Differences in cumulative R_n were tested by assuming only random error in R_n measurements, where the sum of square errors was generated from the estimated random error associated with each measurement (*cf.* Moncrieff et al., 1996). A conservative random error estimate equal to the midday flux value was used. If we treat the cumulative sum as a mean, when summed over the growing season there was no significant difference in cumulative R_n between the INT and DRY sites (Table 2), while $R_{n,\mathrm{wet}}$ was only significantly less than the other sites in 2010 (2010: $p = 0.03$, $n = 8303$; 2011: $p = 0.16$, $n = 8039$). Integrated across daily or larger time scales, we can assume the differences in incoming solar radiation and the radiative sky temperature are negligible given the close proximity of the three study sites. As such, the differences in net radiation between sites are most likely attributed to surface albedo and the radiative temperature of the surface.

3.2.3. Albedo

Due to the strong variation of albedo with the angle of incident solar radiation, average values were calculated based on measurements within 2 h of solar noon. During the 2011 growing season, daily variability of albedo was relatively small at the WET site, where a significant positive trend ($p \ll 0.01$, $n = 148$) was evident throughout the year (Fig. 5). In contrast, both INT and DRY had greater short-term variability, where a small negative and non-significant trend was apparent over the growing season for INT ($p = 0.002$, $n = 140$) and DRY ($p = 0.04$, $n = 148$) respectively. Due to

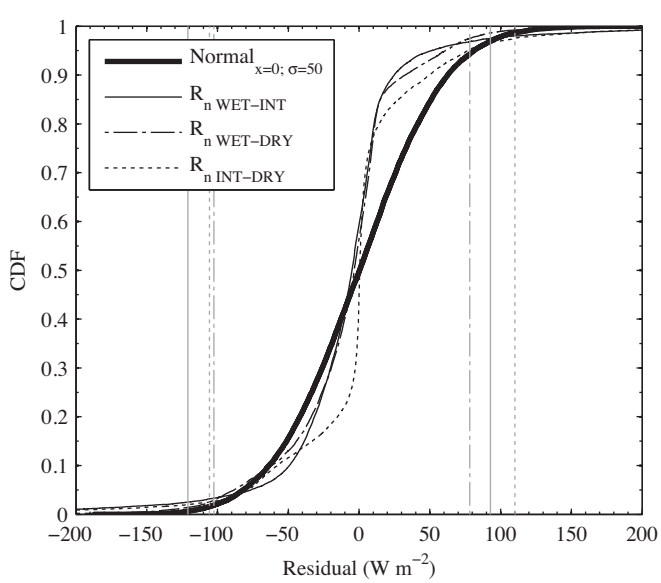

Fig. 4. Empirical cumulative distribution functions of residual half-hourly R_n values between all sites from DOY = 120–300, for 2010 and 2011. Vertical bars represent the 2.5th and 97.5th percentiles.

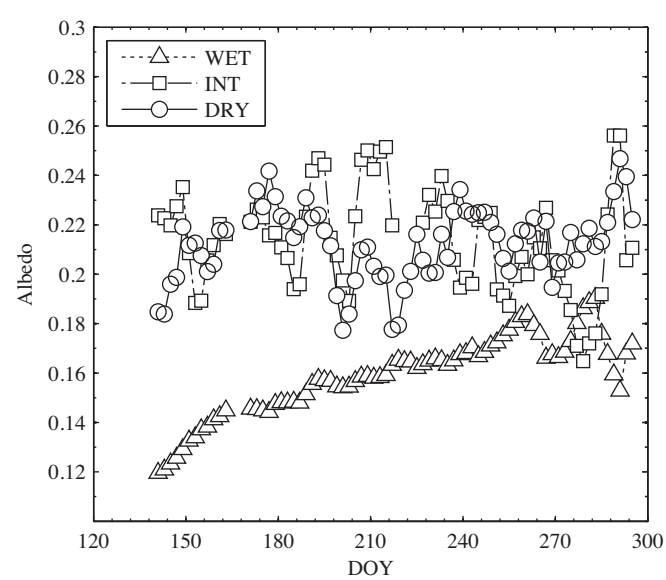

Fig. 5. 7-day running average midday albedo across all study sites during the 2011 growing season.

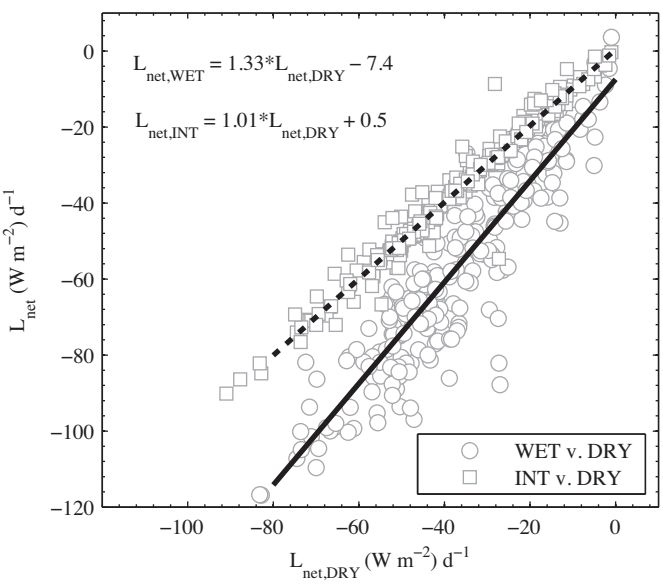

Fig. 6. Correlation of daily average long-wave radiation against the DRY site for 2010 and 2011 growing season data combined. The linear relation between sites is represented by the solid (WET vs. DRY) and dashed (INT vs. DRY) lines. For clarity, the relation between WET and INT has been omitted due to the close correspondence between INT and DRY.

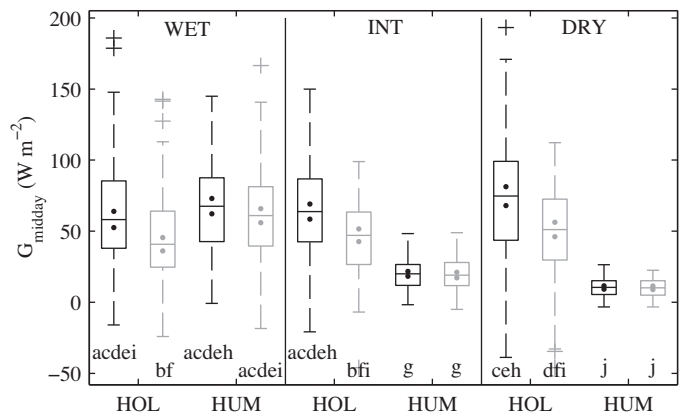

Fig. 7. Boxplots comparing daytime ground heat flux for both 2010 (black) and 2011 (grey). Standard error estimate of the median is represented by the solid dots, and outliers by the crosses. Letters bellow each boxplot indicate where there are significant differences in the median.

mid-season equipment problems with the downwelling shortwave radiometer in 2010, full growing season albedo was not available. However, average midday albedo at the WET site was 0.21 and 0.15 in the early and late growing season of 2010, compared to 0.13 and 0.17 in 2011. The change in albedo over the same period was small for both the INT (2010: 0.23 and 0.21; 2011: 0.21 and 0.21) and DRY sites (2010: 0.22 and 0.19; 2011: 0.20 and 0.21).

3.2.4. Long-wave radiation

All else being equal, a simple radiation balance approach suggests that a lower average albedo, such as that of the WET site, will be balanced by a higher surface temperature, and thus greater outgoing long-wave radiation. The magnitude of the daily average net long-wave radiation (L_{net}) was comparatively high at the WET site, such that the slope of the relation between $L_{net,WET}$ and $L_{net,DRY}$ ($p \ll 0.01$, slope $\pm$SE = 1.33 ± 0.034, n = 275) was significantly greater than a 1:1 ratio (Fig. 6). The degree of correlation of $L_{net,DRY}$ with $L_{net,WET}$ was only somewhat lower (R^2 = 0.841) than with $L_{net,INT}$ (R^2 = 0.966). In addition to the strong linear relation between $L_{net,INT}$ and $L_{net,DRY}$ over both growing seasons, the slope of the relation (p = 0.3034, slope $\pm$SE = 1.01 ± 0.006, n = 276) was not significantly different from 1:1. These results lump both day and nighttime measurements, as well as any changes throughout the season. Qualitatively, in both study years there was no strong seasonal trend in the difference in L_{net} between the INT and DRY sites, nor were there frequent large half-hourly differences (RMSE = 7.5 W m^{-2}, MAD = 4.2 W m^{-2}). The magnitude of half-hourly differences between $L_{net,WET}$ and $L_{net,DRY}$ were greater (RMSE = 35.1 W m^{-2}, MAD = 25.5 W m^{-2}), but the difference was largely attributable to daytime (RMSE = 50.1 W m^{-2}, MAD = 38.7 W m^{-2}) versus nighttime measurements (RMSE = 23.3 W m^{-2}, MAD = 18.1 W m^{-2}). Based on the Stefan–Boltzman law, differences in L_{net} between sites are related to the surface temperature given their close proximity to one another. Although the surface radiative temperature integrates both the canopy and peat surface temperature, in the absence of canopy temperature measurements, differences in peat temperature measured at a depth of 0.01 m provide a qualitative indirect comparison. For simplicity, we focus here on the differences in near-surface temperature between the WET and DRY sites. Again, the difference in half-hourly values was separated by day (K_{ex} > 100 W m^{-2}) and nighttime (K_{ex} < 10 W m^{-2}), where K_{ex} is the modelled clear-sky incoming short-wave radiation over the growing season. WET hummock temperatures were higher compared to the DRY site with a mean daytime and nighttime difference ($\pm$STDV) of 3.0 ± 2.7 °C and 5.3 ± 2.3 °C, respectively in 2010, and similarly for 2011: 4.5 ± 2.1 °C; and 4.1 ± 3.1 °C. In comparison to the mean difference in hummock temperatures, the difference in near-surface hollow temperatures between the WET and DRY sites tended to be smaller during both the day (2010: 1.9 ± 2.2 °C; and 2011: 0.7 ± 2.1 °C) and night (2010: 0.0 ± 2.0 °C; and 2011: 0.7 ± 1.7 °C).

3.2.5. Ground heat flux

Differences in peat thermal properties between sites and microforms associated with θ can have a strong influence on ground heat flux (G) where, for example, higher average near-surface θ at the WET site should lead to higher thermal admittance and a greater partitioning to G. As a result of the relatively small microtopographic variation and WTD, near-surface $G_{WET,hum}$ was consistently greater than $G_{INT,hum}$ and $G_{DRY,hum}$, with values similar to G_{hol}. As a result, average midday G_{WET} was higher compared to the other sites (Fig. 7). $G_{INT,hum}$ and $G_{DRY,hum}$ were numerically similar since they were consistently dry throughout both study years, yet they were significantly different as a result of near-surface $\theta_{DRY,hum}$ being consistently lower while having similar temperature profiles. Overall, when the proportion of hummocks and hollows was factored in, a greater proportion of R_n was partitioned to G_{INT} than G_{DRY}. Overall, the energy available (R_a) for turbulent fluxes is the remainder of R_n which is not consumed by G (Table 2). The resulting cumulative daytime R_a over the growing season was 1.63 × 10^3 MJ, 1.83 × 10^3 MJ, and 1.86 × 10^3 MJ for the WET, INT, and DRY sites over 2010. Similarly, the corresponding values for 2011 were 1.77 × 10^3 MJ, 1.90 × 10^3 MJ, and 1.92 × 10^3 MJ. These results indicate that in both study years, there was less available energy for turbulent fluxes at the WET site. Therefore, in order to have higher ET at the WET site, partitioning to LE must be disproportionally greater than the relative differences in R_a between sites.

3.2.6. Energy partitioning

There was a relatively large range of variability in daily energy partitioning across all sites in both study years, where the ratio of H to LE (β) was calculated based on the sum of daytime flux values.

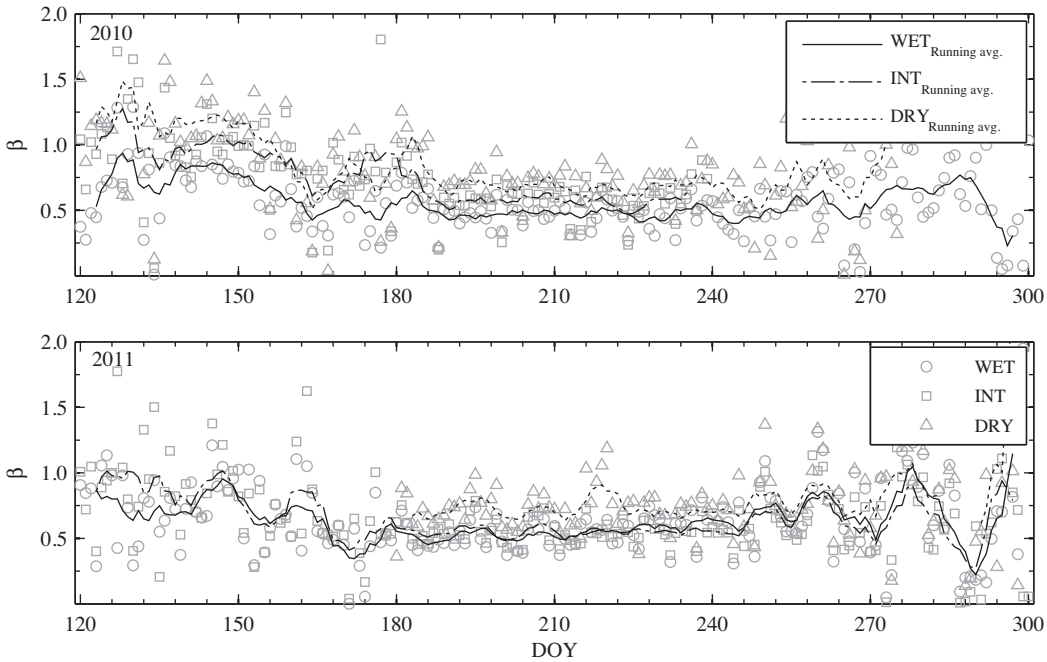

Fig. 8. Daily Bowen ratio for all three sites over the 2010 and 2011 growing seasons. Lines represent the 7-day running average Bowen ratio for WET (solid), INT (dash-dot), and DRY (dashed) sites.

Using a 7-day running average, distinct trends emerge from the data (Fig. 8). In 2010, β was fairly distinct at all three sites where the WET site tended to have the lowest average β, whereas the DRY site had the highest average β, with the INT site taking on intermediate values. However, in 2011, there was a much closer correspondence between the running average β at the WET and INT sites. While, β ratio values were not available during the early 2011 growing season at the DRY site due to instrument failure, during the mid growing season, β tended to be higher at the DRY site, with average values becoming more similar towards the end of the growing season. Independent of the differences between sites, there was a common seasonal trend in 2010, where β values tended to decrease towards mid growing season and then remained relatively stable. A muted version of the same early season trend occurs in 2011, where average mid to late growing season β value increased progressively, though only by a small amount.

3.3. Controls on evaporation

3.3.1. Resistance

Under near-neutral stability, the linear relation (±SE) between u_*/k and u_z [Eq. (5)] yielded the following estimates of z_0 for 2010: $\mathrm{WET}_{z_0} = 0.052 \pm 0.003$; $\mathrm{INT}_{z_0} = 0.065 \pm 0.027$; $\mathrm{DRY}_{z_0} = 0.11 \pm 0.01$ and 2011: $\mathrm{WET}_{z_0} = 0.062 \pm 0.009$; $\mathrm{INT}_{z_0} = 0.078 \pm 0.013$; and $\mathrm{DRY}_{z_0} = 0.079 \pm 0.06$. In addition to the data on average microtopographical variation, estimates of z_0 suggest that the WET site was less aerodynamically rough compared to the other sites. All else being equal, this would imply greater aerodynamic resistance to evaporative losses at the WET site. However, when the distributions of measured wind speed are considered, the average wind speed value was ~26% higher at the WET site compared to the other sites in both study years. As a result of the negative correlation between wind speed and r_a, median midday $r_{a,\mathrm{WET}}$ was often significantly less than $r_{a,\mathrm{INT}}$ and $r_{a,\mathrm{DRY}}$ ($p \ll 0.01$) as a result of small standard error estimates, with the exception of $r_{a,\mathrm{WET}}$ and $r_{a,\mathrm{INT}}$ in 2010 ($p = 0.08$). Nevertheless, median daytime values within sites and between years were fairly similar ($r_{a,\mathrm{WET}}$ – 55.5 s m^{-1}, 54.2 s m^{-1}; $r_{am,\mathrm{INT}}$ – 56.5 s m^{-1}, 52.2 s m^{-1}; and $r_{am,\mathrm{DRY}}$ – 57.5 s m^{-1}, 58.2 s m^{-1}).

Despite the differences in r_a being statistically significant, the small absolute difference between sites and years would have a small effect on differences in total ET. The variability in r_s between sites, however, tended to be greater compared to r_a. Overall, if we consider median daytime r_s values between sites and years, $r_{s,\mathrm{WET}}$ was 42.3 s m^{-1} and 44.3 s m^{-1}, $r_{s,\mathrm{INT}}$ was 55.2 s m^{-1} and 42.3 s m^{-1}, and $r_{s,\mathrm{DRY}}$ was 87.3 s m^{-1} and 74.3 s m^{-1} for 2010 and 2011, respectively. Qualitatively, r_s at all sites trended downwards during the 2010 growing season, while only the WET site varied seasonally in 2011, with a slight increase towards the end of the growing season. Using a simple linear regression, all negative trends in 2010 were significantly different from zero ($p \ll 0.01$). In 2011 only $r_{s,\mathrm{WET}}$ had a slope significantly different from zero ($p \ll 0.01$), whereas $r_{s,\mathrm{INT}}$ and $r_{s,\mathrm{DRY}}$ were not ($\rho_{\mathrm{INT}} = 0.449$ and $\rho_{\mathrm{DRY}} = 0.02$). A Jarvis-type phenomenological model for stomatal resistance r_{st} is typically written as a function of solar radiation, D, and some metric of water availability. Given an appropriate scaling factor for LAI, r_s is related to r_{st} and peat surface resistance. Since incoming solar radiation does not vary greatly between sites, the relation between r_s and D was investigated using equation [Eq. (6)]. Comparing the two growing seasons, the relation between r_s and D only changed appreciably at the WET site where bin-averaged r_s values shifted higher for most D classes in 2011 (Fig. 9). In both growing seasons, bin-averaged r_s values were often highest at the DRY site, where based on the standard error estimates, the difference was only significant between the WET and DRY sites in 2010. Overall, the difference between the WET and INT sites was less distinct, where only r_s in low D classes were significantly different in 2010. The D–r_s relation only tends to increase at low D and remains relatively constant over a range of D classes. The relative insensitivity of r_s to D would suggest that the sites are decoupled from atmospheric demand. The Jarvis–McNaughton decoupling coefficient (Ω) can be used to investigate the degree to which ET is linked to D through strong turbulent mixing (low Ω values) or R_a (high Ω values). Broadly, all three sites have high Ω-values in both study years, where median Ω_{WET} was 0.83 and 0.79, Ω_{INT} was 0.76 and 0.80, and Ω_{DRY} was 0.71

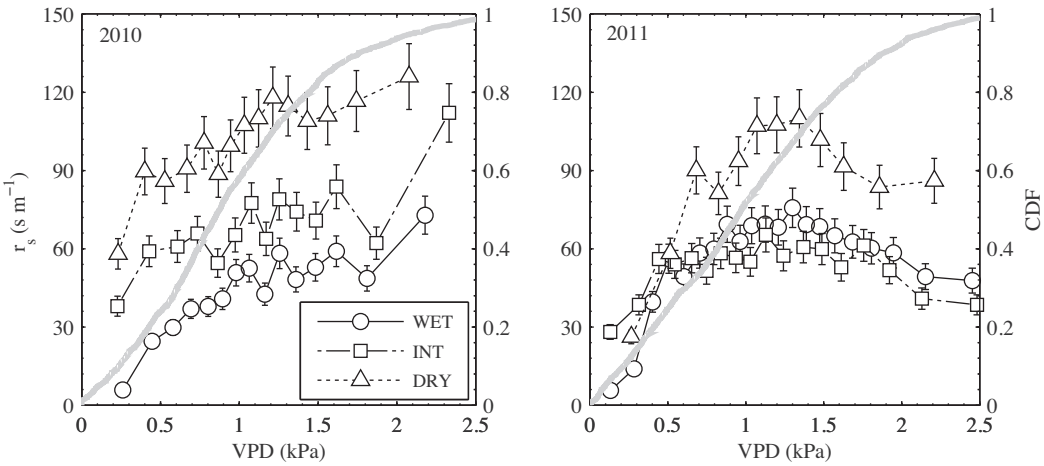

Fig. 9. Median half-hourly daytime (K_{dwn} > 100 W m^{-2}) bulk surface resistance for 2010 (a) and 2011 (b) binned based on vapour pressure deficit with 100 values per bin. Error bars represent the standard error estimate of the median. The grey line represents the cumulative distribution of *D*.

and 0.73. The sites exhibit variable seasonal trends in Ω, where at the WET site Ω tended to increase early in the growing season to a mid-summer high. In 2010, Ω remained high until the end of the growing season, while in 2011 Ω followed more of a parabolic trend by decreasing gradually from the mid-summer high. The trend in Ω for both INT and DRY sites in both study years more closely resembled that of the WET site in 2011, with a shallow seasonal parabolic trend.

3.3.2. Water table depth

Because WTD integrates the variable behaviour of hummocks versus hollows, and *Sphagnum* versus vascular vegetation it was chosen as a simple metric to investigate the effect of moisture availability on ET. ET was first normalized by potential ET (PET) so that the seasonal effect of changing R_a was removed (Fig. 10). Overall given the magnitude of standard error estimates, ET/PET did not tend to vary significantly over the range of measured WTD for all sites over both growing seasons. Qualitatively, ET/PET appeared to be more responsive to WTD class in 2010, where values decreased slightly when WTD was below the average peat surface. In 2011, however, there was no apparent trend with WTD. In 2010, each site had a distinct modal peak in the WTD distribution towards the respective high ends on the WTD distributions, whereas the WTD distributions were relatively broad and flat in 2011. In 2010, this limited direct comparison of the ET/PET vs. WTD relation over the same WTD classes. In 2011, comparing the region of overlap between the WET and INT sites, ET/PET tended to be lower for INT. The same relation was investigated but with wet canopy conditions removed, where the vascular canopy was assumed to be wet during and for several hours following rainfall events. The same general patterns exist for dry canopy conditions in both study years.

4. Discussion

4.1. Evapotranspiration

A direct comparison of peatland ET rates between studies is problematic due to differences in hydrology (e.g. fen vs. bog), vegetation type (e.g. sedge vs. *Sphagnum*-dominated or treed vs. open), growing season R_n associated with latitude, and prevailing synoptic weather conditions. In a mesocosm study, total ET from fen plots was found to be greater compared to bog plots over the snow-free season (Bridgham et al., 1999), where the difference was largely associated with the lack of *Sphagnum* cover in the fen plots and low early-season ET. This is in contrast to our study sites which all have near-total groundcover by *Sphagnum*. Synoptic weather conditions can play an important role where, for example, coastal peatlands tend to have relatively low ET regardless of WTD due to small vapour gradients associated with the low *D* of offshore winds (Sottocornola and Kiely, 2010; Liljedahl et al., 2011). Thus, comparing *Sphagnum*-dominated peatlands of a similar latitude and general synoptic weather conditions, maximum ET at our sites were similar to a northern Minnesota open fen (4.8 mm d^{-1} – Kim and Verma, 1996) and a southern Ontario bog (4–5 mm d^{-1} – Lafleur et al., 2005).

In peatlands, R_n is often the dominant control on ET; however, there are physiological limitations to transpiration imposed by vascular vegetation due to the response of r_s to *D* (Humphreys et al.,

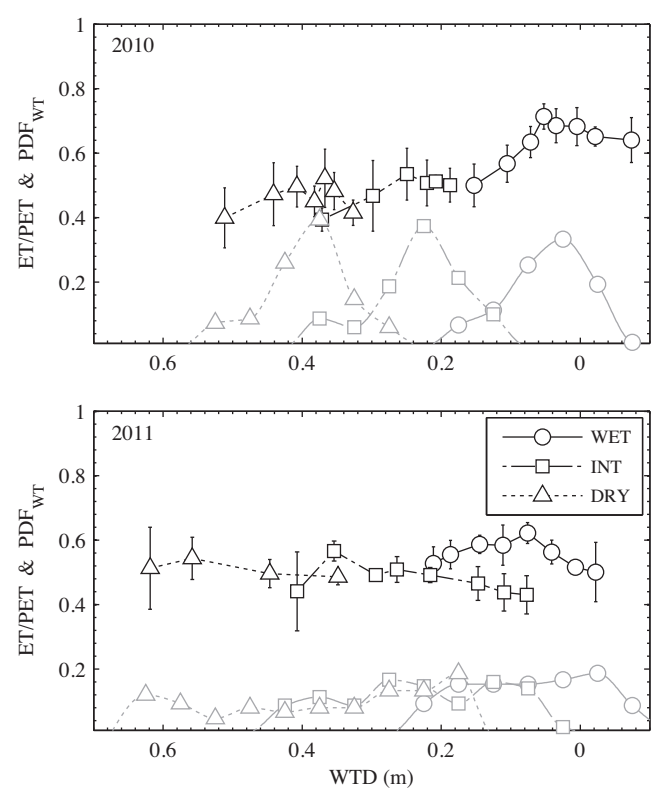

Fig. 10. Relation between normalized ET and bin-averaged WTD for 2010 and 2011 growing seasons. Grey curves represent the relative frequency distribution of WTD.

2006) particularly as vascular LAI and canopy height increases. A strong positive correlation between ET and R_n has been shown on both a daily (Sottocornola and Kiely, 2010; Wu et al., 2010) and seasonal (Brummer et al., 2012) timescale for peatland sites. Moreover, Sonnentag et al. (2010) demonstrated that R_n explained twice the variance in ET compared to D at a minerotrophic fen. Given our high Ω values, the strong control of R_n on ET was also evident at our study sites. However, because the WET site had lower R_n compared to the other two sites, and D was the same across our sites due to their close proximity to one another, these controls on ET cannot explain the differences observed, particularly during the 2010 growing season.

4.2. Albedo and surface temperature

Although R_n is similar between sites, differences in surface albedo which is linked to surface wetness may affect surface temperature, and thus may help to explain the differences in ET between our study sites. In the absence of disturbance, surface albedo at a northern boreal fen has been shown to vary in response to leaf phenology, ranging between 0.09 and 0.18 during the growing season (Lafleur et al., 1997). A similar growing season trend is reported by Kellner (2001) for a Swedish bog, but with albedo being on average 0.01 lower on wet versus dry days. Moreover, a qualitative relation between albedo and peat wetness was suggested for another bog in Central Sweden where daily albedo was shown to gradually increase over the growing season from 0.12 to 0.17 as the peat progressively dried (Phersson and Pattersson, 1997). Although plant phenology could play a role in the seasonal development of albedo at the WET site, 2010 albedo data suggests WTD is a more important factor. In particular, due to the relatively low albedo of open water (Oke, 1987), ponding in hollows would result in a lower albedo at the WET site; where ponding area for our sites could be estimated from WTD and the proportion of hummocks and hollows. Average early and late 2010 growing season albedo values coincided with a WTD trend that was initially low and remained high during the mid and late growing season. In 2011, the growing season trend in albedo at the WET site qualitatively corresponds with a decreasing WTD, where early season WTD was above the mean peat surface. Due to the large proportion of the WET site covered by hollows, a high WTD would coincide with significant surface ponding. This is in contrast to the INT and DRY sites were the WTD was consistently below the mean peat surface throughout the study period, and where hollows represent a decreasing proportion of the surface. There was no trend in surface albedo at the INT or DRY site that could be clearly linked to vascular plant phenology or WTD, where the higher average albedo at the INT and DRY sites is consistent with a lack of ponding at these two sites. Despite similar covariance in albedo at the INT and DRY sites, there was no definitive correlation with surface wetness or rainfall events. Although there was no apparent trend in albedo at the INT or DRY sites associated with WTD, the relatively stable $\theta_{-0.05\,\mathrm{m}}$ for hummocks suggests that *Sphagnum* was not desiccating, thus there was likely no whitening response as has been observed by other studies (McNeil and Waddington, 2003).

Although a lower albedo does not necessarily lead to greater available energy for turbulent fluxes, all else being equal, a lower albedo will result in a higher surface temperature. If we assume that the difference in average incoming long-wave radiation is negligible between sites when integrated over a daily time step, we can infer that differences in daily net long-wave radiation can be attributed to differences in outgoing long-wave radiation (i.e. surface temperature). The lower average daytime net long-wave radiation at the WET site implies a higher integrated surface temperature. Higher near-surface (0.01 m depth) peat temperatures were also measured in both hummocks and hollows at the WET site which would help contribute to the apparent higher surface radiative temperature at the WET site. The degree and importance of surface heating on energy partitioning depends on the thermal conductivity of the surface and the LAI of the vascular canopy. For a saturated high porosity surface, such as in peatlands, a relatively large portion of additional energy associated with a lower albedo will be used in heating the soil profile due to the high thermal admittance of water versus air, thus increasing potential evaporation due to the effect of temperature on the saturation vapour pressure. Due to the dependence of G on near-surface θ, all else being equal, peatlands with a greater proportion of hummocks, or conversely, peatlands with greater microtopographic variation will tend to have lower average G. At both the INT and DRY sites, the effect of the contrast in surface wetness is apparent in the significant difference in median values ($\alpha = 0.05$) between G_{INT} and G_{hol} (Fig. 7). Kellner (2001) similarly showed higher G as a proportion of R_n in hollows versus hummocks. In comparison, the small average microtopographic variation between hummocks and hollows at the WET site lead to no significant difference in G due to similar high θ in both microforms. When combined with a slightly lower R_n, the higher average G of the WET site leads to lower available energy for turbulent fluxes at the WET site.

4.3. Energy partitioning

Differences in energy partitioning between the study sites cannot be explained by higher average midday near-surface temperature at the WET site alone using a 'big-leaf' model. Assuming the surface vapour pressure is near saturation, all else being equal, a higher surface temperature would increase β in favour of the sensible heat flux. Despite low values of $\theta_{-0.05\,\mathrm{m}}$, the assumption of near-saturated vapour pressure at the surface may not be unreasonable for hummocks at the INT and DRY sites. Over the course of the growing season, daily $\theta_{\mathrm{INT,hum},-0.05}$ and $\theta_{\mathrm{DRY,hum},-0.05}$ did not change substantially at the daily time step in response to evaporative losses. Furthermore, the decrease in $\theta_{\mathrm{INT,hum},-0.05}$ and $\theta_{\mathrm{DRY,hum},-0.05}$ in 2011 with an increasing WTD was only slight. This suggests that either: (i) a negligible amount of water is being lost from the moss surface; (ii) capillary transport is able to replenish evaporative losses at a daily time step; or (iii) an extreme hydrological gradient exists between the moss surface and the $\theta_{-0.05\,\mathrm{m}}$ measurement level. Differences in energy partitioning can perhaps be better explained by dual-source models where the lower LAI at the INT and DRY sites limit transpiration in comparison to the WET site, while increasing the relative contribution of moss surface evaporation. Evaporative losses from the *Sphagnum* surface have been shown to be relatively high for an LAI of 0.4–0.7 $\mathrm{m^2\,m^{-2}}$ (Kim and Verma, 1996), but the contribution of moss evaporation to total ET is reduced when surrounded by vascular vegetation (Heijmans et al., 2004) where our LAI values are much higher than those reported by Kim and Verma (1996). Moreover, despite the potential small contribution from the moss surface given our relatively high LAI values, land surface schemes such as CLM include a moisture-dependent empirical term to limit surface evaporation which is meant to be a surrogate for molecular diffusion in the unsaturated zone (Lawrence et al., 2011). Thus, we would expect that the greater proportion of hummocks and low near-surface θ at the INT and DRY sites would serve to further limit total moss evaporation, and thus measured ET.

4.4. Aerodynamic and surface resistance

In general, our z_0 values are higher than the range of values of 0.02–0.03 reported for a Swedish bog (Kellner, 2001), and are at the higher end of the range reported for a Siberian bog (Shimoyama et al., 2004). In a west Siberian bog, a qualitative relation was shown

to exist between vascular development and increasing z_0 which, all else being equal, would enhance ET (Shimoyama et al., 2004). However, because the average microtopographic variation is different between our sites, it is difficult to separate the effect of vascular vegetation and microtopography on z_0. Nevertheless, the difference in the magnitude of z_0 is proportional to the difference in microtopographic variation. The lower surface roughness of the WET site corresponds with r_a values which are higher across wind speed classes. In the absence of any other resistance term, the higher r_a would lead to lower ET at the WET site. However, in addition to the shading effect of vascular vegetation, a higher LAI would increase resistance to moss evaporation by reducing near-surface turbulent transport (Campbell and Williamson, 1997; Heijmans et al., 2001), thus counteracting increases in transpiration.

Although it has been suggested that the similarity in ET between peatland sites is the result of similarity in r_s–D relations despite differences in vegetation types (Humphreys et al., 2006), our results suggest that the difference in peatland ET between sites is the result of the variable response of r_s to D despite having the same vegetation community composition. Again, the dependence of r_s on D using Eq. (6) should be noted. Since there are no large differences in other component of the combination equation, r_s is forced to correlate with difference in ET between sites. Nevertheless, a number of peatland studies show a strong response of r_s to D at low D, whereas r_s tends to be less variable at higher D. Our results show a similar pattern where the average r_s only varies significantly below a daily average D of 0.5 kPa. The difference in average r_s may be related to differences in stomatal resistances, which have been shown to vary between plant functional groups (Korner et al., 1979). While each site has the same basic community composition, the correlation between microtopography and plant functional group results in a greater proportion of graminoids at the WET site and shrubs at the INT and DRY sites. Admiral and Lafleur (2007) show that shrub species on a raised bog generally have higher stomatal resistance compared to other wetland species. A higher stomatal resistance combined with lower LAI at the INT and DRY sites would have a compound effect in reducing r_s. In addition to the potential effect of variable stomatal resistance between plant function groups, the r_s–D relation does not account for any temporal relation between WTD, LAI, and D. D has a strong seasonal progression with low average daily values in the winter and high values in mid-summer due to the non-linear positive correlation with temperature. Therefore, if water availability influences r_s, then it is not simply the degree of water availability that will affect r_s, but also the temporal sequence of variability due to the regular seasonal progression of D.

4.5. Evapotranspiration and water table position

The relation between WTD and ET in peatlands was supported by earlier research (Ingram, 1983), and has been a continuing area of study, in part, due to the relative complexity of modelling physical peat hydrology. However, due in part to the non-linear relation between WTD and θ (Kellner and Halldin, 2002), and heterogeneity in peat surface dynamics (Waddington et al., 2010), there remains no definitive link between ET and WTD in peatlands. For example, a number of studies have demonstrated that there is no significant relation between ET and WTD when ET is normalized by PET or R_n (Shimoyama et al., 2004; Humphreys et al., 2006; Paramentier et al., 2009; Sottocornola and Kiely, 2010; Wu et al., 2010). However, some of these relations were developed for a relatively shallow WTD (i.e. Sottocornola and Kiely, 2010; Wu et al., 2010) while others were in peatlands with near-surface θ retention that was uncharacteristically high (i.e. Paramentier et al., 2009) compared to undisturbed *Sphagnum*-dominated peatlands. Other studies, however, have shown a weak to moderate relation between ET and WTD (Kim and Verma, 1996; Lafleur et al., 2005; Sonnentag et al., 2010), where for a deep WTD, a threshold response may exist which corresponds to the rooting depth of woody vegetation.

In general, our results suggest there was no simple univariate relation between ET/PET and WTD. For the WET site, ET/PET was greater when WTD was above the mean peat surface, but only in 2010. This implies that either WTD (as a proxy for water availability) is only a short-term control on ET that depends on when drying occurs within a growing season, or that WTD was simply better correlated with another ET control in 2010. Given the length of time since disturbance, we would expect that the rooting depth at our sites would be adapted to the average or typical range of WTD (cf. Mitsch and Grosselink, 2000). Consequently, the threshold response of ET/PET to WTD reported by Lafleur et al. (2005) may imply that a threshold response of ET/PET to WTD is predicated upon large inter-annual differences in WTD, rather than due to longer term persistent dry conditions. Under dry conditions, vascular vegetation will respond with increased fine-root production at depth, and a greater proportion of resources allocated to biomass production in stems versus leaves (Murphy and Moore, 2010). Greater fine root production at depth suggests the mechanism which limits any relation between ET/PET and WTD for peatlands with high LAI. Moreover, the lower LAI at the INT and DRY sites, as well as the relatively high PAI:LAI ratio at the DRY site could be the result of past changes in resource allocation between woody and herbaceous structures in response to drying. Therefore the WTD may have an indirect control on ET at our sites through long-term changes in LAI.

5. Conclusions

Overall, despite large differences in WTD, all sites experienced similar energy partitioning with an average β ratio less than one. Significant differences in albedo were likely the result of the timing of shallow WTD (i.e. ponding), but where lower albedo at the WET site was balanced by greater long-wave losses due in part to high nighttime surface temperatures, particularly in hollows. Given the similarity in ET between sites, particularly in 2011, and the relative insensitivity of ET/PET to WTD fluctuations, one can infer a degree of resilience in peatland ecosystems to long-term drying under projected future climate change. Despite the relative insensitivity of r_s to the typical midday range in average D and high Ω associated with the relatively low z_0 of peatland microtopography and short vascular plants, higher D will result in greater ET loss from peatlands and lead to an increased WTD (Roulet et al., 1992). This, however, ignores potential long-term responses of microtopography, shifts in the relative abundance of plant functional groups, or changes in LAI. From the standpoint of vascular vegetation, neither short-term changes in WTD nor near-surface θ are likely to systematically limit transpiration due to the long-term response of rooting depth and fine root growth to average WTD, but where long-term changes in WTD will affect maximum ET through the impact on LAI.

In comparing ET from the WET to INT and DRY sites, our results suggest that the potential long-term effect of WTD on LAI should be incorporated into future land surface schemes since this appears to be one of the primary mechanisms which serves to limit ET. Moreover, microtopography, particularly as it correlates to plant functional groups, needs to be included since, in addition to LAI, differences in stomatal resistance provides the mechanism for explaining difference in r_s where LAI is relatively high. Despite our relatively large LAI values, the results of our study imply a need for a greater understanding of the limitations on water transport processes in the moss layer, particularly as it relates to moss resistance, where we observed that dry hummocks were able to avoid complete desiccation despite low WTD and high daily ET. Vertically-stratified representations of different peatland

vegetation layers in process ecosystem models (e.g., Sonnentag et al., 2008; Lawrence et al., 2011) can help to elicit whether moss evaporation is limited by vascular canopy properties or moisture supply from the underlying peat, where alternative empirical formulations can be used to better understand moss resistance.

Acknowledgements

This research was funded by a Discovery Grant and Discovery Accelerator Supplement from the Natural Science and Engineering Research council of Canada to JMW and by the U.S. Department of Agriculture, grant no. 0220341 to TGP. Additional financial support was provided by the U.S. Department of Energy's Office of Science (BER) through the Midwestern Regional Center of the National Institute for Climatic Change Research at Michigan Technological University. We would like to thank the U.S. Fish and Wildlife Service for permitting access to Seney National Wildlife Refuge, and the staff at SNWR for their kind accommodation. We also thank the two anonymous reviewers who provided many useful comments and valuable suggestions which helped to improve the manuscript. Finally, we would like to thank John Hribljan for his invaluable site selection, as well as Jenna Falk and Reyna Peters for their field assistance.

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