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Water balance of a burned and unburned forested boreal peatland

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Abstract:

We examined the water balance of a forested ombrotrophic peatland and adjacent burned peatland in the boreal plain of western Canada over a 3-year period. Complete combustion of foliage and fine branches dramatically increased shortwave radiation inputs to the peat surface while halting all tree transpiration at the burned site. End-of-winter snowpack was 7–25% higher at the burned site likely due to decreased ablation from the tree canopy at the unburned site. Shrub regrowth at the burned site was rapid post-fire, and shading by the shrub canopy in the burned site approached that of the unburned site within 3 years after fire. Site-averaged surface resistance to evaporation was not different between sites, though surface resistance in hollows was lower in the burned site. Water loss at both burned and unburned sites is largely driven by surface evaporative losses. Evaporation at the burned site marginally exceeded the sum of pre-fire transpiration and interception at the unburned site, suggesting that evapotranspiration during the growing season was 20–40 mm greater at the burned peatland. Although the net change in water storage during the growing season was largely unchanged by fire, the lack of low-density surface peat in the burned site appears to have decreased specific yield, leading to greater water table decline at the burned site despite similar net change in storage.

KEY WORDS Peatlands; wildfire; water balance; boreal; Alberta

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INTRODUCTION

The hydrological response to wildfire disturbance in forested catchments has been widely shown to increase streamflow (e.g. Berndt, 1971; Inbar *et al*., 1998; Neary *et al*., 2004) largely owing to the complete loss of transpiration (Amiro, 2001). Numerous field investigations and modelling studies (e.g. Rouse, 1976; Amiro, 2001; Bond-Lamberty *et al*., 2009) have also shown a reduction in overall evapotranspiration immediately following wildfire in upland boreal forest stands. The decrease in transpiration after wildfire may cause an increase in water storage rather than increased run-off, as is the case in forested hillslopes affected by fire (Inbar *et al*., 1998). Although a majority of studies have focused on streamflow as the main hydrological response to wildfire disturbance, over 20% of the Canadian boreal forest is composed of peatlands (Tarnocai *et al*., 2000), where streamflow or run-off is only a minor component of the water balance, and water storage in the saturated and unsaturated zone represents a better indicator of hydrological response (Van Seters and Price, 2001). This is especially important given that stand-replacing wildfire is the largest disturbance by area in North American peatlands (Turetsky *et al*., 2004).

In forested and afforested peatlands, previous studies have shown trees and shrubs to play a major role in peatland hydrology by simultaneously lowering surface moisture via transpiration and interception (van Seters and Price, 2001) and providing shade that reduces surface evaporative demand on the moss surface (Thompson, 2012). In order to quantify the net impact of trees on the peatland, both the water-losing (transpiration and interception) and water-conserving (shading of the peat surface) impacts of trees need to be considered. Natural, open-canopy forested peatlands have shown a decrease in water table position after wildfire (Simard *et al*., 2007) suggesting that trees may have a water-conserving effect in undisturbed systems. Similarly, Rothwell (1982) observed a 40 cm water table decline after timber harvesting in a forested wetland in Alberta, though the water table was lower at the harvested site compared with unharvested adjacent wetlands only in drier years. However, none of the previously mentioned studies

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specifically accounted for changes in surface evaporation as a result of disturbance and instead, solely examined the net effect of transpiration and surface evaporation processes as reflected in changes in water table. The contribution of specific processes to the net loss of water after wildfire and the ways in which trees impact the water balance of peatlands in the drier and more continental climate of the western Canadian boreal forest remain unknown. Given the strong effect of forest loss (through logging or wildfire) has on increasing water tables, we hypothesize that wildfire in ombrotrophic peatlands in Alberta will have a similar effect of increasing water storage in the system and leading to a rise in the water table in recently burned sites compared with unburned ones.

MATERIALS AND METHODS

Site description

This study was carried in a peatland complex (55.8°N, 115.1°W) 70 km north of Slave Lake, Alberta, Canada. The region is classified as continental mid-Boreal forest (National Wetlands Working Group, 1988). Mean annual precipitation at Slave Lake (1971–2000) was 502 mm, 146 mm of which falls as snow (Environment Canada, 2000). Peatlands in the local area are typically underlain by glaciolacustrine clay deposits (Ferone and Devito, 2004). These low-gradient peatlands are classified as 'flat bogs' and lack surface run-off channels (National Wetlands Working Group, 1988).

The peatland complex is composed primarily of ombrotrophic forested bog that last burned *c.* 1935 (BC35) and an 8 ha portion of the peatland that burned in September 2006 (BC06). Mean combustion depth at BC06 as measured after the fire by the adventitious root method (Kasischke *et al.*, 2008) was 7.5 cm. Hydrometeorological towers were installed in the middle of the BC06 burn scar, and the adjacent unburned stand (BC35) was installed 800 m away in October 2007.

Vegetation at both sites was dominated by *Sphagnum* moss hummocks (height, 0.3–0.4 m; area ≈ 1–4 m^2), as well as low hollows. Hummocks cover 72% of the peatland at BC35 and 62% at BC06. Peat depth averaged 1.7 and 1.6 m at the unburned BC35 and burned BC06 sites, respectively. Hummocks were largely composed of *Sphagnum fuscum* with a shrub canopy dominated by *Ledum groenlandicum* that can reach up to 30 cm in height. Unburned hollow vegetation was a mix of *Sphagnum angustifolium*, lichens and bare peat. After fire, the hummocks sustained little combustion, with most hummocks experiencing less than 2 cm of combustion. Hollows experienced extensive combustion, resulting in a blackened char and ash layer and no remnants of living moss. Hollow surface cover was primarily bare peat with less than 20% *Polytrichum strictum* cover.

Tree cover at both sites is composed entirely of an open canopy of black spruce (*Picea mariana*) with a basal area of 11 $m^2 ha^{-1}$ and an average height of 2.3 m, and all stems at BC06 are fire-killed *P. mariana* snags. Stem density was 19 700 and 16 000 stems ha^{-1} at BC06 and BC35, respectively. Canopy cover (f_{tree}) was 0.27 at BC35 and was calculated as only the area occupied by tree boles (0.001) at BC06.

Water balance

The water balance of both of a low-gradient, ombrotrophic peatlands can be modelled by

$$\Delta S = P - I - E - T \pm Q \pm \Delta\theta \tag{1}$$

where ΔS is change in storage, P is precipitation (mm), I is precipitation interception (mm), E is evaporation (mm), T is transpiration (mm), Q is groundwater flow (mm) and V is the change in unsaturated zone storage (mm).

Using snowmelt as the time for model initiation, water deficits at snowmelt must be accounted for, because snowmelt does not always replenish water deficits in western Canada (Lawson and Dalrymple, 1996). Using the calculated values of ΔS, we modelled water table (WT_m, cm) on a daily time step:

$$WT_m = \frac{(\Delta S)}{(S_y + bS_s)} \pm S_0 \tag{2}$$

where S_y is the depth-averaged specific yield of the peat ($cm^3 cm^{-3}$), b is the thickness of the saturated peat (cm), S_s is the specific storage of the peat and S_0 is the initial moisture deficit (mm). S_y was calculated from laboratory tests from peat at the same sites (Thompson and Waddington, 2013a). S_s was calculated from the relationship between WT and peat surface elevation, using measurements of peat subsidence relative to a 16 mm diameter rebar inserted into the peat down to the clay substrate underlying the peat (Price and Schlotzhauer, 1999). During rain events, the water table rise may be in excess of that predicted by ($S_y + bS_s$) alone, as the addition of water to the capillary fringe causes a rapid rise in the water table with little to no volume of water added (Gillham, 1984). As such, the change in water table in response to a single rain event was modelled on a daily scale:

$$\Delta WT_m = \frac{(P - E - T)}{(S_y + bS_s)} + z_{cf} \tag{3}$$

where z_{cf} is the water table rise due to the capillary fringe which can be estimated from pore size characteristics or water retention tests (Gillham, 1984). Based on the laboratory tests presented in Thompson and Waddington (2013a), we estimated the capillary fringe to be 50 and 25 mm at BC06 and BC35, respectively.

The water table level was continuously logged adjacent to the tower at each site in a hollow using either a Dataflow Systems (Christchurch, NZ) Odyssey capacitance water table logger or an Ott (Kempten, Germany) PLS pressure transducer in a 5-cm diameter PVC groundwater well installed to a depth of 1.5 m. Unsaturated zone water content and storage was monitored using time domain reflectometry probes, see Thompson and Waddington (2013b) for details.

Groundwater flow

Surface run-off channels were not observed in the vicinity of the peatland, so all lateral losses were assumed to be as a result of saturated flow. Steady state groundwater losses from the bog dome (mm d^{-1}) was calculated using the groundwater mound model of Ingram (1982) as modified for variable saturated hydraulic conductivity with depth by Morris and Waddington (2011):

$$Q = \sum_{i=1}^{n} \frac{HK_i \Delta z}{L^2} \tag{4}$$

where H is the height of water table above the base of the peat (m), K_i is the saturated hydraulic conductivity (m d^{-1}), Δz is the thickness of the peat layer (m) and L is the width of the peatland (m). Hydraulic conductivity of the peat was determined using the pump test method of Hvorslev (1951) with peizometers at depths of 0.5, 0.7, 1.0 and 1.2 m below the hollow surface. Although the peatland complex east of Utikuma Lake is over 12 km in width, the local ombrotrophic portion of the complex which both sides reside was estimated from satellite imagery at 1800 m, and the smaller value of L was utilized in the calculations. As the water table declined during the course of the summer, the values of H and Δz were adjusted accordingly for the uppermost layer measured at 0.5 m and representative of 0–0.6 m depths.

Precipitation and interception

In addition to a logging rain gauge at the BC06 tower, throughfall in trees (I_{tree}) was monitored by logging tipping bucket rain gauges, one of which was placed under the dripline of a tree at BC35 and the other at the same distance from a burned tree bole of the same diameter at BC06. Interception by shrubs way from trees (I_{open}) was measured by a series of manual rain gauges deployed away from trees in BC35. Shrub interception was not directly measured at BC06 but was scaled based on the relative difference in shrub LAI. Landscape-scale interception was calculated as

$$I = \left((1 - f_{tree})I_{open}\right) + \left(f_{tree}I_{open}\right) \tag{5}$$

where f_{tree} is the canopy cover, and I_{open} and I_{tree} are the interception away from and underneath the canopy of a conifer tree, respectively.

Evaporation

The Penman–Monteith evaporation model (Monteith, 1965) was used to model peat surface evaporation (E). Net radiation (Q^*) was measured above the canopy with Kipp and Zonen (Delft, Netherlands) NR Lite net radiometers at 10 m elevation. Ground heat flux (Q_g) was calculated using the combination gradient and calorimetric method (Liebethal *et al*., 2005). The ground heat flux was calculated as the gradient across the 10 and 20 cm depth, along with the change in energy storage in the 0–10 cm layer as calculated by thermocouple (20 gauge type T, Omega Engineering: Samford, CT, USA) measurements at 2 and 5 cm. Ground heat flux was measured at one hummock and one hollow at each of the sites.

In order to calculate Q^* at the evaporating peat surface, incoming shortwave radiation at the top of the canopy ($K\downarrow$) was modified by observed light transmittance through the tree canopy (τ_a) obtained via hemispherical photography with a Nikon D80 digital camera and a Sunex Superfisheye lens. Photos were taken on a low tripod at 30-cm elevation, above any shrubs. Canopy photos were analysed via the Gap Light Analyzer software (Frazer *et al*., 2000) at 81 points and averaged across each site. Light transmittance through the shrub canopy (τ_b) was measured using a LP-80 ceptometer (Decagon: Pullman, WA, USA). A simplified version of the LAI model of Norman and Jarvis (1974) was used to calculate LAI from light transmittance beneath the shrubs. Shrub transmittance was combined with values obtained from hemispherical photos taken at the same location in order to calculate a surface-level Q^* at 81 locations at each site:

$$Q^* = K\downarrow \tau_a \tau_b \mathrm{RE} \tag{6}$$

where RE is the summertime understory radiation efficiency of 0.65 for both sites (Thompson, 2012).

Aerodynamic resistance (r_a) of the trees was calculated as a function of wind speed at 10 m height using a wind profile (Allen *et al*., 1998) of three RM Young (Traverse City, MI, USA) model 03101 anemometers. Displacement height (d) was modelled using the simple relation from Abtew *et al*. (1989):

$$d = h_c f_{tree} \tag{7}$$

where h_c is canopy height in metres, and f_{tree} is the tree canopy cover (a value different than canopy openness as measured by a densiometer, Jennings *et al.*, 1999). Aerodynamic resistance from shrubs was not measured via the wind profile technique, but detailed micrometeorological studies have shown that a dense canopy of short shrubs and grasses in treed peatlands confer an r_a of approximately 60 s m^{-1} (Spieksma *et al.*, 1997), which was added to the aerodynamic resistance of BC35. We used a shrub r_a of 35 s m^{-1} at BC06, based on a shrub LAI being half that of BC35.

Surface evaporation was measured at ten locations at BC06 and 11 locations at BC35, each 17 cm in diameter using closed-chamber measurements with a PP Systems EGM-4 infrared gas analyser. The model of Griend and Owe (1994) was used to compute surface resistance (r_s) from surface evaporation. Peat temperature, manually measured at 2-cm depth, was used as the temperature at which to base partial pressure of water vapour in the peat pore space for the resistance model of Griend and Owe (1994). Where 2-cm measurements were missing, a multiple linear regression model based on air temperature and soil temperature at 10 cm was used ($R^2 = 0.77$; $F_{2434} = 747$; $SE_y = 2.6$ °C). A total of 196 surface resistance observations were made between both sites in 2008 and 2009. The geometric mean of hummock and hollow measurements at each site was weighted by the proportion of hummocks ($f_{\text{hum, unitless}}$) and hollows (f_{hol}) on the landscape to derive a site-averaged r_s (s m^{-1}) value

$$r_s = (f_{hum} r_{hum}) + (f_{hol} r_{hol}) \tag{8}$$

where r_{hum} and r_{hol} are the geometric mean of observed surface resistance in hummocks and hollows, respectively (s m^{-1}). Shrubs, primarily *Ledum groenlandicum*, were included in the chamber measurements during measurements of r_s. Supplemental point measurements of shrub stomatal conductance were performed using a Delta-T Devices (Cambridge, UK) AP4 porometer at each site.

Transpiration

Transpiration (T) was modelled as an independent process from surface evaporation. Sapflow observations were made on three trees: one representing the lower 30%, one the middle 40% and one the upper 30% of the within-site distribution of basal area at BC35 using thermal dissipation probes (Dynamax Inc, Houston, Texas, USA). Daily T (mm d^{-1}) was then calculated as the weighted average of the daily sums of sapflow in each of the three trees. Sapflow was only measured in the 2010 field season. For the 2008 and 2009 field season, daily transpiration was modelled using the relationship between transpiration and daily average vapour pressure deficit normalized to the length of the day (D_z; Oren *et al.*, 1996) derived from the 2010 measurements. Transpiration was assumed to begin within 10 days of the last spring snowfall or hard frost, corresponding to Julian Day 143 and 120 in 2009 and 2010, respectively. The Penman–Monteith model was run on a 20-min time step and summed to daily values. The water balance model was run during the 2008–2010 field seasons as a simple spreadsheet. Statistical analyses were performed using the R software (R Core Team, 2013).

Error analysis

Error in the water balance model was accounted for at the daily and annual scale using a Monte Carlo analysis. Daily values of surface resistance for hummocks and hollows at both sites were randomly sampled from the population of observed values. Net radiation was not subject to Monte Carlo error analysis as the net radiometer captured a large area 10 m in radius. Random error was introduced into daily transpiration values by modifying daily transpiration values according to a random sample following the standard error of the regression between transpiration and D_z. At the annual scale, interception error was modelled as the standard deviation of the percentage interception amongst five manual rain gauges at each site, in order to simulate the random error brought about by the location of the rain gauge. Similarly, a standard error of 24% (Cermak *et al.*, 1995) was introduced as a random error in order to account for the small number of sapflow sensors used. Based on repeat measurements of surface moisture at 81 points in a 40 by 40 m area adjacent to the weather station, a surface moisture value for hummocks and hollows at each site was randomly selected following the observed distribution. Change in vadose zone storage in the Monte Carlo analysis was assumed to be proportional to ratio of observed moisture to that randomly selected from the distribution. Using the observations as a deterministic base, the previous errors were introduced randomly and independently in each water balance term at the daily scale, and 40 iterations of each year's water balance were run in order to derive an error estimate of each term and the seasonal change in storage as a whole.

RESULTS

Water table and specific yield

Mean specific yield in the upper 50 cm of hummocks was not significantly different between BC35 and BC06 using a Student's t-test ($t_{38} = 0.51$, $P = 0.615$), with mean S_y at both sites close to 0.50. Similarly, there was a small decrease in mean values of S_y at the hollows from 0.26 at BC35 to 0.21 at BC06, though the difference was not significant according to a Student's t-test ($t_{36} = 0.613$,

$P = 0.526$). When the mean observed S_y values for hummocks and hollows are multiplied by f_{hum} and f_{hol}, site-averaged values of S_y for BC35 and BC06 are calculated to be 0.42 and 0.39, respectively.

Peat depth (b) averaged 1.5 m in the hollows at both sites. Both BC35 and BC06 experienced similar surface elevation change of 5 cm over the course of the 2009 summer. However, the peat elevation change at BC35 occurred during a water table decline of 35 cm, whereas at BC06, the same elevation change occurred over a 50-cm water table decline (Figure 1). The resultant value of bS_s calculated for each site was 0.14 at BC35 and 0.10 at BC06, because the same elevation change required a larger change in water table at BC06.

Snow and throughfall

Snow water equivalent (SWE) in the spring of 2008 was significantly greater at BC06, at 119 mm SWE compared with 89 mm SWE at BC35 ($t_{291} = 8.5$, $P < 0.001$). Mean end of winter snowpack in 2009 measured 96 and 89 mm SWE at BC06 and BC35, respectively, a mean difference of 7%. However, given the equally large variance in snowpack between the sites ($F_{44} = 1.06$; $P = 0.85$), this difference was not significant at a 95% CI ($t_{88} = 1.8$; $P = 0.075$). The winter of 2010 was equal to the driest winter on record, with a complete loss of snow cover by March 1st (Environment Canada, 2010). Snow was a large contributor to the annual water balance of 2008 and 2009, equal to as much as 71% of rainfall received at BC35 and BC06 during the study period from the beginning of May to the end of August.

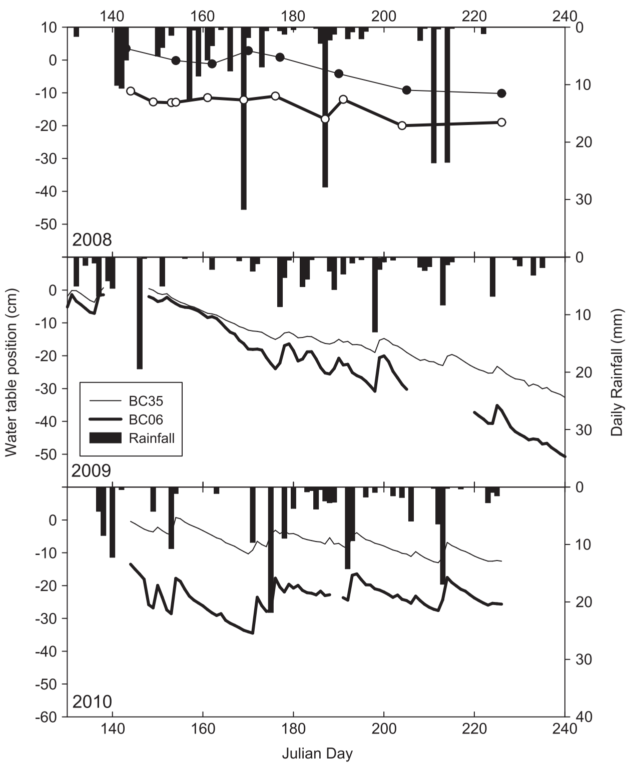

Figure 1. Water table and daily precipitation at peatlands recently burned (BC06) and 75 years since burn (BC35). The water table datum for both sites is the surface of an average hollow. Water table data from 2008 represent point measurements with linear interpolation, whereas 2009–2010 data are end of day values from continuous measurements

Observed cumulative rainfall interception underneath the dripline of black spruce trees in BC35 was 45% and 65% in 2009 and 2010, respectively. Manual rain gauges showed an additional 10% interception by the shrub canopy at BC35, with an estimated 5% shrub interception at BC06. With f_{tree} equal to 0.27, I was calculated to be 20% and 26% of P in 2009 and 2010 at BC35. Although only a single logging rain gauge was placed underneath the spruce canopy, the inclusion of manual rain gauge data allowed for the rainfall uncertainty of 15% per event to be quantified. The observed amount of interception is equal to 9% of E during the study season in 2009 and 24% of E in 2010.

Aerodynamic resistance

Displacement height (d) for the logarithmic wind profile was calculated to be 0.63 and 0.0026 m at BC35 and BC06, respectively. Roughness length for momentum was 0.22 and 0.14 m at BC35 and BCO6, respectively. Median aerodynamic resistance to vapour flux (r_a) was 30 s m^{-1} BC35, similar to 32 s m^{-1} measured at BC06. At the median 10 m elevation wind speed of 2.1 m s^{-1}, r_a was 44 and 45 s m^{-1} at BC35 and BC06, respectively.

Radiation and ground heat flux

Average canopy transmittance at the unburned site was 74% compared with 94% at the burned site. Shrub LAI, equal to 0.84 and 0.44 at BC35 and BC06, respectively, resulted in a further shading of the peat surface by 50% and 33%, respectively. Average net radiation at BC35 was 37% of incoming shortwave radiation at the top of the canopy, whereas at BC06, this proportion was 62%. Q_g in hummocks at both sites average 1.5% of Q^*. Ground heat flux at the hollow at BC35 was on average 6.8% of Q^* but was only 3.2% of Q^* at BC06. When the flux is weighted by f_{hum} and f_{hol}, Q_g of both sites averaged 3% of daily Q^*.

Transpiration

Daily transpiration ranged between 0.2 and 1.2 mm d^{-1}, with an average of 0.7 mm d^{-1}. Cumulative transpiration over study periods in 2008–2010 was 54–75 mm (Table I). Daily transpiration was well-predicted by daily mean vapour pressure deficit (Figure 2; D_z; $R^2 = 0.81$; $F_{36} = 156$; $P < 0.001$). Uncertainty from the regression of sapflow from daily values of D_z resulted in a standard error of approximately 10%.

Table I. Water balance components (mm) of the burned and unburned peatlands during the study periods in 2008–2010

	2008		2009		2010	
Site	BC35	BC06	BC35	BC06	BC35	BC06
Snow	89	119	89	96	~22	~22
S_0	−62	−46	−22	−15	0	−135
P	182	182	135	135	247	247
I	−36 ± 17	−17 ± 16	−34 ± 36	−14 ± 2	−66 ± 20	−27 ± 23
E	−180 ± 9	−244 ± 10	−260 ± 9	−384 ± 13	−252 ± 9	−340 ± 13
T	−62 ± 12	0	−83 ± 19	0	−71 ± 16	0
V	−6 ± 2	−9 ± 5	−7 ± 4	−15 ± 14	−1 ± 1	−3 ± 2
ΔS	−100 ± 22	−85 ± 20	−260 ± 38	−278 ± 27	−140 ± 25	−128 ± 28

Values given are relative to zero at the initiation of the water balance, as shown in Figure 1.
Snow and S_0 are not included in the calculations of ΔS shown here.
Uncertainty is shown as ± 1 standard deviation.

Surface resistance

Median surface resistance (including both peat and shrubs) in the hollows (r_{hol}) did not differ between BC35 and BC06 sites at 78 s m^{-1} (Figure 3), whereas r_{hum} was slightly higher at BC06 (63 s m^{-1}) than at BC35 (56 s m^{-1}). Once weighted for the distribution of hummocks and hollows on the landscape using Equation (8), BC06 had a slightly higher r_s of 69 s m^{-1} compared with 62 s m^{-1} for BC35. Mean stomatal resistance of *Ledum groenlandicum* at BC06 was 542 s m^{-1}, compared with 342 s m^{-1} at BC35. If the peat-shrub system is considered as a parallel circuit analogue, then the high stomatal resistance of the shrubs suggests that shrubs only contribute six and 15% of the total surface conductance at BC06 and BC35, respectively.

Evaporation

Mean daily surface evapotranspiration in the summers of 2008–2010 at BC35 was 2.2 mm d^{-1}, whereas a paired *t*-test of daily evapotranspiration showed BC06 had a significantly larger average daily surface evapotranspiration of 3.4 mm d^{-1} (t_{316} = 48.0, P < 0.001). Evaporation over the duration of the study periods in 2008–2010 (May–August) was 80–160 mm greater at BC06 compared with BC35 (Table I; Figure 4). If the effect of shading from trees was removed and a daily average of 27 W m^{-2} (2.3 MJ m^{-2} d^{-1}) was added to the surface Q^* (Thompson, 2012), then the surface E at BC35 would increase by an average of 0.72 mm d^{-1}, or approximately 60–85 mm over the study season.

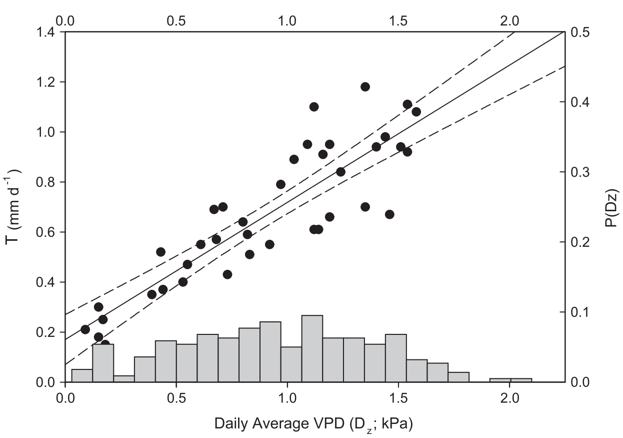

Figure 2. Relationship between measured daily transpiration (mm d^{-1}) scaled from sapflow sensors and daily average vapour pressure deficit (D_z) during 2010. Dashed lines represent 95% CI of the regression line. The frequency distribution of D_z during the period when the sapflow sensors were deployed is shown by the bars along the *x*-axis

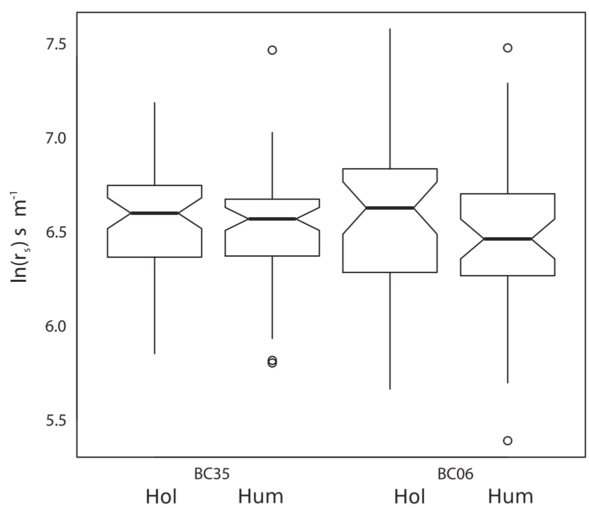

Figure 3. Distribution of surface resistance (r_s) measurements as a function of microtopography and site. The notched boxplots extend $\frac{1.58*IQR}{\sqrt{(n)}}$, where *IQR* is the inter-quartile range, and *n* is the number of observations in the sample, following McGill *et al.* (1978)

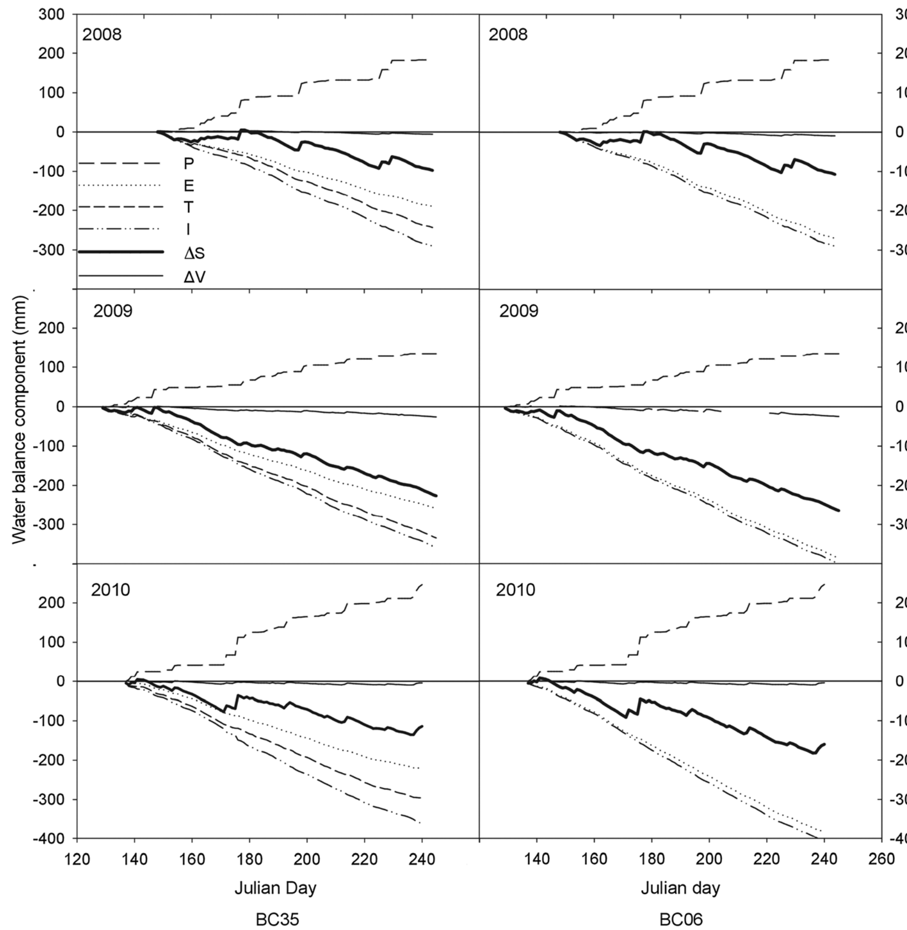

Figure 4. Cumulative seasonal water balance of the burned and unburned peatlands from 2008–2010

Groundwater flow

K_{sat} varied from 51 m d^{-1} for loose peat at 50 cm depth to 0.002 m d^{-1} for peat at 1.2 m depth. Calculated hydraulic gradients from the Ingram model were 0.0015, whereas measured gradients averaged 0.0044 and 0.0020 at BC35 and BC06, respectively. Mean daily Q calculated using Equation (6) was less than 0.1 mm d^{-1}, and thus far below the errors in the other fluxes measured in this study. Using the larger observed hydraulic gradients, this resulted in values of Q still smaller than the error of the larger fluxes.

Storage

Stochastic modelling of change in storage from JD 149–245 using a Monte Carlo method showed nearly identical mean values of ΔS during that period of 238 and 239 mm (Figure 5). Vadose storage change showed the greatest uncertainty in the Monte Carlo simulations, with a coefficient of variation (CV) in the 40 model runs between 0.5 and 0.9, whereas evaporation showed a much lower level of uncertainty with the CV equal to 0.04 at both sites. The CV of *P-I* was greater at BC35 (0.36) compared with BC06 (0.10), which corresponded

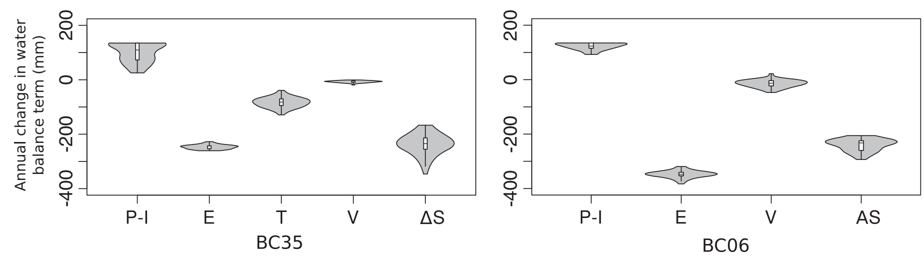

Figure 5. Monte Carlo error analysis of the water balance components from Julian day 129–245, 2009. Data are shown as violin plots with embedded boxplots of the distribution of 40 stochastic iterations of the water balance model. Note that precipitation was not varied, and so the difference of precipitation minus interception (*P-I*) is shown

to similar differences in the magnitude of ΔS (0.16 *vs* 0.09). Optimal water balance model performance was found using single values of $S_y + bS_s$ equal to 0.65 and 0.50 at BC35 and BC06, respectively (Figures 6 and 7), with these optimal values closely matching those measured at both sites.

Although ΔS was well correlated to *WT* decline during rain-free periods without compensating for the capillary fringe, prediction of the *WT* was poor after large rain events (Figure 6). Observed *WT* rise was often 2–3 times what Equation (3) would predict without the inclusion of the capillary rise effect, z_{cf}. Excluding the capillary rise results in an error at the end of the 2010 season of 250 and 125 mm and BC06 and BC35, respectively. With less overall rain and fewer large rain events, the impact of the capillary rise on *WT* fluctuation was only equal to 50 and 25 mm at BC06 and BC35, respectively.

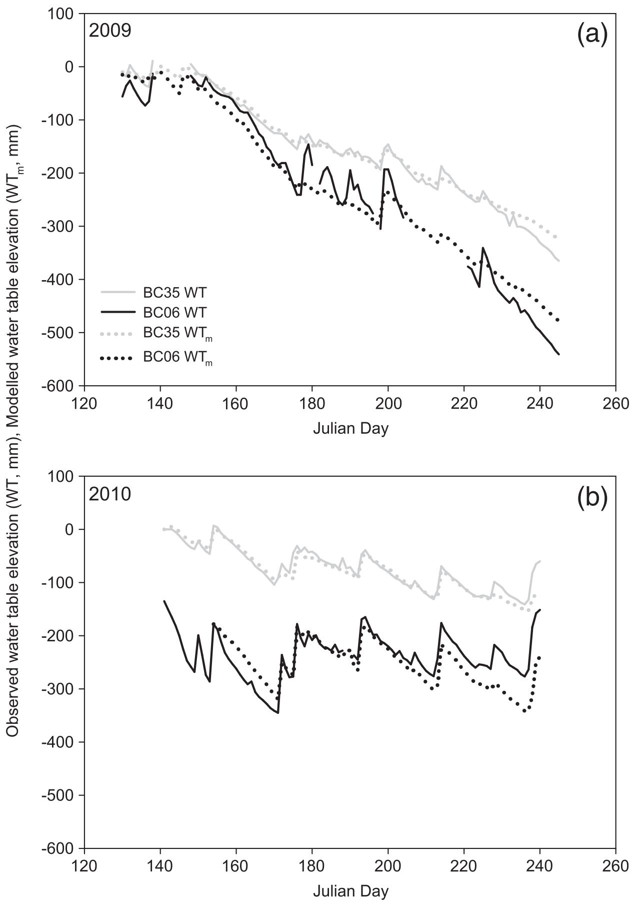

Figure 6. Deterministically modelled water table (as derived from ΔS) plotted against measured water table at BC35 and BC06 during 2009 (a) and 2010 (b)

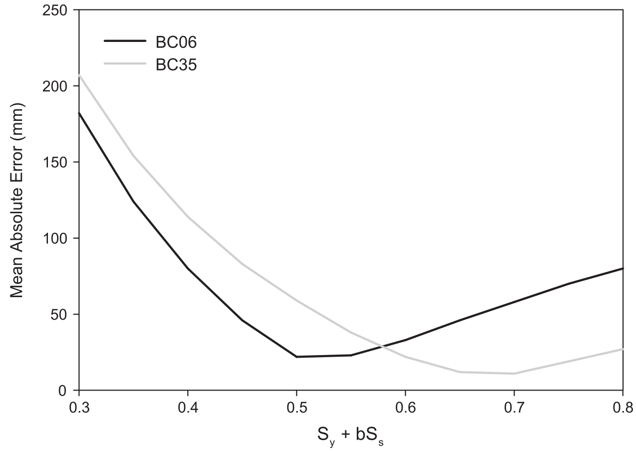

Figure 7. Effect of varying specific yield (S_y) on the accuracy of end of growing season modelled water table in 2009, as measured by mean absolute error, in millimetre

DISCUSSION

Snow

Differences in snowpack between the two sites can largely be attributed to the presence of the unburned conifer canopy at BC35, as conifers such as black spruce have been widely shown to both intercept snowfall (e.g. Davis *et al.*, 1997; Pomeroy *et al.*, 2002) and to enhanced mid-winter ablation via longwave radiation from stems and foliage (Davis *et al.*, 1997). A simple snow ablation model (Pomeroy *et al.*, 2002) that requires only f_{tree} predicts that 12% of the end of season snowpack will ablate.

Snow was a large proportion of the precipitation input in 2008 and 2009, which is not normal for the region (Environment Canada, 2000). In 2009, the amount of snowmelt was sufficient to lead to a wetting up and rise of the water table to surface. In 2010, we were unable to measure the snowpack prior to melt, but it is likely that much of the 20 mm SWE snowpack as measured at Slave Lake was ablated away, and any residual water was minor compared with the large storage deficit from the previous fall.

Surface evaporation

Mean growing season surface evaporation of 3.1 mm d^{-1} at BC06 is similar to the 3.0–3.5 mm d^{-1} of evapotranspiration measured from a closed canopy black spruce forest in northern Manitoba, Canada (Arain *et al.*, 2003). Surface evaporation from BC35 averaged only 1.9 mm d^{-1} and was more similar to measurements of evaporation from the forest floor of closed canopy boreal forests, where *Sphagnum* evaporation is on the order of 1.5 mm d^{-1}, and as little as 0.5 mm d^{-1} in densely shaded areas (Heijmans *et al.*, 2004).

In the measurement of bulk surface resistance (r_s) in peatlands, it is difficult to distinguish moss and shrub contributions to bulk surface resistance without clipping the shrubs or otherwise altering the system. Values of r_s measured in this study are consistent with the lower range of measurements of r_s in a Swedish ombrotrophic peatland by Kellner (2001), who found r_s to be $160 \pm 70\,s\,m^{-1}$, including the shrub canopy. Moreover, in this study, the presence of a generally denser peat owing to the fire disturbance may have increased unsaturated hydraulic conductivity of the peat, thereby likely reducing r_s. This process may be analogous to cutover peatlands, where Petrone *et al*. (2004) measured an increase in evaporation of 5–15% in a cutover peatland compared with an adjacent remediated peatland, despite a 30–35% decrease in surface volumetric water content. The relatively low values of r_s lead to a relative insensitivity of the r_s term in the water balance compared with transpiration and interception (Figure 5). The insensitivity in r_s could be in part due to the presence of lichens, feathermosses and bare peat surfaces, which have high r_s values regardless of *WT* (Bond-Lamberty *et al*., 2011).

Storage and water table modelling

Although the modelled water table presented here fits well with observations, uncertainty remains in both S_y and to a lesser extent evaporation. Laboratory measurements of S_y were limited to six cores from each site, with only triplicate hollow cores at each site that spanned the entire depth of water table variation. The considerable vertical heterogeneity in peat properties (Thompson and Waddington, 2013a) is accentuated by the lack of persistence of microforms after fire, creating a complex pattern of alternating hummock and hollow peat of variable composition and S_y. Moreover, S_y can be expected to decrease with depth (Price, 1997) though the rate of S_y decline with depth would conceivably vary between hummocks and hollows, given their contrasting decomposition rates (Turetsky *et al*., 2008). Encapsulating S_y with a single variable is widely used in wetland water balance models (e.g. Roulet and Woo, 1986; Price and Schlotzhauer, 1999) but is unrealistic under a large *WT* range.

Specific yield and specific storage

Overall, $S_y + bS_s$ was smaller at BC06, as shown by the greater *WT* response to precipitation events (Figure 1). The difference in $S_y + bS_s$ resulted in a greater *WT* response per unit decrease in ΔS. This is particularly important during extended rain-free periods, as the extensive decline of the water table into areas of the peat profile of increasingly smaller S_y (Boelter, 1964) leads to exponentially greater decline in *WT* per unit ΔS. The peat surface at BC06 was more rigid, as evidenced by smaller values of bS_s. The more rigid peat surface was balanced by a larger water table decline over the summer at BC06, resulting in the same peat elevation change at BC35. Though no pre-fire measurements of S_y are available, post-fire irreversible peat compression, similar to that observed in the cutover peatland environment (Kennedy and Price, 2005), may in part be responsible for the smaller S_y values observed here.

Role of shrubs

Shrub recovery in peatlands is rapid after wildfire compared with tree growth, with mean shrub *LAI* having reached 0.44 3 years after fire, compared with an *LAI* of 0.84 in the unburned BC35 site. Given the net effects of high r_c, the 50% reduction in Q^* and the doubling of r_a, shrubs appear to have a significant role in preserving water in the peatland landscape. In the moderate shrub densities observed both before and after fire, there is still sufficient light for *Sphagnum* growth (Bisbee *et al*., 2001), without the dense shrub growth that crowds out *Sphagnum* growth that is seen in cutover peatlands (Farrick and Price, 2009).

Ecohydrological impacts of trees

Transpiration dominates water losses in boreal forest catchments with closed-canopy forests, where soil evaporation is near zero except during snowmelt and after large rain events (Wang, 2008). However, in peatlands, low oxygen in the rooting zone (Silins and Rothwell, 1999) suppresses photosynthesis and potentially the transpiration rate of black spruce trees (Dang *et al*., 1991). The approximately 75-year old black spruce stand at BC35 was a minor contributor to the water balance, transpiring only $75\,mm\,a^{-1}$, equal to only 34% of *E*. In comparison, closed-canopy spruce stands in the boreal forest can transpire upwards of $325\,mm\,a^{-1}$, with only minor surface evaporation (Ewers *et al*., 2005).

The $0.7\,mm\,d^{-1}$ of evaporation saved by the presence of shading of trees at BC35 is similar in magnitude to transpiration in the stand at $0.7\,mm\,d^{-1}$, suggesting that at least in dry periods, the impact of trees on the landscape is water neutral. However, rainfall interception by the same tree canopy at BC35 averaged $0.4\,mm\,d^{-1}$ in the wetter year 2010, effectively tipping the balance towards a net water losing process when interception is considered. In the drier year 2009, interception by trees was only equal to $0.16\,mm\,d^{-1}$, suggesting that trees are a stronger agent of net water loss once rain interception is considered. The additional consideration of up to 25 mm SWE of snowpack ablation during with winter contributes up to another $0.1\,mm\,d^{-1}$ of water loss over an average 200-day snow-free season. However, a few key processes could tip this balance in either direction. An increase in tree density would increase shading only to a point where trees are self-shading, as is the case in closed-canopy forests where the negative feedback of shading is effectively saturated (Gower and Richards, 1990). Once the canopy closes in and *Sphagnum* is outcompeted by feathermosses for light (Bisbee *et al*., 2001), the processes by which *Sphagnum* suppresses the growth of trees (van Breeman, 1995) would also decrease, and a transpiration-driven water table drawdown feedback could take hold (Gorham, 1995).

CONCLUSIONS

Forested peatlands of the boreal plain of western Canada experience periodic wildfire disturbance, which significantly alters the physical structure of the peatland by decreasing shading by trees and creating a more aerodynamically smooth surface. Surface resistance was not controlled by depth to water table, suggesting some degree of resilience of evaporative processes at the peat surface after wildfire disturbance. In the burned site, the high stomatal resistance of shrubs coupled with a significant shrub shading effect provides an effective interim water-conserving mechanism in the absence of a tree canopy.

Trees appear to retain water on the landscape in the absence of rainfall, owing to their modest transpiration and ability to shade the peat surface from additional evaporative demand. These water-conserving effects are likely to be specific to a narrow range of intermediate stem densities and low-growth environments observed in the open-canopy forests typical of forested peatlands. During periods of normal or greater precipitation, rainfall interception by trees could potentially reverse the water-conserving effect of trees relative to an open site due to rainfall interception. An enhanced understanding of tree-moss competition and the implications on the water balance of forested peatlands is essential for predicting water table conditions prior to fire in peatlands, and for estimating the vulnerability of boreal forested peatlands in a changing climate.

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