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DOI: 10.1002/hyp.14132

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Document Version
Publisher's PDF, also known as Version of record

Citation for published version (Harvard):
Kettridge, N, Lukenbach, MC, Hokanson, KJ, Devito, KJ, Petrone, RM, Mendoza, CA & Waddington, JM 2021, 'Regulation of peatland evaporation following wildfire; the complex control of soil tension under dynamic evaporation demand', Hydrological Processes, vol. 35, no. 4, e14132. https://doi.org/10.1002/hyp.14132

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Download date: 15. Jul. 2021
Regulation of peatland evaporation following wildfire; the complex control of soil tension under dynamic evaporation demand

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Abstract

The capability of peatland ecosystems to regulate evapotranspiration (ET) following wildfire is a key control on the resilience of their globally important carbon stocks under future climatic conditions. Evaporation dominates post-fire ET, with canopy and sub-canopy removal restricting transpiration and increasing evaporation potential. Therefore, in order to project the hydrology and associated stability of peatlands to a diverse range of post-fire weather conditions and future climates the regulation of evaporation must be accurately parameterised in peatland ecohydrological models. To achieve this, we measure the surface resistance ($r_s$) to evaporation over the growing season one year post-fire within four zones of a boreal peatland that burned to differing depths, relating $r_s$ to near surface soil tensions. We show that the magnitude and temporal variability in $r_s$ varies with burn severity. At the peatland scale, $r_s$ and near-surface tension correlates non-linearly. However, at the point scale no relationship was evident between temporal variations in $r_s$ and near-surface tension across all burn severities; in part due to the limited fluctuation in near-surface tensions and the precision of $r_s$ measurements. Where automated measurements enabled averaging of errors, the relationship between near-surface tension and $r_s$ switched between periods of strong and weak correlation within a burned peat hummock. This relationship, when strong, deviated from that obtained under steady state laboratory conditions; increases in $r_s$ were more sensitive to fluctuations in near-surface tension under dynamic field conditions. Calculating soil vapour densities directly from near-surface tensions is shown to require calibration between peat types and provides little if any benefit beyond the derivation of empirical relationships between $r_s$ and...
measured soil tension. Thus, we demonstrate important spatiotemporal fluctuations in post-fire \( r_s \) that will be key to regulating post-fire peatland hydrology, but highlight the complex challenges in effectively parameterising this important underlying control of near-surface tensions within hydrological simulations.

**KEYWORDS**
evapotranspiration, fire, negative feedbacks, vadose zone, water repellency, wetland

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### 1 | INTRODUCTION

Peatlands are a key global carbon store, having accumulated \( \sim 500 \) Gt of carbon over millennia (Yu, 2012). They also represent a critical water resource (Holden, 2005), providing the primary landscape water source within the sub-humid climate of the Boreal Plains (Brown et al., 2014; Devito et al., 2017; Gibson et al., 2010). Peatland ecosystems are facing a range of concurrent climate-mediated and land-use change disturbances (Turetsky et al., 2002). These disturbances have the potential to exceed the ecohydrological resilience of northern peatlands (Kettridge et al., 2015) and impact the wider ecohydrological function of the Boreal Plains landscape (Devito et al., 2017; Hokanson et al., 2020). However, peatlands are characterized by an array of negative ecohydrological feedbacks that maintain their characteristic cool and waterlogged conditions, and promote their recovery following disturbance (Belyea & Baird, 2006; Waddington et al., 2015). Evapotranspiration (ET) provides the primary water loss mechanism from these ecosystems within the western Boreal Plains (Brown et al., 2010; Thompson et al., 2015). As such, determining the ecohydrological feedback response of ET to changing climates (Helbig et al., 2020) and disturbances (Bond-Lamberty et al., 2009) is critical to project the future stability and persistence of Boreal Plains peatlands and the associated landscape.

Wildfire represents the largest disturbance to boreal peatlands (Turetsky et al., 2002) and may represent a core catalyst for ecological shifts across the Boreal Plains (Kettridge et al., 2015; Schneider et al., 2016). Wildfire increases potential vaporization by removing the tree and shrub canopies, increasing both the available energy for evaporation (Kettridge et al., 2013) and the connectivity between the atmosphere and evaporating surface (Petrone et al., 2007; Plach et al., 2016). Such increases in PET result in increased evaporation within *Sphagnum* dominated peatlands (Thompson et al., 2014) and, as a result, small increases or decreases in evapotranspiration because of concurrent reductions in transpiration (Morison et al., 2019; Thompson et al., 2014). In comparison, in feather moss dominated peatlands, evaporation can decrease substantially in low severity burns (Kettridge et al., 2017) as a result of increased peat water repellency (Kettridge et al., 2013; Moore et al., 2017); although, this feedback can be exceeded within severe wildfires (Kettridge et al., 2019).

Current studies define the general magnitude and direction of the peatland evaporation feedback following wildfire. However, if we are to represent the absolute response of peatland evaporation to such disturbances under a range of post-fire weather conditions (i.e., both under current inter-annual weather patterns or future climatic conditions), and to embed this evaporation feedback within the wider network of peatland ecohydrological feedbacks that regulate peatland hydrology (Waddington et al., 2015), it is critical that the mechanistic controls on peatland evaporation are accurately parameterised and incorporated into peatland ecohydrological models (e.g., Moore & Waddington, 2015; Nijp et al., 2017; Sonnentag et al., 2008). Within such models, peatland evaporation is simulated through varying derivations of the Penman–Monteith equation, that at their centre can be simplified to a modified form of the Dalton equation (Oke, 1987):

\[
E = \frac{(\rho'_{s} - \rho_{a})}{r_s + r_a},
\]

where \( \rho'_{s} \) is the saturation vapour density of the peat surface, \( \rho_{a} \) the vapour density of the air, and \( r_s \) and \( r_a \) are the surface and aerodynamic resistance, respectively. While the drivers of evaporation [the numerator of Equation (1)] can be obtained from widely available data, uncertainty in the resistances, and notably the surface resistance that regulates the rate of water loss under drying conditions, limits confidence in projections of peatland hydrodynamics under future extremes (Nijp et al., 2017). The surface resistance accounts for the hydration state of the surface (Lehmann et al., 2018) that can be represented by the near surface tensions. Near-surface tension provides a clear indication of this state and strong correlations between peat near-surface tension and \( r_s \) have been demonstrated in peat in the laboratory under steady state conditions; evaporation demand held constant (Kettridge & Waddington, 2014). However, it is unclear the extent to which such laboratory based relationships hold under dynamic field conditions; whether (1) the strength of the feedback response under field conditions differs from that of the laboratory based steady state conditions, (2) short diurnal fluctuations in evaporative demand and episodic rainfall events in the field induce complex spatiotemporal dynamics in observed relationships, and (3) the formation of water repellent peat that provides a diffusion barrier to water transfer within the near surface (Kettridge et al., 2017; Wilkinson et al., 2020) modifies the previously observed relationships.

The surface resistance provides one approach to incorporate the negative feedback regulation of soil water potential on evaporation in the Penman–Monteith equation. Alternatively, if \( r_s \) is assumed to primarily represent the effective reduction in vapour density of the near surface below saturated conditions, evaporation can instead be determined from the difference between the actual vapour density of the peat surface \( \rho_{s} \) and the vapour density of the air:

\[
E = \frac{(\rho'_{s} - \rho_{a})}{r_s + r_a},
\]
Measurements were conducted within the Utikuma Region Study Site (KETTRIDGE ET AL. | 3 of 15). Following rainfall where variations in face tensions providing a direct incorporation of soil tensions within surface is in equilibrium with the near-surface pore water pressure: and the soil surface (Alvenäs & Jansson, 1997); when equal to one the potential between the near-surface (measurement or simulation node) and the soil surface (Alvenäs & Jansson, 1997); when equal to one the surface is in equilibrium with the near-surface pore water pressure:

\[ \psi = \frac{\rho_v - \rho_a}{\rho_a} \]

where \( \psi \) is the soil water potential in the near surface peat, \( M_w \) is the molecular weight of water, \( g \) is the gravimetric constant, \( R \) is the gas constant and \( T \) is the surface temperature (°C) (Philip, 1957). \( e_c \) is an empirical coefficient that corrects for the difference in soil moisture potential between the near-surface (measurement or simulation node) and the soil surface (Alvenäs & Jansson, 1997); when equal to one the surface is in equilibrium with the near-surface pore water pressure:

\[ e_c = 10^{-4 \cdot \psi} \]

where \( \phi_a \) accounts for the difference in tension and \( \delta_i \) accounts for the temporal variation within this correction factor. Previously \( \delta_i \) has been assumed to range between \(-2 \) and \( 1 \) mm and is determined from a simple water balance of the peat surface (Kellner, 2001), with water supply by rainfall, and water loss by evaporation and vapour transfer within the soil. Steady state laboratory based methods (Kettridge & Waddington, 2014) have demonstrated values of \( e_c \) to be consistent with both those originally derived for sand (Alvenäs & Jansson, 1997) and applied within model simulations for peatlands (Kellner, 2001). However, the approach has not been directly assessed under dynamic field conditions within peatlands with variations in \( e_c \), notably for periods following rainfall where variations in \( e_c \) will be attributed to changes in \( \delta_i \).

The aim of this research is to determine the potential of peatland hydrological models to represent the negative feedback response of evaporation to drying of burnt peatlands under dynamic natural environmental conditions. We will: (1) quantify surface resistance to evaporation within a peatland under a range of burn severities, and define its temporal variability in response to diurnal fluctuations in evaporation demand and episodic rainfall events; (2) determine the nature and strength of the relationship between the surface resistance and near-surface tensions under different burn severities, and identify how this differs from steady state laboratory conditions; and (3) quantify the magnitude of the equilibrium coefficient \( e_c \) for different peat burn severities, its consistency with laboratory based estimations of its value, and its response to rainfall events.

**2 | METHODOLOGY**

**2.1 | Study site**

Measurements were conducted within the Utikuma Region Study Area (URSA) in north-central Alberta (56.107° N 115.561° W), within a coarse-textured outwash (Devito et al., 2012; Hokanson et al., 2019). Measurements were undertaken within a small peatland lobe approximately 60 m by 150 m in size, surrounded by aspen forestland (cf. Lukenbach, Hokanson, et al., 2015; Figure 1). The peatland was burnt in May 2011 as part of the ~90 000 ha Utikuma Complex wildfire (SWF-060). Prior to the fire, the peatland was characterized by feather moss (Pleurozium schreberi) and a dense dark spruce tree (Picea mariana) canopy (~7000 stems per hectare). Within the middle area of the peatland Sphagnum fuscum hummocks were interspersed within the feather moss lawns. Depth of burn varied between 0.00 to 1.10 m across the site (Lukenbach, Hokanson, et al., 2015). Following the wildfire the site was classified into two zones (middle and margin) determined both from the distance to the pre-fire peatland-upland interface and the burn severity; higher burn severity within the margin peat (burn depth averaging 0.08 ± 0.01 m and 0.42 ± 0.02 m within the middle and margin peat, respectively; cf. Lukenbach, Hokanson et al. (2015)). The peatland middle was further classified into two zones: hummocks (S. fuscum) and lawns (feather moss). Lawns accounted for approximately two-thirds of this central zone (cf. Lukenbach, Hokanson et al., 2015). The margin was covered only by feather moss pre-fire and was again further classified into two zones: areas that had severely burned to mineral soil or areas in which a peat layer remained. This resulted in four classified zones: middle hummocks, middle lawns, margin peat, margin mineral. The water-table depth differed between these four zones of the peatland, averaging 0.65 m (middle hummocks), 0.41 m (middle lawns), 0.18 m (margin peat), and 0.10 m (margin mineral).

**2.2 | Hydrological and micrometeorological measurements**

ET under defined aerodynamics conditions (within an enclosed chamber, with air mixed by a fan) was measured every hour in each of the designated zones of the peatland throughout the 2012 growing season (May 24th to August 11th), approximately one year following wildfire. Measurements were conducted using an automated version of the chamber approach of McLeod et al. (2004). Three Perspex chambers, with 0.2 m² surface area, were installed within each designated zone. To measure ET, the chamber was closed for 2 min and the air within the chamber continuously mixed by a fan. ET was calculated from the rate of increase in humidity within the closed chamber of known volume (see Kettridge & Waddington, 2014), which was measured using an infra-red gas analyser (LI-COR LI-840). The rate of increase in humidity was determined by fitting a log linear relationship to the humidity difference (difference between the time varying chamber humidity and the constant maximum measured humidity during a given chamber measurement) against time over the first 30 s of each chamber measurement. This log linear relationship was solved to determine the gradient at time zero. Within the low humidity continental climate of the boreal plain, we did not observe extensive fogging within the chamber during the key daytime measurements, and
notably during the early section of the measurement period used for determination of ET. As a result we did not implement a correction factor to take account of this potential sink function within the chamber. Such a constant will influence only the absolute magnitude ET measured and not the relative differences between the chambers and over time which is the focus of this study.
In addition to the ET measurements, peat surface temperatures and air temperatures were measured inside each chamber using type T and type K thermocouples, respectively. Surface temperatures were further supplemented with manual infra-red skin surface temperatures obtained within each chamber throughout the growing season using a FLIR i3 infrared camera. The total resistance to ET ($r_{\text{total}}$, the sum of the surface resistance and aerodynamic resistance) was calculated for each measurement through the inversion of Equation (1) using measurements of surface and air temperature obtained at the point of chamber closure. In laboratory measurements $r_s$ was calculated to equal 62 s m$^{-1}$ by measuring $r_{\text{total}}$ for a peat surface that was saturated to minimize $r_s$ to be close to zero (Kettridge et al., 2017). However, it is recognized that $r_s$ may change both in space and time as a result of variations in surface roughness and boundary layer conditions. Throughout, we therefore present the total resistance to evaporation, with the assumption that changes in $r_{\text{total}}$ are dominated by variations in $r_s$.

The leaf area index (LAI) was determined from the inspection of detailed images taken of the chambers through the growing season, using either the leaf count approach (Strack et al., 2004) or image classification in accordance with Kettridge and Baird (2008) depending on the density or type of vegetation. Vascular vegetation was dominated by *R. groenlandicum* with some *R. chamaemorus* in the middle hummocks and by *Epilobium angustifolium* in the margin peat and mineral zones (Figures S1 and S2). Vascular vegetation was not present within the middle lawns. Stomatal conductance was determined for each dominant species within each chamber. Measurements were performed on three leaves, of three separate plants (where possible), within each chamber using an AP4 Delta-T porometer. Within subsequent calculations we apply lower stomatal conductance of *R. groenlandicum* used for chamber measurements (Chamber 2). Tensions were recorded at 20-min intervals for three months during the study period, through May to August. Automatic tensions correlate strongly with manual measurements of these tensiometers ($R^2 > 0.95, n = 32, p < 0.001$). Logging tensions show consistent patterns between each probe over the measurement period (Figure S4). However, tensiometer 1 better detects the higher tensions during the dry periods of the study (Figure S4) and is applied within the subsequent analysis.

Water table position was recorded at 20-min intervals by capacitance water level recorders (Odyssey Data Recording, Christchurch, New Zealand) that were installed within 0.05 m diameter PolyVinyl Chloride (PVC) wells, one in the middle and on in the margin of the peatland. The depth to water table at locations where the chambers and tensiometers were installed was determined by measuring the water level in the closest well and assuming a flat water table position between the well and the adjacent chamber or tensiometer.

### 2.3 Hydrological and micrometeorological analysis

The control of the different measurement zones (middle hummocks, middle lawns, margin peat, margin mineral) on log $r_{\text{total}}$ were analysed using a general linear model in R with zone as a fixed effect and collar ID as a random effect to account for the lack of independence among collar measurements. To quantify the control of water availability on ET, individual measurements of near-surface tension and $r_{\text{total}}$ were directly correlated. Further, the diurnal variations in near-surface
tensions observed during the intensive measurement period were related to average diurnal variations in $r_{\text{total}}$ observed during the entire study period. The latter approach is applied to average errors in individual $r_{\text{total}}$ measurements; errors in measured $r_{\text{total}}$ particularly result from small-scale spatial variations in surface temperatures and errors in measured rates of ET (notably under periods of low ET). Such errors under dynamic field conditions could be larger compared to steady state laboratory conditions where solar heating was kept to a minimum (Kettridge & Waddington, 2014).

$e_c$ was calculated for each point in time and space where both a measurement of near surface tension and ET were available. As a result, ET was determined at a low temporal interval across each of the surface chamber measurements and at a high temporal frequency for Chamber 2 in the middle hummock zone, utilizing the logging tensiometer measurements. To calculate $e_c$ we apply a form of Equation (2) to the effective humidity gradient between $\rho^{*\text{vs}}$, the saturation vapour density of the peat surface, and $\rho_{\text{vs}}$, the actual vapour density of the peat surface, where:

![Figure 2](#)

**FIGURE 2** Temporal variation in median hourly (9:00 to 18:00) total resistance to evapotranspiration through the four defined zones between day of year 136 and 225 (May 15th and August 12th) 2012. Middle and margin determined by the distance to the pre-fire peatland–upland interface and the burn severity; higher burn severity within the margin peat. Middle further classified into hummocks (Sphagnum fuscum) and lawns (feather moss). Margin divided between areas that severely burned to mineral soil or areas in which a peat layer remained. Each colour represents a separate chamber within each zone; chamber 1, 2 and 3 are white, grey and black, respectively. Error bars represent quartile range. Note difference in scale and log scale of (a) compared to other plots.
\[ E = \left( \frac{\rho_{vs} - \rho_{rs}}{\rho_{vs}} \right) \]  

We rearrange Equation (5) and solved for \( \rho_{rs} \) and subsequent rearrange Equation (3) and solving for e_c.

3 | RESULTS

3.1 | Spatiotemporal variation in \( r_{total} \)

There was a significant effect of measurement zone on median daytime (09:00–18:00) \( r_{total} \). Median daytime \( r_{total} \) for the study period was orders of magnitude higher within the middle lawn (4610 ± 1380 s m\(^{-1}\); median ± SD) than in the middle hummock (242 ± 52 s m\(^{-1}\); \( t = 17.9, p < 0.0001 \)), margin peat (155 ± s m\(^{-1}\); \( t = 16.6, p < 0.0001 \)) and margin mineral (132 ± 15 s m\(^{-1}\); \( t = 16.7, p < 0.0001 \)). Diurnal variations in \( r_{total} \) show consistent patterns between chambers within each zone (Figure 2). Within the middle lawns, average \( r_{total} \) increases between 9:00 and 13:00. The resistance subsequently remains relatively consistent for the remainder of the day. There is considerable scatter around these median values. The high scatter in the middle lawn compared to other measurement regions likely results from the low ET used to calculate \( r_{total} \) and the resultant increased percentage error in measured ET. In comparison, the three other zones show a constant, or declining, median \( r_{total} \) during the first part of the day (9:00–12:00). Median \( r_{total} \) subsequently increases linearly with time through the remainder of the day (12:00 and 18:00).

Over the course of the growing season, \( r_{total} \) does not show a clear long-term trend within any measurement zone. However, moderate but consistent patterns in median daytime \( r_{total} \) are observed within each of the different zones during periods of high evaporative demand after rainfall, exemplified by the days 145–152 of drying that followed a rainfall on day of year 144 (Figure 3). Within the middle lawns, \( r_{total} \) triples over the first four days after rainfall, increasing

![Figure 3](image-url)
from 1000 to 3000 s m\(^{-1}\) (Figure 2a). The rate of increase in \(r_{\text{total}}\) subsequently reduces for the remainder of the rain free period. In comparison, \(r_{\text{total}}\) in the middle hummocks initially remains constant (Figure 2b). Four to six days after rainfall, on day of year 148 and 150, \(r_{\text{total}}\) subsequently increases for two of the chambers, with median daytime \(r_{\text{total}}\) increasing by 2.5–4.5 times its post rainfall values. The burned peat margin shows a consistent increase of \(\sim 50\%\) in \(r_{\text{total}}\) for all three chambers during the dry period (Figure 2c) while \(r_{\text{total}}\) of the mineral margin remains constant through this period (Figure 2d).

3.2 | \(r_{\text{total}}\) versus tension at varying spatiotemporal scales

3.2.1 | Spatial variations

At the peatland scale, zones of high near-surface tension (middle lawns) correspond with high \(r_{\text{total}}\) (Figure 4a). Regions of low near-surface tension (middle hummocks and margins) correspond with low \(r_{\text{total}}\). While there is a general pattern of increasing \(r_{\text{total}}\) with increasing soil tension, due to the clustering of resistance and tension measurements within burn severity classes, an empirical relationship is not clearly evident between near-surface tension and \(r_{\text{total}}\) at the peatlands scale. No relationship between \(r_{\text{total}}\) and water table depth is identifiable at the peatland scale (Figure 4b).

3.2.2 | Temporal variation; growing season

At each measurement location, a direct relationship between concurrent measurements of \(r_{\text{total}}\) and manual measurements of near-surface tension across the growing season is not observed (data not shown). A very weak exponential relationship between concurrent near surface tension and \(r_{\text{total}}\) is evident between logging tensions and \(r_{\text{total}}\) within the single hummock (\(R^2 = 0.13, n = 567, p < 0.001\); excluding \(r_{\text{total}} > 500\) s m\(^{-1}\) and measurements outside the time period 9:00–18:00). High scatter in the relationship may result from uncertainty in the humidity difference between the surface and atmosphere (the numerator within Equation (1)) resulting from small-scale spatial variability in surface temperatures. Within the 0.2 m\(^2\) chambers, standard deviations in surface temperature within individual IR images ranges between 0.4 and 7 °C (\(\mu = 2.3\) °C, \(n = 36\)). The standard deviation is temperature dependent, with higher standard deviations observed at higher average temperatures (\(R^2 = 0.52, p < 0.001, n = 36\)). At 20°C, a 2 °C error in the average surface temperature equates to additive errors in the calculated humidity difference of Equation (1) of 0.0031 kg m\(^{-3}\) K\(^{-1}\) due to errors in the calculation of \(\rho^*_v\) (Oke, 1987). Excluding measurements below a threshold humidity difference of 0.01 kg m\(^{-3}\) and also excluding low ET measurements (<0.2 mm h\(^{-1}\)) where percentage errors will be high, increases the strength of the relationship between \(r_{\text{total}}\) and near surface tension (Figure 5b, \(R^2 = 0.34, p < 0.001, n = 375\)). This relationship is principally driven by a nine day period with high evaporation gradients between DOY 186 and 196 (Figure 5a). This was the only period during the two month installation of the logging tensiometers when tensions exceeded 200 cm on successive days and reached a high of 250 cm (Figure 5a). During this nine day period, a strong exponential relationship is evident between tensions and \(r_{\text{total}}\) (\(R^2 = 0.82, n = 89, p < 0.001\), Figure 5c). Under low positive tensions, \(r_{\text{total}}\) approximates values observed under laboratory conditions at similar tensions. However the gradient of the relationship between \(r_{\text{total}}\) and near surface tension is higher in the field conditions than under steady state laboratory conditions; higher \(r_{\text{total}}\) is observed in the field for a given near surface tension within the peat (Figure 5c).

**FIGURE 4** Median total resistance to evaporation and (a) median near-surface (5 cm depth) tension and (b) water table depth between day of year 136 and 225 (May 15th and August 12th) 2012. Error bars represent quartile range
3.2.3 | Temporal variation: diurnal

The near-surface tension of the middle hummock shows a clear but moderate diurnal variation. The average tension for a given hour of the day ranges from a maximum of 159 cm at 19:00 to a minimum of 122 cm at 11:00. The median \( r_{\text{total}} \) for given hour of the day within the middle hummocks relates with the associated average tensions between 09:00 and 18:00 for each of the three chambers (Figure 6; \( R^2 = 0.87 \), 0.71 and 0.42 for chambers 1, 2 and 3, respectively). In comparison, within the middle lawns, the inverse relationship is observed. Over the extent of the study period, \( r_{\text{total}} \) peaked at 13:00 (Figure 7a), while near-surface tensions measured over the intense 24 h measurement period (normalized between their maximum and minimum values) peak at 00:00 and reach their minimum values at 12:00. In the margin zones, no diurnal variation is observed in near-surface tensions during the intensive 24 h study period (data not shown) despite diurnal variations in margin \( r_{\text{total}} \) (Figure 2c,d).
3.2.4 | Spatiotemporal variation in $e_c$

Within the middle hummock, log($e_c$), equal to $\phi_g \delta_s$/$C_1 \delta_s$, varies between 3.9 and 4.0 (Figure 8) and is at the upper end of the previously parameterised value that assumes near-surface moisture content of the peat is depleted. Log($e_c$) is significantly higher within the margin peat ($t = 8.4$, $p < 0.001$) and the margin mineral ($t = 7.5$, $p < 0.001$) than the middle hummock, with median log($e_c$) ranging between 4.4 and 5.0. Within the middle lawn, log($e_c$) extends further beyond the defined limits of past model parameterisations, with median values equal to 5.5–5.6. Within all zones, the interquartile range of log($e_c$) is less than 0.9. Assuming the maximum $\delta_s$ equal to 2, minimum $\delta_s$ varies between just 1.7 and 1.9 to represent the observed interquartile range in log($e_c$). The logging tensiometer provides a more continuous record of log($e_c$) within the middle hummock. This demonstrates limited daytime variation in log($e_c$) (Figure 9a). This daytime variability in log($e_c$) is broadly equivalent to the multi-day range (Figure 9b). Between day of year 154 and 216, log($e_c$) does fluctuate between 3.5 and 4.0. However, the direct relation between with periods of wetting and drying in response to rainfall is not clearly apparent.

4 | DISCUSSION

4.1 | Vertical hydraulic connectivity and its control on $r_{total}$ dynamics

The magnitude, diurnal fluctuation and the short-term response to drying of $r_{total}$ varies between peatland zones and likely results from difference in the hydraulic connectives between the surface and the saturated peat below (McCarter & Price, 2014). Within the peatland margin burned to mineral soil (mineral margin), the sand profile with comparatively high water retentions, high unsaturated hydraulic conductivities (Carsel & Parrish, 1988; Smerdon et al., 2007) and shallow water table depths (Lukenbach et al., 2016) provides a well-connected system in which water can be supplied to the evaporating surface continuously through periods of high demand. At the other extreme, the middle lawns are highly disconnected vertically to the atmosphere (Lukenbach et al., 2016) as a result of the water repellent layer on singed feather moss (Kettridge & Waddington, 2014) and a deeper water table position. $r_{total}$ increases quickly in response to drying because the supply of water to the peat surface is severely limited. The middle Sphagnum hummocks and margin peat fall between these two extremes. After a period of rainfall, the resistance of the middle hummocks initially remains low because of the high capacity for vertical water transport under unsaturated conditions (McCarter & Price, 2014). The subsequent divergence in the response is likely associated with differences in the burn severity of the hummocks. $r_{total}$ increased to >700 s m$^{-1}$ in a hummock that was more severelyburned, with the removal of the Sphagnum capitula and the likely reduction in the connectivity (Lukenbach, Devito, et al., 2015). In comparison, the other two Sphagnum hummocks appeared visually unaffected by the fire and were better able to supply water to support
evaporative demand. The margin peat showed a moderate capability to supply water to the remaining burned peat surface, with tension increasing consistently during the period of drying.

With evaporation being predominantly from the peat surface (Price et al., 2009), the lag of near-surface tensions to $t_{\text{total}}$ provides an indication of the connectivity within the top peat layer. This lag differs between peatland zones, highlighting differences in the vertical connectivity. Diurnal variations in near-surface tensions within the middle Sphagnum hummocks correlate with average resistances showing a strong vertical connectivity; tensions at a depth of 0.05 m are closely in sync with surface tension. In comparison, the decrease in near-surface tension during the day in the middle feather moss lawns with no vascular vegetation while $t_{\text{total}}$ increases suggest that near-surface tensions are lagged behind and poorly connected to the surface. Either that or: (1) the disconnect between the ceramic cup and the dry hydrophobic peat limits the response rate of the tensiometers, or (2) changes in the temperature of the tensiometer head space are strongly modifying tension readings (Butters & Cardon, 1998; Warrick et al., 1998). The measurement of near surface tensions within peat soils represents a core measurement challenge, with the open structure of the peat limiting contact between the peat and the tensiometers, tensiometers require larger ceramics to maintain contact preventing finer scale measurements within the top centimeter of the peat profile. More novel approaches to the quantification/characterize of the soil tensions and hydrological dynamics in the very near

**FIGURE 8** $\log(e_c)$ across the four different zones of the peatland between day of year 136 and 225 (May 15th and August 12th) 2012 from manual tensiometer measurements and concurrent chamber measurement. White, grey black represent chamber 1, 2 and 3 within each zone, respectively

**FIGURE 9** (a) Daytime variation in $\log(e_c)$ between 09:00 and 18:00 within middle hummock 2 between day of year 154 and 225. (b) Time series of selected data period. Solid lines represent a 48-h running mean of $\log(e_c)$ and hourly rainfall measured at the study site.
surface of this open structure will likely support understanding of the regulation of water loss from these peat systems.

The lack of diurnal variation in the near-surface tension within the margin peat zone, largely as a result of the water table being in close proximity to the peat surface, suggests that observed variations in near-surface tensions result solely from diurnal fluctuations in tensions within the very near-surface of the remaining peat profile or a driven by the comparatively small transpiration component of ET. The nature of the connectivity between near-surface soil tension and surface peat vapour density therefore, represents an important control on evaporation that needs to be both effectively conceptualized and parameterised within hydrological models. To effectively parameterise these reductions in evaporation in response to periods of drying, but also understand how these near-surface peat soils are modified by burning, is therefore key. Although water repellent near-surface peat represents the most extreme (but not unusual) level of modification, it is also important to parameterise the full spectrum of wildfire induced hydrological changes that modify the response of evaporation to drying.

4.2 The control of near-surface tension on evaporation

As hypothesized, at the peatland scale, increases in soil tension between zones of the peatland are associated with increases in \( r_{\text{total}} \). However, despite the controlled nature of this field research, with concurrent measurement of \( r_{\text{total}} \) and near-surface tension, direct relationships between \( r_{\text{total}} \) and near-surface tension were not identifiable across the measurement locations. Here, we consider the extent to which: (a) tensions vary sufficiently over time at individual measurement locations to limit ET, (b) the uncertainty in individual \( r_{\text{total}} \) measurements, and (c) the disequilibrium between surface and subsurface tensions within the dynamic surface boundary under field conditions.

Errors in individual measurements of \( r_{\text{total}} \) are important to consider. These errors result from uncertainty in the humidity difference between the soil and atmosphere and in the measured ET. Errors in the calculated surface humidity of up to ±0.003 kg m\(^{-2}\) result from variability in measured surface temperatures. Within middle hummocks and margin areas, average humidity differences range between 0.008 and 0.02 kg m\(^{-2}\) at 13:00. Spatial variability in the surface temperature therefore results in errors in the humidity gradients of between 15%–38% for the different chambers. Thus, the direct comparison of individual measurements of \( r_{\text{total}} \) to near-surface tensions is unlikely to produce clear relationships when temporal variations in \( r_{\text{total}} \) are small. Further, despite the 40 manual measurements of tensions at each location through the study, periods of high tension were not captured. Within the margin peat and mineral zones, near-surface tensions remained within ±0.05 m of hydrostatic equilibrium for the entire study period (Lukenbach et al., 2016). No variation in \( r_{\text{total}} \) would thus be expected based on laboratory-derived relationships (cf. Kettridge & Waddington, 2014). Furthermore, the 90th percentile of measured tension is 154 cm within the middle hummock zone.

Similarly, the expected range in \( r_{\text{total}} \) from laboratory measurements equates to 85 s m\(^{-2}\). While such a range in \( r_{\text{total}} \) is difficult to measure under field conditions, it is important in terms of water loss from the Boreal Plain landscape where small modifications in \( r_{\text{total}} \) can have important consequences for water conservation within these water limited environments (Devito et al., 2017). Such difficulties can be overcome by targeting periods of high evaporation and humidity gradients when percentage errors are small.

A weak relationship between \( r_{\text{total}} \) and near-surface tension was observed within the Sphagnum hummock; the high number of tensiometer measurements enabled both a larger range in near-surface tension to be observed and the exclusion of \( r_{\text{total}} \) measurements when errors were likely high. The relationship is dominated by a 10-day period of high ET, high humidity differences, and near-surface tensions which exceed 200 cm. During the remainder of the periods, \( r_{\text{total}} \) remains low and is poorly correlated to near-surface tensions. This switching between a low and a high resistance condition that depends on near-surface tension is consistent with laboratory-based observations (Kettridge & Waddington, 2014) and is analogous to the stage 1 and stage 2 evaporation (Lehmann et al., 2008) where high relative constant evaporation (stage 1) transitions to lower rate of evaporation (stage 2) that is controlled by vapour diffusion through the porous media. However, the reason for the period of strong relationship is unclear. While this period does represent the highest observed tensions during the study period, the enhanced relationship between near-surface tensions and \( r_{\text{total}} \) is observed prior to these high tensions being reached. Early during this period, tensions are comparable to times where little if any relationship is observable between near-surface tensions and \( r_{\text{total}} \). In addition, ET is equally composed of transpiration and evaporation within this given hummock, and the observed increases in resistance may also result from a restriction in transpiration. During the period of strong relationship, the relationship observed between \( r_{\text{total}} \) and near-surface tension under dynamic field conditions is more sensitive compared to steady state laboratory conditions; with \( r_{\text{total}} \) increasing to higher values under a dynamic evaporative demand for a given near surface tension. This is indicative of the dynamic nature of the field measurements, with tensions at a depth of 0.05 m being in disequilibrium with the evaporative demand. Therefore, measured tensions in the near-surface are lower for a given \( r_{\text{total}} \).

An approach to embed the connection between near-surface tension and vapour density within the Penman–Monteith equation has been proposed (Kellner, 2001). However, it is clear that this relationship must be directly parameterised for a given soil profile in a manner similar to any identified relationships between near surface tension and surface resistance. \( s_\delta \) appears to not be impacted by precipitation inputs compared to laboratory parameterizations under steady state conditions, but with some variability associated with diurnal evaporation demand. High ET demand and low storage within the near surface may minimize periods of lower \( s_\delta \) under field conditions. While this insensitivity is beneficial for model parameterisations, \( q_\phi \) differs by orders of magnitude between the different zones of the peatland. Measurements within the middle hummock zone are comparable to
both initial applications of this modelling approach (Kellner, 2001) and laboratory based parameterisations (Kettridge & Waddington, 2014). However, \( \phi_g \) is substantially greater within the margin zones and middle lawns. As a result of this variability, it is uncertain as to whether this approach provides any added benefit beyond empirical relationships between surface resistance and near-surface tensions.

4.3 Moving forwards in peatland evaporation simulation

Within peatland hydrological models it is important to recognize that controls on evaporation go beyond water table depth. Effective simulations should acknowledge the important influence peat properties and water repellency can have on spatiotemporal evaporation dynamics. But in these more advanced models that incorporate aspects of vadose zone hydrology, near-surface field based measurements cannot directly quantify surface controls on evaporation without further measurement advances. This does not on its own indicate that laboratory derived parameterizations under pseudo steady-state laboratory conditions will not effectively support the development of peatland hydrological models. Simulated surface tensions can vary substantially within the top 5 cm of the peat profile (McCarter & Price, 2014). The strength and dynamic nature of this near-surface tension gradient may underlie the disconnect between field measurements of surface resistance and near surface tension. Moving forward, simulating the vadose zone hydrology of soil profiles of known hydrophysical properties and driven by measured rates of evaporation would offer the opportunity to infer the relationship between evaporation and resultant simulated surface and near surface tension. While this does not independently derive this critical relationship, such an inverse modelling approach would generate surface resistance-tension relationships that can be evaluated again using those obtained under laboratory and field based conditions, offering a strong opportunity for moving forward.

Alternatively, additional approaches formulated from first principles could offer future promise in representing peatland evaporation. Clear threshold responses, switching between high and low rates of evaporation, have been observed in peat profiles under laboratory conditions (Kettridge & Waddington, 2014), with some evidence of such a response under field conditions within this study that is representative of a transition between stage 1 and 2 evaporation. As a result, the application of evaporation lengths and quantification of this transition may provide a very compelling future approach within peatland ecosystems (Or et al., 2013). To date the application of evaporation lengths has targeted mineral soils using differences in soil types over regional and even global scales (Lehmann et al., 2018) to explore spatiotemporal variations in the evaporation feedback response. The diversity of water retention within peat soils over the decimetre scale explored here transcends much of this diversity in soil properties, in the peat soils themselves (from moss surface to decomposed exposed peat where peat soils still persist, Thompson & Waddington, 2013) to sandy and clay rich mineral soils exposed by complete combustion of the overlying peat. However, traditional and widely applied modelling software can be fairly restrictive in the application of dynamic surface resistance values (e.g., Hydrus) and therefore can perpetuate the application of a simple threshold response (Kettridge et al., 2016; McCarter & Price, 2014), that does not reflect the more complex dynamics observed.

5 CONCLUSIONS

This work illustrates the challenges of accurately incorporating the response of evaporation to peatland drying into numerical modelling frameworks, even when the water table remains comparatively close to the peatland surface. The surface resistance in this study was shown to be neither constant in space nor time, with differences likely meaningfully impacting the ecohydrological response of peatland ecosystems post disturbance. We have shown that the controls occur in the very near-surface, making them difficult if not impossible to measure with traditional field based instrumentation. Laboratory based measurements under steady state conditions (Bond-Lamberty et al., 2011; Kettridge & Waddington, 2014) offer a starting point to parameterise the control of surface tensions on the surface resistance. However, we have shown that modifications are necessary to apply such relationships to the range of conditions observed across field sites. Dynamic conditions are shown here to result in differences in the relationship between measurable near-surface tensions and surface resistances. We suggest that the gradient in soil tensions in the near-surface peat is enhanced under non steady state evaporation conditions observed under field conditions, resulting in an apparent insensitivity of \( r_{\text{total}} \) to variations in near-surface tension. But the connectivity between \( r_{\text{total}} \) and near-surface tensions also shows an apparent temporal switching on and off, turning on for a discrete dry period of time after rainfall. Threshold responses therefore appear to underlie this relationship, occurring only during the highest rates of evaporation observed in a given year; although this likely differs between peatlands and climates, notably future climates.

ACKNOWLEDGEMENTS

Financial support was provided by Syncrude Canada Ltd, Canadian Natural Resources Ltd. (SCL4600100599) industry partners and Natural Sciences and Engineering Research Council–Collaborative Research and Development grant (NSERC-CRDPJ477235-14) of Canada to KJD. We thank two anonymous reviewers for their feedback on an earlier version of the manuscript.

DATA AVAILABILITY STATEMENT

The data that support the findings of this study are available from the corresponding author upon reasonable request.

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SUPPORTING INFORMATION

Additional supporting information may be found online in the Supporting Information section at the end of this article.

How to cite this article: Kettridge N, Lukenbach MC, Hokanson KJ, et al. Regulation of peatland evaporation following wildfire; the complex control of soil tension under dynamic evaporation demand. Hydrological Processes. 2021:35: e14132. https://doi.org/10.1002/hyp.14132