Water budget, performance of evapotranspiration formulations, and their impact on hydrological modeling of a small boreal peatland-dominated watershed

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20 **Abstract**

21 Peatlands occupy around 13% of the land cover of Canada, and thus play a key role in
22 the water balance at high latitudes. They are well known for having substantial water
23 loss due to evapotranspiration. Since measurements of evapotranspiration are scarce
24 over these environments, hydrologists generally rely on models of varying complexity
25 to evaluate these water exchanges in the global watershed balance. This study
26 quantifies the water budget of a small boreal peatland-dominated watershed. We
27 assess the performance of three evapotranspiration models in comparison with *in situ*
28 observations and the impact of using these models in the hydrological modeling of the
29 watershed. The study site (~1-km²) is located in the Eastern James Bay lowlands,
30 Québec, Canada. During summer 2012, an eddy flux tower measured
31 evapotranspiration continuously, while a trapezoidal flume monitored streamflow at
32 the watershed outlet. We estimated evapotranspiration with a combinational model
33 (Penman), a radiation-based model (Priestley-Taylor), and a temperature-based model
34 (Hydro-Québec), and performed the hydrological modeling of the watershed with
35 HYDROTEL, a physically-based semi-distributed model. Our results show that the
36 Penman and Priestley-Taylor models reproduce the observations with the highest
37 precision, while a substantial drop in performance occurs with the Hydro-Québec
38 model. However, these discrepancies did not appear to reduce the hydrological model
39 efficiency, at least from what can be concluded from a 3-month modeling period.
40 HYDROTEL appears sensitive to evapotranspiration inputs, but calibration of model
41 parameters can compensate for the differences. These findings still need to be
42 confirmed with longer modeling periods.

43 **KEYWORDS:** Peatland hydrology; Eddy covariance; Evapotranspiration model;
44 Watershed modeling; Sensitivity analysis.

Introduction

Wetlands are estimated to cover between 5 and 6 million km², equivalent to roughly 4% of the emerged Earth surface (Matthews and Fung, 1987; Aselmann and Crutzen, 1989). In Canada, wetlands cover 14% of the land, of which 90% are peatlands (NWWG, 1997; Price et al., 2005). Canadian peatlands are mostly found in the boreal and subarctic zones (Tarnocai et al., 2005). They are especially dominant in the James Bay and Hudson Bay lowlands, one of the largest peatland area in the world (Gorham, 1991), where peatland is the sole land surface type in many locations (e.g. Glaser et al., 2005; Leclair et al., 2015).

This type of environment is known to transform rapidly under climate changes and fluctuations (Bidgham et al., 2008; Dise, 2009; Runkle et al., 2014), even more so at high latitudes (IPCC, 2013). As northern regions are more and more subject to human development, accurately quantifying the water budget of these environments becomes critical for the perennity of these sensitive ecosystems. With such a high land surface occurrence and sensitivity to climate non-stationarity, boreal peatlands have to be modeled thoroughly by climatologists and hydrologists interested in water pathways across boreal regions.

Considerable efforts must be put into quantifying the actual evapotranspiration (ET_a), as it is the only inherent link between the water and energy budgets, and therefore between hydrology and climatology during snow-free periods. In boreal peatlands, ET_a occupies a fairly variable fraction of the water budget; on an annual basis, cumulative ET_a can amount to between 40% and 85% of the annual precipitation (Verry, 1988; Brümmer et al., 2012). This percentage increases significantly when one considers only the snow-free period, with some examples for

69 which nearly all of the precipitation (P) returns to the atmosphere as ET_a (Isabelle,
70 2014; Runkle et al., 2014). Wu et al. (2010) have shown that ET_a / P also varies
71 greatly between a wet growing season (~62%) and a dry one (~140%). In the James
72 Bay lowlands, ET_a losses are the most important control on freshwater runoff to the
73 saline bay during the summer (Isabelle, 2014). As a result, ET_a has to be properly
74 estimated.

75 However, the precision (related to random error, i.e. variance of the error) and
76 accuracy (related to systematic error, i.e. mean error) of ET_a estimations are often
77 sacrificed at the expense of simpler and faster modeling strategies. This practice raises
78 two questions: (i) which type of model best reproduces ET_a of a boreal peatland? ; and
79 (ii) does a lesser quality in modeled ET_a systematically lead to a poorer agreement
80 between simulated and observed watershed streamflow?

81 A great number of studies have assessed the performance of various ET_a
82 modeling strategies for boreal peatlands, most of them have been reviewed by Drexler
83 et al. (2004). The general consensus is that increasing the complexity of a model leads
84 to a better representation of the ET_a . These authors, however, point out that these
85 more complex models rely on site-specific calibration and the accuracy of the
86 measurements.

87 Meanwhile, several studies have focused on the impacts of the potential
88 evapotranspiration precision (ET_p ; ET when water supply is unlimited at the surface)
89 on hydrological modeling. Andréassian et al. (2004) did a comprehensive review on
90 this subject. Table 1 presents methodological details from the studies listed by
91 Andréassian et al. (2004), but including one more recent paper (Oudin et al., 2005)
92 and this study. The main takeaway is that, most of the time, the precision of input ET_p

93 (and hence of modeled ET_a) does not largely influence the quality of the hydrological
94 modeling. This conclusion seems to be valid whether the temporal (Fowler, 2002;
95 Oudin et al., 2005) and spatial (Andréassian et al., 2004) precisions of ET_p are
96 improved or the precision of the ET_p model is increased (Andersson, 1992;
97 Nandakumar and Mein, 1997; Joukainen, 2000; Kokkonen and Jakeman, 2001; Oudin
98 et al., 2005). Most authors agree that the calibration of the model can compensate for
99 varying ET_p precision. Nevertheless, ET_p models have been shown to be a great
100 source of variability in hydrological predictions under changing climate conditions
101 (Donohue et al., 2010; Seiller and Anctil, 2014). Moreover, some studies found that
102 using a simple temperature-based ET_p model can lead to serious offsets in
103 hydrological predictions, when calibration is virtually impossible (Lofgren et al.,
104 2011; Hoerling et al., 2012).

105 The aforementioned analyses mostly focused on large watersheds and multiple
106 years of data. However, one could think that on reduced temporal and spatial scales,
107 the precision of the ET_a would have a larger influence on the outcome of the
108 hydrological model. Indeed, as smaller scales imply that water flow amplitudes are
109 diminished as a whole, slight variations of one term could substantially affect the
110 water budget, a consideration relevant under operational conditions. An advantage of
111 studying a smaller watershed, even for a short period, is the possibility to use ET_a
112 measurements as input to the hydrological model.

113 Such measurements can be obtained by applying the eddy covariance method,
114 deemed one of the most reliable and accurate approach to measure ET_a (Itier and
115 Brunet, 1996). With an optimal experimental setup, the measured ET_a is then
116 representative of the whole small-scale watershed while accounting for the high ET_a

117 diurnal variability. None of the aforementioned studies have compared the
 118 performance of a hydrological model using sophisticated ET_a measurements or
 119 common, but less-precise ET_p models. Furthermore, to the best of our knowledge, this
 120 type of analysis has not previously been done on boreal peatland-dominated
 121 watersheds. In addition, understanding the impacts of ET_a on the hydrological
 122 modeling of peatlands is critical around James Bay for the ecosystems preservation
 123 and the hydroelectric exploitation of rivers.

124 The research conveyed in this paper quantifies the water budget of a small
 125 peatland-dominated watershed. We have two specific objectives: (i) to compare
 126 several ET_p models with *in situ* observations on a James Bay lowland peatland; and
 127 (ii) to estimate the impact of using ET_p models in lieu of ET_a measurements as inputs
 128 to an hydrological model in that type of environment.

129 **Materials**

130 **Study site: Necopastic peatland**

131 The Necopastic peatland watershed drains 97 ha, of which 63% is occupied by
 132 a peatland (Clerc, 2009), while the rest of the watershed is boreal forest (20%) and
 133 rocky outcrops (17%). The peatland is mostly ombrotrophic, meaning that it receives
 134 all its water and nutrients from precipitation rather than from upstream runoff and
 135 groundwater. Named after the nearby Necopastic River, a tributary to La Grande
 136 River, the Necopastic peatland is located in the James Bay lowlands, northern
 137 Québec, Canada (53°40'28"N; 78°10'14" W, elevation: 105 m ASL). Figure 1a shows
 138 a satellite image of the watershed (Google Earth, 2017) along with the location of the
 139 instruments and other landmarks, with the location in Canada as an inset. Figure 1b
 140 presents a topographical map of the watershed, which has a slope of 0.013 m m⁻¹. The

141 initial watershed boundary was identified by Clerc (2009) using interpretation of
142 airborne image and confirmed on the field with GPS points. As determining
143 watershed extent is especially delicate in low relief environments, the boundaries
144 were further refined by an extended analysis of the digital elevation model of the area.

145 The vegetative cover was thoroughly surveyed by Clerc (2009) and further
146 described by Nadeau et al. (2013). The peatland portion of the watershed is mostly
147 covered with *Sphagnum* mosses, lichens and shrubs, with some scattered spruce,
148 larch, and jack pine trees, 1 to 4 m tall, covering a 1.6 thick peat layer. This peatland
149 is surrounded by a black spruce forest with 6 to 8 m trees. The climate is subarctic,
150 with a strong influence from the nearby James Bay. A mean annual temperature of -
151 2.4 °C and mean annual precipitation of 697 mm (35% as snow) was observed over
152 the period 1977-2011 (Nadeau et al., 2013).

153 **Experimental setup**

154 During summer 2012, the Necopastic peatland watershed was instrumented to
155 measure most components of the energy and water budgets. The main feature of the
156 experimental setup was a 6-m flux tower (see Figure 1c) using the eddy covariance
157 method to measure sensible and latent heat fluxes (H and LE , respectively) and
158 evapotranspiration ($ET_{a,EC}$). It consisted of a three-dimensional sonic anemometer
159 (CSAT-3, Campbell Scientific, USA) equipped with a fine-wire thermocouple and an
160 open-path CO_2/H_2O gas analyzer (LI-7500, LI-COR Biosciences, USA).

161 Turbulent fluxes were computed using EddyPro®, version 5.0 (LI-COR
162 Biosciences, USA), an open-source software designed to process eddy covariance
163 data. The performed data processing is described in Nadeau et al. (2013) and Isabelle
164 et al. (2015). Uncertainties are associated with the use of the eddy covariance method.

Random sampling errors (Finkelstein and Sims, 2002), underestimation of turbulent fluxes with an unclosed energy balance (Foken, 2008), and spatial extrapolation of local measurements to the whole watershed are the most critical. We elaborate on these issues in the *Discussion*.

Net radiation R_n was monitored with a 4-component radiometer (CNR1, Kipp and Zonen, The Netherlands), while soil heat flux G was measured with soil heat flux plates (HFT3, Campbell Scientific, USA). Note that since heat flux plates have been known to perform poorly in *Sphagnum* mosses (Rouse, 1984; Rouse et al., 1987; Halliwell and Rouse, 1987; Petrone et al., 2004), their measurements were verified successfully with the calorimetric method, that is: the sum of the heat energy intake over several layers of a vertical soil temperature profile (Halliwell and Rouse, 1987; Ochsner et al., 2007).

Several other components of the water budget were measured, aside from the ET_a . This budget may be expressed as an equality between incoming (left-hand side) and outgoing and change in stored (right-hand side) water:

$$P = ET_a + Q + S + \text{Residual} \quad (1)$$

where Q is river discharge [mm], and S , change in groundwater storage [mm] over the monitored time interval. Precipitation P [mm] inputs were measured with a tipping-bucket rain gauge in the direct vicinity of the flux tower (CS700H, Campbell Scientific, USA) and cumulated every 30 min.

The water table was monitored with two wells located within a 100-m distance of the flux tower (see Figure 1a,b) and equipped with level sensors (Level Loggers, Solinst, USA). Soil water content in the unsaturated zone was measured with two soil water content reflectometers (CS616, Campbell Scientific, USA). The instruments

189 were installed horizontally at 3 cm and 15 cm depth. As the factory calibration is
190 problematic in peat soils, the instruments were used to compute the apparent dielectric
191 constant of the sampled soil (Hansson and Lundin, 2006). With those values,
192 volumetric water content was obtained with the peat-specific empirical function of
193 Kellner and Lundin (2001).

194 This setup was used to compute the saturated and unsaturated portion of S . The
195 saturated part was taken as the water table height difference between two time steps,
196 multiplied by the soil porosity at the water table depth. The latter was obtained at the
197 3 cm and 15 cm depth as the maximum volumetric water content at saturation, with
198 values of 0.92 and 0.86, respectively. The porosity was assumed constant at 0.86
199 below 15 cm, and a linear interpolation was made between the two known values
200 from 15 cm to 3 cm deep. Water storage in the unsaturated zone was computed taking
201 the difference in water content between two time steps multiplied by the depth of the
202 unsaturated zone. S is then the sum of the unsaturated and saturated parts.

203 The streamflow at the watershed outlet Q was monitored with a 12-in
204 trapezoidal flume (SRCRC, Accura-FloTM, USA) (see Figure 1d). The residual is
205 taken as a water outflow or storage; meaning that a positive value implies that water
206 entering the watershed exceeds that coming out or stored. Obviously, this term
207 accounts for measurement uncertainties, as well as possible other terms that do not
208 appear in Equation 1 such as lateral groundwater flow. However, we note that the
209 incoming part of the latter is usually negligible for ombrotrophic peatlands (Aselmann
210 and Crutzen, 1989). Outgoing lateral groundwater flow could be important, especially
211 in the southwest part of the bog (see Figure 1a) where a channel may form between
212 the ponds present.

Every measured variable was averaged (or summed for P) over a 30-min timescale and stored using a datalogger (CR-5000 for $ET_{a,EC}$ fluxes, CR-3000 for the rest, Campbell Scientific, USA). P and $ET_{a,EC}$ were next summed on hourly and daily time steps for hydrological modeling. The study period extended from 24 June to 27 September 2012. Maintaining such a remote site is very challenging, and as result the experimental period is rather short from a hydrological standpoint. Still, it does not prevent the dataset from being unique and appropriate to answer the proposed scientific questions. The study focuses on summer conditions because it is the period where ET_a has the largest influence on the water budget, especially at these high latitudes.

Methods

Comparison of modeled ET_a with observations

ET_p models

We tested a hierarchy of models of contrasting complexities and development philosophies to estimate ET_p : Penman (1948), Priestley and Taylor (1972) and a temperature-based empirical equation commonly used in hydrological engineering applications in northern Québec (St-Hilaire et al., 2003; Minville et al., 2008, 2009), hereby referenced to as the Hydro-Québec model. As there is a great number of temperature-based ET_p equations with relatively similar efficiencies (Oudin et al., 2005, Seiller and Anctil, 2016), we chose the Hydro-Québec model because it is hypothetically the best-calibrated for the study site region.

The most complex of these models is the Penman model, originally developed to describe evaporation from an open-water surface. The model, also considered applicable over saturated land surfaces, combines energy-balance and mass-transfer

approaches to include contributions from the available energy ($R_n - G$) and unsaturated atmosphere to ET_p . The Penman model can be written as follows:

$$ET_{p,PM} = \frac{1}{L_v} \left[\frac{\Delta}{\Delta + \gamma} (R_n - G) + \frac{\gamma}{\Delta + \gamma} E_A \right] \quad (2)$$

where $ET_{p,PM}$ [$\text{kg m}^{-2} \text{s}^{-1}$] is the water vapor flux; L_v [J kg^{-1}], the latent heat of vaporization of water; Δ [Pa K^{-1}], the slope of saturation vapor pressure versus temperature curve; γ [Pa K^{-1}], the psychrometric constant; R_n [W m^{-2}], the net radiation; G [W m^{-2}], the ground heat flux; and E_A [W m^{-2}], the drying power of the air, defined by Katul and Parlange (1992) as:

$$E_A = \frac{L_v \kappa^2 \bar{\rho} \bar{u} (\bar{q}_* - \bar{q})}{\ln \left(\frac{z_v - d_0}{z_{0v}} \right) \ln \left(\frac{z_m - d_0}{z_{0m}} \right)} \quad (3)$$

where κ is the von Kármán constant ($= 0.4$); $\bar{\rho}$ [kg m^{-3}], the humid air density; $\bar{u}$ [m s^{-1}], the mean wind velocity measured at height z_m [m]; $\bar{q}$ [kg kg^{-1}], the mean specific humidity measured at height z_v [m]; $\bar{q}_* = \bar{q}_*(T_a)$ [kg kg^{-1}], the specific humidity of the air at saturation; and the overbars are time averages over a 30-min period; z_{0m} and z_{0v} [m], the roughness lengths for momentum and humidity, respectively; d_0 [m], the zero-plane displacement height. d_0 , z_{0m} and z_{0v} are estimated with the mean vegetation height at the site ($h_v = 0.5$ m) as $(2/3)h_v$, $0.1h_v$ and $0.01h_v$, respectively (Brutsaert, 1982; 2005).

The Priestley-Taylor model is an empirical simplification of the Penman model valid for wet surfaces under conditions of limited advection. This radiation-based model is often used as the description of equilibrium ET , meaning ET coming from a saturated surface to a saturated atmosphere. It is described as:

$$ET_{p,PT} = \frac{1}{L_v} \left[\frac{\Delta}{\Delta + \gamma} (R_n - G) \right] \quad (4)$$

where $ET_{p,PT}$ [$\text{kg m}^{-2} \text{s}^{-1}$] is also a water vapor flux. Usually, the formulation of $ET_{p,PT}$ includes a scale factor α to convert equilibrium ET to ET_p . We also included an α scale factor to the model, but to convert equilibrium ET to ET_a (more details in the next section).

Hydro-Québec is the government-owned public power utility responsible for large-scale hydropower production in Québec, Canada. To predict reservoir inflows, Hydro-Québec uses a lumped hydrological model named HSAMI (Bisson and Roberge, 1983, Fortin, 2000, St-Hilaire et al., 2003, Minville et al., 2008; 2009). The latter also uses an empirical temperature-based formulation to compute ET_p , hereby designed as Hydro-Québec model:

$$ET_{p,HQ} = \left[0.029718 DTR \exp(0.0342 DTR + 1.216) \right] \quad (5)$$

where DTR [$^{\circ}\text{C}$] is the daily temperature range equal to the difference between the maximum and minimum air temperatures of the day; and $ET_{p,HQ}$ [mm] is the daily cumulative ET_p . To adjust the daily $ET_{p,HQ}$ to an hourly rate, a conversion factor is applied for each hour of the day. This conversion procedure is identical to that used for hourly computations with HSAMI (Fortin, 2000). While daily $ET_{p,HQ}$ has been shown to reproduce observations obtained with energy budget methods fairly well (Dionne et al., 2008; Isabelle, 2014), the conversion procedure to hourly scale adds uncertainty to the already empirical Hydro-Québec equation. Note that this temperature-based ET_p formulation has been shown to seriously overestimate ET_a in future climate predictions (Ludwig et al., 2009).

Conversion to modeled ET_a

For our study, each model described in Equations 2, 4 and 5 is multiplied by a scale factor denoted α . With the three models, α is taken as the ratio ET_a / ET_p to obtain $ET_{a,PM}$, $ET_{a,PT}$ and $ET_{a,HQ}$. In several studies, it is common practice to use such a coefficient for the Priestley-Taylor model, using a value of 1.26 for well-watered surfaces (Priestley and Taylor, 1972; Eichinger et al., 1996). Here, we used α as a calibration parameter to adjust modeled ET_p to observed ET_a values.

In this study, to compare modeled and observed ET_a , we used a fixed value of α for the whole study period. This is a simplification as α is known to vary over time. Over the course of three months, we feel that such variation should be negligible. We tested two of these values for each model: (i) an optimized value (α_{opt}) obtained as the value that minimize the root mean square error between observed ($ET_{a,EC}$) and modeled ET_a values (done for both daily and hourly values); and (ii) a value taken as the average of several α -coefficients obtained in a number of studies on peatland ET_a (α_{lit}) (see the literature review in Table 2). α_{opt} is considered to be the real value of α , while α_{lit} reproduces what a modeler without any direct ET_a observations would probably use. Table 3 presents the α used for each model. Note that as there are not many references on the Hydro-Québec model, we selected an α_{lit} of 0.75, a mean value commonly used in some unpublished HYDROTEL applications.

Performance metrics

To quantify the performance of each modeled ET_a when compared with $ET_{a,EC}$, we used the coefficient of determination R^2 , the root mean squared error (RMSE), the mean bias (MB) and the Nash-Sutcliffe efficiency (NSE) (Nash and Sutcliffe, 1970). R^2 expresses the correlation between modeled and observed ET_a by describing the proportion of the variance in the observations that can be explained by

the model. RMSE is an always positive expression of the error between observations and the model that is given in the units of ET_a , mm. MB differs from RMSE by taking positive or negative values. A perfect model would get both RMSE and MB equal to zero, with an R^2 of 1. RMSE and MB are particularly useful as they can be used to fraction the error between its systematic component (MB^2) and a random component (variance of the error, i. e. $RMSE^2 - MB^2$).

Impacts of ET_a precision on hydrological modeling

Hydrological model: HYDROTEL

To model the Necopastic peatland streamflow, we used the hydrological model HYDROTEL (Fortin et al., 2001a; Turcotte et al., 2003; Turcotte et al., 2007). It is a physically-based semi-distributed model simulating water flows in subsections of the watershed called relatively homogeneous hydrological units (RHHUs). These units are usually delineated using PHYSITEL (Turcotte et al., 2001; Rousseau et al., 2011; Noël et al., 2014), a specialized geographic information system used for watershed and stream network delineation as well as to determine other attributes needed to support distributed hydrological models. To create the RHHUs, PHYSITEL uses a digital elevation model, along with land cover and soil type maps. HYDROTEL has been successfully applied on many midlatitude watersheds (Fortin et al., 2001b; Lavigne et al., 2004; Boucher et al., 2011; Bouda et al., 2012; Ricard et al., 2012; Abaza et al., 2013, 2014; Bouda et al., 2014; Abaza et al., 2015), including the present field site (Clerc, 2009; Jutras et al., 2009). Note that the small Necopastic peatland is modeled with a single RHHU (see Figure 1b).

Once the digital renderings of the watershed landscapes and hydrological routes are made, HYDROTEL simulates six different components of the water cycle: (i) precipitation, by weather data interpolation; (ii) snowpack evolution; (iii) potential

330 evapotranspiration; (iv) soil moisture balance; (v) surface runoff; and (vi) river
331 routing. Each of these components can be simulated with different functions. In our
332 case, we respectively used the following: (i) Thiessen polygons; (ii) no snowpack
333 module necessary (snow-free period); (iii) measured ET_a and the three ET_p models
334 described in the ET_p models section; (iv) three-layer vertical budget; (v) kinematic
335 wave and (vi) kinematic wave. HYDROTEL runs each of these modules sequentially
336 for each RHHU.

337 HYDROTEL can only be given ET_p inputs, from which it will determine ET_a
338 based on soil and vegetation properties. The top soil layer is subject to bare-soil
339 evaporation, computed as a function of ET_p , soil water content and vegetation leaf
340 area index. Soil layers occupied by plant root systems lose water through
341 transpiration, which is calculated as a function of soil moisture and the difference
342 between ET_p and bare-soil evaporation. More details are available in Fortin et al.
343 (2001a).

344 As part of the calibration procedure, HYDROTEL has a built-in set of 26
345 RHHU-independent parameters. One of them is a coefficient to modulate ET_p . Given
346 that this study focuses on the sensitivity of the model to ET_p inputs, the calibration
347 parameter for ET_p within HYDROTEL was set to 1. In our case, the precipitation
348 module used for weather data interpolation is not activated since we have only one
349 weather station (the flux tower, Figure 1c). The snowpack evolution module is also
350 irrelevant given that our study focuses on the snow-free period. Therefore, the number
351 of calibration parameters reduces to 13.

352 Due to the short study period, calibration was performed on the whole dataset
353 (24 June to 27 September 2012), implying that no validation was performed. Taking

the values of calibration parameters found manually by Jutras et al. (2009) for the same watershed as a starting point, we were able to obtain an improved manual calibration for our specific dataset. We obtained this parameter set by a trial-and-error parameter-by-parameter method focused on reducing the NSE between observed and modeled discharges. From this improved calibration parameter set, we used an automatic calibration algorithm to perform the multiple optimization described in the *Sensitivity analysis* section: the dynamically dimensioned search (DDS) algorithm (Tolson and Shoemaker, 2007). In brief, DDS searches for an optimized solution by randomly varying a selection of parameters and keeping only the variations that improve the NSE. The varied parameters are selected randomly, and the size of that varied parameter subset decreases as the algorithm approaches the user-set maximum number of iterations. The simplicity, low computational cost and high performance of DDS makes it ideal for use with HYDROTEL (Huot, 2014).

Given the limited duration of our study period, special care has to be given to the treatment of initial conditions, so that their impact on modeling results are minimized. Here, for each model run, a 1-month spin-up period was used with fixed hydrometeorological inputs. Moreover, the initial conditions of the model are included in the set of calibration parameters, and as such they are optimized. This particular procedure ensures that any modeling artifact related to the initial conditions are minimized. We are aware that the rather short duration of the modeling period is a limitation of the study, and only additional runs spanning over longer time periods will allow confirmation of our conclusions.

376 *Sensitivity analysis*

377 As described in Andréassian et al. (2004), two categories of sensitivity
378 analyses can be used to quantify the impact of the precision of ET_p values on the
379 hydrological modeling: static and dynamic.

380 With the static analysis, a benchmark set of calibration parameter values is
381 obtained with the most precise ET available, in our case $ET_{a,EC}$. The simulated flows
382 obtained with the benchmark calibration can be compared with the observed flows to
383 assess the performance of the simulation. This sets the benchmark performance of the
384 model and is usually done with objective functions (in our case, NSE, see the next
385 section). Once the benchmark calibration was completed, the values of calibration
386 parameters remained constant independently of the ET_p (either one of $ET_{p,PM}$, $ET_{p,PT}$
387 or $ET_{p,HQ}$) model used in subsequent simulations. The performance values of these
388 other simulations were compared with the benchmark performance to assess the
389 sensitivity of the hydrological model to ET_p inputs.

390 The dynamic analysis uses the same benchmark set of calibration parameter
391 values obtained with $ET_{a,EC}$, with associated benchmark performance obtained by
392 comparison of simulated with observed flows. In this instance however, recalibration
393 is permitted, that is the values of the calibration parameters were optimized each time
394 a different ET_p model was used.

395 As Andréassian et al. (2004) mentioned, the two types of sensitivity analysis
396 stem from different modeling philosophies: the static approach assumes the existence
397 of a “true” set of parameter values for a specific watershed, independently of climatic
398 input data, whereas with the dynamic version, the values of the calibration parameters
399 are allowed to “adapt” to varying climate inputs.

400 For this study, we performed both the static and dynamic input-related
401 sensitivity analyses. The static analysis was performed with the benchmark calibration
402 taken as that obtained by DDS using $ET_{a,EC}$ as the input. Since HYDROTEL can only
403 be fed with ET_p values, we generated a synthetic time series, hereby labeled $ET_{p,EC}$.
404 This synthetic ET_p was obtained as the input that ultimately led HYDROTEL to
405 compute a modeled ET_a identical to the observations, $ET_{a,EC}$. Changing only ET_p
406 inputs for subsequent simulations provided information on how sensitive
407 HYDROTEL was with respect to simulated streamflow on small spatiotemporal
408 scales.

409 The dynamic analysis was then performed using DDS to recalibrate the
410 hydrological model each time ET_p inputs were changed. The recalibration allowed us
411 to quantify how HYDROTEL could adapt to these changes in ET_p in terms of
412 calibration parameter values. Both static and dynamic analyses were accomplished at
413 the daily and hourly timescale.

414 ***Performance metrics and comparison criterion***

415 To compare modeled and observed streamflow at the watershed outlet, we
416 used RMSE to assess precision and accuracy of the model and NSE as a skill score.
417 As previously mentioned, the objective function used in both manual and automatic
418 calibration of HYDROTEL is the NSE.

419 To compare performance values of HYDROTEL using $ET_{p,EC}$ with those of
420 HYDROTEL using ET_p models, we need to determine if the differences in
421 performance metrics are significant. This is done with a comparison criteria. For
422 comparisons of NSE, Nash and Sutcliffe (1970) already provided such a criterion
423 named r^2 :

$$r^2 [\%] = \frac{NSE_{\text{mod}} - NSE_{\text{obs}}}{1 - NSE_{\text{obs}}} \times 100 \quad (6)$$

where NSE_{mod} is the NSE obtained with the hydrological model when using modeled ET_p inputs; and NSE_{obs} , the NSE obtained when using $ET_{p,EC}$ inputs. If r^2 is negative, performance of the hydrological model is decreasing with modeled ET_p inputs, and inversely. Senbeta et al. (1999) suggest that values of $|r^2| > 10\%$ indicate a significant difference in performance. Note however that this threshold does not imply a formal statistically significant difference.

To compare RMSE values, we simply use the relative difference between the metrics:

$$\text{RMSE RD} [\%] = \frac{RMSE_{\text{obs}} - RMSE_{\text{mod}}}{RMSE_{\text{obs}}} \times 100 \quad (7)$$

where $RMSE_{\text{obs}}$ is the RMSE obtained with the hydrological model when using $ET_{p,EC}$ inputs; and $RMSE_{\text{mod}}$, the RMSE obtained when using modeled ET_p inputs. Again, negative values indicate a decrease in performance with the use of modeled ET_p as inputs, and inversely.

Results

Measured water budget

Figure 2 presents the components of the water budget (Equation 1) measured at the Necopastic peatland field site. As expected, it highlights the annual cycle of ET_a , peaking at the summer solstice and slowly decreasing towards fall. On 9 September, our instruments measured very low ET_a ; this day was characterized by a constant (but weak) rain all day long, with quite low solar radiation. Such conditions are known to stifle evaporation, and this phenomenon is amplified by shorter days and

446 lower air temperatures. In general, ET_a rates (between 0.5 and 4.3 mm d⁻¹) are in line
447 with those reported for summer at other boreal peatlands (Kellner, 2001; Humphreys
448 et al., 2006; Wu et al., 2010; Brümmer et al., 2012; Runkle et al., 2014).

449 Figure 2 also illustrates how every term of the water budget reacts to
450 precipitation. ET_a decreases during precipitation, but increases shortly after. S and Q
451 are very sensitive to precipitation inputs, with a reaction time within a 30-min period.
452 This reaction time is consistent with the watershed 10 to 20-min concentration time
453 calculated using classical empirical formulations (Kirpich, 1940; Chow, 1962; Watt
454 and Chow, 1985).

455 Streamflows exiting the watershed, Q , range between 0.3 and 6.7 mm d⁻¹,
456 while ET_a rates are between 0.5 and 4.3 mm d⁻¹. In the absence of precipitation, these
457 water outflows are in general very well balanced by a consequent decrease in
458 groundwater storage, leading to almost no residual (i.e. with the average of daily
459 values when $P = 0$: $ET_a = 2.75$ mm; $Q = 0.62$ mm; $S = -3.31$ mm; Residual = 0.06
460 mm). However, during rainfall, the groundwater storage increase exceeds $P - ET_a -$
461 Q , leading to a negative residual (i.e. with the average of daily values when $P \neq 0$: $P =$
462 3.36 mm; $ET = 2.16$ mm; $Q = 0.99$ mm; $S = 1.89$ mm; Residual = -1.68 mm). This
463 lack of equilibrium may be caused by the non-measured presence of groundwater
464 inflows to the peatland; although measurement uncertainties to every water budget
465 term may also be tied to such behavior.

466 To ascertain the influence of groundwater flow to Q , we have analyzed five
467 rainfall events (3, 11, 14, 30 July and 5 August) and performed a hydrograph
468 separation into baseflow and quickflow. We used the standard graphical method
469 outlined by Linsley et al. (1975), where baseflow during the hydrograph rise and fall

470 following a precipitation event is assumed to vary exponentially. Quickflow is
471 presumed to happen between the start of the rainfall event and the inflection point of
472 the recession limb. For these events, the proportion of Q attributable to baseflow
473 varied between 60% and 88%. Also, in the presence of quickflow (between 10 h and
474 32 h after the start of the rainfall events), the cumulated discharge amounted for 10%
475 to 25% of the total precipitation, meaning that the latter mostly infiltrates the peatland
476 and is not directly discharged.

477 The few instances where the residual takes high negative values (e.g. 3 July,
478 17 August, and 7 September) are caused by an important increase in groundwater
479 storage that exceeds the incoming precipitation. Underestimation of incoming
480 precipitation under windy conditions is a well-documented problem (e.g. Rodda and
481 Dixon, 2012) that can be causing this feature. A positive peak of the residual is also
482 noticeable at the end of the season (23 September). On this day, a 20-mm rainfall
483 occurred while the water table was at its highest point of the season (~30 mm below
484 the surface), implying that there may have been runoff not measured by the
485 trapezoidal flume.

486 Figure 3 presents each term of the water budget on a cumulative basis, thus
487 expressing the total seasonal volumes of water carried by each of these processes. The
488 most noticeable feature is that ET_a is the largest term of the water budget with a
489 cumulative value of 224 mm. P follows with 203 mm, and the other terms are much
490 lower (Q , 78 mm; and S , 4 mm). The total seasonal residual is of -106 mm, meaning
491 that measured water fluxes exiting the watershed during the summer also exceed those
492 coming in. Again, a physically-reasonable hypothesis for this phenomenon can be the
493 presence of groundwater flow coming into the peatland from areas where volumes of
494 stored water from earlier snowmelt or rain are larger.

495 **Comparison of modeled ET_a with observations**

496 This part of our analysis describes the performance of each modeled ET_a with
497 respect to the observations from the eddy flux tower. Figure 4 presents side-by-side
498 comparisons of modeled (y-axis) and observed (x-axis) ET_a for the three models
499 (Penman, Priestley-Taylor, Hydro-Québec), two α -coefficients (α_{opt} and α_{lit}), and two
500 time steps (hourly and daily). Evidently, R^2 does not change with the different α -
501 coefficients as it is invariant to a multiplication by a constant. Potential outliers
502 calculated with the Cook's distance method (Cook, 1977; 1979; Montgomery et al.,
503 2001) are identified as "X". Note that these outliers are not removed to calculate
504 performance metrics and linear regressions. Table 4 presents the related performance
505 metrics. For most modeled ET_a and α -coefficients, most of the error is random (i.e.,
506 $MB^2 \sim 0$, $RMSE^2 \sim \text{variance of the error}$), meaning that the different versions of
507 modeled ET_a express a variation in precision while accuracy is almost optimal.
508 Finally, Figure 5 shows the time series of cumulative observed and modeled ET_a for
509 both α -coefficients.

510 The Priestley-Taylor model seems to best represent observed ET_a both at the
511 daily and hourly time scales, closely followed by the Penman model, whereas the
512 Hydro-Québec is markedly behind, as shown by the R^2 , RMSE and NSE values.
513 Indeed, at the daily scale, with α_{opt} , R^2 varies between 0.90, 0.76 and 0.62, RMSE,
514 between 0.50, 0.54 and 0.82, while NSE varies between 0.74, 0.70 and 0.32, for
515 Priestley-Taylor, Penman and Hydro-Québec, respectively (Table 4). The hourly scale
516 yields similar results. However, the Penman model is the least biased (e. g. on a daily
517 scale, MB: -0.10 vs. -0.20 and -0.28 for Priestley-Taylor and Hydro-Québec,
518 respectively). On a daily basis, the Penman model performance are fairly sensitive to
519 outliers. Indeed, if the latter are excluded, some performance metrics increase above

those of the Priestley-Taylor model without its own outliers (e. g. R^2 : 0.83 vs. 0.89;
 RMSE: 0.47 vs. 0.51 mm; NSE: 0.78 vs. 0.74 for Penman without outliers and
 Priestley-Taylor without outliers, respectively).

Figure 4 clearly shows that the Hydro-Québec model obtains the poorest
 agreement with observations, as shown by the lower R^2 values (i.e. daily scale: 0.62;
 hourly scale: 0.69) and the other performance metrics (Table 4), especially on a daily
 scale. These results are expected from a temperature-based modeled ET_a . The model
 however obtains the best linear regression slope on a daily basis (0.99), but the large
 negative y-intercept reflects the important bias of the model.

Our results demonstrate that the α_{lit} scale factors are marginally decreasing the
 quality of estimation of the Penman and Priestley-Taylor models when compared to
 α_{opt} . This can be seen in the values of every performance metrics (see Table 4), where
 daily values decrease marginally while hourly values are still reasonably affected.

Figure 5 displays the differences in seasonal evaporated water at the watershed
 scale between modeled ET_a and observations. On a daily basis, total evaporated water
 of modeled ET_a are much closer to the 224 mm (see dashed horizontal line in Figure
 5) of observed $ET_{a,EC}$ when α_{opt} is used, especially for the Penman and Priestley-
 Taylor models. The Hydro-Québec model is unaffected by the change in α -coefficient
 since values are almost identical (see Table 3). The same can be said for the Priestley-
 Taylor model on a daily basis. An opposite behavior can be seen on an hourly basis,
 where Penman and Priestley-Taylor depths of water are more realistic with α_{lit} . With
 α_{opt} , the worst model is the Priestley-Taylor model, underestimating the total
 evaporated height of water by about 50 mm. It appears to diverge from the Penman

543 model around the middle of August (Figure 5c). This behavior is visible with α_{lit} , and
544 on a daily basis, where it diverges from observations around the same time.

545 **Impacts of ET_a precision on hydrological modeling**

546 Figure 6 shows the hydrographs of observed and modeled streamflow when
547 $ET_{p,EC}$ and each version of modeled ET_p are used as input. Table 5 presents the values
548 of performance metrics for each combination of timescale and ET_p inputs, and for
549 both static and dynamic sensitivity analyses.

550 HYDROTEL reproduces observed discharges fairly well when fed the best
551 representation of ET_p possible (e.g. at the daily scale, with $ET_{p,EC}$, RMSE: 0.32 mm;
552 NSE: 0.89). Figure 6 shows this as well, although the model seems slightly
553 overreactive to precipitation and dry periods. Visual inspection of the hydrographs
554 shows that the curves obtained with modeled ET_p follow the same behavior, showing
555 that the model dynamic remains the same with different ET_p inputs.

556 Table 6 presents a detailed summary of the differences in watershed model
557 efficiency (via the comparison criteria RMSE and r^2), when varying the ET_p inputs
558 with the Penman, Priestley-Taylor and Hydro-Québec models. Bolded values of r^2
559 imply a significant difference according to Senbeta et al. (1999) criteria ($>|10\%|$). The
560 table shows results for both static and dynamic sensitivity analyses.

561 The static sensitivity analysis (Table 6, left side) shows that HYDROTEL
562 performance significantly decreases when ET_p inputs are changed, especially with the
563 Priestley-Taylor and Hydro-Québec models. When the Penman model is used,
564 streamflow simulated with HYDROTEL are in closer agreement with observed flows,
565 but still significantly off. With the modeled ET_p and without recalibration (see dashed
566 colored lines in Figure 6), the hydrographs simulated by HYDROTEL start to diverge

567 from the curve generated with $ET_{p,EC}$ (see dashed black line in Figure 6) around 5
568 July, with a positive shift of around 0.2-0.5 mm d⁻¹, depending on the ET_p model.
569 These results provide information on the internal sensitivity of HYDROTEL to ET_p
570 inputs, which is high without recalibration.

571 This last conclusion does not systematically emerge from the dynamic
572 sensitivity analysis (Table 6, right side). When recalibration is allowed with each ET_p
573 input, it compensates accordingly and bring performance values closer to, or even
574 over those obtained with $ET_{p,EC}$. Indeed, the hydrographs obtained by HYDROTEL
575 with modeled ET_p during the dynamic sensitivity analysis (see solid colored lines in
576 Figure 6) are virtually superimposed over the curve simulated with $ET_{p,EC}$ (black
577 dashed line in Figure 6). Results show that HYDROTEL has an efficiency more
578 comparable to the Priestley-Taylor model than with $ET_{p,EC}$ (e.g., at the hourly scale,
579 RMSE: 0.015 vs 0.015 mm; NSE: 0.88 vs 0.87). Meanwhile, the use of the Penman
580 model creates a very slight drop in HYDROTEL performances. Surprisingly, with the
581 temperature-based Hydro-Québec model, simulated flows have the same agreement
582 with observed flows than with $ET_{p,EC}$ at both daily scale and hourly scale (e.g., at the
583 hourly scale, RMSE: 0.015 mm; NSE: 0.87).

584 Discussion

585 Comparison of modeled ET_a with observations

586 The fact that the Priestley-Taylor model performs similarly to the Penman
587 model on a side-by-side comparison with observations seems to contradict the general
588 idea that the efficiency of a model is proportional to the number of key physical
589 processes considered. Indeed, the Penman model is expected to be more accurate than
590 the Priestley-Taylor model by taking into account the effects of vapor pressure deficit

591 and, hence, air advection: a crucial component of the ET_a process (e.g. Donohue et al.,
 592 2010). However, Drexler et al. (2004) mention that the Penman model can be
 593 questionable when applied to peatlands because of a high surface vegetation
 594 variability.

595 On a Swedish peatland, Kellner et al. (2001) reported that a low coupling
 596 between ET_a and ambient vapor pressure deficit is the reason behind the efficiency of
 597 the Priestley-Taylor model with a fixed α -coefficient. A similar phenomenon happens
 598 at our field site: ET_a and ambient vapor pressure deficit are poorly correlated
 599 ($R^2 = 0.43$). However, a previous study at this particular field site determined that ET_a
 600 has a good correlation ($R^2 = 0.70$) with the difference in humidity between the soil
 601 surface and the air aloft (Isabelle et al., 2015). Thus, the surface humidity appears to
 602 have a much more important control on ET_a than the humidity of the ambient air.
 603 However, the fact that the Priestley-Taylor model neglects air advection is a possible
 604 explanation for the visible differences in ET_a after mid-August, when net radiation
 605 starts to decrease substantially at these latitudes.

606 According to our results, the Hydro-Québec model is the least precise of our
 607 modeled ET_a when compared to observations. However, it seems that the empirical
 608 constants of the Hydro-Québec model and its functional form are well-suited for the
 609 subarctic regions east of the James Bay, since R^2 , RMSE and NSE values are still
 610 reasonable. Isabelle (2014) shows that this model does not work as well on other
 611 peatlands in the St. Lawrence River lowlands or Western Great Plains of Canada,
 612 characterized by much different climates.

613 The good performance of Penman and Priestley-Taylor models with α_{lit} are not
 614 surprising since the respective values of α_{opt} and α_{lit} are relatively close, especially on

615 a daily basis (Table 3). Plus, it appears that our method to obtain α_{opt} on an hourly
616 basis produces values that underestimates total seasonal volumes. If we instead
617 compute α as the ratio between observed and modeled total seasonal volume, we
618 obtain values of 0.59, 0.92 and 0.87 for the Penman, Priestley-Taylor and Hydro-
619 Québec models, respectively. However, these values yield much lower side-by-side
620 performance (results not shown).

621 Nevertheless, this is excellent news to modelers that rarely have access to
622 detailed ET_a measurements for a given boreal peatland site. For them, the use of a
623 fixed α scale factor to convert ET_p to ET_a with values of the order of 0.63 for Penman
624 and 0.84 for Priestley-Taylor seems indicated in boreal peatlands, especially to
625 reproduce seasonal volumes. The same cannot be said for the Hydro-Québec model.
626 Caution needs to be exercised if the temperature-based model is to be used to model
627 ET_a , as a coarse adjustment can lead to serious seasonal volume offset.

628 **Impacts of ET_a precision on hydrological modeling**

629 In some cases, our results show that HYDROTEL can generate slightly more
630 representative streamflows with ET_p models than with observations. The ensuing
631 HYDROTEL efficiency highlights the capacity of the calibration parameter values to
632 adapt to different hydrometeorological inputs. For example, despite some important
633 differences in precision between $ET_{a,EC}$ and $ET_{a,HQ}$ (e.g., at the hourly scale and with
634 α_{opt} , RMSE: 0.063 mm; NSE: 0.63), recalibration of HYDROTEL provides a way to
635 find a set of parameter values capable of compensating for those differences,
636 providing satisfying streamflows. This adaptive behavior can be generalized to all
637 studied ET_p models. This feature is especially highlighted in Figure 6, where flows
638 generated during the dynamic analysis with every ET_p models are virtually identical to
639 those obtained with $ET_{p,EC}$. Of course, the calibration procedure becomes very

640 efficient considering that it is performed using the whole sampling period. Our
641 observations may not hold at larger time scales. However, the results follow a similar
642 trend at the hourly scale, where a much larger data sample is used.

643 It is also important to remember that $ET_{a,EC}$ observations are not a perfect
644 representation of ET_a . The method of Finkelstein and Sims (2001) estimates that eddy
645 covariance uncertainties related to random sampling errors amount to about 12% of
646 the season total evaporated water. Furthermore, some uncertainties also lie in the
647 assumption that $ET_{a,EC}$ is representative of the whole watershed ET_a . Using the Kljun
648 et al. (2004) model, the flux tower mean footprint for the whole season is of the order
649 of 1 ha while the watershed area is 97 ha. Our assumption stems from the fact that
650 most of the Necopastic peatland watershed is covered with the same type of surface as
651 that found within the footprint of the flux tower (peatlands covering 63% of the
652 watershed area). The other land covers of the watershed area, mostly boreal forest,
653 probably have different ET_a rates. Another possible source of uncertainties for eddy
654 covariance measurements lies in a well-documented problem: the failure to close the
655 energy balance (Foken, 2008). Our observations show that turbulent fluxes ($H + LE$)
656 account for only 70% of the total available energy ($R_n - G$) for our whole study
657 period, a value within the range of turbulent fluxes observed at other sites (Wang and
658 Dickinson, 2012). This means that $ET_{a,EC}$ may underestimate actual rates. These
659 uncertainties on observed ET_a may play a role in the improvements of HYDROTEL
660 efficiency using ET_p models.

661 Several other studies have reported that the precision of ET_p values does not
662 excessively affect the efficiency of a given hydrological model (see Table 1).
663 However, none of them have used ET_a observations as the best estimates to model
664 streamflow. According to our results, the modeled ET_a are reasonably close to the

665 observed rates. These estimations are good enough for short-term streamflow
666 simulations, especially considering that the calibration of the hydrological model is
667 able to adapt to different ET_p inputs. Discrepancies in observed and modeled ET_a rates
668 are also uninfluential compared to precipitations driving the hydrological model.

669 This study focused on much smaller spatiotemporal scales than what was
670 previously reported. The fact that the time of concentration of the watershed is
671 relatively short required an hourly computational time step. ET_a rates vary greatly
672 during the day, meaning that hourly estimates could be relevant for streamflow
673 modeling. This is especially true for peatlands, known to lose a lot of water through
674 ET_a . At the hourly timescale, the modeled ET_a reproduced observations with good
675 precision, albeit not as good as at the daily timescale. This good performance is seen
676 even for the empirical Hydro-Québec model, which was originally designed for a
677 daily timescale. Meanwhile, HYDROTEL has no difficulty to cope with different
678 hourly ET_p inputs for modeling hourly streamflow during a single summer.

679 These results appear to confirm those of previous studies summarized in Table
680 1. The precision of ET_p does not seem to have an impact on the efficiency of
681 streamflow modeling, even at smaller spatial and temporal scales, and even if precise
682 ET_a observations are used as benchmark. These results are more indicative than
683 conclusive, given that they were obtained at a single field site over a relatively short
684 study period.

685 On the surface, this seems to point to the use of more simple temperature-
686 based ET_p models for hydrological modeling or forecasting. However, these models
687 need to be used cautiously as their calibration (i.e., empirical constants and functional
688 forms) still needs to be optimal. Spatiotemporal transferability of the calibration of

689 these temperature-based ET_a formulations are also questionable (Mauser and Bach,
690 2009; Lofgren et al., 2011; Hoerling et al., 2012), including for the Hydro-Québec
691 model (Ludwig et al., 2009).

692 Conclusion

693 The goal of this study was to compare three ET_p models of contrasting
694 complexities with precise field observations, and to assess the impacts of using these
695 models on the efficiency of a physically-based hydrological model. The study focused
696 on a small watershed (0.97 km²) dominated by a boreal peatland, a type of
697 environment prone to substantial evapotranspiration. Our analysis relied on an
698 original dataset of eddy covariance fluxes and other meteorological and hydrological
699 variables measured during summer 2012.

700 As expected, in terms of side-by-side comparisons between observed and
701 modeled ET_a , the combinational and radiation-based models (Penman and Priestley-
702 Taylor) had better performance than the temperature-based Hydro-Québec model.
703 However, a previously reported low coupling between ET_a and ambient vapor
704 pressure deficit benefited the Priestley-Taylor model when compared to the more
705 comprehensive Penman model. Nevertheless, the Priestley-Taylor model seemed to
706 diverge from observed data by the end of the summer, seriously underestimating ET_a .
707 Results of the Penman and Priestley-Taylor models were adjusted from ET_p to ET_a
708 according to an α -coefficient coming from a literature review on boreal peatlands
709 (Penman: 0.63; Priestley-Taylor: 0.84), but it did not affect the models respective
710 efficiencies, compared to an optimized α -coefficient. On an hourly basis, it was even
711 a better value to compute seasonal evaporated volumes.

712 While each ET_p model reproduced observed values with different precisions,
713 they did not have an important impact on the hydrological modeling of streamflows at
714 the watershed outlet when compared to those simulated using ET_a observations. A
715 static sensitivity analysis established that HYDROTEL can be very sensitive to ET_p .
716 This sensitivity was greatly diminished in the dynamic sensitivity analysis, when
717 recalibration was performed with each ET_p inputs. In other words, the sensitivity was
718 almost nullified by adjusting the values of the calibration parameters, at least on a
719 very short hydrological modeling period. Altogether, the results of this study
720 illustrated that the set of calibration parameters has a substantial short-term capacity
721 to offset the precision of ET_p inputs.

722 This study illustrates that, for a small peatland-dominated watershed and a
723 short modeling period, the precision of ET_p inputs does not seem to affect the
724 modeling of streamflows at the outlet with HYDROTEL. This conclusion still needs
725 to be confirmed with a longer modeling period to affect operational modeling
726 strategies. The simplicity of a temperature-based ET_p model appears enticing,
727 provided that it is properly calibrated. However, the transposability of the above
728 results to larger spatiotemporal scales still needs to be confirmed.

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Draft

1060 **Tables**1061 Table 1: Literature review of the impacts of ET_a or ET_p formulations in hydrological modeling.

Reference	Location(s) (Number of sites)	Area [km ²]	Variations in ET	Impacts on hydrological modeling performances
Parmelee (1972)	Pennsylvania, U.S.A. (5), North Carolina, U.S.A. (4)	8-534	Random and systematic errors	Bias has an impact, but not random fluctuations
Andersson (1992)	Sweden (3)	4216-8484	7 different ET_p models	Similar performances
Paturel et al. (1995)	Ivory Coast (5)	368-4700	Systematic errors in ET_p and P	Smaller error with ET_p than with P
Nandakumar and Mein (1997)	Victoria, Australia (22)	0.02-5.2	Systematic error in ET_p and P	Smaller error with ET_p than with P
Joukainen (2000)	Finland (1)	14000	Added parameters to convert ET_p to ET_a	Similar performances
Kokkonen and Jakeman (2001)	North Carolina, U.S.A. (1), Western Australia (1)	0.5-0.8	Different conversion from ET_p to ET_a	Similar performances
Fowler (2002)	Auckland, New Zealand (1)	N/A	Actual ET_p vs. long-term average ET_p	Similar performances
Vázquez and Feyen (2003)	Belgium (1)	586	3 different ET_p models	Strong performance differences
Andréassian et al. (2004)	Massif Central, France (62)	5-89	Improved areal ET_p representation	Similar performances
Oudin et al. (2005)	Australia (8), U.S.A. (79), France (221)	5-9387	Interannual mean ET_p vs. temporally varying ET_p	Similar performances
This study	Québec, Canada (1)	0.97	ET_a measurements vs. 3 ET_p models	Similar performances

1062

Table 2: Review of α scale factors (taken as ET_a/ET_p) for the Penman and Priestley-Taylor models for

1063

various peatlands.

Penman model		
Reference	Study site	α
Koerselman and Beltman (1988)	Quaking fens (The Netherlands)	0.77
Lafleur and Roulet (1992)	Hudson Bay Lowland fens (Canada)	0.75
Campbell and Williamson (1997)	Raised peat bog (New Zealand)	0.34-0.77
Kellner (2001)	Stormossen mire bog (Sweden)	0.61-0.77
Lafleur et al. (2005)	Mer Bleue bog (Canada)	0.44-0.59
Humphreys et al. (2006)	West-East range of peatlands (Canada)	0.38-0.64
Wu et al. (2010)	Salmisuo mire, fen and bog (Finland)	0.59-0.69
	Average α_{lit} used in this study	0.63
Priestley-Taylor model		
Reference	Study site	α
Kellner (2001)	Stormossen mire bog (Sweden)	0.71-0.85
Petrone et al. (2004)	Bois-des-Bel restored peatland (Canada)	0.67-0.87
Admiral et al. (2006)	Mer Bleue bog (Canada)	0.50-1.50
Humphreys et al. (2006)	West-East range of peatlands (Canada)	0.82-1.05
Petrone et al. (2007)	Wetland-Pond complex (Canada)	0.69
Parmentier et al. (2009)	Horstermeer Polder peatland (The Netherlands)	0.75
Sonnentag et al. (2010)	Sandhill fen (Canada)	0.79-1.04
Brümmer et al. (2012)	Western Peatland and Mer Bleue bog (Canada)	0.60-0.80
Runkle et al. (2014)	Ust-Pojeg mire complex raised bog (Russia)	0.97-1.06
	Average α_{lit} used in this study	0.84

1064

1065 Table 3: Values of α scale factors (ET_a / ET_p) used for each model.

Daily scale		
Model	α_{opt}	α_{lit}
Penman	0.56	0.63
Priestley-Taylor	0.84	0.84
Hydro-Québec	0.76	0.75
Hourly scale		
Model	α_{opt}	α_{lit}
Penman	0.49	0.63
Priestley-Taylor	0.69	0.84
Hydro-Québec	0.69	0.75

1066

1067 Table 4: Performance metrics between observed and modeled ET_a values with α_{opt} and α_{lit} .

Daily scale (α_{opt} ; α_{lit})				
Model	R ²	RMSE [mm]	MB [mm]	NSE
Penman	0.76	0.54 ; 0.62	-0.10 ; 0.17	0.70 ; 0.61
Priestley-Taylor	0.90	0.50 ; 0.50	-0.20 ; -0.19	0.74 ; 0.74
Hydro-Québec	0.62	0.82 ; 0.82	-0.28 ; -0.30	0.32 ; 0.32
Hourly scale (α_{opt} ; α_{lit})				
Model	R ²	RMSE [mm]	MB [mm]	NSE
Penman	0.82	0.048 ; 0.062	-0.017 ; 0.007	0.78 ; 0.64
Priestley-Taylor	0.87	0.047 ; 0.056	-0.025 ; -0.007	0.79 ; 0.71
Hydro-Québec	0.69	0.063 ; 0.064	-0.019 ; -0.013	0.63 ; 0.62

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1070 Table 5: Performance metrics between observed and modeled streamflow when using various ET_p
1071 observations and models.

Daily scale					
Static sensitivity analysis			Dynamic sensitivity analysis		
Model	RMSE [mm]	NSE	Model	RMSE [mm]	NSE
Observed	0.32	0.89	Observed	0.32	0.89
Penman	0.37	0.85	Penman	0.34	0.87
Priestley-Taylor	0.51	0.72	Priestley-Taylor	0.32	0.89
Hydro-Québec	0.48	0.75	Hydro-Québec	0.32	0.89
Hourly scale					
Static sensitivity analysis			Dynamic sensitivity analysis		
Model	RMSE [mm]	NSE	Model	RMSE [mm]	NSE
Observed	0.015	0.87	Observed	0.015	0.87
Penman	0.017	0.84	Penman	0.016	0.86
Priestley-Taylor	0.022	0.73	Priestley-Taylor	0.015	0.88
Hydro-Québec	0.021	0.75	Hydro-Québec	0.015	0.87

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1073 Table 6: Performance metrics difference between observed and modeled streamflow when using
1074 various ET_p models.

Daily scale					
Static sensitivity analysis			Dynamic sensitivity analysis		
Model	RMSERD [%]	r^2 [%]	Model	RMSERD [%]	r^2 [%]
Penman	-13.89	-31.21	Penman	-5.56	-15.16
Priestley-Taylor	-58.33	-150.31	Priestley-Taylor	0.00	1.17
Hydro-Québec	-50.00	-123.68	Hydro-Québec	0.00	-2.60
Hourly scale					
Static sensitivity analysis			Dynamic sensitivity analysis		
Model	RMSERD [%]	r^2 [%]	Model	RMSERD [%]	r^2 [%]
Penman	-12.20	-24.96	Penman	-4.88	-9.43
Priestley-Taylor	-46.34	-117.27	Priestley-Taylor	2.44	2.93
Hydro-Québec	-39.02	-98.57	Hydro-Québec	0.00	-0.24

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1076 Figures

1077 Figure 1: (a) Satellite image of the watershed (adapted from Google Earth, 2017) with instruments
1078 location; (b) Topographical map of the watershed with instruments location; (c) Flux tower at the site
1079 (20 July 2012); (d) Trapezoidal flume at the outlet (adapted from Clerc, 2009).

1080 Figure 2: Seasonal variation of each fluxes of the water budget at the Necopastic peatland watershed.
1081 Air temperature T_a is shown on the first plot (precipitation P) and water table depth WTD is shown on
1082 the 4th plot (groundwater storage change S).

1083 Figure 3: Seasonal variation of cumulative fluxes of water at the Necopastic peatland watershed.

1084 Figure 4: Comparison between modeled and observed ET_a for: (a) daily and (b) hourly scales. The first
1085 column is Penman; the second, Priestley-Taylor, and the third, Hydro-Québec. The first row of each
1086 timescale uses α_{opt} for model computation, while the second row uses α_{lit} . The solid lines are least
1087 squared regressions of the data with statistics shown on the graphs, while the dashed lines are 1:1 lines.
1088 Potential outliers are shown with a cross ("x").

1089 Figure 5: Time series of cumulative $ET_{a,EC}$ and modeled ET_a using: a) α_{opt} (daily time steps); b) α_{lit}
1090 (daily time steps); c) α_{opt} (hourly time steps); and d) α_{lit} (hourly time steps). The dashed line is the total
1091 cumulative evaporated water from observations $ET_{a,EC}$ (224 mm).

1092 Figure 6: Hydrographs of observed and modeled streamflow. The latter are those obtained when using
1093 $ET_{p,EC}$ (black dashed) and every version of modeled ET_p (Penman: blue; Priestley-Taylor: red; Hydro-
1094 Québec: grey), during either static (dashed lines) and dynamic (full lines) analysis. Precipitation is
1095 shown on top histogram.