Taliks: A tipping point in discontinuous permafrost degradation in peatlands

Élise G. Devoie\textsuperscript{1}, James R. Craig\textsuperscript{1}, Ryan F. Connon\textsuperscript{2}, William L. Quinton\textsuperscript{2}

\textsuperscript{1}Department of Civil and Environmental Engineering, University of Waterloo, Waterloo, ON, Canada, N2L 3G1
\textsuperscript{2}Cold Regions Research Centre, Wilfrid Laurier University, Waterloo, ON, Canada, N2L 3C5

Key Points:

• The formation of an isolated talik results in accelerated and likely irreversible permafrost thaw
• A combination of field data and 1-D modelling is used to investigate the formation of isolated taliks
• Soil moisture, snowcover, and advection affect the formation and development of isolated taliks

Corresponding author: Élise Devoie, egdevoie@uwaterloo.ca
Abstract
Taliks (perennially thawed soil in a permafrost environment) are generally found beneath water bodies or wetlands, and their development and evolution in other environments is poorly documented. Sustained isolated taliks between frozen surface soils and permafrost have been observed at the Scotty Creek Research Station in the discontinuous permafrost region of the Northwest Territories, Canada. These taliks have been expanding both vertically and laterally over the past decade of monitoring. The main controls on expansion are thought to be 1) the availability of energy, determined by incoming radiation and advective heat flux, 2) the ability to transfer this energy to the freezing/thawing front, determined by the thermal conductivity (soil properties and moisture content), and 3) the presence and thickness of the snowpack. These controls are investigated using data collected in the field to inform a 1-D coupled thermodynamic freeze-thaw and unsaturated flow model. The model was successfully used to represent observed thaw rates in different parts of the landscape. It is found that high soil moisture, deeper snowpacks, and warmer or faster advective flow rates all contribute to accelerated talik growth and subsequent permafrost degradation. Simulations show that slight perturbations of available energy or soil properties, such as an increase in average surface temperature of 0.5 °C or a 1 cm change in SWE, can lead to talik formation, highlighting the vulnerability of this landscape to changes in climate or land cover.

1 Introduction

Permafrost regions are very sensitive to changes in climate, especially those classified as discontinuous or sporadic (Stendel & Christiansen, 2002; Zhao et al., 2010; Chasmer et al., 2011). Climate warming trends have been shown to cause permafrost degradation and loss, resulting in subsidence, wetland expansion, and landscape transition (Rowland et al., 2010; Walvoord & Kurylyk, 2016; Carpino et al., 2018). One of the mechanisms for permafrost degradation is cited as ‘active layer thickening’, driven by increases in mean annual air temperature, precipitation, and anthropogenic or natural land cover change or disturbance (Shiklomanov et al., 2012; Bonnaventure & Lamoureux, 2013). This process is indicative of permafrost degradation in continuous permafrost, where the active layer, or layer which freezes and thaws annually, is defined by the late summer maximum depth of thaw (Burn, 1998). However, in areas where permafrost is degrading, especially at the southern limit of permafrost, the active layer can either be determined by the depth
of thaw or the late winter refreeze depth if a talik exists between the permafrost table and the active layer (Connon et al., 2018). In the second case, the depth of thaw exceeds the refreeze depth, leaving a perennially thawed region between the base of the active layer and the top of the degrading permafrost body.

Taliks are typically documented beneath or adjacent to water bodies such as wetlands or lakes (Bonnaventure & Lamoureux, 2013; Rowland et al., 2010; Woo, 2012), while the formation and evolution of shallow taliks isolated from wetland features (hereafter referred to as isolated taliks) have received relatively little attention. Though briefly mentioned in field literature (e.g., Fisher et al. (2016)), the factors controlling the formation of isolated taliks have not been thoroughly investigated in thermal modelling literature. A comprehensive review of current thermal models was completed by (Kurylyk & Watanabe, 2013); talik modelling was not mentioned. This omission is likely due in part to scale, where large scale models do not resolve the relatively local process of talik formation, e.g. Stendel and Christiansen (2002).

Freeze-thaw models based on analytic or semi-analytic solutions of the Stefan problem (e.g. Hayashi et al. (2007); Hinzman et al. (1998); Krogh et al. (2017); Semenova et al. (2013); Woo et al. (2004); Zhang et al. (2003)) are unable to represent the three-tiered system (permafrost-talik-active layer) present in talik formation. These models may be inappropriate for modelling degrading permafrost at the local scale because they assume a linear temperature profile between the soil surface and the (single) freeze/thaw front. Existing continuum models (that do not assume a linear temperature profile; e.g. McKenzie and Siegel (2007); Zhang et al. (2008); Frampton et al. (2012); Daanen et al. (2008); Karra et al. (2014); Schaefer et al. (2009); Jorgenson et al. (2010); McGuire et al. (2016)) have thus far focussed on longer-term lateral permafrost extent, water seepage, carbon storage, permafrost resilience or other processes that do not distinguish controls on isolated talik formation or evolution in peatland environments.

The initiation of an isolated talik has been simulated by Frampton et al. (2012); Atchley et al. (2016); Yi et al. (2014); Endrizzi et al. (2014); Rawlins et al. (2013); Jafarov et al. (2013); Brown et al. (2015) and Walvoord et al. (2019), where controls on the active layer such as saturation, snow cover, shading, forest fire and ponded water were quantified. These studies shed light on processes governing active layer thickness as defined by maximum thaw depth, but talik formation was not their focus. Nickolsky et al.
(2016) simulated talik formation by the end of the century in the Alaskan North Slope under RCP 4.5 and 8.5 greenhouse gas scenarios, driven by changes in climate forcing, but moisture, snowpack and incoming radiative effects were not distinguished. Evans and Ge (2017) simulated supra-permafrost layer thickening due only to increased mean annual air temperature in the aim of quantifying the effect of changes in frozen ground regimes on groundwater discharge.

The modelling efforts presented in this paper focus on the formation, evolution and persistence of isolated taliks at the local scale. This work aims to build on existing work in discontinuous permafrost presented by Kurylyk et al. (2016) and Langford et al. (2019). Kurylyk et al. (2016) detail the lateral permafrost thaw and vertical thaw at the bottom boundary of permafrost bodies while Langford et al. (2019) simulated a single peat plateau in the research basin. The detailed 3-D modelling work presented in these studies did not address vertical thaw at the top of a permafrost body, or the formation of isolated taliks within permafrost plateaux.

It is proposed here that talik formation plays an important threshold-based (as defined by Grosse et al. (2016)) thermodynamic role in the initiation of permafrost degradation at the interior of permafrost bodies in discontinuous permafrost regions, and ultimately the hydrologic evolution of permafrost environments in the peatlands region of the southern Taiga Plains, Canada. The impacts of unsaturated soil conditions, moisture migration due to temperature and pressure gradients, lateral advection, and insulation due to variations in annual snowfall will be considered to explicitly identify drivers for talik formation. The subsequent permafrost degradation rate once a talik is formed is also presented for this ecosystem-protected permafrost environment. The objectives of this study are to (1) use a 1-D model to identify the conditions under which isolated taliks are likely to form, (2) assess the extent and controls on the rate of talik formation, and their consequences for permafrost evolution and (3) evaluate the mechanisms by which talik formation can lead to a ‘tipping point’ condition at which permafrost recovery becomes unlikely.

2 Study Site

Field studies were completed at the Scotty Creek Research Station (SCRS), located approximately 50 km south of Fort Simpson (MAAT -3.2 °C) in discontinuous permafrost
peatlands (described in Quinton et al. (2019)). The site is ideal for the study of permafrost degradation because it includes permafrost in a variety of degradational stages including stable permafrost features, features with isolated and connected taliks, and permafrost-free wetland features (Connon et al., 2018). The peat deposit in this study site ranges from 2 m to 8 m in thickness, and overlays clay, silt/clay, and low-permeability glacial till (Quinton et al., 2019). High permeability peat soils drain readily when the water table is below the ground surface, leaving a relatively dry insulating surface peat layer that preserves permafrost underlaying peat plateaux (Quinton et al., 2009). Though insulated from high summer temperatures, the permafrost in this discontinuous permafrost site is warm, and a significant portion of it is within the zero curtain (undergoing phase change) as shown in figure 1(a). Peat plateaux are elevated above the surrounding permafrost-free wetlands due to the subsidence which accompanies permafrost thaw, and the loss of segregated ground ice. Plateaux have a relatively dry vadose zone, allowing them to support a black spruce canopy (Quinton et al., 2009). These plateaux are surrounded by two wetland types that dominate the landscape: collapse scar bogs and poor fens that act mainly as water storage features, and channel fens that act as the low-gradient routing feature in this flat, high-storage landscape (Quinton et al., 2009; Gordon et al., 2016).

In this landscape of ephemerally interconnected wetlands, lateral movement of water through the portions of the active layer that remain saturated most of the year is common, and may help explain variability in permafrost degradation rates across the landscape (Connon et al., 2015). Peat plateaux adjacent to wetland features have been observed to degrade more quickly (both vertically and laterally) than plateaux with isolated taliks (McClymont et al., 2013; Baltzer et al., 2014); it is postulated that lateral advection through the talik plays a considerable role in determining the rate of permafrost degradation.

3 Methods

3.1 Field Methods

This study focuses on the application of a 1-D vertical freeze/thaw model informed by boundary conditions and validation data collected at the SCRS to simulate a set of representative soil columns in a discontinuous permafrost environment. Temperature data were measured using Campbell Scientific CS107 or CS109 thermistors or Onset HOBO
U12 4-channel thermistors and loggers. These were installed in depth profiles approximately every 5 - 20 cm from a depth of 5 cm to 50 cm, where the observation data used to generate boundary condition were measured at either a depth of 5 cm or 10 cm. Exact measurement depths and spacing varied across the measurement sites.

The permafrost that underlays peat plateaux in the landscape can be classified either as degrading or stable. The degrading boundary condition describes a plateau with a talik that remains perennially thawed. The stable boundary condition describes a plateau in which the permafrost is not actively degrading, and there is complete refreeze of the active layer in most if not all years. Figure 1(b) shows surface temperature data collected in both conditions from which surface temperature boundary conditions were constructed.

Data to inform the mean and variance of the stable temperature boundary condition were collected at five different sites in the area of interest, three of which were located on the same permafrost plateau (Indicated as white circles on figure 1(c)). Sites with moss and lichen ground cover were likewise represented; at least three of the sites developed a talik before the end of the data record. Data up to two years preceding talik development were included in the dataset used to generate boundary conditions for the stable condition. Talik formation was inferred if any one of the thermistors in the vertical profile did not drop below the zero curtain for 5 or more consecutive days. The data after the formation of a talik was combined with temperature data collected along an abandoned winter road (an anthropogenic cut line where permafrost has degraded visible in figure 1(c)) to inform the degrading temperature boundary condition (those representing unsteady warming conditions) as observed in figure 1(b). Both field data time-series included data gaps due to instrument malfunction, but provided a roughly 10-year data record.

Deep soil temperature data were collected at two sites with degrading permafrost using RBR deep thermistors with one metre spacing from 2 m to 8 m below the ground surface white circles on figure 1(c). The temperature of lateral flow through a talik was assigned based on the average daily temperature at 40 cm - 50 cm below the ground surface in wetlands in the landscape. Subsurface temperature data for a bog were only available for a single sampling location (labelled ‘bog station’ in figure 1(c)), though these data were consistent over the more than 10 year period of record. Similar data were available for a channel fen (the ‘fen station’), though the period of record was only four years,
and contained several data gaps. Temperature data collected at three other locations in a fen using HOBO U20 pressure transducers and temperature loggers were consistent with this shorter data series. Lateral advective flow rates ($q_y$) were estimated based on available hydraulic gradients measured using sets of HOBO U20 pressure transducers installed in adjacent features (shown as white circles in figure 1(a)). Soil moisture boundary conditions were based on measured moisture content at a depth of 20 cm using Campbell Scientific CS 615 and Campbell Scientific CS 616 water content reflectometers located at meteorological station indicated as black triangles in figure 1(a). Probes were calibrated according to the manual using soil samples collected at the SCRS. Model parameters such as saturated hydraulic conductivity, porosity, thermal conductivity, field capacity, and others listed in table S1 of the supplementary material were fixed at representative values, either measured in the field, or found in literature (e.g. thermal conductivity of peat), without model calibration.

Validation data included temperature data measured in the same locations as above, but at a depth of 50 cm, as well as frost table measurements along 9 transects and one grid, (black lines and grid in figure 1(c)) documenting lateral and vertical permafrost thaw rates. Monitoring points were located along transects traversing permafrost plateaux and intersecting the border of bogs and fens in order to contrast permafrost degradation adjacent to different wetlands. The transects were established in 2011, and permafrost degradation along these transects has been measured annually since 2015, as described in detail by Connon et al. (2018).

In typical stable permafrost environments where the ground annually refreezes to the permafrost table, active layer depth can be measured at the end of the thaw season, but taliks are prevalent at the SCRS, resulting in an active layer depth that can only be determined by knowing the (hard to measure) refreeze depth. In the instance of a talik, the active layer is therefore measured in early spring before ground thaw begins by drilling through the frozen soil to find the base of the frozen layer as detailed in Connon et al. (2018).

3.2 Modelling Methods

Two coupled differential equations are used here to represent the problem of heat and water transport in freezing soils. These equations are solved in one dimension us-
Figure 1. (a) Trumpet plot for ground temperatures measured at the SCRS, indicating maximum and minimum soil temperature at various depths. Dashed line illustrates the theoretical curve for this site. Data below the talik sit within the freezing point depression indicating permafrost undergoing phase change. (b) Range of soil temperatures measured on a stable peat plateau (stable) and a degrading peat plateau in the presence of a talik (degrading) at a depth of 5 - 10 cm below soil surface. Note that the moisture content increases with the formation of a talik, which is reflected in the duration of the zero-curtain periods. (c) Map of study site including frost table transects and grid, thermistor sites where soil temperature is measured in profile, water level recorders used to establish hydraulic gradient and subsurface temperature and meteorological stations monitoring a suite of climate variables including soil temperature and moisture described in Field Methods section (3.1).

The unsaturated Richards’ equation (represented in mixed form here, similar to Celia et al. (1990)):

\[
\left( \frac{\partial \theta_l}{\partial \psi} + S_s \right) \frac{\partial \psi}{\partial t} + \left( \frac{\rho_i}{\rho_l} \theta_i + \theta_l \right) \frac{\partial F(T)}{\partial t} = \frac{\partial}{\partial z} \left( K(\psi, T) \frac{\partial \psi(z, T, \theta)}{\partial z} \right)
\]

where \( \theta \) [-] represents water saturation, \( \psi \) [m] is soil matric potential, \( S_s \) [m\(^{-1}\)] is specific storage, \( t \) [s] is time, \( \rho \) [kg m\(^{-3}\)] is density, \( F \) [-] is the temperature-dependant ice fraction, \( T \) [°C] is temperature, \( z \) [m] is vertical distance, and \( K \) [m s\(^{-1}\)] is hydraulic conductivity. The hydraulic conductivity is given using the van Genuchten model with peat-
specific parameters (van Genuchten, 1980). The hydraulic conductivity is modified by an empirical relationship describing the impedance of ice content to water movement through the partially-saturated soil, as presented by Kurylyk and Watanabe (2013). The subscripts \(l\) and \(i\) refer to liquid water and ice phases respectively. The less common term \(\left(\frac{\varepsilon}{\varepsilon_{th}} + \theta_l\right)\frac{\partial F(T)}{\partial t}\) on the LHS is a source/sink term arising from the inclusion of phase change. Vapour flux is not included in this model; Putkonen (1998) deemed it unimportant (especially in wet soils) as compared to the other processes occurring in an unsaturated 1-D freezing or thawing soil column. The main role of vapour flux is to deliver water to the freezing front, which is accomplished instead by the Clausius Clapeyron (CC) relation (Karra et al., 2014). The CC relation is included in the \(\frac{\partial \theta_l}{\partial \psi}\) term in equation 1 when the temperature is in the freezing range, assumed to be 0 °C to -0.05 °C (equation 2). This equation is very similar to the form of the Clausius relationship presented in Kurylyk and Watanabe (2013). The terms \(g\), \(L\) and \(\Delta \rho^{-1}\) refer to gravity [\(\text{m s}^{-2}\)], the latent heat of fusion of water [\(\text{J kg}^{-1}\)] and the change in specific volume associated with the phase change [\(\text{m}^3 \text{ kg}^{-1}\)] respectively.

\[
\frac{\partial \theta_l}{\partial \psi} = \frac{d\theta_l}{d\psi} \frac{d\psi}{dz} = \frac{d\theta_l}{d\psi} \left(1 - \frac{\partial \psi(\theta)}{\partial z} + \frac{1}{\rho g T \Delta \rho^{-1}} \frac{\partial T}{\partial z}\right)
\]

Heat transport in the porous media is governed by:

\[
\left[\frac{c_p}{\rho} + L \frac{dT}{dT} \rho \theta \right] \frac{\partial T}{\partial t} = \left(\frac{\partial}{\partial z} \left(\lambda_b \frac{\partial T}{\partial z}\right) - c_l \rho \frac{\partial q_y T}{\partial z}\right) + q_y \rho_w c_w (T_{in} - T)
\]

in which the parameters \(c\) [\(\text{J kg}^{-1}\text{C}^{-1}\)], \(\eta\) [\(\cdot\)], \(\lambda_b\) [\(\text{J m}^{-1}\text{s}^{-1}\text{C}\)], and \(q\) [\(\text{m s}^{-1}\)] refer to bulk heat capacity, porosity, bulk thermal conductivity and flow rate of liquid water respectively. The subscripts \(z\) [\(\text{m}\)] and \(y\) [\(\text{m}\)] refer to the vertical and horizontal directions, and \(T_{in}\) [\(^\circ\text{C}\)] is the temperature of water laterally entering the soil column driving advection. Note that water may be supplied laterally to the column via the final \(q_y\) term, with the flux given by \(q_y = -K(\psi, T) \frac{dh}{dy}\), where the gradient is fixed for each simulation, and \(K\) values reflect the impedance due to ice content. This term is used only when the soil column contains a talik connected to a wetland feature. The inclusion of this source term allows the 1-D vertical model to represent lateral water movement in short-term simulations of permafrost evolution. This method is not appropriate for long-term change detection where lateral permafrost thaw is expected.

These equations have individually been solved elsewhere, and the uncoupled formulations (i.e. unfrozen Richards’ and saturated conductive heat transport) were separately benchmarked against results from Kurylyk et al. (2014) and Celia et al. (1990).
No existing analytical models are available to benchmark the coupled set of equations. The relationships are coupled using operator splitting, first solving the unsaturated Richard’s equation and then the heat transport equation in each time step, using an approach similar to Harlan (1973). The specific storage $S_s$ is dependent on ice content, $\theta_i$, which is determined from the temperature $T$, as is the hydraulic conductivity, $K$. Moisture migration due to temperature gradients near the freezing temperature is allowed using the CC relationship. The inclusion of this process changed model predictions less than 1% in saturated conditions, but reduced model stability and increased computation time. Bulk parameters are calculated based on the ice/water/air fractions in the soil matrix, where the heat capacity ($c$) and density ($\rho$) are calculated using the volumetrically weighted arithmetic mean of saturation values determined by the unsaturated Richards’ equation, and the thermal conductivity $\lambda_b$ is calculated using a volumetrically weighted geometric mean, as suggested by Kurylyk et al. (2014). The physical properties of the assumed soil column are homogenous, except those which depend on the water or ice saturation of the profile, which are allowed to vary with soil water or ice content. The omission of depth-dependant parameters most affects the hydraulic conductivity in the top 0.3 m of the soil profile, as the other modelled parameters are relatively constant with depth (Quinton et al., 2008). This may speed the equilibration time of the profile when subject to specified water flux conditions due to higher hydraulic conductivity near the surface, but is not expected to drastically affect model results and still permits the comparison of various drivers of permafrost degradation.

As seen in figure 2, the soil column is discretized into 2 cm elements for all simulations. For simulations where the soil column is saturated (i.e. not those investigating unsaturated conditions), a 0.48 h time step is used. In unsaturated conditions this is refined to 0.24 h to ensure convergence, and is further refined when moisture migration due to temperature gradients is included to 0.12 h. In all cases, the thermal operator converged at each time step. In the unsaturated case with moisture migration due to temperature gradients, the moisture operator (equation 1) failed to converge no more than 0.5 % of the simulation time. This lack of convergence occurred in the shoulder seasons, when the hydraulic conductivity is modified by ice content and very sharp gradients due to relatively rapid movement of the freeze/thaw front were observed as see in figure 1 (b).
3.3 Boundary and Initial Conditions

As a peatland, the study site maintains a water table near the ground surface. The water table is found within the top 0 - 30 cm of the soil profile in the wetlands and in depressions atop peat plateaux. The sloping edges of the peat plateaux are able to drain to adjacent wetlands, and the interior of the plateau can drain to internal depressions. In these locations, the water table is often found just above the frost table at a depth of 30 - 80 cm below the ground surface depending on time of year. A depth of - 50 cm is representative of a well-drained area for most of the thawing season.

The surface boundary condition for the Richards’ equation used here is a specified flux condition. Saturated conditions are simulated with a no-flow surface condition imposed on an initially saturated soil column. For unsaturated conditions, the average depth to water table of - 50 cm on peat plateaux is used as an initial condition. A constant spec-
ified flux condition derived from water level records consists of the removal of 10 cm of water between mid-May and the end of June, and the addition of the same amount of water between mid-August and the end of September. A no-flow condition is imposed at the base of the soil column, where it is assumed that permafrost is always present. A diagram of the model setup including both thermal and water content initial and boundary conditions is included in figure 2.

Thermal boundary conditions are given by the ground surface temperature as modified by the presence of the snowpack, generated using a seasonal autoregressive moving average model following the method outline by Hipel and McLeod (1994) constructed using soil temperatures collected in the field. This allowed to generate multiple independent temperature boundary conditions with the same statistical properties as the collected field data. It was necessary to use a continuous dataset from the same field site to appropriately capture the data covariance. For the stable case, only a four year time-series was available for building the autoregressive model. In the degrading case, a longer 10 year data set was available, though it had a data gap of approximately six months. Therefore only a four-year subset was used to generate the autoregressive model, while the entire series was tested for stationarity. Both datasets were tested for stationarity: it was found that a stationary model was sufficient for stable permafrost, while a linear trend of + 0.055 °C/year was apparent in the deseasonalized data for the degrading case. This positive trend was applied when generating ensemble temperature data for this boundary condition. Boundary conditions realizations for either 5 - year or 10 -year simulations were sampled from the stable and degrading models reproducing seasonal trends and data variance observed in the field, while incorporating inter-annual variability. The bottom thermal boundary condition is a zero heat flux condition. Field data show that the permafrost is effectively isothermal at 0 °C to a depth of 8 m, and simulations have a maximum vertical depth of 3 m, so the geothermal gradient (of approximately 0.08 W/m² (McClymont et al., 2013)) is neglected, see figure 1(a).

Heat flux at the ground surface is a clear direct driver of permafrost degradation (e.g. Walvoord et al. (2019)). Changes in tree canopy, surface albedo and ground cover contribute to modified ground heat flux (Quinton et al., 2019). This change is incorporated into model simulations as an increase in soil surface temperature in the summer. It would not be expected that the duration of the summer and winter season or the overwinter temperatures should be affected due to a change in canopy, unless there are changes
to the snowpack due to interception. In the winter season, the ground heat flux is controlled by snow accumulation. Snowcover, and inter-annual variability in snowcover, has important implications for the ground thermal regime. End of season snow surveys are conducted annually across all landcover types at the SCRS providing an estimate of snow water equivalent (SWE), or total water stored on the landscape in the form of snow.

Once a talik exists in the soil profile it is possible for it to connect adjacent wetland features. In the case of a connected talik, lateral advection through the talik is parameterized by treating through-flow as a distributed source/sink term in the 1-D energy balance. The lateral hydraulic gradient contributing to $q_y$ in equation 3 is specified as 0.002 in the case of a connection to a collapse scar wetland and 0.007 in the case of a channel fen. The gradients are selected based on water level data collected in the field. Actual field data from talik features show variations and even reversals in hydraulic gradient over the year, but for simplicity these are omitted in favour of the average observed differences between wetland types. The range of inflow temperatures was derived from measurements in a collapse scar wetland and a channel fen at a depth of 50 cm below the ground surface. Sensitivity to changes in gradient are computed to put this assumption into context. For simplicity, this study neglects any freezing point depression, and assumes phase change occurs between 0 and -0.05 °C, while data collected at the field site indicates a very small freezing point depression.

Two sets of initial temperature conditions are used in model simulations: a plateau condition and a talik condition. These conditions describe (generalized) field observations of mid-winter soil temperatures on a stable peat plateau and a plateau with a talik respectively. The plateau initial temperature profile is a uniformly frozen soil column at a constant temperature of -0.1 °C. The talik condition is initialized to the same temperature as the plateau condition, except from a depth of 0.75 m to 1.5 m which is initialized at a temperature of 0 °C and zero ice content as seen in figure 2.

4 Results and Discussion

4.1 Model Evaluation

After the model was successfully benchmarked against both thermal and water-content models from literature (Kurylyk et al., 2014; Celia et al., 1990) (see supplementary material), the performance of the model was evaluated in relation to soil temperatures and
refreeze depths measured at the SCRS. This ensured that the governing processes in this field site were represented adequately, and the approximated depth-homogenized soil properties were appropriate.

### 4.1.1 Soil Temperature

The model was first tested by comparing modelled and measured temperatures at depth (40 cm - 50 cm below ground). Stable boundary conditions were applied to an unsaturated plateau initial condition with water table initially 50 cm below the soil surface to best represent a permafrost plateau. All model simulations are summarized in table S2 of the supplementary materials. A contour plot of temperature evolution for the stable peat plateau boundary condition, as well as a comparison of modelled and measured temperatures at approximately 45 cm is shown in figure 3 (a) and (b). These results were obtained without model calibration. Modelled soil temperatures adequately represent the zero curtain period both during spring melt and winter freeze up. Measured data at all depths have high variability relative to modelled data because they are aggregated from data collected at 5 different sites over approximately 10 years (with data gaps) including inter-site variability, while simulated boundary conditions are constructed from data collected at a single instrumented site.

### 4.1.2 Refreeze Depth

Though there is good agreement between modelled and measured soil temperature at depth, latent heat represents a large fraction of the system’s energy storage and transfer (Hayashi et al., 2007), so the depth of freeze/thaw is also compared to field measurements of active layer (figure 3 (c)). Field measurements of maximum thaw depth in the absence of a talik reported 59 ± 4.5 cm (n = 106) and 60 ± 4.1 cm (n = 99) in 2016 and 2017 respectively, while the maximum refreeze with a talik was 58 ± 12.2 cm (n = 120) and 59 ± 11.5 cm (n = 48) in 2017 and 2018 respectively. Numbers in parentheses indicate the number of point measurements. Modelled soil columns with a talik align well with the measured value of refreeze over a talik: 61 ± 6.7 cm (n = 12), while simulations without a talik very slightly under-estimate the maximum thaw depth: 56 ±10.5 cm (n = 15). Given the agreement between modelled and measured soil temperatures and active layer depths, the model was deemed sufficient for the evaluation of controls on talik formation.
Figure 3. Modelled stable permafrost plateau. (a) shows freeze-thaw cycles over 5 year simulation. (b) Range of measured and modelled temperatures at 40 cm below ground surface for the soil column. (c) Measured refreeze and underlaying permafrost is shown on the left, sorted by increasing depth to permafrost (n = 48). Dashed line indicates average refreeze (over a talik). Box and whisker plots indicate the measured and modelled variance in active layer (refreeze) depth (n = 12).

The benchmarked and verified model was used to 1) assess the relative influence of controls on talik formation from a continuous permafrost state (section 4.2) and, 2) determine how these factors affect permafrost degradation rates once a talik is formed (section 4.3).
4.2 Controls on Talik Formation

Conditions favourable to talik formation on a permafrost plateau are identified based on the a) soil moisture, b) soil surface temperature, and c) snow cover. All simulations of talik formation were subject to the plateau initial condition and (unless otherwise stated) the stable boundary condition. Average snow conditions for the study site (115 mm SWE) are used in all simulations except those testing the impact of changes in SWE.

4.2.1 Unsaturated Soil Conditions

To determine the impact of different soil moisture conditions on talik formation, a modelled saturated soil column was compared to three unsaturated columns. For the unsaturated cases, the water table was initially set to a depth of 0.25 m, 0.5 m, and 1 m, and then subject to a specified flux boundary condition, with and without moisture migration due to temperature gradients.

Figure 4 compares the unsaturated (a) and saturated (b) cases. It can be seen by examining the 0 °C isotherm that in the unsaturated case there is complete refreeze of the soil column each winter, while the freezing front penetrates the soil column further than the thawing front. This can be compared to the saturated case in figure 4(b), in which a talik forms. The unsaturated surface condition in the summer plays a key role in insulating the permafrost, protecting it from degradation evident in figure 4(b) as the expanding region in the zero curtain (between 0 and -0.05 °C).

The response of the mean temperature in the top 1 m of soil, and talik thickness to changes in soil moisture are presented in panels (c) and (d) of figure 4. The mean temperature and the volumetric water content are evaluated over the course of the entire 5-year simulation, while the maximum talik thickness is evaluated in the final winter of the simulation. Part (c) of the figure shows that as the moisture content increases, the mean annual soil temperature decreases, likely because more energy is required to heat a wetter soil profile. As the water table moves further from the soil surface, it becomes more difficult to draw the water to the freezing front, insulating the soil column overwinter. The saturated case is drastically different from the unsaturated case because it is not subject to the seasonal water flux at the soil surface. This flux plays an important role in cooling the profile as it increases the thermal conductivity over winter and decreases it during the warmest part of the summer. The inclusion of the CC relation
Figure 4. Soil temperatures over 5 year simulation with stable boundary condition for (a) data-driven unsaturated soil conditions with initial water table 25 cm from soil surface and (b) hypothetical saturated conditions. Dashed line indicates the maximum thaw contour. Panels (c) and (d) present the respective response in mean temperature of near-surface soil and talik thickness to changes in soil moisture (labelled by initial position of the water table below the ground surface).

in the simulations (grey triangles) obfuscates the moisture effects because the temperature gradient draws water to the soil surface, bringing the upper soil near saturation. With the CC relation, the applied boundary and initial conditions are not impactful on simulation results, which are shown as an average with standard deviation.
Only the saturated soil column develops a talik without the addition of moisture migration to the freezing front, and though the least saturated case has a warm mean annual temperature, it does not have sufficient heat capacity to offset the over-winter freezing of the soil column. The decreased soil moisture also leads to very low thermal conductivity, limiting the ability of this heat to penetrate the soil column and cause permafrost thaw. The active layer in these simulations is deeper, but is not saturated and stores no more energy than the saturated conditions. In all cases involving the CC relation, a talik was formed. This is thought to be due to the near-saturated conditions induced in the near-surface during both freezing and thawing. The migration of moisture to the thawing front overcomes the imposed summer drying and winter wetting, leading to near-saturated conditions. However, the simulations without the CC relation are consistent with moisture conditions observed in the field and should therefore be considered when understanding the impact of soil moisture on talik formation.

In the field it is thought that taliks are associated with increased soil moisture, and differences in soil moisture are the driving control on differentiating land cover types, where the presence of (dry) lichen indicates higher permafrost stability, as reported by Grant et al. (2017); O’Donnell et al. (2009) and many others. These simulations seem to confirm this field observation. Other modelling studies have found that soil moisture is less impactful on active layer evolution and talik formation when compared to changes in snowfall, but they do not include the observed seasonality in soil moisture leading to dry summer conditions and wet winter conditions (Atchley et al., 2015). The inclusion of this seasonality alleviates the competing processes of increasing thermal conductivity due to increasing moisture content, with a concomitant increase in latent heat required to fully freeze/thaw the soil column observed in other modelling studies (Atchley et al., 2015).

### 4.2.2 Temperature

Summer soil surface temperatures were altered to reproduce changes in heat flux into the ground representative of differences in incoming radiation or surface albedo. Both the stable and degrading boundary conditions were considered. Results are reported in terms of annual mean surface temperature. Figure 5 shows the thickness of talik formed due to changes in mean annual surface temperature. Both responses are near linear, with the degrading condition forming a notably thicker talik at cooler temperatures. The shape of the temperature profile drives the degrading condition to store more heat despite the
same mean annual temperature in both simulations. The *degrading* condition differs most prominently from the *stable* condition in that the zero-curtain period is longer. This delay in freezing and thawing may lead to a net energy gain to the system, where the zero-curtain does not provide a temperature gradient sufficient to freeze the soil in the fall, while the (shorter) spring zero-curtain period occurs earlier than in the stable case, leaving the summer thaw season approximately the same length in both cases.

### 4.2.3 Snow Cover

It is widely understood that snow is a highly effective insulator due to its large (60-90%) air volume. When there is an early or thicker than usual snowpack, the soil temperatures are warmer than usual (Woo, 2012). Though somewhat data-limited at this field site (established in 2006), figure 6 (c) shows that this relationship holds true at the SCRS. The relationship between snow water equivalent (SWE) and soil temperature was used to test the impacts of increased snowfall by uniformly increasing the average overwinter temperature for the *stable* condition where the empirical relation from figure 6 (c) was used to determine the impact on winter soil temperatures. An increase in soil surface temperature derived from the linear relationship between SWE and overwinter...
soil temperature (figure 6(c)) was applied to only the first winter (figure 6(a)) and then
the first and second winter (figure 6(b)). Though there is no immediate talik develop-
ment in the first case, it should be noted that the talik formed in the fourth winter is
actually due to the additional energy stored in the soil profile from the first high snow
winter, as without it, the simulation results in figure 3(a). The second case clearly con-
firms that increased snow cover can be a trigger for talik formation. These results were
also found in the study by Atchley et al. (2015) and Atchley et al. (2016), in which it
was shown that increased snow cover could lead to talik formation near Barrow, AK. Sim-
ilarly to that study, increased snow resulted in an extension of the zero-curtain period,
and a decrease in the maximum depth reached by the freezing front in high-snow years
(Atchley et al., 2015). Some climate change scenarios predict higher snowfall in the study
region (Solomon et al., 2007). Snow depth may also be increased locally by talik forma-
tion which leads to the development of local depressions where snow will preferentially
accumulate, a forward feedback mechanism. These depressions and accompanying in-
creased snow depth have been observed at the SCRS. This is similar to sites with shrub
development, where Jafarov et al. (2018) have shown that the preferential snow accu-
mulation within shrub patches can lead to the formation of isolated taliks that evolve
from confined taliks to through taliks.

The opposite process, in which low-snow years can reverse the formation of a talik
is less likely. A decrease in SWE of equal magnitude to the initial increase is required
to reverse the formation of a talik. Similarly to other regions in which the overwinter snow-
pack undergoes transformations such as sublimation and redistribution, the statistical
distribution of annual SWE near the study site is skewed. It shows a tail toward high
snow years, and a higher probability of just below average low-snow years (Shook et al.,
2015). Therefore it is unlikely to see an equal magnitude SWE deficit.

The sensitivity of talik formation to snow cover was tested by modifying the sta-
ble case with winter temperatures shifted resulting in a mean surface temperature of 2°C
as the original saturated stable condition resulted in talik formation with no change in
SWE. Figure 6(d) indicates the necessary deviation from mean SWE (temperature shift
applied only in the second winter) to form a talik is approximately 2 cm.
4.3 Talik Evolution

Soil moisture, temperature and snowfall conditions have been identified under which taliks are formed, initiating permafrost degradation. Once a talik is present in the soil column, field observations indicate that it is likely that permafrost will continue to degrade. The sensitivity of this permafrost degradation to the same conditions studied above is presented here. These results are extended to include the impacts of advection through the newly formed lateral flow path. Simulations in the following sections are run with the talik initial condition, and (unless otherwise specified) the stable boundary condition so as to be directly comparable to the simulations of talik formation. Average snow accumulation...
conditions for the study site are used in all simulations except those testing the impact of changes in SWE.

4.3.1 Soil Moisture

The impact of soil moisture on average temperature in the top 1 m of the soil column is very similar in the talik condition and the plateau condition, especially when the CC relation is not applied (figure 7 (a)). With a talik, there is less temperature-driven movement of water in near-surface, resulting in similar trends in temperature as the case neglecting the CC relation. In all cases, the mean temperature is higher than without a talik.

The impact of unsaturated conditions on the thickness of the talik is quite remarkable. All unsaturated conditions without the CC relation, and the condition with lowest soil moisture including the CC relation resulted in talik thinning and evidence of permafrost recovery. This contrasts the saturated conditions (and near-saturated conditions with temperature-driven moisture migration) in which the talik thickens over the 5-year simulation.

4.3.2 Temperature

An increase in talik thickness was found in the case initialized with a talik for all surface temperatures above 2°C (Figure 7 (c)). A decrease in mean annual temperature up to 1°C did not seem to lead to permafrost recovery as the talik thickness remained constant. Simulated increased summer temperatures result in faster permafrost degradation beneath a pre-existing talik than over a plateau forming a talik.

4.3.3 Snow

The relationship between snowfall and permafrost degradation is very similar before and after the formation of a talik, i.e., small increases in SWE help to insulate the ground leading to increased talik thickness. Though the trend remains the same, an increase in talik thickness is observed in all cases tested, while in the simulation started with the plateau condition only showed talik development after a SWE increase of 20 mm (figure 7 (d)). Even under low-snow conditions, the talik in the profile is observed to thicken by 4 cm over the course of the simulation, with no evidence of permafrost recovery.
**Figure 7.** (a) Sensitivity of average soil temperature of top 1 m of soil profile (only in the case of soil moisture). Change in talik thickness due to (b) soil moisture, (c) average surface temperature, and (d) changes in SWE for talik initial condition.

### 4.4 Advection

Here, three hypothetical soil columns with the talik initial condition and degrading boundary condition are modelled to examine the role of advection: 1) an isolated talik (without advection), 2) a talik with a hydrological connection to a collapse-scar wetland, and 3) a talik with a hydrological connection to a channel fen. A hydraulic gradient of 0.002 was applied in the case of the bog and 0.007 in the case of a fen, while the isolated...
talik was not subjected to any lateral flow. Figure 8 (a), (b) and (c) shows that although all three simulations result in permafrost degradation, advection accelerates thaw rates and increases the sensible heat stored in the soil profile. As reported by Sjöberg et al. (2016), the abrupt changes in temperature of water flowing through the soil column in a fen observed at the SCRS can have significant impacts on thaw rates because of the high thermal gradients induced. In both cases including advection it is apparent that there is overwinter permafrost thaw in taliks adjacent to wetlands, which is not present without advection (see annotations on figure 8(c)). This overwinter degradation has been observed in the field in a talik connecting a fen on one side to a bog on the other. Overwinter permafrost degradation can be seen as a tipping point, where permafrost recovery is highly implausible as degradation occurs year-round. No plausible conditions could be identified through modelling that led to permafrost recovery in this case, nor have there been any instances of permafrost recovery observed under these conditions in the field.

Figure 8. Soil temperatures as a result of advection representative of (a) an isolated talik, (b) a talik adjacent to a bog, or (c) a fen. Rate of vertical permafrost degradation determined by temperature (d) for a constant gradient of $7 \times 10^{-3}$, and hydraulic gradient (e) where the inflow temperature is unaltered from field measurements.
The inclusion of plausible advection rates roughly doubles the modelled permafrost degradation rate relative to that with only conductive input (from 9.1 cm/yr to 17.5 cm/yr). McKenzie and Voss (2013) also report that conduction and advection have similar magnitude effects on permafrost thaw rates, though their analysis focussed on taliks beneath lakes and groundwater exchanges were the source of (vertical) advection. Lateral flows as the source of advection similar to the ones modelled here were considered by de Grandpré et al. (2012), where advection contributed a significant amount of heat to thaw permafrost under a road bed, though the relative contribution of advection was not computed.

The assumption of a constant gradient in the simulations of a bog and a fen neglects the variability in this data. In the case that a talik connects two wetland features, this assumption is reasonable. Though there may be seasonal changes in pressure (especially during the freshet), there is enough water stored in each feature to sustain flow over the winter without large changes in gradient. However, a small isolated talik adjacent to a wetland (with a single connection) likely does not have the storage capacity to sustain flows over winter, so the impacts of advection would be less in these cases. Observed rates of permafrost degradation at the SCRS are slower in these cases.

To better understand the dependence of thaw rate on temperature and rate of advection, 42 simulations with consistent stable boundary and talik initial conditions were completed for a range of advective flow rates and incoming temperatures. Figure 8(d) and (e) shows the change in permafrost degradation rate due to increasing mean advection temperature and hydraulic gradient. Permafrost degradation rate for each simulation was calculated as the slope of a linear fit to the maximum annual depth to permafrost over a 10 year simulation. The inflow temperature was sampled between $T_{in} - 1 \, ^\circ C$ and $T_{in} + 3 \, ^\circ C$, where $T_{in}$ is the mean daily inflow temperature representative of a bog which never drops below 1.2 $^\circ C$.

Thaw rates are comparably sensitive to changes in hydraulic gradient and advection temperature, where an increase in gradient or temperature leads to a clear increase in permafrost thaw rate, as may be expected from examining equation 3. Both relations are linear in the range unaffected by changes in hydraulic conductivity due to partial ice saturation. As seen in figure 8(e), at low flow rates, pore ice formation limits the hydraulic conductivity and thus the impact of hydraulic gradient on thaw rate is suppressed (up to a gradient of approximately 0.005). In the field, this decrease in hydraulic conduc-
tivity may lead to a resulting increase in hydraulic gradient as water movement becomes limited, pushing advection out of this non-linear, low-flow region. Increases in hydraulic gradient may also be expected during the freshet when ice and snow prevent overland flow and result in ponded water on the landscape, though many flow pathways would be clogged with frozen pore water. Such complicated interplay between partially frozen soils and groundwater flow is similarly documented in de Grandpré et al. (2012).

5 Synthesis

5.1 Drivers of Talik Formation

A comparison of figures 4, 5 and 6 indicate that soil moisture conditions and changes in SWE have effects of similar magnitude on the formation of taliks over the expected range of parameter values observed in the field. The magnitude of the resulting change in talik thickness is similar to that observed for a change in ground surface temperature of 0.5 °C. This suggests that with projected climate warming in Northern Canada (Solomon et al., 2007), talik formation is likely regardless of changes in soil moisture and SWE regimes.

Model simulations indicate that small variability in forcing even over a single season can lead to permafrost degradation and talik formation. This sensitivity to perturbation is not unique to the SCRS, where the permafrost can be classified as ecosystem protected, as defined by Shur and Jorgenson (2007). Robinson and Moore (2000) describe how changes to the surface layer of peat due to wildfire can lead to permafrost degradation, but other disruptions to it such as an abnormally wet summer, the removal of a tree canopy, or anthropogenic compaction of the surface layer are also very likely to lead to talik formation (through the process shown in figure 4).

Though soil moisture, ground heat flux and SWE are thought to be the main drivers of talik formation, there were other factors which should receive attention. One such factor is the timing of the snow-covered period as a control on the refreeze process. Though not simulated here, it is anticipated that a prolonged delay between the onset of freezing conditions and the arrival of substantial snowfall would result in enhanced refreeze depths as the absence of snow would not only increase the conductive heat transfer through the soil surface, but would also leave the bare surface exposed to convective heat transfer. Measurements and simulations by Zhang et al. (2007) confirm this, showing earlier and deeper refreeze in a year with late snow onset. This is observed in the field as the
sloping edges of channel fens which are scoured by wind in early winter generally exhibit
refreeze about 10 cm deeper than other, more protected, areas in the landscape.

5.2 Talik Formation as a Tipping point

Modelling results from section 4.3 demonstrate that once a talik is present in the
soil column, only changes to soil moisture are able to initiate permafrost recovery in the
studied discontinuous permafrost peatlands environment. The other cases including cooler
mean annual surface temperatures and decreases in SWE (within ranges observed at the
SCRS) did not lead to any talik thinning. Comparing results presented in sections 4.2
and 4.3 leads to the generalization that changes in mean annual temperature results in
increased permafrost degradation rates once a talik is formed. Changes to soil moisture
and SWE caused slower talik thickening after talik formation due to the increased soil
column depth, but did show more sensible heat storage.

Advection modelling tests (section 4.4) showed that lateral flow through a talik sig-
nificantly increases thaw rates which has been verified by field observations. Such flow
pathways are permanently active year-round and can contribute significantly to permafrost
thaw in a manner that is likely irreversible, since advection can only supply energy to
frozen ground. McClymont et al. (2013) document advection as a dominant control on
permafrost thaw, but in this work, as in others (e.g. Walvoord and Kurylyk (2016)), ad-
vection is the result of flows along the edges of peat plateaux, along channel fens, or in
the moat of collapse scar bogs. Here it is proposed that advection through a talik con-
necting wetland features (i.e. a bog connected to a fen via a talik) would also lead to ther-
mal erosion of permafrost at the top of a permafrost body. This has the potential for
a much higher hydraulic gradient and subsequent flow rate than may be expected in a
channel fen in this low-relief landscape (with a typical gradient of 0.0032, Stone (2018)
as compared to a gradient of 0.01 measured between features connected by a talik), es-
pecially over winter when ice and snow-load can increase subsurface pressure in the land-
scape.

Given the drivers of talik formation and evolution identified above, the model was
applied using observed field conditions to predict likely changes in permafrost in differ-
ent parts of this landscape. These conditions are the plateau initial condition with sta-
boundary condition (Stable Plateau) and talik initial condition with degrading boundary conditions (Degrading Talik).

5.2.1 Stable Plateau

None of the 20 realizations of surface boundary conditions constructed using the seasonal autoregressive model of the stable condition discussed in section 3.3 with an initial water table at a depth of 0.5 m below the surface resulted in the formation of a talik in the 10 years modelled. The simulations did however show potential for active layer thickening, as reported in Shiklomanov et al. (2012). There was an increase in maximum thaw depth (with complete refreeze) of approximately 0.3 cm/year over the entire 10 year simulation, with the same average rate in only the last 5 years of simulation for the stable condition. Figure 9(a) shows the distribution of rates of permafrost degradation for this stable boundary condition on the left for the entire 10-year simulation (grey), and then only for the last 5 years of simulation (black). Agreement between these supports an appropriate selection of initial conditions. The small annual loss of permafrost in a 10 year simulation indicates that the current condition of peat plateaux results in permafrost degradation. Given long enough periods forced with these consistent boundary conditions, it is expected that the permafrost will begin to degrade, as discussed for this field site in Quinton et al. (2019). This is analogous to results found by Briggs et al. (2014) who shows that though permafrost aggradation has been observed in draining lakes, this phenomenon is only transitional as the permafrost is an artefact of the groundwater regime and shading from shrubs, and it is expected to thaw within the decade.

At the SCRS there is clear evidence of degrading permafrost. Evidence of aggrading permafrost is sparse and unconfirmed, suggesting that permafrost formation may not be possible in the current climate unless significant dewatering of the landscape occurs. Though sections of permafrost peat plateaux appear stable, they are extremely vulnerable to change as the underlaying permafrost is not in equilibrium with the current climate. Given the slow rate of permafrost degradation identified from models with the current boundary conditions, it is only a matter of time before this ecosystem protected permafrost degrades even without the additional mechanisms discussed above. Using tree canopy as a proxy for permafrost (Carpino et al., 2018), aerial imagery from this study site was compared from 1947 to 2008, showing a $38 \pm 8\%$ decrease in permafrost coverage (Quinton et al., 2011), indicating that this degradation is already underway. This
Figure 9. (a) Stacked relative frequency distribution of permafrost degradation rates modeled with stable boundary condition (left) and degrading condition (right). Vertical dashed lines indicate the mean permafrost degradation rate for the entire simulation (10 years) or for only the last 5 years of simulation. (b) Talik formation (for the degrading boundary condition) shown as a temperature contour plot. Note that the incomplete refreeze in the sixth winter initiates the formation of a talik, which then provides a previously absent pathway for advection.

is echoed in the findings of Kwong and Gan (1994) describing a northward moving southern limit of sporadic and discontinuous permafrost due to increases in mean annual temperature.

5.2.2 Expanding Talik

Figure 9 (a - degrading) shows the response of an initially talik-free system to the degrading boundary condition. The formation of a talik increases permafrost thaw rates, while also increasing the variability in thaw rate. The variance in simulated thaw rates is significantly greater in the case with a talik for two possible reasons: 1) These simulations include the formation of the talik, as shown in figure 9(b), which is sensitive to changes in boundary conditions; and 2) The formation of a talik allows for advection which is further modified by the presence of ice in the soil column affecting the permeability. It is not appropriate to compare the permafrost degradation rate in the stable case directly to the talik case as the boundary conditions in each scenario differ. The impact of talik formation on permafrost thaw can, however, be discussed by focusing only on the talik condition simulations.
An example of talik formation can be seen in figure 9(b). The first five winters of this simulation involve complete refreeze and subsequent cooling of the underlaying permafrost. As soon as a talik is formed in year six, the underlaying permafrost remains near the measured freezing point depression and above the temperature at which the freezing function used in the model reaches residual unfrozen saturation. This simulation aptly demonstrates that permafrost degradation clearly accelerates once a talik forms, as can be seen to the right of figure 9 (a - degrading). This is anticipated for three reasons: 1) the activation of an advective flow pathway, 2) a reversal of the temperature gradient, and 3) (in the case of field measurements) subsidence of the ground surface, resulting in a positive feedback leading to increased soil moisture.

The presence of a talik with a temperature at or above the zero-point depression alters the ground temperature profile such that the surface of a permafrost body always experiences a positive (or zero) temperature gradient (Connon et al., 2018). The permafrost at or below the freezing point depression is always colder than the overlaying thawed talik, and consequently gains thermal energy year-round. This can be seen in figure 1(a), which shows the (co-linear) maximum and minimum annual temperatures measured over a talik. The talik is apparent in the figure as the region (in grey) that never cools below the zero point depression, but is warmed in the summer. Below this region, the permafrost is essentially isothermal in the zero curtain, indicating that it is undergoing phase change and is unable to lose energy to the atmosphere. In combination with advection, this can lead to the overwinter permafrost thaw observed in the SCRS. One such talik feature connecting a bog and fen has exhibited 21 cm of thaw between August 2016 and May of 2017 and 13 cm of thaw between September 2017 and April of 2018.

In the field, the positive feedback of permafrost degradation after talik formation is furthered by an increase in soil moisture due to ground surface subsidence. As shown in section 4.2.1, this increase leads to slightly enhanced thaw rates, but increased soil moisture also counteracts the only realistic modelled conditions able to promote permafrost recovery - an unsaturated soil column. More energy is able to reach the permafrost table both due to the increased thermal conductivity, as well as a thinning canopy as the black spruce (Picea mariana) suffer from water-logging of their root networks (Quinton & Baltzer, 2013). The depressions formed allow for preferential accumulation of snow, which was shown to trigger permafrost thaw. In this sense, the formation of a talik can
be seen as a tipping point leading to accelerated permafrost thaw rates with little chance of recovery.

5.3 Permafrost Degradation in the Landscape

Given the impacts of soil moisture, advection and the existence of taliks on permafrost degradation rates, it is expected that the rates of permafrost degradation should differ across landscape features. Sjöberg et al. (2016) suggest that the relative importance of thaw mechanisms including conduction and vertical or lateral advection vary both seasonally and across different peatland landscape features. To better capture thaw in each part of the landscape, a final modelling experiment was undertaken. In this experiment, advection was forced with temperatures based on profiles observed in a bog and a fen at the SCRS, similar to the simulations used to generate figure 8(a), (b) and (c). Temperature boundary conditions were consistent with the stable condition for the plateau landscape type, while the three talik conditions (isolated, connected bog and connected fen) used the a boundary condition constructed using the autoregressive model constructed for the talik condition. Modelled thaw rates are compared to thaw measured as change in end-of-season depth to permafrost using a frost probe at different locations in the landscape. This comparison is shown in figure 10, where field data were categorized according to the type of talik. Note that the high variance in measured thaw rates in connected taliks adjacent to fens and bogs is likely due to variations in flow rate and temperature at different monitoring locations. The modelled thaw rates for taliks assume a saturated soil column. This explains the slight over-estimation of thaw rates in isolated taliks, which are generally wetter than the surrounding stable plateau, but are rarely completely saturated in the field. Permafrost degradation is incrementally faster as advection rates and temperatures increase in bogs and fens, as demonstrated in figure 8.

The slow but positive thaw rate observed on a ‘stable’ plateau is indicative of a system in disequilibrium with the climate. The gradual permafrost loss either as active layer thickening or as talik formation points toward an eventual near-complete loss of permafrost from the system. Once a talik is formed, the rate of permafrost degradation is notably more rapid due to the combined effects of higher thermal conductivity, advection, canopy degradation, thermal storage in taliks, and ground surface subsidence.
Figure 10. Comparison of permafrost thaw rates in different portions of the landscape. Box and whisker plot shows spread in measured talik development data, while triangles show average modelled value for each type of talik. A total of 78 locations were measured semi-annually over the course of 8 years to generate the permafrost degradation rate data. Note that points fall outside of 1.5 times the interquartile range, the upper bound of the whiskers.

At the landscape scale, permafrost loss is likely to emanate outward from existing wetland features, especially fens that have higher flow rates and temperatures, while preferentially forming the hydrologic connections between wetlands where the hydraulic gradient is greatest, as documented by Connon et al. (2015). Not only will the edges of permafrost cored peat plateaux be eroded, but depressions are likely to grow into isolated taliks which will expand and interlink, leaving small isolated hummocks of permafrost as described in Quinton et al. (2018). These isolated permafrost features are documented in the sporadic permafrost region by Woo (2012).

This transition is complex and difficult to incorporate into model boundary conditions as they are currently implemented, hence simulation duration is limited to a decade. If these more realistic boundary conditions could be represented, a smooth transition between the stable and degrading case would be expected. This boundary condition would include gradual wetting of the soil profile due to subsidence, increase in surface temperature because of canopy loss, increased snow accumulation due to depression formation and lateral flow rates and temperatures varying based on upstream conditions. Under these more realistic boundary conditions, long-term simulations would be meaningful, but it is likely that the conclusions regarding the positive feedbacks associated with per-
mafrost thaw drawn here would still be valid, though more specific predictions of thaw rates would be possible. Further work is needed to better quantify these rates at a longer time scale, and it is anticipated that multi-dimensional modelling would be necessary to represent this system adequately.

6 Conclusion

This study used a combination of extensive field data and 1-D modelling to investigate the formation of isolated taliks beneath the active layer. A freeze-thaw model was developed that reproduces temperature data and refreeze depths in discontinuous permafrost peatlands. This model was used to provide a rigorous evaluation of the controls on isolated talik formation, which is prevalent in discontinuous permafrost environments and can be driven by soil moisture, snow accumulation, and/or seasonal temperature trends. Soil moisture, ground heat flux, snowcover, and advection were all found to affect the formation of taliks in different contexts. It was deemed difficult to identify which factor is dominant as they often occur simultaneously and are inter-dependent. However, wet conditions, deep snow, and warmer soil surface temperatures were found to increase the probability and rate of isolated talik formation. Isolated talik formation was shown to be a tipping point in permafrost degradation, leading to accelerated permafrost thaw which is unlikely to recover once the talik is formed. Once a talik is formed, permafrost degradation can be accelerated by subsurface flows through the talik (especially if it forms a pathway between wetland features), increased soil moisture, increased ground heat flux, and snow accumulation due to a critical lack of energy loss from the permafrost to the atmosphere overwinter. Only unsaturated conditions (highly unlikely in permafrost degrading within a wetland system) were found to lead to permafrost recovery, resulting in accelerated and potentially irreversible permafrost thaw in this environment. It may therefore be pertinent to consider talik formation in models of other permafrost environments, especially at larger scales where these processes are often neglected, but may lead to significantly different permafrost conditions.

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