

**Examining the spatial distribution of soil moisture and its relationship to vegetation and permafrost dynamics in a Subarctic permafrost peatland**

**By**

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## ABSTRACT

### **Examining the spatial distribution of soil moisture and its relationship to vegetation and permafrost dynamics in a Subarctic permafrost peatland**

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Landscape change is occurring rapidly in the permafrost peatlands of the Canadian Subarctic due to rising mean annual air temperatures. Manuscript 1 examines the statistical characteristics and environmental controls of near-surface soil moisture at the field scale. Soil moisture standard deviation increased with mean moisture, while the coefficient of variation and skew decreased. Soil moisture was correlated to frost table depth and the general topographical differences between the elevated plateaus and low-lying fens and bogs. In Manuscript 2 *Picea Mariana* sap velocity, available water and atmospheric conditions were measured in a 100 m<sup>2</sup> plot. Sap velocity was primarily controlled by atmospheric conditions and antecedent soil moisture, indicating that the study site experienced drought stress. The findings of this research will aid in advancing our understanding of the processes governing forest browning, hydrological modeling of cold regions, and the accuracy of remotely sensed soil moisture products.

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## TABLE OF CONTENTS

|  |             |
|--|-------------|
| <b>ACKNOWLEDGEMENTS .....</b>  | <b>iii</b>  |
| <b>LIST OF TABLES .....</b>  | <b>vi</b>   |
| <b>LIST OF FIGURES .....</b>   | <b>vii</b>  |
| <b>LIST OF EQUATIONS.....</b>  | <b>viii</b> |
| <b>Chapter 1.0: Introduction .....</b>   | <b>1</b>    |
| <b>Chapter 2.0: Statistical characterization of soil moisture variability in a Subarctic<br/>permafrost peatland and its relationship to permafrost, vegetation and topography .....</b> | <b>5</b>    |
| <b>Abstract.....</b>   | <b>5</b>    |
| <b>2.1 Introduction.....</b>   | <b>5</b>    |
| <b>2.2 Methods.....</b>  | <b>8</b>    |
| 2.2.1 Study Site .....   | 8           |
| 2.2.1.1 Subarctic peatlands .....  | 8           |
| 2.2.1.2 Study site.....  | 10          |
| 2.2.2 Experimental design and data collection .....  | 10          |
| 2.2.3 Data Processing.....   | 11          |
| 2.2.4 Statistical Analysis.....  | 12          |
| <b>2.3 Results and Discussion.....</b>   | <b>13</b>   |
| 2.3.1 Characterization of soil moisture spatial variability .....  | 13          |
| 2.3.1.1 Statistical descriptors .....  | 13          |
| 2.3.1.2 Trends in soil moisture variance at the field-scale.....   | 15          |
| 2.3.1.3 Frost table variance .....   | 18          |
| 2.3.2 Controls on soil moisture variability.....   | 20          |
| <b>2.5 Conclusion .....</b>  | <b>23</b>   |
| <b>Chapter 3.0: The impact of temperature-induced drought stress on the environmental<br/>controls of <i>Picea mariana</i> sap velocity in a permafrost peatland .....</b>               | <b>25</b>   |
| <b>Abstract.....</b>   | <b>25</b>   |
| <b>3.1 Introduction.....</b>   | <b>26</b>   |
| <b>3.2 Methods.....</b>  | <b>28</b>   |

|   |           |
|---|-----------|
| 3.2.1 Study Area .....  | 28        |
| 3.2.1.1 Peatlands .....   | 28        |
| 3.2.1.2 Scotty Creek.....   | 28        |
| 3.2.2 Site Instrumentation.....   | 29        |
| 3.2.2.1 Measurement of sap flow.....  | 29        |
| 3.2.2.2 Measurement of atmospheric variables .....  | 29        |
| 3.2.2.3 Measurement of ground variables.....  | 30        |
| 3.2.3 Statistical Analysis.....   | 30        |
| 3.2.3.1 Examining environmental controls of sap velocity .....                                | 31        |
| 3.2.3.2 Identification of climate change induced drought stress in <i>Picea mariana</i> ..... | 31        |
| <b>3.3 Results .....</b>  | <b>32</b> |
| 3.3.1 Environmental controls on sap velocity.....   | 32        |
| 3.3.2 Sap velocity response under drought stress .....  | 34        |
| <b>3.4 Discussion .....</b>   | <b>35</b> |
| 3.4.1 Environmental controls of sap velocity .....  | 35        |
| 3.4.2 Influence of drought stress on <i>Picea mariana</i> .....                               | 37        |
| 3.4.3 Influence of climate change on <i>Picea mariana</i> .....                               | 39        |
| <b>3.5 Conclusion .....</b>   | <b>40</b> |
| <b>Chapter 4.0: Summary and Conclusion .....</b>  | <b>42</b> |
| <b>Chapter 5.0: References .....</b>  | <b>45</b> |

## LIST OF TABLES

|  |           |
|--|-----------|
| Table 2.1: Comparison of radiation ( $W/m^2$ ) calculations between the model and observations at two stations at Scotty Creek. Radiation was averaged for the entire day (E.D.) and daylight hours (7:00 to 20:00) (D.H.). Measurements were not available for the last 2 dates in 2013 so the closest sunny day (August 23, 2013) is presented instead. .... | <b>13</b> |
| Table 2.2: Statistics of the near surface (0 – 5 cm) volumetric soil moisture ( $m^3/m^3$ ) variance on a peat plateau for each sampling day.....  | <b>14</b> |
| Table 2.3: Comparison of soil moisture statistical properties in various studies. Studies included measurements in mineral and organic soils with and without permafrost.....  | <b>17</b> |
| Table 2.4: Frost table depth mean and standard deviation (to nearest cm), coefficient of variation and skew statistics on a peat plateau each measurement day. ....  | <b>18</b> |
| Table 2.5: Pearson’s r correlation coefficients for the associations between volumetric soil moisture and various environmental controls on a peat plateau. Significant correlations are denoted with an asterisk.....   | <b>22</b> |
| Table 2.6: Comparison of the mean and standard deviation of soil moisture in soils under moss versus lichen.....   | <b>23</b> |
| Table 3.1: Summary of best-fit statistics for the various ARIMA models developed to predict sap velocity.....  | <b>33</b> |
| Table 3.2: Significance values for each individual regressor in the best fitting ARIMA models for sap velocity. Soil moisture is only significant at 10 cm when lagged one day.....  | <b>33</b> |
| Table 3.3: Linear regression statistics between the measured variables and sap velocity at monthly intervals. ....   | <b>34</b> |

## LIST OF FIGURES

- Figure 2.1: a) Oblique aerial photograph of the study plot showing the three peatland landscape features: channel fens (light green areas), flat bogs (yellow areas), and plateaus (forested dark green areas). b) The study site (in red) nested within the larger pre-established grid. Measurements were taken every 20 m. North is at the top of the plot. .... **9**
- Figure 2.2: Distribution of near surface (0 – 5 cm) volumetric soil moisture measurements ( $\text{m}^3/\text{m}^3$ ) taken on each sampling day. The boxplot shows the interquartile ranges of measured soil moisture, with the dark horizontal line depicting the median, and the asterisk the mean. The whiskers show the range of mean values within 1.5 times the interquartile range, with outliers identified by the circles. Saturated areas have been excluded. .... **15**
- Figure 2.3: Relationships between daily mean volumetric soil moisture and the a) standard deviation, b) coefficient of variation, and c) skew. These were calculated for each sampling day using all the measurements taken on plateaus within the plot. .... **16**
- Figure 2.4: Plots of the probability density (histogram) of volumetric soil moisture ( $\text{m}^3/\text{m}^3$ ) on each sampling day. Soil moisture was positively skewed in all cases. .... **19**
- Figure 2.5: Distribution of frost table measurements taken on each sampling day. Fens and bogs do not have underlying permafrost or the permafrost is greater than 2 meters deep. .... **20**
- Figure 3.1: The relationships between sap velocity and a)  $R_n$  ( $R^2=0.58$ ,  $p<0.01$ ), and b) vapour pressure deficit ( $R^2=0.60$ ,  $p<0.01$ ) over the course of the study period. .... **35**
- Figure 3.2: The relationships between sap velocity and soil moisture at 5 cm ( $R^2=0.13$ ,  $p<0.05$ ), 10 cm ( $R^2=0.48$ ,  $p<0.05$ ), and 20 cm ( $R^2=0.41$ ,  $p<0.05$ ) over the study period. .... **36**
- Figure 3.3: Variation in environmental processes over the 72-day study period (June 12 – August 22). a) Soil moisture at 5, 10 and 20 cm depths. Bars are precipitation. b) Interpolated frost table depth and measured water table depth. c) Calculated net radiation ( $R_n$ ) and vapour pressure deficit ( $D$ ). d) Pattern of average plot sap velocity compared with the Priestley-Taylor potential evapotranspiration (PET). .... **38**

**LIST OF EQUATIONS**

|   |           |
|---|-----------|
| Equation 2.1: Beven & Kirkby 1979's wetness index .....   | <b>12</b> |
| Equation 3.1: Priestley-Taylor equation for potential evapotranspiration.....   | <b>29</b> |
| Equation 3.2: Heat pulse sap velocity ( $\text{cm hr}^{-1}$ ) equation used in Sap Flow Tool (ICT<br>International, Armidale, Australia)..... | <b>30</b> |

## Chapter 1.0: Introduction

The Subarctic region, located immediately below the Arctic, is particularly sensitive to climate warming as dramatic changes in permafrost extent, and the storage of water and carbon could occur with minimal increases in mean annual air temperatures (Prowse et al, 2009). Research in the last few decades has already shown distinct landscape changes across the Canadian Subarctic (Quinton et al, 2011). For example, Quinton & Baltzer (2013) observed increasing soil moisture and active layer thickness, Quinton et al. (2011) identified a 38% decrease in permafrost cover from 1948 to 2008 in the Lower Liard Valley, Northwest Territories, and Barber et al. (2000) found decreases in tree growth in northern Boreal forests. Numerous other researchers have stated analogous indicators of changing climatic conditions (e.g. Prowse et al, 2009; Vitt et al, 2000; Robinson & Moore, 2000; Jorgenson & Osterkamp, 2005; Hayashi et al, 2007; Lloyd & Bunn, 2007; Verbyla, 2008).

Peatlands, defined as wetlands that have accumulated organic material exceeding 40 cm in depth (NWWG, 1988), constitute 90% of Canada's wetlands and cover large portions of northern North America within the high-, mid- and low-boreal, taiga and tundra (Zoltai & Tarnocai, 1975). Subarctic Boreal peatlands are globally important as they sequester carbon (Robinson & Moore, 2000), influence the hydrological cycle (Quinton & Roulet, 1998), and affect water chemistry and vegetation (Price et al, 2005). These Subarctic peatlands are also often underlain by permafrost (Robinson et al, 2003). Permafrost occurs in soils that remain below 0 °C for over two years; however seasonal thaw may occur within the top few centimeters to meters of soil which is referred to as the active layer (Muller, 1945). The presence of permafrost is dependent not only air temperature (Halsey et al, 1995), but also vegetation and soil properties (Williams & Burn, 1996). While annual air and soil temperatures may exceed those required for permafrost formation, the insulating properties of peat allow for the persistence of frozen ground (Wright et al, 2009). Subarctic peatland complexes consist of several different types of peatlands, notably channel fens, flat bogs, and peat plateaus, which are defined based on their biophysical properties (Aylesworth & Kettles, 2000). Currently, as a result of climate warming peatland complexes are increasingly being transformed as the permafrost

supported peat plateaus collapse into saturated permafrost-free channel fens and flat bogs (Quinton et al, 2011).

In recent years, interest in soil moisture – vegetation – atmosphere interactions has increased as researchers seek to understand the importance of soil moisture in the global system (Seneviratne et al, 2010). Soil moisture is a key component of land surface – atmosphere interactions, and has major influences on permafrost degradation (Wright et al, 2009; Quinton et al, 2011), as well as forest productivity and atmospheric fluxes (Baltzer et al, 2014; Barber et al, 2000). Ground surface temperatures are determined through the relationship between sensible and latent heat exchanges, where wetter soils increase the latent heat flux, decreasing the surface temperature. Wetter soils also result in large evapotranspiration rates that provide water and energy to the atmosphere fueling cloud formation and storms (Seneviratne et al, 2010). The soil moisture state and content further influences the amount of precipitation that can be stored before inducing overland flow (Seneviratne et al, 2010). For areas underlain by thawing permafrost, wetter soils coincide with the highest rates of permafrost thaw. This is due to the significantly larger thermal conductivity of water relative to air, which increases the rate and magnitude of heat transfer through the soil to the frost table, the impermeable boundary within the active layer that separates the overlying thawed soil from the underlying frozen soil (Hayashi et al, 2007). The frost table is commonly defined as the zero-degree isotherm, although due to a freezing-point depression the actual location of the frost table may occur only approximately at this isotherm (Carey & Woo, 2000). Numerous studies, such as Famiglietti et al. (2008), Western & Blöschl, 1999, Petrone et al. (2004), Brocca et al. (2007), Wright et al. (2009), and Quinton et al. (2011), have shown that soil moisture is highly variable across space and time in agricultural and Subarctic environments.

Characterizing the spatial variance of soil moisture allows proper representation of available water over a landscape (Famiglietti & Wood, 1994), especially for climate, hydrological and ecological modeling (Rodrigues-Iturbe, 2000), as well as remote sensing applications (Charpentier & Groffman, 1992). Soil moisture variance has been shown to follow predictable trends when examined over a range of moisture conditions (Famiglietti et al, 2008; Brocca et al, 2007). Standard deviation has often been reported to be highest at intermediate

moisture contents, overall representing a convex upward curve, while the coefficient of variation and skew both decrease with increasing moisture content (Famiglietti et al, 2008; Baroni et al, 2013; Brocca et al, 2007; Petrone et al, 2004). Contradictory findings on the normality of soil moisture measurements have been commonly reported. Hawley et al. (1983) and Anctil et al. (2002) found soil moisture was normally distributed at the field scale when the influence of topography was minimal. In studies where the redistribution of moisture was controlled by variability in other landscapes processes, such as topography, vegetation, precipitation and soil type, non-normal distributions of soil moisture were observed (Famiglietti et al, 1999; Famiglietti et al, 2008; Petrone et al, 2004). At the field scale soil moisture is primarily controlled by soil type, vegetation and topography, and by precipitation, slope, and aspect at large regional scales (Famiglietti et al, 2008; Petrone et al, 2004). In addition, Wright et al. (2009) found that in permafrost peatlands frost table depth was also a significant control on soil moisture variability. However, a major limitation of the literature on soil moisture variability is that the bulk of the literature has been exclusively conducted in agricultural fields with relatively homogeneous mineral soils with minimal and well-studied vegetation.

Vegetation changes have been observed over the Subarctic (Verbyla, 2008; Llyod & Fastie, 2012; Jorgenson & Osterkamp, 2005), notably Normalized Difference Vegetation Index (NDVI) trends have emerged showing distinct patterns in changing biomass (Verbyla, 2008). Increases in biomass have been limited to wetter and cooler regions that are energy limited, whereas declines have been noted in the dry continental interior of the Subarctic. Mean annual temperatures, and thus evapotranspiration rates, are expected to rapidly increase in the Subarctic; however, this may not necessarily be matched by a corresponding increase in total precipitation (Jorgenson & Osterkamp, 2005). Carey & Woo (2001) found that evapotranspiration was one of the major components of the water balance in Subarctic Yukon. Due to the impermeable nature of the frost table, the water table is perched within the soil layers above this boundary (Quinton & Hayashi, 2005). The elevation of the frost table therefore defines the minimum elevation of the water table. Decreasing frost table elevation, and associated water table, coupled with increasing evapotranspiration demands may result in significant moisture deficits for Subarctic forests. This temperature induced drought stress is a prominent hypothesis explaining the observed declines in forest biomass, productivity and carbon sequestration (Barber et al, 2000).

In addition, Ford et al. (2005) and Bovard et al. (2005) found that during an agricultural drought, transpiration rates become increasingly dependent on available soil moisture, as opposed to radiation and vapour pressure deficit. Agricultural drought specifically refers to soil moisture conditions that do not meet vegetation requirements at a certain location and point in time. As such, soil moisture is an essential component of models used to predict the response of northern Boreal forests to climate change.

To date, little research has examined soil moisture - vegetation - atmosphere interactions and their spatial distribution in a Subarctic permafrost peatland. Therefore the objectives of this thesis were to:

1. Statistically characterize the spatial distribution of soil moisture and its relationship to permafrost, vegetation, and topography, and to
2. Examine the importance of the relationship between soil moisture, and frost and water table depth on *Picea mariana* sap velocity.

Advancing our understanding of the spatial variability of soil moisture and its feedbacks to other environmental processes will benefit the monitoring and management of permafrost, and northern water and carbon cycles in a changing climate.

## **Chapter 2.0: Statistical characterization of soil moisture variability in a Subarctic permafrost peatland and its relationship to permafrost, vegetation and topography**

### **Abstract**

The Canadian Subarctic is experiencing one of the highest rates of climate change, impacting local and global cycles of water, energy, and carbon. Furthermore, soil moisture has been linked to increases in permafrost degradation. Extensive research has been done on soil moisture variability in mineral soils, while little is known for peat soils underlain by permafrost. Characterizing soil moisture variability in Subarctic peatlands will allow better understanding of the distribution of soil moisture and related processes. Soil moisture and frost table depth were measured in 2012 and 2013 within a 140 by 520 m plot. Multiple topographical indices and incident radiation were calculated for the study plot. Leaf Area Index (LAI) was measured at the end of August 2013. The relationship between mean soil moisture and its standard deviation, coefficient of variation, and skew generally agreed with trends observed in mineral soils. Standard deviation increased with mean moisture, while the coefficient of variation and skew decreased. All days of soil moisture measurements were positively skewed. On plateaus, soil moisture was positively correlated to frost table depth attributable to the feedback processes between soil moisture and permafrost degradation, where wetter soils increase thaw rates. Micro-topography, radiation and LAI were not significant controls on soil moisture; however, higher mean moisture contents were found under mosses than under lichens, though this may be more indicative of general topographical differences. These findings will improve the remote sensing of soil moisture in Subarctic peatlands allowing for the acquisition of accurate large-scale soil moisture retrievals. These products will provide large spatial or temporal measurements over the large inaccessible regions of the Canadian north for environmental monitoring and modelling.

### **2.1 Introduction**

Near surface soil moisture is highly spatially variable (Famiglietti et al, 2008; Western & Blöschl, 1999) and governs many landscape processes, including evapotranspiration, runoff, infiltration, groundwater recharge, root water uptake, and atmospheric variability (Seneviratne et al, 2010; Vereecken et al, 2010). In permafrost regions, soil moisture also plays a critical role in

determining the rate and extent of permafrost thaw (Wright et al, 2008; Quinton et al, 2011), and is essential for non-vascular plants, such as mosses, and shallowly-rooted shrubs and trees (Oechel & Sveinbjornsson, 1978) common in northern latitudes (Zoltai & Tarnocai, 1975). Large-scale permafrost degradation has occurred dramatically in the last half century with the greatest rates observed in the Subarctic (Jorgenson & Osterkamp, 2005; Beilman & Robinson, 2003; Prowse et al, 2009). In the southern wetland-dominated portion of the Canadian Subarctic, permafrost is found predominately within peat plateaus, elevated features surrounded by saturated, permafrost-free channel fens and flat bogs (Quinton et al, 2011; Quinton et al, 2003).

Understanding the spatial variability of soil moisture within a distinct environment, allows better representation of its distribution over the landscape than solely the mean moisture value from a representative point measurement or remotely sensed pixel (Famiglietti & Wood, 1994). Characterizing the variability will improve water and energy balance models (Brocca et al, 2007; Hawley et al, 2007; Famiglietti et al, 2008), as well as the accuracy of remotely sensed soil moisture products (Charpentier & Groffman, 1992) ideal for monitoring soil moisture, and associated permafrost degradation, over vast inaccessible regions such as the Canadian north (Seneviratne et al, 2010). However, contradictory findings of the patterns and controls of soil moisture variability are prevalent in the literature (Famiglietti et al, 2008; Famiglietti et al, 1998; Hawley et al, 1983), although some of these differences can be explained by the specific environmental conditions and sampling design of individual studies (Famiglietti et al, 1998). Soil moisture variability is determined using statistical properties, notably: the probability density function (PDF); mean; standard deviation (SD); coefficient of variation (CV); skew (SK), and regression analyses. Geostatistical methods may also be used, but are not examined in this study.

Regardless of the contradictions in the literature, general patterns of soil moisture variability have emerged from which scaling properties have been determined. Standard deviation has been shown to peak at intermediate moisture contents with the overall trend representing a convex upward curve. The coefficient of variation and skew both decrease with increasing moisture content (Famiglietti et al, 2008; Baroni et al, 2013; Brocca et al, 2010; Brocca et al, 2012; Brocca et al, 2007; Petrone et al, 2004). Hawley et al. (1983) and Anctil et al.

(2002) found soil moisture was normally distributed in flat agricultural fields, although the PDF, and therefore normality, may be influenced by topography (Brocca et al, 2007; Hawley et al, 1983) or wetness conditions (Brocca et al, 2012; Brocca et al, 2007). As such, some studies have also reported non-normal moisture distributions (Famiglietti et al, 1999; Famiglietti et al, 2008; Petrone et al, 2004). These trends are the result of variations in soil texture, topography, vegetation, and precipitation affecting the rate and pattern of infiltration and redistribution (Brocca et al, 2007; Hawley et al, 1983; Famiglietti et al, 2008). In peat soils, the concept of soil texture cannot be readily transferred as these soils are composed of decomposing and living vegetation rather than individual grains. Therefore, soil porosity is used as a surrogate.

Landscape features and processes exert controls on the spatial distribution of soil moisture. Soil porosity, topography, and vegetation regulate infiltration, runoff and evapotranspiration thus influencing the mean and variation of soil moisture (Hawley et al, 1983; Famiglietti et al, 2008). Topography itself encompasses aspect, slope and the location on the slope (hereby referred to as relative elevation). Aspect controls the amount of incident radiation the ground surface is receiving thereby controlling evapotranspiration and energy fluxes (Ried, 1973). Slope controls the redistribution of water where steeper slopes are often drier than flatter areas (Hawley et al, 1983). Relative elevation describes the difference between the top and bottom of a slope. In studies by Hawley et al. (1983), Engstrom et al. (2005) and Henniger et al. (1976) soil moisture content was consistently higher at the bottom of the slope even in fields with a gentle topography. However, many frequently used terrain indices do not inherently describe redistribution mechanisms, therefore numerous authors have preferred to use Beven & Kirkby (1979)'s wetness index, slope or aspect (Brocca et al, 2007). In addition to topography, the overall wetness also influences the controls on soil moisture. In wet conditions, particularly after a large rainfall, soil moisture variability is controlled by soil porosity whereas during dry conditions variability is controlled by relative elevation, radiation (as related to aspect), and vegetation (Famiglietti et al, 1998; Hawley et al, 1983; Teuling & Troch, 2005).

Overall, while soil moisture variability in mineral soils has been well established in the literature (e.g. Crow et al. (2012)), few studies have examined soil moisture variability in organic soils and even fewer within northern environments (Ancitil et al, 2002; Petrone et al, 2004;

Chapin et al, 2000). Few studies are available that characterize soil moisture spatial variability at relatively small scales (< 500 m<sup>2</sup>) and its related processes in a peatland underlain by permafrost. Previous studies by Hewelke et al. (2014) and Szatyłowicz et al. (2007) similarly investigated soil moisture variability in organic peat soils; however, these studies differ with respect to experimental design and disturbance, scale, soil porosity and the presence of permafrost. Therefore the objectives of this study are 1) to characterize the spatial variability of near surface soil moisture, and 2) to determine the relationship between soil moisture and the distribution of permafrost, vegetation, and topography in a Subarctic peatland.

## **2.2 Methods**

### 2.2.1 Study Site

#### *2.2.1.1 Subarctic peatlands*

Subarctic peatland complexes consist of three main landscape features: channel fens, flat bogs and peat plateaus (Zoltai & Tarnocai, 1975) (Figure 2.1a). Channel fens are vegetated channels on average 50 to 100 meters across (some may be almost 1 km wide) without underlying permafrost (Quinton et al, 2009) that transmit water through the basin (Quinton et al, 2003). A floating peat mat 0.5 to 1 meter thick typically covers the channel surface supporting *Carex spp.* (minerotrophic sedges), and various species of grasses, shrubs, and plants including *Typha latifolia* (common bulrush), *Equisetum fluviatile* (water horsetail), and *Larix laricina* (Tamarack) (Quinton et al, 2009; Zoltai & Tarnocai, 1975). Flat bogs, also referred to as collapse scars, are *Sphagnum spp.* (moss) dominated, permafrost-free features that range in size from small isolated depressions within a plateau to encompassing multiple plateaus (Quinton et al, 2009; Zoltai & Vitt, 1995). Bogs primarily store water and are less hydrologically connected than channel fens as fill and spill processes only occur during the spring freshet and large precipitation events (Quinton et al, 2003; Connon et al, 2014). In some cases, bogs may be completely hydrologically isolated losing water only to internal drainage. Peat plateaus are peat mounds underlain by permafrost, and standing on average one to two meters above the surrounding fens and bogs (Robinson & Moore, 2000; Vitt et al, 1994). Peat plateaus act as runoff generators that rapidly infiltrate water from precipitation due to the large near-surface hydraulic conductivity and porosity into an unsaturated zone within the active layer. The sloped flanks of the plateaus result in a hydraulic gradient that redistributes water throughout the active

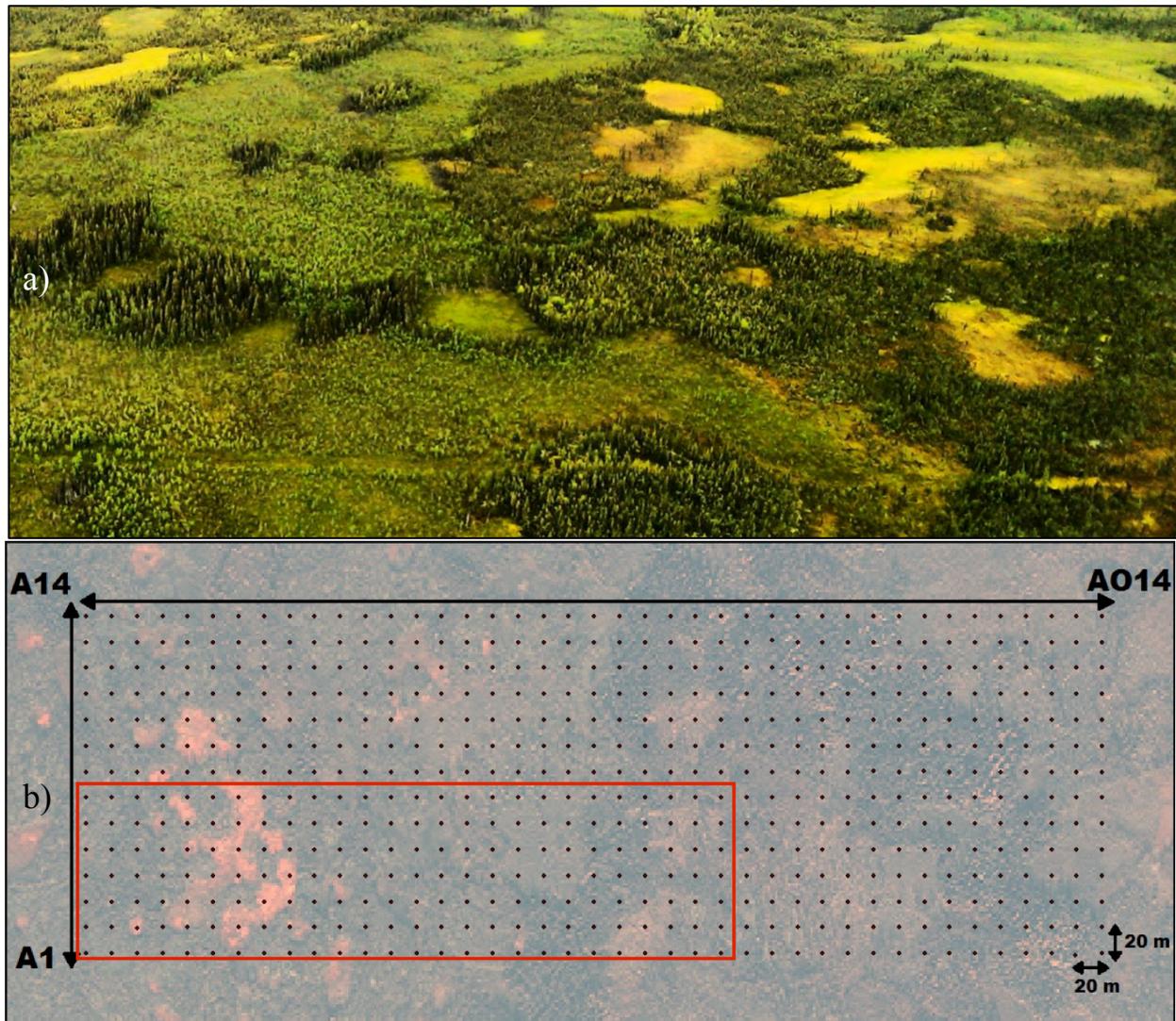


Figure 2.1: a) Oblique aerial photograph of the study plot showing the three peatland landscape features: channel fens (light green areas), flat bogs (yellow areas), and plateaus (forested dark green areas). b) The study site (in red) nested within the larger pre-established grid. Measurements were taken every 20 m. North is at the top of the plot.

layer and directs the infiltrated precipitation into adjacent fens and bogs (Quinton et al, 2003; Quinton & Gray, 2003). The plateau interior is relatively flat, or slightly concave in cases where a bog is contained within the plateau. Plateau vegetation includes *Picea mariana* (Black spruce), *Cladina mitis* (Reindeer lichen), *Pleurozium schreberi* (feather moss), *Rhododendron groenlandicum* (bog Labrador tea), *Andromeda glaucophylla* (bog rosemary) (Quinton et al, 2009), and *Sphagnum spp.* Peat plateaus cover approximately 60% of the study plot.

### 2.2.1.2 Study site

Research was conducted at Scotty Creek (61°18'N, 121°18'W), a 152 km<sup>2</sup> peatland-dominated basin south of Fort Simpson, a town in the Dehcho region of the Northwest Territories, Canada at the confluence of the Liard and Mackenzie Rivers. This region is within the discontinuous permafrost zone and has a dry continental climate with short summers and long, cold winters. Mean annual air temperature is just below 0 °C (MSC, 2015) and average annual precipitation is approximately 369 mm of which 46% is snow (Quinton et al., 2009). Scotty Creek has a low drainage density of 0.016 km km<sup>-2</sup> and relatively uniform topography with elevations within the actual study plot ranging less than 3 meters between the highest and lowest points (Hayashi et al, 2004). Peat is the dominant soil type and covers the entire study plot. The peat soils present consist of a surface layer of living and partially decomposed vegetation over approximately 8 meters of increasingly decomposed peat (Aylesworth & Kettles, 2000; Quinton et al, 2009). Mineral soils are present beneath the peat layer. Scotty Creek has a wet soil regime immediately after snowmelt at the onset of active layer development, followed by a dry soil regime as the active layer thickens (Quinton & Gray, 2003; Quinton et al, 2009).

### 2.2.2 Experimental design and data collection

A 140 x 520 m subset of a larger research plot (data not published) was used with previously placed and georeferenced stakes located every 20 meters (Figure 2.1b). The plot encompassed all three described peatland features (i.e. fens, bogs and plateaus) and is undisturbed. Near surface volumetric soil moisture (0 – 5 cm) (SM) was collected at each stake using a Stevens Hydra Probe (Stevens Water Monitoring Systems Inc, Oregon, USA) or ThetaProbe Soil Moisture Sensor (Delta-T Devices, Cambridge, UK) on 3 days during the summer of 2012 (July 19, July 31 and August 2) and 4 days in 2013 (May 27, May 31, August 27 and August 28). Thirty-eight gravimetric soil moisture samples were taken using an aluminum core (6 cm in height and 7.5 cm in diameter corresponding to the approximate sampling volume measured by both soil moisture probes) and used to determine a calibration equation for the peat soils at the study site. Three soil moisture measurements were also taken around the extracted core using both the Stevens Hydra Probe and ThetaProbe. The moisture content of the gravimetric samples varied from very dry to nearly saturated. Volumetric soil moisture was calculated using similar methods as Rowlandson et al. (2013). The samples were

weighed with a Mettler Toledo PB1502-S precision balance and then dried in a Fisher Scientific Isotemp lab oven at 50 °C for 4 days. A calibration equation was determined for both soil moisture probes based on the relationship between the measured and calculated volumetric moisture contents. The equation for the Stevens Hydra Probe had a  $R^2$  of 0.90 and RMSE of 0.0488  $\text{m}^3/\text{m}^3$ , and the ThetaProbe equation had a  $R^2$  of 0.92 and RMSE of 0.0436  $\text{m}^3/\text{m}^3$  (Merchant, 2015). The depth to the frost table, the impermeable boundary within the active layer separating the overlying thawed soil from the underlying frozen soil (Hayashi et al, 2007), was measured using a graduated steel rod on July 19 2012, and May 27, 31 and August 27 2013 at the same locations as soil moisture. Measurements are presented as positive values corresponding to the distance between the ground surface and the frost table (e.g. a frost table depth of  $x$  cm indicates that the frost table is  $x$  cm below the ground surface). Leaf Area Index (LAI) was collected at the end of August 2013 using a LAI-2200 Plant Canopy Analyzer (LI-COR, Lincoln, USA). However, due to a processing issue only data from four of the seven transects was usable. A one-meter resolution LiDAR image was acquired from 2010 and used to derive a digital elevation model (DEM) and subsequent topographical, wetness and radiation indices. Fens and bogs are permafrost-free, saturated features and therefore soil moisture measurements were not taken in these locations. As such, additional measurements at plot stakes in fens or bogs were removed from the analysis as they may have a significant impact on observed trends. Current remote sensing techniques could be modified for use in Subarctic peatlands to account for the differences in soil moisture variance between plateaus, fens and bogs when developing the final product.

### 2.2.3 Data Processing

Various indices were developed from the LiDAR DEM using Whitebox Geostatistical Analysis Tools v 3.2.1, and open-source geomatics software, including difference from mean elevation (DFME), topographical ruggedness index (TRI) (Riley et al, 1999), and the wetness index (WI) (Beven & Kirkby 1979). DFME is calculated as the difference between the elevation at a point and its specified local neighbourhood. The DFME was determined based on the values in a 3 by 3 window around the central cell. This was thought to better account for the relative elevation of a point as micro-topography plays an important role in redistribution in this environment (Wright et al, 2009). The TRI was also calculated as a comparable index, which

measures the root mean square deviation and residuals between each point and its neighbours (Riley et al, 1999). The wetness index describes the likelihood of a point to be saturated given the specific contributing area and local slope (Beven & Kirkby 1979). The wetness index used was calculated as

$$WI = \ln \frac{a}{\tan b} \quad [ 2.1 ]$$

where  $a$  is the specific contributing area, and  $\tan b$  is the local slope in degrees (Beven & Kirkby 1979). Slope was calculated following Horn's (1981) 3<sup>rd</sup> order finite difference method. The specific contributing area was determined using the D-infinity flow accumulation algorithm, which allows for flow divergence (Tarboton, 1997), and was shown to be the accumulation index most highly correlated to soil moisture in peat soils (Sørensen et al. 2006).

Daily incident radiation was calculated for each cell in the DEM on each of the 7 measurement days using the model presented in Dingman (2008). The model is based on the physical transfer of solar radiation through the atmosphere accounting for the influence of surface slope and aspect. A number of additional meteorological parameters are required in the model including albedo (determined based on the values presented by Betts & Ball (1997), optical air mass and attenuation of radiation from dust (from Bolsenga (1964)), and the dew point temperature at the Fort Simpson Airport (Environment Canada, 2015). It should be noted that this model does not account for the effects of vegetation. Model outputs were compared to radiation measurements taken at two stations Scotty Creek; one in an open bog and one on a plateau with a canopy (Table 2.1). Modelled radiation was similar to daytime (7:00 - 20:00) average radiation on each of the sampling days, although there are notable differences, specifically on August 2 2012, likely due to atmospheric conditions (i.e. clouds) and vegetation.

#### 2.2.4 Statistical Analysis

Boxplots of soil moisture and frost table depth were developed using R v 3.1.2 (R Development Core Team, 2014). Plots of the probability density function of soil moisture were similarly constructed for each sampling period. Soil moisture standard deviation, coefficient of variation, and skew were calculated for the plot on each sampling day and plotted against mean

Table 2.1: Comparison of radiation ( $W/m^2$ ) calculations between the model and observations at two stations at Scotty Creek. Radiation was averaged for the entire day (E.D.) and daylight hours (7:00 to 20:00) (D.H.). Measurements were not available for the last 2 dates in 2013 so the closest sunny day (August 23, 2013) is presented instead.

| Date         | Model | Bog (E.D.) | Bog (D.H.) | Plateau (E.D.) | Plateau (D.H.) |
|--------------|-------|------------|------------|----------------|----------------|
| July 19 2012 | 313   | 166        | 323        | 160            | 306            |
| July 31 2012 | 286   | 162        | 295        | 158            | 286            |
| Aug 2 2012   | 277   | 100        | 208        | 104            | 205            |
| May 27 2013  | 329   | 175        | 349        | 151            | 294            |
| May 31 2013  | 333   | 131        | 275        | 116            | 236            |
| Aug 27 2013  | 232   | NA         | NA         | NA             | NA             |
| Aug 28 2013  | 231   | NA         | NA         | NA             | NA             |
| Aug 23 2013  |       | 127        | 233        | 130            | 231            |

soil moisture, similar to the analysis done by Famiglietti et al. (2008). Coefficient of variation in particular is an important statistical descriptor of soil moisture as it allows comparison between samples with different means (Brocca et al, 2012). As well, plots of the coefficient of variation and mean soil moisture tend to have a predictable decreasing exponential trend even when there are few samples (Bell et al, 1980). These plots allow researchers to determine under what moisture conditions predicted soil moisture would be more prone to estimation errors (Famiglietti et al, 2008). Frost table depth mean, standard deviation, coefficient of variation, and skew were also calculated. Soil moisture, frost table depth, DFME, TRI, WI, radiation, and LAI were correlated using Pearson's r for overall average values at each individual point in the study plot.

## 2.3 Results and Discussion

### 2.3.1 Characterization of soil moisture spatial variability

#### 2.3.1.1 Statistical descriptors

Soil moisture in the study plot remained relatively low throughout the study period, where mean volumetric soil moisture did not exceed  $0.31 \text{ m}^3/\text{m}^3$  for non-saturated areas (Figure 2.2, Table 2.2). Greater ranges in moisture contents were observed when mean soil moisture was higher, specifically in May after snowmelt, and after large precipitation events (August 28, 2013) (Figure 2.2). Based on precipitation records from the Fort Simpson Airport (MSC, 2015), the

period between May and August in 2012 was wetter than the same time period in 2013. As well, soil moisture in 2013 was less variable and closer to a normal distribution. However, this may be affected by the difference in sampling dates between years. Overall, total precipitation received during the study period was approximately average for Scotty Creek in 2012, while slightly less than average in 2013. Mean plateau soil moisture ranged from 0.17 to 0.31  $\text{m}^3/\text{m}^3 \pm 0.11$  to 0.22 (Table 2.2). Soil moisture coefficient of variation and skew ranged from 0.52 – 1.21 and 1.04 - 3.25 respectively (Table 2.2). Statistical descriptors of soil moisture variance, as well as related site characteristics, are reported from notable studies in Table 2.3 for comparison with the results presented here. Organic soils overall appear to have higher mean moisture contents and variance than mineral soils, though the mean moisture content found in this study is relatively drier than those stated in comparable studies in organic soils. This is likely due to the removal of saturated areas in this analysis. Peat is highly porous, and therefore able to store more water. However, due to this high porosity, water in near surface soils is rapidly redistributed into lower soil layers near the frost table (Quinton & Gray, 2003). Soil moisture variance is greatest in organic soils underlain by permafrost, as shown in this study and by Engstrom et al. (2005). Porosity and hydraulic conductivity decrease with depth (Quinton & Gray, 2003), therefore variability in the thickness of the active layer can result in varying soil properties, and thus soil moisture, throughout the landscape.

Table 2.2: Statistics of the near surface (0 – 5 cm) volumetric soil moisture ( $\text{m}^3/\text{m}^3$ ) variance on a peat plateau for each sampling day.

| <b>Date</b>  | <b>Mean</b> | <b>SD</b> | <b>CV</b> | <b>SK</b> |
|--------------|-------------|-----------|-----------|-----------|
| July 19 2012 | 0.17        | 0.20      | 1.21      | 2.80      |
| July 31 2012 | 0.17        | 0.15      | 0.92      | 3.25      |
| Aug 2 2012   | 0.15        | 0.13      | 0.91      | 2.96      |
| May 27 2013  | 0.26        | 0.22      | 0.84      | 1.75      |
| May 31 2013  | 0.21        | 0.17      | 0.79      | 1.96      |
| Aug 27 2013  | 0.17        | 0.11      | 0.63      | 1.04      |
| Aug 28 2013  | 0.31        | 0.16      | 0.52      | 1.83      |

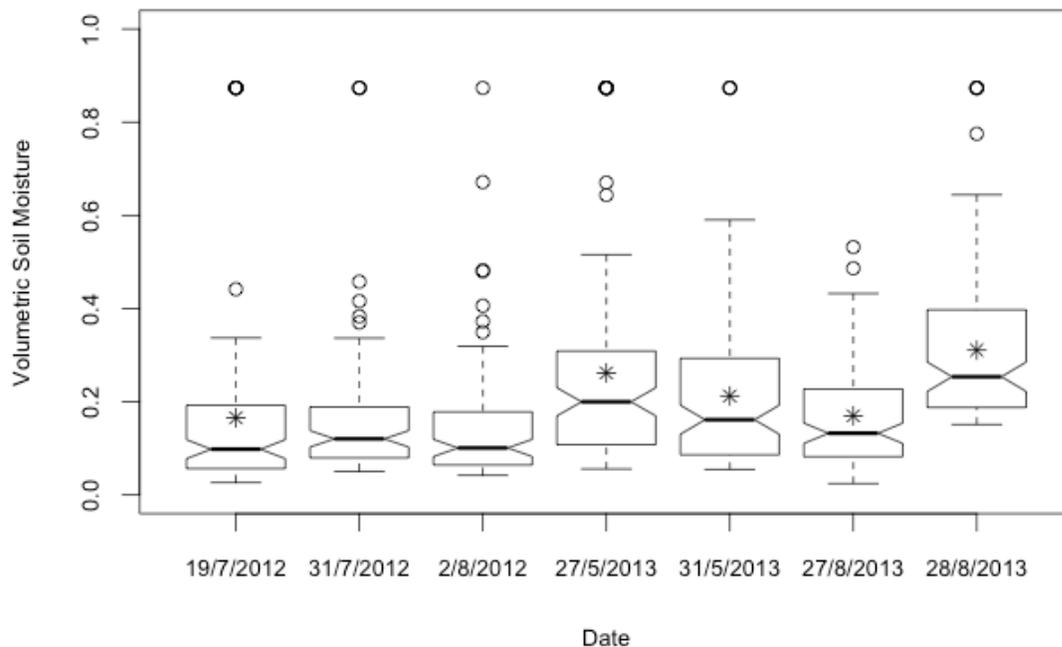


Figure 2.2: Distribution of near surface (0 – 5 cm) volumetric soil moisture measurements ( $\text{m}^3/\text{m}^3$ ) taken on each sampling day. The boxplot shows the interquartile ranges of measured soil moisture, with the dark horizontal line depicting the median, and the asterisk the mean. The whiskers show the range of mean values within 1.5 times the interquartile range, with outliers identified by the circles. Saturated areas have been excluded

### 2.3.1.2 Trends in soil moisture variance at the field-scale

General trends of soil moisture variance over the entire study site showed that soil moisture standard deviation increased with increasing mean moisture (Figure 2.3a); however, the mean soil moisture values are all clustered in the lower range of values. The coefficient of variation and skew decreased with mean soil moisture, which was also found by Brocca et al. (2007), Famiglietti et al. (2008), and Baroni et al. (2013) (Figures 2.3b and 2.3c). These relationships, coupled with the difference in variance statistics between 2012 and 2013 (Table 2.2) suggest that soil moisture coefficient of variation and skew are at a minimum in wet conditions, while standard deviation is lowest in dry conditions. Extending this analysis across the range of observable moisture conditions will greatly assist in characterizing soil moisture variance in a Subarctic peatland. The probability function distributions of soil moisture were highly positively skewed and non-normal for all measurement days (Figure 2.4). Days following precipitation remained skewed, but were shifted such that the lowest moisture contents were slightly wetter than dry days. While the topic of normality in soil moisture studies is often

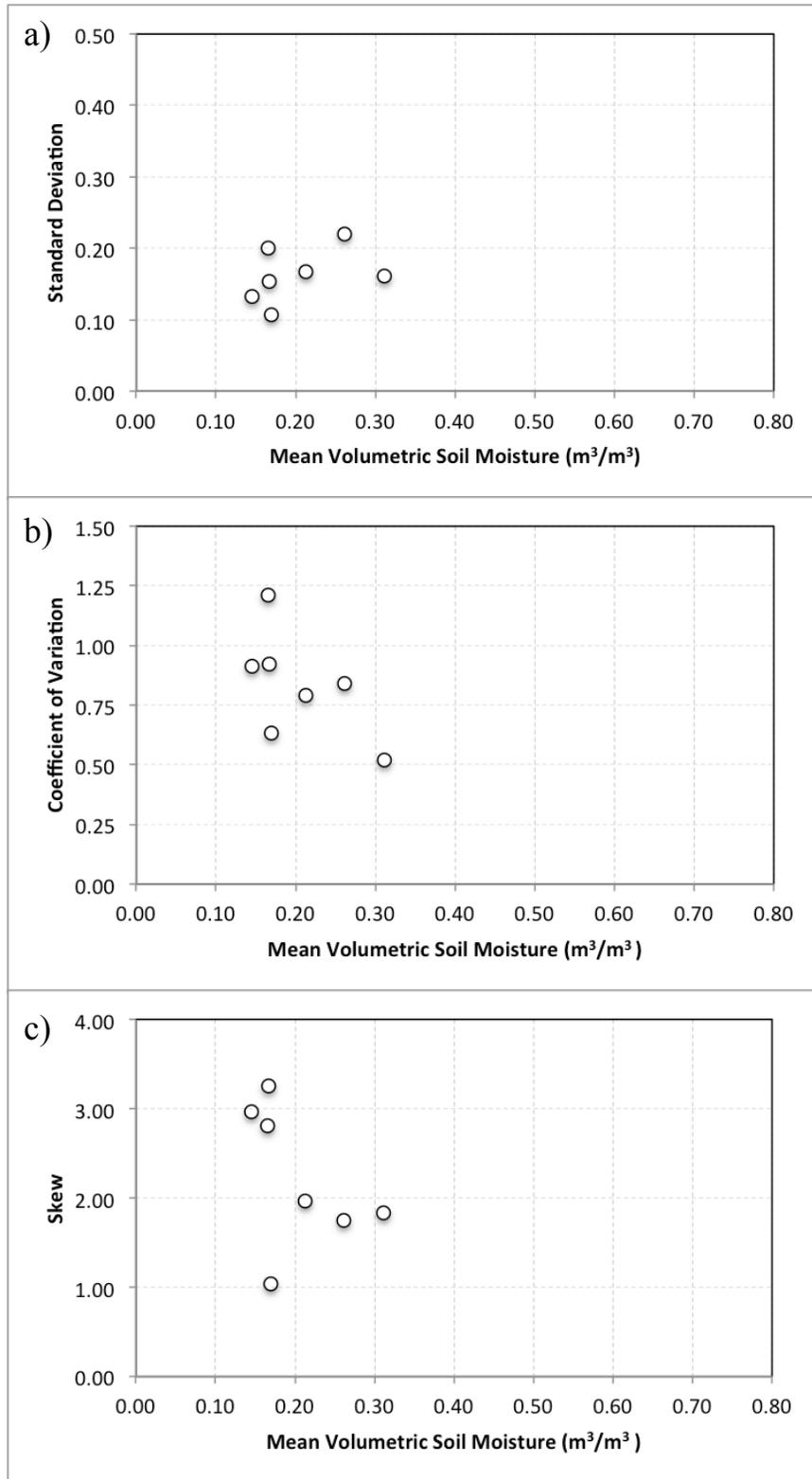


Figure 2.3: Relationships between daily mean volumetric soil moisture and the a) standard deviation, b) coefficient of variation, and c) skew. These were calculated for each sampling day using all the measurements taken on plateaus within the plot.

Table 2.3: Comparison of soil moisture statistical properties in various studies. Studies included measurements in mineral and organic soils with and without permafrost.

| <b>Study</b>            | <b>Soil</b> | <b>Scale</b>        | <b>Depth (cm)</b> | <b>Permafrost</b> | <b>Mean</b> | <b>SD</b>   | <b>CV</b>   |
|-------------------------|-------------|---------------------|-------------------|-------------------|-------------|-------------|-------------|
| Hawley et al, 1983      | mineral     | 76.3 ha             | 15                | No                | 0.25        | 0.05        | 0.21        |
| Famiglietti et al, 1998 | mineral     | 200 m <sup>*</sup>  | 5                 | No                | 0.37        | na          | 0.21        |
| Brocca et al, 2010      | mineral     | 60 x 50 m           | 15                | No                | 0.09 - 0.52 | 0.01 - 0.05 | na          |
| Brocca et al, 2012      | mineral     | 242 km <sup>2</sup> | 15                | No                | 0.08 - 0.44 | 0.02 - 0.06 | 0.10 - 0.35 |
| Yang et al, 2011        | mineral     | 80 x 100 m          | 10                | Yes               | 0.36        | 0.05        | 0.15        |
| Engstrom et al, 2005    | organic     | 0.5 km <sup>2</sup> | 7                 | Yes               | 0.52        | 0.17        | 0.33        |
| Anctil et al, 2002      | organic     | 450 m <sup>*</sup>  | 5                 | No                | 0.32        | 0.05        | 0.17        |
| Petrone et al, 2004     | peat        | 1200 m <sup>2</sup> | 12                | No                | 0.78        | 0.1         | 0.13        |

contradictory in the literature, soil moisture in this study never resembled a normal distribution regardless of wetness conditions. However, the soil studied was either relatively uniformly dry or saturated; whereas Famiglietti et al. (1999) and Brocca et al. (2012) found that soil moisture followed a normal distribution at intermediate moisture contents. Hawley et al. (1983) found that soil moisture was normally distributed only in homogeneous fields; therefore the heterogeneity (topography, frost table depth, and vegetation) present at Scotty Creek likely plays an important role in the redistribution of soil moisture across the landscape resulting in the non-normal distributions observed (Brocca et al, 2007). Based on QQ-plots, soil moisture more closely follows a lognormal distribution than normal at Scotty Creek; however, there is still notable skew in the distribution mostly due to a few exceedingly wet locations. This lognormal distribution may be the result of the high near-surface hydraulic conductivity observed on plateaus resulting in exceptionally dry surface soils.

Table 2.4: Frost table depth mean and standard deviation (to nearest cm), coefficient of variation and skew statistics on a peat plateau each measurement day.

| <b>Date</b>  | <b>Mean</b> | <b>SD</b> | <b>CV</b> | <b>SK</b> |
|--------------|-------------|-----------|-----------|-----------|
| July 19 2012 | 47          | 20        | 0.44      | 3.37      |
| May 27 2013  | 30          | 13        | 0.44      | 2.77      |
| May 31 2013  | 34          | 11        | 0.34      | 3.71      |
| Aug 27 2013  | 75          | 34        | 0.46      | 1.61      |

### 2.3.1.3 Frost table variance

The range of measured frost table depths is presented on each of the sampling days in Figure 2.5. Mean frost table depth, as expected, increased over the summer with a maximum mean depth of approximately 75 cm, as well as standard deviation (Table 2.4). Maximum frost table depth in this study was greater than depths reported by Wright et al. (2009) for another site at Scotty Creek between the years 2002 and 2006. The coefficient of variation remained constant over the study period (Table 2.4). The large range and standard deviation indicate that frost table depth is also highly spatially variable across the landscape; however, outlier measurements of shallower frost table depth, especially in July or August (see Figure 2.5) are likely due to

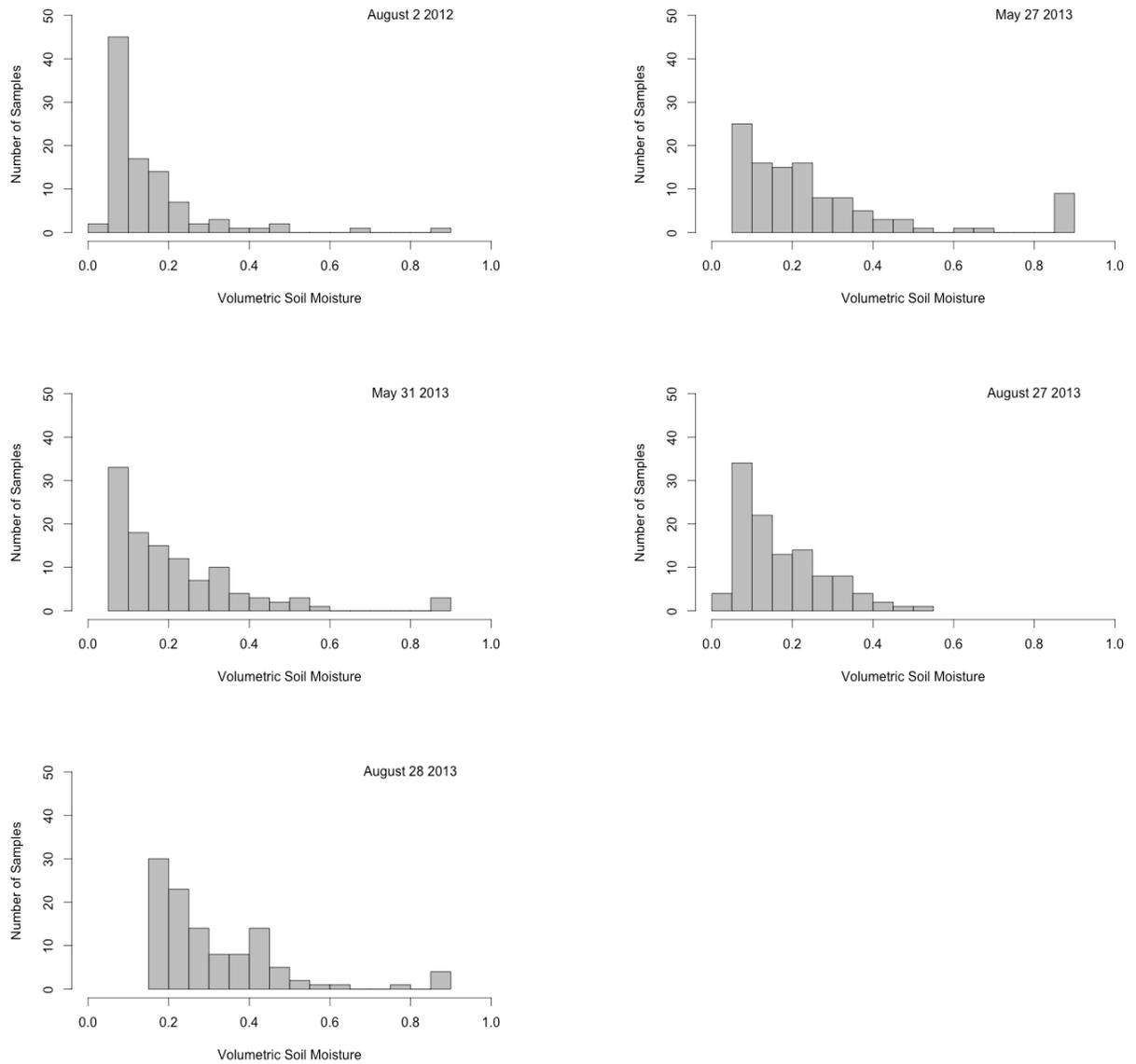


Figure 2.4: Plots of the probability density (histogram) of volumetric soil moisture ( $m^3/m^3$ ) on each sampling day. Soil moisture was positively skewed in all cases.

measurement error and may bias these findings. As the frost table thaws, it is not uncommon for dense root patches or ice lenses in the active layer to obstruct the insertion of the probe. Care should be taken to ensure the probe has reached the frost table in future studies.

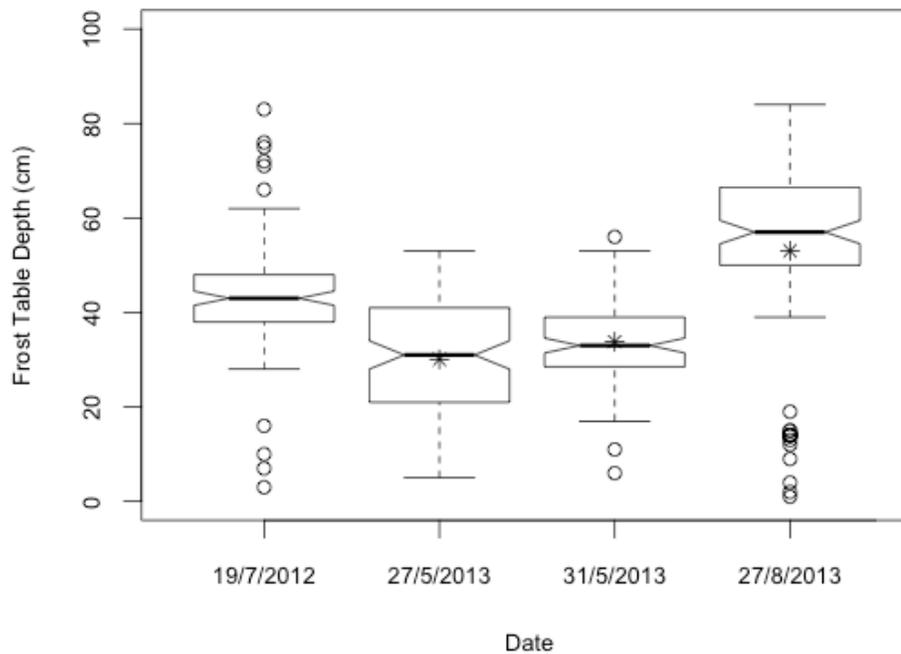


Figure 2.5: Distribution of frost table measurements taken on each sampling day. Fens and bogs do not have underlying permafrost or the permafrost is greater than 2 meters deep.

### 2.3.2 Controls on soil moisture variability

Soil moisture redistribution is controlled by various environmental factors, including soil porosity, vegetation, topography and radiation (Hawley et al, 1983; Famiglietti et al, 2008). Table 2.5 provides a correlation matrix between soil moisture, and frost table variance statistics, LAI, radiation and multiple topographical indices. No relationship was apparent between soil moisture and LAI, micro-topography, and radiation. Instead, soil moisture was primarily related to frost table depth, where high moisture contents are associated with deeper frost tables.

In subarctic peatlands, the elevation of the frost table primarily controls subsurface hydrology, as hydraulic conductivity decreases as a function of depth at Scotty Creek (Quinton

& Gray, 2003; Quinton & Hayashi, 2005). Thus frost table elevation determines the rate and magnitude of flow, storage, and overall basin hydrology. Carey & Woo (2001) at Wolf Creek in Subarctic Yukon, and McNamara et al. (1998) at Kuparuk River Basin in Arctic Alaska found similar relationships. For example, shallow frost table depths are associated with higher runoff rates due to the large hydraulic conductivity at the near surface. Conversely, as the frost table lowers into the zone of low hydraulic conductivity the rate of flow decreases and more water is stored within the active layer. Due to the impervious nature of the frost table, a perched water table may form in the layer immediately above this boundary (Quinton & Hayashi, 2005). As the frost table deepens, the water table also increases in depth, thickening the unsaturated zone within the active layer. The lowering of the water table consequently also leads to a smaller hydraulic gradient between the plateau and adjacent fens and bogs, further decreasing subsurface flow. As well, Wright et al. (2009) found that depressions in the frost table caused preferential accumulation of water leading to increased rates of permafrost thaw generating a positive feedback loop. This is due to the presence of water increasing the rate and magnitude of heat transfer through the soil (Hayashi et al, 2007). While the near surface soil layers become uniformly dry, as seen in Figure 2.2, the outliers of higher soil moisture promote preferential thaw instigating the soil moisture – thaw feedback process.

Plateaus are elevated runoff generators that transmit water across the frost table and into the adjacent low-lying fens and bogs (Quinton et al, 2003; Quinton & Gray, 2003). It is these general topographical differences that are related to whether an area will be overall wetter (fens and bogs) or drier (plateaus). Of these three peatland types an unsaturated zone only occurs within plateaus, as the water table is near or above the ground surface in fens and bogs (Quinton et al, 2003). As shown, plateau micro-topography alone does not greatly impact soil moisture redistribution on the plateaus themselves (Table 2.5). This may be in part due to the inability of the topographical indices used to capture the governing redistribution mechanisms (Brocca et al, 2007), or the micro-topography itself. As well, Wright et al. (2009) found that the topography of the frost table, rather than that of the ground surface, primarily influences soil moisture redistribution. It is interesting to note that variance in frost table depth was moderately correlated to TRI and LAI, which therefore may be indirectly related to soil moisture.

Table 2.5: Pearson's r correlation coefficients for the associations between volumetric soil moisture and various environmental controls on a peat plateau. Significant correlations are denoted with an asterisk.

|         | Mean SM | SD SM | CV SM | SK SM  | Mean FT | SD FT  | CV FT  | SK FT | WI     | TRI    | DFME   | LAI   | Mean Rn |
|---------|---------|-------|-------|--------|---------|--------|--------|-------|--------|--------|--------|-------|---------|
| Mean SM | 1.00    | 0.78* | 0.00  | -0.41* | 0.31*   | 0.33*  | 0.32*  | 0.21* | 0.03   | -0.05  | 0.15   | -0.22 | 0.07    |
| SD SM   | 0.78*   | 1.00  | 0.58* | -0.05  | 0.19*   | 0.18   | 0.17   | 0.11  | -0.02  | -0.03  | 0.11   | -0.03 | 0.00    |
| CV SM   | 0.00    | 0.58* | 1.00  | 0.49*  | -0.12   | -0.14  | -0.11  | -0.07 | -0.08  | 0.00   | 0.00   | 0.26  | -0.07   |
| SK SM   | -0.41*  | -0.05 | 0.49* | 1.00   | -0.07   | -0.14  | -0.20* | -0.10 | 0.02   | 0.00   | -0.05  | -0.11 | -0.10   |
| Mean FT | 0.31*   | 0.19* | -0.12 | -0.07  | 1.00    | 0.86*  | 0.56   | 0.15  | 0.14   | -0.16* | 0.11   | -0.13 | 0.06    |
| SD FT   | 0.33*   | 0.18  | -0.14 | -0.14  | 0.86*   | 1.00   | 0.89*  | 0.19* | 0.17   | -0.24* | 0.04   | -0.03 | -0.01   |
| CV FT   | 0.32*   | 0.17  | -0.11 | -0.20* | 0.56*   | 0.89*  | 1.00   | 0.17  | 0.13   | -0.23  | -0.01  | 0.02  | -0.06   |
| SK FT   | 0.21*   | 0.11  | -0.07 | -0.10  | 0.15    | 0.19*  | 0.17   | 1.00  | 0.02   | -0.01  | 0.00   | 0.28  | 0.02    |
| WI      | 0.03    | -0.02 | -0.08 | 0.02   | 0.14    | 0.17   | 0.13   | 0.02  | 1.00   | -0.76* | -0.31* | 0.02  | -0.04   |
| TRI     | -0.05   | -0.03 | 0.00  | 0.00   | -0.16   | -0.24* | -0.23* | -0.01 | -0.76* | 1.00   | 0.29*  | -0.08 | -0.07   |
| DFME    | 0.15    | 0.11  | 0.00  | -0.05  | 0.11    | 0.04   | -0.01  | 0.00  | -0.31* | 0.29*  | 1.00   | -0.14 | 0.49*   |
| LAI     | -0.22   | -0.03 | 0.26  | -0.11  | -0.13   | -0.03  | 0.02   | 0.28* | 0.02   | -0.08  | -0.14  | 1.00  | 0.00    |
| Mean Rn | 0.07    | 0.00  | -0.07 | -0.10  | 0.06    | -0.01  | -0.06  | 0.02  | -0.04  | -0.07  | 0.49*  | 0.00  | 1.00    |

*Picea mariana*, the dominant tree species, is relatively intolerant of waterlogging and thus unable to survive in saturated areas, confining them to plateaus (Islam & Macdonald, 2004). While no relationship was found between soil moisture and LAI on plateaus, Lafleur et al. (2005) found that *Sphagnum spp.* can absorb 20 times their dry weight in water and retain capillary water well above the water table. As such, soils underlying mosses are generally wetter than soils underlying lichens. Therefore, a two-tailed unpaired t-test was performed to compare measured mean volumetric soil moisture under moss versus lichen over the study site. Soil moisture was found to be statistically different ( $p < 0.05$ ) under the two vegetation types, where higher mean moisture conditions and standard deviation were similarly found under mosses (Table 2.6). Wright et al. (2009) found similar results at Scotty Creek. It may be the case that examining solely LAI as a measurement of vegetation discounted the role of these species on soil moisture variability. However, lichens generally occur in well-drained soils, whereas mosses are found in wetter areas. Therefore, topographical position may be the underlying control on soil moisture rather than solely vegetation. The lack of relationship between soil moisture and radiation is likely due to the inability of the model to account for the effects of vegetation on radiation reaching the ground surface.

Table 2.6: Comparison of the mean and standard deviation of soil moisture in soils under moss versus lichen.

|             | <b>Moss</b> | <b>Lichen</b> |
|-------------|-------------|---------------|
| <b>Mean</b> | 0.23        | 0.17          |
| <b>SD</b>   | 0.14        | 0.07          |

## 2.5 Conclusion

Overall, the trends of moisture variability observed in this study agreed with previously identified trends in mineral soils suggesting existing scaling properties developed in mineral soils may be applied to peat soils. However, near surface soil moisture variability in organic soils, especially at Scotty Creek, is much larger than that for mineral soils which may limit the transferability of previously defined relationships. Standard deviation increased with mean volumetric soil moisture, while the coefficient of variation and skew decreased. The PDF of soil

moisture was positively skewed on all sampling days. As soil moisture influences runoff, evapotranspiration, and permafrost degradation, it is important to fully characterize moisture relationships in order to understand and predict future change in these regions (Seneviratne et al, 2010; Wright et al, 2009). Peat plateau soil moisture at Scotty Creek was primarily related to frost table depth ( $r = 0.31$ ,  $p < 0.05$ ), where deeper frost table depths corresponded with wetter soils. This relationship is due to the high thermal conductivity of water leading to increased thaw rates, creating a feedback process. Further characterizing this relationship may better capture frost table depth and variance across the landscape for monitoring of changing permafrost and associated processes. In addition to this relationship, unless the water supply is unlimited, deeper thaw will eventually result in lower water tables and thus drier near surface soils. However, deeper frost tables may also accumulate water from upslope areas transforming the plateau into a bog. The potential switch from a positive to negative feedback process between frost table depth and soil moisture should be investigated in future research. No relationship was found between soil moisture and LAI, radiation or plateau micro-topography. Mean soil moisture varied depending on the presence of moss versus lichen, although this may be due to underlying topographical conditions controlling the initial establishment of certain species.

Subarctic peatlands are sensitive to disturbances, and considerable changes to permafrost extent and freshwater discharge have been documented in the last few decades (Connon et al, 2014). High spatial and temporal resolution data sets contribute extensive information on soil – water – atmosphere processes; however, ground based monitoring is rare due to the intensity and extensive resources required. Since it not feasible to acquire high-resolution data sets over large remote regions such as the Subarctic, remote sensing platforms, for example the newly launched Soil Moisture Active Passive (SMAP), are becoming increasingly of interest for use in these environments. The statistical characterization of soil moisture variability allows for better representation of its distribution across a landscape, improving the accuracy of remotely sensed products. Continued monitoring of soil moisture in northern latitudes is vital for the continued study of the feedbacks to permafrost degradation and carbon sequestration in the face of anthropogenic development and climate change.

## **Chapter 3.0: The impact of temperature-induced drought stress on the environmental controls of *Picea mariana* sap velocity in a permafrost peatland**

### **Abstract**

The Canadian Subarctic is presently experiencing one of the highest rates of mean annual air temperature increase; however, this has not been matched by a corresponding increase in precipitation. As such, increased evapotranspiration due to higher temperatures, while water is available, is hypothesized to result in moisture deficits, eventual declines in evapotranspiration, and an overall decrease in forest productivity termed browning. Furthermore permafrost degradation may exacerbate drought by positioning the water table below the rooting zone drying out the upper soil layers, although this decline may be offset by subsequent ground subsidence. This research aims to examine the environmental controls on sap velocity and the influence of temperature-induced drought stress on *Picea mariana* in a Subarctic peatland. Three *Picea mariana* were instrumented with Heat Ratio Method (HRM) sap flow sensors. Measurements of sap velocity were averaged across the 100 m<sup>2</sup> study plot. Soil moisture at 5, 10, and 20 cm depths were recorded at an in-situ station adjacent to the site. Frost table was measured across 24 transects, and water table depth was monitored at a well within the plot. Atmospheric variables were acquired from a nearby meteorological station. Autoregressive integrated moving average (ARIMA) models were developed to predict sap velocity from a variety of environmental controls, and to determine the dominant controlling processes. The relationships between sap velocity, net radiation (Rn), vapour pressure deficit (*D*) and soil moisture were further examined to identify the magnitude and timing of drought stress. Net radiation and vapour pressure deficit were the primary controls on daily sap velocity; however, antecedent soil moisture improved model best-fit statistics. Sap velocity was correlated to Rn, *D*, and soil moisture during the entire study period indicating *Picea mariana* experienced drought-stress, although the exact timing could not be determined. Drought stress may further develop in these environments as temperatures continue to increase, potentially impacting tree growth, evapotranspiration, and the carbon and hydrological cycles.

### 3.1 Introduction

The response of high northern latitude boreal forests to climate change is relatively uncertain (Shaver et al, 2000) and widely debated in the literature. Remotely sensed Normalized Differenced Vegetation Index (NDVI) trends, depicting changes in biomass over time, indicate highly varied responses of the Boreal forest across North America (Verbyla, 2008). Increasing productivity (generation of biomass), termed greening, is primarily limited to Arctic and coastal regions (Verbyla, 2008) that are energy-limited with ample available moisture. In these regions, increases in growing season length are advantageous for forest productivity (Myneni et al, 1997). Decreasing biomass and productivity, termed browning, occurs predominantly within the warmer and drier Subarctic Boreal forest (Verbyla, 2008), where *Picea glauca* (White Spruce), *Picea mariana* (Black Spruce) and *Pinus banksiana* (Jack Pine) show the greatest growth declines (Barber et al, 2000; Lloyd & Bunn 2007; Baltzer et al, 2014). Multiple hypotheses attempt to ascertain the browning mechanism. The most notable of these are temperature-induced drought stress caused by increased evapotranspiration (Barber et al, 2000; Lloyd & Fastie, 2002), and direct temperature stress due to temperatures above the optimal range (approximately 24 °C for *Picea spp.*) (D'Arrigo et al, 2004). Other hypotheses include pollution (Wilson & Elling, 2004), ozone depletion (Briffa et al, 2004), and global dimming (D'Arrigo et al, 2008).

Over the next century, ambient air temperatures are predicted to increase across the Boreal forest, with the highest rates expected to occur in the Subarctic region (Jorgenson & Osterkamp, 2005). The Subarctic region is underlain by discontinuous permafrost, which is defined as ground that has been frozen for over two consecutive years (Muller, 1945). Permafrost, and the active layer, provides support for the ground surface and aboveground biomass, as well as controls available water and nutrients, rooting depth, and soil temperature (Ford & Bedford, 1987; Woo, 1992). Additionally, a local water table is often found perched above the frost table, the impermeable boundary within the thawing active layer (Quinton & Hayashi, 2005), which can provide plant available water during the growing season. The frost table is commonly defined as the zero-degree isotherm within the active layer (Carey & Woo, 2000). In energy-limited systems (as opposed to moisture-limited), evapotranspiration is largely controlled by atmospheric conditions such as radiation and vapour pressure deficit ( $D$ ). As  $D$  increases, a progressively larger moisture gradient is formed between the moist leaf interior and

the relatively dry air, drawing water out of the leaf through transpiration. Plants close their stomata during high  $D$  to avoid internal moisture stress (Lambers et al, 1998). However, as temperatures continue to increase, and thus evapotranspiration, by no account has this been met with a similar rise in total annual rainfall (Quinton et al, 2009; Connon et al, 2014). These then resulting moisture deficits, lead to decreased sap flow and tree growth (Barber et al, 2000). In fact, available moisture may decrease as much as 20% due to temperature-induced increases in evapotranspiration (Schneider & Rosenberg, 1989; Mitchell et al, 1990) causing an agricultural drought. Specifically, agricultural drought refers to instances of insufficient soil moisture to sustain vegetative demands. Already, studies at Wolf Creek in Subarctic Yukon show that evapotranspiration constitutes one of the greatest water losses from organic soils underlain by permafrost (Carey & Woo, 2001). Increasing temperatures, in addition to increases in evapotranspiration, have lead to greater rates of permafrost thaw and subsequent deeper water tables. Subarctic permafrost landscapes are often dominated by *Picea mariana*, which are rooted in the top 20 cm of soil (Lieffers & Rothwell, 1987; Patankar et al, 2015). As such, increasing frost and water table depths may have dramatic impacts on moisture availability in these forests, unless subsequent ground subsidence offsets the lower water table. Furthermore, trees become less sensitive to changes in radiation and  $D$  during moisture stress, and thus transpiration is primarily controlled by available water (Oren et al, 2001; Lopez et al, 2007; Granier & Loustau, 1994).

In the last half century, rising mean annual air temperatures in the Subarctic (Quinton & Hayashi, 2005; Robinson & Moore, 2000) have effected forest structure and function (Baltzer et al, 2014; Prowse et al, 2009), carbon uptake (Robinson & Moore, 2000; Barber et al, 2000; Van Herk et al, 2011), the rate and pattern of permafrost degradation (Baltzer et al, 2013; Vitt et al, 2000; Prowse et al, 2009), and the hydrological cycle (Quinton & Baltzer, 2013; Connon et al, 2014). Therefore the objectives of this research are to 1) examine the environmental controls of sap flow, and 2) identify the impact (timing and magnitude) of a) the declining frost and water table and b) temperature-induced drought stress on sap flow in Subarctic *Picea mariana* forests underlain by permafrost. In order to accomplish these objectives, sap velocity will be modeled as a function of various environmental factors. Then the relationships between sap velocity and key atmospheric parameters and soil moisture will be examined under different moisture conditions.

## 3.2 Methods

### 3.2.1 Study Area

#### 3.2.1.1 Peatlands

Peatlands are wetlands that have accumulated organic material exceeding 40 cm in depth (NWWG, 1988), and in the Subarctic discontinuous permafrost zone, permafrost is found almost exclusively in these peatlands (Quinton et al, 2003). Subarctic peatlands consist of three main landscape features: channel fens, flat bogs and peat plateaus (Zoltai & Tarnocai, 1975). Channel fens and flat bogs are permafrost free features where the water table is at or above the ground surface (Quinton et al, 2003; Zoltai & Tarnocai, 1975; Zoltai & Vitt, 1995). Peat plateaus are forested features that are supported by a permafrost core, elevating them one to two meters above the surrounding wetlands (Robinson & Moore, 2000; Vitt et al, 1994). Dominant plateau vegetation includes: *Picea mariana*, *Cladonia mitis* (Reindeer lichen), *Pleurozium schreberi* (feather moss), *Rhododendron groenlandicum* (bog Labrador tea), *Andromeda glaucophylla* (bog rosemary) (Quinton et al, 2009), and occasionally *Sphagnum spp.* (mosses) (R. Warren, personal obs.).

#### 3.2.1.2 Scotty Creek

Research was conducted at Scotty Creek (61°18'N, 121°18'W), Northwest Territories, Canada, a 152 km<sup>2</sup> peatland-dominated basin within the Lower Liard River valley and Mackenzie River basin. The region has a dry continental climate with short summers and long, cold winters with average annual temperatures just below 0 °C (MSC, 2015). Annual temperatures have risen approximately 2 °C in the last century (Quinton et al, 2009). Average annual precipitation is 369 mm of which 46% is snow (Quinton et al, 2009; MSC, 2015). Plateaus consist of a 15 to 20 cm layer of living and partially decomposed vegetation over 2 to 8 meters of dark, highly decomposed peat, which sits on mineral soils (Aylesworth & Kettles, 2000; Quinton et al, 2009). Scotty Creek has a distinct wet regime immediately after snowmelt when the water table is high and there is minimal storage capacity within active layer. This is then followed by a dry regime as the thaw depth reaches its maximum, and the water and water tables decline (Quinton & Gray, 2003; Quinton et al, 2009).

### 3.2.2 Site Instrumentation

#### 3.2.2.1 Measurement of sap flow

A 100 m<sup>2</sup> plot was established in June 2014 with one edge along a small isolated bog and the rest extending into the adjacent plateau interior. Three *Picea mariana* of similar size (6.1 to 7.4 cm diameter at breast height) were instrumented with two Heat Ratio Method (HRM) sap flow sensors at breast height (approximately 1.3 meters off the ground). These sensors are composed of two temperature probes and a heat pulse needle, which are inserted into the xylem. The heat pulse needle is positioned equidistant between the two temperature probes. The heat pulse needle then emits a pulse at a preconfigured frequency and the ratio of the temperature increase is measured by two thermistors within the temperature probes. Burgess et al. (2001) provide a more detailed explanation. Heat pulse sap velocity (cm hr<sup>-1</sup>) was calculated using the Sap Flow Tool software (ICT International, Armidale, Australia) equation

$$V_s = \frac{k}{0.0025} \times B \times \frac{\rho_b(c_w + m_c + c_s)}{\rho_s \times c_s} \times V_h \quad [ 3.1 ]$$

where  $V_s$  is the sap velocity (cm hr<sup>-1</sup>),  $k$  the thermal diffusivity (cm<sup>2</sup> s<sup>-1</sup>), 0.0025 the reference thermal diffusivity (cm<sup>2</sup> s<sup>-1</sup>),  $B$  the wound correction factor,  $\rho_b$  the density of wood (kg m<sup>-3</sup>),  $c_w$  the specific heat capacity of wood,  $c_s$  the specific heat capacity of sap,  $\rho_s$  the density of water,  $m_c$  the water content of sapwood, and  $V_h$  the measured heat pulse velocity (cm hr<sup>-1</sup>). Only sap velocity measured at the first thermistor (inserted 0.75 cm into the xylem) was used, as the second thermistor (inserted 2.25 cm into the xylem) was located in the heartwood. Sap velocity was used in the analysis as it is easily comparable between different trees and can be related to transpiration.

#### 3.2.2.2 Measurement of atmospheric variables

Net radiation (Rn), air temperature, relative humidity, and precipitation were measured continuously at a monitoring station located in a neighboring bog. Actual and saturated vapour pressure were calculated from air temperature and relative humidity and used to determine the vapour pressure deficit. Potential evapotranspiration was calculated using the Priestley-Taylor equation

$$PET = \alpha \left[ \frac{1}{\lambda} \times \left( \frac{s \times (R_n - G)}{s + \gamma} \right) \right] \quad [ 3.2 ]$$

where PET is the potential evapotranspiration ( $\text{mm day}^{-1}$ ),  $\alpha$  the Priestley-Taylor coefficient,  $\lambda$  the latent heat of vaporization,  $s$  the slope of the saturation vapour pressure-temperature relationship,  $G$  the ground heat flux ( $\text{W m}^{-2}$ ) and  $\gamma$  the psychrometric constant. The Priestley-Taylor coefficient was determined to be 0.69 based on values stated in relevant literature (Gong et al, 2012; Wight et al, 2008) and the total evapotranspiration over a footprint encompassing plateaus, fens and bogs measured at a nearby eddy covariance tower. As the Priestley-Taylor equation does not account for water loss from trees, calculated potential evapotranspiration is likely underestimated for our plateau study site. However, due to the relatively open canopy of plateaus the underestimate may not be as significant as that for higher stand density portions of the Boreal forest (Wright et al, 2008).

### 3.2.2.3 Measurement of ground variables

Frost table depth was measured to the closest centimeter with a graduated steel rod across four 1.5 m transects in 10 cm increments that intersected each instrumented root. Twenty-four transects were established in total. Frost table depth was recorded on 15 days throughout the study period and averaged across the plot. Volumetric water content (VWC) in  $\text{m}^3/\text{m}^3$  was continuously measured using Stevens Hydra Probes (Stevens Water Monitoring Systems Inc, Oregon, USA) at 5, 10 and 20 cm depths at an in situ station adjacent to the study plot. Station measurements were calibrated using a calibration equation previously developed for peat soils at Scotty Creek (Merchant, 2014) using the method of Rowlandson et al. (2013). The height of the water table was measured at a well installed within the study plot using a HOBO U20 Water Level (Onset Computer Corporation, Massachusetts, USA).

### 3.2.3 Statistical Analysis

Statistical analyses were performed using IBM SPSS Statistics v 22 (IBM Corp, 2013) and R v 3.1.2 (R Development Core Team, 2014). Sap velocity at breast height was averaged from the instrumented trees in order to construct a continuous time series for the study period as sensors periodically stopped and started recording. A linear trend was fitted to the frost table

measurements ( $R^2 = 0.99$ ,  $p < 0.001$ ) and interpolated to create a continuous series of frost table depth over the study period. Daily time series were constructed for each of the remaining variables by averaging sub-hourly measurements from June 12 to August 22, 2015.

### *3.2.3.1 Examining environmental controls of sap velocity*

Multivariate time series analysis is useful for identifying patterns and relationships in time series data, specifically; autoregressive (AR) integrated (I) moving average (MA) models, hereby referred to as ARIMA models, were used. ARIMA models were developed in order to account for the serial dependence (i.e. the autocorrelation of consecutive measurements) and trends inherent in the observed variables. They consisted of an autoregressive term ( $p$ ; describes the correlation between consecutive observations), the trend term ( $d$ ; identifies and removes underlying trends to make the series stationary), and a moving average term ( $q$ ; accounts for the persistence of a random shock) (Brockwell & Davis, 2002; Tabachnick & Fidell, 2012). The residuals of an ARIMA model should resemble random error and have no full or partial autocorrelations, or trends remaining in the time series (Tabachnick & Fidell, 2012). With the aim of assessing whether changing moisture conditions affect daily sap velocity, various combinations of the measured environmental processes were added as regressors to the ARIMA model following a similar approach to Ford et al. (2005) and Patankar et al. (2015). Model residuals were checked to ensure they closely resembled random errors. Cross-correlations of sap velocity and  $R_n$ ,  $D$ , soil moisture, and precipitation identified soil moisture as being more highly correlated to sap velocity at negative lags, indicating that soil water availability was most closely associated with the succeeding day's sap velocity. The lagged relationship likely accounts for the time it takes for water to be taken up by the roots and transmitted through the xylem to the HRM sap flow sensors located at breast height. As such, additional ARIMA models were constructed to account for this lagged relationship.

### *3.2.3.2 Identification of climate change induced drought stress in *Picea mariana**

Trees regulate transpiration in relation to available water, which in turn controls sap velocity (Lopez et al, 2007; Granier & Loustau, 1994; Oren et al, 2001). Research by Bovard et al. (2005), Rutter (1968) and Van Herk et al. (2011) have shown that during drought conditions the relationship between sap flow and radiation (Oren et al, 2001), and  $D$  (Lopez et al, 2007;

Granier & Loustau, 1994) weakens. Additionally, studies by Granier & Loustau (1994) and Bauerle et al. (2002) state that soil moisture only controlled sap flow during a drought. Sap velocity was therefore correlated with  $R_n$ ,  $D$  and soil moisture over the study period. These relationships were then also examined for each month. A weak relationship between sap velocity and the atmospheric conditions suggests of the importance of drought stress in this environment. By further examining the changing strength of the relationship each month we are able to identify the exact timing moisture stress manifests during the season.

### 3.3 Results

#### 3.3.1 Environmental controls on sap velocity

The Akaike information criterion (AIC), which is a measure of the relative goodness-of-fit and complexity of a model, and the corrected AIC (AICc), which penalizes models for extra parameters, were calculated and used to identify the best performing ARIMA model. The agreement between the observed and model fitted values was also computed and presented in Table 3.1. Models including  $R_n$  and  $D$  were able to account for the greatest amount of variance, while lagging soil moisture greatly improved AICc values. Net radiation,  $D$  and antecedent soil moisture were the best predictors of sap velocity; however, upon examination of parameter significance, the inclusion of soil moisture at any depth or lag with the exception of lagged 10 cm moisture was not significant ( $p > 0.05$ ). The lagged 10 cm soil moisture coefficient was significant at  $p < 0.001$ . Significance statistics of the individual regressors for the best performing models (in bold) are presented in Table 3.2. Net radiation and  $D$  were always significant predictors. Therefore  $R_n + D +$  lagged 10 cm soil moisture constituted the best fitting model based on the AICc and accounted for 83% of the variance, while the second best model ( $R_n + D$ ) explained 86% of the variance. Including all measured variables in the model did not improve model performance. While soil moisture may partially control sap velocity, its role is minor compared to  $R_n$  and  $D$ . Similarly, Bovard et al. (2005), Ford et al. (2005), and Patankar et al. (2015) all found that sap flow is primarily controlled by atmospheric processes though available water may influence long-term trends.

Table 3.1: Summary of best-fit statistics for the various ARIMA models developed to predict sap velocity.

| <b>Model</b>  | <b>AIC</b>    | <b>AICc</b>   | <b>R<sup>2</sup></b> | <b>p-value</b>   |
|---|---------------|---------------|----------------------|------------------|
| Sap velocity  | 9.44          | 10.04         | 0.48                 | <0.001           |
| 5 cm soil moisture                                  | 8.91          | 9.83          | 0.50                 | <0.001           |
| 10 cm soil moisture                                 | 14.12         | 14.71         | 0.40                 | <0.001           |
| 20 cm soil moisture                                 | 9.65          | 10.58         | 0.49                 | <0.001           |
| Precipitation                                       | -14.39        | -13.47        | 0.64                 | <0.001           |
| dFT   | -0.7          | 0.21          | 0.52                 | <0.001           |
| Rn  | -55.39        | -54.46        | 0.81                 | <0.001           |
| <i>D</i>  | -32.33        | -31.97        | 0.72                 | <0.001           |
| <b>Rn + <i>D</i></b>                                | <b>-73.93</b> | <b>-72.16</b> | <b>0.86</b>          | <b>&lt;0.001</b> |
| Rn + 5 cm soil moisture                             | -54.51        | -53.19        | 0.80                 | <0.001           |
| Rn + 10 cm soil moisture                            | -49.37        | -47.09        | 0.78                 | <0.001           |
| Rn + 20 cm soil moisture                            | -57.85        | -56.07        | 0.82                 | <0.001           |
| Rn + Precipitation                                  | -59.57        | -57.24        | 0.84                 | <0.001           |
| Rn + dFT  | -58.57        | -57.66        | 0.78                 | <0.001           |
| <i>D</i> + 5 cm soil moisture                       | -30.44        | -29.83        | 0.72                 | <0.001           |
| <i>D</i> + 10 cm soil moisture                      | -27.28        | -26.68        | 0.65                 | <0.001           |
| <i>D</i> + 20 cm soil moisture                      | -37.72        | -36.41        | 0.75                 | <0.001           |
| <i>D</i> + Precipitation                            | -39.88        | -39.27        | 0.75                 | <0.001           |
| <i>D</i> + dFT                                      | -39.01        | -38.41        | 0.71                 | <0.001           |
| dFT + Precipitation                                 | -22.18        | -20.89        | 0.65                 | <0.001           |
| <b>Rn + <i>D</i> + 5 cm soil moisture</b>           | <b>-72.9</b>  | <b>-70.58</b> | <b>0.86</b>          | <b>&lt;0.001</b> |
| Rn + <i>D</i> + 10 cm soil moisture                 | -55.09        | -53.8         | 0.78                 | <0.001           |
| Rn + <i>D</i> + 20 cm soil moisture                 | -65.82        | -64.9         | 0.83                 | <0.001           |
| Rn + <i>D</i> + Precipitation                       | -72.17        | -69.85        | 0.86                 | <0.001           |
| Rn + <i>D</i> + dFT                                 | -68.1         | -66.81        | 0.82                 | <0.001           |
| Rn + <i>D</i> + soil moisture + Precipitation + dFT | -69.6         | -65.2         | 0.85                 | <0.001           |
| <b>Rn + <i>D</i> + lagged 5 cm soil moisture</b>    | <b>-77.98</b> | <b>-76.2</b>  | <b>0.86</b>          | <b>&lt;0.001</b> |
| <b>Rn + <i>D</i> + lagged 10 cm soil moisture</b>   | <b>-78.75</b> | <b>-77.84</b> | <b>0.83</b>          | <b>&lt;0.001</b> |
| Rn + <i>D</i> + lagged 20 cm soil moisture          | -71.42        | -70.5         | 0.84                 | <0.001           |

Table 3.2: Significance values for each individual regressor in the best fitting ARIMA models for sap velocity. Soil moisture is only significant at 10 cm when lagged one day.

| <b>Model</b>                               | <b>Regressor Significance</b>                                 |
|--|---|
| Rn + <i>D</i>                              | Rn (<0.001); <i>D</i> (<0.001)                                |
| Rn + <i>D</i> + 5 cm soil moisture         | Rn (<0.001); <i>D</i> (<0.001); soil moisture ( <b>0.33</b> ) |
| Rn + <i>D</i> + lagged 5 cm soil moisture  | Rn (<0.001); <i>D</i> (<0.001); soil moisture ( <b>0.06</b> ) |
| Rn + <i>D</i> + lagged 10 cm soil moisture | Rn (<0.001); <i>D</i> (<0.001); soil moisture (<0.001)        |

### 3.3.2 Sap velocity response under drought stress

Radiation ( $R^2=0.58$ ,  $p<0.01$ ) (Figure 3.1a) and  $D$  ( $R^2=0.60$ ,  $p<0.01$ ) (Figure 3.1b) were strongly correlated to sap velocity during the study period. A power function rather than a linear model better described the relationship between seasonal  $D$  and velocity (Figure 3.1b). Oren et al. (2001) similarly found a non-linear relationship. The strength of the relationships increased each month from June to August (Table 3.3). Bovard et al. (2005) observed a similar relationship between  $D$  and sap flux attributed to available moisture. In wet soils, the  $D$  – sap flux relationship was linear at all values of  $D$ . In dry soils, sap flux linearly increased with  $D$  while  $D$  was less than 1 kPa. When  $D$  was greater than 1 kPa, sap flux was either unresponsive to further increases in  $D$  or declined. Soil moisture was correlated to sap velocity when it was lagged one day (Figure 3.2); however, this may be influenced by sap velocity and moisture following a similar seasonal decline. Examining the monthly relationships revealed weak correlations with 10 cm soil moisture ( $R^2 = 0.22$ ,  $p<0.01$ ) in July, and 20 cm soil moisture in July ( $R^2 = 0.20$ ,  $p<0.05$ ) and August ( $R^2 = 0.18$ ,  $p=0.05$ ) (Table 3.3). Traditional linear regression analysis often breaks the assumptions of independence among samples in time series analysis, and as such it is likely that the reported  $R^2$  statistics are artificially inflated (Shumway & Stoffer, 2006). Therefore the relationships between sap velocity and soil moisture may be less than reported.

Table 3.3: Linear regression statistics between the measured variables and sap velocity at monthly intervals.

| <b>Variable</b>              | <b>Month</b>  | <b>R<sup>2</sup></b> | <b>p-value</b>   |
|------------------------------|---------------|----------------------|------------------|
| Rn                           | <b>June</b>   | <b>0.44</b>          | <b>0.002</b>     |
|                              | <b>July</b>   | <b>0.68</b>          | <b>&lt;0.001</b> |
|                              | <b>August</b> | <b>0.79</b>          | <b>&lt;0.001</b> |
| $D$                          | <b>June</b>   | <b>0.41</b>          | <b>0.003</b>     |
|                              | <b>July</b>   | <b>0.48</b>          | <b>&lt;0.001</b> |
|                              | <b>August</b> | <b>0.58</b>          | <b>&lt;0.001</b> |
| 10 cm soil moisture (lagged) | June          | 0.07                 | 0.29             |
|                              | <b>July</b>   | <b>0.22</b>          | <b>0.007</b>     |
|                              | August        | 0.14                 | 0.09             |
| 20 cm soil moisture (lagged) | June          | 0.04                 | 0.45             |
|                              | <b>July</b>   | <b>0.20</b>          | <b>0.01</b>      |
|                              | <b>August</b> | <b>0.18</b>          | <b>0.05</b>      |

### 3.4 Discussion

#### 3.4.1 Environmental controls of sap velocity

Scotty Creek experienced average daily air temperatures of 18 °C, with a maximum of approximately 30 °C, and 171 mm of rainfall during the 72 day study period (Figure 3.3a). Potential evapotranspiration peaked in late-June to early-July along with Rn and  $D$ , whereas sap velocity appears to have peaked in early-June when or before measurements began (Figure 3.3d). Examination of the residuals showed that the greatest deviations between observed sap velocity and calculated potential evapotranspiration occurred at the beginning of the measurement period. Due to the relatively short growing period and shallow rooting system, *Picea mariana* have been shown in previous studies, to begin transpiring at the onset of soil thaw, which occurs mid- to late-May at Scotty Creek, in order to capture the influx of nutrients and water from snowmelt (Jones, 1999). The discrepancy between observed sap flow and potential evapotranspiration, especially at the beginning of the study period, suggests there may be other processes controlling transpiration rates besides radiation, such as thaw depth and available water. Some of the largest recorded total annual snowfalls between 1898 and 2012 have occurred in the past decade (Connon et al, 2014). While this provides a significant amount of water during the freshet, it is not stored within plateaus due to the thin active layer and high hydraulic conductivity during snowmelt. As well, increasing snowfall would result in deeper end of season frost tables (Wright et al, 2009).

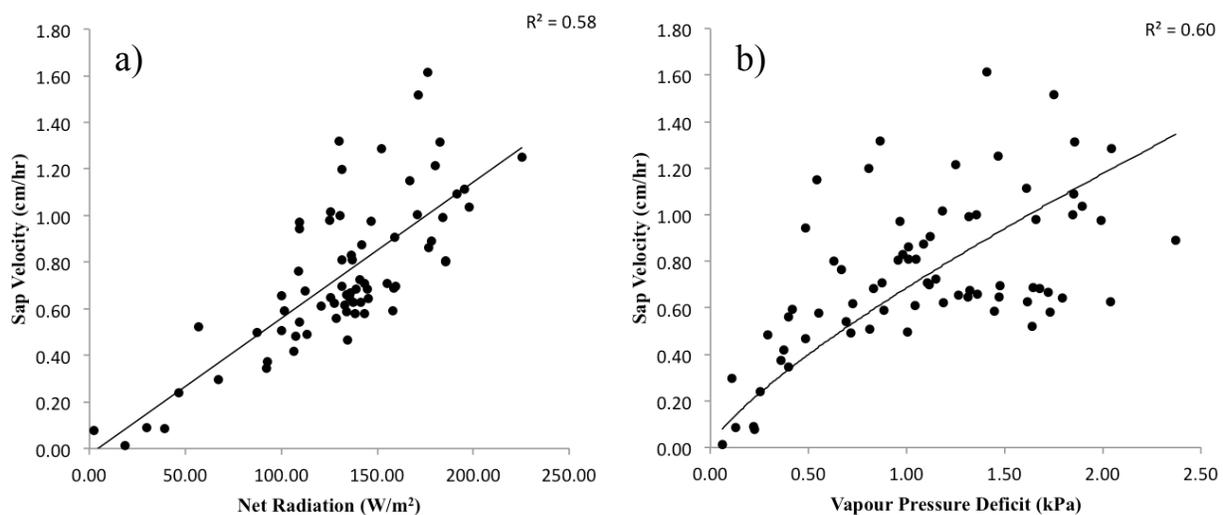


Figure 3.1: The relationships between sap velocity and a) Rn ( $R^2=0.58$ ,  $p<0.01$ ), and b) vapour pressure deficit ( $R^2=0.60$ ,  $p<0.01$ ) over the course of the study period.

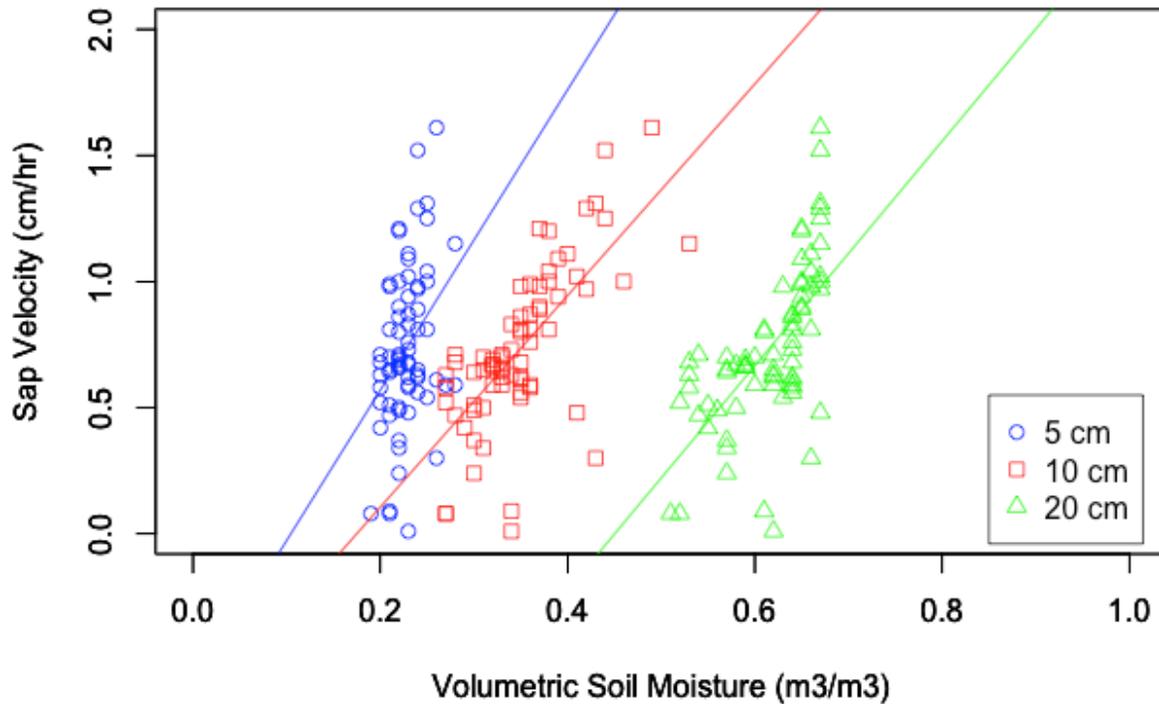


Figure 3.2: The relationships between sap velocity and soil moisture at 5 cm ( $R^2=0.13$ ,  $p<0.05$ ), 10 cm ( $R^2=0.48$ ,  $p<0.05$ ), and 20 cm ( $R^2=0.41$ ,  $p<0.05$ ) over the study period.

Active layer thaw was likely initiated mid- to late-May and reached an average depth of approximately 52 cm (Figure 3.3a), with a maximum depth of 82 cm measured at the edge of the study plot adjacent to the bog, on August 18. While the frost table is near the ground surface at the beginning of the season, the water table, which is perched above it, remains in or near the rooting zone of *Picea mariana* (Figure 3.3b). However, as the frost table declines it positions the water table well below this zone. Precipitation events during the study period caused instances of reduced sap velocity with spikes in soil moisture content and the height of the water table as seen in Figures 3.3a and 3.3b. During precipitation the water table rapidly rises, due to low drainable porosity of peat (Quinton & Gray, 2003), back to the rooting zone. However, during rainfall events, atmospheric conditions do not often facilitate transpiration processes, and as such may inhibit sap flow on these days. While *Picea mariana* is sensitive to root waterlogging, it would only produce a temporary impact on tree functioning, as water is rapidly drained laterally over the frost table into adjacent wetlands due to the high hydraulic conductivity (Quinton & Gray, 2003). Root waterlogging may be a more significant control on sap velocity for trees on the edge

of laterally thawing plateaus (Patankar et al, 2015). It should also be noted that frost table depth is highly spatially variable and may have a local effect on plant water availability.

The ARIMA analysis identified  $R_n$  and  $D$  as the dominant controls of observed daily sap velocity of *Picea mariana* at the study site (see Figure 3.3c for the seasonal pattern of  $R_n$  and  $D$ ), which has similarly been shown in a variety of species and locations across the globe regardless of the presence of permafrost (Bovard et al, 2005; Ford et al, 2005; Granier et al, 1996; Lopez et al, 2007; Oren et al, 2001; Patankar et al, 2015; Van Herk et al, 2011). Soil moisture currently does not appear to exert a dominant control on daily sap velocity, but contributes to the overall seasonal trend as sap velocity and soil moisture both systematically decline over the growing season (Figures 3.3a & 3.3d). As well, the inclusion of 10 cm soil moisture greatly improved the ARIMA predictions of sap velocity. In a study by Small & McConnell (2008), spruce forests in general did not show a distinct association between soil moisture and sap velocity, nor was  $R_n$  or  $D$  able to account for meteorological restrictions on transpiration. In permafrost regions, previous studies such as Baltzer et al. (2013), Iwata et al. (2012) and Patankar et al. (2015) found that while atmospheric conditions controlled daily sap flow, declining moisture due to the declining frost table was responsible for the total amount of observed flow, which decreased throughout the growing season. Scotty Creek experiences a wet period immediately after the initiation of soil thaw followed by a systematic dry-down. A comparison of the controls on sap flow during the wet and dry periods may assist in better understanding the role of soil moisture on sap flow. Converse to previously mentioned research, Sugimoto et al (2002) proposed that thawing permafrost could supply water during drought; however, Lopez et al (2007) found, and other studies have suggested, that thawing permafrost does not offset drought conditions, but in contrast exacerbates them (Patankar et al, 2015; Quinton & Baltzer, 2013). Therefore future vertical permafrost degradation (or active layer thickening) in these environments may induce severe moisture deficits. However, ground subsidence as a result of permafrost thaw may offset drought by keeping the water table near or within the rooting zone.

#### 3.4.2 Influence of drought stress on *Picea mariana*

Various studies, including Bovard et al. (2005), Oren et al. (2001) and Granier & Loustau (1994), show that during drought conditions soil moisture becomes the primary control of sap

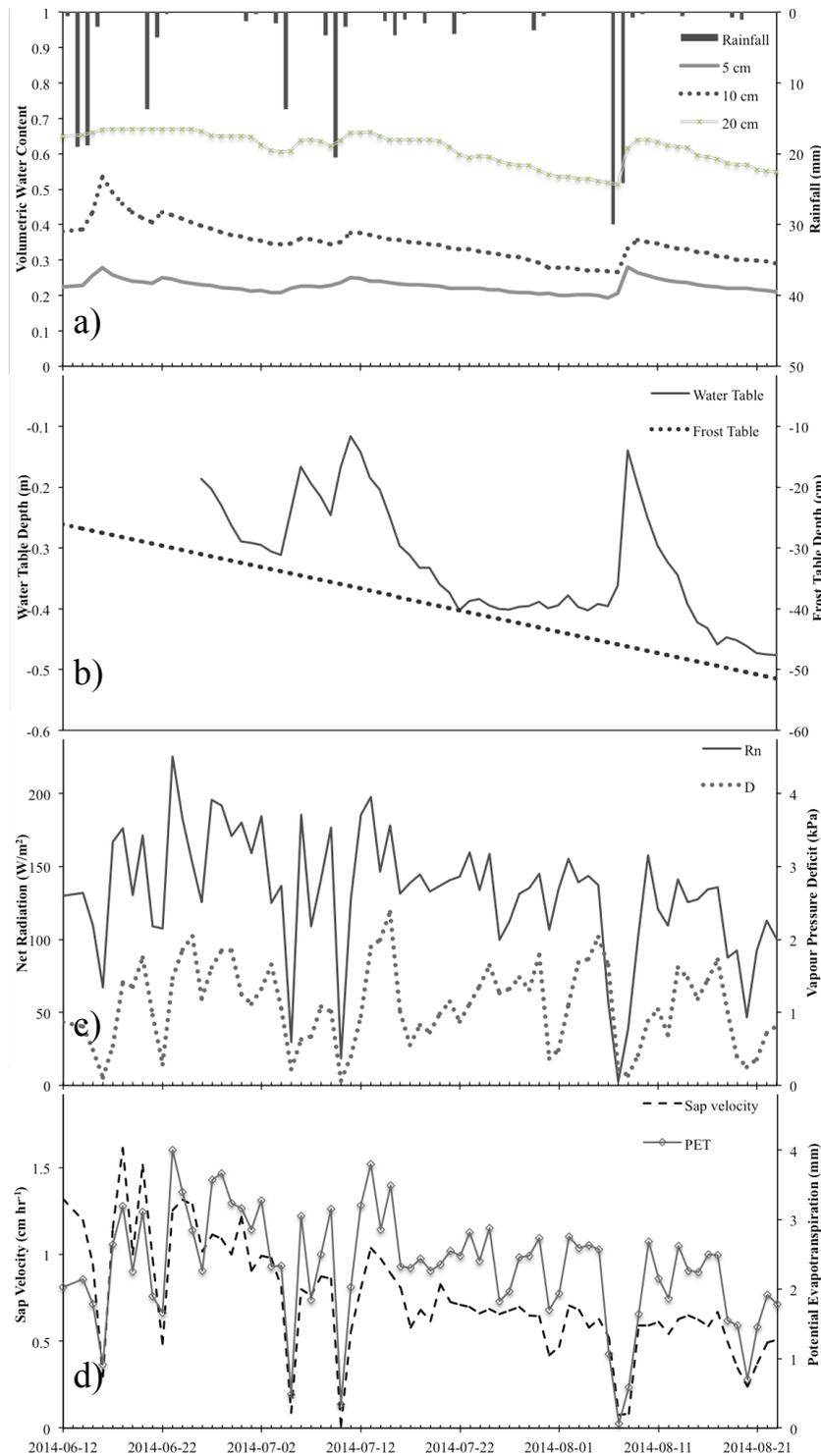


Figure 3.3: Variation in environmental processes over the 72-day study period (June 12 – August 22). a) Soil moisture at 5, 10 and 20 cm depths. Bars are precipitation. b) Interpolated frost table depth and measured water table depth. c) Calculated net radiation ( $R_n$ ) and vapour pressure deficit ( $D$ ). d) Pattern of average plot sap velocity compared with the Priestley-Taylor potential evapotranspiration (PET).

flow, while the influence of  $R_n$  and  $D$  is reduced. Atmospheric conditions at Scotty Creek were correlated with sap velocity throughout the study period (Figure 3.1), and these relationships improved each consecutive month, whereas soil moisture was found to moderately influence sap velocity (Table 3.2). This may be due to the volumetric water content over the growing season never falling below ~20%; however, moisture contents below 20% are not commonly observed at Scotty Creek in non-surface soils (> 5 cm depth). Moreover, shallowly rooted species, such as *Picea mariana*, should be more sensitive to drought conditions (Bovard et al, 2005). The study site experienced a prolonged period of drying over the course of the growing season, which accounts for the stronger relationship between moisture and sap velocity over the entire study period. Overall, the study site did not experience major drought conditions, although the moisture deficits observed did impact sap velocity. As such, drought stress in permafrost regions may manifest over the subsequent years as a continual decline in daily sap flow in *Picea mariana* as the active layer thickens drying out the surface layers. The peak and decline in flow may also start occurring earlier in the growing season. The development of drought conditions in permafrost regions will require the inclusion of soil moisture in predictive models of forest evapotranspiration, including ARIMA modeling.

#### 3.4.3 Influence of climate change on *Picea mariana*

Active layer thickness has dramatically increased in the past few decades (Beilman & Robinson, 2003; Prowse et al, 2009; Quinton et al, 2011), and predictions estimate a complete loss of permafrost in the southern discontinuous zone in the next century, with organic regions being especially sensitive (Prowse et al, 2009). More than 38% of the permafrost at Scotty Creek has been lost since 1947 (Quinton et al, 2011), though predominantly along plateau edges resulting in forest fragmentation and loss (Baltzer et al, 2014; Quinton et al, 2011). Lateral permafrost thaw, as opposed to thickening of the active layer as discussed in this study, also contributes to sap flow declines through the collapse of plateaus into adjacent wetlands and may be a more significant factor in understanding the impact of climate change in permafrost peatlands. Ground subsidence and subsequent waterlogging from lateral thaw has already been shown to decrease sap flow and transpiration rates (Baltzer et al, 2013; Patankar et al, 2015), and radial growth (Baltzer et al, 2014) in individual trees at Scotty Creek. At Scotty Creek, *Picea mariana* may thus be exposed to both drought and waterlogging conditions depending on their

location within the plateau, with interior trees experiencing moisture stress and edge trees experiencing waterlogging. However, while plateau interiors are currently undergoing drought, eventually permafrost degradation and subsequent ground subsidence will transform these features into saturated bogs. Therefore Boreal browning may be the result of these various mechanisms acting to decrease *Picea mariana* productivity and overall forest biomass. Future work should focus on up-scaling our knowledge of the response of individual trees to forest-wide response to changing moisture and permafrost conditions.

### **3.5 Conclusion**

The Subarctic boreal forest is currently experiencing rapid rates of climate change inducing large scale permafrost degradation (Jorgenson & Osterkamp, 2005), as well as declines in tree growth (Barber et al, 2000). Subarctic peatlands are unique as raised permafrost mounds support a *Picea mariana* dominated forest, which are surrounded by wetlands. The shrinkage of plateaus from lateral permafrost thaw leads to an overall decrease in forest cover from plateau collapse and subsequent tree mortality. Increasing active layer thickness may further induce stress by positioning the water table below the rooting zone of *Picea mariana* limiting soil moisture. Sap velocity was seen to peak and quickly decline shortly after the initial thaw and wet period, weeks before the peak in net radiation and potential evapotranspiration. Velocities remained low throughout the latter portion of the growing season when available water was steadily reduced during the dry period. As temperatures continue to increase without a similar increase in precipitation, declining frost and water tables may push these environments into severe drought conditions resulting in widespread decreases in sap flow, and overall productivity causing forest browning. Atmospheric conditions proved to be the primary drivers of sap velocity even in permafrost environments, though soil moisture was shown to be a secondary control. The comparison of sap velocity between the wet and dry regimes may further elucidate the role of soil moisture and the frost table on *Picea mariana* water use. The continued measurement of evapotranspiration, such as from eddy covariance techniques, could provide a long term monitoring of the impact of climate change on forest productivity. Additionally, the characterization of the spatial distribution of soil moisture in Subarctic peatlands will assist in advancing remotely sensed soil moisture products, providing high spatial and temporal resolution data across northern environments. Rooting zone soil moisture can be modeled from near-surface

remotely sensed soil moisture, allowing monitoring and prediction of forest evapotranspiration based on the identified relationships between moisture and sap flow. These efforts will improve our understanding of the response of the Boreal forest to climate change, especially the mechanisms behind the forest browning.

## Chapter 4.0: Summary and Conclusion

Peatlands extend across much of the Subarctic Canada, where climate change has occurred dramatically over the last century. Increasing mean annual temperatures have impacted local and global hydrology, climate, permafrost extent, forest productivity, and carbon cycling, as well as northern infrastructure and livelihoods. While temperatures have risen, a corresponding increase in precipitation will not necessarily occur. Permafrost peatland landscape processes are intricately tied to the distribution of soil moisture; however, the prediction of future conditions and implications is widely debated. The aim of this thesis was to further characterize the distribution and controls of soil moisture in these environments for the monitoring of vegetation and permafrost dynamics. Soil moisture was found to be more highly spatially variable in peat soils, especially if they were underlain by permafrost, than in previously researched agricultural mineral soils. While more variable, the trends of variability over a range of moisture conditions were similar to those observed in mineral soils suggesting similar methods may be used for up-scaling and associated modeling. Standard deviation increased, and coefficient of variation and skew decreased with increasing mean moisture content. At the field scale (Chapter 2.0) soil moisture variability was related to the depth to the frost table, such that mean soil moisture increased with increasing frost table depth. The high thermal conductivity of wet soils resulted in a feedback process where higher moisture increases local thaw depth leading to preferential flow into these depressions, further deepening the frost table. As the frost table is impermeable, a water table may sit above this layer providing plant available water during the growing season. Large topographical differences (i.e. plateaus versus fens or bogs) determined whether an area is saturated or unsaturated; however, micro-topography, LAI and radiation were not significantly related to soil moisture at this scale.

In recent years, NDVI trends have shown areas of browning occurring in Subarctic Boreal forests attributed to temperature induced drought stress. Rising mean annual temperatures also lead to an increase in evapotranspiration; however, without a subsequent increase in available water affected areas will experience moisture deficits and stress. In permafrost regions, changing annual temperature and moisture regimes will also have profound impacts on permafrost subsistence, which will in turn alter current soil – water – vegetation dynamics. For

example, *Picea mariana*, the dominant tree species, is rooted in the top 20 cm of soil and as such increasing frost table depth situates the water table below the rooting zone limiting forest-wide evapotranspiration. In this case, at the plot scale (Chapter 3.0) drier moisture conditions are associated with greater frost table depths over the growing season. Additional research at Scotty Creek has documented radial growth declines in *Picea mariana* over the past few decades. During drought, evapotranspiration is primarily related to soil moisture rather than atmospheric conditions. *Picea mariana* sap velocity during the 2014 growing season at Scotty Creek was partially controlled by soil moisture indicating that this region has begun to experience drought stress. Current atmospheric conditions, and antecedent soil moisture were found to be the primary controls on *Picea mariana* sap velocity. Continuous monitoring of surface soil moisture in this environment will allow better predictions of forest evapotranspiration, and associated carbon processes.

The continued measurement of soil moisture could provide long term monitoring of soil – water – atmosphere processes for examining the effect of climate change on ecosystem functioning. However, ground based techniques do not provide large spatial or temporal measurements due to their intensive nature. Remote sensing is highly advantageous for acquiring these large spatial and temporal data sets, especially in remote regions such as the Canadian Subarctic. However, the statistical characterization of soil moisture variability, a necessary component for the development of soil moisture monitoring networks or interpreting the accuracy soil moisture retrievals, has not been widely studied in this region. Once better understood, soil moisture products developed for this region can then be used for the modeling and forecasting of environmental responses, such as permafrost loss and forest productivity, in the face of climate change. Specifically, accurate soil moisture measurements will allow a better understanding of the physiological response of the Boreal forest to temperature induced drought stress.

In conclusion, soil moisture plays an important role in Subarctic peatlands. *Picea mariana* are experiencing reductions in evapotranspiration as a result of drought stress, which may be modeled using remotely sensed soil moisture products. However, the distribution of soil moisture in peatlands may not necessarily be modeled using previously identified relationships

found for mineral soils due to its higher variability. The presence of permafrost adds a complex feedback process controlling subsurface flow and redistribution, permafrost thaw, and evapotranspiration warranting continued research.

## Chapter 5.0: References

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