MODELLING SOIL TEMPERATURE ON THE BOREAL PLAIN WITH AN EMPHASIS ON THE RAPID COOLING PERIOD

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ABSTRACT

To accurately model soil temperatures on the Boreal Plain, factors that influence fine-grained soils during the rapid cooling period must first be identified. The effects of air temperature, soil moisture and snow depth were quantified at 0.1 and 0.5 m depths for 14 sites encompassing five treatment types: three upland burned, three upland harvested, three upland conifer, three upland deciduous and two wetland. In the absence of snow from September to October at the 0.1 m depth, air temperature was identified as the most important parameter, explaining approximately 70% of the variation in soil temperature for upland and wetland sites. At the same depth in the presence of snow from November to December, soil moisture was more important. At a deeper soil depth (0.5 m), soil moisture was identified as the most important parameter regardless of snow cover, explaining from 63 to 91% of the variation in soil temperatures for upland and wetland sites. The presence of snow was a significant factor influencing soil temperatures, but snow depth was not. Further, the soil temperature algorithms of SWAT were tested using one site of each treatment type at 0.1, 0.5 and 1.0 m depths. The algorithms utilized by SWAT were able to reproduce seasonal trends in soil temperatures adequately for the spring, summer and autumn seasons, with only a slight increase in the lag coefficient parameter. During winter months, the SWAT algorithms tended to predict soil temperatures that were consistently lower than measured data. Further development to the SWAT soil temperature algorithms is required to represent better the important insulating effect of snowpack.

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CHAPTER 1: INTRODUCTION

1.1 RESEARCH OBJECTIVES

When studying the hydrology, nutrient cycles and bioindicators in forested watersheds with cold climates, it is important to understand the annual freeze-thaw (FT) cycles within the soils. However, to model soil temperatures in cold climates, one must first understand which factors influence soil temperatures during the rapid cooling period, when air temperatures fall below freezing for long periods of time.

This research was conducted as part of the Forest Watershed and Riparian Disturbance (FORWARD) project, which is a collaboration of researchers, forestry industries, First Nations communities and both Federal and Provincial government agencies. FORWARD participants have been working in the Swan Hills study area, in northern Alberta on the Boreal Plain for over ten years. The main objectives of FORWARD are to: 1) collect watershed data (vegetation, soils, meteorology, surface water quality and quantity and bioindicators) to adapt an appropriate hydrological model to predict the effects of watershed disturbance (MacDonald et al., 2008; Watson et al., 2008), and 2) apply this knowledge to decision support tools to aid in watershed and forest management planning (Smith et al., 2003; Prepas et al. 2008). FORWARD researchers have been collecting hourly soil temperature data at multiple depths and at 14 sites across the study area, along with climate data, which provides an ideal dataset for an in-depth study on the fine-grained forested soils of the Boreal Plain. Data from selected sites were used to model soil temperatures throughout the year, along with the algorithms from the Soil and Water Assessment Tool (SWAT), which was developed for the Blackland Prairies ecoregion in southeast Texas.

The objectives of this study were to: (1) identify differences and similarities between the agricultural sites where the SWAT model was developed and the study area in the Swan Hills, (2) identify and quantify factors that influence soil temperature during the rapid cooling period, (3) explain differences in soil temperatures among burned, harvested, conifer and deciduous stands, and between upland and wetland sites, and (4) test and modify as needed the soil temperature submodel within SWAT for northern climates.

1.2 Thesis Outline

The following provides an outline of subsequent chapters:

Chapter 2 presents a summary of the literature that identifies similarities and differences between soil characteristics, vegetation biomass and climate in the Swan Hills and southeast Texas where SWAT was developed. This will aid in determining if the soil temperature model will work in cold climates and if modifications to the model will likely be required.

Chapter 3 deals with identifying and quantifying factors which influence soil temperatures during the rapid cooling period from September to December. All 14 study sites were used to determine the effect of air temperature, soil moisture and snow depth. Differences among treatment types (upland burned, upland harvested, upland conifer, upland deciduous, and wetland) were also studied.

Chapter 4 studies the suitability of the SWAT soil temperature algorithms to model soil temperatures on the Boreal Plain. One representative site for each of the five treatments was selected and a sensitivity analysis was conducted. Sensitive parameters were modified to increase the efficiency of the model for Boreal Plain conditions.

Chapter 5 provides a general conclusion of the findings within this work and recommendations for further studies.

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CHAPTER 2: MODELLING THE ANNUAL FREEZE-THAW CYCLE OF BOREAL PLAIN SOILS: A REVIEW

2.1 INTRODUCTION

In forested watersheds where air temperatures fall below freezing for long periods of time, particularly those with deep, fine soils, a thorough knowledge of the annual FT cycle of the soils is a key requirement when modelling hydrology, nutrient cycles and bioindicators. Similarly, this information is essential for planning forestry-related operations (e.g., harvesting, replanting) during FT periods. Although several soil freezing models are readily available (Flerchinger and Hardegree, 2004; Bond-Lamberty et al., 2005; Vattenteknik, 2008), a model has yet to be identified that can accurately model soil temperatures on the Canadian Boreal Plain with minimal input data requirements. The deterministic modelling efforts within the FORWARD project have focused on adapting SWAT for northern climates. SWAT was developed in southeast Texas (Blackland Prairies ecoregion) and is a physically-based model designed for large agricultural watersheds to predict hydrological, sediment and nutrient exports over long periods of time (Neitsch et al, 2005). SWAT can predict soil temperature; however conditions where the model was developed differ significantly from the FORWARD study area, which is located on the Boreal Plain ecoregion in the Swan Hills, approximately 225 km northwest of Edmonton, Alberta (Figure 2-1). A thorough understanding of both agricultural and forested site conditions is required to determine if the SWAT soil temperature submodel is suitable for application in the northern climate of the boreal forest. The following is a compilation of literature that describes the soil characteristics, vegetation biomass and climatic conditions for agricultural and forested sites, as well as provides an overview of other soil temperature models.

Figure 2-1: Swan Hills study area location (star) on the Boreal Plain of Canada (shaded area).

2.2 Soil Characteristics

Many countries have developed their own systems to identify major soil characteristics and classify soils. For example, the United States Soil Classification System classifies soils based on mineral, metal and organic content, moisture, fertility, morphology and texture (Pidwirny, 2006). In the Blackland Prairies of Texas where SWAT was developed, soils are mainly chernozemic (highly fertile organic soils that are sometimes referred to as black soils) and vertisolic (highly expansive dark clays that retain large amounts of water) (World Wildlife Fund, 2001; Texas Parks and Wildlife Department, 2007). The Canadian System of Soil Classification is based on parameters similar to the United States, with the inclusion of permafrost (Pidwirny, 2006). On the Boreal Plain, dominant soils are luvisolic; these are phosphorus-rich, have a high clay content and are derived from eluvial (*in situ* weathering) and illuvial (movement of soil particles to deeper soil horizons by water percolating downward) processes (Government of Canada, 1994).

Soils play an important role in agricultural systems, because they act as the foundation for vegetation growth. Certain key characteristics are required for a soil to be used for agricultural purposes: a good balance of nutrients, moisture and microorganisms, and an absence of unsuitable chemicals (Acton and Gregorich, 1995). Farming practices will affect these characteristics over time, as nutrients and soil moisture are absorbed by crops and percolate out through runoff. Agricultural soils have a tendency to be more homogenous and their characteristics are better known than forest soils, which can vary significantly on a spatial scale. Many other factors that affect soil health occur naturally, however some of these are exaggerated by tillage and other farming practices. These include erosion from wind and water, salinization (Acton and Gregorich, 1995), soil respiration (Han et al., 2007) and soil compaction. Temporal variations (daily, seasonally, or yearly) commonly occur, resulting in additional changes to soils. For example, the quantity of soil erosion is much higher in the initial stages of crop growth and after cultivation, when there is less vegetation to protect the bare soil (Kilmer, 1982). Seasonal variation in soil respiration follows the same pattern as fluctuation in soil temperature, indicating that soil temperature is a controlling factor (Valentini et al., 2000; Han et al., 2007). The soil temperature-respiration relationship is also influenced by biotic factors during the growing season: when soil temperatures are high, crop roots and microbes also contribute to the overall soil respiration process (Han et al., 2007). Forest soils are also affected by these factors. However, management practices take place over much longer periods of time. For a forested site on the Boreal Plain, the crop rotation age is a minimum of 55 years, depending on the region and stand type (Millar Western Forest Products Ltd., 2007). By comparison, cultivation occurs one to several times a year on agricultural sites, depending on the crop and weather patterns. Consequently, conventional farming practices tend to negatively affect soil health more than harvesting of forested sites on the Boreal Plain.

Differences exist between agricultural soils and mineral boreal forest soils, based upon morphology of the sites, texture, organic matter content, bulk density and water-retention capability (Wall and Heiskanen, 2003). For instance, agricultural soils in Finland tend to have higher organic matter content, be more fine-textured, be better sorted and have lower air-filled porosity than forest soils (Wall and Heiskanen, 2003). Low air-filled porosity in agricultural soils with high clay content causes drainage problems and reduces crop yield (Schlenker et al., 2007). Low air-filled porosity also has a detrimental effect on tree growth because there is not enough oxygen available for uptake by vegetation, although with proper drainage systems in place, this problem can be overcome. Both sufficient aeration and moisture content are important for agricultural, as well as forest soils, to sustain growth.

When studying soil temperatures, certain soil characteristics are important. For example, soil density (weight over volume, *W/V*) and porosity (volume of air and water over total volume, $(V_a + V_w)/V$) affect the damping (wetting) depth of a soil (Das, 2004) (Figure 2-2). Although freezing enlarges pore sizes to some extent, thawing causes soils to shrink. The shrinkage rate is higher for fine-grained soils than for coarse-grained soils (Aluko and Koolen, 2001). This shrinking effect could reduce infiltration of water in fine-grained soils, decrease soil temperature, and reduce porosity. In forests, the air-filled porosity needed to support tree growth must be at least 20% (Warkentin, 1984). Harvested sites during snowmelt tend to have slightly higher moisture content while soil temperature has a tendency to be higher when compared to treed sites (Whitson et al., 2005). Forest removal is also known to increase soil moisture by reducing the rate of evapotranspiration (Bosch and Hewlett, 1982).

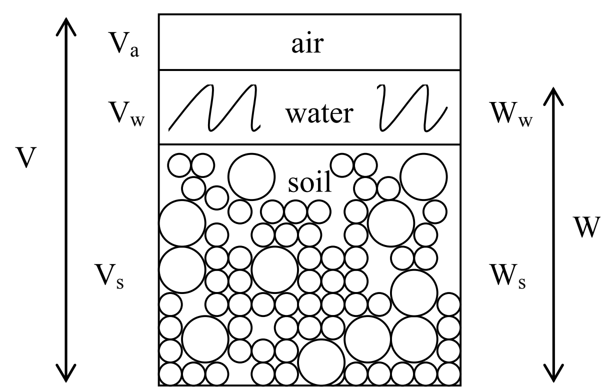

Figure 2-2: Soil phase diagram (V = volume, W = weight, a = air, w = water, s = soil). Modified from Das (2004).

2.3 VEGETATION BIOMASS

On agricultural fields, vegetation biomass refers to the crop, but in forests this term could refer to trees, shrubs, lower vegetation, litter, as well as downed woody debris (Figure 2-3). Agricultural sites are generally monocultures; therefore the amount of vegetation biomass is similar from one location to the next for a given crop. However in forested sites, the diversity of trees, shrubs and lower vegetation communities is much higher, which leads to a more complex environment across both the horizontal and vertical landscape. The understory in forests, which includes lower vegetation and litter, insulates soils since it reduces the amount of light that reaches the soil surface. Its proximity to the ground also traps long wave radiation being emitted at ground level. Stand height, structure and leaf-area index (LAI) also affect soil temperature regimes (Bond-Lamberty et al., 2005). Generally, vegetation biomass is most abundant and provides the most insulation in summer months on shallow south- or west-facing slopes.

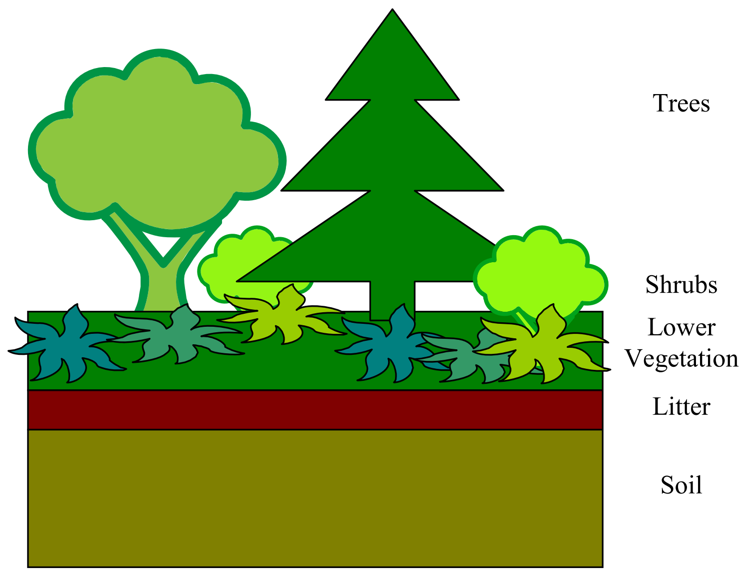

Figure 2-3: Vegetation biomass in a mixed-forest setting.

The amount of vegetation biomass varies temporally and spatially on the Boreal Plain. For example, canopy cover increases for most plant species during the spring months (April to May), peaks in the summer (June to August) and has a decreasing transition period in the autumn (September to October) (Barr et al., 2004). During the winter months (November to March), the canopy cover is appreciably lower after leaf off, an effect that is more pronounced in deciduous-dominated sites. Lower vegetation also changes with the seasons, with an increase in biomass in summer months, a decline during the autumn and relatively low biomass in winter (Barr et al., 2004). Other factors affecting differences in vegetation biomass from one site to another include slope, aspect, type of disturbance (e.g., wildfire, forest harvest, disease and insect outbreaks), time since disturbance and type of treatment applied after disturbance (Fowler and Helvey, 1981). Wildfire is the most important landscape-scale disturbance on the Boreal Plain. For example, in the province of Alberta between 2003 and 2007, wildfires burned an average of nearly 119 000 ha of land each year (Government of Alberta, 2008*a*). The wood products industry is also important. Of the 38 million ha of forested land in Alberta, more than 21 million ha are leased to forestry companies under the terms of Forest Management Agreements (Government of Alberta, 2008*b*). In the study area, the mean annual vegetation biomass for three plots during the first three years after harvest ranged from 338 to 1343 kg ha^{-1} (MacDonald et al., 2007), whereas for a mature forest stand of the same study area, the biomass was 30 to 800 times higher (MacDonald et al., 2007). These considerable ranges are a good example of the spatial variation within a region.

2.4 CLIMATE

Climate and weather each play a role in defining the environment of a site, because they affect the soils, vegetation biomass, hydrology and wildlife. Climate can be defined as weather averages and variations over a long period of time, whereas weather can vary on a daily basis. For example, long-term climate normals (1971-2000) for Whitecourt, Alberta, indicate that it has a cool-temperate climate with an annual average daily temperature of 2.6^{o}C and 201 days of below 0^{o}C temperature (Environment Canada, 2008). The average annual precipitation in Whitecourt is 578 mm, 24 percent of which falls as snow (Environment Canada, 2008). By comparison, the region for which the SWAT model was developed in southeast Texas has an average annual temperature of 18.8^{o}C and only 31 days of below 0^{o}C temperatures throughout the year (based on 1959-2004 climate normals) (National Weather Service Forecast Office, 2007). The average annual precipitation in Texas is 910 mm (1971-2000 climate normals) (Southern Regional Climate Center, 2008).

Dynamics involved in FT cycles affect thermal and geotechnical properties of soils. When soils freeze in the autumn and thaw in the spring, an active layer forms (the depth at which soil temperatures fluctuate above and below freezing (Andersland and Ladanyi, 1994; Smith, 1996)) (Figure 2-4). The depth of this layer varies according to its location and can vary from centimetres to metres in both warm and cold regions (Zhang et al., 1997). Multiple FT cycles within the active layer will increase the hydraulic permeability, diffusivity and void ratio (volume of voids divided by the volume of solids) for dense soils, and decrease the resilient modulus (a measure of elasticity), undrained shear strength and thermal conductivity (Efimov et al., 1980; Qi et al., 2006).

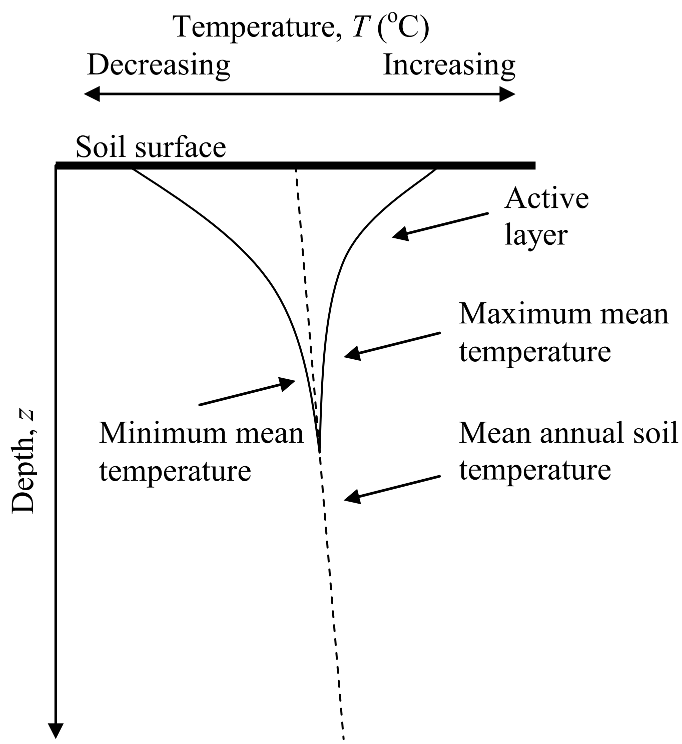

Figure 2-4: Ground temperature regimes for cold regions. Modified from Smith (1996).

Climate and seasonal temperature variations play an important role in the management of agricultural crops, because they affect hydrology and nutrient availability, which have a direct influence on crop yield. To maximize farm income, agricultural sites are managed to produce the highest crop yield; this includes selecting the best suited crop for the site characteristics. In New Zealand, climate and soils data were used to design maps to aid in the selection of appropriate locations for specific crops, to facilitate agricultural decision making for various climate change scenarios (Wratt et al., 2006). Climate trends in Alberta from 1950 to 1998 show an increase in air temperature, a decreased in precipitation (as opposed to the 5 to 35% increase for the rest of Canada) and a decrease in winter and spring snowfall (Zhang et al., 2000). These trends result in less spring and early-summer runoff coming from snowmelt available for the growing season.

Opportunities may arise from a warmer climate such as longer growing seasons and the possibility for diversified crops, however there are many risks associated with a warmer climate. These include more expensive irrigation and increased incidence of fires and pest infestations (Government of Alberta, 2007).

Climate variations are also important factors for the management of forested regions. If climate warming leads to shorter and warmer winters, the incidence of pest infestations may increase (Government of Alberta, 2007). For example, the elevational and latitudinal range of the mountain pine beetle (*Dendroctonus ponderosae* (Hopkins)) in western Canada is limited by winter air temperatures below -40°C (Safranyik and Linton, 1998). In Alberta, it has been estimated that 97.5% of the beetle population must die from cold weather each year to maintain the population at current levels, and this level of mortality cannot be achieved with warmer winters (Government of Alberta, 2009). Climate change can also increase the frequency, length of season, size, intensity (measure of vegetation mortality) and severity (measure of heat transfer to soils) of wildfires (Li et al., 2000; Westerling et al. 2006; Government of Alberta, 2007). Warming temperatures, increased risk of drought frequency and early spring snowmelt were strongly associated with an increased risk of wildfires across the western United States (Westerling et al., 2006). In terms of management of forests for timber supply, climate not only interacts with these natural disturbances, but it influences the ability to mitigate the effects of forest operations. In the forestry industry, compaction due to machinery operating on deep fine-grained soils, such as the Luvisols that dominate Boreal Plain uplands, has been a concern for half a century in North America (McNabb et al., 2001). Therefore, harvesting is often conducted during winter months on frozen ground, and thus climate warming would shorten the window of opportunity for heavy machinery to access the fibre supply. A warming climate could not only promote pest infestations and wildfire, which would limit the amount of wood available for harvest, but it could shift forests (that are considered to be carbon sinks) toward acting as sources of carbon dioxide to the atmosphere.

Snow cover has a critical role to play in cold regions because it creates an insulating layer over the soil, but snow cover patterns are projected to be altered by climate change. As discussed earlier, vegetation biomass provides insulation for the soil but, in the winter months, the insulating effect of snow is greater than that of vegetation biomass. Snow is a good insulator because of its low thermal conductivity and because of its ability to reflect light (Zhang et al., 1997). Further, the insulation from a snowpack is nonlinear and a threshold snow depth exists in which any additional snow does not provide an increased insulation effect (Henry, 2008). A warmer climate is widely believed to decrease the depth of snowpack and decrease the snow-covered period and in turn cause average soil temperatures to decrease during winter months (Zhang et al., 1997; Hardy et al., 2001; Whitson et al., 2004; Mellander et al., 2006; Mellander et al., 2007; Henry, 2008). A survey of weather station data from 31 sites across Canada (3 of which are within 330 km of the study area) suggests that warmer winters will increase the number of FT cycles over a period of one year (Henry, 2008). A two-year experimental study in New Hampshire, which involved removing accumulated snow in early winter months to simulate the delayed onset of snowfall in winter projected to occur with climate change, showed a strong relationship between snow depth and frost depth (Hardy et al., 2001). For all measured depths, treated sites were substantially colder than the reference sites, supporting the hypothesis that snow could maintain soil temperature above the freezing point (Hardy et al., 2001) (Figure 2-5).

The environment can also modify the snowpack via mechanisms associated with albedo, a unitless measure of a surface's reflectivity that ranges from 0 to 1, as well as via physical mechanisms. For fresh snow, albedo ranges between 0.8 and 0.9 (Gray and Landine, 1987; Pohl and Marsh, 2006), indicating high reflectivity. However, impurities in snow like dust and forest litter can decrease snow albedo (Melloh et al., 2001). Thin and discontinuous forest litter on and within snow can reduce albedo to as low as 0.3 for aged snow (Barry et al., 1990). This reduction for aged snow relative to fresh snow is quite substantial, since the albedo of a ground surface was reported to be 0.17 by Gray and Landine (1987) and 0.27 by Flerchinger et al. (1996). Since spectral irradiance of littered snow is reduced when compared to that of clean snow, littered snow melts more rapidly (Melloh et al., 2001). The percentage of forest cover also affects albedo values and for fresh and non-littered snow, the albedo in an open area is higher than in forests (Melloh et al., 2002). Forest cover and slope direction affect snowpack temperature (Wu and Johnston, 2007). Snow depth also increases with increasing openness of a stand and with decreasing LAI and tree height (Mellander et al., 2007) (Figure 2-5). The snowpack can be as much as 37% deeper in open sites than forested sites, because tree branches and leaves intercept snowfall, and either retain snow or provide a site for sublimation of snow (Troendle and Reuss, 1997; Pomeroy et al., 2002). This is especially true for conifers (Parviainen and Pomeroy, 2000).

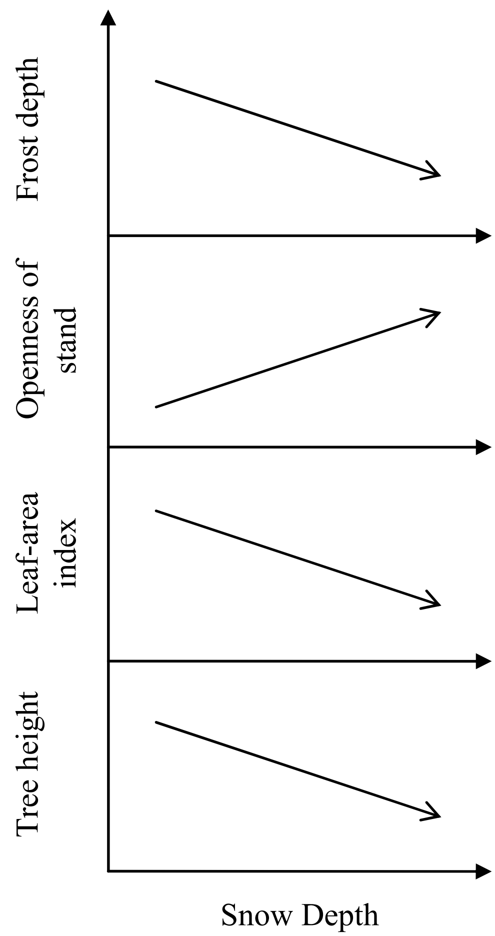

Figure 2-5: Schematic demonstrating the general relationship between various factors and snow depth as described in Hardy et al. (2001) and Mellander et al. (2007).

When considering snowmelt, the albedo of the snow is a key factor, but albedo is difficult to predict because its value changes over time. Snow albedo decreases as the snow pack ages due to changes in wetness, particle size, surface roughness and direction of radiation, but when new snow falls, the albedo returns to the 0.8 to 0.9 range. An albedo decay function, based upon measurements and point estimates of incoming and reflected shortwave radiation, provided reasonably good agreement between predicted and observed data for pre-melt, melt and post-melt periods and for deep (> 20 cm) and shallow (≤ 20 cm) snow cover in the Canadian prairies (Gray and Landine, 1987). This model was also used to simulate the variability of snowmelt over an arctic catchment. It provided accurate results, except in shrub-dominated tundra, which was attributed to the high spatial variability and different snow accumulation patterns in this region (Pohl and Marsh, 2006). The simultaneous heat and water (SHAW) model was tested against data from Minnesota. It consistently overestimated snow albedo, but predicted the surface temperature of the soil well (Flerchinger et al., 1996). Although many attempts have been made to model the variation in snow albedo over time, none have yet found an efficient solution.

2.5 MODELLING THE FREEZE-THAW PROCESS

When modelling soil temperature, air temperature is a key input parameter. The two main approaches for quantifying the influence of air temperature on soil temperature are termed previous-day and degree-day. The previous-day approach uses the temperature from the previous day or a weighted value of the previous few days (Neitsch et al., 2005). In some cases, a lagging factor between the previous and current days' temperatures is used. The degree-day approach is a measure of heating or cooling (Schlenker et al., 2007). A base temperature is selected and each degree above or below that base temperature accounts for one positive or negative unit, respectively. These units are summed for a given period of time to calculate the total heating or cooling degree-days.

Several models have been developed for cold regions to predict the freezing and thawing of soils. In the past, researchers have generally opted to develop simple models that focus on a few specific components of the overall system, rather than developing complex models which require definition of more data inputs and parameters. A recent approach to modelling the freezing and thawing of soils is to capture the spatial and temporal variations within a specified geographical area, which should improve model precision compared to conventional models (Sadler et al., 2007). However, this approach is more involved and requires larger data sets. In addition, many models focus mainly on frost heaving, to support efforts to protect pipelines, buildings, roadways and other infrastructure (Jumikis, 1982; Thomas and Tart, 1984; Kudryavtsev, 2004; Michalowski and Zhu, 2006). Many of these models utilize the Stefan equation, which is a simple method to estimate the depth of frost in soils (Jumikis, 1982; Smith, 1996; Neitsch et al., 2005; Hayashi et al., 2007). The Stefan equation was derived from the Neumann equation, which predicts the movement of the thawing interface based on a large range of parameters. The simplified Stefan equation (Eq. 2-1) estimates the thaw depth, z (m), as a function of unfrozen thermal conductivity, k_u ($W \cdot m^{-1} {}^oC^{-1}$), air thawing (or freezing) index, *TI* or *FI* ($^oC \cdot day$), a surface correction factor for the thawing index, n and specific volumetric latent heat of fusion, L ($kJ \cdot m^{-3}$) (Smith, 1996).

Eq. 2-1 $$z = 13.15\left(\frac{k_u \cdot n \cdot TI}{L}\right)^{0.5}$$

The modified Berggren formula also calculates frost depth in multi-layered soil systems by applying a correction factor to the Stefan equation that accounts for the initial uniform soil temperature, soil freezing temperature and soil volumetric heat capacity (Andersland and Ladanyi, 1994). Using the ground surface temperature and the bulk thermal conductivity (estimated from soil moisture content), Hayashi et al. (2007) developed a simple method, similar to the Stefan equation, which estimates the depth of the frost table. They found that the primary factor responsible for the difference in thawing rates between sites was bulk thermal conductivity, which indicates that thawing rates are greatly dependent on soil moisture. Other approaches include a numerical simulation model to predict infiltration of water into frozen soils (Zhao and Gray, 1997) and a finite difference model to verify the effect of air temperature and snow cover on the active layer and permafrost (Zhang et al., 1997). In Manitoba, a simple model was developed that predicted soil temperature of a boreal forest based on past air temperature and LAI (Bond-Lamberty et al., 2005). This model provided reasonably good results when compared to observed data for the study sites, however it was not expected to be widely applicable for other regions. Predictions made by complex models are often only slightly more accurate than the predictions made by simple models. In addition, complex models require additional data that are not always readily available, thereby making them less practical for use in industry.

Other comprehensive models include: the coupled heat and mass transfer (COUP), SHAW (described above) and SWAT. COUP utilizes many input parameters; these include soil (water retention curve, hydraulic and thermal conductivity and heat capacity), vegetation (vertical root distribution, surface resistance, water uptake, transpiration and albedo) and meteorological data (precipitation, air temperature, air humidity, wind speed and cloudiness). COUP outputs include snow depth, soil temperature, soil moisture, ice content, heat and water flow, heat storage, snow water equivalents, surface runoff, drainage flow and deep percolation to ground water (Mellander et al., 2006; Vattenteknik, 2008). Water uptake of various stands in boreal forests in northern Sweden were modelled using COUP and after a colder than average winter, soils thawed later in the spring and water uptake of trees was reduced (Mellander et al., 2006). Input parameters to SHAW are similar to those of COUP and include initial conditions (snow depth, snow density, soil temperature and water content), soil properties (bulk density and saturated conductivity), vegetation (biomass residue, LAI, plant height, rooting depth and plant albedo), meteorological data (air temperature, wind speed, humidity, precipitation and solar radiation) and other site characteristics (slope, aspect and surface roughness). SHAW outputs include surface energy flux, water balance, snow depth and soil temperature, water, ice and solute content profiles (Flerchinger et al., 1996). The SHAW model was used to predict soil temperature and moisture at shallow depths ($\leq$ 50 cm) in an agricultural field in southwest Idaho, where it provided accurate results for soil temperature, but showed some limitations for modelling soil moisture (Flerchinger and Hardegree, 2004). The SWAT model is described in detail in the next section.

2.6 THE SOIL AND WATER ASSESSMENT TOOL

The SWAT model was selected by FORWARD researchers because it is easy to use, it uses readily-available data as input, it has a GIS interface that facilitates integration of data for large-scale applications and it provides graphical input and output (Putz et al., 2003). SWAT was developed in the early 1990's by agricultural engineers and agronomists in the United States Department of Agriculture (USDA) (Arnold et al. 1998) and has undergone many changes since its conception (Comis, 2002; Gassman et al., 2007). This model incorporates many components of a watershed, including hydrology, soils, vegetation, pollutants and land use. SWAT was developed to assess the impact of management (i.e. agricultural practices) and climate on water quality and quantity for various sized watersheds (Arnold et al., 1998; Arnold and Fohrer, 2005).

The soil temperature submodel in SWAT calculates soil temperatures at a specific depth based on the current day's air temperature and the previous day's soil temperature (Neitsch et al., 2005). It utilizes multiple input parameters that can be roughly separated into three categories: soil characteristics (soil bulk density, moisture content, depth of the soil layer and depth from the ground surface to the bottom of the soil profile), cover (above ground vegetation biomass and residue and snow) and climate (daily mean, maximum and minimum air temperatures, solar radiation and plant and snow albedo). To calculate soil temperatures on the current day $T_{soil}(z,d_n-1)$ (°C), SWAT utilises a weighted temperature based on soil temperature on the previous day $T_{soil}(z,d_n)$ (°C) and soil surface temperature on the current day T_{ssurf} (°C) at a specified depth z (mm) and day d_n, a lag coefficient which controls the influence of the previous day's temperature λ, the average annual air temperature T_{Aair} (°C), and a depth factor that accounts for the influence of depth below ground surface of the soil layer df (Eq. 2-2) (Neitsch et al., 2005).

Eq. 2-2 $$T_{soil}(z,d_n) = \lambda \cdot T_{soil}(z,d_n-1) + (1-\lambda)\cdot \left[df \cdot (T_{AAir} - T_{ssurf}) + T_{ssurf}\right]$$

Vegetation biomass, snow cover and albedo vary, although most of the input parameters are fixed or measured values. The insulation factor bcv, which is calculated from vegetation biomass CV (kg·ha^{-1}) or snow SNO (mm) is accounted for by a weighting factor (0 for bare soil and approaches 1 as cover increases) (Eq. 2-3).

Eq. 2-3 $$bcv = \max\left\{\frac{CV}{CV + \exp(7.563 - 1.297\cdot 10^{-4}\cdot CV)}, \frac{SNO}{SNO + \exp(6.055 - 0.3002\cdot SNO)}\right\}$$

The model selects the highest insulation value from vegetation biomass and snow to account for maximum possible insulation. This insulation factor is used, along with mean, maximum and minimum air temperature, to calculate T_{ssurf} (from Eq. 2-2). The default lag coefficient used by the model is 0.8.

For snow, the routines in SWAT rely on snow already present on the ground: SNO (mm), precipitation R_{day} (mm), sublimation E_{sub} (mm) and snowmelt SNO_{mlt} (mm) (Eq. 2-4).

Eq. 2-4 $$SNO = SNO + R_{day} - E_{sub} - SNO_{mlt}$$

The snowmelt is controlled by a melt factor b_{mlt} (mm·day^{-1}·^{o}C^{-1}), which varies according to the time of year, the fraction of the site covered by snow sno_{cov}, the snowpack and maximum air temperatures, T_{snow} and T_{mx} (^{o}C), respectively, and the base temperature above which snowmelt is allowed, T_{mlt} (^{o}C) (Eq. 2-5) (Neitsch et al., 2005).

Eq. 2-5 $$SNO_{mlt} = b_{mlt} \cdot sno_{\mathrm{cov}} \cdot \left(\frac{T_{snow} + T_{mx}}{2} - T_{mlt} \right)$$

SWAT has been applied to various watersheds in the United States (Gassman et al., 2007). For example, SWAT was used to demonstrate that changes in water balance components (evapotranspiration, soil water storage and water yield) associated with variability in soils and climate of six different watersheds in Texas were more pronounced in wet climates with heterogeneous soils (Muttiah and Wurbs, 2002) (Figure 2-6; Table 2-1). In addition, runoff increased with homogeneity of clay and clay loam soils (Muttiah and Wurbs, 2002). The SWAT model has also been applied successfully to numerous other agricultural-dominated watersheds in Australia, Canada, China, India, New Zealand, and in many African and European countries (Gassman et al., 2007).

Although the SWAT model was developed for croplands, it has shown great potential for application in other areas. In most cases however, modifications have proven to be necessary (Table 2-1). For example, Fontaine et al. (2002) determined that SWAT was unable to properly simulate the snowmelt process in alpine watersheds in Wyoming, which are located approximately halfway between Alberta and Texas (Figure 2-6; Table 2-1). Modification of the snowfall and snowmelt routines increased the versatility of the model. Similarly, application of the model to both forested- and wetland-dominated watersheds in northern Michigan required alteration of snowmelt parameters (Wu and Johnston 2007) (Figure 2-6; Table 2-1). Modifications to the model's snowmelt algorithm were also important when modelling snowmelt and baseflow within a watershed in Pennsylvania containing 75% fragipan soils, which restrict water percolation (Peterson and Hamlett, 1998) (Figure 2-6; Table 2-1). For a watershed in New

York with soils similar to those in Pennsylvania, SWAT was unable to model lateral flow when soils were frozen (Tolson and Shoemaker, 2007). Instead, water remained in the soil profile until saturation, when soils were thawed water percolated to the next soil layer. As a result, water did not reach the channel until after the soil thawed, which resulted in large predicted peak flow values in the spring that were not observed at the streamflow gauging station. The model was modified by forcing it to predict lateral flow and percolation when the soil was frozen, just as it would in summer months (Tolson and Shoemaker, 2007) (Figure 2-6; Table 2-1). Finally, in a Finnish watershed, Francos et al. (2001) replaced the snowmelt routine by a snowmelt submodel and included a new weather generator, which had previously been calibrated for their watershed to better represent local conditions. The new snowmelt submodel used a degree-day method to calculate snowmelt, capable of being set to different values based on land cover (Francos et al., 2001) (Table 2-1). Therefore, for successful application of the SWAT model in cold regions, modifications were required for snowmelt and subsurface hydrology routines.

Since SWAT was developed in part based on the Erosion/Productivity Impact Calculator (EPIC) (Williams et al., 1984), modifications made to the latter are also important. When comparing the soil temperature algorithms of SWAT and EPIC, the equations are similar, with the exception of the inclusion of a 5-day moving average to simulate heat stored in the soil and the lag coefficient in EPIC being set to a value of 0.5 (compared to 0.8 in SWAT) (Potter and Williams, 1994). Testing of the soil temperature algorithms of the EPIC model with data from Iowa, North Dakota and Texas showed that the model did not accurately simulate daily temperature variability (Potter and Williams, 1994) (Figure 2-6; Table 2-1). Two modifications were required to reduce the damping effect. First, the 5-day moving average was reduced to 2 days to maintain a small heat storage effect for cloud-covered days and second, *bcv* (insulation parameter, see Eq. 2-3) was maintained below 0.19. Both changes provided more accurate results. However, it should be noted that no changes were made with respect to the effect of snow on *bcv* (Potter and Williams, 1994). Application of the EPIC model to two agricultural sites in Eastern Ontario demonstrated that the model was unable to accurately model the soil temperature (Roloff et al., 1998) (Figure 2-6). The soil temperature algorithm was also tested in the Peace River region of Alberta, Canada, where EPIC reproduced the increasing soil temperature trends well, but was much weaker at predicting decreasing soil temperature trends (Puurveen et al., 1997) (Figure 2-6;

Table 2-1). A major limitation with SWAT (and EPIC) for cold regions is that it does not model snow processes accurately. This, in turn, affects the soil temperature predictions. To accurately model the FT cycles of cold regions with fine-grained soils, it is expected that further modifications need to be incorporated into the SWAT model.

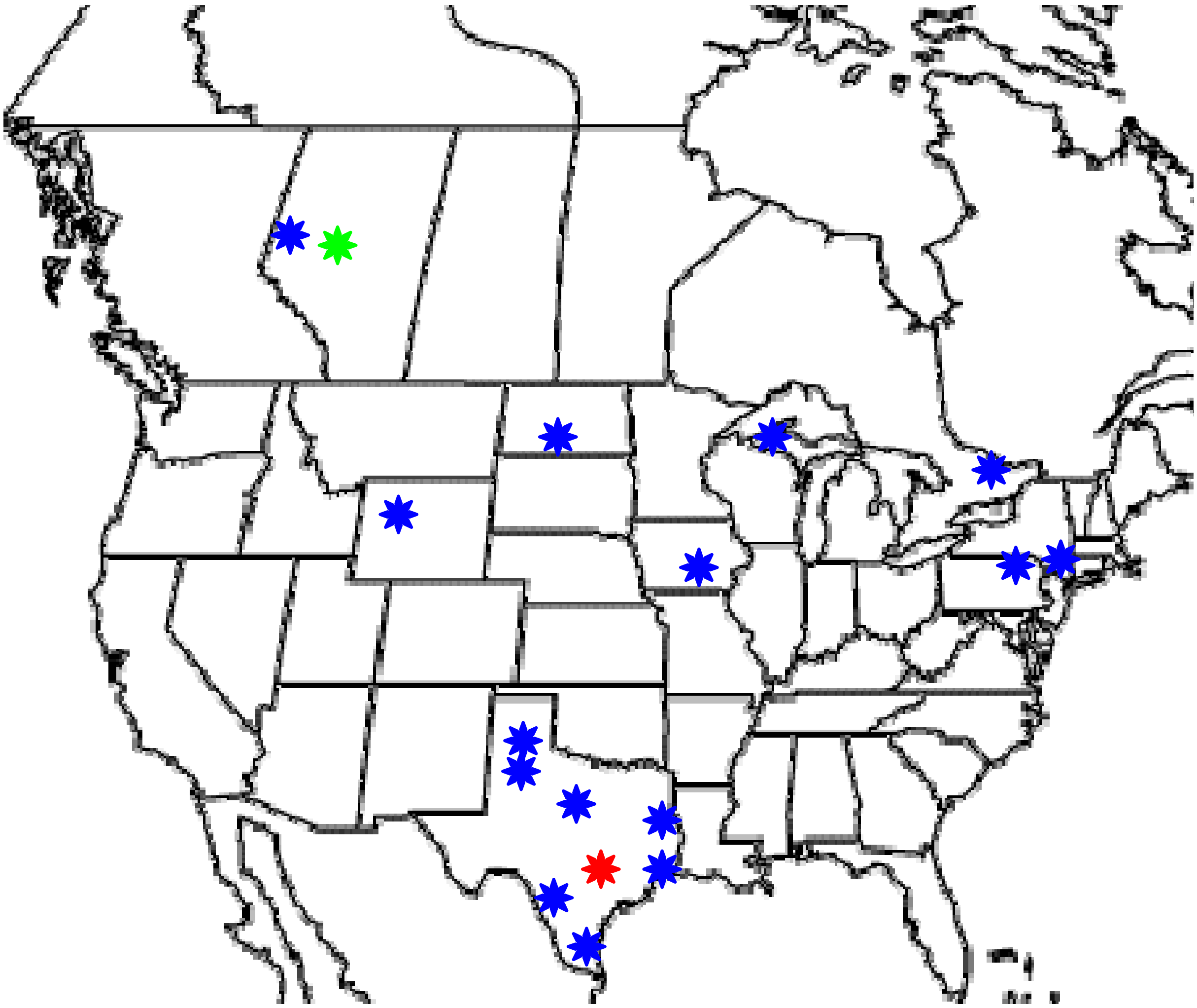

Figure 2-6: Location of study area near Whitecourt, Alberta (green), study area where SWAT model was developed near Temple, Texas (red), and other North American sites where SWAT or EPIC have been applied (blue).

Table 2-1: Summary of modifications made to SWAT as reported in the literature.

Location	Area (km^2)	Annual Precipitation (mm)	Annual Air Temperature (^oC)	Soils / Geology*	Vegetation / Land-use	Modifications	Reference
Ontonagon River, Michigan	3460	1000 to 800	-2.5 to 0	Rock, sandstone, limestone and dolomite.	Forest Wetland/Lake	Snowmelt parameters.	Wu and Johnston, 2007
Upper Wind River Basin, Wyoming	4999	1250 to180	n/a	Varies between bedrock, glacial till, and sand.	Alpine tundra Forest Range (with decreasing altitude)	Snowfall and snowmelt routines.	Fontaine et al., 2002
Ariel Creek, Pennsylvania	39.4	966	7.7	Glacial till, glaciofluvial material and lacustrine deposits. 75% of soils contain fragipans.	Residential Forest Cropland Wetland/Lake	Subsurface hydrology and snowmelt routines recommended.	Peterson and Hamlett, 1998
S. Laguna Madre, Lower Angelina, Upper Sabine, Leon, White, Hondo, Texas	7505 5046 3375 7812 4376 2807	663 1235 1041 767 515 732	n/a	SCL, FSL FSL, LFS FSL LFS Clay loam GRV-L CBV-C	Range, agriculture Forests, pasture Forests, pasture Pasture, ag. Range Agriculture, range Range, agriculture	Water balance components.	Muttiah and Wurbs, 2002
Cannonsville Reservoir, New York	1178	1100	n/a	n/a	Forest (59%) Agriculture (26%) Farmland (10%)	Lateral flow and percolation routines for frozen soils.	Tolson and Shoemaker, 2007
Kerava River, Finland	400	n/a	n/a	Mainly moraine, sand, clay, peat.	Farmland Residential	Replaced snowmelt routine.	Francos et al., 2001
La Glace, Alberta	n/a	475	1.9	Sand, silt, Chernozems	Agriculture	Evaluated EPIC snowmelt routine.	Puurveen et al., 1997
Mandan, N. Dakota Boone, Iowa Bushland, Texas	n/a	550 1016 532	5.3 10.1 13.4	Loam Loam Clay loam	Agriculture	Evaluated EPIC soil temperature routines.	Potter and Williams, 1994

n/a = not applicable
* SCL = sandy clay loam, FSL = fine sandy loam, LFS = loamy fine sand, GRV-L = gravelly loam, CBV-C = very cobbly clay

2.7 CONCLUSION

Many differences were identified between the study sites in the Swan Hills and agricultural lands in southeast Texas where SWAT was developed:

- Soils on the Boreal Plain are mainly fine-grained Luvisols, compared to highly fertile dark Vertisols on the Blackland Prairies in Texas.
- Vegetation in boreal forests is highly complex with overlapping canopy layers and a high degree of spatial variability among sites due to anthropogenic and natural disturbance regimes. In contrast, monocultures of agricultural sites generally have simple vertical structure and similar amounts of vegetation biomass from one location to the next.
- The Swan Hills area is also 16^{o}C colder (mean annual air temperature), dryer (578 vs 910 mm mean annual total precipitation) and snow depth was over 8 times greater than the southern United States.

From these differences, one can conclude that modifications to the SWAT soil temperature submodel will be necessary if it is to be used in the northern climate of the Boreal Plain. Based on the literature, representation of the snowpack and its influence upon soil temperature seems to be a very significant problem when attempting to use the SWAT model in cold regions.

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Chapter 3: Quantifying the Effects of Air Temperature, Soil Moisture and Snow Depth on Soil Temperatures During the Rapid Cooling Period, on the Boreal Plain

3.1 Introduction

To understand streamflow and water quality dynamics within the boreal forest, and to predict how landscape disturbance alters these dynamics, factors that influence water movement through watershed soils are crucial. In boreal forests, the timing and rate of soil freezing in autumn affect the relative volume of water moving over the soil surface and through shallow and deep subsurface flowpaths. When soils are frozen, porosity is reduced and ice lenses form, which prevents infiltration of water to deeper layers, and surface or saturated overland flow can then occur (Carey and Woo, 2001). Thus, the moisture content of the soil at the time of freezing plays a part in determining the degree to which infiltration is reduced in frozen soils relative to unfrozen soils (Kane and Stein, 1983). In association with these processes, soil freezing patterns also affect solute concentrations in water moving over and through soils, because they influence the type of soils the water will encounter and the contact time between water and soil. In general, litter and shallow soil horizons in the boreal forest are richer in organic compounds and labile phosphorus and nitrogen than deeper mineral soil strata (Schoenau et al., 1989; D'Arcy and Carignan, 1997; Anderson and Lepistö, 1998). It follows that lengthening the contact period for water moving through shallow soils would increase concentrations of these solutes. Thus, the annual soil freeze and thaw cycle in the boreal forest exerts a strong influence on water movement patterns, nutrient concentrations in soil water, and water leaving the watershed in the stream channel.

Many parameters interact to influence the cooling of soil temperatures in autumn. Air temperature and soil moisture regulate soil temperature (Smith, 1996; Martinez et al., 2008), as well as interacting with each other. As air temperatures in autumn decrease, dry soils cool proportionally and heat stored within the soils is lost to the atmosphere; this relationship depends on the soil's thermal conductivity. Thermal conductivity is higher for soil than water (Arenson and Sego, 2004); as air temperatures cool in the fall, soils with higher soil moisture cool more slowly relative to dryer soils that have more voids filled with air. In addition, the presence of a snowpack decreases the amount of solar radiation reaching the soil surface, since snow (particularly fresh snow) has high albedo. In opposition to the albedo effect, which limits direct radiant heating of soil, a snowpack also insulates soils, preventing heat loss from the soils to the atmosphere (Hardy et al., 2001). The insulating effect keeps soils warm during winter but the albedo effect delays thawing in the spring. Vegetation cover is a fourth factor that can influence soil temperatures in autumn. Similarly to a snowpack, vegetation directly decreases the amount of solar radiation that reaches the soil surface. For example, forest harvesting in a boreal forest jack pine stand in northern Ontario increased soil temperatures by as much as $6^{o}C$ in summer months (Bhatti et al., 2000). Further, the forest canopy interacts with the snowpack by altering snow accumulation patterns on the forest floor (Metcalfe and Buttle, 1998). Snow interception rates are high for conifer trees and snow depth increases with horizontal distance outward from tree trunks to approximately 3 m, depending on the size of the tree (Faria et al., 2000). As winter progresses, wind can alter the snow distribution patterns on the ground and may clear snow from open areas. During spring, the canopy cover shields the snow from the sun and snowmelt is delayed compared to open sites. Forest harvest experiments in aspen stands in northern Minnesota demonstrated that snowmelt was more rapid and occurred earlier after tree removal (Verry et al., 1983). Although the effects of air temperature, soil moisture, snow depth and vegetation on soil temperatures have been studied individually, there is need for a study that integrates all four factors to determine how they interact.

In the boreal forest, snowmelt can be an important annual hydrologic event. Snowmelt runoff accounted for up to 72% of annual streamflow in the central Canadian Boreal Shield (Beall et al., 2001). Although snowmelt comprises a smaller proportion of annual streamflow on the western Canadian Boreal Plain, where annual snowfall accounts for less than 30% of annual precipitation, snowmelt runoff is important for nutrient movement in watersheds. For example, the snowmelt period was estimated to account for 50% of the nitrate export from Boreal Plain watersheds (Pelster et al., 2008). Therefore efforts to quantify disturbance (forest harvest or wildfire for example) impacts on hydrologic regimes and the quality of receiving waters have tended to consider the timing and rate of snowpack depletion (Hardy et al., 1998; Parviainen and

Pomeroy, 2000; Jones and Pomeroy, 2001). However, a key factor for predicting snowmelt in spring is quantifying the formation of ice in soil layers during the cooling period in autumn (Metcalfe and Buttle, 2001). Further, analysis of soil temperatures during the cooling period is also important for forest products industries operating in the boreal forest, where winter harvesting on frozen soils is commonly employed to limit soil compaction. Ground-based logging equipment increases soil bulk density and strength and decreases air-filled porosity, infiltration rate and hydraulic conductivity, which in turn can negatively affect tree regeneration and growth (Elliot et al., 1998; McNabb et al., 2001).

This study is part of the Forest Watershed and Riparian Disturbance project (FORWARD), a collaborative effort by researchers, forest-related industries, First Nations communities and Federal and Provincial governments focused on two boreal forest study areas in Canada: the western Boreal Plain and the central Boreal Shield. This study is based in the Boreal Plain study area, situated in the Swan Hills of Alberta (Figure 3-1). The main objectives of FORWARD are to collect meteorological and watershed data (vegetation, soils, surface water quality and quantity and bioindicators) to adapt hydrological and water quality models to predict the effects of watershed disturbance, and to apply this knowledge to develop decision support tools to aid in forest management planning (Smith et al., 2003; Prepas et al., 2008*a*). One focus of the modelling efforts within FORWARD has been adapting SWAT, a deterministic hydrological model, for northern climates and forest conditions. To adequately modify the model, parameters that influence soil temperature must be identified. The objectives of this study were to: (1) quantify the effects of air temperature on soil temperature in autumn, (2) quantify the combined effect of air temperature, soil moisture and snow depth on soil temperature, and (3) explain differences in soil temperature among burned, harvested, conifer and deciduous stands, and between upland sites and wetland sites.

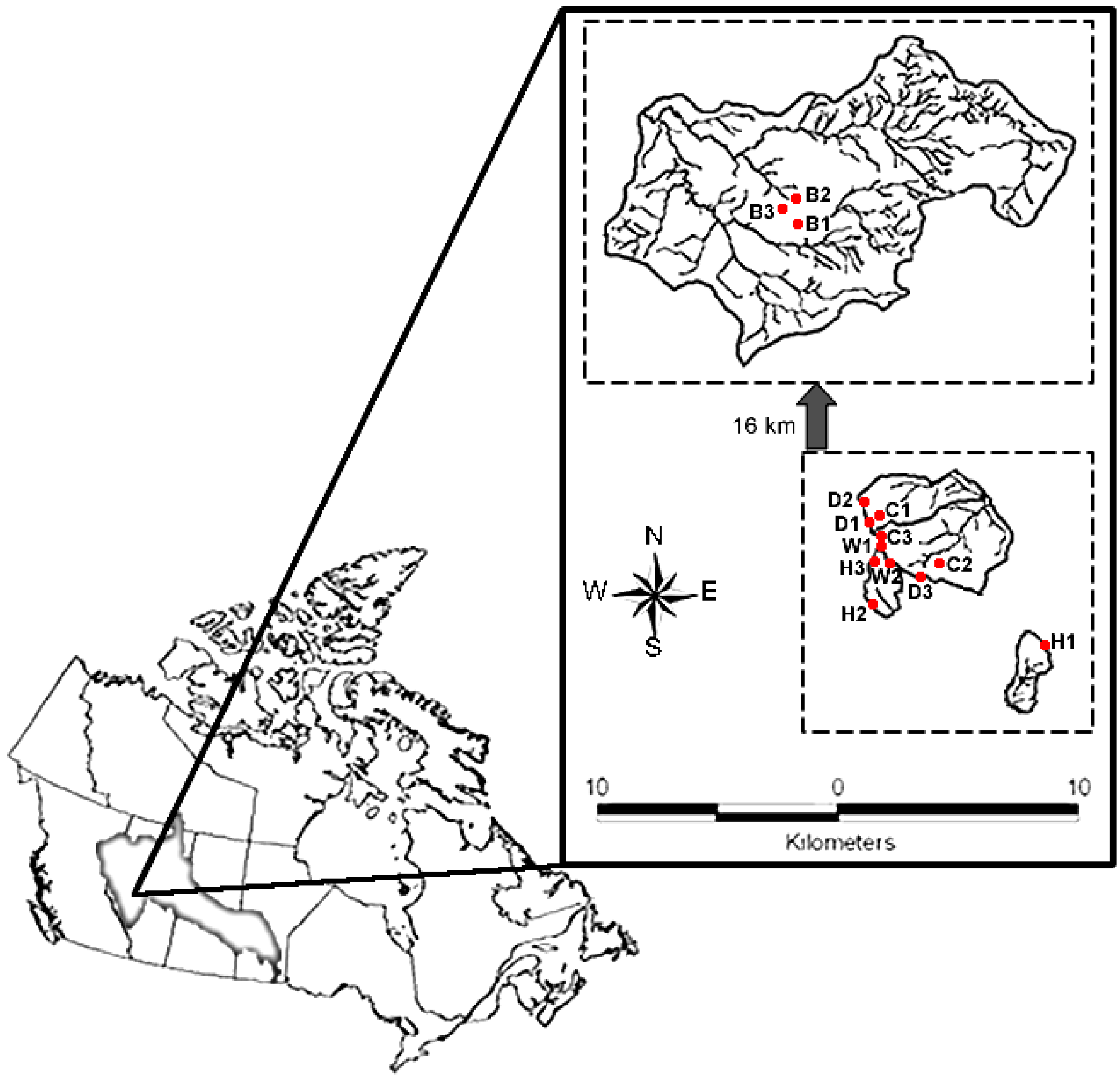

Figure 3-1: Location of study sites within the Boreal Plain (see Table 3-2 for site descriptions).

3.2 Site Description

The Swan Hills study area is approximately 225 km northwest of Edmonton, Alberta, Canada (Figure 3-1). The five first- to third-order study watersheds constitute a subset of the FORWARD research watersheds and cover an area of 182 km^2 (Figure 3-1). The climate of the Boreal Plain is sub-humid (Zoltai et al. 1998); long-term climate normals (1971 to 2000) for precipitation and mean annual air temperature are 578 mm and 2.6^{o}C, respectively, with 24% of total precipitation falling as snow (Environment Canada, 2008). Generally, this area is snow-covered for six to seven months of the year, beginning in October or November (Environment

Canada, 2008). Among the three study years and during the four-month study periods (September to December), 2008 was relatively dry (precipitation was approximately 70% of the climate normal), whereas precipitation in 2007 was similar to the climate normal and 2006 had nearly double the normal precipitation (Table 3-1). Snowfall was lower than normal in 2007 and almost twice as much as normal in 2006. In terms of the mean air temperature, 2006 and 2008 were cooler than the climate normal, and 2007 was 4.6°C warmer than normal (Table 3-1).

Table 3-1: Climate data for each study year and long-term means (1971-2000) from 1 September to 31 December, followed by annual means (1 January to 31 December) in brackets. Data from Environment Canada Whitecourt A station (Environment Canada, 2008).

Year	Total Precipitation (mm)	Total Snowfall (mm)	Mean Air Temperature (°C)
2006	231 (516)	125 (163)	-1.5 (3.6)
2007	125 (533)	12 (94.3)	5.9 (3.8)
2008	92 (438)	data missing	-0.4 (2.7)
Long-term mean	128 (578)	57 (139)	1.3 (2.6)

Upland soils across the study area are mostly fine-grained Orthic Gray Luvisols overlaying deep glacial tills (Whitson et al., 2003) and wetlands consist mainly of peat (Ecological Stratification Working Group 1996). Peat depth varies, but is generally > 0.4 m (Couling et al., 2007). Tree cover is mainly mixed-wood, consisting primarily of black spruce (*Picea mariana* (Mill.) BSP), trembling aspen (*Populus tremuloides* Michx.), lodgepole pine (*Pinus cortorta* Dougl. ex Loud.var. latifolia Engelm) and tamarack larch (*Larix laricina* (Du Roi) K. Koch) (Ecological Stratification Working Group 1996). Other species present in the area include jack pine (*Pinus banksiana* Lamb.), balsam poplar (*Populus balsamifera* L.) and white spruce (*Picea glauca* (Moench) Voss) (Smith et al., 2003). Conifer and deciduous stands across the study area are dominated by lodgepole pine and trembling aspen, respectively (Table 3-2). Most wetlands are treed fens (Couling et al., 2007), dominated by black spruce and/or lodgepole pine (Table 3-2).

Table 3-2: Sample site descriptions

Site ID	Watershed	Cover type	Canopy Cover (%)	Stand Characteristics Species (Cover %)	Elevation (m)	First date of sampling
B1	Goose	Upland Burned	0	n/a	987	9 Aug 2006
B2	Goose	Upland Burned	0	n/a	985	9 Aug 2006
B3	Goose	Upland Burned	0	n/a	991	13 Jul 2006
H1	Kashka	Upland Harvested	0	n/a	1043	1 Aug 2006
H2	Millions	Upland Harvested	0	n/a	991	6 Aug 2006
H3	Millions	Upland Harvested	0	n/a	1016	10 Aug 2006
C1	Thistle	Upland Conifer	92	PL(90) AW(10)	1028	20 Oct 2005
C2	Willow	Upland Conifer	97	PL(90) AW(10)	1005	28 Oct 2005
C3	Willow	Upland Conifer	89	PL(90) AW(10)	1054	3 Nov 2005
D1	Thistle	Upland Deciduous	84	AW(70) SB(20) PL(10)	1053	27 Oct 2005
D2	Thistle	Upland Deciduous	96	AW(60) SB(20) PL(20)	1026	3 Nov 2005
D3	Willow	Upland Deciduous	95	AW(90) PL(10)	1013	28 Oct 2005
W1	Willow	Wetland	69	SB(80) LT(10) PL(10)	1026	28 Jun 2006
W2	Millions	Wetland	71	SB(100)	1028	10 Aug 2006

* PL: lodgepole pine, AW: trembling aspen, SB: black spruce, LT: tamarack larch.

3.3 DATA COLLECTION

Eight reference sites (three upland conifer, three upland deciduous and two wetland) and six disturbed or open sites (three upland burned and three upland harvested) were selected within the FORWARD watersheds (Figure 3-1; Table 3-2). Both wetlands were identified as peatlands, with peat depths ranging from 1.6 to 2.8 m. There were slight differences in elevation and slope (range from 0.01 to 0.04%) among sites (Table 3-2). The burned sites were a result of a severe wildfire in the Virginia Hills in June 1998 (Prepas et al., 2003), where nearly 727,000 hectares of forest burned (Government of Alberta, 2008). Regrowth since the wildfire mainly consists of lodgepole pine with some trembling aspen; tree height ranges from 0.5 to 2 m. The harvested sites were the result of logging activity that occurred between January and March of 2004. All harvested sites received the same treatment after harvest, namely mechanical site preparation (shear blading and Donaren mounding) between March and August 2004, and aerial application of the herbicide glyphosate (2.1 kg/ha) in August 2004 (Prepas et al., 2008*b*). Regrowth since harvesting is sparse and consists of lodgepole pine and white spruce; tree height ranges from 0.5 to 1 m.

Sample sites were set-up in the autumn of 2005 and summer of 2006 and data collection occurred during the autumn to winter transition period (September to December) of 2006, 2007 and 2008. Soil temperature was measured and recorded either by water/soil temperature sensors and HOBO® dataloggers (Onset Computer Corporation, Bourne, MA, USA) or YSI thermistors (YSI Incorporated, Yellow Springs, OH, USA) and Data Dolphin dataloggers (Optimum Instruments Inc., Edmonton, AB, Canada). Measurements were taken hourly at 0.1 and 0.5 m depths for soils and at 2 m above ground level for air temperature (Figure 3-2). Soil moisture content was measured with theta probes (Delta-T Devices Ltd., Burwell, Cambridge, UK) and recorded by either a DL6 soil moisture datalogger (Delta-T Devices Ltd.) or a Data Dolphin datalogger at 0.1 and 0.5 m depths (Figure 3-2). Theta probes were calibrated for mineral or organic soils and measure volumetric moisture content. Snow depth was measured a total of 44 times in 2006 and 2007 during regular site visits. Data from burned and harvested sites were pooled, since site access limited the amount of snow depth data that could be collected. The canopy cover was measured using a spherical densiometer (Forest Densiometers, Bartlesville, OK, USA) in August 2008. Measurements were taken at each site facing the four cardinal directions, and averages of the measurements were taken as the percent cover of each site (Table 3-2).

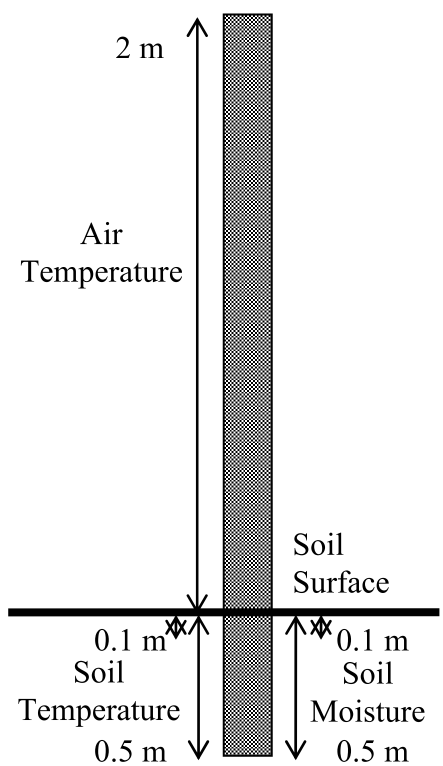

Figure 3-2: Vertical profile of soil temperature, air temperature and soil moisture probes.

3.4 DATA ANALYSIS

For each site, mean daily air temperature and soil temperature and moisture at each depth were calculated. Daily snow depth for 2006 and 2007 was estimated using a linear regression ($P < 0.001$) between snow depth measured at the sites and snow depths for the same dates recorded at the Whitecourt A station (Environment Canada, 2008), located 50 km southeast of the FORWARD study area (Figure 3-3). The intercept was forced through 0. Daily snow depth estimates were modified by calculating the error between estimated and measured values on the 44 sampling dates, conducting a linear interpolation of this correction error between winter sampling dates and then applying the correction error to the estimated depth. The maximum snow depth for upland conifer, upland deciduous, wetland and pooled burned and harvested was recorded for each of the two years. Due to the small data set (2 replicates per treatment for 2006 and 2007), no comparisons of maximum snow depth were made among treatments. However, maximum snow depths for the two years were compared using a paired *t*-test.

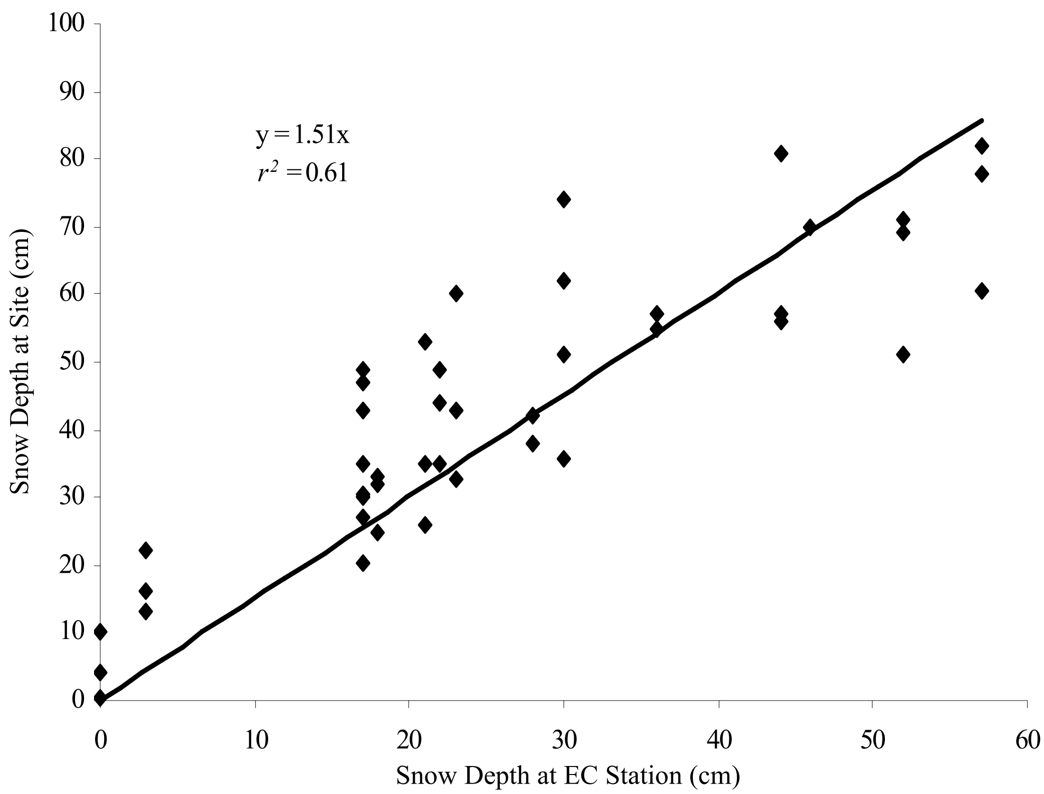

Figure 3-3: Relationship between measured snow depth at study sites and Environment Canada (EC) Whitecourt A station (*P* < 0.001) (Environment Canada, 2008).

Mean 1 September to 31 December air temperature and soil temperature and moisture at 0.1 and 0.5 m depths were calculated for each site for each of the three years. Treatment means were calculated by pooling the three years, therefore the wetland treatment had six replicates (two sites for three years) and each of the four upland treatments had nine replicates (three sites for three years). In some cases, several days of data were missing and a four-month average could not be calculated adequately; therefore the entire site for that year was taken out of the analysis. A one-way Analysis of Variance was used to compare these means among the five treatments, followed by the Tukey test for unequal sample sizes (Daniel, 1995) if $P < 0.05$. Variation around the mean was expressed in terms of Standard Error (*SE*). A paired *t*-test (alpha 0.05) was conducted grouping all treatments to compare soil temperature and soil moisture at 0.1 *versus* 0.5 m depths.

Simple linear regressions were conducted to determine relationships between daily air temperature (independent variable) and daily soil temperature (dependent variable) at 0.1 and 0.5 m soil depths. Slopes of regression lines describing these relationships were compared between upland conifer and deciduous treatments, and between burned and harvested treatments with two-tailed *t*-tests and if slopes were similar, intercept tests were conducted (Zar, 1996). In wetlands, the presence of porous peat at 0.1 m depth meant that soil moisture was not representative of a wetland, and in most cases was dryer than the upland sites at the same depth. Therefore, to include wetlands in this analysis, it was also conducted at 0.5 m depth for all treatments. Multiple linear regressions were carried out for each year to determine the interactions between air temperature, soil moisture and snow depth (independent variables) and soil temperature (dependent variable) (2006-2008 in the absence of snow, 2006-2007 in the presence of snow). Only two representative treatments (upland conifer and wetland) were analyzed this way, to avoid repeatedly using snow depth data in the pooled datasets. Means of percent contribution to variation in soil temperature for all years were calculated and these results presented in terms of general trends. An alpha value of 0.01 was chosen for regression analysis and regression *t*-tests, given the large size of the dataset.

3.5 RESULTS

Mean air temperatures during the period of rapid cooling from September to December differed among cover types. Mean air temperatures in harvested sites were 1.0, 1.1, 1.1 and 1.8^{o}C cooler than burned, conifer, deciduous and wetland sites, respectively (P < 0.001) (Figure 3-4). Mean air temperatures at wetlands were 0.8, 0.7 and 0.7^{o}C warmer than burned, conifer and deciduous sites, respectively (P < 0.001) (Figure 3-4). Mean air temperature was not detectably different between burned, conifer and deciduous sites. Maximum snow depth in 2006 (59.2 ± 1.8 cm) was deeper than in 2007 (41.0 ± 1.8 cm) (P < 0.001).

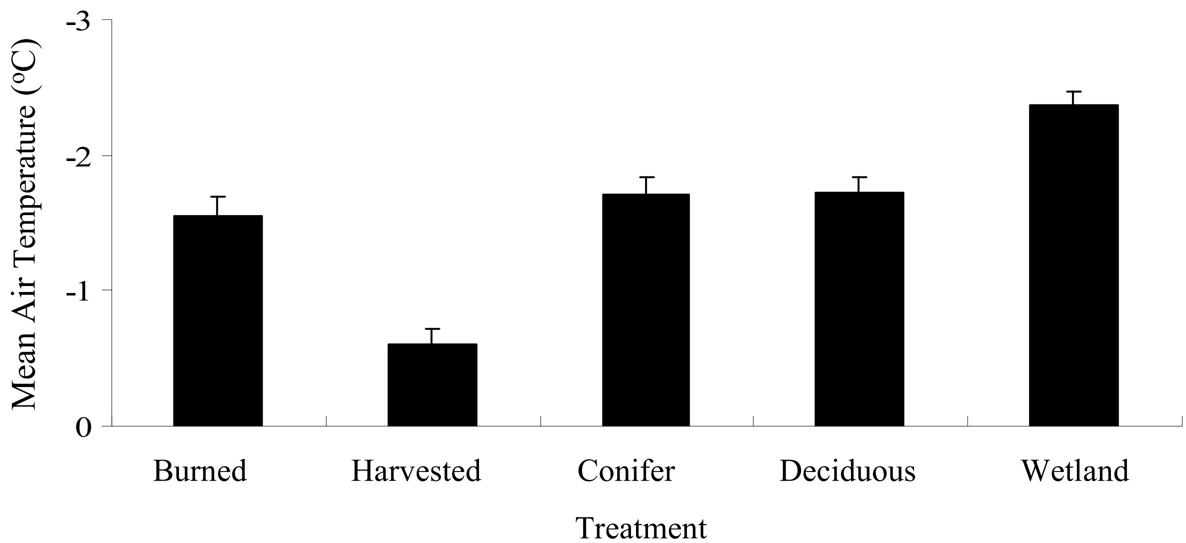

Figure 3-4: Mean (± Standard Error) air temperature from 1 September to 31 December 2006-2008 under five vegetation cover types. Note y-axis is inverted and negative.

3.5.1 SOIL TEMPERATURE AT 0.1 M DEPTH

Mean soil temperatures during the rapid cooling period in autumn followed different patterns in the absence (September and October) and presence (November and December) of snow. In the absence of snow, mean soil temperatures at harvested sites were 0.6, 1.0, 1.0 and 1.1^{o}C warmer than at burned, conifer, deciduous and wetland sites respectively, and burned sites were 0.6^{o}C warmer than wetland sites (P < 0.001) (Figure 3-5a). The mean soil temperature did not differ between conifer and deciduous sites (Figure 3-5a). In the presence of snow, mean soil temperatures did not differ among treatments (P = 0.43) (Figure 3-5b). Therefore in the absence of snow, soil temperatures depended upon treatment, whereas in the presence of snow, treatment did not affect soil temperature at 0.1 m depth.

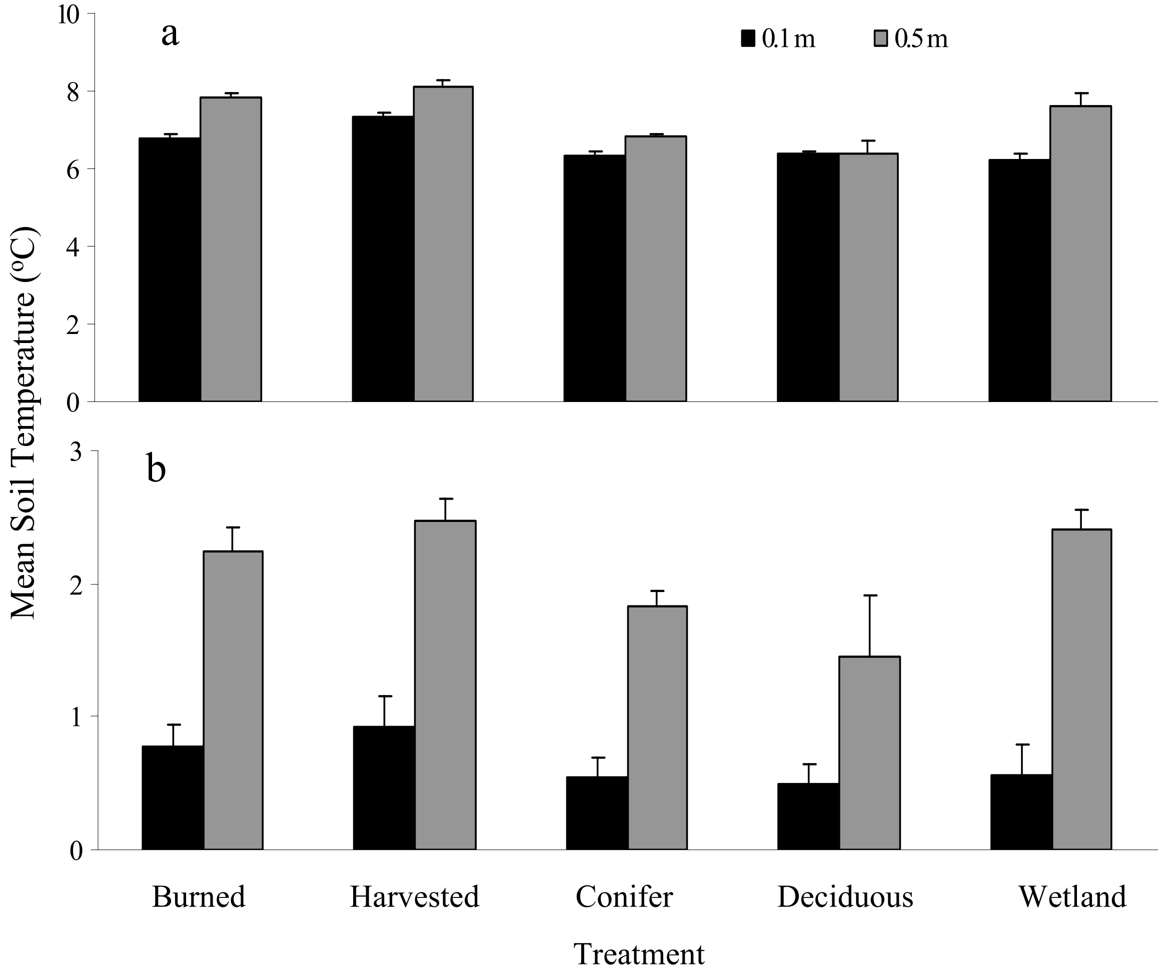

Figure 3-5: Mean (± Standard Error) soil temperature from (a) 1 September to 31 October 2006-2008 and (b) 1 November to 31 December 2006-2008, under five vegetation cover types at 0.1 and 0.5 m soil depths.

Over the four month period, a positive relationship existed between soil temperature at 0.1 m depth and air temperature for burned and harvested sites (i.e. in the absence of tree cover). It was evident that as air temperature cooled in the autumn, the soil temperature declined until an air temperature of approximately -5°C, at which point the soil temperatures reached a plateau (Figure 3-6; note x axis is inverted to show the cooling trend). This coincided with the presence of snow; therefore the data were separated into snow-free and snow-covered periods (delineation indicated by vertical line in Figure 3-6).

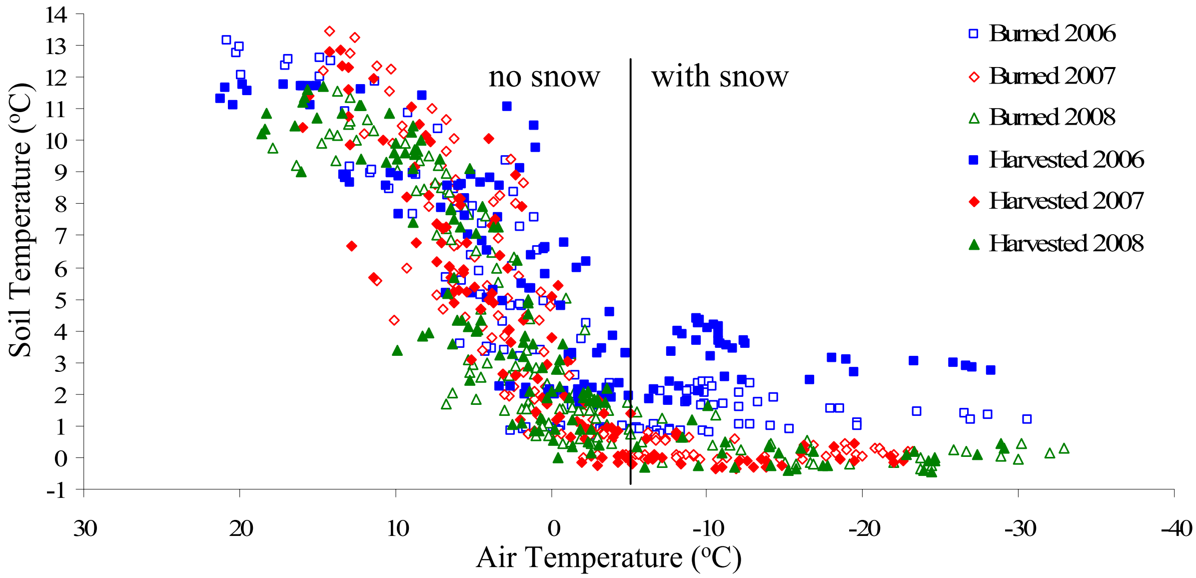

Figure 3-6: Mean daily soil temperature at 0.1 m depth *versus* mean daily air temperature at 2 m above ground surface for burned and harvested sites in 2006, 2007 and 2008. Vertical line shows an approximate demarcation of absence and presence of snow. Note x-axis is inverted to show the cooling trend.

When the relationships between soil and air temperature are compared for all five treatments, three distinct groups emerged: upland disturbed, upland treed and wetlands. As determined by two-tailed *t*-tests, slopes and intercepts of the lines describing relationships between soil temperature and air temperature did not differ between burned and harvested sites in the absence ($P = 0.29$ and 0.80, respectively) or presence of snow ($P = 0.67$ and 0.83, respectively), therefore these data were pooled into two datasets (upland disturbed without snow, upland disturbed with snow). In the absence of snow, a much larger percentage of the variation in soil temperature could be explained by air temperature than in the presence of snow (Figure 3-7). Similarly, the slopes and intercepts of the regression lines for soil temperature *versus* air temperature were similar for conifer and deciduous sites in the absence ($P = 0.14$ and 0.95, respectively) and presence of snow ($P = 0.92$ and 0.87, respectively), therefore these data were also pooled into two datasets (upland treed without snow, upland treed with snow).

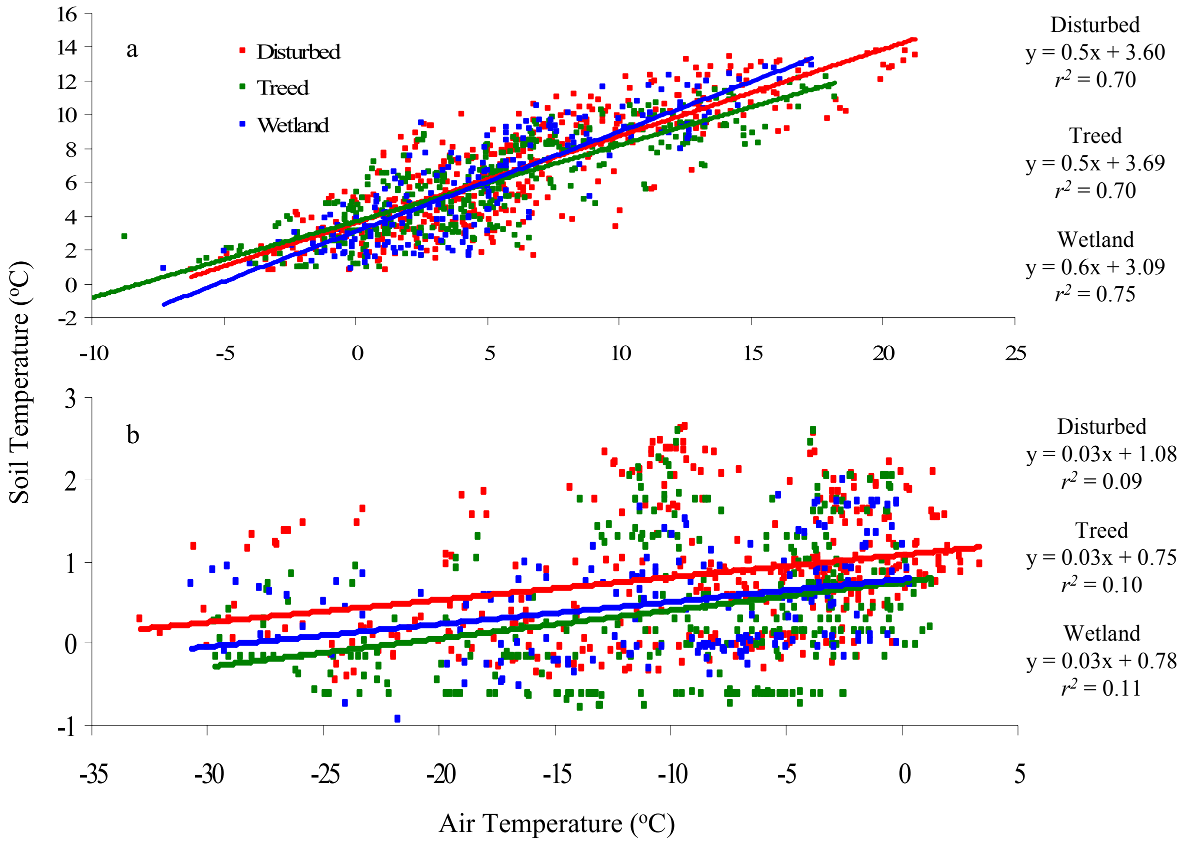

Figure 3-7: Mean daily soil temperature at 0.1 m depth *versus* mean daily air temperature at 2 m above ground surface for disturbed, treed, and wetland sites in (a) the absence and (b) the presence of snow from 1 September to 31 December 2006-2008. For all regressions $P < 0.001$. Note change in scale for both axes.

Relative to the untreed disturbed sites, the presence of trees in upland sites affected the relationship between air and soil temperature. In the absence of snow, the slope of the regression line was steeper for disturbed and wetland sites than for treed sites ($P = 0.007$ and < 0.001, respectively) (Figure 3-7a). There were no detectable differences in slopes and intercepts between disturbed and wetland sites ($P = 0.014$ and 0.57, respectively). In the presence of snow, regression slopes did not differ among disturbed, wetland, and treed sites ($P > 0.35$), but the intercept for disturbed sites was 0.3°C warmer than treed and wetland sites ($P < 0.001$) (Figure 3-7b). Intercepts of treed and wetland sites were similar ($P = 0.14$). Thus, in the absence of snow, upland tree cover affected the rate of change in soil temperature with each decrement in air temperature, whereas the disturbed and wetland sites behaved similarly to each other. In the presence of snow, soil temperature at disturbed sites were warmest for a given air temperature, whereas soil temperatures at upland treed and wetland sites were similar for a given air temperature.

At the 0.1 m depth, mean soil moisture from September to December for the three years was 79, 58 and 67% higher in burned, conifer and deciduous sites, respectively, compared to wetlands (P = 0.002) (Figure 3-8). To determine how air temperature interacted with soil moisture and snow depth to influence soil temperature, multiple regressions were conducted in both the absence and presence of snow for upland (represented by upland conifer) and wetland sites. In the absence of snow, air temperature and soil moisture accounted for 69.8 ± 5.1% and 3.2 ± 2.0%, respectively, of the variation in soil temperature in upland sites among the three years ($P < 0.001$) (Figure 3-9a). In 2006, there was no relationship between soil temperature and soil moisture at these upland sites ($P = 0.42$). In wetlands in the absence of snow, air temperature and soil moisture explained 74.8 ± 5.4% and 7.7 ± 2.8% of the variation in soil temperature, respectively, among the three years ($P < 0.001$) (Figure 3-9a). Snow cover altered these patterns. For upland sites in the presence of snow, air temperature, soil moisture and snow depth explained 2.0 ± 1.0%, 91.2 ± 3.3%, and 0.6 ± 0.6% of the variation in soil temperature, respectively ($P < 0.001$) (Figure 3-9a). In 2006, there was no relationship between soil temperature and snow depth at upland sites ($P = 0.08$). For wetlands in the presence of snow, air temperature, soil moisture and snow depth explained 37.1 ± 19.9%, 38.5 ± 22.3%, and 1.3 ± 1.3% of the variation in soil temperature, respectively ($P < 0.001$) (Figure 3-9a). In 2007, there was no relationship between soil temperature and snow depth at wetland sites ($P = 0.49$). For upland and wetland sites, air temperature and soil moisture were the most important factors in the absence and presence of snow, respectively.

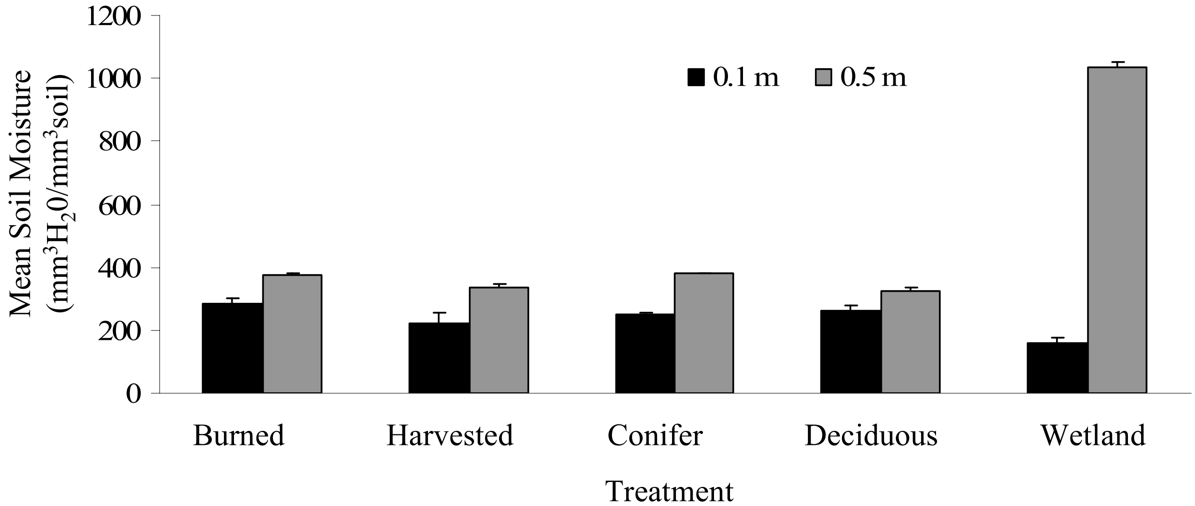

Figure 3-8: Mean (± Standard Error) soil moisture from 1 September to 31 December 2006-2008 under five vegetation types.

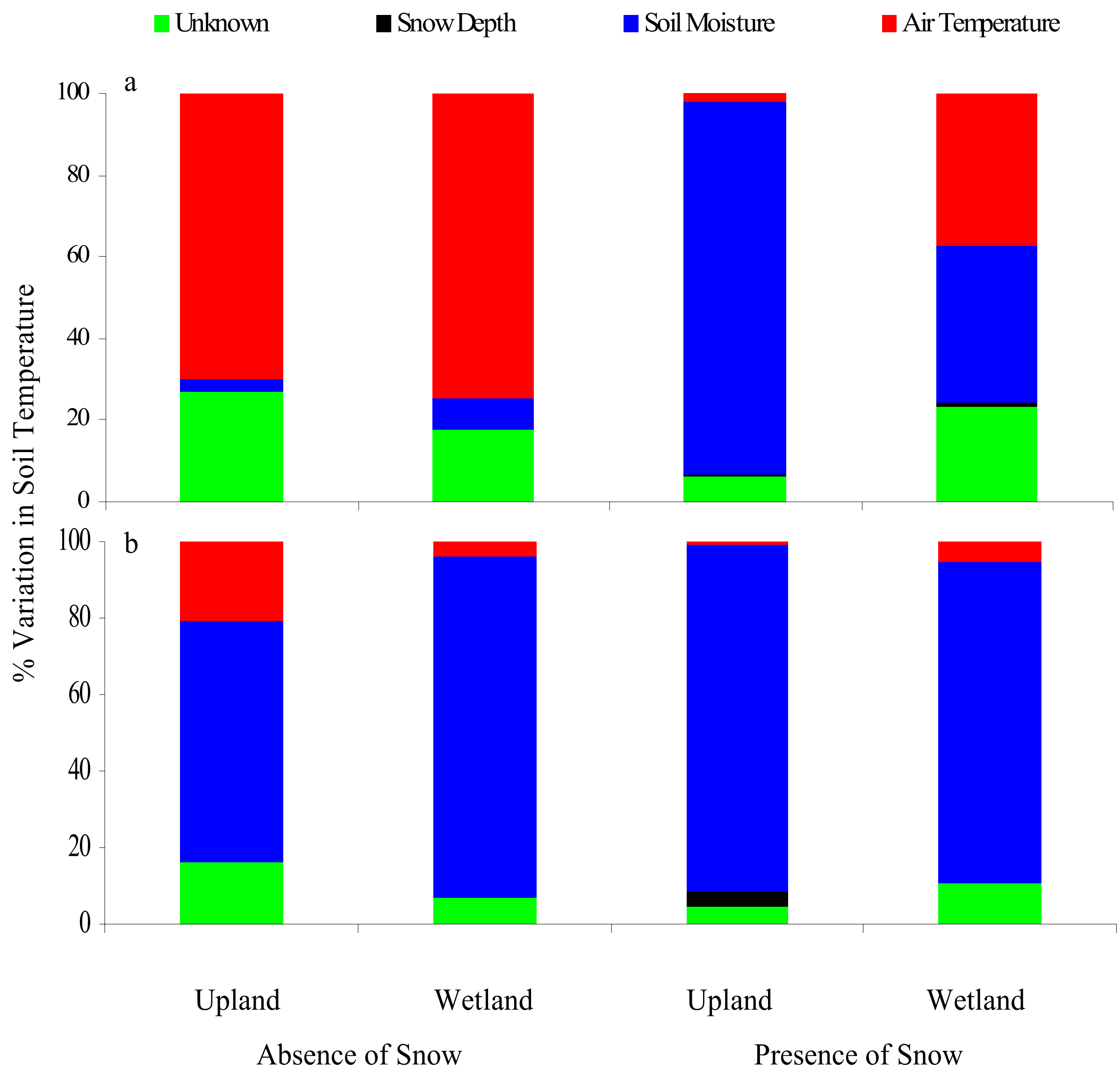

Figure 3-9: Percentage of contribution to soil temperature based on multiple linear regression analyses for upland conifer and wetland sites at (a) 0.1 m and (b) 0.5 m depths.

3.5.2 Soil Temperature at 0.5 m Depth

Mean soil temperatures for all treatments grouped were warmer at the 0.5 m than 0.1 m depth for September to October (P = 0.03) (Figure 3-5a) and November to December (P = 0.001) (Figure 3-5b), with a more dramatic difference between the two depths later in the autumn. As with the 0.1 m depth, mean soil temperatures at the 0.5 m depth followed separate patterns in the absence and presence of snow during the rapid cooling period, coinciding with an air temperature of approximately -5°C (Figure 3-10). However, differences among treatments did follow the same pattern as at the 0.1 m depth. For September to October, mean soil temperatures in burned sites were 1.0 and 1.4°C warmer than conifer and deciduous sites, respectively, warmer in harvested than conifer (1.2°C) and deciduous sites (1.7°C) and 1.2°C warmer in wetland than deciduous sites ($P < 0.001$) (Figure 3-5a). For November to December, mean soil temperatures were the same for all cover types, with the exception that harvested sites were 1.0°C warmer than deciduous sites ($P = 0.02$) (Figure 3-5b).

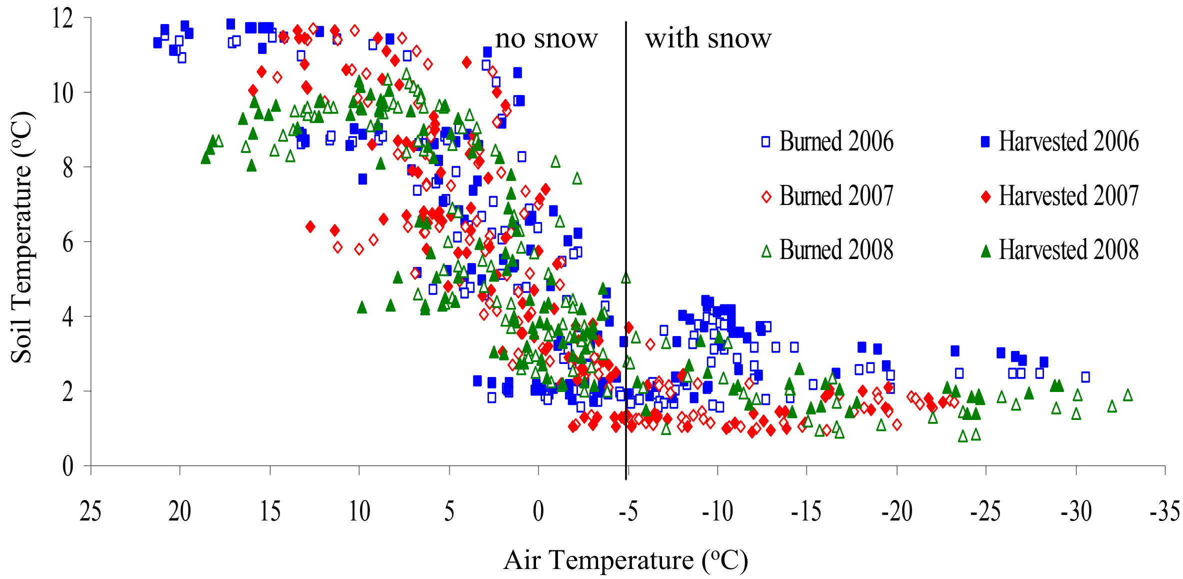

Figure 3-10: Mean daily soil temperature at 0.5 m depth *versus* mean daily air temperature at 2 m above ground surface for burned and harvested sites in 2006, 2007 and 2008. Vertical line shows an approximate demarcation of absence and presence of snow. Note x-axis is inverted to show the cooling trend.

A positive relationship was detected between soil temperature and air temperature for all treatments. Similar to the 0.1 m depth, the data for each treatment were separated into two sets, for the absence and presence of snow (Figure 3-10; note x axis is inverted to show the cooling trend). In the absence of snow, for each unit of air temperature decrease, soil temperature at 0.5 m cooled less than at 0.1 m ($P < 0.001$). In the presence of snow, slopes of the soil temperature *versus* air temperature lines were similar at 0.1 and 0.5 m depth ($P = 0.70$), however the intercept of shallow soils was 1.5°C cooler than deeper soils ($P < 0.001$). In the absence of snow, the relationship between soil temperature at 0.5 m depth and air temperature (Figure 3-11a) explained 20% less variation when compared to the same regressions at the 0.1 m depth (Figure 3-7a). In the presence of snow, the relationship between soil and air temperature explained only 3% less variation at the 0.5 m depth (Figure 3-11b) than the 0.1 m depth (Figure 3-7b). The slopes and intercepts of the lines describing relationships between soil temperature and air temperature did not differ between burned and harvested sites in the absence (P = 0.40 and 0.49, respectively) or presence of snow (P = 0.29 and 0.03, respectively), therefore these data were pooled into two datasets (disturbed without snow, disturbed with snow). Slopes of soil temperature *versus* air temperature lines did not differ between conifer and deciduous sites in the absence (P = 0.77) or presence of snow (P = 0.83), however intercepts of conifer sites were 0.6 and 0.4^{o}C warmer than deciduous sites, in the absence and presence of snow, respectively (P < 0.001). These data were therefore treated separately (conifer without snow, conifer with snow, deciduous without snow, deciduous with snow). In the absence of snow, the treatments (disturbed, conifer, deciduous and wetland) shared the same slope (P > 0.30) but for all cases intercepts were different (P < 0.005) (Figure 3-11a). The intercept from disturbed sites was 0.7, 1.3 and 0.3^{o}C warmer than conifer, deciduous, and wetland sites, respectively (P < 0.005) (Figure 3-11a). In the presence of snow, there were no differences in the slopes and intercepts of disturbed and wetland regressions (P = 0.17 and 0.86, respectively), while all other treatments shared the same slope (P > 0.17) and had different intercepts (P < 0.001) (Figure 3-11b). Therefore in the absence of snow and for a given air temperature, soil temperatures were warmest for disturbed, wetland, conifer and deciduous sites, and in the presence of snow, disturbed and wetland sites were warmest, followed by conifer and deciduous sites.

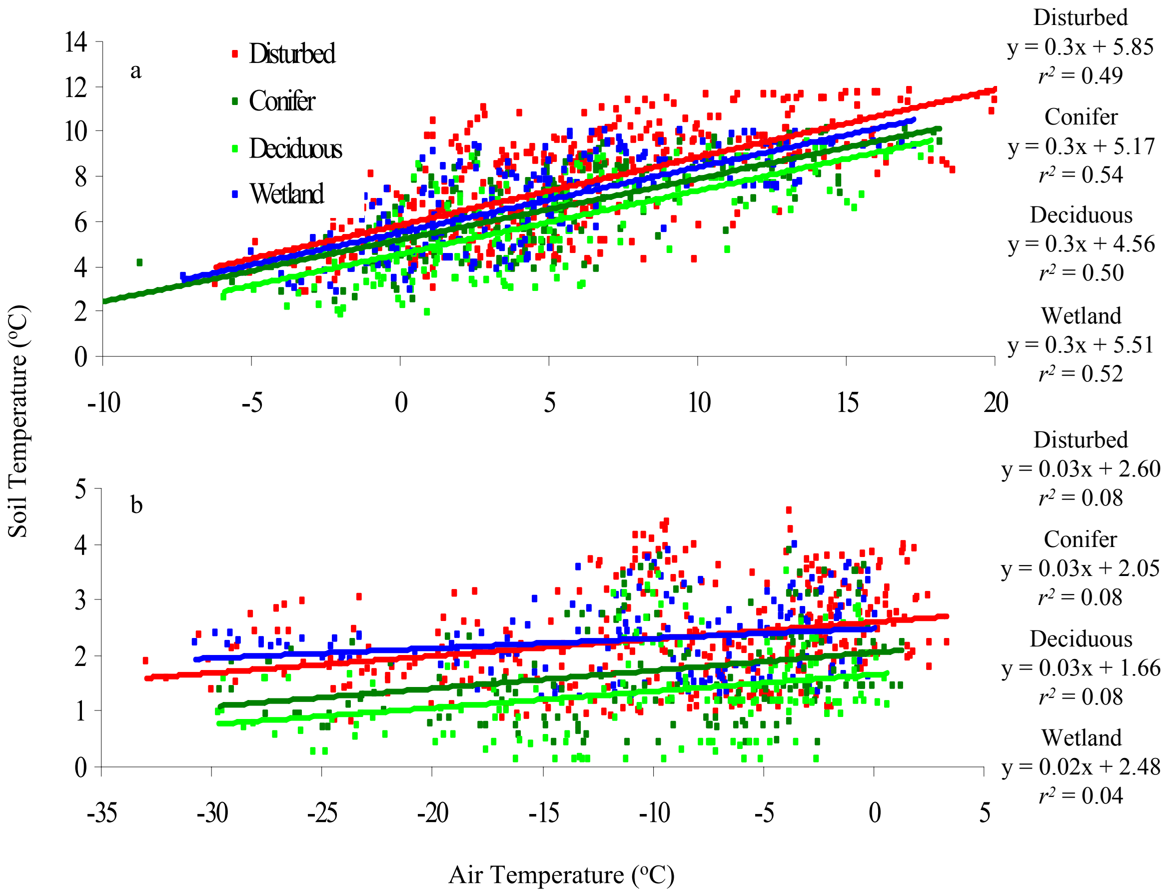

Figure 3-11: Mean daily soil temperature at a 0.5 m depth *versus* mean daily air temperature at 2 m above ground for disturbed, treed and wetland sites in (a) the absence and (b) the presence of snow from 1 September to 31 December 2006-2008. For all regressions *P* < 0.001. Note change in scale.

For all cover types, soil moisture for all treatments grouped was higher at the 0.5 m depth when compared to 0.1 m ($P < 0.008$) (Figure 3-8). Mean soil moisture from September to December was 3.0 ± 0.1 times higher in wetlands than all other treatments, and soil moisture in burned and conifer sites was 16 and 17% higher, respectively, than deciduous sites ($P < 0.001$) (Figure 3-8). In the absence of snow for upland sites, air temperature and soil moisture explained 20.9 ± 20% and 62.5 ± 29% of the variation in soil temperature ($P < 0.001$) (Figure 3-9b). In wetlands, air temperature and soil moisture accounted for 4.0 ± 1.8% and 89.2 ± 5.9% of the variation in soil temperature in the absence of snow (Figure 3-9b). In the presence of snow, air temperature, soil moisture and snow depth at upland sites explained 0.6 ± 0.6%, 91.2 ± 0.6%, and 4.0 ± 0.8%, respectively, of the variation in soil temperature ($P < 0.001$) (Figure 3-9b). In 2006, there was no relationship between air and soil temperature at 0.5 m ($P = 0.09$). For wetland sites, soil temperature was only explained by air temperature (5.2 ± 1.2%) and soil moisture (83.9 ± 11.5%) ($P < 0.001$), because there was no relationship with the snow depth for either year ($P > 0.08$) (Figure 3-9b). Thus in either the absence or presence of snow, soil moisture was the most important factor affecting soil temperatures at the 0.5 m depth.

3.6 DISCUSSION

The simplest scenario analyzed in this evaluation of factors that influence soil temperature was disturbed (open) upland sites in the absence of snow. Positive linear relationships between air and soil temperature at 0.1 m depth at these sites support similar observations for bare soils at an agricultural site in Germany (Langholz, 1989). For each decrement in air temperature, disturbed upland soils at 0.1 m cooled 1.7 times more than soils at 0.5 m. During the cooling of fine-grained soils at shallow depths, heat flow increases due to large differences between air and soil temperatures. At deeper soil depths, these differences in temperature are reduced and heat flow decreases, therefore soil temperatures cool more slowly (Smith, 1996). The decreasing cooling rate in deeper soils may also be related to the thermal gradient of soils (that decreases with soil depth) and higher soil moisture (wetter soils have lower diffusivity, a lower rate of temperature change) (Smith, 1996). Increased soil moisture at deeper soil depths also alters the thermal conductivity of the bulk soil by filling voids with water rather than air. Consequently, with increasing depth and soil moisture, the cooling rate of soils decreases.

Under snow cover, the decrease in soil temperature with each unit decrease in air temperature was only 6 (0.1 m) to 10% (0.5 m) of the decrease observed when there was no snow cover. In cold regions, snow cover has a critical role to play in soil temperature dynamics. It creates an insulating layer over the soil due to its low thermal conductivity and reduces heat exchange between the soil and atmosphere (Zhang et al., 1997; Mölders et al., 2003). This is particularly important in disturbed sites, where there is no vegetation to intercept snow or shade snow from solar radiation. Although snow depth was deeper in 2006 than 2007 in the present study, there were no evident differences in soil temperature between the two years. In addition, snow depth accounted for little or none of the variation in soil temperature at either 0.1 or 0.5 m in the multiple regression analyses. Both of these observations support the concept that insulation from a snowpack is nonlinear and that a threshold snow depth exists beyond which the insulation from any additional snow does not provide an increased insulation effect (Henry, 2008). However, one could argue that snow density would be a better measure of the insulation from a snowpack. For this study snow density was not measured. Snow depth measured in snow water equivalents could provide a better indication of insulation.

In the absence of snow, air temperature was the most important factor influencing soil temperature in the upland and wetland sites at the 0.1 m depth, whereas at 0.5 m, soil moisture became more important. The reduced importance of air temperature as a determinant of soil temperature with increasing soil depth was also observed in an agricultural site in Germany (Langholz, 1989). The stronger relationship between air and soil temperatures for wetlands (when compared to upland sites) at the 0.1 m depth may explain why in the multiple linear regressions, soil moisture was more important in upland sites than wetland sites. Snow depth was relatively unimportant when compared to air temperature and soil moisture in the multiple linear regressions at both soil depths. This shows that any relationships between soil temperature and snow depth in simple linear regressions are linked to other parameters and that snow depth alone cannot be used as an effective measure for predicting soil temperatures. However, the presence of snow itself was important, since for all cases (upland and wetland soils at both depths), soil moisture became the most important factor influencing soil temperatures when snow was present. The snowpack insulates soils from cold air, reducing its effect on soil temperatures. Deep snowpacks in the spring can delay soil thawing and increase overland flow, resulting in lower concentrations of organic compounds being transported to streams. As soils freeze, individual water particles move towards the freezing layer by cryogenic suction (Michalowski and Zhu, 2006) and liquid water content decreases with decreasing temperature, which reduces the overall mobility of soil water because infiltration and percolation of water within frozen soils are inefficient (Mölders et al., 2003). In the presence of snow, the snowpack decreased the effect of air temperature and increases the effect of soil moisture on regulating soil temperatures at both soil depths.

The presence of trees was associated with cooler soil temperatures for a given air temperature compared to disturbed sites. Unlike the 0.1 m depth, where there were no differences between conifer and deciduous sites, intercepts of conifer sites were 0.4^{o}C warmer than deciduous sites at the 0.5 m depth. Warmer soils under conifer cover relative to deciduous cover could be explained by stronger insulation from the thicker canopy during winter months, because deciduous leaf senescence begins in late August to early September in the boreal forest. During the freezing period in the presence of snow, soil temperatures of treed sites were again cooler than disturbed sites (which coincides with the trend in air temperature). This might be attributable to the open canopy in disturbed sites that allows soils to accumulate more heat during the summer months before snowfall.

At the 0.1 m depth, shallow peat from wetlands was dryer than the upland disturbed and treed soils. Wetlands store large amounts of water at deeper depths but near the surface, peat moisture content fluctuates because evaporation from this shallow layer is high (Hökkä et al., 2008). Peat has a lower thermal conductivity when compared to fine-grained soils and when dry, it releases heat quickly during winter and insulates soils during summer (Smith, 1996). This could explain the stronger relationship observed between soil and air temperatures in the absence of snow for wetlands than for any other cover type at the 0.1 m depth. At the 0.5 m depth, deep peat was significantly wetter and soil moisture exceeded 100% on a volume basis; this could be achieved because voids are completely saturated with water and the organic matter itself contains moisture. As moisture increases, diffusivity decreases and soils retain more heat, which explains the strong relationship between soil temperature and moisture in the deeper soil layer of a wetland (Figure 3-12).

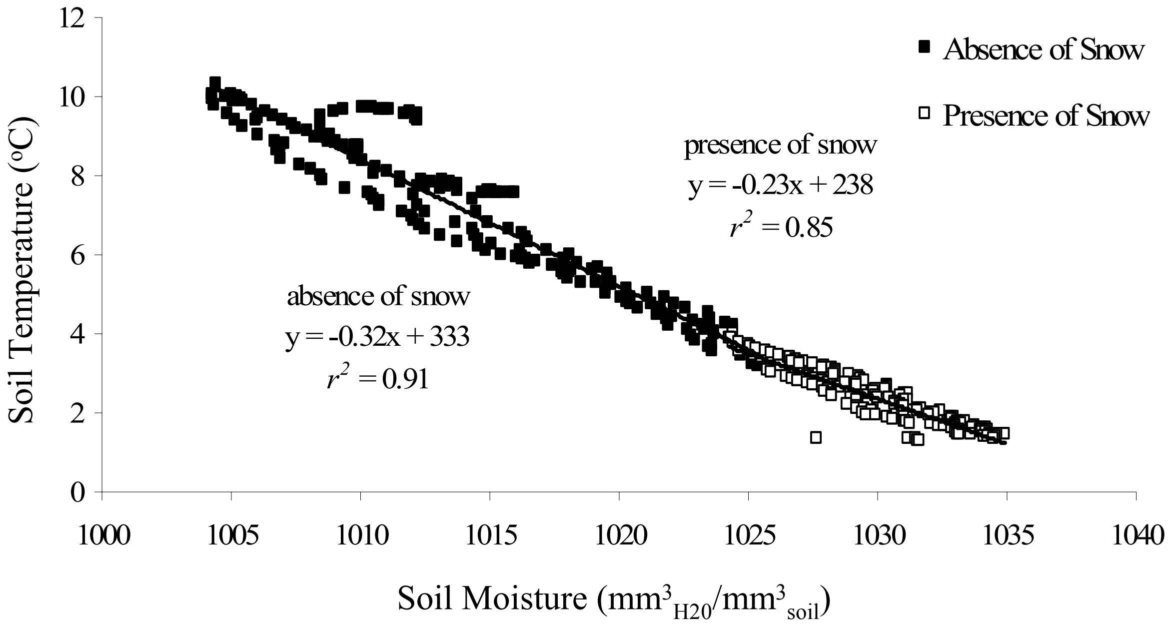

Figure 3-12: Relationship between wetland soil temperature and soil moisture at 0.5 m depth. For all regressions *P* < 0.001.

This analysis was a first attempt for the FORWARD study group to use an extensive data set to determine factors that affect soil temperatures and explain an important element in hydrological modelling: soil freezing. Movement of water in forests during unfrozen and frozen conditions is highly dependent on the presence of frozen soils. Linking modelling to forest management requires additional consideration of soil freezing as it relates to water movement. Forest removal is known to increase soil moisture by removing an important sink for moisture: evapotranspiration (Bosch and Hewlett, 1982). For the forest industry operating on the Boreal Plain, compaction due to machinery on these fine-grained soils has been a concern for half a century (McNabb et al., 2001). For this reason, winter harvesting on frozen soils is common; therefore knowing the timing of freeze/thaw cycles is important. Daily and yearly variations in air temperature, soil moisture and snow depth still pose a problem in this type of analysis. To provide more confidence in the soil temperature predictions and to develop a tool to pinpoint when soils freeze, long-term studies are still required. Although linear relationships were identified in this study, it is possible that non-linear functions might better explain variations observed in soil freezing patterns in these boreal forest soils. Within the FORWARD project, the importance of snow's effect on soil temperatures needs to be addressed in the deterministic modelling. Although the current analysis would suggest that a simple snow presence/absence approach is sufficient, a more detailed analysis of snow depth would determine if there is a threshold depth beyond which the influence of snow depth plateaus. Further, snow density measurements might clarify or strengthen relationships between snow and soil temperatures. By combining these findings with the hydrological modelling that has already been conducted (Putz et al., 2003; Watson et al., 2008), we will be better able to project streamflow year-round and perform complete annual water and nutrient budgets for disturbed and forested watersheds.

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CHAPTER 4: ANNUAL FREEZE/THAW TEMPERATURE CYCLE IN SOILS ON THE CANADIAN BOREAL PLAIN: COMPARISON OF SWAT PREDICTIONS TO MEASURED DATA[1]

4.1 INTRODUCTION

A critical element in modelling the quantity and quality of water in streams draining undisturbed and disturbed forested watersheds is to understand how the annual freeze/thaw cycle affects soil temperatures. Frozen soils are important factors in forest hydrology, because they prevent infiltration of water to deeper layers and may cause surface or saturated overland flow (Carey and Woo, 2001). Soil freezing patterns also affect solute concentrations in water moving over and through soils, because they influence the type of soils the water will encounter and the contact time between water and soils. In boreal forests, the timing and rate of soil freezing affect the relative volume of water moving over the soil surface and through shallow and deep subsurface flowpaths in the autumn and following spring.

The SWAT model has been applied across the United States and around the globe to predict water movement in watersheds. However, SWAT was developed for agricultural lands in the southern United States (on the Blackland Prairies of southeast Texas), where soils are considerably different from those in boreal forests. Therefore it is expected that the soil temperature submodel within SWAT will require some modifications if it is to be applied to boreal forest conditions, as has been demonstrated in other studies in cold regions, where changes were made to the snowmelt and subsurface hydrology routines when frozen soils were present (Fontaine et al., 2002; Tolson and Shoemaker, 2007). The soil temperature algorithms in SWAT are similar to and based upon those in the Erosion/Productivity Impact Calculator (EPIC) (Williams et al., 1984). However, EPIC includes a 5-day moving average to simulate heat stored in the soil, which is absent in SWAT. In addition, the lag coefficient (that controls the influence of the previous day's temperature on the current day) in EPIC is set to a value of 0.5, compared

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[1] A version of this chapter was submitted to the 5th International SWAT Conference, 3-7 August 2009, Boulder, CO. Authors for this paper were: J.A. Bélanger, G. Putz, B. Watson and E.E. Prepas.

to 0.8 in SWAT (Potter and Williams, 1994). It is critical that the SWAT algorithms be tested stringently for cold regions because of differences in the hydrologic cycle in these areas compared to warmer climates.

This study is part of the Forest Watershed and Riparian Disturbance project (FORWARD), a collaborative effort by researchers, forest-related industries, First Nations communities, and Federal and Provincial governments in Canada. The main objectives of FORWARD are to collect watershed data (vegetation, soils, weather, surface water quality and quantity, and bioindicators) to adapt hydrological and water quality models to predict the effects of watershed disturbance, and to apply this knowledge to decision support tools to aid in forest management planning in Canada (Smith et al., 2003; Prepas et al., 2008*a*). One focus of the modelling efforts within FORWARD has been adapting the SWAT hydrological model for northern climates and forest conditions. The objective of this study was to verify the suitability of the soil temperature algorithms within SWAT to model soil temperatures in cold regions, specifically, the Canadian Boreal Plain, for one representative example of each of the following five forest land classification types: upland burned, harvested, conifer and deciduous forest and wetland forest.

4.2 MODEL DESCRIPTION

SWAT is a semi-distributed, physically based hydrological model that operates on a continuous daily time-step (Neitsch et al. 2005). SWAT was developed in the early nineties for agricultural land by the USDA. It is foremost a watershed hydrology model and the water balance is the driving force for all functions, which include hydrology, sediment quantity and transport, and pesticide and nutrient (nitrogen, phosphorus) concentrations. The model calculates forest canopy water storage, infiltration, lateral subsurface flow, ground water flow, surface runoff (using either SCS curve number procedure (USDA Soil Conservation Service, 1972) or Green & Ampt infiltration method (Green & Ampt, 1911)) and potential and actual evapotranspiration (where potential evapotranspiration is calculated by the Penman-Monteith (Monteith, 1965), Priestley-Taylor (Priestley and Taylor, 1972) or Hargreaves method (Hargreaves et al., 1985)). Redistribution is also calculated and both the variable storage coefficient (Williams, 1969) and Muskingum (Overton, 1966) methods are used as routing equations.

To facilitate the modelling process, SWAT allows users to divide watersheds into multiple subbasins. The model then partitions the subbasins into different hydrologic response units based on their vegetation type, soil characteristics and land use. This allows the model to group similar sites together. SWAT also separates the soil profiles into up to ten different layers to accurately simulate belowground processes (Neitsch et al. 2005).

4.3 SWAT SOIL TEMPERATURE SUBMODEL

The soil temperature submodel in SWAT utilizes multiple input parameters that can be roughly separated into three categories: soil characteristics, vegetation cover and climate (Table 4-1; Figure 4-1). To calculate the soil temperature (T in ^{o}C) at a specific depth in the soil (z) on the current day ($T_{soil}(z, d_n)$), SWAT utilizes a weighted soil temperature based on the soil temperature on the previous day ($T_{soil}(z, d_{n-1})$) and the soil surface temperature on the current day (T_{ssurf}):

Eq. 4-1
$$T_{soil}\left(z, d_n\right) = \lambda \cdot T_{soil}\left(z, d_{n-1}\right) + \left(1-\lambda\right) \cdot \left\{df \cdot \left(T_{AAir} - T_{ssurf}\right) + T_{ssurf}\right\}$$

where λ is a lag coefficient that controls the influence of the previous day's temperature on the current day (set at 0.8 in SWAT), df is a depth factor, which accounts for the influence of the depth of soil below the surface, and T_{AAir} is the average annual air temperature based on long-term data (Neitsch et al., 2005).

The depth factor (df) accounts for the influence of the depth of soil below the surface:

Eq. 4-2
$$df = \frac{zd}{zd + \exp\left(-0.867 - 2.078 \cdot zd\right)}$$

where zd is a depth ratio, calculated with the following equation:

Eq. 4-3
$$zd = \frac{z}{dd}$$

where z is the depth at the center of the soil layer in mm and dd is the damping depth in mm.

The damping depth is obtained from:

Eq. 4-4 $$dd = dd_{\max} \cdot \exp\left[\ln\left(\frac{500}{dd_{\max}} \right) \cdot \left(\frac{1-\varphi}{1+\varphi} \right)^2 \right]$$

where dd_{max} is the maximum damping depth in mm and φ is a scaling factor. The maximum damping depth is a function of the soil bulk density, ρ_b in $Mg \cdot m^{-3}$:

Eq. 4-5 $$dd_{\max} = 1000 + \frac{2500\rho_b}{\rho_b + 686 \cdot \exp(-5.63\rho_b)}$$

The scaling factor (φ) is calculated from the following equation:

Eq. 4-6 $$\varphi = \frac{SW}{(0.356 - 0.144\rho_b) \cdot z_{tot}}$$

where z_{tot} is the depth from the soil surface to the bottom of the soil profile in mm and SW is the soil moisture in mm.

The soil surface temperature is calculated from:

Eq. 4-7 $$T_{ssurf} = bcv \cdot T_{soil}(1, d_{n-1}) + (1 - bcv) \cdot T_{bare}$$

where bcv is above ground insulation and T_{bare} is the temperature of bare soil in ^oC. The above ground insulation is calculated by a weighting factor that ranges from 0 (for bare soil) and 1 (100% cover):

Eq. 4-8 $$bcv = \max\left\{\frac{CV}{CV + \exp\left(7.563 - 1.297 \cdot 10^{-4} \cdot CV\right)}, \frac{SNO}{SNO + \exp\left(6.055 - 0.3002 \cdot SNO\right)}\right\}$$

where CV is above ground vegetation biomass and residue in kg·ha^{-1} and SNO is snow-water equivalents in mm.

The temperature of bare soil is calculated from:

Eq. 4-9 $$T_{bare} = T_{av} + \varepsilon_{sr} \frac{\left(T_{mx} + T_{mn}\right)}{2}$$

where T_{av}, T_{mx} and T_{mn} are average, maximum and minimum air temperatures in ^{o}C and ε_{sr} is a radiation term, calculated empirically from:

Eq. 4-10 $$\varepsilon_{sr} = \frac{H_{day} \cdot \left(1 - \alpha\right) - 14}{20}$$

where H_{day} is solar radiation in MJ·m^{-2}d^{-1} and α is surface albedo, which varies according to the cover present at the surface. The albedo for a snow-covered surface is set as 0.8 in the SWAT model. When vegetation is present and snow-water equivalents are below 0.5 mm, the value of albedo depends on cover and is calculated using the following equation:

Eq. 4-11 $$\alpha = \alpha_{plant} \cdot \left(1 - cv_{sol}\right) + \alpha_{soil} \cdot cv_{sol}$$

where α_{plant} is the plant albedo, α_{soil}, is the soil albedo, and cv_{sol} is a biomass factor. In the SWAT soil temperature submodel, the plant and soil albedo values were set to 0.23 (Neitsch et al., 2005) and 0.13 (URS Corporation, 2002) respectively. The biomass factor is calculated based on aboveground vegetation biomass and residue (CV from Eq. 8) (Neitsch et al., 2005):

Eq. 4-12 $$cv_{sol} = \exp\left(-5.0 \times 10^{-5} \cdot CV)\right)$$

Table 4-1: SWAT soil temperature model input parameters.

Parameter	Unit	Symbol
Aboveground vegetation biomass and residue	$kg{\cdot}ha^{-1}$	CV
Albedo	unitless ratio	α
Daily mean temperature	^oC	T_{av}
Daily maximum temperature	^oC	T_{mx}
Daily minimum temperature	^oC	T_{mn}
Depth at the center of the soil layer	mm	z
Depth of soil profile	mm	z_{tot}
Soil bulk density	$Mg{\cdot}m^{-3}$	ρ_b
Soil moisture	mm	SW
Solar radiation	$MJ{\cdot}m^{-2}d^{-1}$	H_{day}
Snow cover	mm	SNO

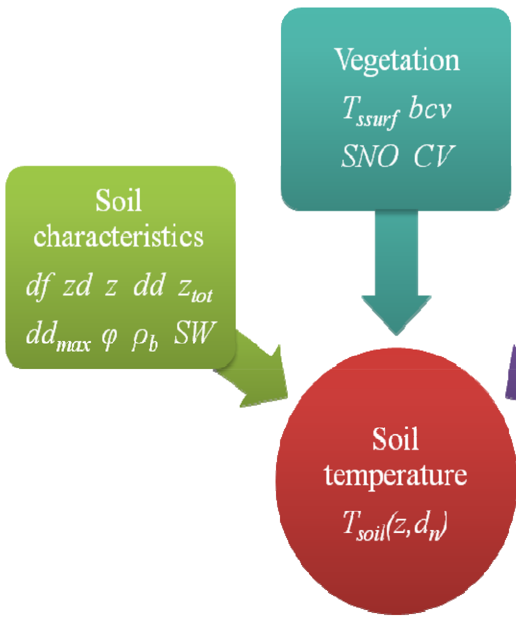

Figure 4-1 Schematic of SWAT soil temperature submodel.

4.4 Model Testing

The SWAT soil temperature algorithms were set up in a Microsoft Excel spreadsheet and all input parameters were integrated into the model. A sensitivity analysis was conducted by varying the following parameters by ± 20%: aboveground vegetation biomass and residue, depth of the soil profile, bulk density, snow equation coefficients and lag coefficient. The model was modified by changing the most sensitive parameters and other default values. Original (default parameter values) and modified (varied parameter values) SWAT soil temperature submodels were run for each site at various depths and treatments. Daily differences between predicted and observed data were calculated and maximum differences for summer and winter seasons were determined. Model outputs were compared by calculating the correlation coefficient (r^2), mean absolute error (*MAE*), and root mean square error (*RMSE*).

4.5 SITE DESCRIPTION

The study area is located in the Swan Hills which is approximately 225 km northwest of Edmonton, Alberta, Canada (Figure 4-2). The four first- to third-order watersheds used in this study constitute a subset of the watersheds monitored by FORWARD and cover an area of 178 km^2 (Figure 4-2; Table 4-2). The climate of the Boreal Plain is sub-humid (Zoltai et al. 1998); long-term climate normals (1971 to 2000) for precipitation and mean annual air temperature are 578 mm and 2.6^{o}C, respectively, with 24% of total annual precipitation falling as snow (Environment Canada, 2008). Generally, this area is snow-covered for six to seven months of the year, beginning in October or November (Environment Canada, 2008). All three study years were dryer than normal; precipitation in 2006, 2007, and 2008 was 11, 8, and 24% lower than normal, respectively. Snowfall was 17% higher than normal in 2006 and 32% lower in 2007 and snowfall in 2008 was normal. In terms of the mean annual air temperature, 2006, 2007 and 2008 were 1.0, 1.2 and 0.1^{o}C warmer than the climate normal (Environment Canada, 2008).

Table 4-2: Sample site descriptions (see also Figure 4-2).

Site ID	Watershed	Cover type	Stand Species Cover (Approx. %)*	Canopy Cover (%)	Elevation (m)	First date of sampling
B	Goose	Upland Burned	n/a	0	991	13 Jul 2006
H	Millions	Upland Harvested	n/a	0	1016	10 Aug 2006
C	Thistle	Upland Conifer	PL(90) AW(10)	92	1028	20 Oct 2005
D	Thistle	Upland Deciduous	AW(70) SB(20) PL(10)	84	1053	27 Oct 2005
W	Willow	Wetland	SB(80) LT(10) PL(10)	69	1026	28 Jun 2006

* PL: lodgepole pine, AW: trembling aspen, SB: black spruce, LT: tamarack larch.

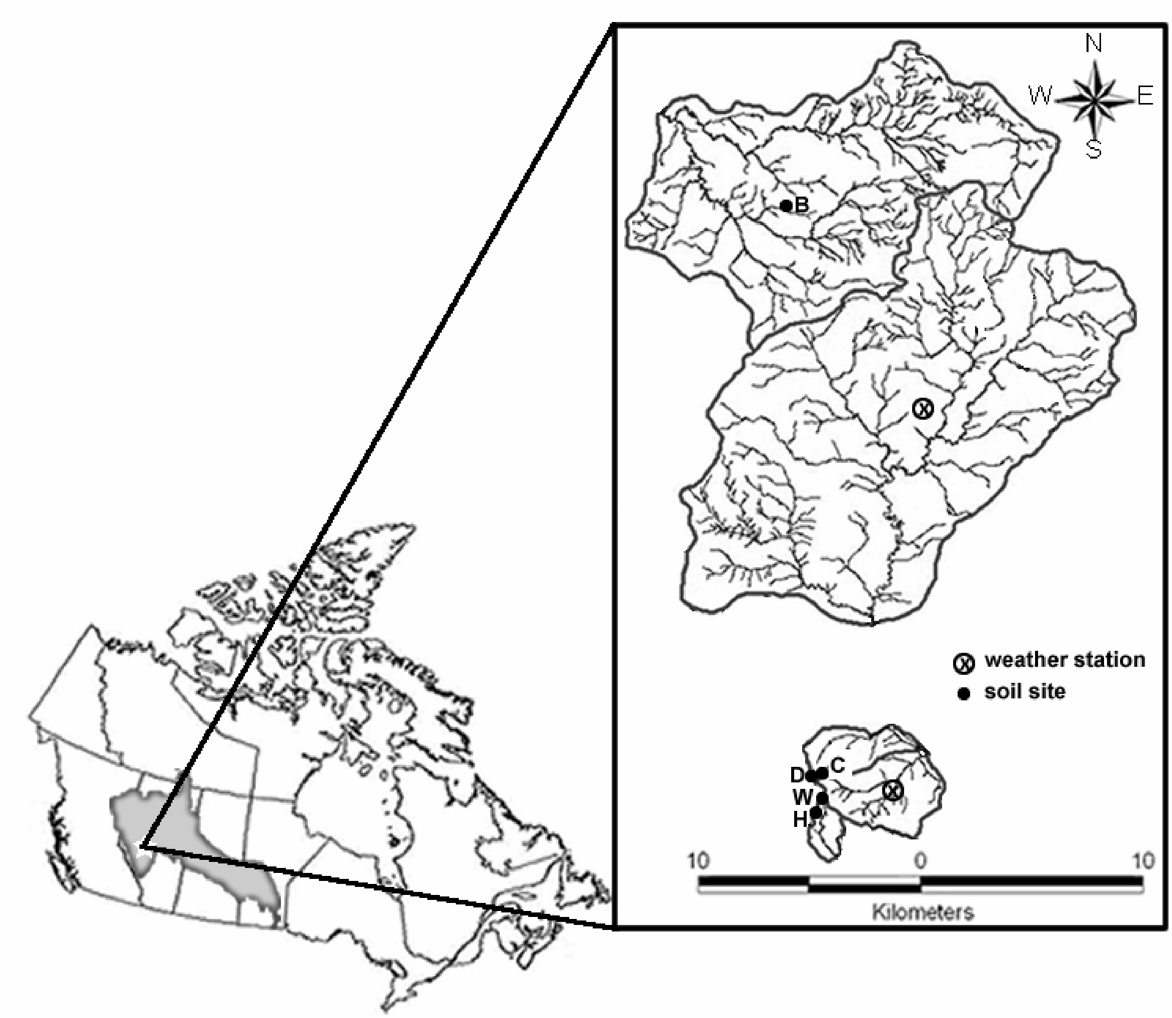

Figure 4-2: Location of burned (B), harvested (H), conifer (C), deciduous (D) and wetland (W) soil sites within the FORWARD study area.

Upland soils across the study area are mostly fine-grained Orthic Gray Luvisols overlaying deep glacial tills (Whitson et al., 2003), and wetlands mainly consist of peat (Ecological Stratification Working Group 1996). Peat depth varies, but is generally > 0.4 m (Couling et al., 2007). Conifer and deciduous stands across the study area are dominated by lodgepole pine (*Pinus cortorta* Dougl. ex Loud.var. latifolia Engelm) and trembling aspen (*Populus tremuloides* Michx.), respectively (Ecological Stratification Working Group, 1996) (Table 4-2). Wetlands are treed fens (Couling et al., 2007), dominated by black spruce (*Picea mariana* (Mill.) BSP) and include some lodgepole pine and tamarack larch (*Larix laricina* (Du Roi) K. Koch) (Ecological Stratification Working Group, 1996) (Table 4-2).

4.6 DATA COLLECTION

Five sites (one each of: upland burned, upland harvested, upland conifer, upland deciduous and wetland) were selected within the FORWARD watersheds (Figure 4-2). The wetland is a peatland, with peat depth of 2.8 m. There are slight differences in elevation and slope (range 0.01 to 0.04%) among sites (Table 4-2). The burned site is a result of a severe wildfire in the Virginia Hills in June 1998 (Prepas et al., 2003). Regrowth since the wildfire mainly consists of lodgepole pine with some trembling aspen; tree height ranges from 1.5 to 2 m. The harvested site is the result of logging activity that occurred in January 2004. This site received post-harvest mechanical site preparation (shear blading and Donaren mounding) between March and July 2004, and aerial application of the herbicide glyphosate (2.1 kg/ha) in August 2004 (Prepas et al., 2008*b*). Regrowth since harvesting is sparse and consists of lodgepole pine and white spruce (*Picea glauca* (Moench) Voss); tree height ranges from 0.5 to 1 m.

Sampling sites were set-up in the autumn of 2005 and summer of 2006. Soil temperature was measured hourly and recorded either by water/soil temperature sensors and HOBO® dataloggers (Onset Computer Corporation, Bourne, MA, USA) (conifer and deciduous sites) or YSI thermistors (YSI Incorporated, Yellow Springs, OH, USA) and Data Dolphin loggers (Optimum Instruments Inc., Edmonton, AB, Canada) (burned, harvested, and wetland sites). Measurements were taken at 0.1, 0.5 and 1 m depths for soil temperature and at the ground surface and 2 m above ground level for air temperature (Figure 4-3). Soil moisture content was measured hourly with theta probes (Delta-T Devices Ltd., Burwell, Cambridge, UK) and recorded by either a DL6 soil moisture datalogger (Delta-T Devices Ltd.) or a Data Dolphin datalogger at 0.1, 0.5 and 1 m depths (Figure 4-3). Snow depth was measured a total of 44 times in 2006 and 2007 during regular site visits. Air temperature (at 2 m above ground level), precipitation and solar radiation data were collected hourly at two FORWARD weather stations located in the study area (Figure 4-2). For each sampling site, precipitation and solar radiation data from the nearest weather station were used in the modelling. Also, in the case of missing air temperature data, values from the nearest soil temperature site or weather station were used. Measurements of aboveground vegetation biomass and residue from the study area were obtained from an earlier study (MacDonald et al., 2007). Canopy cover was measured using a spherical densiometer (Forest Densiometers, Bartlesville, OK, USA) in August 2008. Measurements were taken at each site facing the four cardinal directions, and averages of the measurements were taken as the percent cover of each site (Table 4-2). Mean soil bulk density was measured as 1.5 $Mg \cdot m^{-3}$ with the exception of wetlands at the 0.1 m depth, where bulk density was 0.3 $Mg \cdot m^{-3}$ (I. Whitson, Univ. Alberta, pers. comm.).

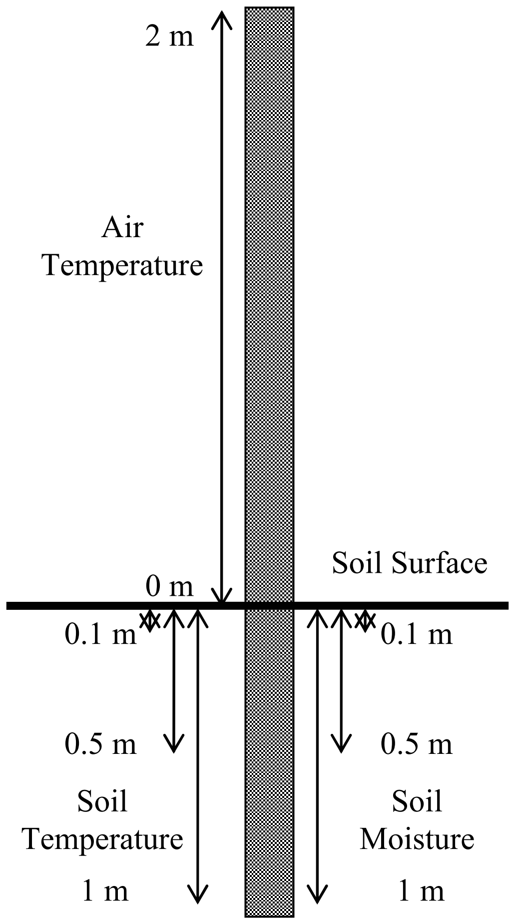

Figure 4-3: Vertical profile of soil temperature, air temperature and soil moisture probes.

4.7 DATA ANALYSIS

For each site, mean, maximum and minimum daily air temperature, and mean daily soil temperature and moisture at each depth were calculated. Daily snow depth (in cm) was estimated using a linear regression ($r^2 = 0.61$; $P < 0.001$) between snow depth measured at the sites and snow depths for the same dates recorded at the Whitecourt A station (Environment Canada, 2008), located 50 km southeast of the FORWARD study area (Eq. 4-13).

Eq. 4-13 *estimated snow depth at study sites = 1.51 * snow depth at Whitecourt*

The intercept was forced through 0 in the equation above. Daily snow depth estimates were modified by calculating the error between estimated and measured values on the 44 sampling dates, conducting a linear interpolation of this correction error between winter sampling dates and then applying the correction error to the estimated depth.

4.8 RESULTS AND DISCUSSION

Given the broad range in daily air temperature conditions on an annual basis on the Boreal Plain (i.e. from a mean daily minimum of -17^{o}C in January to a mean daily maximum of 22^{o}C in July (Environment Canada, 2008)), differences between predicted and observed soil temperatures were identified separately for the summer (Jun-Aug) and winter (Nov-Mar) months. For all sites and depths, the original SWAT model predicted soil temperatures that were 5.2 (conifer) to 17.3^{o}C (burned) warmer than observed in summer and 4.1 (conifer) to 7.8^{o}C (burned) cooler than observed in winter (Table 4-3; Figure 4-4; Appendix). The latter outcome was unexpected, since air temperatures in the study area fall below freezing 201 days each year (Environment Canada, 2008), compared to 31 days in southeast Texas (National Weather Service Forecast Office, 2007). A time-shift was also identified in transition periods. Predicted soil temperatures decreased sooner in the autumn and increased sooner in the spring than was observed. For the original SWAT model, over the three year period for each site and depth, the *MAE* ranged from 1.8 (wetland) to 4.3 (burned)) (Table 4-3). The time-shift increased with soil depth and was larger for treed sites than open sites (Figure 4-4; Appendix). Cooler than predicted air temperatures in the spring transition period and summer could be attributed to the vertically complex, overlapping forest canopy in these boreal forest sites, which likely provided more shade than the vegetation in croplands, for which the model was developed. In higher latitudes compared to southeast Texas, the low angle of the sun over the horizon also reduces the amount of solar radiation reaching the ground. Conversely, warmer than predicted air temperatures in the autumn transition period and winter could be due to the insulation effect of aboveground vegetation biomass, particularly at conifer sites. During winter months, differences are likely the result of insulation by snow, which reaches a depth of 22 cm on average in January in the study area (Environment Canada, 2008), compared to 2.5 cm for southeast Texas (City-Data, 2008). Most snow-related equations in the original SWAT model were probably not subjected to the same testing rigour for a wide range of conditions as the other components of the model because at southern latitudes, below freezing days are few and deep snowpack is absent. Modifications were often required when the model was applied in other cold regions, such as Pennsylvania (Peterson and Hamlett, 1998), Michigan (Wu and Johnston, 2007), Wyoming (Fontaine et al., 2002) and Finland (Francos et al., 2001).

Table 4-3: Maximum differences in summer and winter soil temperatures, and error measurements between predicted and observed model outputs for original and modified SWAT model outputs at 0.1, 0.5 and 1.0 m depths for each site.

Treatment	SWAT model	Lag coefficient	Maximum difference in summer (°C)	Maximum difference in winter (°C)	r^2	*MAE*	*RMSE*
			0.1 m				
Burned	Original	0.80	11.6	-7.8	0.90	4.3	4.9
	Modified	0.95	6.7	-6.7	0.93	3.4	4.1
Harvested	Original	0.80	7.9	-5.0	0.89	2.3	2.8
	Modified	0.95	6.8	-5.0	0.90	1.8	2.3
Conifer	Original	0.80	13.1	-4.8	0.83	3.0	3.8
	Modified	0.96	7.3	-4.8	0.90	2.3	2.7
Deciduous	Original	0.80	17.3	-4.6	0.81	2.9	3.8
	Modified	0.96	14.9	-4.8	0.91	2.2	2.7
Wetland	Original	0.80	9.9	-4.3	0.91	1.8	2.2
	Modified	0.94	6.8	-3.8	0.94	1.4	1.7
			0.5 m				
Burned	Original	0.80	9.2	-7.2	0.77	3.7	4.1
	Modified	0.97	2.5	-4.6	0.96	1.9	2.4
Harvested	Original	0.80	7.3	-5.7	0.78	2.6	3.1
	Modified	0.96	1.8	-4.6	0.91	1.3	1.8
Conifer	Original	0.80	10.2	-4.4	0.70	2.6	3.1
	Modified	0.97	3.8	-2.9	0.92	1.1	1.4
Deciduous	Original	0.80	11.6	-4.2	0.63	2.6	3.2
	Modified	0.97	4.9	-3.4	0.90	1.2	1.5
Wetland	Original	0.80	7.9	-4.6	0.66	2.2	2.7
	Modified	0.97	1.9	-1.2	0.96	0.7	0.8
			1.0 m				
Burned	Original	0.80	7.3	-5.9	0.54	3.0	3.4
	Modified	0.97	1.7	-3.3	0.91	1.7	2.0
Harvested	Original	0.80	5.2	-6.7	0.60	2.6	3.2
	Modified	0.96	2.0	-3.9	0.88	1.8	2.1
Conifer	Original	0.80	9.3	-4.1	0.52	2.0	2.5
	Modified	0.97	5.3	-1.5	0.88	0.8	1.1
Deciduous	Original	0.80	9.1	-4.7	0.43	2.2	2.7
	Modified	0.97	4.9	-2.3	0.82	1.1	1.3
Wetland	Original	0.80	6.0	-4.8	0.26	2.1	2.6
	Modified	0.97	1.6	-2.7	0.74	1.1	1.3

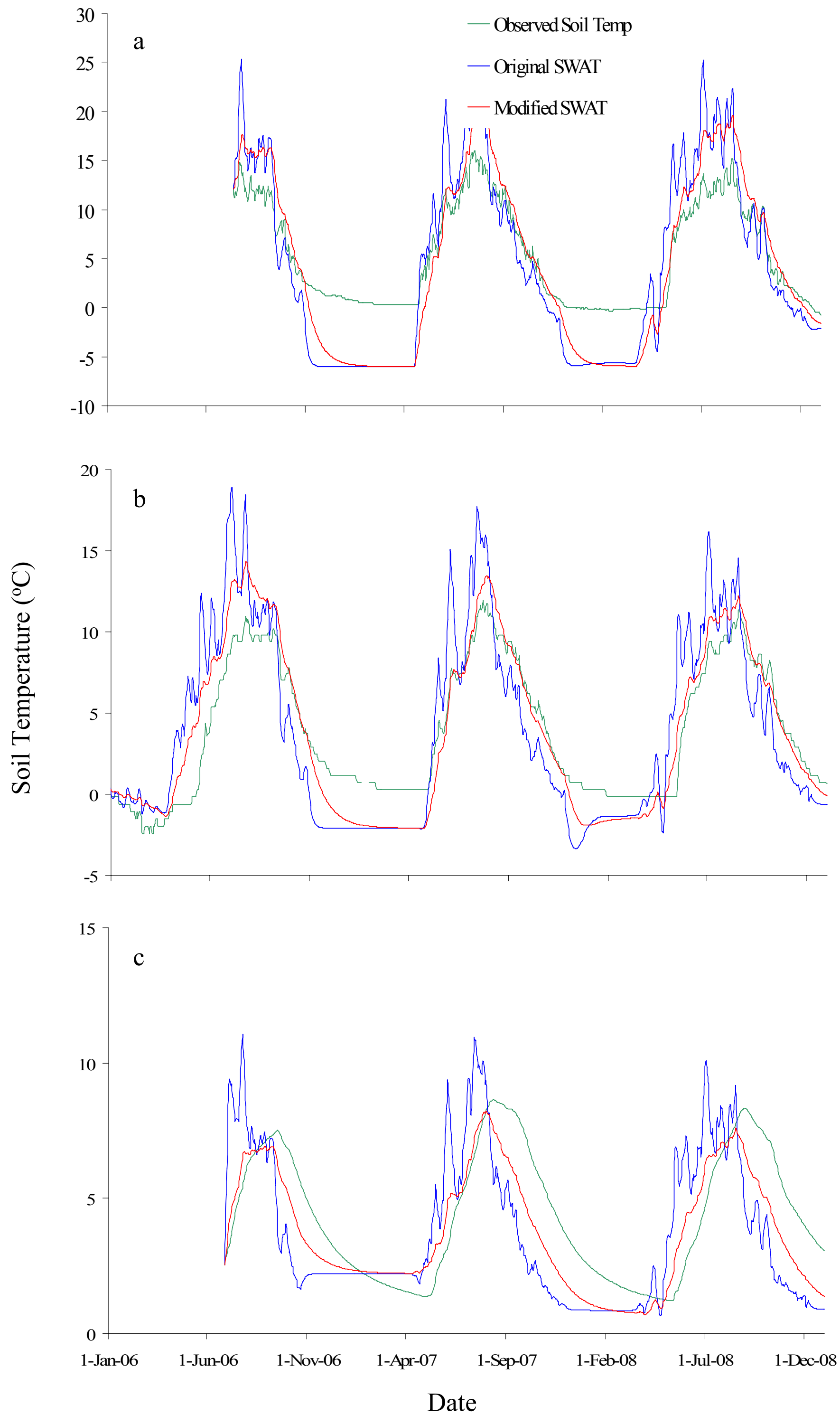

Figure 4-4: Observed data and original and modified SWAT model outputs for the (a) burned site at 0.1 m, (b) conifer site at 0.5 m and (c) wetland site at 1.0 m. Results for each site and depth are located in Appendix. Note the differences in the y-axis scale.

Many factors were considered to address the problems in the original SWAT soil temperature submodel, including: albedo, soil bulk density, depth of the soil profile, vegetation biomass and the lag coefficient. The default albedo in the original SWAT model of 0.8 for a snow-covered surface is not representative of northern disturbed or undisturbed forested sites, because it does not account for vegetation in treed sites and smaller shrubs emerging from the snowpack in open sites. Reported albedo values for this region of the boreal forest are 0.43 for treed sites (conifer, deciduous and wetland) and to 0.66 for open sites (harvested and burned) (URS Corporation, 2002), which are much lower than the value provided by the SWAT model. As a result, the first modification to the SWAT model was to alter the albedo value for a snow-covered surface based on cover type.

The sensitivity analysis found that soil bulk density, depth of soil profile and vegetation biomass cover did not significantly affect SWAT model results, as long as the parameters were varied within a representative range of values for this region. However, the lag coefficient (recall fixed at 0.8 in original SWAT model; see equation 4-1) was identified as a sensitive parameter. The second modification to the SWAT model was to increase the lag coefficient for all treatments and depths (Table 4-3; Figure 4-4; Appendix). The lag coefficient that provided the best overall fit for each scenario was selected (Figure 4-5). In all cases, the lag coefficient increased from the 0.1 m depth to the 0.5 and 1.0 m depths, which demonstrates that the previous day's soil temperature was more important at deeper soil depths (Table 4-3). This finding, along with an increasing time-shift with increasing soil depth, shows that the overlying soil layer also acts as an insulator at deeper depths. By increasing lag coefficients, maximum differences between predicted and observed soil temperatures were reduced by half in the summer (4.9 ± 0.9^{o}C, mean $\pm$ *SE*) and by 30% in the winter (3.7 ± 0.4^{o}C) (Table 4-3). The time-shift present in the spring and autumn transition periods was also reduced (Figure 4-4; Appendix). By making both changes to the model (decreasing the albedo and increasing the lag coefficient), the overall efficiency of the modified SWAT model increased; *MAE* decreased and ranged from 0.7 (wetlands) to 3.4 (burned) (Table 4-3). Still, the modified soil temperature model predicted soil temperatures that were cooler than observed during the winter months when snow was present.

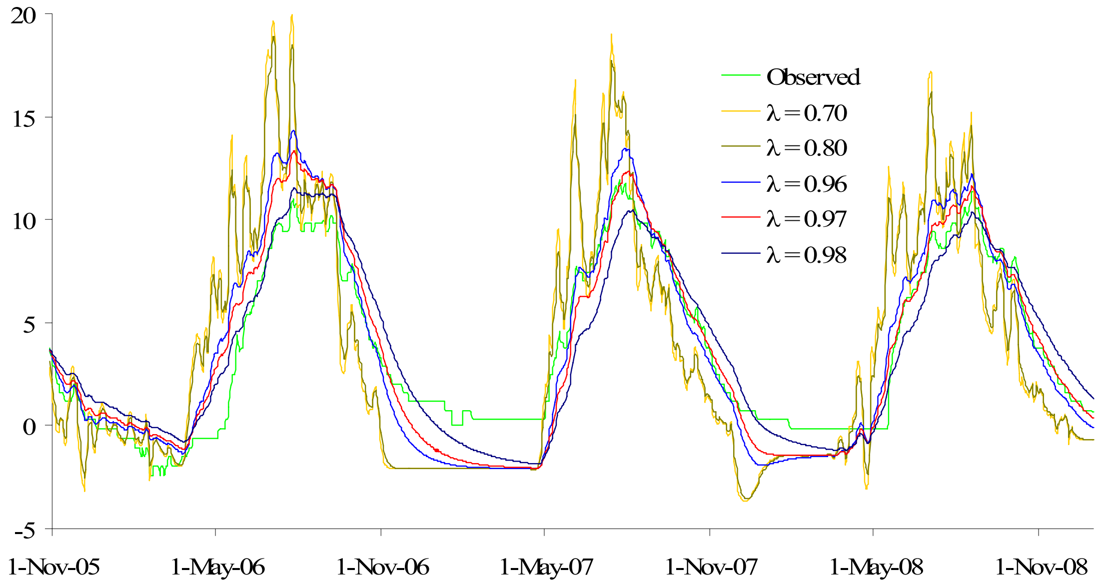

Figure 4-5: Example of sensitivity analysis for the lag coefficient (λ) for a conifer sites at 0.5 m depth.

Following modifications to the model, predicted soil temperatures from January to March of 2006 at the conifer site (Figure 4-4b) were only slightly colder than observed in comparison to the same periods in 2007 and 2008. This is because there was very little snow on the ground in the winter of 2005/2006 and the modified SWAT model was more capable of accurately modelling soil temperatures. This emphasizes the importance of considering the insulation effect of deep snowpacks (> 20 cm). Although modifications to the snow albedo and lag coefficient increased the efficiency of the model throughout the spring, summer and autumn, observed soil temperatures during winter months were still generally warmer than predicted.

4.9 CONCLUSION

The comprehensive soil temperature dataset from the FORWARD study sites provided an opportunity to contribute to the reworking of SWAT for cold climates. On the Boreal Plain, the presence of snow drastically decreased the cooling effect of air temperature on soil temperatures in autumn and winter, and from the literature one can surmise that snow also decreases the warming effect of air temperature on soil temperature in early spring. The insulation from heat transfer provided by a relatively deep snowpack is not reflected in the original SWAT model. To adequately model forested sites in northern climates with SWAT, modelling efforts need to incorporate the insulating effect of snow. By changing the lag coefficient, the modified SWAT model was able to effectively model soil temperatures in the spring, summer and autumn seasons in selected sites on the Boreal Plain where snow was absent. However, to effectively simulate soil temperatures under a snowpack during winter months, further modifications to the model must be incorporated. Recommended modifications consist of testing the equation presently in SWAT that considers the insulating effect of snow and, if this proves insufficient, to include another parameter to account for deep and persistent snowpacks in northern climates.

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CHAPTER 5: CONCLUSIONS AND RECOMMENDATIONS

This study was a first attempt to identify factors influencing soil temperatures in the FORWARD study area on the Boreal Plain. Further, to the best of my knowledge, this was the first study that dealt with testing the soil temperature submodel within SWAT in the cold region of the Boreal Plain. Previous studies south of the study area focussed on hydrological aspects of SWAT and identified changes that were required in the snowmelt and subsurface hydrology routines, but none of these explicitly looked at the soil temperature modelling subroutine.

A comprehensive review of the literature identified differences in soil characteristics, vegetation biomass and climate between the study site and the Blackland Prairies of southeast Texas, where SWAT was developed. The upland study sites had fine-grained soils and had complex overlapping vegetation in treed sites (conifer, deciduous and wetland) and were relatively open in disturbed sites (burned and harvested) compared to the monocultures of agricultural sites that have less variation in the amount of biomass from one location to another. The area was also dryer (578 *versus* 910 mm mean annual precipitation) and had a much cooler climate (2.6 *versus* 18.8^{o}C mean annual air temperature) when compared to southeast Texas. Perhaps the most important difference was the presence of a deep and persistent snowpack at the study sites, which was absent in southeast Texas (22 vs. 2.5 cm mean January snow depth).

The complete dataset (three upland burned, three upland harvested, three upland conifer, three upland deciduous sites and two wetland sites) was used to quantify the effects of air temperature, soil moisture and snow depth on soil temperatures. In the absence of snow for shallow soil depths (0.1 m), air temperature was identified as the most important parameter explaining approximately 70% of the variation in soil temperature for upland and wetland sites. When snow was present, soil moisture was more important at the 0.1 m depth. At a deeper soil depth (0.5 m), the overlying soil layer and in the presence of snow, a deep snowpack, provided increased insulation from the cold air and soil moisture was identified as the most important parameter explaining from 63 to 91% of the variation in soil temperatures for upland and wetland. Further, it should be noted that no distinction was observed between burned and harvested sites and they were pooled into a set of “disturbed” sites. Conifer and deciduous sites also behaved similarly at the 0.1 m depth (they were pooled as “treed” sites), but their effects on soil temperature at the 0.5 m depth were different. The presence of snow was significant when identifying and quantifying factors that affect soil temperatures; although snow depth was relatively unimportant using the multiple linear regression approach.

Snow was also important when testing the SWAT soil temperature submodel. To test the submodel, one representative site of each treatment type (upland burned, harvested, conifer and deciduous forest, and wetland) was used. Two modifications were made to the model, the albedo for a snow-covered surface was decreased to provide a more representative value for the study area and the lag coefficient (that controls the influence of the previous day’s soil temperature on the current day) was increased. With these modifications, SWAT effectively predicted soil temperatures during the spring, summer and autumn seasons. However, the presence of snow in the study area (as identified in the literature review and shown to be an important factor influencing soil temperatures in Chapter 3) was reflected in the model output. In the presence of a shallow snowpack (< 20 cm) during winter months, as was observed from 1 January to approximately 31 March 2006, SWAT adequately modelled soil temperatures. However, the following winters, which had deep and persistent snowpacks (> 20 cm), SWAT persistently underestimated soil temperatures relative to observed data. To address this problem when modelling soil temperatures in cold climates, further work is needed to determine the effect of a deep and persistent snowpack. It is recommended that the equation used by SWAT to calculate the insulation effect of snow on soils be analysed. Previous studies have changed snow-related and subsurface hydrology equations, consequently modifying the snow insulation equation may not be sufficient to address the problem. In the event that the original model equation cannot be modified, another function should be incorporated within SWAT that considers the non-linearity of snow as an insulator of Boreal Plain soils.

Further recommendations include: analysing the soil temperature data to identify trends on the Boreal Plain in the late-winter and spring seasons, measuring snow depth and snow density more frequently throughout the winter season and collecting long-term data at these sites, spanning 10-30 years. These would enable proper calibration and validation of the SWAT soil temperature submodel and could also allow for the development of tools for the forestry industry that could identify when frozen soils were present.

APPENDIX

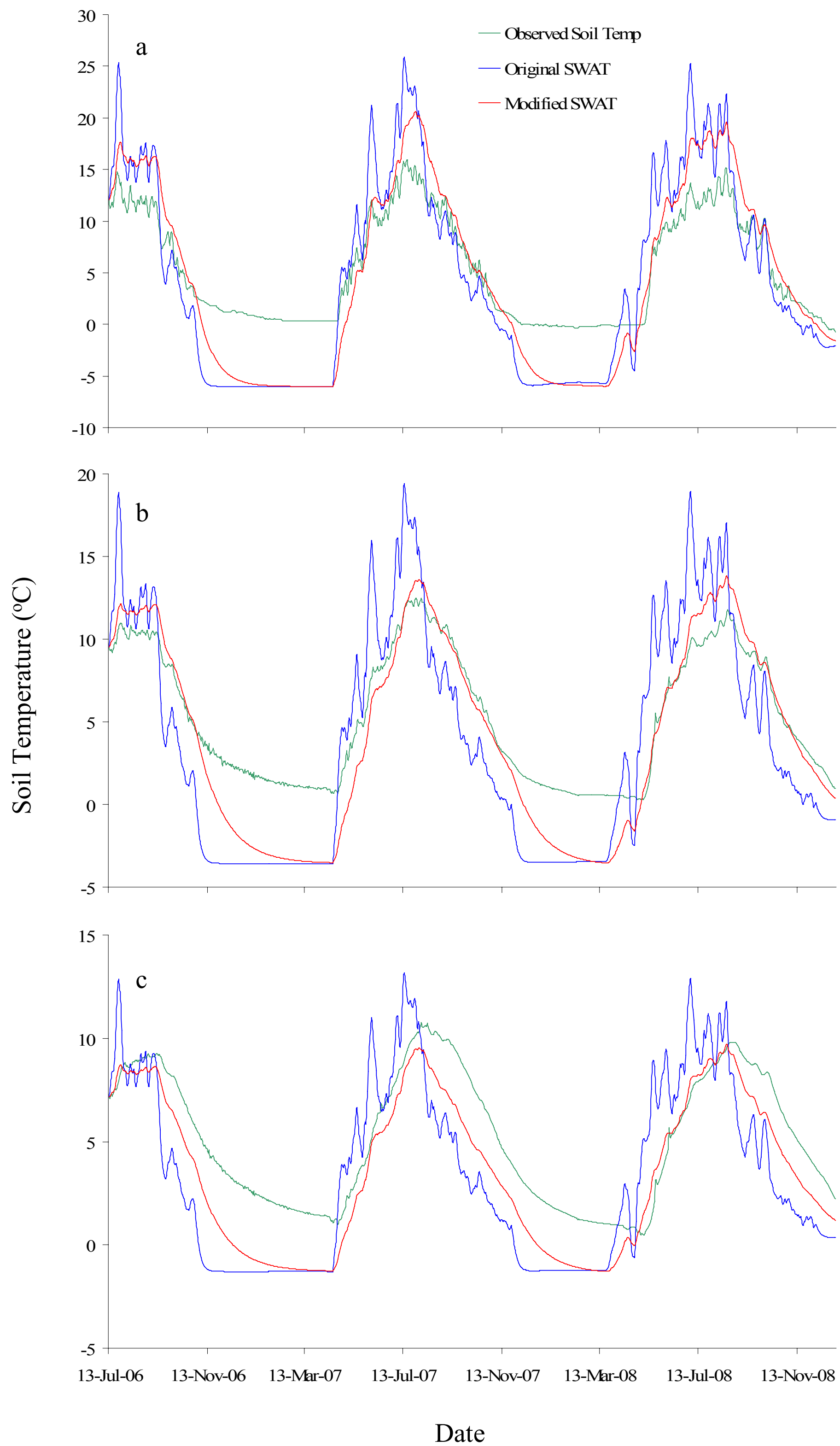

Figure A-1: Observed data and original and modified SWAT model outputs for the burned site at (a) 0.1 m, (b) 0.5 m and (c) 1.0 m. Note the difference in the y-axis scale.

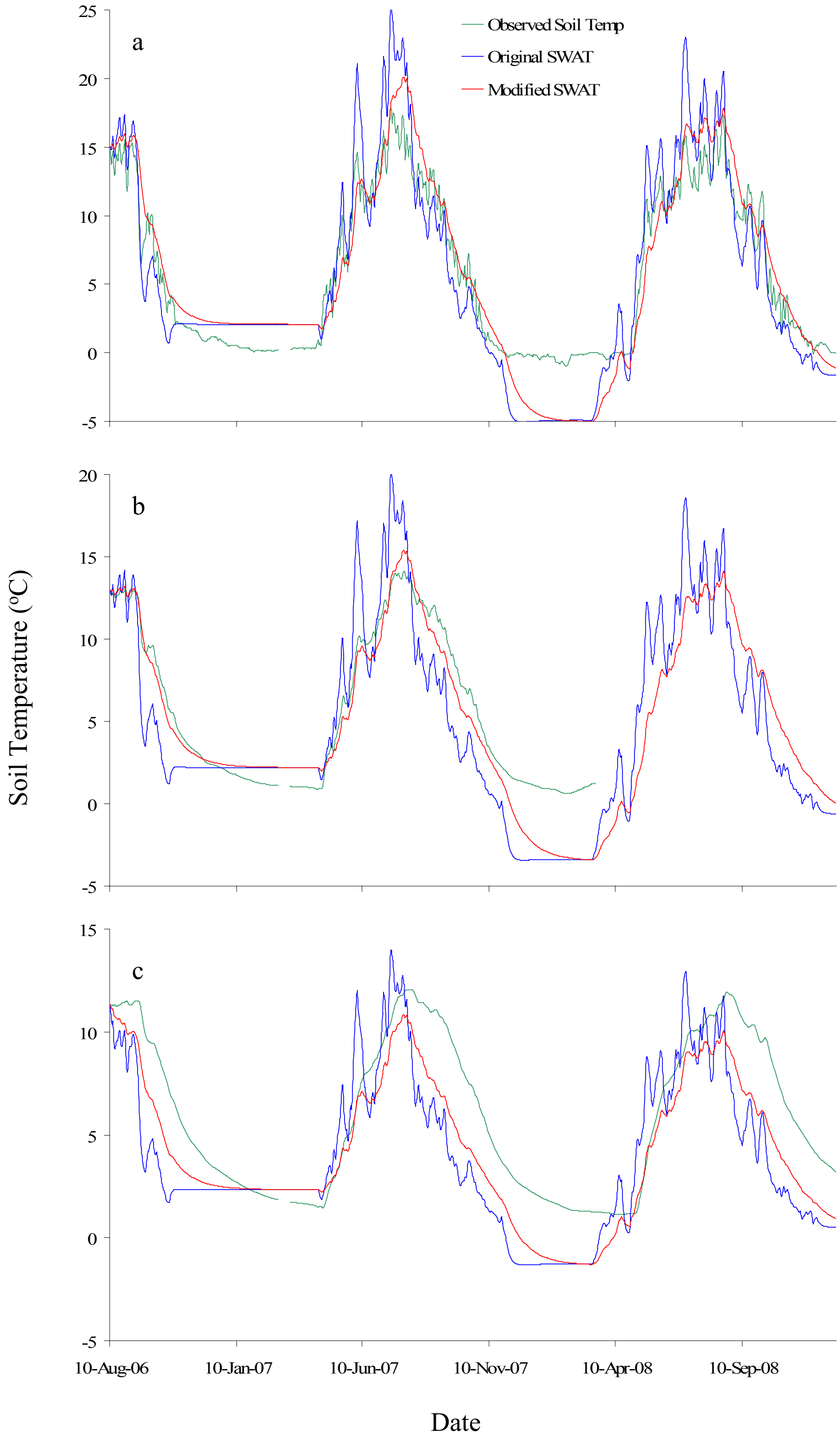

Figure A-2: Observed data and original and modified SWAT model outputs for the harvested site at (a) 0.1 m, (b) 0.5 m and (c) 1.0 m. Note the difference in the y-axis scale.

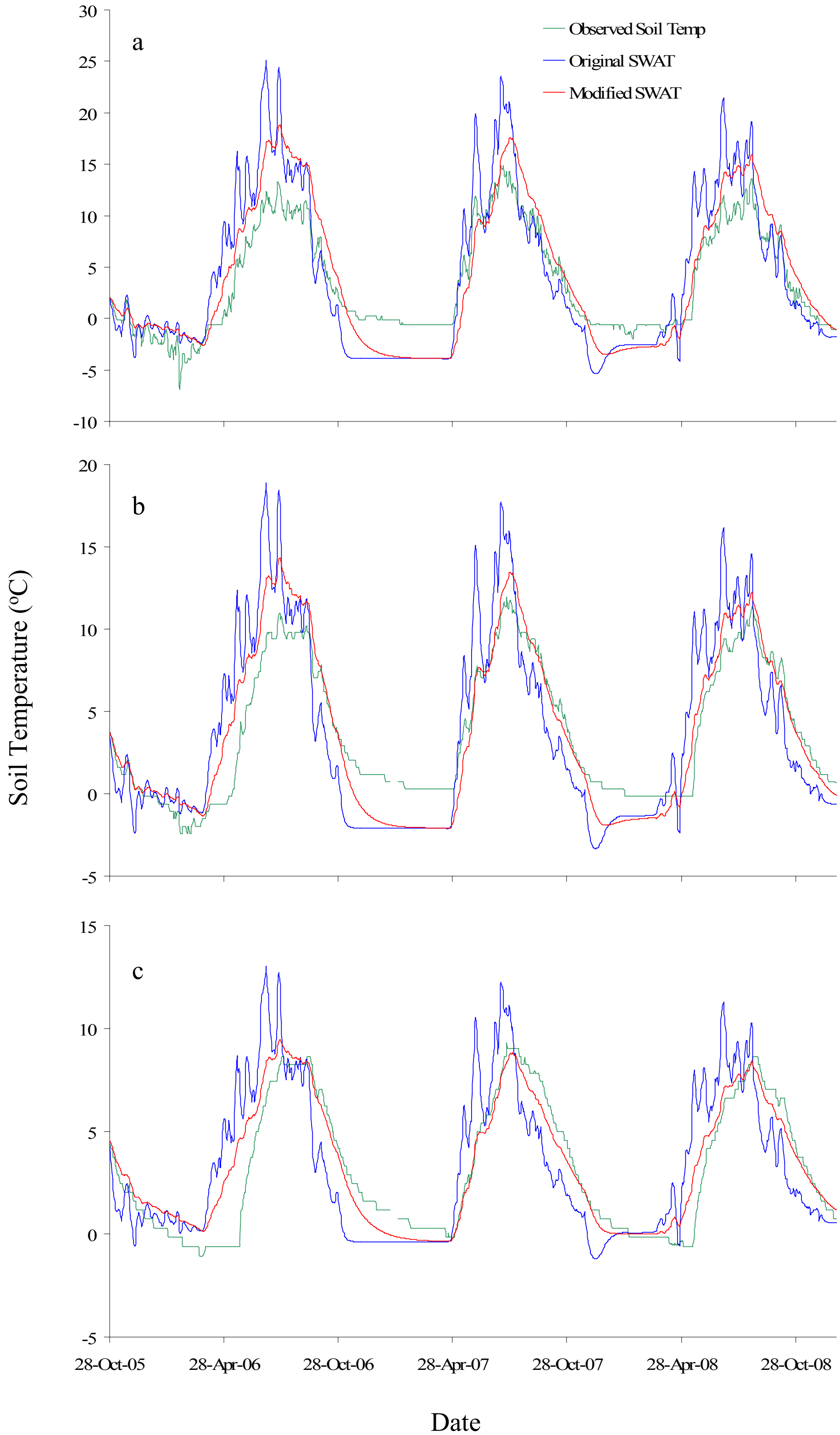

Figure A-3: Observed data and original and modified SWAT model outputs for the conifer site at (a) 0.1 m, (b) 0.5 m and (c) 1.0 m. Note the difference in the y-axis scale.

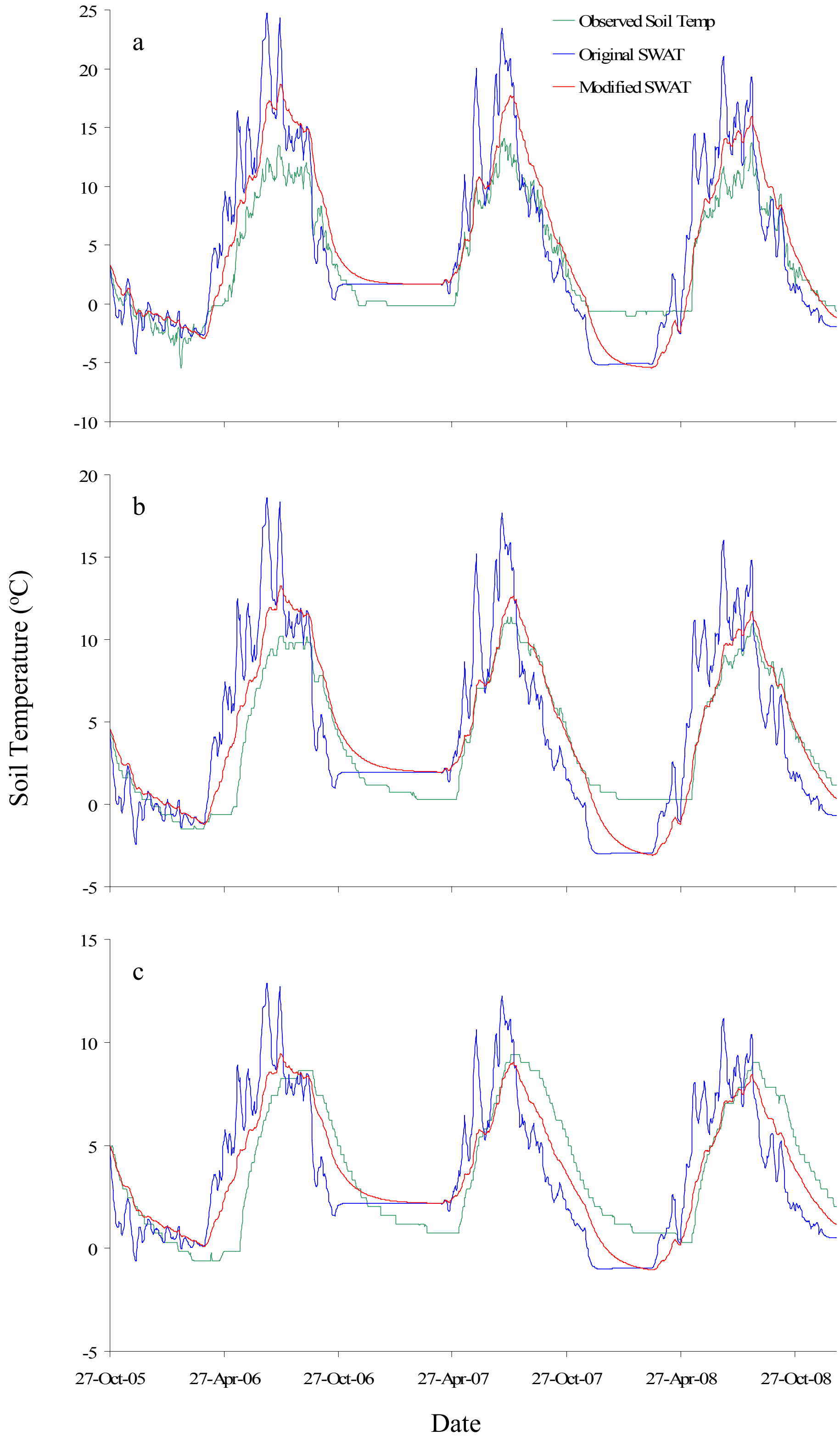

Figure A-4: Observed data and original and modified SWAT model outputs for the deciduous site at (a) 0.1 m, (b) 0.5 m and (c) 1.0 m. Note the difference in the y-axis scale.

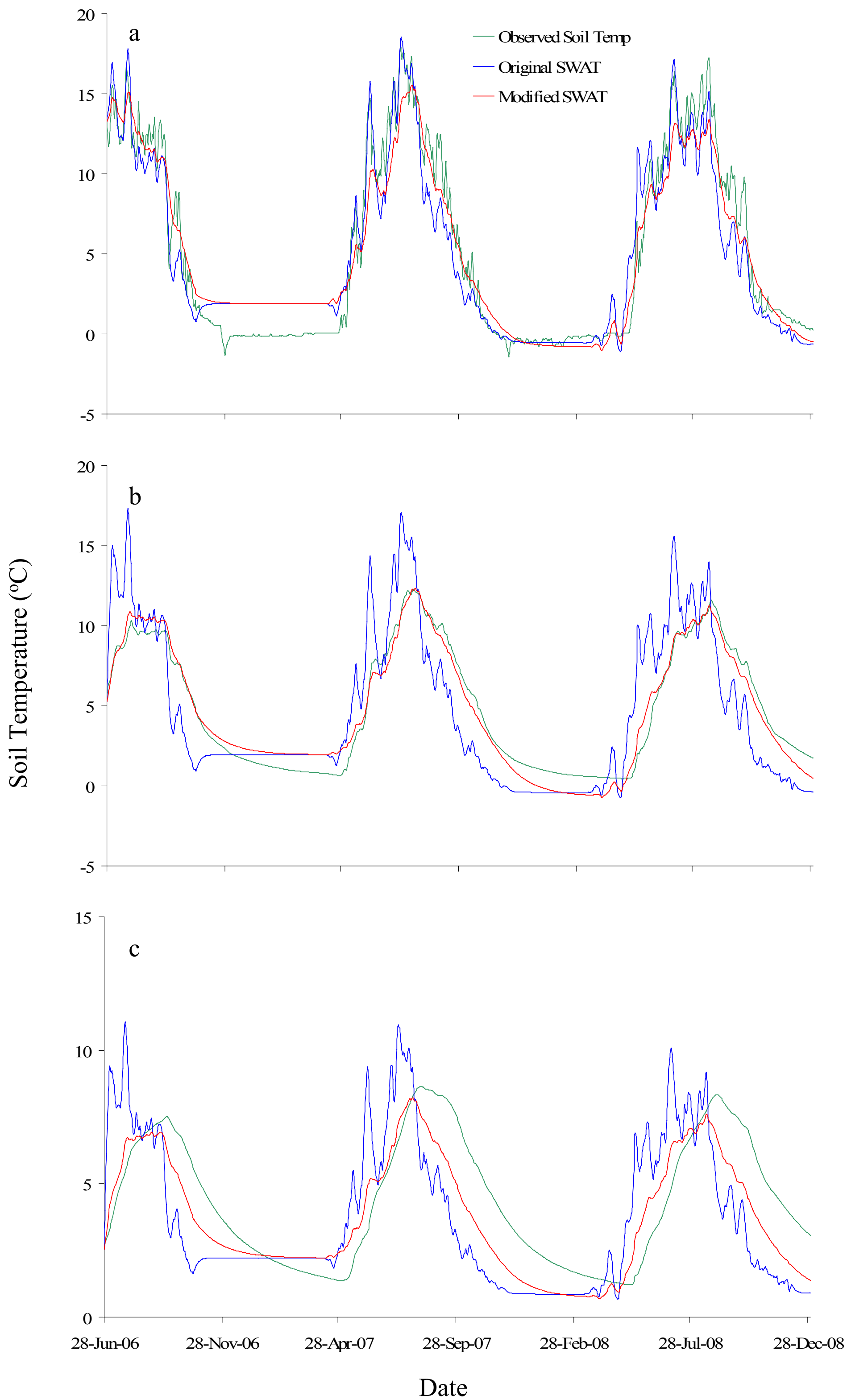

Figure A-5: Observed data and original and modified SWAT model outputs for the wetland site at (a) 0.1 m, (b) 0.5 m and (c) 1.0 m. Note the difference in the y-axis scale.