PERSISTENCE OF SOIL MOISTURE IN THE CARIBOO MOUNTAINS, BRITISH COLUMBIA

By

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Abstract

In the Cariboo Mountains of British Columbia, nearly 50% of the annual precipitation falls as snow. Over the winter, snow accumulates, is redistributed and metamorphoses until temperatures warm and the spring melt begins. In about 2-3 weeks, the water stored in the seasonal snowpack is released at the surface to infiltrate the soil and runoff into local streams and rivers. Soil moisture is an integral component of the hydrologic cycle and knowing the contribution and persistence of soil moisture from snowmelt is important to understanding the local hydrology.

Here are presented hydrometeorological data spanning from 17 July 2008 to 14 July 2009 from a cut-block site in the Mt. Tom Forest Management Area, south east of Prince George, British Columbia. Data from this period are compared to the 1971-2000 climate normals at Barkerville. The winter of 2008-2009 saw higher than average snowfall with average monthly snowfall of 58 cm compared to 44 cm for 1971 to 2000. Maximum accumulation reached 156 cm on 2 April 2009. Temperatures for the 2008 to 2009 season were cooler than average (1.9°C) at both Mt. Tom (0.4°C) and Barkerville (0.7°C). There are no long term data available for Mt. Tom so climate comparisons are based on data collected at Barkerville.

Persistence of soil moisture has implications for plant and tree growth as well as for the length of the work season for industry. With increased industrial activity and the current mountain pine beetle problem in the Cariboo region, it is likely that this area will undergo significant deforestation in years to come. Coupled with a wetter climate, as projected by climate change analysts, the capacity of the soil to retain water could become more important when trying to reduce impacts of industry, such as erosion and compaction of soil.

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Chapter 1: Introduction

1.1 Motivation

Snow and soils are intrinsically linked to each other and the climate system. In regions where snow cover exists on a seasonal basis snow has the potential to prolong cooler air temperatures and recharge soil moisture in the spring. In the mid-latitudes, the link between snow and soil moisture is especially pronounced, owing to climate memory imposed on the soil by anomalies in snow cover extent and duration (e.g. Koster and Suarez 2001, Shinoda 2001).

In spring, snowmelt contributes to surface runoff, increased flows in rivers and tributaries and soil moisture recharge. The ability of the soil to retain and transport this water is of importance to hydrologic modelling and climate change research. It has been demonstrated that snow cover and subsequent soil moisture conditions have the potential to set up large-scale teleconnections between areas that receive annual snowfall and other remote regions that may not (e.g. Hahn and Shukla 1976, Meehl 1994). Studies on this topic typically rely on data from modelling studies. These data are only as reliable as the models used to generate them and there is a need to validate the models with ground based data.

In the Cariboo Mountains of British Columbia, snow cover is highly variable both spatially and temporally. With an average 50% of the annual precipitation in this region falling as snow, the contribution of snowmelt to rivers, soil moisture and groundwater recharge is of great concern for planning and management of water resources. Understanding how climate change and land use changes may alter water storage capabilities in and around the region is important to decisions that will affect the local economy and management practices. With the Mountain Pine Beetle (MPB) epidemic, it is likely that the local groundwater levels will rise as there is less demand for water from local flora (Helie et al. 2005). In turn, increased soil moisture levels may have negative impacts such as soil erosion and flooding.

Precipitation is an important input of water in any terrestrial ecosystem. In the Cariboo Mountains snow comprises a significant portion of this input. The snowmelt period begins as early as late March and ends in mid to late June, during which a large volume of water is released to the surface. The capacity of the soil to retain moisture in the spring and through the summer has ecological and social implications. Understanding the connection between snowmelt and subsequent soil moisture conditions can aid in forecasting water supplies for the upcoming spring/summer as well as help assess the possibility of natural disasters.

As the snowmelt signal propagates through the soil it can be traced via in-situ measurements of volumetric soil water content. These data allow us to characterize how snowmelt contributes to soil moisture recharge in spring. Tracking the response of the soil to different precipitation events gives insight into the capacity of an area to accept, store and use water from the soil. The meteorological station set up at the Northern Hydrometeorology Group's (NHG) research site in the Mt. Tom Forest Management Area is implemented with three soil moisture probes at varying depths in the soil: 10 cm, 20 cm and 35 cm. The data collected will be used to characterize the propagation of the snowmelt signal in the soils of a sub-alpine spruce ecosystem.

The purpose of this study is to characterize the persistence of the snowmelt signal in soils of the Cariboo Mountain. The data were collected from only one point in space so the main focus of this study is on methods used to glean temporal information from the soil moisture data collected at three depths. We are especially interested in determining how long moisture from the snowmelt period contributes to the soil moisture budget and whether or not the signal can be detected later in the season. We use daily average soil moisture and precipitation data to assess the persistence of moisture from precipitation events, assuming that precipitation is a stochastic event. We then assess the persistence of the snowmelt signal using autocorrelation functions to characterize how the signal propagates through the soil. These methods are described in greater detail in the data and methods section of this thesis.

1.2 Objectives

The objectives of this study are:

1) To observe the temporal variability of soil moisture at the Mt. Tom site (17 July 2008 to 14 July 2009) using three Campbell Scientific CS616 Water Content Reflectometers at 10 cm, 20 cm and 35 cm depth;
2) To characterize the contribution of snowmelt to soil moisture recharge using data collected on-site;
3) To determine the persistence, if any, of water from snowmelt in the soil profile at Mt. Tom.

The first objective involves characterizing the response of soil to precipitation events versus its response to the influx of water input from snowmelt. Comparing daily average soil moisture values during the fall freeze-up, or antecedent period, to those obtained during the snowmelt period will allow us to characterize whether or not there is any difference between the memory of the soil to individual precipitation events and snowmelt. Looking closer at the relationships between soil moisture and environmental factors such as air temperature, soil temperature, snow depth and precipitation will help us assess what factors affect soil moisture most significantly, highlighting important relationships. It is hypothesized that we shall see significant relationships between soil moisture and precipitation events in the fall freeze-up period and not in the spring melt period. It is also hypothesized that there will be significant relationships between soil moisture, air temperature and snow depth in the spring melt period.

The second objective involves exploring the data obtained to see how snowmelt contributes to soil moisture recharge. In some areas, ice within the soil may impede soil moisture recharge. Slope and surface conditions could lead to surface runoff. Although we will not be able to quantify the contribution as a whole to the area, it is possible for us to see whether or not there is an influx of moisture during the snowmelt season. It is hypothesized that we will see an increase of soil moisture corresponding to higher average daily air temperatures and that soil moisture will be higher than average during the spring melt period.

The persistence of soil moisture will be characterized by autocorrelation which will determine if there is any lag in the response of soil moisture to moisture input from precipitation events and snowmelt with depth. We shall examine the similarity of soil moisture at each depth using cross-correlation analysis.

1.3 Outline

An extensive literature review in Chapter 2 will explore the properties of snow and soils that dictate the relationship between them and local climate. There will be a brief review of the climate system including a look at the hydrological cycle, surface radiation budget and energy balance. These concepts are integral to understanding the interactions between snow, soils and climate. Snow contributes to precipitation recycling and climate memory, both of which will be discussed in the literature review. Persistent anomalies in snow cover can set up large-scale teleconnections, with conditions in one area affecting remote regions. This is the last topic of the literature review.

Chapter 3 provides a detailed look at the study site, the Mt. Tom Forest Management Area. This chapter gives a description of the local climate based on data collected from the site and compared to data from a long term station located near Barkerville. Information about the local ecosystem, soils and climate will be included here along with maps, photographs of the station and soils, and other points of interest.

Data and Methods are discussed in Chapter 4. All field methods will be described in detail along with any rationalization for sampling methods. Detailed explanation and description of analyses performed will be given in this chapter. Scripts used to run analyses in statistical software programs will be mentioned here.

Results and Discussion (Chapters 5 and 6) will focus on the relationship of different physical and meteorological variables to the propagation of the snowmelt signal. Data from the late-summer and fall freeze-up are compared to data from the snowmelt period. Significant correlations and trends between measured variables are presented and discussed. A summary and conclusion will reiterate the main findings and any take home messages.

Chapter 2: Background and Literature Review

2.1 Climate

All components of the environment interact to produce climatic conditions that are unique to different ecosystems. The main components of the climate system are the lithosphere (rocks, soils, and minerals), the hydrosphere (rivers, lakes, and oceans), the atmosphere (air, trace gases, and aerosols), the biosphere (plants, animals, and humans) and the cryosphere (snow, ice, and glaciers) (Figure 2.1). Each component of this system has a distinct climate regulated by different surface properties and characteristics. Depending on the environment, different factors will limit precipitation, heat exchange and the radiation budget. For example, in wet environments, precipitation is likely to be limited by the radiation budget and the advent of convective processes. In drier environments, precipitation may be limited by the availability of water from the atmosphere and soil. Knowing the relative importance of each climate variable for different ecosystems or regions is important in making informed projections about future weather and climate.

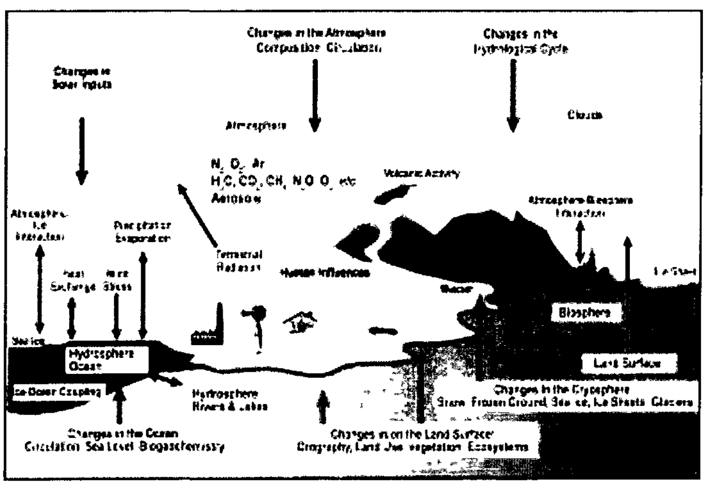

Figure 2.1- Schematic of components of climate system, IPCC 2001

Mid-latitude climates represent a transition zone between tropical environments and polar environments. The climate in mid-latitude regions is largely dependent on the soil-plant-atmosphere feedback (Kim and Wang 2007). Evapotranspiration is the process linking the soil, plants and the atmosphere, and it is largely dependent on soil moisture and the energy budget. Convective processes, dependent on vertical thermal gradients, entrain water from the surface to be condensed into clouds and precipitation at higher levels in the atmosphere. It is through these convective processes that soil moisture has the greatest effect on precipitation. The controlling factors for convective processes change depending on the scale to be considered (Anderson et al. 2003). With respect to modelling, the response of the surface to atmospheric processes, at smaller scales (ecosystem), evapotranspiration is limited by water availability and vegetation whereas at larger scales (regional) the surface energy budget becomes more important (Anderson et al. 2003).

2.2 The Hydrological Cycle

The hydrological cycle describes how water is recycled through the earth system and the processes that are involved in moving water from one place to another. Water exists in the atmosphere, in oceans, rivers, streams and lakes, in soil and aquifers, stored in glaciers and snow, and in vegetation and other living beings. As it cycles through each component of the earth system it either uses or releases energy as it changes from one state to another.

In the Cariboo Mountains, snow covers the surface for several months every year, generally from mid to late November to late May-early June, representing a seasonal store of water that replenishes soil moisture, stream and river discharge in the spring (MacLeod and Déry

2007; Tong et al. 2009). As the snow melts, some water infiltrates the soil and percolates through to replenish ground water. Some water is stored on the surface in lakes and rivers. Water on the surface or stored as soil moisture is available for evaporation. This becomes the link between the radiation budget, energy budget and water cycle and hence is accounted for in the surface storage equation.

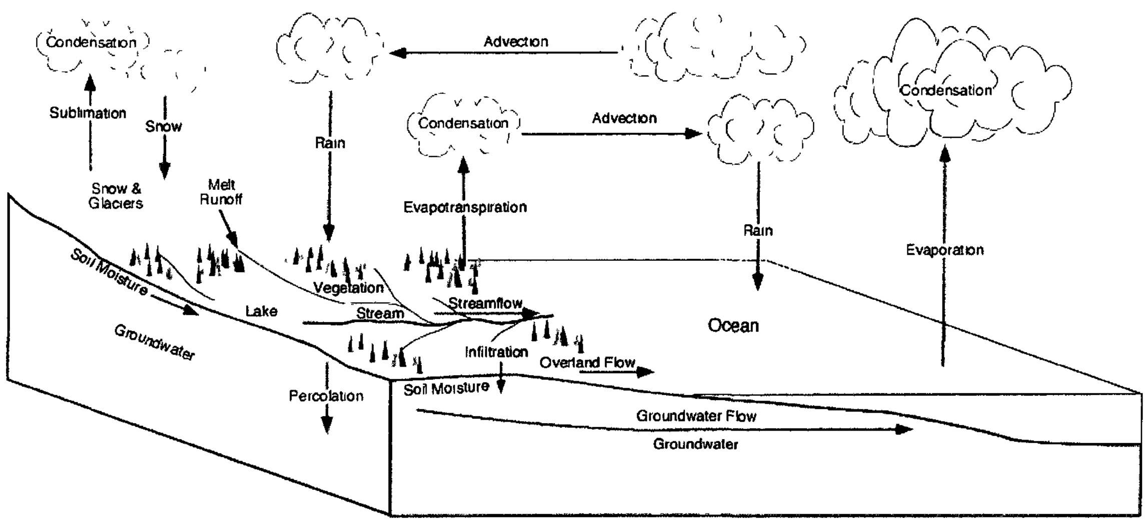

Figure 2.2- Schematic diagram of the hydrological cycle, IPCC 2001

Surface storage can be described by the following equation:

$$\Delta S = P - Q - ET \qquad (1)$$

where ΔS (mm) is a change in surface storage (including soil moisture), P (mm) is precipitation, Q (mm) is runoff and ET (mm) is evapotranspiration (Bowling et al. 2003; Déry et al. 2005a). A more detailed version of this equation would incorporate properties of the surface such as water holding capacity and the average degree of saturation as seen in the work of Koster and Suarez (2001) as well as track the phase of water (Oke 1987).

2.3 The Surface Radiation Budget

The surface energy balance can be separated into two components: the surface radiation budget and the energy budget. The surface radiation budget depends on the amount of incoming solar radiation $K\downarrow$ (W m^{-2}) and the amount that is reflected from the surface $K\uparrow$ (W m^{-2}) back into space, as well as the amount of incoming $L\downarrow$ (W m^{-2}) and outgoing longwave $L\uparrow$ (W m^{-2}) radiation. Incoming shortwave radiation (mainly in the visible spectrum) has relatively high energy associated with its emitting source, the sun. A portion of the incoming shortwave radiation is absorbed by the surface and then emitted as longwave radiation in the infrared spectrum. The radiation absorbed is largely dependent on the reflectivity or albedo (α) of the surface, which is the ratio of reflected shortwave radiation to incoming shortwave radiation. The surface radiation balance may be then expressed as:

$$\begin{aligned} Q^* &= (K\downarrow - K\uparrow)+(L\downarrow - L\uparrow) \\ &= K\downarrow (1-\alpha) + L\downarrow - L\uparrow \\ &= K^* - L^* \qquad (2) \end{aligned}$$

where K^* is net shortwave radiation and L^* is net longwave radiation. The net excess or deficit of radiation (Q^*) is used to determine the turbulent heat fluxes and hence, the energy available to heat or cool the surface (Q_G).

2.4 The Energy Balance

Energy is partitioned, on a global scale. The source of energy is the sun, emitting shortwave and high energy radiation. A portion of this radiation is reflected right back to space by clouds, particles in the atmosphere or the land surface itself. Some is absorbed to heat the atmosphere, clouds, ocean and land surface. Once absorbed, radiation from the sun can be radiated back to space as longwave energy as described above. The connection between water and energy via the latent heat flux is accounted for in this budget with 23% of the incoming solar energy that is absorbed by land and oceans is carried to clouds and atmosphere via latent heat in water vapour. When this vapour condenses, latent heat is released to heat the atmosphere.

The surface energy balance for a bare surface can be expressed as:

$$Q^{*}=Q_H + Q_E + Q_G + Q_M + \Delta Q_S \qquad (3)$$

where Q^* (W m^{-2}) is the net radiation, Q_H (W m^{-2}) is sensible heat, Q_E (W m^{-2}) is latent heat, Q_G (W m^{-2}) is heat being stored in or lost from the surface, Q_M (W m^{-2}) is heat available for snowmelt and ΔQ_S (W m^{-2}) is the heat stored within the surface, as is the case with a snow covered surface (Oke 1987). A snowpack at 0°C is said to be ripe. The addition of energy to a ripe snowpack will result in snowmelt.

A useful value in describing the partitioning of energy into the sensible and latent heat fluxes is the dimensionless Bowen Ratio β (Oke 1987):

$$\beta = \frac{Q_H}{Q_E} \qquad (4)$$

Values less than unity signify that more energy is being transferred to the atmosphere as latent heat rather than as sensible heat. Values greater than one imply that the sensible heat flux is dominant. The Bowen Ratio can be useful in determining the climate of an area based on latent and sensible heat exchange. For example, in very dry, hot environments, the sensible heat flux dominates year round, leading to higher values of β in desert areas. Table 2.1 lists β values for different environments.

Table 2.1- Bowen ratio for different environments, Oke 1987.

Bowen Ratio	Environment
0.1	Tropical Oceans
0.1-0.3	Tropical wet jungles
0.4-0.8	Temperate forests and grassland
2.0-6.0	Semi-arid regions
>10.0	Deserts

In areas that receive annual snowfall, the Bowen Ratio can change significantly from winter to spring/summer. A snow cover reflects a large fraction of incoming solar radiation and is decoupled from the ground below, significantly reducing the sensible heat flux. The latent heat flux depends on the thermal gradient between the surface and the air above and absolute humidity tends to be low owing to cooler winter temperatures despite the readily available water stored in snow. Sensible and latent heat fluxes are then dependent not only on the temperature and humidity of an environment but also on properties of the surface that can change drastically from one season to the next, where seasonality exists.

2.5 Properties of Snow vs. Soil

Snow has an albedo between 0.45 and 0.95 and therefore reflects considerably more radiation from the surface. Soil generally has an albedo ranging between 0.05 to 0.4 so a larger portion of radiation is absorbed at the surface, heating the ground and above surface air temperatures (Oke 1987). The albedo of snow is highly variable in time and space, constantly changing as the result of wind, temperature fluctuations, and deposition of particulates onto the surface as well as other disturbances. As the snow metamorphoses and is redistributed, its properties change. Soil is not subject to the same metamorphic processes as snow, and tends to have a more constant albedo throughout the year. Areas that have seasonal snowfall will undergo significant fluctuations in surface energy conditions from one season to the next. Vegetation also changes the albedo of the surface however, the change from a bare soil surface to a surface covered by vegetation is less pronounced as from a vegetated surface to a snow covered surface. Any disturbance or interruption to these processes will affect local climates and quite possibly affect vegetative and other biological interactions.

The heat capacity (amount of heat required to raise a unit volume of a substance by 1°C) and thermal conductivity (ability of a substance to conduct heat) of snow are low as compared to soil. The heat capacity of soil increases as water infiltrates so that wetter soils can retain or lose more heat than drier soils in the same environment. Retention of heat within the soil would prolong warmer temperatures via land-atmosphere energy exchange processes. Snow does not have a similar capacity to force warming in near surface temperatures. It is a reservoir for heat but is limited by the change of phase from ice to water that occurs at 0°C. At this point, the snowpack begins to melt as energy is used to convert snow to water.

Heat capacity limits the thermal regime of the surface and air temperatures above it. Consider a permafrost environment where the ground is perennially frozen. That portion of the soil that freezes and thaws on an annual cycle, the active layer, would require a significant amount of cooling to freeze up in the fall. With warmer temperatures in the north, it is expected that the active layer will deepen and that there will be a significant loss or degradation of permafrost (IPCC 2007). The loss of permafrost would result in warmer soil temperatures leading to warmer air temperatures via the sensible heat flux and evapotranspiration. A positive feedback could emerge whereby warmer temperatures persist for longer periods in polar and alpine environments.

The physical characteristics of snow are highly variable in space and time. Successive precipitation events cause layering within the snowpack. Temperature changes can alter the structure of snow, deforming layers within the pack or leaving surface hoar in the right conditions. Snow is also mobile, meaning that snow particles can be transported by wind and deposited elsewhere (Déry et al. 2004). Areas prone to the accumulation of snow are somewhat predictable in space, owing to topographic controls on snow distribution (Litaor et al. 2008, Tong et al. 2009).

The snowpack is seasonal and variable from year to year. This high degree of variability is difficult to simulate in climate models and likewise difficult to predict, although its importance in our day to day lives is becoming exceedingly more important as water resources become scarce (Barnett et al. 2005). The capability of snow to transport water is largely dependent on the density of the snow pack as well as temperature, both within and around the snow pack. Thermal gradients, liquid water availability and the presence of impermeable layers (ice) govern the movement of melt water through a snow pack. In soils, water movement is generally controlled by hydraulic gradients which are limited by soil physical properties related to particle size distribution and porosity and gravity. The percentage of water remaining in the soil after it has drained freely for 2-3 days following saturation is called field capacity (Hillel 1998). Scientists use this measure to describe soil moisture conditions in time and between different areas.

2.6 Interactions between Snow, Soils and Climate

Linking the water and the energy balance is the process of evapotranspiration that releases or stores heat during the change of state of water to vapour. The equation below describes the latent heat flux associated with the change of state of water to vapour via evapotranspiration:

$$Q_E = L_v \times ET \qquad (5)$$

where L_v is the latent heat of vaporization (2.5×10^6 J kg $^{-1}$ at 0°C). Equation 5 describes the energy required to change water to vapour from a surface without snow cover. When snow is present on the surface, the energy requirements to change snow to vapour (sublimation) are greater because the latent heat of sublimation at 0°C is 2.8×10^6 J kg^{-1}. The difference explains the energy required to melt snow, the latent heat of fusion L_f, which is 0.3×10^6 J kg $^{-1}$. While snow cover is present on the surface, cooler temperatures persist above the snowpack. Once snow cover has melted away, energy that was previously used to melt snow is made available to heat the surface.

The addition or loss of water can alter the albedo, thermal conductivity and heat capacity of the surface. These properties limit convection and conduction within the boundary layer thus affecting local climatic processes. In areas that receive seasonal snow cover, the link between these two budgets will vary significantly from winter to summer as the conversion of snow to water requires less heat than the conversion of water to vapour. Properties of the surface affect energy exchange such that wetter (drier) conditions induce cooler (warmer) above surface air temperatures. Furthermore, the direct relationship between air temperature and humidity regulates the amount of water vapour available for precipitation, influencing local and regional precipitation regimes.

Generally, factors governing the climate system tend to vary latitudinally (Koster et al. 2000; Kochendorfer and Ramirez 2005; Bosilovich and Chern 2006). In northern latitudes, snow plays a major role in modulating the surface energy and water budgets (Barnett et al. 2005). A large volume of water is stored as snow and ice during fall, winter and spring. This mass, accumulated over many weeks and months, is released quickly during the melt period in spring to early-summer, thereby replenishing water stores for summer evapotranspiration and human consumption (Barnett et al. 2005; Déry et al. 2005a).

2.7 Teleconnections

It has long been recognized that dominant modes of climate persist through feedbacks and that conditions in one region may affect conditions in adjacent or even remote regions (see, for example, Hahn and Shukla 1976; Jacobs and DeBruin 1992; Meehl 1994; Pal and Eltahir 2001; Shinoda 2001; Kochendorfer and Ramirez 2005; Kim and Wang 2007). Regional circulation patterns influence the overall global circulation. In this way, the climate of one area can greatly affect the climate in a region far away. The lagged effects of anomalous conditions in one remote location on another are referred to as teleconnections (Hahn and Shukla 1976; Barnett et al.1989; Bamzai and Shukla 1999; Ye and Bao 2001; Cohen and Saito 2003; Déry et al. 2005a). Fluctuations in the dominant modes of climate, whether forced by stochastic events or prolonged atmospheric anomalies, will force changes in the circulation affecting localized and faraway regions alike.

A classic example is by Hahn and Shukla (1976) who reported that Eurasian snow cover in winter/spring is correlated with the strength of the Indian summer monsoon the following summer. Bamzai and Shukla (1999) later describe a mechanism for this relationship by which higher (lower) than normal snow cover in western Eurasia causes wetter (drier) soil conditions. Wetter (drier) soil will lead to cooler (warmer) air temperatures and a weaker (stronger) Indian monsoon rainfall (Bamzai and Shukla 1999).

The Bowen ratio is a useful concept in explaining this phenomenon. In wet conditions the Bowen ratio decreases as more energy is partitioned towards the latent heat flux, energy is used to evaporate water from the surface, and then energy is released upon condensation. In drier conditions the sensible heat flux dominates as heat is transferred directly from the atmosphere to the surface. Less water is available to vaporize so sensible heat exchange is dominant. Positive anomalies in snow cover depth and extent prolong the dominance of the latent heat flux as wetter conditions persist for longer in the spring, preventing warming of the surface (Shinoda 2001).

It has been demonstrated that positive (negative) Eurasian snow cover anomalies force cooler (warmer) than normal above surface temperatures in northern Canada (Déry et al. 2005a). Correlations exist between Eurasian snow cover and Canadian snow water equivalent and river discharge (Cohen and Saito 2003; Déry et al. 2005a). This research suggests that Eurasian snow cover may be used as a predictor of snow water equivalence (SWE) and river discharge in Canada the following spring. Pan-Arctic teleconnections have the potential to affect the climate at more southerly latitudes as temperature gradients change, altering the meridional transport of heat (Déry et al. 2005b).

The extent and lifetime of the snowpack are of great interest to understanding teleconnections. The greater the extent of snow and the longer it stays on the surface, the longer relatively cool air temperatures are able to persist. The melt period can be very short in contrast to the duration of snow cover. Moving northward, days become longer faster and radiation becomes more intense towards the summer solstice. The persistence of cooler air temperatures affects conditions downwind, setting up large-scale teleconnections between regions that are connected by atmospheric circulation. Inevitably, the entire globe is affected by atmospheric circulation such that long-term anomalies have the potential to cause persistent, large-scale perturbations in climate.

There has also been discussion of the relationship between soil moisture and the North American Monsoon (NAM) (Small 2001; Zhu et al. 2005; Grantz et al. 2007). The hypothesis investigated here is that more (less) snow in the southwestern United States and mountains of Colorado and Utah leads to more (less) soil moisture in Arizona and western New Mexico the following spring. In particular, warmer tropical Pacific sea surface temperatures (SSTs) and cooler SSTs over the Northern Pacific in the antecedent winter/spring are correlated with wetter than normal conditions over the south-western desert of the United States and drier than normal conditions in the Pacific Northwest (Grantz et al. 2007). Wetter than average antecedent soil conditions decrease heating at the surface thereby delaying the onset of the monsoon rains. More (less) soil moisture induces lower (higher) spring surface temperatures thus weakening (strengthening) the summer monsoon (Zhu et al. 2005; Grantz et al. 2007).

Climate change and land use changes are likely to alter the precipitation regime in the Cariboo Mountains. Such changes could impact boundary layer climates and possibly influence teleconnections. Wetter conditions, as projected for this area, could lead to cooler conditions as water from the surface is evaporated, storing energy as latent heat in water vapour. Wetter, cooler conditions could arise as result of this positive feedback. Evaporation is a convective process leading to more unstable conditions that favour precipitation. This leads to the next topic of discussion which describes the process of precipitation recycling.

2.8 Precipitation Recycling

Precipitation recycling is defined as the contribution of local evaporation to precipitation (Eltahir and Bras 1996). Precipitation recycling is a scale dependent process. On a global scale, it is not necessarily obvious that this process occurs; however, focusing in on a regional or basin scale, this process becomes useful in elucidating sources of water for local convection and rainfall (Eltahir and Bras 1996). Precipitation recycling is most obvious in regions where soil moisture contributes significantly to the water cycle, enhancing climate memory as it is replenished by recycled precipitation (Szeto 2002). The processes and factors affecting regional precipitation recycling include evapotranspiration, runoff, precipitation, the contribution of snow to the regional hydrologic cycle and the role of soil moisture as a source of water within a region (Szeto 2002).

Antecedent terrestrial surface conditions affect subsequent precipitation in regions that experience seasonal rainfall patterns and where there is dependence on soil moisture as a source for evapotranspiration (Meehl 1994; Koster et al. 2000; Small 2001; Bosilovich and Chern 2006; Zhu et al. 2005). In an experiment using atmospheric general circulation models it was found that knowledge of antecedent soil moisture conditions contributed significantly to the predictability of precipitation in transition zones between wet and dry climate zones (Koster et al. 2001). Predictability in these areas is higher than particularly wet or dry regions because of the seasonality of precipitation and its variability over time. Where conditions are consistently wet or dry there is little to no periodicity in the precipitation regime making trends difficult or impossible to determine.

This work compliments earlier research by Trenberth (1999) who found that moisture recycling via local evaporation was highest in regions that are least affected by moisture transport from surrounding regions. The moistening effect is greatest in June, July and August, in the Northern Hemisphere. The effects are most notable in the Northern Hemisphere between approximately 38°N and 58°N and in the Amazon basin (Trenberth 1999). Mt. Tom, in the study region, is within this latitudinal range at 54°N.

In terms of estimating the contribution of precipitation recycling to annual precipitation (also called annual precipitation ratio), studies have focused on large, continental basins in the Northern Hemisphere (see, for example, Meehl 1994; Eltahir and Bras 1996; Trenberth 1999; Szeto 2002; Bosilovich and Chern 2006). In particular, studies have focused on the Amazon Basin (Eltahir and Bras 1996; Trenberth 1999; Bosilovich and Chern 2006), the Mississippi Basin (Eltahir and Bras 1996; Trenberth 1999; Bosilovich and Chern 2006) and the Mackenzie River Basin (Szeto 2002; Bosilovich and Chern 2006).

Estimates of annual precipitation recycling ratios for the Amazon, Mississippi, Eurasia and Sahel are reported by Eltahir and Bras (1996). Earlier estimates by Molion (1975) show a high precipitation recycling ratio for the Amazon of 56% compared to later estimates that range between 25-35% (Brubaker et al. 1993; Eltahir and Bras 1994; Eltahir and Bras 1996). The estimated recycling ratio in the Mississippi region is about 24% compared to 13% in Eurasia and 35% in the Sahel (Brubaker et al. 1993). Over the Mackenzie Basin, the estimated annual recycling ratio is about 20-25% (Szeto 2002; Bosilovich and Chern 2006). This value is close to the values obtained for the Amazon and Mississippi basins, although annual evapotranspiration is lower in the Mackenzie than in either of these regions. Szeto (2002) attributes this to the unique topography of the Mackenzie Basin, which enhances the persistence of either low or high pressure systems.

These findings highlight the significance of surface characteristics (vegetation, soil physical properties, etc...) on controlling the process of evaporation. Other studies have investigated how land use changes can alter precipitation recycling, with a focus on deforestation as well as irrigation (Eltahir and Bras 1996). Deforestation decreases evapotranspiration by plants as the surface is depleted of vegetation. Irrigation increases evapotranspiration as more water is artificially supplied to support the process. Increased (decreased) evaporation influences surface humidity and temperature thus influencing local convective processes including precipitation formation. More (less) evaporation from the surface increases (decreases) the moisture in the boundary layer, leading to more (less) convective precipitation. Changes in surface conditions, especially over large scales, have the potential to disrupt these processes affecting regional precipitation patterns. Thus is the case in the area of interest where thousands of acres of forest will deteriorate as result of the MPB epidemic and deforestation associated with it (Helie et al. 2005, Winkler and Boon, 2010).

Deforestation greatly alters the albedo, denuding the surface of trees and shrubs that would use water and reflect radiation keeping the ground cooler in summer. Changes in the land surface caused by deforestation and MPB are expected to yield higher baseflows and change the timing, quantity and quality of water input to rivers and groundwater storage. It has been noted that groundwater levels are rising as the result of decreased evapotranspiration as more and more trees are affected by MPB (Winkler and Boon 2010). Convective precipitation is likely to increase as a result of deforestation so precipitation recycling may become more prevalent in this region (Helie et al. 2005). The area of interest in this study is slightly higher in elevation than the areas that are most affected by MPB; however the snow that accumulates at this site makes its way to streams and rivers at lower elevations contributing to flow and water storage throughout the watershed.

Changes in vegetation and forest canopy impact interception as well. As the forest canopy deteriorates due to MPB, less snow is intercepted in the winter. Reduced interception will increase delivery of water to the surface and groundwater.

2.9 Climate Memory

The ability of the atmosphere, land surface and water surfaces to retain characteristics of anomalous events is referred to as climate memory. Certain features of the earth system retain evidence of events for longer than others. Compare how quickly a thunderstorm dissipates to the time it might take for an area to recover from drought. In this case we are comparing the memory of the atmosphere to the memory of the soil. Memory within the soil and land surface is much greater than that of the atmosphere and is therefore a useful tool in weather forecasting and climate projections (Koster et al. 2000; Koster and Suarez 2004; Koster et al. 2004; Koster et al. 2010; Schlosser and Milly 2002; Shinoda 2001, Small 2001; Wu et al. 2002).

Soil moisture anomalies persist for longer periods in northerly latitudes due to decoupling of the surface from the atmosphere by snow cover during winter (Barnett et al. 1989; Delworth and Manabe 1989; Yeh et al. 1984). The snow cover prevents the loss of soil moisture through evaporation and increases the albedo of the surface, so the land surface absorbs little, if any, energy from incoming solar radiation (Barnett et al. 1989). Evaporation is limited by the thermal gradient and moisture holding capacity of the atmosphere (Yeh et al. 1984); thus in colder months, it does not contribute substantially to precipitation. Cooler air temperatures persist for longer when the snowpack is more extensive (Ellis and Leathers 1998). When the snow melts, wetter soils regulate above ground temperatures such that wetter soils induce near surface atmospheric cooling (Barnett et al.1989). Wet soil moisture anomalies correspond to cooler air temperatures and often persist owing to a reduction in evaporation. Dry soil anomalies also persist when less moisture is available at the surface. In these conditions, the surface warms to greater temperatures and the water stored is evaporated more efficiently.

Snow greatly impacts the thermal conditions of the ground beneath it, as well as above surface air temperatures. The high albedo of snow means that a greater portion of incoming shortwave radiation is reflected back to space than is absorbed by the snowpack (Oke 1987; Stieglitz et al. 2003; Zhang 2005). Having a low thermal conductivity, the snowpack conducts less energy to the ground, a process dependent on snow depth, such that the soil is generally warmer than the air temperature during winter (Zhang 2005). Generally, a snowpack will insulate the soil from changes in air temperature. The snowpack also acts as a heat sink because it requires so much energy for this change of phase to occur (Oke 1987; Zhang 2005). When the snowpack melts, this sink is lost and short wave radiation heats the ground instead of the snowpack. The overall effect of the snowpack on the ground thermal regime is dependent on the timing, duration, density and thickness of the snowpack as well as local meteorological conditions, topography and vegetation (Zhang 2005).

Soil moisture memory ties into local and regional climate predictions through relationships that cause anomalous conditions to persist. Convective instability in the boundary layer can cause wet soil conditions to persist. In areas characterized by wetter soil conditions convective processes will be forced as wetter soil will lead to a cooler, more shallow and moist boundary layer. A shallower boundary layer corresponds to a lower level of free convection as well as an increased net radiative flux (Oke 1987; Schar et al. 1999; Pal and Eltahir 2001). These conditions favour convective instability, which leads to precipitation formation. Through these processes, climate memory is retained and conditions persist, affecting downwind areas and contributing to climate change via teleconnections.

Of interest to this study are characteristics of the snowmelt period, how they are retained in the soil, and how long the snowmelt signal can be detected. With changes in the precipitation regime and land use in the adjacent area, knowing how the soil responds to precipitation and snowmelt inputs might help with forest management and in planning for new infrastructure such as roads. Vegetation is also likely to be affected by changes in the form and timing of precipitation and melt processes.

2.10 Soil Moisture and Groundwater Recharge and Quantification

Excess soil moisture is generally considered to be a contributing factor to groundwater recharge (Rushton et al. 2006). Much of our freshwater resources are fed in part by groundwater, so the contribution of precipitation to soil moisture and subsequently groundwater recharge is of great interest to many scientists.

The difference between infiltration and evapotranspiration yields estimates of groundwater recharge; however due to spatial variability in soil conditions and a host of other factors, these estimates are prone to significant error and are therefore unreliable. Models must incorporate important processes and take into consideration some heterogeneity to provide better estimates. These factors may differ from region to region making it impossible for any single model to provide a global estimate. The following is a look at different approaches to this issue, from methods of estimating recharge to looking at soil moisture memory. These concepts are intrinsically linked as they both consider the role of near surface soil storage (of water) in the processes of infiltration and evapotranspiration.

There have been many approaches to estimating groundwater and soil moisture recharge, none without their own inherent error. Estimates can be made by monitoring the level of the water table, use of lysimeters, and use of chemical tracers (Murray and Buttle 2003; Rushton et al. 2006). Less direct methods include estimating infiltration based on unsaturated flow equations and hydraulic properties and/or storage properties (Vigiak et al. 2006) or crop stress and yield. The soil moisture balance technique is another method for estimating recharge, depending on estimates of evapotranspiration and moisture holding properties of the soil (Rushton et al. 2006).

Some studies show evidence of infiltration occurring in frozen soils (Sutinen et al. 2008) or during warm periods in the winter. The snow provides an insulating layer keeping the near surface soil from freezing in the winter. In fact, measurements have shown that the soil temperature at 12 cm depth hovers near 1.5°C with a snow cover equal to or greater than 20 cm thickness (Oke 1987). This implies that unfrozen water that percolates to the bottom of the snowpack has the potential to infiltrate the soil, contributing to soil water and groundwater recharge even during the winter. Antecedent conditions must be considered as well as they will greatly affect the capacity of the soil to accept and retain water during the melt period.

There is a lack of sufficient ground-based data to initiate and support modeled outcomes. Having ground based measurements from areas where evapotranspiration contributes significantly to local precipitation will improve weather forecasting in these and adjacent regions as well as provide more certainty with respect to projections of climate change.

With the advent of better remote sensing technology, soil moisture may be estimated using surface radiometric temperature data and other forms of remotely sensed data (Crow et al. 2008). In earlier studies, advances were made relating vegetation cover and surface temperature data to estimate soil moisture content (Vicente-Serrano et al. 2004; Wang et al. 2007). Wang et al. (2007) show that soil moisture can be estimated from a combination of microwave and optical/infrared remote sensing data. Normalized Difference Vegetation Index (NDVI) and land surface temperature data from the Moderate Resolution Imaging Spectroradiometer (MODIS) are used to generate soil moisture estimates at a horizontal resolution of 1 km (Wang et al. 2007).

These methods are not without their own uncertainties, which are often difficult to resolve due to spatial heterogeneity of land surface features. For example, soil dielectric properties are directly related to soil moisture content and hence are used to estimate this parameter. Many models used to estimate soil moisture from soil dielectric properties do not account for the role of soil texture in determining dielectric properties leading to errors in estimations of soil moisture content modelled from these data (Fernandez-Galvez 2008). Since these features may vary significantly over relatively small areas, mapping soil moisture at very fine scales is nearly impossible, especially in complex and mountainous terrain.

Characterizing soil moisture conditions involves some analysis of its temporal variability. This has been achieved using a number of different analytical tools such as wavelet analysis (Lauzon et al. 2004; Parent et al. 2006), cross-spectral analysis (Wu et al. 2002), and autocorrelation (Deliberty and Legates 2008). These tools share the common purpose of evaluating persistence in a time series. For Mt. Tom, precipitation events occurring during or after the snowmelt period must be removed, or filtered, from the soil moisture data to characterize the snowmelt signal without the added noise from random precipitation events.

Chapter 3: Site Description

3.1 Mt. Tom

The Mt. Tom Forest Area is located south-east of Prince George, north of Wells, BC and can be accessed via Naver Rd. just south of Hixon, BC (see Figure 3.1) . The area lies within the Quesnel (Cariboo) Forest District and is used for a number of different activities, primarily logging and mining. There are currently a number of ongoing studies in the area, including the Mt. Tom Adaptive Management Trial that is managing forests in an effort to protect mountain caribou populations while sustaining logging practices in the region.

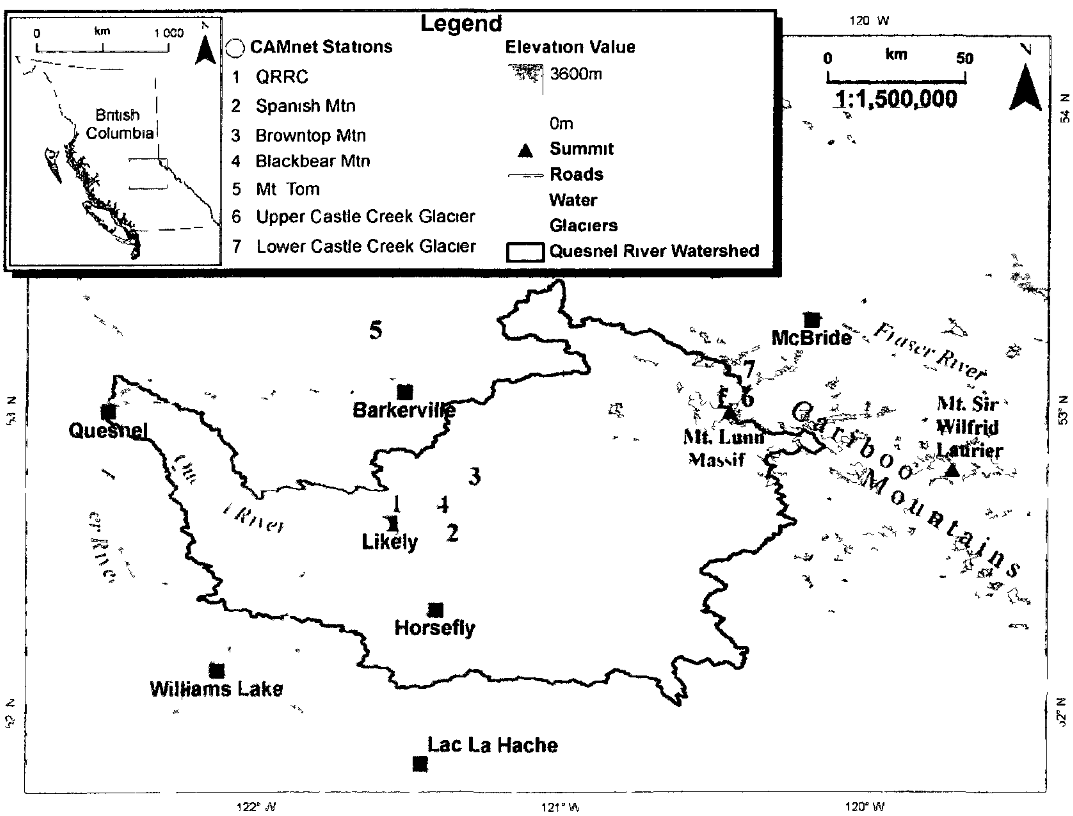

Figure 3.1- CAMnet stations, map by Theo Mlynowski, (2009)

The meteorological station at Mt. Tom supports ongoing research by the NHG regarding local hydrological and atmospheric processes. It was installed in the fall of 2007 and was dismantled in July 2009 to provide equipment for other studies in September 2009. The station is part of the Cariboo Alpine Mesonet (CAMnet; see Figure 3.1), a network of meteorological stations set up by the NHG to monitor the water budget of the Quesnel River Basin (MacLeod and Déry 2007; Déry et al. 2010); however, the Mt. Tom site is in the adjacent Willow River watershed. The coordinates of the station are 53° 11.54' N, 121° 39.82' W and it is at 1489 m above sea level (a.s.l.). The slope of the terrain at the meteorological station is about 3° in a southwest direction with the surrounding area having variable steepness (Kara Przeczek, UNBC, personal communication).

For this study, the meteorological station was set up in a cut block approximately one hectare in area, that was logged in 2006. The surface is littered with logs and other screef left behind from clear cut logging practices. Some shrubs and grasses grow and there are spruce saplings that were planted in 2007 throughout the cut block (Pat Teti, Ministry of Forests, BC, personal communication). As discussed in the previous chapter, vegetation affects the albedo of the surface and hence plays a large role in the surface energy budget and atmospheric processes occurring at the surface. Aspect and slope also greatly affect the amount of insolation available to heat the surface or melt snow. At Mt. Tom, snow cover persists longer in forested areas and on north-facing slopes and melts sooner on south facing slopes or well drained, open areas (Kara Przeczek, UNBC, personal communication).

3.2 Biogeoclimatic Zone Designation

The abbreviated designations in the legend of Figure 3.2 are Englemann Spruce - Subalpine Fir (ESSF), Sub-Boreal Spruce (SBS), Interior Mountain-heather Alpine (IMA). The lower case letters in the biogeoclimatic zone designation are to categorize soil conditions and climate (wet, cool, damp, cold, etc.). The area in which the meteorological site is located is classified as ESSFwk1 or ESSFwc3 (see Figure 3.2), according to biogeoclimatic zones designated by DeLong (2003). ESSF signifies that the primary vegetation in this zone is Engelmann Spruce (*Picea englemannii*) and Sub-Alpine Fir (*Abies lasiocarpa*) (DeLong 2003). The ESSFwk1 zone is characterized as being very cold and very wet (DeLong 2003). The ESSFwc3 sub-zones are characterized as being the wettest subzones in the Cariboo Forest District and always lie between ESSFwk1 at lower elevations below 1500 m and alpine tundra (AT) at higher elevations, generally above 1800 m (Steen and Coupe 1997). ESSFwc zones are colder than ESSFwk zones but warmer than AT zones. The site of the meteorological station is at 1489 m elevation, and hence lies at the transition between these two sub-zones.

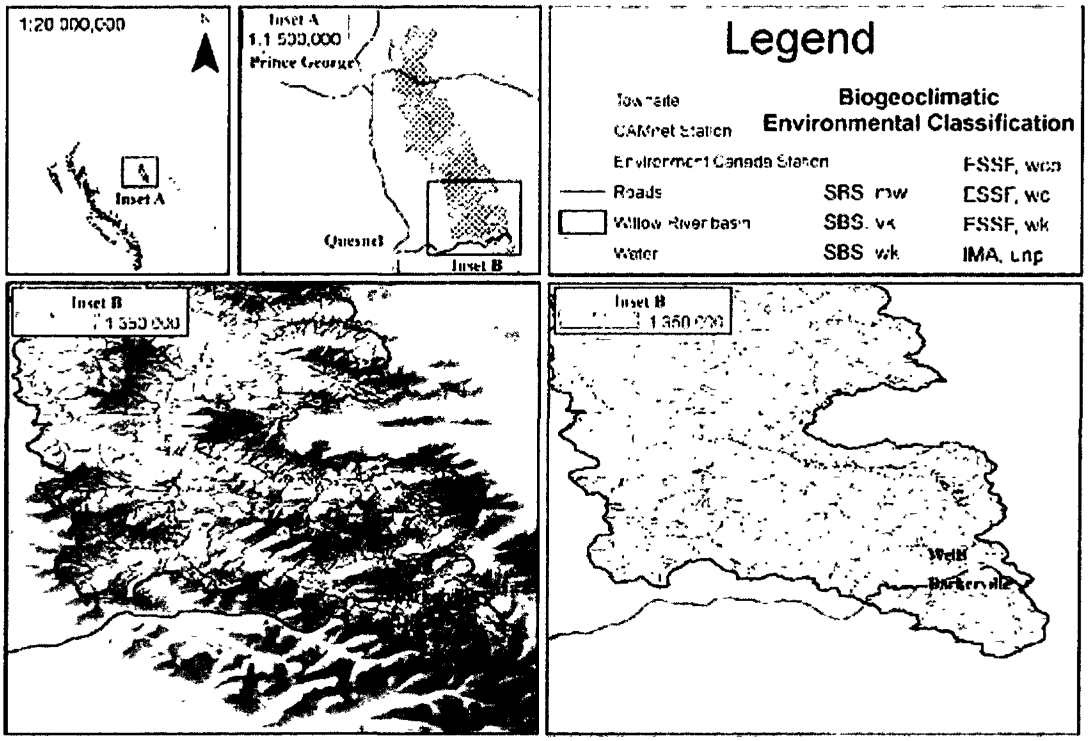

Figure 3.2 Map of study area, BEC designations are explained in Chapter 3.2, map by Theo Mlynowski, (2009)

3.3 Soil Landscape

The soil profile is visible along road cuts throughout the forest area. During a site visit on 16 June 2008 with Dr. Paul Sanborn and Ekaterina Daviel the following observations of spatial variability in soil characteristics were made. Near the meteorological tower the profile is generally homogeneous, varying mostly in moisture content due to topographic variability. The soil at this site is typical of the surrounding area, and would generally be classified as a Humo-Ferric Podzol (British Columbia Soil Survey 1985, Valentine et al. 1978). This classification is consistent with conditions that were observed on 16 June 2008. The photos and descriptions below are from the study area.

Figure 3.3 shows a drier profile. Note how deep the roots penetrate the soil and the lighter color of the soil. This soil was dry to touch and would conduct water with greater ease than the soil in the second photo. This soil was described as a Humo-Ferric Podzol on sandy gravelly glaciofluvial parent material. The sandy, gravelly matrix drains more easily than the till found in the second pit.

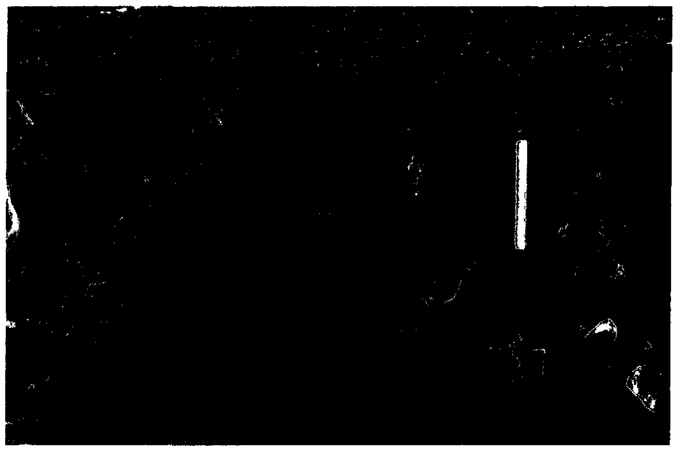

Figure 3.3- Soil profile of drier soils at Mt. Tom, 53° 11' 22.9"N 121° 39' 44.6"W, 1483 m photo taken 16 June 2008 by Paul Sanborn

Figure 3.4 shows a profile that is typical of the soils at Mt. Tom. They tend to be moist and void of roots below 30-50 cm depth. Moist conditions have been noted on several outings to Mt. Tom from October 2007 to September 2008. The soil in this pit was described as a Humo-Ferric Podzol on loamy, morainal parent material. There was compact till at 70 cm depth in the profile.

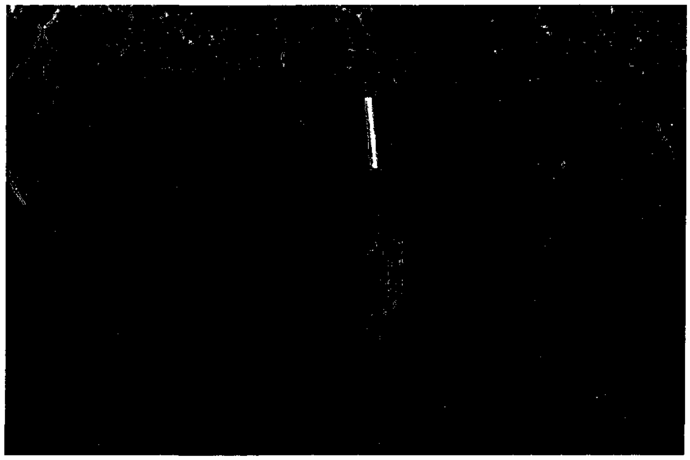

Figure 3.4- Soil profile at Mt. Tom, 53° 11' 19.7"N 121° 39' 26.8"W, 1508 m, photo taken 16 June 2008 by Paul Sanborn

The forest floor depth varies between 5-10 cm thickness. This layer consists of a matted organic layer of decomposing leaves and other matter. Beneath this layer is an Ae horizon ranging between 1-4 cm in thickness. The Ae horizon was greyish and moist at the time of visit. The next layer is described as either a Bf or a Bhf horizon, and is 10-15 cm thick with very few aggregates. A carbon analysis would be needed to make the distinction between these two designations. Following the Bf/Bhf layer is a modified B layer (Bm) about 15-20 cm sub-angular blocky structure. Next is the BC horizon which is transitional to the parent material. This layer is well structured and slightly fractured. Clay films are visible along the fractures indicating downward movement of water through the cracks. Roots are generally confined to the upper 30 cm of the profile with the exception of drier areas where the roots must grow deeper to reach water stores. A brief description of what the horizon designations signify can be found in Table 3.1 below. These definitions are compiled from the Canadian System of Soil Classification (Soil Classification Working Group 1998).

Table 3.1- Horizon descriptors and definitions adapted from the Soil Classification Working Group, 1998

Descriptor	Definition
A	surface mineral horizon
B	mineral horizon
C	mineral horizon at depth, relatively unaffected by pedogenic processes
e	eluviated horizon
f	enriched with Fe or Al, up to 5% organic carbon
g	grey horizon
h	enriched with organic matter, more than 5% organic carbon is Bhf
m	modified by chemical processes changing color and/or structure

Areas with better drainage generally were drier with deeper root penetration. The soil material was sandier and rockier in these areas, allowing for easier drainage but less water retention. Inundated areas along gullies and water channels were generally organic and mesic with some peat and vegetation that can survive in wet conditions.

The soil hydraulic properties can be estimated using the algorithms of Saxton et al. (1986) that are based on percent sand and percent clay composition. The soils at Mt. Tom are described as having percent composition of about 32% sand and 8% clay (British Columbia Soil Survey 1985). Plugging these values into the online soil hydraulic properties calculator based on the aforementioned algorithms yields an estimated field capacity of 27% and a saturation capacity of 42% for the soils at Mt. Tom .

3.4 Local Climate

Barkerville is about 20 km southeast of the Mt. Tom area where our meteorological station was located (see Figure 3.2). The Environment Canada station at Barkerville is at 1289 m in elevation, making it a reasonable station with which to compare our data. Looking at the 30-year climate normals (1971-2000) for Barkerville allows us to characterize the 2008-2009 season at Mt. Tom. It should be noted that data from Barkerville are expected to have slightly warmer average air temperatures because it is approximately 200 m lower in elevation compared to Mt. Tom.

Average monthly mean air temperatures at Barkerville range from -8.8°C in January to just over 12°C in July for the period from 1971-2000 (see Table 3.2) with an average annual air temperature of 1.9°C. Fall and winter temperatures (October to April) show much greater variation than spring and summer temperatures (April to October) with standard deviations ranging between 1.7°C in March to 3.6°C in January compared to 1.2°C in July and 1.9°C in September.

Table 3.2- Average, maximum, minimum and standard deviation of monthly air temperature (°C) at Barkerville, 1971 to 2000, available online at Environment Canada's Climate Archive, Environment Canada (2010)

	Aug	Sept	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Year
Avg	12.1	8.1	2.6	-4.4	-8.2	-8.8	-5.9	-2.9	1.4	6.2	9.8	12.3	1.9
Max	19.0	14.3	7.6	-0.1	-3.8	-4.0	-0.5	2.9	7.4	12.6	16.1	19.0	7.5
Min	5.2	1.9	-2.3	-8.7	-12.6	-13.6	-11.3	-8.7	-4.6	-0.1	3.5	5.6	-3.8
St Dev	1.4	1.9	1.7	2.9	3.4	3.6	3.1	2.0	1.5	1.6	1.4	1.2	1.2

Average annual precipitation is about 1014 mm with 481 mm falling as snow and the remaining 533 mm falling as rain (Table 3.3). Thus almost 50% of the annual precipitation in this area falls as snow. June and November are the wettest months of the year, each with monthly precipitation averaging about 102 mm, with the majority of precipitation in November falling as snow.

Table 3.3- Average monthly rainfall (mm) and snowfall (mm water equivalent) at Barkerville, 1970-2000, Environment Canada (2010)

	Aug	Sept	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Year
Rainfall (mm)	85.1	74.4	57.1	14.4	6.2	6.5	3.4	8.1	20.4	59.8	101	96.3	533.0
Snowfall (mm W.E.)	0.0	3.1	28.6	87.4	102.5	92.5	60.9	59.5	35.5	10.0	0.8	0.0	481.0

Chapter 4: Data and Methods

4.1 The Meteorological Station

The station at Mt. Tom is equipped with a number of instruments and sensors that monitor meteorological conditions such as wind speed, air temperature and humidity. These instruments are all wired to the Campbell Scientific CR10X Data logger that collects data every 15 minutes. Table 4.1 summarizes the instruments present at this site and the tower itself is shown in Figure 4.1. These data can be downloaded from the station using the Loggernet software, with a connector cable from the computer to the data logger.

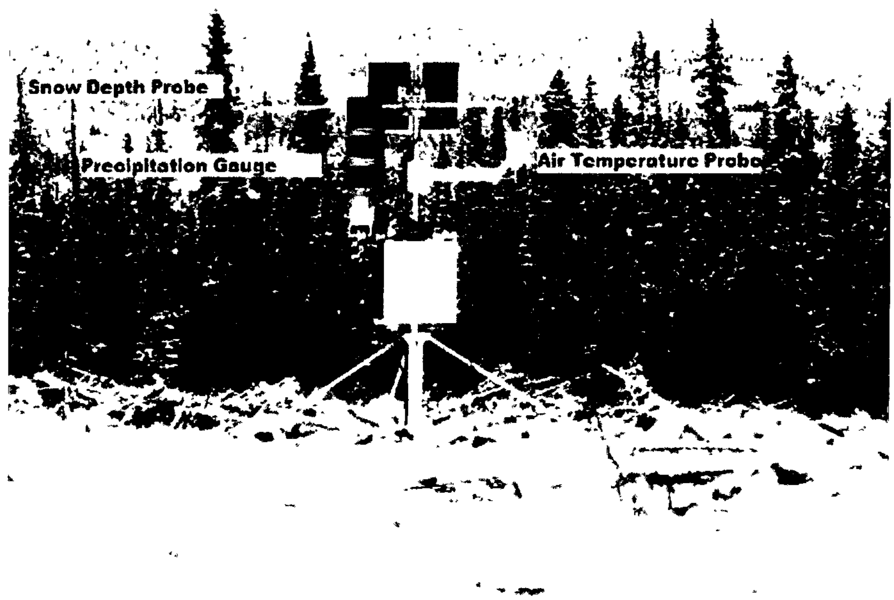

Figure 4.1 - Photo of the meteorological station at Mt. Tom facing South taken on 19 October 2007 by Tullia Upton.

Table 4.1- Instruments installed at Mt. Tom.

Instrument	Variable Measured	Accuracy
CR10X Data logger	stores measurements	N/A
107B Temperature Probe	soil temperature at 12 cm	± 0.9°C
CS616 Water Content Reflectometer	soil moisture at 10 cm, 20 cm and 35 cm	± 2.5% Volumetric Water Content
SR50A Sonic Ranger	snow depth	± 0.4% of distance to target
RM Young Wind Monitor	wind speed and direction	± 0.3 m s^{-1}
HMP45C212 Temperature and RH probe	temperature and relative humidity	± 0.1°C/ ± 2% (0-90% RH), ± 3% (90-100% RH) at 20°C
TE525WS Tipping Bucket	precipitation	± 1% of precipitation rate up to 2.54 mm hr^{-1}
CMP3 Pyranometer	incoming solar radiation	± 5%

Of particular interest to this study are variables that are known to affect near-surface atmospheric processes: soil moisture, snow depth, precipitation, air temperature and relative humidity. Each of these instruments is described in brief, below.

4.1.1 Soil Moisture

Soil moisture is measured as volumetric water content (VWC) and is expressed as a percent in decimal form. VWC (or soil water content) is the volume of pore space within the soil that is occupied by water. The total volume of soil in unsaturated conditions is given by the following equation:

$$V_T = V_w + V_s + V_a \quad (6)$$

where, V_w is the volume of space occupied by water, V_s is the volume of soil particles and V_a is the volume of space in the soil occupied by air. Volumetric water content is then simply:

$$\theta = \frac{V_w}{V_T} \qquad (7)$$

Soil moisture data were sampled every minute and averaged over 15 minutes. The instruments deployed to measure VWC are Campbell Scientific® Water Content Reflectometers (CS616). They consist of two steel rods 30 cm long running parallel to each other. These rods are attached to a circuit board that sends out an electric pulse, according to the interval specified in the customized computer program that connects the instrument with the data logger. The pulse travels down the rod and returns to the circuit board, which measures the time it takes to do so. This period is directly related to the water content of the soil.

The pit was dug to about 50 cm depth and was about 1 m across. The probes were installed at depths of 10 cm, 20 cm and 35 cm in the soil column. Originally, the probes were meant to be spaced approximately 5 cm apart horizontally, and 10 cm apart vertically. Rocks in the soil prevented this spacing so they were inserted as close to these spacings as was possible. The first two probes at 10 cm and 20 cm depth are offset from each other by approximately 5-10 cm horizontally. The third probe at 35 cm depth is almost directly beneath the second (see Figure 4.2).

The probes are horizontally offset from each other by 2-10 cm to prevent interference between instruments and to try and minimize overall disturbance to the soil layer they were inserted in. This set-up has been used in other studies (Sutinen et al. 2008) but was adapted to the Mt. Tom site as best as was possible.

The CS616 probes do not measure frozen water content and hence, frozen water within the soil will not be accounted for until it melts. At the Mt. Tom site, the soil temperature at 12 cm depth did not drop below 1.5°C for the entire winter season. Since the first probe is at 10 cm depth we can assume that the water at this depth was not frozen during the winter and hence our measurements represent the true soil water content for the entire period of time that data were collected.

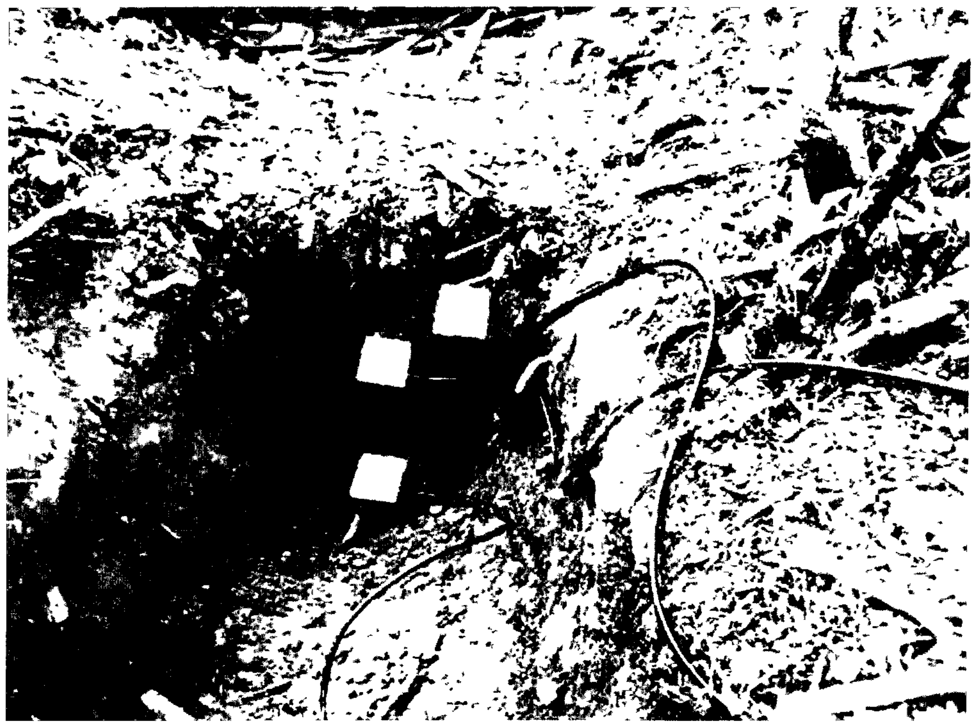

Figure 4.2- Soil moisture sampling Pit at Mt. Tom, installed 16 July 2008, photo taken by Tullia Upton.

Data were sampled every minute and then averaged over 15 minute periods for this study. This interval should capture short-lived precipitation events and allows for a better representation of the temporal variability associated with soil moisture. Hourly and daily averages can be computed from these values. All other instruments at this station reported data in 15 minute intervals. Data were periodically collected from the field using the LoggerNet software designed to work with the Campbell Scientific® data loggers.

4.1.2 Snow Depth

Snow depth was measured using the Campbell Scientific® SR50A (Sonic Ranger). This instrument emits an ultrasonic pulse that travels from the sensor to the surface and back. The time it takes to do so is converted, using complex equations, into a depth of snow. The measurement must be corrected for the effect of air temperature on the speed of sound in air (Campbell Scientific 2006). The data from Mt. Tom show many inconsistencies and are difficult to filter. For this reason, they are supplemented with snowfall data from nearby Barkerville where snowfall measurements are taken twice daily. These data are reliable and allow for comparison of the 2008 to 2009 season with climate normals (1971 to 2000) established at Barkerville.

The SR50A only provides a depth to target estimation and hence does not provide information on SWE or snow density. Snow surveys as described below are done to obtain estimates of SWE. These values are then used to find the density of the snow through the following equation:

$$\rho_{(snow)}\ (kg\ m^{-3}) = [\ SWE\ (m)\ /\ snow\ depth\ (m)]^{-1}\ x\ \rho_{\ (water)}\ (kg\ m^{-3}) \quad (8)$$

Snow density changes over the winter as snow undergoes morphological changes resulting from temperature fluctuations and compaction under its own weight. Sublimation could also contribute to loss of water from the snow pack.

SWE is an important factor in determining the volume of water stored and available from snowmelt. It is determined using a Federal snow corer and spring scale that are calibrated to give SWE of a snow core based on weight. The accuracy of the instrument is ±0.5 mm. The tube is inserted and forced through the snow pack. A sample that has not reached the surface of the ground (i.e. there is no soil visible in the bottom of the core) is not acceptable and must be resampled. The tube is weighed with the snow inside and values are recorded in a table. SWE is read from the scale on the tube and can be determined by subtracting the weight of the tube from the weight of the snow plus the tube.

Twenty samples were taken on 6 April 2009 to determine the average density of the snow pack at Mt. Tom. Snow depth measurements were taken along with SWE measurements. This process was completed only once in April 2009 and hence it is not possible to comment on the evolution of the snowpack with respect to SWE or density. The SWE data collected only provide an estimate of snowpack conditions for 6 April 2009. This is close to the date of peak accumulation at Barkerville, which was 156 cm on 2 April 2009. Snow density is generally less variable spatially than snow depth or SWE, therefore, it is assumed that the 6 April 2009 snow density measurements are representative of the season.

4.1.3 Air Temperature and Relative Humidity

The HMP45C212 air temperature and relative humidity probe was housed in a Gill radiation shield to prevent erroneous air temperature measurements. This probe made concurrent measurements of humidity and air temperature at the same temporal resolution as all other variables being monitored.

4.1.4 Precipitation

Measuring precipitation at this remote site was particularly challenging and the data collected are not considered reliable. The precipitation gauge at Mt. Tom is a TE525WS Tipping Bucket Rain Gauge. It measures precipitation in increments of 0.254 mm of water. Every 0.254

mm of water collected by the bucket causes a 'see-saw' device that collects the water to tip. It may take more than one short-lived precipitation event to cause a 'tip'. The problem this poses is that shorter, less intense events may actually not be represented by the gauge until further precipitation causes sufficient accumulation.

There were other problems associated with the precipitation gauge at Mt. Tom. The gauge was attached to the mast of the meteorological station. This could have affected the catch of precipitation if the gauge was obstructed by other instruments on the tower. Also, the gauge is quite small and unshielded such that it was likely affected by wind. Unless the air was very still, it is likely that precipitation accumulation was being underestimated because the precipitation tended to fall sideways as opposed to straight down into the gauge.

There were also problems with water freezing in the gauge. Adding antifreeze to the gauge in the fall season is meant to prevent freezing of winter precipitation; however, there is still some freezing that occurs and hence, precipitation events are not recorded at the actual time they occur.

These data were deemed insufficient due to the unreliability of the instrument. Data from Barkerville were thus used to supplement precipitation data from Mt. Tom. The data from this site were collected twice daily so that there are data for accumulation in the morning vs. night. Values are given in centimetres for the volume of liquid precipitation or the depth of snow. Daily snowfall and daily snow depth measurements are provided for the winter season.

4.2 Methods

4.2.1 Data Downloading and Quality Control

The period for which reliable data were obtained from Mt. Tom spans from 17 July 2008 to 14 July 2009. Data were downloaded from the CR10X Campbell Scientific Data logger using the Loggernet program. They were labelled and stored for future use. The first step was to organize the data and undertake their quality control. This involved checking for any anomalies and rectifying any spurious data. Generally, missing data were replaced by taking the average of the values preceding and following the missing data. Anomalous values arose in some data, mostly the snow depth data. The snow depth data from Mt. Tom were supplemented with data from Barkerville.

4.2.2 Selecting Antecedent and Snowmelt Periods

The antecedent period was chosen to be 15 August 2008 to 20 November 2008. This period is characterized by cool, damp conditions and the onset of below freezing air temperatures and solid precipitation (snow). The period was chosen to represent conditions prior to fall freeze-up. The cut-off date for the season was chosen to be about 10 days after the daily average soil temperature stopped fluctuating by more ±1°C. This occurred around 10 November 2008. After this date, snow depth was almost consistently above 15 cm and the soil temperature at 12 cm depth in the soil near 2°C.

We compare these data with the snowmelt period beginning 1 April 2009 to 15 June 2009 (the snowmelt period). This period is characterized by decreasing snow depth and daily average air temperatures above freezing. In the snowmelt period the soil temperature remained at about $1^{\circ}C$ until the snow depth decreased to about 20 cm depth. In 2009, the snow cover had melted at the site around 22 May. Soil water content seems to return to average values around 5 June 2009. After this date, increases in daily average soil moisture clearly correspond to precipitation events.

4.2.3 Computing Averages and Correlations

Once sorted and quality controlled, the data were transferred into the R statistical software version 2.11.1 (R Development Core Team, 2008). This program is used to calculate hourly and daily averages from the 15 minute data in a sequence loop command. The data presented here are based on daily averages computed from values logged every 15 minutes on the CR10X Data logger. Pearson product-moment correlation coefficients, autocorrelation and cross-correlation coefficients at the 95% confidence level were computed in R (R Development Core Team 2008).

Autocorrelation describes persistence in a successive list of values, in this case soil moisture values. The statistical test for autocorrelation in this study assesses the persistence of the soil moisture from snowmelt. Autocorrelation was calculated from daily average soil moisture time series at a lag period of one to twenty days. The coefficient of autocorrelation, ρ, with values between -1 and +1 is defined by the following equation:

$$\rho(\tau) = \frac{\gamma(\tau)}{\gamma(0)} \qquad (9)$$

where τ is the lag (units of time) and γ is the covariance of a stationary process. The covariance is determined from the auto-covariance function of the time series X_t, with mean μ given by:

$$\gamma(\tau) = \varepsilon((X_t - \mu)(X_{t+\tau} - \mu)^*)$$

$$= Cov(X_t, X_{t+\tau}) \qquad (10)$$

Autocorrelation can be used to assess periodicity in time series data. Positive values of ρ_{IX} indicate positive correlation in the time series; values near or equal to zero indicate no correlation and values less than zero indicate that values of the time series become less and less related with the passing of time. If values of ρ are consistently close to zero, then there is no periodicity to the data, and they are assumed to be random. Higher or lower values of ρ suggest some trend in the data.

Cross-correlation analyses are performed for the three depths of soil moisture measurements. Cross-correlation evaluates similarities in the periodicity of two different time series, X_τ and Y_τ. When values are significantly positive or negative then we identify the time lag at which the two series are most highly related. This offers insight into the time lag between response of the soil to moisture inputs at one depth compared to the other depths. The cross-correlation is given by the equation:

$$\rho_{xy}(\tau) = \frac{\gamma_{xy}(\tau)}{\sigma_X \sigma_Y} \qquad (11)$$

where σ_X and σ_Y are the standard deviations of processes X_τ and Y_τ.

Chapter 5: Results

This section is organized to facilitate comparison of soil moisture conditions between the fall freeze-up, referred to as the antecedent period, versus conditions in the spring or snowmelt period. Before we look at these two periods, we will examine the data from 17 July 2008 to 14 July 2009. This is the period for which reliable data from Mt. Tom are available.

5.1 Mt. Tom and Local Climate

The 2008 to 2009 season overall had cooler air temperatures at 0.4°C for the annual daily average compared to 1.9°C for the 1971-2000 period at Barkerville. The slightly lower average is expected as Mt. Tom is about 200 m higher in elevation than Barkerville. Air temperatures were warmer than average during the early winter months with average monthly air temperatures ranging from about -6.0°C in January, the coldest month for this year, to just over 1.2°C in December in the 2008-2009 winter season.

Despite this being a particularly warm winter in the earlier months, February to June were slightly cooler than average with monthly mean air temperatures hovering closer to about 7°C for most of the summer compared to the warmer climate normals from Barkerville. For the 2008-2009 period, air temperatures fluctuated more than average with maximum and minimum daily average air temperatures ranging between -28.7°C in March to 19.8°C in August. Average daily air temperatures at Barkerville from 1971 to 2000 show less variability than for the 2008-2009 period at Mt. Tom, ranging between -13.6°C in January to 19.8°C in July and August.

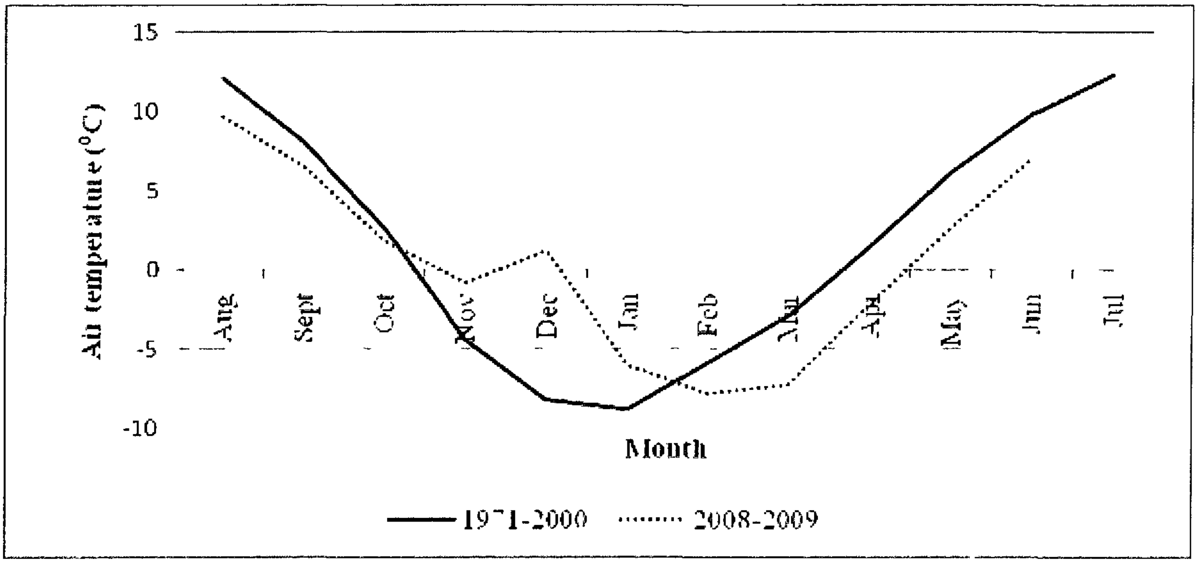

Figure 5.1- Average monthly air temperature (°C) for each month at Barkerville (1971-2000) compared to Mt. Tom (2008-2009)

Apart from February, snowfall at Barkerville was greater than average in every month of the 2008 to 2009 season. Higher than average snowfall may have contributed to lower than average air temperatures in the spring and early summer (see Figure 5.1), or it could have been a cooler spring. The mean monthly snowfall from September to June was approximately 44 cm per month according to climate normals compared to an average 58 cm of snowfall per month at Barkerville for the same period.

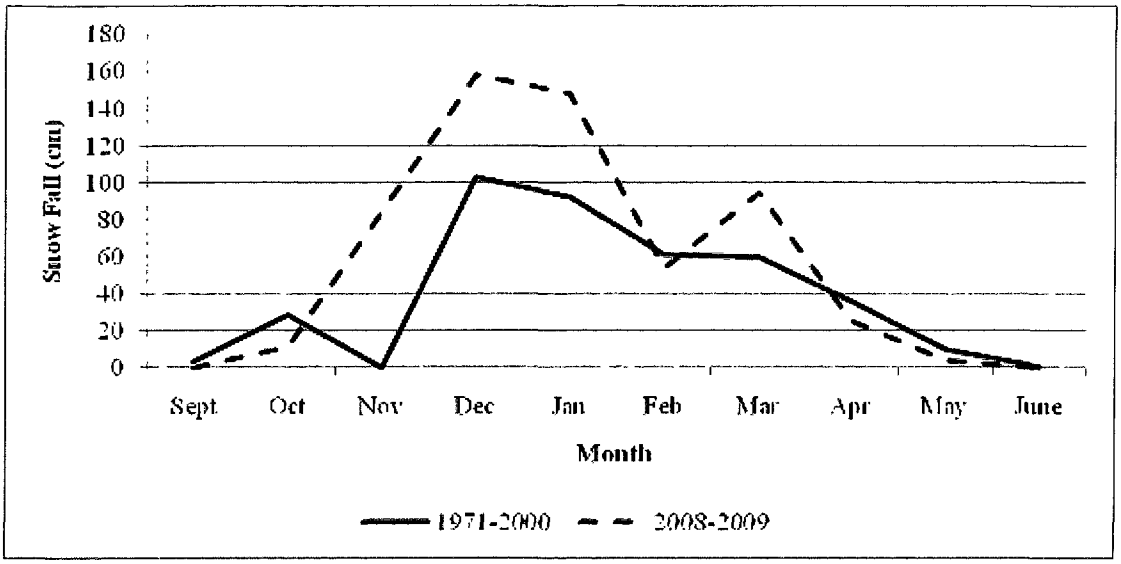

Figure 5.2 Average monthly snowfall (cm) at Barkerville (1971 to 2000) compared to 2008-2009 winter season

5.2 Mt. Tom Data (July 2008 to July 2009)

Daily average air temperatures from 14 July 2008 to 17 July 2009 reached a daily maximum of 19.8°C on 18 August 2008 and a minimum of -28.7°C on 10 March 2009. The average daily air temperature for the year was 0°C. There was an anomalously warm day on 19 January 2009, when the average daily air temperature was over 11°C. Temperatures at Mt. Tom from July 2008 to July 2009 were warmer than average in the early winter season from November to January, and colder than usual from February onwards.

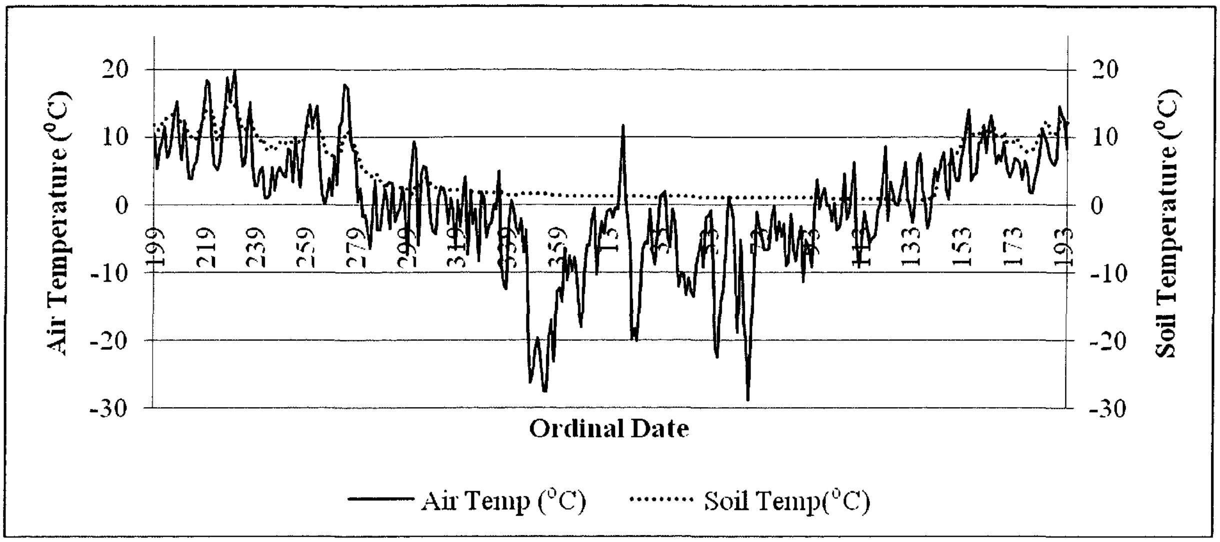

Figure 5.3 Daily average air temperature and soil temperature at 12 cm depth, Mt. Tom, 17 July 2008 to 14 July 2009

Soil temperature fluctuates throughout the summer and spring when there is no snow cover present (Figure 5.3). Between approximately 20 November 2008 (ordinal date (OD) 325) and 20 April 2009 (OD 110) the soil temperature decreased slowly from about 2°C to 0.5°C, a difference of only 1.5°C in 6 months. The sharp change in soil temperature that occurs when the snow pack melts provides an independent estimation of when snow leaves the surface at Mt. Tom. Once the snow depth decreases, soil temperature shows a steep rise to higher average temperatures ranging from approximately 7.5°C to over 15°C for the spring and summer months.

In late September temperatures begin to decline rapidly until snow cover is sufficiently deep to insulate the soil from fluctuating air temperatures. Further evidence of this is seen in mid-January when air temperatures are above 0°C for a few days (Figure 5.3) and even reach average daily temperatures above 10°C but no response is seen in the soil temperature. Interestingly the daily average soil temperature (4.3°C) is greater than the daily average air temperature (2.5°C) in the snowmelt period.

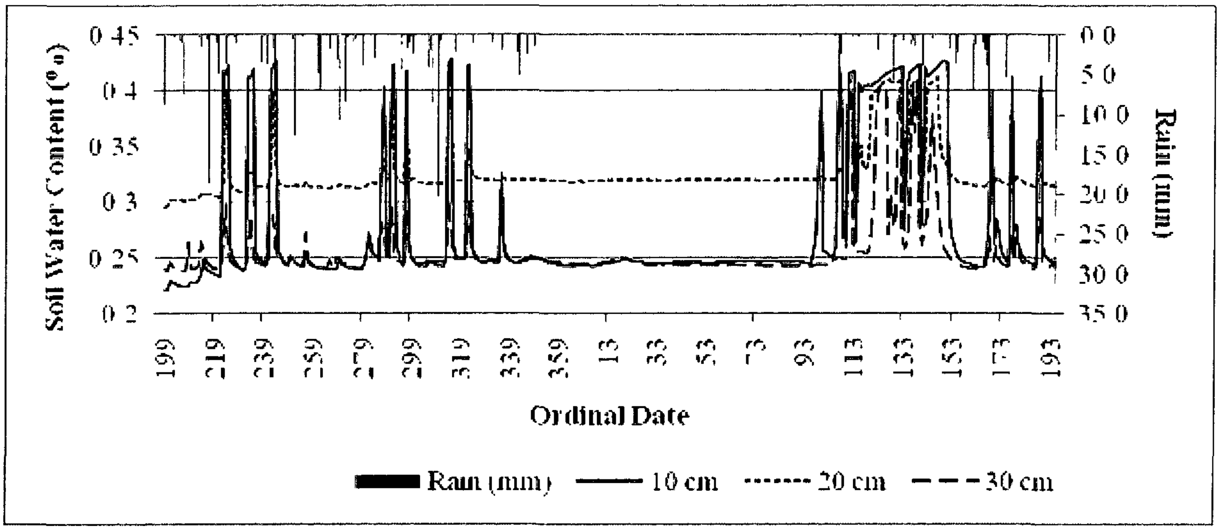

Figure 5.4- Daily average soil water content (%) and rainfall (mm) at Mt. Tom measured at 10 cm, 20 cm and 35 cm depth from 17 July 2008 to 14 July 2009.

It is clear that soil moisture responds to precipitation events in the summer with increases in soil moisture lasting only a few days at most during this season. During the winter, there is little to no change in soil moisture (see Figure 5.4). Increases are noted again in the spring melt season and are often sustained for longer periods at this time of year.

Average volumetric water content (VWC) measured for the period from 17 July 2008 to 14 July 2009 was 28% at 10 cm depth, 33% at 25 cm depth and 26% at 35 cm depth. These values were considerably higher during the snowmelt period at 33%, 35% and 28%, respectively. Values were only slightly lower than average in the antecedent period: 27% at 10 cm, 32% at 20 cm and 25% at 35 cm. These values are consistent with expected hydraulic properties of soils with approximately 30% sand and 10% clay by composition.

5.3 Frequency Distribution

The daily average soil moisture data are skewed to the right but show some bimodal tendency in that values are either low or high and do not frequently occur in the middle range (see Figure 5.5).

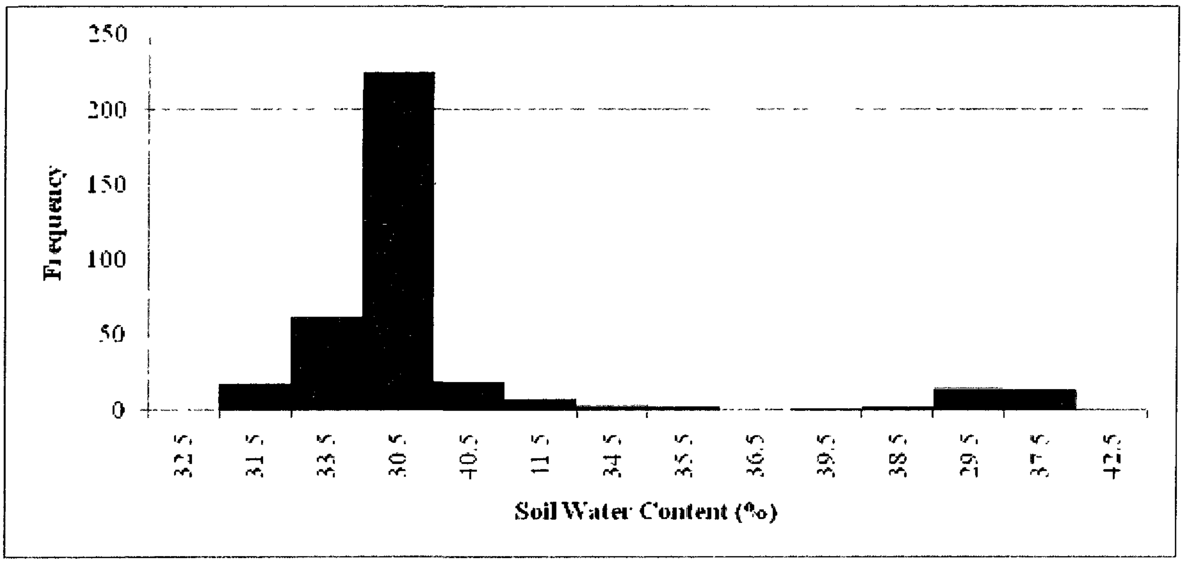

Figure 5.5- Histogram of daily average soil moisture for July 2008- July 2009 data from Mt. Tom, data from 20 cm depth

The 15 minute data collected from the data logger exhibit the same tendency towards average value with very few values in the middle range (see Figure 5.6). The data are heavily biased by the winter soil moisture values that remained at the pre-winter levels. Only one season of data prevents any definitive assessment of how data would be distributed in the long term (i.e. over several years).

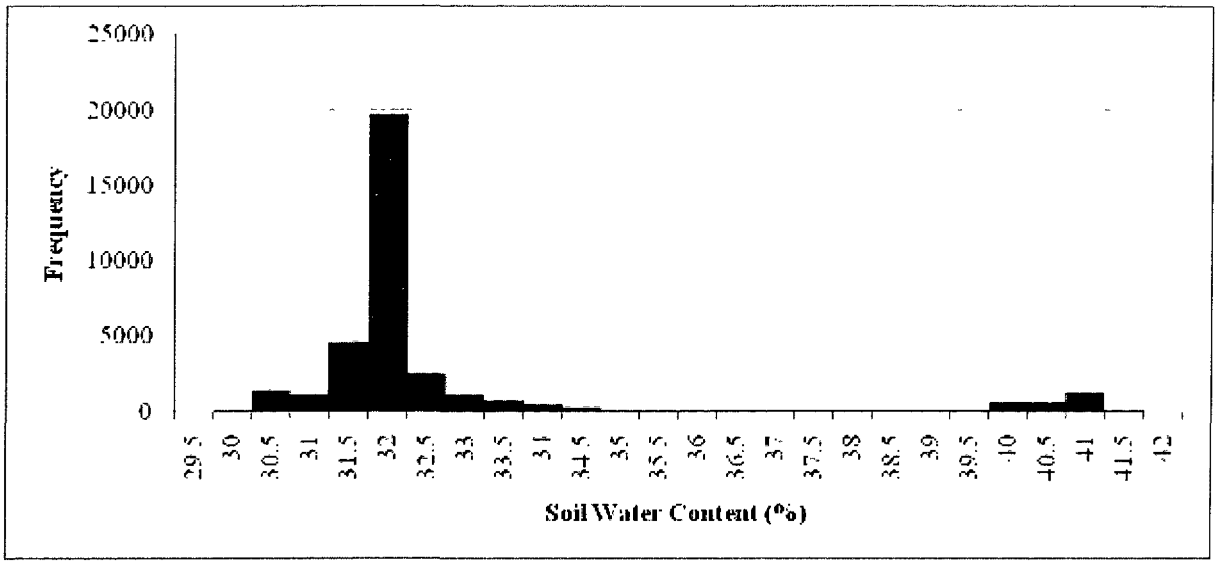

Figure 5.6- Histogram of 15 minute data for the July 2008- July 2009 period at Mt. Tom from 20 cm depth

5.4 Antecedent Conditions

Figure 5.7 shows that the first snowfall event recorded at Barkerville occured on 11 October 2008 (OD 310). This is when snow depth measurements begin for the 2008-2009 winter season. The precipitation data presented in this figure were collected from Barkerville, which is about 20 km to the south east of the Mt. Tom site. Soil temperature data from Mt. Tom taken at 12 cm depth are also plotted in this figure.

Figure 5.8 shows the soil moisture and precipitation data for the antecendent period. From the data, it is clear that increases in soil moisture content at Mt. Tom correspond to precipitation or snowmelt events at Barkerville, however, the magnitude of the soil moisture response compared to the magnitude of precipitation events is not consistent. Several factors could contribute to this inconsistency. For example, the distance between the Mt. Tom and

Barkerville meteorological stations prevents a more accurate description of the effects of the soil at Mt. Tom to local precipitation events.

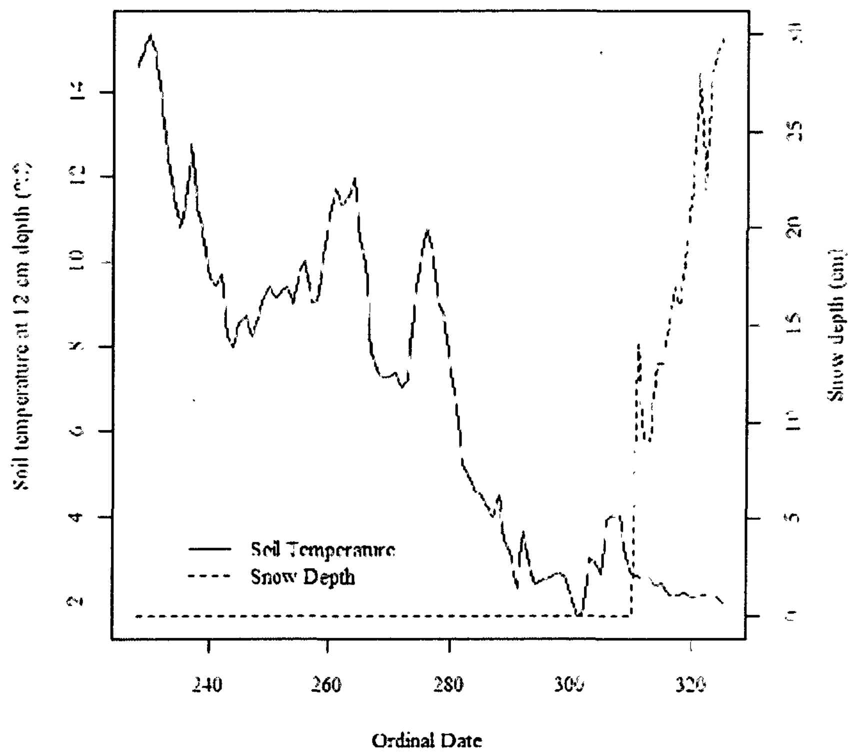

Figure 5.7- Daily average soil temperature at 12 cm depth and snow depth (cm) for the antecedent period at Mt. Tom, first snowfall occurs on ordinal date 310

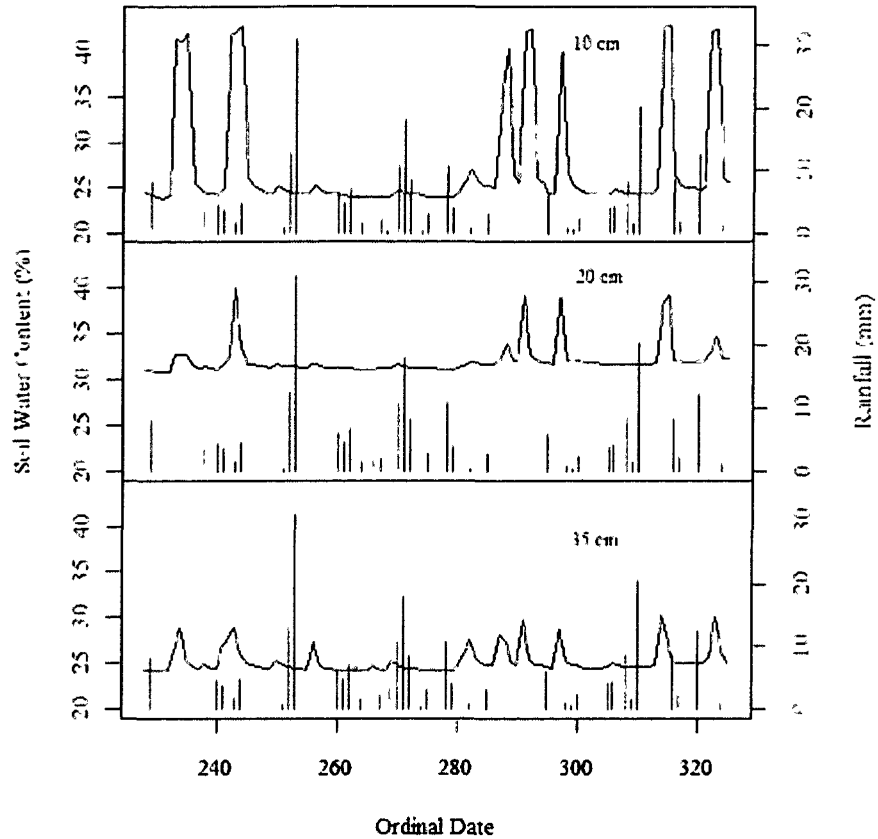

Figure 5.8- Soil water content (%) and rainfall (mm) from 15 August 2008 to 20 November 2008 (OD 228 to 325).

A few patterns to note are the slight delay, of about one day, in the soil moisture response to precipitation events. In addition, soil moisture remains higher for about 3-4 days following any given precipitation event, which is typical for well drained soils. Poorly drained soils would retain water for longer, increasing persistence of soil moisture. Air temperature and snow depth data (Figures 5.9 and 5.10) suggest that this increase may be the result of a short melt event occurring before the winter season. Daily average air temperature is above 0°C while snow depth decreases from about 27 cm to 17 cm from 18 November to 20 November.

Soil moisture does not remain above the average values for longer than three to four days following any of the precipitation events. The magnitude of an event does not change the response of the soil to preciptitation and we see that after any given rainfall event, there appears to be no substantial rise in soil moisture at any depth. We also see that soil moisture no longer fluctuates after snow cover reaches a depth of approximately 35 cm around 4 December 2008. It is approximately at this date that we see little to no fluctuation in soil temperature at 12 cm depth as well.

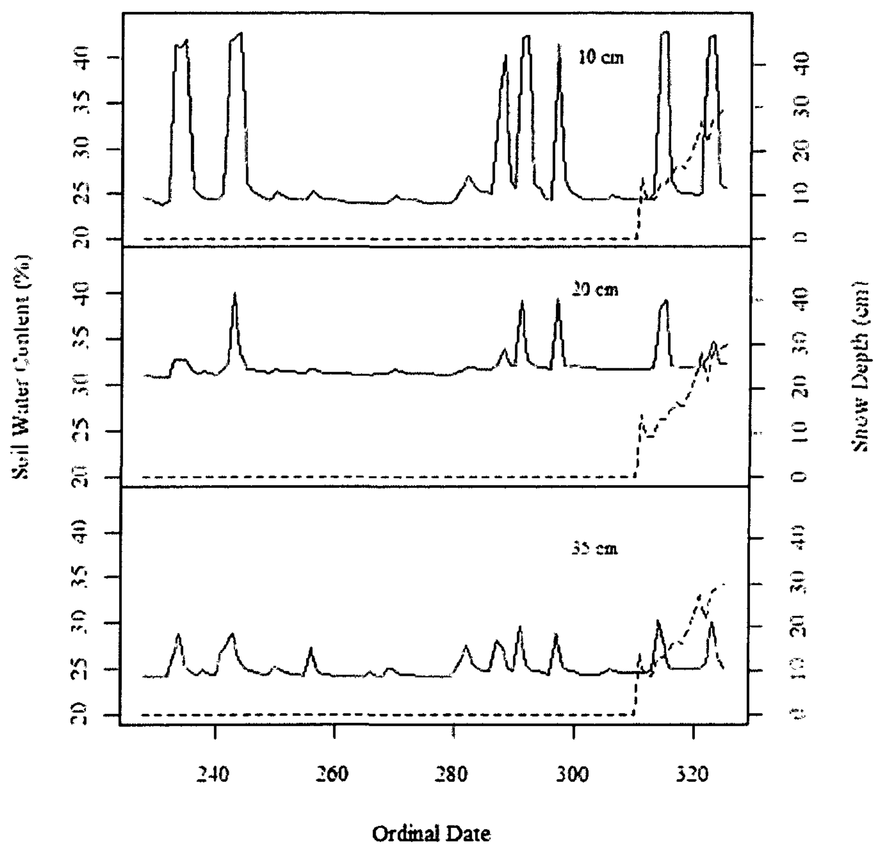

Figure 5.9- Snow depth (cm) (dashed line) and soil water content (solid line) 15 August 2008 to 20 November 2008 (OD 228 to 325).

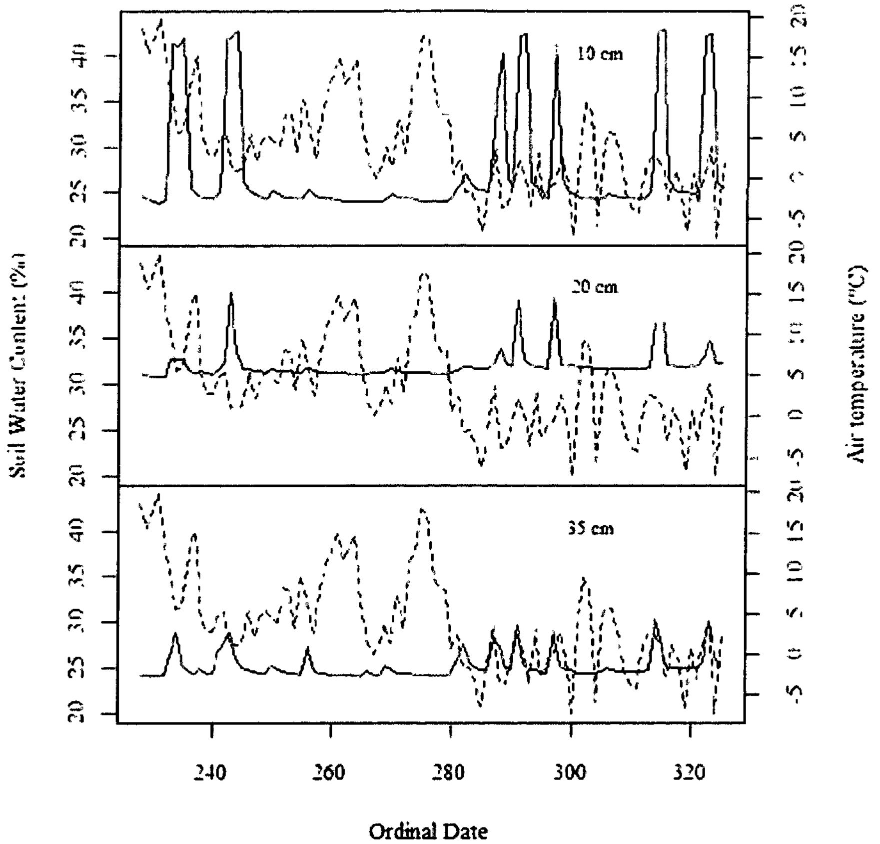

Figure 5.10- Soil water content (black) and daily average air temperature (dashed), 15 August 2008 to 20 November 2008 (OD 228 to 325).

In the antecedent period, there is significant correlation (at the 95% significance level) found between daily average soil temperature and air temperature (ρ = 0.65), soil temperature and snow depth (ρ = -0.83), snowfall and air temperature (ρ = -0.39), snowfall and soil temperature (ρ = -0.30) and between soil moisture at all depths (Table 5.1). Significant negative correlations exist between snow depth and soil moisture (ρ = -0.83) so that soil temperature decreases as snow depth increases. When over 20 cm of snow had accumulated, the soil temperature at 12 cm depth did not fluctuate by more than 0.5°C until snow cover decreased again in the melt period.

Table 5.1 Pearson correlation coefficients (upper diagonal) and p-values (lower diagonal) of daily averages for the antecedent period (15 Aug 2008 - 20 Nov 2008), italics denote significant p-values (at 95% confidence level)

	AT	ST	SD	RF	SF	SM10	SM20	SM35
AT		0.65	-0.43	0.20	-0.39	-0.08	-0.09	0.00
ST	*0.00*		-0.83	0.22	-0.30	-0.27	-0.32	-0.22
SD	0.01	*0.00*		-0.02	0.56	-0.26	-0.26	-0.31
RF	0.04	0.02	0.90		-0.10	-0.16	-0.13	-0.15
SF	*0.00*	*0.00*	*0.00*	0.29		-0.03	-0.01	-0.09
SM10	0.40	0.00	0.11	0.09	0.75		0.78	0.81
SM20	0.33	*0.00*	0.11	0.17	0.88	*< 2.2e-16*		0.82
SM35	0.98	0.02	0.06	0.12	0.32	*< 2.2e-16*	*< 2.2e-16*	

AT = air temperature (°C) ST = soil temperature (°C) SD = snow depth (cm)
RF = rainfall (mm day^{-1}) SF = snowfall (cm day^{-1})
SM = soil water content at 10 cm, 20 cm and 35 cm depth

5.6 Snowmelt Period

Soil temperature during the snowmelt period increased rapidly after the onset of snowmelt at a rate of approximately 1°C day^{-1} for about 10 days, until reaching a high for the snowmelt period of approximately 10°C at 12 cm depth (Figure 5.11).

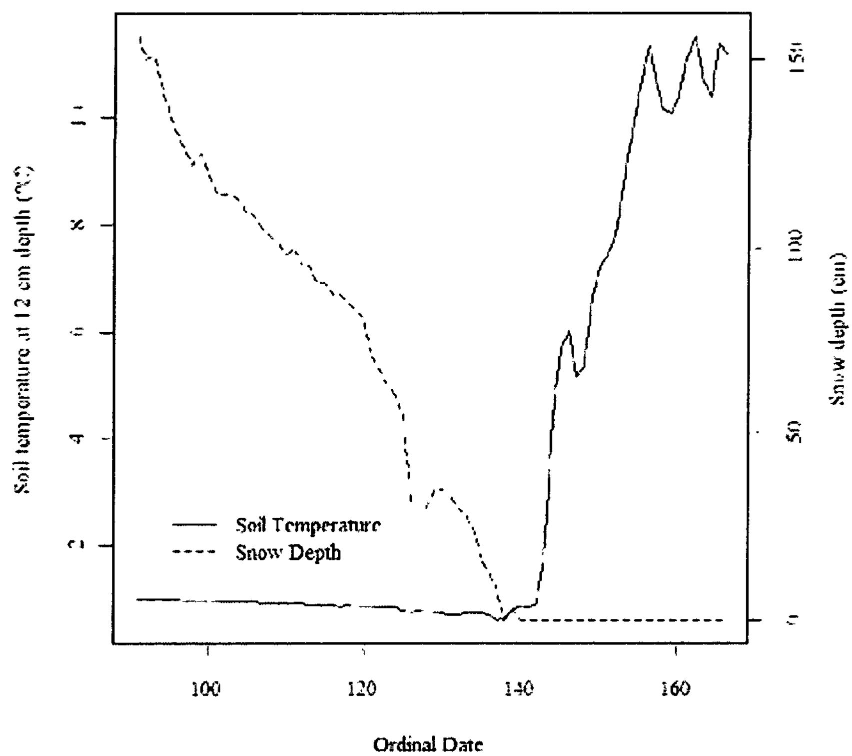

Figure 5.11- Soil temperature at 12 cm depth and snow depth 1 April 2009 to 15 June 2009 (OD 91 to 166)

Figure 5.11 shows daily average snow depth data from Barkerville along with soil temperature from Mt. Tom for the snowmelt period. Although it would be expected that snow at Mt. Tom would persist longer into the spring than at Barkerville, we do see that the soil at Mt. Tom shows a more rapid change in temperature when snow depth decreases to about 25 cm. This relationship is used to determine the date when snow cover leaves the surface at Mt. Tom.

The response to snowmelt is first observed around 8 April when there is a steep increase in soil moisture at 10 cm depth. There is no response observed at 20 cm or 35 cm depth at this time. This increase is accompanied by a liquid precipitation event and daily average air temperatures hovering near 0°C. Cooler daily air temperatures following the precipitation event could lead to freezing in the upper soil profile preventing melt water from infiltrating to greater depths.

We begin to notice an increase in soil moisture again on 17 April 2009 (OD 107) (see Figure 5.12). The increase is first noted at 10 cm depth and then a few days later (around 21 April, OD 111) at 20 cm and 35 cm depth. The daily average air temperature on 21 May (OD 141) was about 6°C with a maximum of about 9°C. This would have caused a considerable amount of melt, which is evident by the rapid increase in soil moisture seen at all 3 depths.

Between 27 April and 14 May (OD 117 and 134), the snow depth decreased from 87 to 27 cm. During this period, soil moisture at 10 cm depth was consistently higher than average at about 40% VWC. Soil moisture at 20 cm and 35 cm depths fluctuates during the same period, with values between 33% and 41% at 20 cm depth and between 25% and 40% at 35 cm. Daily average air temperatures in this period fluctuate between -5.8°C and 8.6°C with lower average air temperatures corresponding to lower soil moisture values in the lower depths.

The snow depth data from Barkerville show that snow had completely melted by 20 May 2009 (OD 140). The soil moisture data at Mt. Tom register higher than average soil moisture values up until about 4 June 2009. It is likely that there was still snow cover at Mt. Tom longer into the spring than at Barkerville.

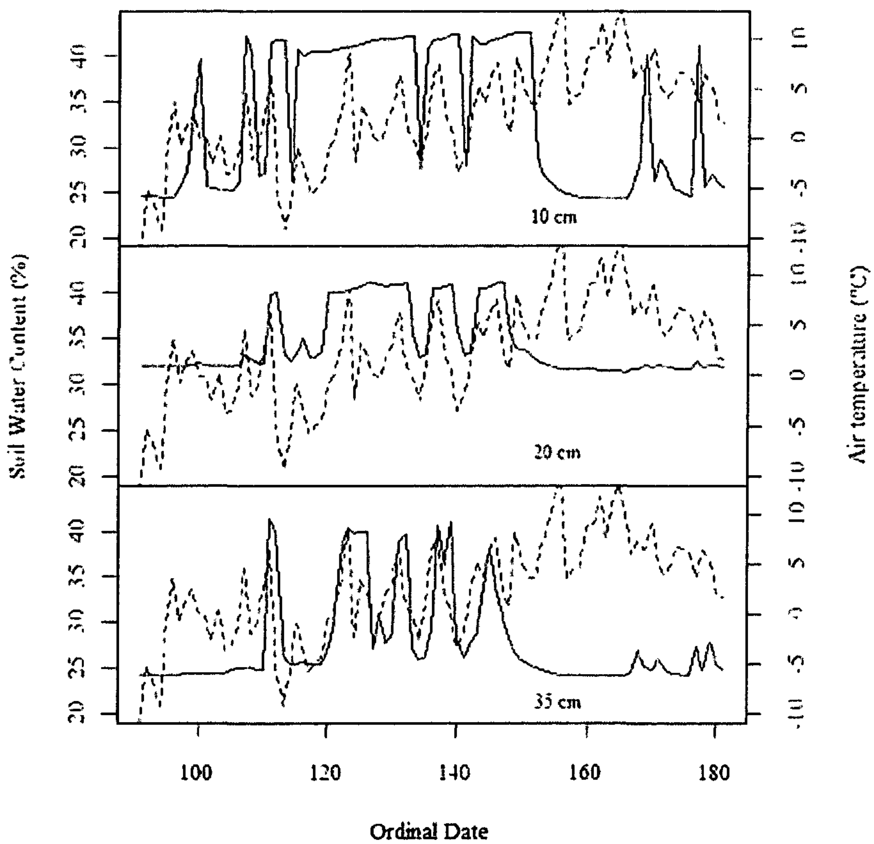

Figure 5.12- Soil moisture (solid) at 10 cm, 20 cm and 35 cm depth and air temperature (dashed) from 1 April 2009 to 15 June 2009 (OD 91 to 166).

Soil moisture is much higher than average during the snowmelt period and the response at each depth is different. While the snowpack ripens to 0°C, we may not necessarily see evidence of snowmelt in the soil moisture profiles, as melt water may refreeze lower within the pack. The top layer is the first to respond to snowmelt. In the top layer, we see a rapid increase in soil moisture on a few occasions when the air temperature is above 0°C for more than two consecutive days. The response is less pronounced in the second layer; however, the average VWC is higher at that depth. In the third layer, we do not see an increase in soil moisture until about 20 April 2008. The increase happens sooner in the upper levels: on about 10 April in the first layer and about 8 April in the second layer.

During the snowmelt period, there is significant correlation between soil temperature and air temperature ($\rho = 0.71$), air temperature and snow depth ($\rho = -0.44$), soil temperature and snow depth ($\rho = 0.94$), soil temperature and soil moisture at 10 cm ($\rho = -0.37$), at 20 cm ($\rho = -0.46$) and at 35 cm depth ($\rho = -0.48$), and snow depth and soil moisture at 10 cm ($\rho = -0.56$), 20 cm ($\rho = -0.73$) and 35 cm depth ($\rho = -0.71$). There is also significant correlation between soil moisture at 10 cm and 20 cm depths ($\rho = 0.84$), 20 cm and 35 cm depths ($\rho = 0.63$) and 10 cm and 35 cm depths ($\rho = 0.76$) (see Table 5.2).

Table 5.2- Correlation coefficients (upper diagonal) and p-values (lower diagonal) of daily averages for the snowmelt period (1 Apr 2009 - 15 Jun 2009, OD 91 to 166), *italics denote significant p-values (at 95% confidence level)*

	AT	ST	SD	RF	SF	SM10	SM20	SM35
AT		0.71	-0.44	0.12	0.01	0.03	-0.03	-0.15
ST	*0.00*		0.94	0.05	-0.26	-0.37	-0.46	-0.48
SD	*0.00*	*< 2.2e-16*		-0.30	0.39	-0.56	-0.73	-0.71
RF	0.26	0.66	0.03		-0.06	0.14	0.11	0.06
SF	-0.28	0.01	*< 2.2e-16*	0.58		-0.15	-0.11	-0.13
SM10	0.78	*0.00*	*0.00*	0.17	0.15		0.84	0.63
SM20	0.76	*0.00*	*0.00*	0.29	0.28	*< 2.2e-16*		0.76
SM35	0.16	*0.00*	*0.00*	0.57	0.23	*0.00*	*< 2.2e-16*	

5.7 Autocorrelation and Cross-Correlation

Figure 5.13 was generated using the autocorrelation function in R. The function calculates autocorrelation at a set lag interval of one or more days. The autocorrelation plots of soil water content (Figure 5.13) in the antecedent period show slightly different lag times at each depth. At 10 cm depth, the autocorrelation function is significant on day 2 and day 9, same as at 35 cm depth. There is negative autocorrelation at all depths, most notably at 10 cm and 35 cm depths.

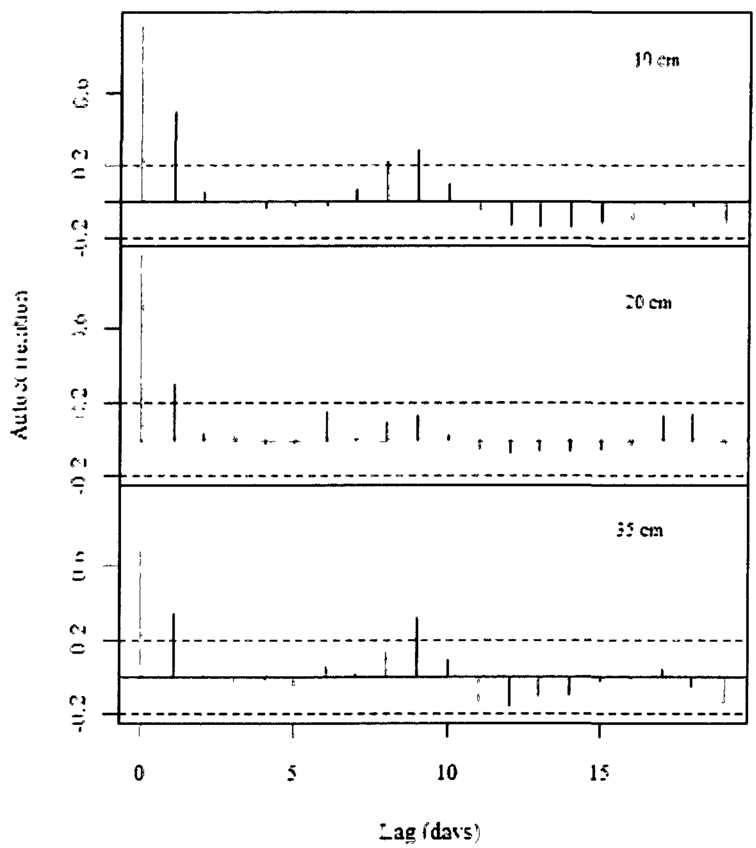

Figure 5.13- Autocorrelation functions of soil water content, 15 August 2008 to 20 November 2008, dashed lines indicate 95% confidence intervals.

The cross-correlation analyses for the antecedent period (Figure 5.14) show that the soil water contents at each depth are most highly related to each other at a time lag of one day at all depths and again at nine days between 20 cm and 35 cm and 10 cm and 35 cm depths.

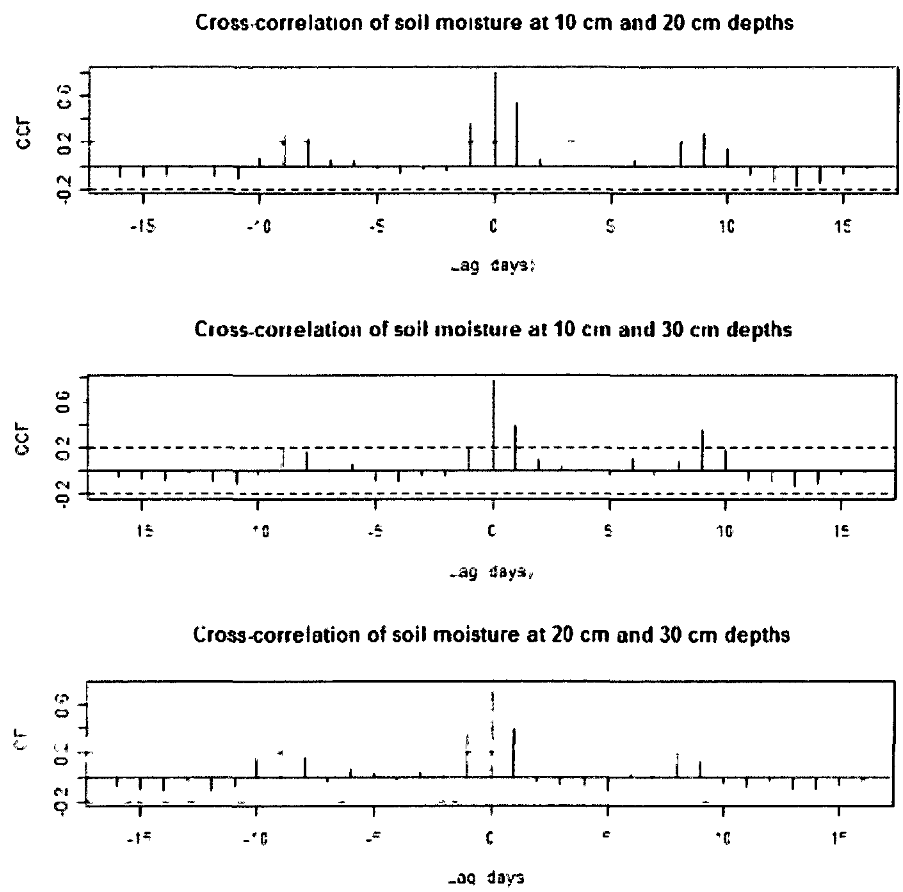

Figure 5.14- Cross-correlation of soil water content at 10 cm, 20 cm and 35 cm depth, 15 August 2008 to 20 November 2008, dashed lines indicate 95% confidence intervals.

Autocorrelation in the spring melt period is much higher than in the fall freeze-up, decreasing gradually over two weeks at 10 cm and 20 cm depth. At 35 cm depth the autocorrelation becomes insignificant at a lag time of 4 days, and again at a lag of 9-12 days. After a lag of 15 days there is negative autocorrelation at 35 cm depth. There is significant autocorrelation at 10 cm and 20 cm depths until up until a lag time of 13 days. After this point, the soil water content at these depths are no longer significantly correlated (Figure 5.15).

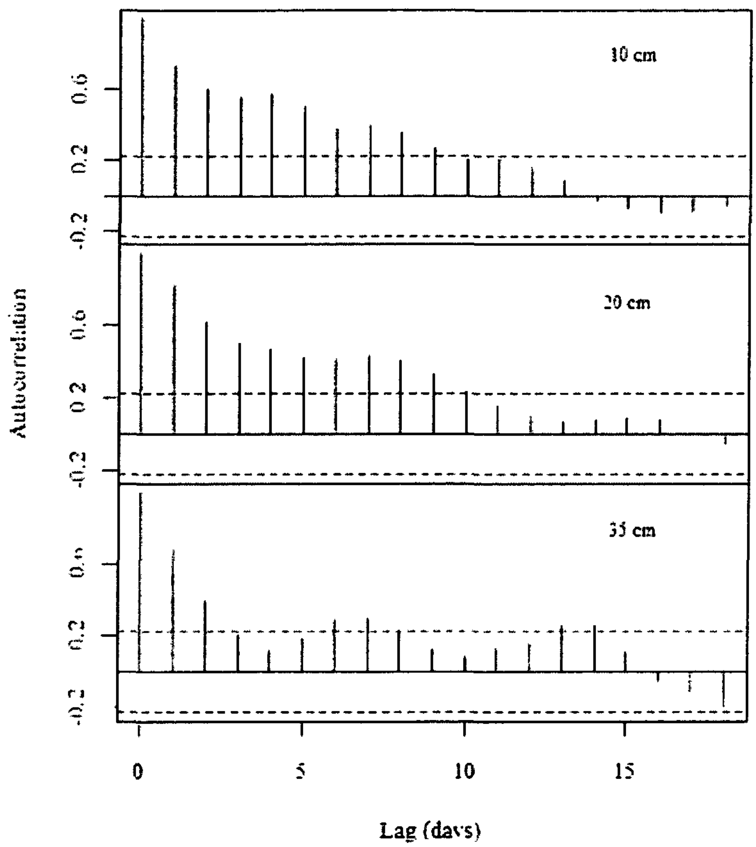

Figure 5.15- Autocorrelation functions (ACF) of soil water content at 10 cm, 20 cm and 35 cm depth, 1 April 2009 to 15 June 2009, dashed lines indicate 95% confidence intervals.

There is significant correlation between the soil water content at all depths at lag times from one to 10 days (see Figure 5.16). After 10 days, there is no significant correlation between soil water content at 20 cm and 35 cm depths; however, there is significant correlation again at a lag time of 12 to 15 days.

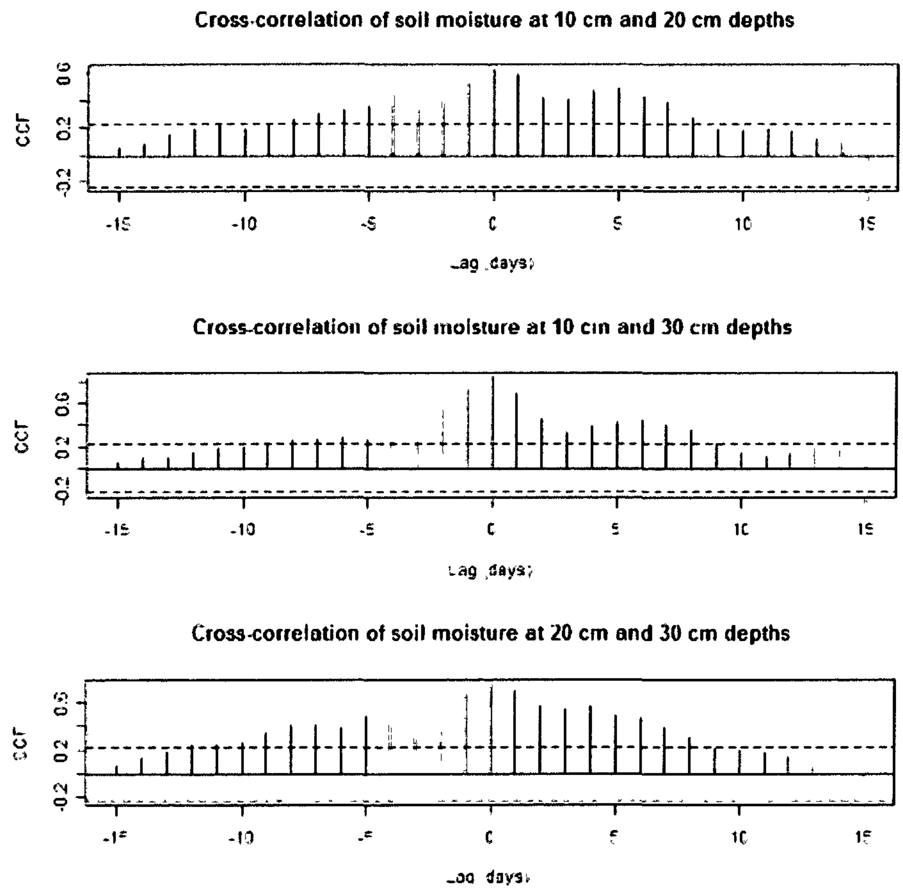

Figure 5.16- Cross-correlation of soil water content at 10 cm, 20 cm and 35 cm depth, 1 April 2009 to 15 June 2009, dashed lines indicate 95% confidence intervals.

For the year as a whole, there is significant autocorrelation at all depths up to a lag time of 22 days (see Figure 5.17). Also, we see a clear sinusoidal pattern to the autocorrelation function at all depths with autocorrelation being greatest at a lag time of about 7 days and again at about 17 days in the upper two depths, and at day 13 at 35 cm depth.

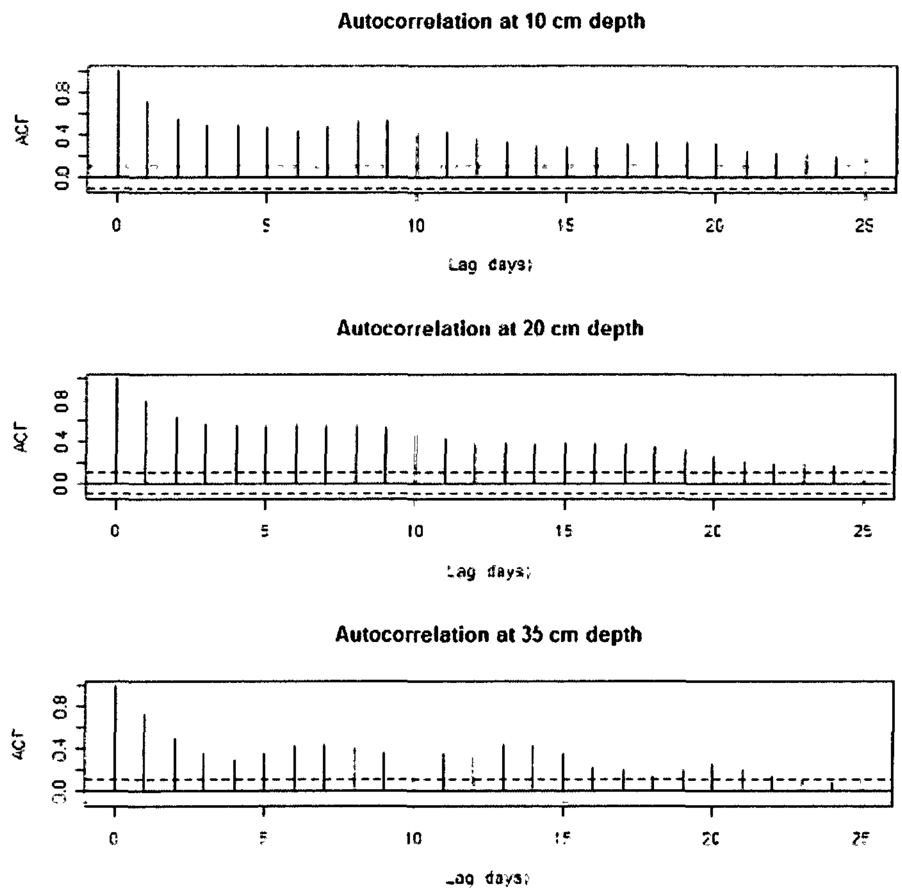

Figure 5.17- Autocorrelation of soil water content at 10 cm, 20 cm and 35 cm depth, 17 July 2009 to 14 July 2009, dashed lines indicate 95% confidence intervals.

There is also significant cross-correlation in soil water content between all depths up until 22 days, with autocorrelation gradually decreasing in a sinusoidal pattern. Peaks are more clearly delineated in the relationship between soil water content at 20 cm and 35 cm depth with higher significant correlation at lag times of 6 and 14 and about 20 days. In the cross-correlation of soil water content for the full time series at 10 cm and 20 cm depths, peaks occur at lag times of approximately 8 and 18 days, a little later than between 20 cm and 35 cm depth (see Figure 5.18).

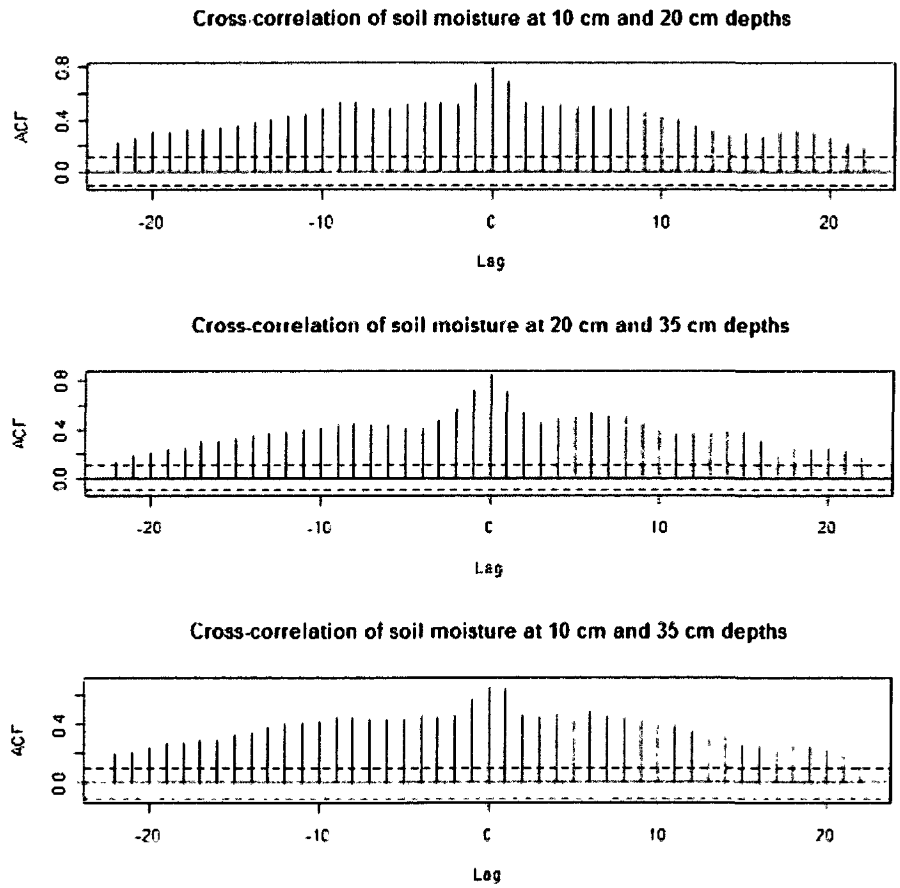

Figure 5.18 Cross-correlation of soil water content at 10 cm, 20 cm and 35 cm depth, 17 July 2009 to 14 July 2009, dashed lines indicate 95% confidence intervals.

5.8 Snow Water Equivalence and Snow Depth

The snow depth data collected from Mt. Tom are prone to error. The SR50A sonic ranger that measures depth consistently recorded negative values as well as spurious data. Also, as it records a depth to target, it was likely to register an increase in snow depth during summer when vegetation grows. For this reason, snow depth data were supplemented with data from Barkerville.

Data collected from a snow course survey done at Mt. Tom on 6 April 2008 show a SWE of 440 mm of water and an average snow depth of 153 cm in and around the clear cut where the meteorological tower was situated. This corresponds to a snow density of 0.290 g m^{-3}, approximately 30% the density of water. The snow depth measured at Barkerville for the same day was 130 cm. The maximum snow depth recorded at Barkerville was on 1 April at 156 cm. Twenty point measurements of SWE and snow depth were done at the clear-cut site. This leaves us with some confidence that we captured a reasonable representation of the maximum SWE at Mt. Tom for this season.

Chapter 6: Discussion

6.1 Persistence of Soil Moisture at Mt. Tom

In this study autocorrelation is used to determine persistence of anomalously high or low soil moisture values. Considering the definition and estimate of field capacity, we see that the soil moisture at Mt. Tom does not drop below field capacity at any time during the year. This means that the soils at Mt. Tom are generally wet, indicating persistence of wet conditions. Free draining soils tend to return to field capacity within two to three days of a rainfall event (Veihmeyer and Hendrickson, 1931). Significant autocorrelation in the soil moisture series beyond 3-4 days indicates persistence of anomalously high (low) soil moisture conditions at this site.

There is significant autocorrelation of soil moisture at the Mt. Tom site detected in the autocorrelation analysis of the daily average soil moisture data from the antecedent period, seen at about 8-9 days at 10 and 35 cm depths only. This is not to imply that the soils here are not wetter than other soils at similar elevation and latitude, but that anomalous moisture conditions are not persistent in the upper 35 cm of the soil profile. Significant autocorrelation of soil moisture persists for about two weeks looking at the entire time series from 17 July 2008 to 14 July 2009. Comparing the antecedent fall period with the spring melt season we see that soil moisture does not exhibit the same pattern of autocorrelation in the fall as compared with the spring. In the spring, autocorrelation of the soil moisture series is significant up until about 10-13 days, depending on what depth is being considered.

In the fall, there is no significant autocorrelation in the soil moisture series past about two days. This indicates that there is little to no persistence of anomalously high soil moisture at this time of the year. For the year as a whole there is significant autocorrelation up until about 20 days or so. These values are likely biased by the winter time data. In the winter, the soil moisture and soil temperature values are static as the soil becomes decoupled from atmospheric processes by snow cover. The autocorrelation coefficients of soil moisture through the winter are therefore higher (i.e. close to unity) than in other seasons when there is no snow cover. When the autocorrelation coefficients are calculated for the year as a whole the winter values have a strong influence on the calculated coefficients for the remainder of the year.

The persistence of significant autocorrelation observed in the snowmelt period is attributed to the melting of the snow pack, which takes place from mid-April to 20 May 2009. After this date, soil moisture values remain well above average and slowly decline until approximately 4 June 2009. The snow cover at Mt. Tom had melted by 20 May 2009 so we see that there is some persistence of the snowmelt signal in the soils at this site. Sampling to greater depths would allow us to assess whether or not the signal propagates the soil or whether it dissipates at shallow depths. It is expected that soil moisture infiltrates to greater depths than was observed at Mt. Tom. The snowpack delivers a continual supply of water to the soil during the melt period and hence it is this supply that accounts for the persistence observed in the spring melt season, and not the soil properties themselves.

Most studies on the persistence of soil moisture that utilize ground based data have focused on agricultural environments, with well developed soils such as in Illinois (Wu et al. 2002), Oklahoma (Deliberty and Legates 2008), Orgeval, France (Lauzon et al. 2004), and southern Quebec (Parent et al. 2006). These studies sampled soil moisture to greater depths than was possible at Mt. Tom and hence, give a much better picture of infiltration processes and water storage.

There are limited studies on soil moisture persistence in mountainous terrain but those that exist highlight the significance of terrain features on spatial variability of soil moisture. One notable study by Williams et al. (2009), working out of the Dry Creek Experimental Watershed near Boise, Idaho, attempted to characterize the temporal and spatial variability in a semi-arid mountainous area. The authors show that the amount of snow present on the surface in the spring as well as terrain and soil properties are significantly related to soil moisture in the catchment with wet (dry) areas staying wetter (drier) than average throughout the year.

The work of Williams et al. (2009) agrees with previous work done by Litaor et al. (2008) in the high-elevation alpine tundra region of Niwot Ridge, Colorado. Litaor et al. (2008) also found that soil moisture was significantly correlated ($\rho = 0.7$ at the 99% significance level) with snow accumulation and terrain factors. The Niwot Ridge study was focused more on the controls of species diversity of herbaceous plants as opposed to soil moisture persistence in particular; however, soil moisture is a limiting factor of plant growth and hence very important to plant diversity, especially in a changing climate where ecotones are expected to change.

A study by Szeto (2002) suggests that precipitation recycling, and hence soil moisture persistence, is significant in the Mackenzie River Basin (MRB). This is a northern watershed spanning a large geographic area stretching over 15 degrees of latitude and approximately 20% of Canada's landmass. What is interesting to note from this study is that the spatial distribution of the mean seasonal precipitation recycling ratio is lowest over mountainous terrain to the west of the basin, but that these areas correspond to the highest estimated annual runoff ratios for sub-basins of the MRB. This hints at the importance of monitoring the contribution of snowmelt from mountainous regions to the tributaries and sub-basins at lower elevations as it is this contribution that is responsible for the increase in moisture availability in the period with the strongest recycling ratios (April to June).

Our study at Mt. Tom did not assess the same factors as the Szeto (2002) study that focused on atmospheric conditions as opposed to terrestrial conditions. Having similar seasonal climates, it is possible that if these data were available and similar analyses were performed for the Fraser River Basin, similar results as seen in the Mackenzie River Basin would be observed.

Downstream from the Mt. Tom area are hemlock and cedar forests that are wetter environments year round, compared to the higher elevations. This area is referred to as the Interior Cedar-Hemlock (ICH) zone. These species have a higher demand for water than the spruce and fir species that are characteristic of the ESSF zone at Mt. Tom. The moisture received from higher elevations sustains the water need of the trees in the ICH zone through the spring and summer as there is continual supply of soil moisture from snow and ice melt. Steeper topographic gradients contribute to easy movement of melt water through the soil and over the surface via gravity.

6.2 Factors Affecting the Persistence of Soil Moisture at Mt. Tom

There is a number of factors that could affect the persistence of soil moisture at Mt. Tom. More importantly, there are several factors that could affect the detection of such persistence that are associated with the experimental design and analyses chosen to describe the data. These are discussed in the following sections.

6.2.1 Slope and Steepness

Slope might affect movement of meltwater into low lying areas so that higher soil moisture values do not necessarily coincide with areas of greater snow depth (Williams et al. 2009). Steepness of the slope would contribute to redistribution of snow and water during the melt season with a tendency for meltwater to pool in low lying areas (Tong et al. 2009; Williams et al. 2009). Such is the case at Mt. Tom where waterlogged soils are found in low lying areas and along natural and man-made drainage routes.

The lack of persistence in soil moisture during the antecedent period at this site might in part be attributed to the steep gradient of the surrounding area. Although the meteorological station was set in a relatively flat area with a grade of about 3°, the surrounding terrain is steep and the tendency is for the ground to slope away in a southwest direction from the station. Even a 1° slope will encourage good drainage (Parent et al. 2006). Steep terrain will promote greater surface runoff and drainage through the subsurface via gravity. Differences in snow accumulation on south facing versus north facing slopes also limits the contribution of moisture from snow to soil.

Slope aspect would also affect soil moisture variability as north facing slopes are generally cooler and wetter than south facing slopes, which receive more sunlight. Snow takes longer to melt on north facing slopes so we would expect that snowmelt affects soil moisture on north facing slopes for a longer period compared to south facing slopes, in the Northern Hemisphere. Observations of radiation on different slopes and aspects from 0° to 60° latitude north show that radiation varies through the year and that slope and aspect both play significant roles in radiation budgets. For example, a north facing slope of 0° at 50° latitude will receive an average of 618 Cal cm^{-2} day^{-1} on both north and south facing slopes, compared to 485 Cal cm^{-2}

day^{-1} on a north facing slope and 710 Cal cm^{-2} day^{-1} on a south facing slope of 15° grade. On December 22 at the same slopes would receive 105 Cal cm^{-2} day^{-1} at 0° grade compared to 3 Cal cm^{-2} day^{-1} on the north facing slope of 15° and 211 Cal cm^{-2} day^{-1} on a south facing slope of 15° (Buffo et al. 1972).

To date, most studies of soil moisture have been carried out in agricultural areas (see for example Lauzon et al. 2004; Parent et al. 2006; Wu et al. 2002). These surfaces are relatively flat and do not necessarily drain as easily as some of the soils at Mt. Tom. Topography affects the movement of water through a watershed and there is likely to be greater movement of water within the soil if the terrain is sloping in any direction.

A recent study by Williams et al. (2009) conducted in a small, snow-dominated, mountainous region attempted to determine the relative importance of topographic controls on snow distribution and subsequent soil moisture. The authors found that static properties such as slope and soils control the spatial variability of snow and soil moisture and that the timing and form of precipitation influences their spatial variability throughout the year. It is noted that ridge tops and steep areas have lower than average soil moisture throughout the year agreeing with results found by Litaor et al. (2008) who found a significant correlation (r^2=0.9, at the 99% significance level) between SWE and terrain factors that limit exposure and degree of shelter. This is not to be confused with the previous coefficient reported relating soil moisture with snow accumulation.

6.2.2 Surface and Atmospheric Conditions

The study area is transitory between wetter environments at lower elevations and drier, alpine tundra environments at higher elevations. As well, it is between the wetter, more humid coastal climate to the west and the drier, inland environments to the east and north (see Figure 3.2). Studies such as the Global Land Atmosphere Coupling Experiment (GLACE) have shown that it is in areas that are transitions between wetter and drier environments where soil moisture plays the greatest role in local precipitation feedbacks (Koster et al. 2004). In these areas, soil moisture data may be used to make better predictions of summer precipitation (Eltahir and Bras 1996; Jacobs and DeBruin 1992; Kochendorfer and Ramirez 2005; Pal and Elthair 2001) and summer air temperatures (Durre et al. 2000; Fisher et al. 2007; Jaeger and Seneviratne 2010; Lorenz et al. 2010).

The fraction of precipitation variance that is accounted for by differences in soil moisture values (Ω - dimensionless) and the rate of evaporation (cm day^{-1}) are used to determine how soil moisture contributes to predictability of local meteorological conditions (Koster et al. 2004). Areas where soil moisture contributes to greater predictability of precipitation have high values of both Ω and evaporation. These correspond to areas of intermediate soil wetness where the degree of saturation is between approximately 20-40% (Koster et al. 2004). The Mt. Tom area has average soil wetness values in this range (between 0.26 and 0.33) and is also within or adjacent to an area of low to moderate land surface-atmosphere coupling. The areas in question in the study by Koster et al. (2004) would be of greater spatial extent than the Mt. Tom area. Topographic controls become less apparent when considering larger expanses of land versus small areas that would be represented by a single monitoring point in a vast area like Mt. Tom.

In areas where soil moisture shows persistence of at least three months, land surface-atmosphere coupling may lead to significant local precipitation recycling (Robock et al. 2000). It has also been found that soil moisture affects near surface air temperatures such that drier soils are generally associated with higher daily maximum temperatures (Durre et al. 2000; Lorenz et al. 2010) and vice versa. The positive feedback whereby drier (wetter) soils cause warmer (cooler) air temperatures to persist is caused by evapotranspiration (Durre et al. 2000; Fisher et al. 2007; Lorenz et al. 2010). These effects are best understood via the Bowen ratio and the separation of latent and sensible heat fluxes. In this way, the persistence of soil moisture conditions, wet or dry, directly affects air temperatures and hence boundary layer conditions that contribute to convective precipitation (Fisher et al. 2007).

Pal and Eltahir (2001) describe the mechanisms for the positive feedback between soil moisture and precipitation. Higher than average soil moisture increases the moist static energy flux, reducing the height of the boundary layer, which in turn increases the moist static energy per unit mass of air. This results in an increase in convective rainfall.

With global climate change, precipitation is expected to fall more often as rain as opposed to snow (Barnett et al. 2005). This may have serious implications in snow-dominated areas where snow accounts for a significant input of water to the surface. Timing and the form of precipitation could mean less infiltration to soils, as runoff is expected to increase if precipitation occurs more frequently as rain (Barnett et al. 2005; 2008). The effect of rain-on-snow through the winter could lead to an earlier snowmelt season. With an earlier melt, we would expect soils to be drier earlier in the spring/summer. There is potential for the feedback to be amplified by the warmer air temperatures projected by climate change models.

Surface conditions change the storage capacity of the soil as well as albedo, directly affecting evaporative rates and turbulent fluxes at the surface (Jacobs and DeBruin 1992; Murray and Buttle 2005). Vegetation on the surface determines water demands for evaportranspiration, another important component of the surface water budget. The relationship between soil moisture, vegetation and evapotranspiration affects water availability and storage on the surface, directly influencing regional boundary layer conditions (Jacobs and Debruin 1992).

Although long term persistence of anomalous soil moisture conditions was not detected at this site, it might be possible that we would find different results if nearby forested areas were instrumented. Plants and forest litter retain water and shade from trees, preventing heating at the surface. Soil moisture measurements taken from a soil profile within a forest plot would likely be quite different from those taken in the cutblock at Mt. Tom. Murray and Buttle (2005) report greater infiltration in harvested versus forested slopes. They acknowledge that harvesting of timber contributes to greater runoff to streams and increases erosion.

The MPB epidemic could have similar effects on runoff and erosion (Rex and Dubé 2008). With fewer trees to take up water from the soil it is likely that soil moisture will increase and that runoff to streams will increase as well. Dead roots will not retain soil to the same degree as living roots with mycorhizae and organisms that associate and/or connect with roots and soil.

Additional instrumentation would provide better insight into the different factors that affect land-surface interactions in the Cariboo Mountains. Local evaporation in this area is subject to change due to land use (deforestation) and climate change, which is expected to yield warmer average air temperatures and hence less snowfall as precipitation will be more likely to fall as rain (Barnett et al. 2005; 2008).

6.2.3 Subsurface Soil Conditions

6.2.3a Antecedent Moisture Conditions

Antecedent soil moisture conditions impose limits on the rate of infiltration of meltwater in the spring. A drier than average soil will take longer to 'wet-up' than a soil that is primed by antecedent moisture (Williams et al. 2009). This phenomenon is referred to as soil moisture hysteresis and is dependent on factors such as particle size and porosity, antecedent soil moisture, air entrapment and liquid-solid interactions.

The study at Mt. Tom did not really allow us to truly characterize the long-term soil moisture normals at the site but we did have some data from the previous years, measuring average moisture of the upper 30 cm. The probe was deployed in a vertical orientation, as opposed to the current set-up with three probes inserted horizontally at 10, 20 and 35 cm depths. The data from the previous set-up is in agreement with the data from this study, with similar average soil moisture observed for the 30 cm profile as for all three depths.

6.2.3b Soil Physical Properties

Hydraulic conductivity depends on the structure and degree of saturation of the soil. A dry soil will transmit water differently than one that is moist (Williams et al. 2009). This site and the surrounding areas have been disturbed by tree harvesting. This activity has likely altered the natural flow of water within the management area by contributing to compaction and by changing the surface cover. Also, compaction from machines and logging activity has potential to reduce porosity in the upper layers of the soil profile which could contribute to decreased infiltration. The instruments at Mt. Tom were installed in a cut-block site where the surface had been disturbed by machinery. Tracks in the soil and debris on the surface contribute to uneven ground, pooling in low areas and compaction of the upper soil layers. The soils at the site of the meteorological tower are turned over, furrowed and or packed down in areas, contributing to changes in the flow of water through this area.

Subsurface soil conditions contribute to the rate of infiltration of water into and through the soil. The depth of the root zone and soil physical properties such as hydraulic conductivity, limit how easily water moves within the soil profile. Land use and forestry management practices can greatly affect subsurface soil conditions. In general, deforestation alters surface conditions by denuding the surface and compacting the soil (Sidel et al. 2006). These practices have the effect of decreasing infiltration (increasing surface runoff) by decreasing the hydraulic conductivity of the soil as soil aggregates become compacted and less permeable to water.

We see in our results that soil moisture is generally higher in the second layer at Mt. Tom than it is in the upper and lower layers (refer to Figure 5.8). It is possible that water is more easily retained in this layer, or that there is some bias in the instrumentation. The second layer in the soil profile from about 10-15 cm depth was classified as a Bf or Bhf horizon, meaning that it is enriched with aluminum and organic material. It is generally accepted that organic material contributes to higher water retention in soils.

The layer beneath was classified as a Bm horizon, implying that it has been modified by chemical and/or physical weathering processes. A sub-angular blocky structure would account for a greater capacity to retain and transport water, compared to the second layer. Sub-angular blocky structure within the soil leaves gaps and channels for water to be transported with ease. The second layer at the site had fewer aggregates and less pore space for easy transport of water. It is likely that these physical characteristics contributed to the higher soil moisture values measured at this depth.

6.2.4 Time Scale

Data in this study were analysed on a daily time scale, to give daily average soil moisture values. An hourly time scale would provide better insight into changes in soil moisture throughout the day as opposed to a week or a month. Likewise, monthly data would remove daily variability and smooth out fluctuations from individual precipitation events. Looking at the data on an hourly time scale would allow for better characterization of infiltration rates whereas a monthly time scale would better characterize the annual cycle of soil moisture. The effects of time averaging are also of importance when interpreting results. For example, averaging the 15 minute data to find daily average soil moisture has the effect of smoothing the data so that diurnal fluctuations in soil moisture are removed.

Researchers use different time scales to characterize the progression of soil moisture depending on the processes they are interested in observing. For example, to examine the effects of transpiration on the soil water budget, Daly et al. (2004) employ a model that works on an hourly time scale. Daily, weekly or monthly time scales do not capture the processes that are most significant to transpiration (Daly et al. 2004). Koster and Suarez (2004) suggest that the timescale used to model land-surface feedbacks affects the outcome of the model, citing that any given variable could be affected by the temporal variability of other factors.

Parent et al. (2006) observe the temporal variability of soil moisture on timescales from one hour to two weeks during the growing season from mid-June to August 2003. They found that the greatest variation (85%) in near-surface soil moisture at their site in southern Quebec (latitude 46°N, longitude 72°W) was on the order of 48 hours to two weeks. Their site is on a crop field in a temperate climate, not necessarily comparable to the Mt. Tom site but this study does demonstrate that the timescale used for sampling and analysis limits the information that can be gathered from tests of persistence.

The frequency of observations along with the spatial distribution of sampling points limits the type of analysis we can do and the objectives we can set for the project. Wu et al. (2002) employ a data set of soil moisture from 17 sampling points in Illinois, 11 depths at each site spanning 16 years. These data allow them to characterize the response of soil moisture to precipitation on a long-term basis.

6.3 Limitations of the Study

6.3.1 Length of Time Series

The results presented in this study are based on only one year of data. A longer time series of soil moisture would facilitate a more robust analysis of trends and any long-term periodicity in soil moisture. Unfortunately, budget constraints and time limits did not allow for instrumentation of nearby forested plots or for longer soil moisture instrumentation and hence annual and spatial variability are poorly represented by the results in this study. There are no supplemental soil moisture data available from the nearby Barkerville weather station and soil moisture is not typically measured at other meteorological weather stations in the region.

Having only one year's worth of data prevents assessment of long term variability of soil moisture. We do not see how anomalously wet or dry years affect persistence and we are unable to assess whether soil moisture conditions in the fall affect persistence in the spring and early summer. Without a longer series, we are unable to assess how differences in the amount of annual snowfall are manifested in the soil profile. The amount of annual snowfall and air temperature throughout the melt season affect the duration of the snowmelt period, hence directly affecting the persistence of soil moisture in the spring.

Other studies with longer time series are able to detect how snow cover and soil moisture affect large-scale teleconnections (see for example Barnett et al. 1989; Cohen and Saito 2003; Déry et al. 2005a; Meehl 1994; Robock et al. 2000), and precipitation recycling (Bosilovich and Chern 2006; Brubaker et al. 1993; Eltahir and Bras 1996). Many of these studies are made possible by data generated through reanalysis of meteorological observations (see for example Robock et al. 2000) while other datasets are generated through climate simulations modelling with land-surface models and general circulation models. With the development of new technologies in the field of radar and satellite imagery, soil moisture data will be available at greater resolutions and longer timescales (Vicente-Serrano et al. 2004). Remote sensing is a useful tool in determining these data at better resolution over both space and time.

6.3.2 Spatial Variability

Spatial variability of soil moisture can vary considerably within any environment; however, in a mountainous watershed, topography imposes limitations of the spatial distribution of snow, water, and subsequently soil moisture (Tong et al. 2009). Studies in the Dry Creek Environmental Watershed near Boise, Idaho have shown that the spatial distribution of soil moisture is significantly correlated to the spatial distribution of snow, slope, soil texture and soil depth (Williams et al. 2009). As soil moisture is a limiting factor for plant growth, any changes in its spatial variability are important to ecosystems, having the ability to change vegetation regimes and essentially the soils themselves (Litaor et al. 2008).

In the Mt. Tom study there was only one point measurement of hydrometeorological conditions, making it difficult to assess spatial variability in the area. It is also difficult to say that the site chosen was representative of the entire area, being set in a clear cut surrounded by other patches on varying slopes and aspects. Visual observations confirm that there are areas with waterlogged and well drained soils. Figures 3.3 and 3.4 show examples of relatively well drained soils, with Figure 3.4 demonstrating that rockier, more well drained soils are drier and have deeper rooting zones than soils that have fewer rocks and pebbles. Figure 6.2 demonstrates how soils can vary over short distances, making it difficult to rely on point measurements to capture anything more than temporal variability in specific conditions.

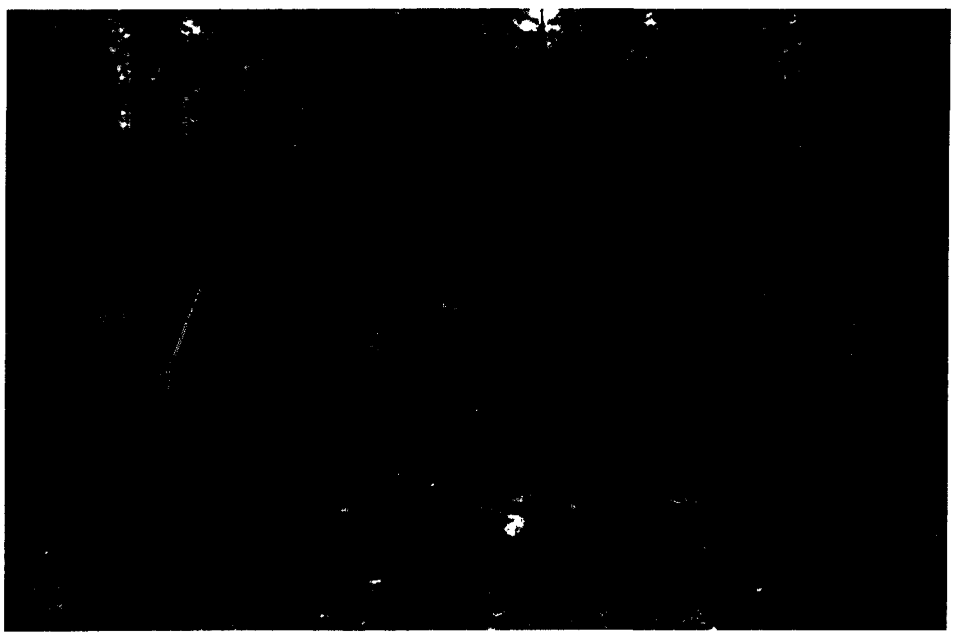

Figure 6.1 Soil profile as observed along logging road cutblocks in the Mt. Tom Forest Management Area, photo taken by Paul Sanborn, June 2007

Due to financial and time constraints, data were not obtained from nearby forested areas or from opposing aspects, except on one outing where cores were taken with an auger to assess soil moisture using the oven dry method. Samples taken with the corer were from variable depths and did not represent soil moisture conditions for each depth separately. Also, there was only one set of point measurements taken during the course of this study, leaving nothing for comparison between seasons or time.

A network of monitoring stations throughout the area would allow for some assessment of spatial variability of soil moisture; however, the equipment is costly and this is not feasible. Remote sensing has emerged as a resource to fill these missing data. With further technological advances, more reliable and current data will become available. Such is the mandate of the NASA Soil Moisture Active Passive (SMAP) mission that uses satellites with high-resolution radar and radiometer measurements to map soil moisture and freeze-thaw states over large expanses of land (see http://science.nasa.gov/missions/smap/) for more detail).

The role of soil moisture in forcing local climatic conditions is apparent in studies that utilize land surface schemes and general circulation models to project and model past and future changes in climate. Soil moisture affects evapotranspiration and hence boundary layer conditions that contribute to precipitation. In areas where precipitation is limited by local convective processes as opposed to advection of moisture from adjacent regions, monitoring soil moisture is of greater importance than it is in areas where it does not limit evapotranspirtation, such as rain forests and deserts.

6.3.3 Depth of Sampling

It is difficult to say whether or not the data gleaned from the CS616 probes are of any use in describing actual soil moisture levels in each horizon. One would expect to see more of a lag in soil moisture with depth and also decreasing soil moisture content with depth. Soil horizons were difficult to distinguish and although there are clear cuts showing well developed horizons, we must remember that the soil at the site was disturbed by logging and planting and therefore did not display the same characteristics as would be seen in surrounding soils.

Other studies that utilize ground based data have sampled much greater depths than was possible at the Mt. Tom site. A study by Wu et al. (2002) used data that samples to 2 m depth from 17 stations around Illinois. The data were taken from 11 layers and span from February 1981 to August 1996. These soils are deeper than would be expected at Mt. Tom as they are in an agricultural area, but being able to sample to greater depths would allow detection of how persistence dissipates with depth. The data from Wu et al. (2002) clearly show how soil moisture changes with depth and how wet and dry signals are manifest in the soil. The results demonstrate that soil moisture at depth is less variable than near the surface (amplitude damping). Also, we see that the timing of lowest and highest average soil moisture values shifts on a monthly timescale with depth (phase changing).

There are groundwater data available for Barkerville from 1967 to present. These data may be useful for relating ground water and soil moisture fluxes to determine how well they are connected. The groundwater readings are taken from a well at about 1525 m elevation so that readings are giving the depth to water level below ground surface (bgs). The lowest level recorded in the well was 13.1 m (bgs) for April 1980. The highest level recorded was 3.2 m (bgs) in May 1992. These data are available online at http://www.env.gov.bc.ca/wsd/data_searches/obswell/wellindex.html. Long term soil moisture would be needed to assess the contribution of soil moisture to groundwater and vice versa. Flow directions and dynamics would also need to be investigated to see whether there is a causal relationship between soil moisture and groundwater levels.

Data from another study set in mountainous terrain employed ground based data taken with Campbell Scientific Time Domain Reflectometers (TDR) from five depths at 5, 14, 45, 75 and 105 cm (Williams et al. 2009). This is not as deep as the Illinois study and these data were obtained from only one site in the catchment area. Other measurements were taken with a portable TDR which probes the soil to no more than 30 cm depth to assess spatial variability of soil moisture in the upper soil profile.

Parent et al. (2006) used similar TDR probes set up at seven sites along a 90 m transect to measure soil moisture at 20 minute intervals from 5-25 cm depth. They used these data to characterize the temporal variability of soil moisture at the surface on short timescales ranging from one hour to two weeks. Another study by Lauzon et al. (2004) used data from three rows of instruments, two rows with instruments at 5, 15, 25, 35 and 45 cm depths and one row with instruments at 55, 75, 95, 115, 135 and 155 cm depths also to characterize soil moisture conditions on a monthly to yearly scale.

Chapter 7: Conclusion

Soil is the interface between the land and the atmosphere. Processes occurring at the surface link the hydrological and energy budgets, directly affecting local weather and climate. Often, large scale surface anomalies in one area may persist via feedback processes, eventually affecting conditions in nearby or remote areas. For example, snow depth over Eurasia is known to affect the Indian summer monsoon. These lagged effects, whereby conditions in one region impact conditions in another region and season, are referred to as teleconnections.

Apart from these large scale teleconnections, there exist links between conditions at much smaller scales. Within a basin, local precipitation recycling could account for a significant portion of surface water input to the atmosphere. Convective processes depend on moisture and heat exchange at the surface. Local meteorological conditions are largely result of climate and surface conditions that contribute to heat and moisture fluxes. Precipitation regimes are expected to be altered by climate change and projections show that climate and vegetation will change as well. With these changes, precipitation recycling could become more significant in areas where precipitation occurs on a more seasonal timescale.

In snow dominated regions, the timing and magnitude of surface water input from snowmelt is of great importance to vegetation and water supplies. Soil water conditions in the fall freeze-up can limit infiltration of meltwater in the spring, subsequently affecting moisture conditions into the spring and summer. In mountainous terrain, the distribution and amount of snow play a large role in local water budgets. A significant portion of the world population relies on water from mountain streams and rivers fed from snow and ice. As climate change alters precipitation regimes, the input of water from snow will change in many regions. Understanding how these changes influence water availability, agriculture and other land uses is becoming more important as we learn to adapt to climate change.

This study was set in the Cariboo Mountains, British Columbia. The data come from the Mt. Tom Forest Management Area where the Northern Hydrometeorology Group deployed a meteorological station measuring air temperature, relative humidity, soil temperature at 12 cm depth, snow depth, precipitation, wind speed and direction. Soil moisture was measured at 10, 20 and 35 cm depths. Precipitation and snow depth data from the tower at Mt. Tom were deemed unfit for analysis due to instrument error and were supplemented with data from the nearby Environment Canada tower at Barkerville. The data from the SR50 were sporadic and did not agree with snow depth data from Barkerville. Abrupt changes in snow depth at Mt. Tom do not correspond to other meteorological factors and hence these data were not used in this thesis.

Reliable meteorological data for Mt. Tom were obtained from 17 July 2008 to 14 July 2009. The meteorological tower was dismantled on 16 July 2009. Data from the study period at Mt. Tom were compared to climate normals (1971 to 2000) from Barkerville to characterize the season with respect to average climate in the area.

The climate in the study area is wet and cold with annual average air temperatures around 1.9 °C and about 1000 mm yr^{-1} of precipitation. Nearly half of the average annual precipitation in the area falls as snow, with an average 480 mm yr^{-1} (from 1971 to 2000) reported as snow water equivalent. The area is within the Engleman Spruce-Subalpine Fir biogeoclimatic zone, which is described as a wet and cold climate with forests dominated by spruce and fir trees as well as willow bushes. The 2008 to 2009 study period saw slightly higher than average snowfall and cooler temperatures in the mid to late winter season.

7.1 Temporal Variability of Soil Moisture at Mt. Tom

The response of the soil to precipitation events during the course of the summer and fall shows that the soils here are easily drained. After a precipitation event, soil moisture at all three depths returns to average values after approximately 2-3 days. This is likely due to the steepness of the terrain (with approximately 3° grade at the site and steeper on the surrounding slopes) as well as the structure of the soil, which is rocky, allowing for easy drainage along rock faces and channels in the soil. Water collects in low lying areas and runs off into streams and creeks that drain into the Willow River, part of the Fraser River Watershed.

Soil moisture is generally constant at Mt. Tom with average soil moisture of 0.28 at 10 cm depth, 0.33 at 25 cm and 0.26 at 35 cm depth for the period from 17 July 2008 to 14 July 2009. These values do not change significantly when comparing the data from the entire year with data from the antecedent period. In this time period, the average soil moisture content is 0.26 at 10 cm depth, 0.32 at 25 cm depth and 0.25 at 35 cm depth. Soil moisture is higher in the melt period compared with the year as a whole or compared to the antecedent period. From 1 April 2009 to 15 June 2009, average soil water content was 0.33 at 10 cm, 0.35 at 20 cm and 0.28 at 35 cm depth. The average soil moisture values measured in the field indicate the field capacity of the soils at Mt. Tom.

The maximum soil moisture was 0.43 at 10 cm depth and 0.41 at both 20 and 35 cm depths. These values were typical during large magnitude rainfall events and during the snowmelt period, times when surface water would have been most abundant. The lowest soil moisture values were typically found in the winter when the soil did not receive inputs from precipitation or during short-lived snowmelt events throughout the winter. Soil moisture was also lower than average in the month of July 2008 for a period of approximately ten days. Lower soil moisture values in the summer tend to coincide with higher air temperatures.

7.2 Persistence of Soil Moisture at Mt. Tom

Persistence was detected by calculating autocorrelation functions for two separate periods. The first period, called the antecedent or fall freeze-up period, was from 15 August 2008 to 20 November 2008. After 20 November, the snow depth was deep enough to prevent any fluctuation in soil temperature at 12 cm depth. The second period was called the spring melt period and it was chosen to represent the time during which snow depth decreases to zero and soil moisture responds to individual precipitation events instead of snowmelt. For this study, this period was chosen to be from 1 April 2009 to 15 June 2009.

Using autocorrelation we find that in the period from 17 July 2008 to 14 July 2009 there is persistence of autocorrelation of soil moisture on the order of 10-13 days in the snowmelt period at Mt. Tom. This persistence is attributed to the continuous input of water from snowmelt during this time. In the period prior to the onset of continuous snow cover from 15 August 2008 to 20 November 2008 (the antecedent period) significant autocorrelation of soil moisture lasts no more than two days. For the year as a whole, we find autocorrelation on the order of about 10-13 days; however, these data are weighted by the winter period, when soil moisture at all three depths remains at average values and the soil is decoupled from the atmosphere by the seasonal snow pack. During this time, we see no change in soil moisture. Values rise again when snowmelt begins and the snow pack has diminished enough to allow melt water to infiltrate the soil.

The lack of persistence of anomalous soil moisture at other times of the year is likely due to the steepness of the terrain and the site. It is also likely that the experimental design and lack of instrumentation prevented proper detection of significant persistence of moisture from snow melt or high magnitude rainfall events. It is speculated that with deeper sampling we could obtain a better idea of how soil moisture moves through the soil to groundwater supplies and how persistence might change with depth. Lateral flow could be observed as well, but the equipment and space on the data logger were not available for this study. Also, monitoring evaporative fluxes would allow for assessment of how soil moisture contributes to the boundary layer moisture and local convection.

Using cross-correlation we find that soil moisture at 10 and 35 cm is more in phase than the soil moisture at 10 and 20 cm depth or 20 and 35 cm depth. This has no real significance to this study. Further research into lateral flow and differences in hydraulic conductivity of the different layers in the soil profile would enhance understanding of these phenomena; however, this is presently beyond the scope of this research.

7.3 Future Direction and Knowledge Gaps

For future studies on the persistence of snowmelt it would be useful and interesting to assess surface evaporation. Having measurements for temperature and humidity at two separate heights above the ground allows for calculations of the moisture flux above the surface. This would aid in evaluating whether or not evaporation contributes significantly to local precipitation. Mt. Tom is in an area that currently receives ≈50% of its annual precipitation as snow; however, it is expected this fraction will decrease in a warmer climate. With the surface exposed for a longer period through the year and with a thinner snow cover there is potential for changes in evaporative fluxes and surface water storage.

In forests where interception plays a role in storing snow, sublimation might contribute to local humidity and hence could affect precipitation in the winter. It would be interesting to assess whether or not sublimation contributes significantly to snowfall in the Cariboo Mountains, or elsewhere for that matter.

Adjacent to the Willow Basin is the Crescent Creek Basin, part of which flows through the Interior Cedar/Hemlock (ICH) zone near McBride, BC. The ICH zone is fed year round by melt water from the icefields at higher elevations. It is this meltwater that supports a lush and ancient cedar hemlock forest. Current research by the NHG is exploring movement of soil moisture through the soils in this area.

Ideally, there would be a long term soil moisture time series that we could use to better assess persistence of anomalous soil moisture and what other meteorological factors might contribute to greater or less persistence. Furthermore, a longer time series would allow for assessment of how anomalously wet or dry periods are manifested in the soil and whether or not land-surface interactions in this region would cause these anomalies to persist. Soil moisture has the greatest effect on local climate where seasonal patterns to precipitation exist. These areas include mid-latitudes and areas where precipitation is non-stationary.

There is no assessment of spatial variability of soil conditions and how different soils might affect the overall persistence of soil moisture in the region. In the Mt. Tom area there are several different land conditions spread over a relatively small spatial area in complex, mountainous terrain. It is impossible to represent the true diversity of soil conditions in the area and hence, a site was chosen that was neither too wet nor too dry and was in relatively flat terrain between north and south facing slopes. The site was also chosen because of its accessibility. The research team did not possess the resources to take regular measurements of soil moisture from multiple sample points around the area, and the length of the cables and cost of instruments and other equipment prevented continuous sampling of soil moisture in forested versus open, clear cut areas.

With the current MPB epidemic it is expected that local soil moisture conditions will likely increase as result of decreased demand for water by vegetation and due to decreased interception of snow by trees. Erosion tends to increase as result of deforestation and with the added threat of MPB this could become an important issue in years to come. Soil erosion slows vegetation from recolonizing clear cut sites. If there is no soil on the surface there is no medium for roots to take hold.

Further research in nearby areas would also provide knowledge of spatial variability and differences in surface storage throughout BC. Having greater representation of soil moisture conditions over the range of ecosystems in this province would facilitate better understanding of how soil moisture limits vegetation and the land-atmosphere interactions that determine ecosystem diversity. There is such a wide range of environments in BC from the dry interior around the Okanagan, Kelowna and Kamloops to the wet, coastal rainforests. The Cariboo Mountains exhibit extreme diversity in ecosystems with icefields, cedar and hemlock forests as well as the spruce and fir forests more typical of drier environments. With mountains imposing control on local weather patterns, predicting how climate change will affect soil water budgets is difficult, especially without ground based data to support models.

This study provides some information on the temporal variability of soil moisture and opens a discussion on the factors that contribute to seasonal, weekly and daily variations in this parameter. Focussing on the snowmelt season and comparing it with the antecedent fall freeze-up allows for characterization of the soil at Mt. Tom with respect to its ability to store moisture for future use by plants, trees and wildlife.

We find that the capacity of the soil to store moisture is limited by a number of factors such as snow accumulation, soil physical properties, and terrain features such as slope. With greater investigation and data that cover a larger spatial extent, it would be possible to narrow down the factors that limit soil moisture in the Mt. Tom area.

For now, without more data, we conclude that soil moisture at the site persists for approximately 3-4 days after any given precipitation event throughout the year except during the snowmelt period, during which soil moisture values remain well above average. This behaviour is typical of well drained soils. Once snow cover has melted completely, which is determined by observing when soil temperature data begin to fluctuate more than 0.5°C, soil moisture stays above average for no longer than 3 to 4 days following the end of any individual precipitation event. The fact that soil moisture does not drop below field capacity at any depth in the rooting zone is significant from an ecological point of view. Vegetation in this area is adapted to these wet conditions.

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