Effect of Vegetation on the Accumulation and Melting of Snow at the TW Daniels Experimental Forest

Vinod Mahat
Utah State University

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EFFECT OF VEGETATION ON THE ACCUMULATION AND MELTING OF SNOW
AT THE TW DANIELS EXPERIMENTAL FOREST

by

Vinod Mahat

A dissertation submitted in partial fulfillment of the requirements for the degree of

DOCTOR OF PHILOSOPHY

in

Civil and Environmental Engineering

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2011
ABSTRACT

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by

Vinod Mahat, Doctor of Philosophy
Utah State University, 2011

Major Professor: David G. Tarboton
Department: Civil and Environmental Engineering

Snow melt is an important component of Western US water resources, accounting for about 50-80% of the annual runoff. Prediction of runoff from snowmelt in heterogeneous watersheds requires the quantification of physical processes accounting for the effects of forest canopy on snow accumulation, melt and sublimation. The forest canopy intercepts snowfall that resulting in smaller snow accumulations in forest area than in open area. The forest canopy also modifies the energy exchange between snow surface and the atmosphere, and alters the sublimation and melting of sub-canopy snow relative to open area. This dissertation has examined ways to improve snowmelt modeling capability to better account for canopy effects and has presented enhancements to an energy balance model that include i) an improved representation of the transmission of radiation through the canopy, ii) an improved representation of the atmospheric transport of heat and water vapor between the snow on the ground, in the canopy and the atmosphere above, and iii) an improved representation of the processes of
canopy snow interception and unloading. These enhancements were evaluated against 4 years of field data (2006-2010) collected at the TW Daniels Experimental Forest (TWDEF) located 30 miles N-E of Logan. Observations included continuous automated climate and snow depth measurements supported by periodic field measurements of snow water equivalent and temperature in four different vegetation classes (grass, shrubs, coniferous forest, deciduous forest). The enhanced canopy components were included into the Utah Energy Balance Snowmelt model and provide improved capability to predict the surface water input and runoff from snowmelt in heterogeneous watersheds using a parsimonious approach that can be used with practically available information.
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measurements of snow water equivalent and temperature in four different vegetation classes (grass, shrubs, coniferous forest, deciduous forest). The enhanced canopy components were included into the Utah Energy Balance Snowmelt model and provide improved capability to predict the surface water input and runoff from snowmelt in heterogeneous watersheds using a parsimonious approach that can be used with practically available information.
DEDICATION

This work is dedicated to my loving family.
ACKNOWLEDGMENTS

My deepest gratitude is to my advisor, Dr. David G. Tarboton. I thank him for setting up high academic standards for his students and guiding them to meet those standards. I have been amazingly fortunate to have an advisor like him; without his support this dissertation could not have been completed. I thank him for financial support and his generous availability despite his busy schedule to provide stimulating suggestions and encouragements during the work. I deeply appreciate his persistence in pushing me to re-examine, rethink and rewrite so as to produce the best work possible. He has been an excellent advisor. I hope that one day I would become as good an advisor to my students as David has been to me.

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I dedicate this dissertation to my wife and kids, and my family in Nepal for their unconditional love and support in every way possible throughout the process of this course, this dissertation and beyond. None of this would have been possible without their love and care. Their love and support helped me overcome setbacks and stay focused on my graduate study.

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Vinod Mahat
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CHAPTER 1
INTRODUCTION

Snow melt is an important component of Western US water resources. In the Western US, snowmelt accounts for about 50-80% of the annual runoff. The processes of snow accumulation and melt in open areas are understood for a range of climates and well represented in numerical models [Anderson, 1976; Jordan, 1991; Marks et al., 1992; Price and Dunne, 1976; Tarboton and Luce, 1996]. Prediction of the evolution of snow packs and runoff from the melting of these in forested areas is complex [Storck et al., 2002]. The presence of vegetation on the watershed influences radiation and turbulences and that have strong impact on the energy balance and snowmelt below the canopy, and reduces the amount of snowfall/rainfall reaching the ground by interception. These interactions of vegetation with snow influence the timing, quantity and duration of snowmelt and need to be understood in assessing and forecasting water supplies from the melting mountain snowpack.

There have been number of studies focused on snow-vegetation interactions in the past [e.g. Bartlett et al., 2006; Ellis and Pomeroy, 2007; Essery et al., 2003; Hedstrom and Pomeroy, 1998; Koivusalo, 2002; Link and Marks, 1999; Storck et al., 2002; Tribbeck et al., 2004; Wigmosta et al., 1994]. But, when more than thirty-three snowpack models were included in a snow model inter-comparison project (SNOWMIP2) in 2007, no universal ‘best’ model was found for all sites or locations. Comparison of model performance at different sites showed less consistency at forest sites than in open sites [Rutter et al., 2009]. This could be because of the poor representation of the vegetation dependent snow processes in the models as current
models either overlook or use simplified representations of physical processes controlling the accumulation and melt of snow in the forest canopy.

In the history of snowmelt modeling, early snow simulation models used the temperature index approach [Anderson and Crawford, 1964; Rockwood, 1964]. These models use air temperature as an index to quantify energy exchange across the snow-air interface. Early energy balance models include models developed by US Army Corps of Engineers [1956], Anderson [1976], Humphrey and Skau [1974], Outcalt et al. [1975], Obled [1973], and Price and Dunne [1976]. These models are detailed single or multiple layers models which include net radiation transfer, latent and sensible heat transfer, heat transfer by rain water, and the change in snow heat storage in an energy balance. Snow properties and processes are described by processes of precipitation, compaction and settling; and snow grain diameter, albedo, snow density and thermal conductivity of snow. These models are designed to operate in open areas where no vegetation canopy is present, and do not have vegetation component.

US Army Cold Research and Engineering Laboratory Model, SNTHERM [Jordan, 1991; Jordan et al., 1986]; The Simultaneous Heat and Water, SHAW [Flerchinger and Saxton, 1989]; point or distributed energy balance models, SNOBAL [Marks, 1988] and ISNOBAL [Marks et al., 1999], and Utah Energy balance physically-based model, UEB [Tarboton and Luce, 1996] are often cited recent snowmelt models in snow hydrology. The modeling approach of these models is similar to that of Anderson. These models are also designed to operate in open areas, and cannot be used to investigate the impacts of canopy impact on snow processes.
Some snow models that account for canopy effects on snow processes are the Canadian Land Surface Scheme, CLASS [Verseghy, 1991; Verseghy et al., 1993], the Distributed Hydrology Soil Vegetation model, DHSVM [Wigmosta et al., 1994] and the Variation Infiltration Capacity model, VIC, originally developed by Liang et al. [1994]. These models are extensively parameterized spatially distributed models, and require extensive inputs. In mountainous regions the distributed data for model input and verification is limited. As a consequence, many distributed parameters are often adjusted. In this case the model parameters are frequently not physically based or not clearly related to watershed properties.

Based on the assessment of snowmelt models and prior work on snow vegetation interactions it is evident that the representation of canopy processes in snowmelt models remains an unsolved problem. Though there are some models that do consider the snow-vegetation interaction processes, but the usefulness of these models has been questioned due to problems of over-parameterization, parameter estimation and validation limitations. This dissertation focuses on the modeling of snow vegetation interactions to quantify snow accumulation and melt in a heterogeneous watershed using a parsimonious approach that can be used with practically available information. Parsimony in terms of model complexity and data requirements is a major concern. We selected the UEB model as starting point for our work. The UEB model is parsimonious, in its focus on surface mass and energy exchange without becoming overparameterized by representing within snow multilayer complexity. This limits the number of parameters and state variables within the model.
With the goal of improving snowmelt modeling capability in a forested environment to predict the surface water input and runoff from snowmelt using a parsimonious approach, we developed and evaluated a vegetation component for the UEB snowmelt model that includes

1. Improved representation of the transmission of radiation through the canopy,
2. Improved representation of the atmospheric transport of heat and water vapor between surface snow, canopy snow and the atmosphere above, and
3. Improved representation of processes of canopy snow interception and unloading.

With the addition of this new vegetation component, the formerly single layer model has become a two-layer model that represents the surface and the canopy snow separately. Figures 1.1 and 1.2 show schematic diagrams of the new two layer UEB model. The model is driven by modest canopy (leaf area index, canopy height and canopy cover) and meteorological data (precipitation, air temperature, relative humidity and wind speed).

This dissertation also reports field observations of snow in four different vegetation classes at the TW Daniels Experimental forest (TWDEF) located 30 miles N-E of Logan, Utah that have been used to examine the effects of vegetation on snow and to evaluate the model enhancements.

Towards achieving the goal, this dissertation offers solutions to these five related questions

a) How does the surface water input (rain + snowmelt) differ between vegetation types and why?

b) How does snow accumulation differ between vegetation types and why?
c) Does interception result in lower accumulation in more dense forests?

d) Does redistribution enhance snow accumulation in open settings?

e) What are the losses to sublimation?

This dissertation contains five chapters including introduction chapter (Chapter 1) and summary chapter (Chapter 5). The other three chapters represent the core contributions of this dissertation and are formatted as papers for publication. Each of these is outlined in the following paragraphs.

Chapter 2 focuses on how to better represent the penetration of radiation through a forest canopy to better estimate the beneath canopy radiation that drives the energy balance and snowmelt in a forested area. In snow melt modeling, previous studies [e.g. Ellis and Pomeroy, 2007; Essery et al., 2003; Koivusalo, 2002; Link and Marks, 1999] have used Beer's law to attenuate the solar radiation penetrating a canopy. In Beer's law solar radiation is decreased exponentially with the increase in path length of the absorbing medium without accounting for scattering [Monteith and Unsworth, 1990]. Some more detailed, multiple-layer radiation models [e.g. Dickinson, 1983; Flerchinger and Yu, 2007; Flerchinger et al., 2009; Norman, 1979; Sellers, 1985; Zhao and Qualls, 2005, 2006] that account for the radiation multiple scattering demand information about leaf density, gaps, inclination, orientation, distribution etc. which are not easily available.

In Chapter 2, we developed a two stream canopy radiation transmission model that explicitly accounts for radiation scattering, absorption and reflections by leaves. In order to maintain a sufficient level of parsimony in the model we considered the whole canopy as a single layer and took an approach that uses minimum number of vegetation parameters (i.e. leaf area index, canopy coverage fraction and canopy height) to model
radiation transfer. Multiple reflections of direct and diffuse solar radiation between the canopy and surface were also modeled; however the multiple reflections of longwave radiation between the canopy and surface were ignored as plants strongly absorb longwave radiation. Shortwave and longwave radiation that transmitted through the canopy combined with other energy fluxes provides the net energy that drives snowmelt beneath and within the canopy.

The model was first evaluated at an open site where measurements of four radiation components (incoming and outgoing solar and longwave radiation), surface temperature and SWE were compared with model simulations. This comparison that showed a good agreement between modeled and observed values served to check the model representations of could cover, albedo, thermal conductivity and emissivity. The model was then run for conifer and deciduous forest canopies, driven by open area measurements taken as representative of above canopy inputs (solar and longwave radiation and meteorology) to estimate the net radiation and SWE beneath the canopies. Model simulations of these were compared with beneath canopy measurements inside the conifer and the deciduous forest at the TW Daniels Experimental Forest.

Chapter 3 focuses on the improved representation of the atmospheric transport of heat (sensible heat) and water vapor (latent heat) between the surface snow, canopy snow and the atmosphere above in an energy balance snowmelt model. Previous studies focused on snow-vegetation interactions [e.g. Bartlett et al., 2006; Ellis and Pomeroy, 2007; Ellis et al., 2010; Essery et al., 2003; Koivusalo, 2002; Link and Marks, 1999; Tribbeck et al., 2004] have indicated the importance of radiation and turbulent fluxes in snow cover mass and energy balance. In snow hydrology, canopy radiation transmission
and below canopy radiation have been measured and validated [e.g. Ellis et al., 2010; Pearson et al., 1999; Tribbeck et al., 2004] but the measurements and validation of turbulent fluxes (above and below the canopy) in the forested environments have rarely been done.

In Chapter 3, we developed and evaluated a turbulent flux model for computing sensible heat and latent heat and subsequent vapor losses from the snow in the canopy and the snow at the surface below the canopy. The approach was to use the temperature and vapor pressure differences between the snow surface, canopy air and the snow in the canopy with the resistances. The model solves the two energy balance equations: one at the canopy and the other at the surface to estimate the canopy and the surface temperatures and the vapor pressures with the inputs of above canopy air temperature and vapor pressure. Resistances were estimated using leaf area index and wind profile which is assumed to be logarithmic above the canopy, exponential within the canopy and again logarithmic over the snow surface on the ground, following Choudhury and Monteith [1988] and Dolman [1993]. The flux model was evaluated by making comparison of model simulated values with eddy covariance measurements at Niwot Ridge AmeriFlux site at Colorado and wind speed measurements at TWDEF and Niwot Ridge.

Chapter 4 focuses on how to represent physical process of snow interception, sublimation, unloading and melt from the canopy that impacts SWE on the below canopy snowpack beneath. The presence of vegetation on the watershed not only alters the turbulence and radiation that have strong impact on the energy balance and snowmelt below the canopy but also reduces the amount of snowfall/rainfall reaching the ground by interception. Intercepted snow in the canopy either sublimates [Lundberg and Halldin,
or unloads as mass [Mackay and Barlett, 2006] or melt water drip within the canopy. Sublimation reduces the amount of snow available beneath the canopy, and unloading whether it is mass release or melt from the canopy affects the depth and density of snow beneath the canopy. All these processes affect SWE in the snowpack below the canopy, but numerical models that explicitly represents these processes are rare [Andreadis et al., 2009].

To estimate the canopy snow interception, sublimation, mass unloading and melt, we distinguished surface snow that has accumulated on the ground surface, either beneath a canopy or in the open, from the canopy snow that is held above the surface on vegetation after having been intercepted. Sublimation from the canopy snow and surface were estimated separately. We also partitioned the canopy mass release and melt water drip from the canopy sublimation. We used the approach developed by Hedstrom and Pomeroy [1998] to estimate the canopy snow interception and mass unloading. Melt water drip was quantified based on canopy energy balance, using energy flux (radiation, sensible and latent heat) exchanges between the canopy and the atmosphere. This canopy process model was added to the Utah Energy Balance (UEB) snowmelt model [Tarboton and Luce, 1996; Tarboton et al., 1995] to obtain the interception and sublimation loss for the prediction of SWE beneath the canopy, that we compare to observations made in the location under the conifer and deciduous canopies at the TW Daniels Experimental Forest in Northern Utah.

The research described in the dissertation advances the understanding of snow-vegetation interactions by investigating the canopy impacts on snow accumulation and melt processes using the field study and modeling approach. The model developed in
this work comprehensively describes the physical processes of canopy radiation transmission; snow interception, sublimation, mass unloading and melt in forested areas. Representation of canopy processes in a snowmelt model has enabled more accurate forecasts of surface water input from the accumulated snow and snowmelt which are responsible for runoff and water supply in the semi-arid climates of the western US where most of the water comes from the melting mountain snowpack. The model has been evaluated and parameterizations are tuned based on the measurements available from the TWDEF and Niwot Ridge AmeriFlux study sites. While it is rarely possible to constrain by measurements all the processes in a model, the comprehensive set of measurements available at these study sites has given a high level of confidence to the model.

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**Figure 1.1.** Schematic illustration of two layer UEB model physics and parameterizations of energy balance. Here $Q_{si}$ is solar radiation incident at the canopy. $Q$ may denote any of $Q_b, Q_d, Q_{li}, Q_{p}, Q_{c}, Q_{le}$ or $Q_{net}$. $Q_b, Q_d$ and $Q_{li}$ denote direct fraction of solar radiation, diffuse fraction of solar radiation and incoming longwave radiation incident at the top of the canopy, respectively. $Q_p$ is energy advected due to precipitation, and $Q_c$ and $Q_{le}$ are longwave radiation emitted from the canopy and the snow surface, respectively. $Q_{net}$ is net radiation absorbed at the snow surface. $f_1, f_2$ and $f_3$ are net fractions of radiation absorbed at the snow surface snow, in the canopy, and radiation lost to the sky, respectively. $Q_h$ and $Q_e$ are total fluxes of sensible heat and latent heat from the atmosphere which is partitioned into $Q_{hc}$ and $Q_{ec}$, the sensible and latent heat flux contributions to the forest canopy, and $Q_{hs}$ and $Q_{es}$, sensible and latent heat flux contributions to the surface. The units of all these energy terms is $(kJ \ m^{-2} \ hr^{-1})$ per unit of horizontal area. $T_{ac}$ is canopy air temperature ($^\circ C$), $T_s$ is surface temperature ($^\circ C$), $T_c$ is canopy temperature ($^\circ C$), $T_a$ is above canopy air temperature ($^\circ C$), $e_{ac}$ is canopy air vapor pressure (Pa), $e_a$ is above canopy air vapor pressure (Pa) and $e_s(T_s)$ and $e_c(T_c)$ are surface and canopy saturated vapor pressures calculated as functions of snow and canopy temperatures, respectively.
Figure 1.2. Schematic illustration of two layer UEB model physics and parameterizations of mass balance. The units of all these mass terms is \((\text{m hr}^{-1})\) for fluxes and \((\text{m})\) of water equivalent for storages.
CHAPTER 2

CANOPY RADIATION TRANSMISSION FOR AN ENERGY BALANCE SNOWMELT MODEL

Abstract

To better estimate the radiation energy within and beneath the forest canopy for energy balance snowmelt models, a two stream radiation transfer model that explicitly accounts for canopy scattering, absorption and reflection was developed. Upward and downward radiation streams represented by two differential equations using a single path assumption were solved analytically to approximate the radiation transmitted through or reflected by the canopy with multiple scattering. This approximation results in an exponential decrease of radiation intensity with canopy depth, similar to Beer’s law for a deep canopy. The solution for a finite canopy is obtained by applying recursive superposition of this two stream single path deep canopy solution. The radiation model was included in a distributed energy balance snowmelt model and results compared with observations made in three different vegetation classes (open, coniferous forest, deciduous forest) at a forest study area in the Rocky Mountains in Utah, USA. The model was able to capture the sensitivity of beneath canopy net radiation and snowmelt to vegetation class. The model slightly over predicted net radiation beneath the forest canopy during early winter however the late winter prediction was better. The good prediction of net radiation during the late winter resulted in satisfactory predictions of snowmelt from the forested areas. These results suggest that reasonable predictions of snowmelt from forested areas, and the sensitivity of snowmelt to forest type and canopy
density (leaf area) can be obtained with the parsimonious (one layer two stream) representation of canopy radiation transmission given here.

2.1. Introduction

Snow accumulation, melt and sublimation processes are different for open and forest sites. Vegetation and land cover influences snow processes making it difficult to predict snowmelt which is responsible for water supply in much of the world, including the mountainous regions of the western U.S. where this study was conducted. The processes of snow accumulation and melt in open areas are understood for a range of climates and well represented in numerical models [Anderson, 1976; Bartlett and Lehning, 2002; Jordan, 1991; Lehning et al., 2002; Marks et al., 1992; Price and Dunne, 1976; Tarboton and Luce, 1996; Wigmosta et al., 1994]. Prediction of the evolution of snow packs in forested areas is more complex [Storck et al., 2002]. The forest canopy intercepts snow fall, attenuates radiation, and modifies the turbulent exchanges of energy and water vapor between snow in and under the canopy and the atmosphere, thereby affecting snow accumulation and melt.

Radiation is the main energy that drives the energy balance and snowmelt. This paper focuses on how to represent the penetration of radiation through a forest canopy in an energy balance snowmelt model. The input of solar radiation to the ground surface whether in the open or beneath the canopy varies depending on solar angle and azimuth as well as cloudiness and topography (slope and aspect) [Link et al., 2004; Stähli et al., 2009]. Net radiation at the snow surface then depends on reflection from the surface, governed by the surface albedo as well as scattering and multiple reflections between the snow surface and canopy. Surface albedo depends on coverage by snow (coverage is
patchy when the snow is shallow and surface rough), snow surface grain size which is related to age and the presence of dust or litter on the surface (how fresh and clean is the snow) [Hardy et al., 2000; Jordan, 1991].

A number of techniques have been used to model radiation beneath forest canopies. Ellis and Pomeroy [2007], Essery et al. [2003], Koivusalo [2002] and Link and Marks [1999] used Beer's law to attenuate the solar radiation penetrating a canopy. In Beer's law solar radiation is decreased exponentially along the path through the absorbing medium without accounting for scattering [Monteith and Unsworth, 1990]. Depending on the density of the canopy, multiple scattering may increase the irradiance reaching the surface as compared to Beer's law, by up to 100% [Nijssen and Lettenmaier, 1999]. Transmission of solar radiation through the canopy varies with leaf area as well as canopy gaps [Hardy et al., 2004]. Essery et al. [2008], Hardy et al. [2004] and Hu et al. [2010] derived leaf area and canopy gap fraction by analyzing hemispherical photographs and calculated canopy radiation transmission from this information.

Dickinson [1983] and Sellers [1985] developed a two stream approximation for radiation transfer through the atmosphere or a vegetation canopy which includes multiple scattering [Dickinson, 1983; Sellers, 1985]. In this two stream approximation, upward and downward radiation is expressed using two differential equations quantifying the change in downward and upward radiation due to interception, absorption and scattering. This approach applies to integrated quantities as opposed to angular dependent intensities [Meador and Weaver, 1980] and neglects anisotropy that may result due to angular effects in scattering.
A more detailed approach was taken by Li et al. [1995] and Ni et al. [1997] in the Geometric-optical and radiative transfer (GORT) model which accounts for the three dimensional geometry of the forest canopy and includes multiple scattering within and beneath the canopy. The GORT model is computationally expensive and also requires parameters such as crown geometry and foliage area volume density that are difficult to measure in the field [Hardy et al., 2004].

There are many other single or multiple-layer radiation transfer models [e.g. Flerchinger and Yu, 2007; Flerchinger et al., 2009; Ni et al., 1997; Niu and Yang, 2004; Norman, 1979; Roujean, 1996; Zhao and Qualls, 2005, 2006]. These models often require canopy information (e.g. leaf density, inclination, orientation, crown diameter and depth etc.) that is hard to obtain. Furthermore some of these models have only been tested in agricultural environments and have not been evaluated for the purpose of snowmelt prediction beneath a forest canopy.

In mountainous regions data for model input and verification are limited. As a consequence, many parameters are often adjusted. Thus the usefulness of detailed models has been questioned due to problems of over-parameterization, parameter estimation and validation limitations. There are very few models that can be used when input data are limited, and are transportable and applicable at different places with little calibration. Efforts to develop simplified approaches to model radiation beneath the canopy accounting for multiple scattering of radiation include Nijssen and Lettenmaier [1999], Tribbeck et al. [2004], and Yang et al. [2001]. Nijssen and Lettenmaier’s [1999] model provides a solution for infinitely deep canopy while Tribbeck et al.’s [2004] model assumes radiation scattered by the canopy is reflected equally in upward and downward
directions and does not account for within canopy scattering. Yang et al. [2001] present a
simplified two stream approach but their model requires vegetation geometry
information.

The purpose of this paper is to present and evaluate a simple model to estimate
beneath canopy radiation that drives the energy balance and snowmelt beneath the forest
canopy using a parsimonious approach that can be used with practically available
information. Parsimony in terms of model complexity and data requirements is a design
consideration, striving for the best possible physical representations given typical
practically available data. The forest canopy is modeled as a single layer with parameters
leaf area index and canopy cover fraction quantifying the radiation attenuation. A two
stream radiation transfer model that explicitly accounts for canopy scattering, absorption
and reflection is used. Upward and downward radiation streams represented by two
differential equations using a single path assumption were solved analytically to
approximate the radiation transmitted through or reflected by the canopy with multiple
scattering. This approximation results in an exponential decrease of radiation intensity
with canopy depth, similar to Beer’s law for a deep canopy. The solution for a finite
canopy is obtained by applying recursive superposition of this two stream single path
deep canopy solution. The parameters required are the same parameters that are used in
Beer’s law, but the theoretical foundation of the model has been improved in that
multiple scattering and a finite canopy depth are represented. The new radiation transfer
approach was added to the Utah Energy Balance (UEB) snowmelt model [Tarboton and
Luce, 1996; Tarboton et al., 1995] to model snow energy and mass balances within and
beneath the canopy driven by inputs of radiation and weather from above the canopy. At
the surface UEB focuses on surface mass and energy exchanges without becoming overparameterized by attempting to represent within snow multilayer complexity. The added canopy component similarly uses a canopy parameterization that strives for a good physical representation of the processes involved without requiring hard to quantify information on canopy structure and leaf orientation.

2.2. Study Site

Field measurements were carried out at the TW Daniels Experimental Forest (TWDEF; http://danielforest.usu.edu) located about 30 miles Northeast of Logan, Utah (Figure 2.1). TWDEF comprises an area of 0.78 km² at an elevation of approximately 2700 m. It lies at 41.86° North and 111.50° West. The TW Daniels Experimental Forest is on the divide of the watershed that contributes to the Logan River and Bear Lake. Average annual precipitation is about 950 mm of which about 80% is snow. The maximum snow depth can reach 5 m in the area where snow drifts occur. Vegetation is comprised of deciduous forest (Aspen), coniferous forest (Engelmann spruce and subalpine fir), open meadows consisting of a mixture of grasses and forbs, and shrub areas dominated by sagebrush.

Instrumentation was installed starting in 2006 to monitor weather and snow within four different vegetation classes: grass, shrubs, coniferous forest, and deciduous forest; and includes twelve weather station towers (three replicates in each vegetation class), one central tower (in shrub area) with more comprehensive radiation instrumentation and one SNOTEL station in a clearing within the coniferous forest. The following automated data were collected:
• Continuous measurements of snow depth (Judd communications depth sensor) at each of the twelve stations.

• Continuous measurements of weather: temperature and humidity (Vaisala HMP50); wind (Met One, 014A); net radiation, (Kipp & Zonen NR-Lite) at one station in each vegetation class. These instruments were placed at heights above the ground of about 2.5 m in conifer, 4.5 m in deciduous and 4 m in shrub sites so as to remain above the deep snow that accumulates in the deciduous and shrubs areas.

• Four separate radiation components: downward and upward shortwave and long wave (Hukseflux, NR01 4-way radiometer) and snow surface temperature (Apogee Instrument, IRR-PN) at the centralized weather station.

• The standard suite of SNOTEL observations at the adjacent SNOTEL site, from which we used precipitation. This SNOTEL site was installed in summer 2007, so its data are first available for the 2007/8 winter.

Slope and aspect were determined from a 1 m resolution digital elevation model constructed from bare earth points classified from an airborne LiDAR survey of the site. Table 2.1 lists the site information, and in addition to these parameters includes parameters used with other aspects of the model that are not the focus of this paper.

Field observations roughly every two weeks for four winters (2006/7 - 2009/10) comprised two snow pits: one in the shrub area (Pit 1, Figure 2.1) and the other in a conifer clearing (Pit 2, Figure 2.1), and snow depth at multiple locations in all four vegetation classes. Within each snow pit samples were taken at 10 cm vertical intervals over the entire snow pit depth using a 250 cm³ stainless steel cutter to derive the snow
density. The density measured at the pit in the shrub area was used to represent both shrub and grass areas. Both shrub and grass are regarded as open because during the winter snow season snow completely covers the shrubs. Snow density measured in the conifer clearing was used to represent forested areas (both conifer and deciduous). These density values were used with the depth measurements at multiple locations to derive the snow water equivalent (SWE). Temperature was also measured at the surface and at 10 cm vertical intervals over the entire snow pit depth. These temperature measurements were used to derive the energy content of the snow. Numbered snow survey points (Figure 2.1) show locations where the depth measurements were made across the four vegetation classes.

2.3. Model Description

The UEB snowmelt model [Tarboton and Luce, 1996] is a physically-based point energy and mass balance model for snow accumulation and melt. Snowpack is characterized using three state variables, namely, snow water equivalent, \( W_s \), (m), the internal energy of the snowpack and top layer of soil, \( U_s \), (kJ m\(^{-2}\)), and the dimensionless age of the snow surface used for albedo calculations. The UEB model is a single layer model. \( U_s \) and \( W_s \) are predicted at each time step based on the energy balance. Details of the original UEB model formulation are given by Tarboton et al. [1995], Tarboton and Luce [1996] with enhancements for the calculation of surface temperature using a modified Force-Restore approach given by Luce and Tarboton [2010] and You [2004].

In this paper we present the canopy radiation transmission component of an enhanced UEB model that includes representation of canopy processes. The canopy
component is modeled as a single layer, which added to the original single layer UEB model results in a two-layer model that represents the surface and the canopy intercepted snow separately. Energy balances are solved iteratively for each layer to provide outputs of surface temperature, canopy temperature and the other energy fluxes that are based on canopy or surface temperature. The quantity and state of snow in the canopy is represented by a new state variable, canopy snow water equivalent ($W_c$). We assume that the energy content of intercepted snow in the canopy is negligible so canopy temperature, including snow in the canopy, is assumed to adjust to maintain energy equilibrium, except when this requires canopy temperature to be greater than freezing when snow is present in the canopy, in which case the extra energy drives the melting of snow in the canopy.

### 2.3.1. Shortwave Radiation

#### 2.3.1.1. Partitioning of Radiation

The incoming solar radiation reaching the canopy surface, $Q_t$ is partitioned into direct and diffuse components, $Q_d$ and $Q_d$, as these components penetrate the canopy separately. $AT$ is the fraction of top of atmosphere total radiation reaching the top of the canopy either measured or estimated from diurnal temperature range using the procedure of Bristow and Campbell [1984]. This is split into direct radiation fraction, $AT_d$ and diffuse radiation fraction $AT_d$. Cloudiness fraction, $C_f$, is estimated from $AT$ using an the empirical relationship provided by Shuttleworth [1993]. We assume that when the sky is clear ($C_f = 0$) that a fraction $\lambda$ of $AT$ is direct. The value of $\lambda$ may be estimated based on
scattering and absorption properties of the cloud free atmosphere and is due to water vapor, dust and other scatterers in the atmosphere. We assume that when the sky is completely cloudy \( C_f = 1 \) that all radiation is diffuse. Using these as boundary conditions and assuming linear variation of each factor with \( C_f \) (Figure 2.2) leads to

\[
AT_b = \lambda AT_c (1 - C_f) \tag{2.1}
\]

\[
AT_d = AT - AT_b \tag{2.2}
\]

where \( AT_c = \max (AT, a_s + b_s) \) is the clear sky transmission factor. \( a_s + b_s \) is the fraction of extraterrestrial radiation reaching the surface on clear days. \( \text{Shuttleworth} [1993] \) recommended \( a_s = 0.25 \) and \( b_s = 0.5 \) for settings where no actual solar radiation data are available.

Once \( AT_b \) and \( AT_d \) are estimated, the total incoming radiation can be partitioned into direct and diffuse parts

\[
Q_b = \frac{AT_b}{AT} Q_i \tag{2.3}
\]

\[
Q_d = \frac{AT_d}{AT} Q_i \tag{2.4}
\]

### 2.3.1.2. Canopy Radiation Transmission

We develop the canopy radiation transmission model in three steps. First the attenuation of incident radiation due to interception, but not scattering is quantified. This results in an exponential decrease of radiation intensity with depth into the canopy (Beer's law). Next we consider scattering using a two stream approach for an infinitely deep canopy. This results in a modified exponential attenuation. In the third step we consider
a finite canopy with downward radiation incident at the top and upward radiation incident at the bottom. The direct and diffuse fractions of radiation transmitted through the canopy in the first step without scattering are represented by $\tau_b''$ and $\tau_d''$, respectively. $\tau_b'$ and $\tau_d'$ denote the direct and diffuse fraction when there is scattering but for a deep canopy, and $\tau_b$ and $\tau_d$ denote direct and diffuse fraction when there is scattering and the canopy is finite. The approach used is general such that it can be applied with both direct and diffuse radiation, and shortwave and longwave radiation, but with different scattering parameters. In this general approach we use $Q$ to represent radiation that may be direct, $Q_a$, diffuse, $Q_d$ or longwave, $Q_l$, and $Q_o$ to represent the value of this at the top of the canopy.

2.3.1.2.1. Radiation Transmission Without Scattering (Beer's Law)

In considering the penetration of light through a canopy the interception of a beam at zenith angle $\theta$ by an incremental layer of vegetation results in reduction in intensity given by

\[ dQ = -QG\rho \frac{dy}{\cos \theta} \]  

(2.5)

where $Q$ is radiation intensity, $\rho$ is the leaf density, $y$ is distance measured vertically downward from the top of the canopy and $G$ is a leaf orientation factor quantifying the average area of leaves when viewed from direction $\theta$. Here $G$ is assumed constant (i.e. independent of $\theta$). Integrating this from the top of the canopy downwards results in Beer's law (Figure 2.3)
The non-scattering transmission factor is thus given by
\[ \tau''_b = \frac{Q}{Q_o} = \exp(-K_b \rho y) \]  

where \( K_b = G / \cos \theta \) groups leaf orientation and zenith angle into a single parameter which is referred to as the black body attenuation coefficient because it describes the attenuation when the leaves are perfect radiation absorbers (black bodies). \( \rho y \) gives the leaf area index of canopy above point \( y \).

### 2.3.1.2.2. Radiation Transmission with Scattering in an Infinitely Deep Canopy

The attenuation in equation (2.7) does not consider scattering of light intercepted by the canopy. To account for scattering we use an approximation following Monteith and Unsworth [1990] that radiation from an incremental layer is scattered equally in an upward and downward direction and that scattering is along the same path as the incoming light. This approximation, strictly true only for leaves oriented perpendicular to the light beam, has been suggested and used as reasonable approximation for other angles to obtain analytic results [Goudriaan, 1977; Monteith and Unsworth, 1990] where otherwise radiation in multiple directions would need to be modeled. With this approximation streams of both downward and upward radiation need to be considered, hence the name two stream model, leading to

\[ -dU = -UK_b \rho dy + UK_b \rho \frac{\alpha}{2} dy + QK_b \rho \frac{\alpha}{2} dy \]  

\[ (2.8) \]
In these equations $\alpha$ is the leaf scattering coefficient, $Q$ and $U$ are intensity of the downward and upward beams, respectively (Figure 2.4). These equations account for the reduction in intensity of each beam due to interception, similar to Beer’s law, but with scattering from each incremental layer (as opposed to the canopy as a whole in the RM model, Tribbeck et al., 2004) assumed to be half upward and half downward. These equations are referred to as the Kubelka and Monk equations [Monteith and Unsworth, 1990]. Note that these are written for $y$ positive in the downward direction.

The pair of differential equations (2.8) and (2.9) have a general solution (see Appendix A)

$$Q(y) = \frac{1}{2} \left[ C_1 \left( 1 - \frac{1}{k'} \right) \exp \left( k' K_s \rho y \right) + C_2 \left( 1 + \frac{1}{k'} \right) \exp \left( -k' K_s \rho y \right) \right]$$  \hspace{1cm} (2.10)

$$U(y) = \frac{1}{2} \left[ -C_1 \left( \frac{1}{k'} + 1 \right) \exp \left( k' K_s \rho y \right) + C_2 \left( \frac{1}{k'} - 1 \right) \exp \left( -k' K_s \rho y \right) \right]$$  \hspace{1cm} (2.11)

where $C_1$ and $C_2$ are integration constants and $k' = \sqrt{1 - \alpha}$.

For an infinitely deep canopy with $y=0$ at the top of the canopy, a beam penetrating the canopy is reduced to zero ($Q=0$) when $y \to \infty$ (measured downward). This condition results in $C_1 = 0$. With this boundary condition, equations (2.10) and (2.11) reduce to

$$Q(y) = \frac{C_2}{2} \left( 1 + \frac{1}{k'} \right) \exp \left( -k' K_s \rho y \right)$$  \hspace{1cm} (2.12)
\[ U(y) = \frac{C_2}{2} \left( \frac{1}{k'} - 1 \right) \exp(-k' K_y \rho y) \]  \hspace{1cm} (2.13)

These represent an exponential decrease in light intensity into the canopy similar to equation (2.7) but with the exponent reduced by a factor \( k' \). \( k' \) quantifies the effect of multiple scattering on light penetration. The value of \( C_2 \) is related to the top boundary condition, \( Q_o \). The deep canopy solution, equation (2.12), yields the deep canopy multiple scattering transmission factor

\[ \tau_b' = \frac{Q(y)}{Q_o} = \exp(-k' K_y \rho y) \]  \hspace{1cm} (2.14)

This is a modification to Beer's law for radiation transmission of a single beam accounting for scattering.

The upward reflection factor giving the fraction of radiation reflected back from a deep canopy with multiple scattering, \( \beta' \) can be estimated using equations (2.12) and (2.13) as

\[ \beta' = \frac{U(y)}{Q(y)} = \frac{1 - k'}{1 + k'} \]  \hspace{1cm} (2.15)

The above is for a single beam. For diffuse radiation the approach is to recognize that it is comprised of single beam components from each direction \( Q(\theta) \). The component of each of these normal to the surface is integrated over the hemisphere. With this approach diffuse radiation above and in the canopy is given by \( \int_{\Omega} Q(\theta) \cos \theta d\Omega \) and \( \int_{\Omega} Q(\theta) \tau_b' \cos \theta d\Omega \), respectively. In this integral \( \tau_b' \) depends on \( K_b \) which is function of
Using these integrals, the transmission factor for diffuse radiation, $\tau_d'$ may be expressed as

$$
\tau_d' = \frac{\Omega \int Q(q) \tau_d' \cos \theta \, d\Omega}{\Omega \int Q(q) \cos \theta \, d\Omega}
$$

(2.16)

where $Q(\theta)$ is the radiance of the sky from the direction $\theta$, $d\Omega = \sin \theta d\theta d\phi$ is the solid angle for integration over the hemisphere, $\theta$ is the zenith angle in the range $(0, \pi / 2)$ and $\phi$ is the azimuth angle in the range $(0, 2\pi)$.

Assuming that radiation in the canopy is isotropic, $Q(\theta) = Q$, a constant; the solution to this equation [Nijssen and Lettenmaier, 1999] is

$$
\tau_d' = \left[1 - k' G \gamma y \exp(-k' G \gamma y) + (k' G \gamma y)^2 E_i(1, k' G \gamma y)\right]
$$

(2.17)

where $E_i(n, x)$ with $n$ a nonnegative integer is the exponential integral, defined as

$$
E_i(n, x) = 2 \int_1^\infty \frac{\exp(-xt)}{t^n} \, dt
$$

(2.18)

Because diffuse radiation is just an integral of direct beam components over the hemisphere, the upward diffuse radiation reflection factor for a deep canopy is also given by equation (2.15).

2.3.1.2.3. Radiation Transmission with Scattering in a Finite Canopy

The radiation transmission factors shown in equations (2.14) and (2.15) above are for an infinitely deep canopy. We obtain the solution for a finite canopy by recursive superposition of the deep canopy solution (Figure 2.5). At depth $y$ into a deep canopy, the solution is
where \( \tau'(y) \) may be \( \tau_b' \) from equation (2.14) or \( \tau_d' \) from equation (2.17).

Now suppose the canopy has a finite depth, \( D \), and incident radiation, \( Q_o \), at the top with no incident radiation from below the base. At the base, \( y = D \), the upward radiation \( U \) should be zero rather than \( U_1(D) \) given by equation (2.20). This can be obtained by adding (superposing) a solution for radiation input \( -U_1(D) \) at the base.

Applying equations (2.14) and (2.15) but for \( -U_1(D) \) incident from below, we get

\[
U_2(y) = -U_1(D)\tau'(D-y) = -\beta'Q_o\tau'(D)\tau'(D-y) \tag{2.21}
\]

\[
Q_2(y) = -\beta'U_1(D)\tau'(D-y) = -(\beta')^2Q_o\tau'(D)\tau'(D-y) \tag{2.22}
\]

This would result in \( Q_2(0) = -\beta'U_1(D)\tau'(D) = -(\beta')^2Q_o(\tau'(D))^2 \) at the top where \( y=0 \).

As before the top boundary condition \( Q_2(0) \) should be zero. This necessitates superposing another solution using incident radiation input of \( -Q_2(0) \) at the top, which gives

\[
Q_3(y) = -Q_2(D)\tau'(y) = (\beta')^3Q_o(\tau'(D))^2\tau'(y) \tag{2.23}
\]

\[
U_3(y) = -\beta'Q_2(D)\tau'(y) = (\beta')^3Q_o(\tau'(D))^2\tau'(y) \tag{2.24}
\]

Continuing this process recursively, the finite depth solution is

\[
Q(y) = Q_1(y) + Q_2(y) + Q_3(y) + \ldots \tag{2.25}
\]

\[
U(y) = U_1(y) + U_2(y) + U_3(y) + \ldots \tag{2.26}
\]
These infinite series can be evaluated to give

$$Q(y) = Q_o \left[ \frac{\tau'(y) - (\beta')^2 \tau'(D)\tau'(D - y)}{1 - (\beta')^2 (\tau'(D))^2} \right]$$  \hspace{1cm} (2.27)

$$U(y) = Q_o \left[ \frac{\beta \tau'(y) - (\beta')^2 \tau'(D)\tau'(D - y)}{1 - (\beta')^2 (\tau'(D))^2} \right]$$  \hspace{1cm} (2.28)

Using equations (2.27) and (2.28), the finite canopy transmission and reflection factors, $\tau$ and $\beta$, can be calculated as

$$\tau = \frac{Q(D)}{Q_o} = \frac{\tau'(D)[1 - (\beta')^2]}{1 - (\beta')^2 (\tau'(D))^2}$$  \hspace{1cm} (2.29)

$$\beta = \frac{U(0)}{Q_o} = \frac{\beta'[1 - (\tau'(D))^2]}{1 - (\beta')^2 (\tau'(D))^2}$$  \hspace{1cm} (2.30)

Equations (2.29) and (2.30) can be used for both direct and diffuse radiation. The fraction of direct radiation transmitted through the canopy, $\tau_h$, and diffuse radiation transmitted through the canopy, $\tau_d$, can be calculated using $\tau' = \tau'_h$ and $\tau' = \tau'_d$, respectively in equation (2.29). Similarly the direct and diffuse fractions of radiation reflected back from the canopy in an upward direction, $\beta_h$ and $\beta_d$ can be calculated using $\tau' = \tau'_h$ and $\tau' = \tau'_d$, respectively, in equation (2.30).

In evaluating equations (2.29) and (2.30) in the direct radiation case, using equation (2.14)

$$\tau'_h(D) = \frac{Q(D)}{Q_o} = \exp(-k' K_h \rho D) = \exp(-k' \frac{G}{\cos \theta} LF)$$  \hspace{1cm} (2.31)

Here $\rho D$, the area of leaves over the full canopy depth $D$ has been replaced by $LF$ where $L$ is the tree level leaf area index and $F$ the canopy cover fraction accounting for the fact
that trees may not completely cover the domain. The product $LF$ is effectively a canopy level leaf area index. We assume a constant leaf orientation factor, $G=0.5$, representing isotropic leaf orientations.

In the diffuse radiation case, using equation (2.17)

$$\tau_d'(D) = \left[ (1 - k'GpD) \exp(-k'GpD) + (k'GpD)^2 E, (1, k'GpD) \right]$$

$$= \left[ (1 - k'GLF) \exp(-k'GLF) + (k'GLF)^2 E, (1, k'GLF) \right]$$

(2.32)

We treat $G$, $L$, and $F$ as constants, neglecting any effects canopy intercepted snow may have on canopy radiation transmission and reflectance.

Figure 2.6 compares the transmittance of direct and diffuse solar radiation calculated using the two stream approach (equation 2.29) with the transmittance of radiation calculated using Beer’s law (equation 2.7) as a function of zenith angle. A significant increase in transmittance over the Beer’s law attenuation occurs due to multiple scattering in the canopy.

2.3.2. Longwave Radiation

Longwave radiation originates from three possible sources: the sky, snow surface and the canopy. Longwave radiation from each of these sources is considered to be diffuse radiation that penetrates through or is scattered by the canopy according to diffuse radiation transmission processes. However the scattering of longwave radiation is much less than that of shortwave radiation because the leaf scale reflectance for longwave, $\alpha = 1 - \varepsilon_c$, is very close to 0, where $\varepsilon_c$ is canopy emissivity. Longwave radiation emitted by the canopy, $Q_{lc}$ is calculated as $\varepsilon_c \sigma T_c^4 (1 - \tau_d)$, where $\sigma$ is the Stefan–Boltzmann constant ($5.67 \times 10^{-7}$ W m$^{-2}$ K$^{-4}$), $T_c$ is the canopy temperature in Kelvin and
(1 − τ_d) accounts for the fraction of the canopy exposed. The longwave radiation emitted from the atmosphere, Q_li and snow surface, Q_le, are calculated as \( \varepsilon_a \sigma T_a^4 \) and \( \varepsilon_s \sigma T_s^4 \), where \( \varepsilon_a \) and \( \varepsilon_s \) are air and snow emissivity, and \( T_a \) and \( T_s \) are air and snow surface temperatures, respectively.

We use Satterlund's parameterization [Satterlund, 1979] of air emissivity for clear sky conditions

\[
\varepsilon_{acls} = 1.08 \left[ 1 - \exp\left( -\left( \frac{e_a}{100} \right)^{\frac{T_a}{2016}} \right) \right]
\] (2.33)

where \( e_a \) is air vapor pressure (Pa). To adjust for cloud cover we use

\[
\varepsilon_a = C_f + (1-C_f)\varepsilon_{acls}
\] (2.34)

where \( C_f \) is the cloud cover fraction.

2.3.3. Multiple Reflections Between the Canopy and Surface

The above canopy transmission parameterization represents multiple scattering within the canopy. There is however the opportunity for light to reflect multiple times between the canopy and surface. This section describes how these multiple reflections are numerically evaluated.

For solar radiation we treat the canopy as a single layer with internal multiple scattering accounted for as described above. When each component of the solar beam (direct and diffuse) impacts the canopy; part of it is absorbed, part is reflected and part is transmitted. The reflected part is lost upwards. The transmitted part is absorbed or reflected at the surface; and the part reflected from the surface is again absorbed,
transmitted or reflected by the canopy leading to multiple reflections between the canopy and surface. These multiple reflections are assumed to be diffuse and the reflection by or transmission through the canopy is calculated using $\tau$ and $\beta$ from equations (2.29) and (2.30). Radiation that is reflected from the surface is calculated using snow surface albedo, $A$, which is modeled based on snow surface age and depth [Tarboton and Luce, 1996; Tarboton et al., 1995]. The effects of forest litter on the beneath canopy snow albedo are not modeled.

After multiple reflections the overall fractions of solar radiation from above transmitted and reflected by the canopy, $f_1$ and $f_3$ (Figure 2.7) are given by

$$f_1 = \frac{(1-A)\tau}{1 - A\beta(1-\tau_d)}$$  \hspace{1cm} (2.35)

$$f_3 = \frac{(1-A)\beta + A\tau\beta_d}{1 - A\beta_d(1-\tau_d)}$$  \hspace{1cm} (2.36)

Here $\tau$ and $\beta$ are direct or diffuse factors depending on whether the incident radiation is direct or diffuse. The fraction of radiation intercepted by the canopy, $f_2$ can be calculated subtracting equations (2.35) and (2.36) from 1. Summing up fractions from both direct and diffuse beams yields

$$Q_{sns} = f_{1b}Q_b + f_{1d}Q_d$$  \hspace{1cm} (2.37)

$$Q_{cns} = f_{2b}Q_b + f_{2d}Q_d$$  \hspace{1cm} (2.38)

$$Q_{rns} = f_{3b}Q_b + f_{3d}Q_d$$  \hspace{1cm} (2.39)
where $Q_{snes}$, $Q_{cns}$ and $Q_{rns}$ are sub-canopy net solar radiation, canopy net solar radiation and reflected solar radiation lost upwards, respectively. Here subscripts $b$ and $d$ in $f_1$, $f_2$ and $f_3$ refer to direct and diffuse solar radiation, respectively.

For longwave radiation we ignore multiple reflections as both plants and snow strongly absorb longwave radiation (absorptivity equal to emissivity close to 1). Like shortwave radiation, longwave radiation from all three sources is partitioned into fractions: $f_1$ (absorbed at surface), $f_2$ (absorbed in canopy) and $f_3$ (lost to sky).

Summing up fractions from all sources yields

\[ Q_{slnl} = f_{1i}Q_{sl} - Q_{ie} + f_{ie}Q_{le} + f_{1e}Q_{le} \]  
(2.40)

\[ Q_{cnl} = f_{2i}Q_{cl} + f_{2e}Q_{le} + f_{2c}Q_{lc} - 2Q_{le} \]  
(2.41)

\[ Q_{rnl} = f_{3i}Q_{sl} + f_{3e}Q_{le} + f_{3c}Q_{lc} + Q_{le} \]  
(2.42)

where $Q_{slnl}$, $Q_{cnl}$ and $Q_{rnl}$ are sub-canopy net longwave radiation, canopy net longwave radiation and reflected net longwave radiation lost upwards, respectively. The subscripts $i$, $e$, and $c$ in $f_1$, $f_2$ and $f_3$ are used to represent the radiation from sky, snow surface and from the canopy, respectively. The fractions $f_1$, $f_2$ and $f_3$ for longwave radiation are calculated as

\[ f_{1i} = \varepsilon_s \tau_d, f_{2i} = (1 - \tau_d)\varepsilon_c + \tau_d(1 - \varepsilon_s), f_{3i} = (1 - \tau_d)(1 - \varepsilon_c), \]

\[ f_{3e} = \tau_d, f_{2e} = (1 - \tau_d)\varepsilon_c, f_{1e} = (1 - \tau_d)(1 - \varepsilon_c), \]

\[ f_{1c} = \varepsilon_s, f_{3c} = \tau_d(1 - \varepsilon_s) \text{ and } f_{2c} = (1 - \tau_d)(1 - \varepsilon_s)\varepsilon_c \]  
(2.43)
With emissivities close to 1 there is a very small error in these equations that we neglect due to the neglect of multiple reflections. The net longwave and shortwave radiation calculated here are used with other energy fluxes in the snowmelt model energy balance equations to provide the net energy that drives the snowmelt in the open, beneath the canopy or within the canopy.

2.4. Model Application

Simulations were performed for the period of January 2008 to July 2008, December 2008 to July 2009 and January 2009 to July 2010 to estimate the radiation and snowmelt in the open and within and below the deciduous and coniferous forest using the hourly meteorological inputs of precipitation, temperature, wind speed and relative humidity. For forested areas, the open site meteorological variables are assumed representative of conditions at a height of 2 m above the forest canopy. Wind speeds within and beneath the canopy were calculated working downwards from above the canopy using exponential and logarithmic wind profiles [Bonan, 1991; Koivusalo, 2002]. Input precipitation data were taken from the SNOTEL site located in a small opening in the conifer forest and the other meteorological input data were obtained from the shrubs B (SB) open site (Figure 2.1). Leaf area index values for conifer and deciduous forest were chosen based on the ranges of values that are found in the literature, but with adjustments within these ranges to fit our data. Canopy coverage fraction was estimated based on our field observations of the canopy, but not on formal measurements. Thermal conductivity of snow and soil were adjusted (calibrated) to obtain a better match between modeled and observed surface temperature at the central open site for the whole simulation period. This adjustment was needed to correctly estimate the energy fluxes,
including longwave radiation that is based on surface temperature. The thermal conductivity parameters obtained from calibration at the open site were used in both open and forest settings. By calibrating thermal conductivity at the open site we separate the calibration issue from the evaluation of the canopy radiation model that is the main focus of this paper. Other parameters follow the original UEB model [Tarboton and Luce, 1996; You, 2004] presented in Table 2.2.

The model is able to predict the SWE, snow surface temperature, snow average temperature, canopy wind speed, radiation, energy fluxes and interception for both open and forest areas. Measurements of the four radiation components and surface temperature were available for the years 2009 and 2010, but not 2008 at the open site. We compare measured and modeled radiation components for 2009 and 2010 to validate the models calculation of open (above canopy) radiation. We then drive the model by inputs of measured open incoming shortwave and longwave radiation for 2009 and 2010, and modeled incoming shortwave and longwave radiation in 2008. 2008 serves as a check of the more complete model including atmospheric radiation parts. We evaluate the modeling of canopy radiation transmission processes by comparing modeled and observed below canopy net radiation and SWE. The SWE comparisons serve as an aggregate test of all aspects of the model, not limited to correct radiation transmission.

2.5. Simulation Results

2.5.1. Four Radiation Components and Surface Temperature

The four radiation components were continuously measured at the central tower site for two winter periods 2008/9 and 2009/10 and compared with model outputs.
aggregated to daily time scale so as to mask the effect of diurnal fluctuations and to better see daily total comparisons (Figure 2.8). Simulated values of solar radiation (incoming and reflected) and longwave radiation (incoming and outgoing) compared well with the observations for the two years with measured radiation data (2009 and 2010). The modeled incoming radiation that tracks observations reasonably well confirms cloud cover and atmospheric transmittivity parameterizations based on diurnal temperature range. The modeled outgoing radiation that tracks observation reasonably well serves to check the model albedo and surface temperature and emissivity representations. The high correlation and modest BIAS and RMSE values in scatterplots (Figure 2.8), relative to the ranges of these measurements also confirm the model effectiveness. Some of the differences may also be due to measurement errors such as the sensor sometimes having snow on it in this winter environment.

The outgoing longwave radiation, and many other fluxes at the snow surface, are functions of the snow surface temperature, which itself results from the balance of energy fluxes to and from the surface. This is why the representation of surface temperature by a snowmelt model is important. The model predictions of surface temperature at the open (snow/shrub) central tower site compared reasonably well to measured values (Figure 2.9). Some zero values seen in the model differed from observed values (Figure 2.9) because of the model retaining snow a few days longer than was observed. Temperature values above 0 °C occur on days without snow.

2.5.2. Net Radiation

The model simulation of below canopy net radiation was compared with the net radiation measured below conifer (CA) and deciduous (DB) forest canopy, aggregated to
daily time scale (Figure 2.10). The model predictions of net radiation followed the below canopy net radiation measurements reasonably well with correlation of about 0.90 for both forest types (Figure 2.10). Also, small BIAS and RMSE values were observed. In Figure 2.10 the scatterplots for 2008 where inputs were modeled radiation are separated from scatterplots for 2009 and 2010 where inputs are measured longwave and shortwave radiation from the open site. Both the time series and scatterplots for 2008 showed that the predictions of below canopy net radiation from the modeled above canopy radiation were not significantly different than those predicted using the measured above canopy radiation as input. For all these results the modeled net radiation tended to have a slight overprediction bias compared to the measurements in the early period. However the model showed relatively good agreement with observations during spring, which is important for calculating melt. The BIAS and RMSE values were found slightly higher for deciduous forest in comparison to conifer forest.

To further evaluate the model, modeled beneath canopy daily net radiation versus open area daily net radiation was compared to observations of the same quantities each year for both conifer and deciduous forest (Figure 2.11). The solid lines in the Figure are linear least square fits constrained to go through the origin: red for simulated and black for the measured values, with the slopes given in the plots. These graphs show what fraction of open net radiation is measured beneath the canopy and how the model is able to represent this for both coniferous and deciduous forest. These figures indicate a slight over prediction bias in the model.

The original UEB model uses linear relationships to reduce shortwave, longwave or net radiation beneath the canopy based on forest cover fraction, \( F \). Wind speed and the
corresponding heat and vapor fluxes are reduced by factor \((1-0.8F)\). Table 2.3 compares the new model simulated radiation with old model results. The old model predictions of beneath canopy radiation, especially the longwave radiation is very low. The old model reduces the incoming longwave radiation beneath the canopy, however the beneath canopy longwave radiation increases because of the higher emissivity of the canopy in comparison to the atmosphere. Also, while calculating beneath canopy radiation, this model does not consider leaf area index and provides similar solutions for two different forest types with different leaf area with same canopy cover fraction.

2.5.3. Snow Water Equivalent

Snow depths were monitored in the field by manually probing depth at twenty one locations and automatically with snow depth sensors mounted in twelve weather station towers (Figure 2.1). These depths were used with pit snow densities to derive snow water equivalent (SWE) in each of the vegetation classes. The model was initialized with measured SWE values (from snow depth sensors) on April 1st and run for the period between April 1st to June 30th to simulate the SWE values that were compared with observations made in the open, and beneath the deciduous and coniferous canopies (Figure 2.12). The simulation period was chosen to cover the melt period only, because the canopy radiation transmission is dominant in driving snowmelt, while other processes like interception and sublimation are more important earlier in the snow season. The observed SWE values (from depth sensors) below the conifer and deciduous forest are averages of the measurements in each forest type. The observed SWE for the open area is taken from a single site (SB) chosen because this site was least affected by wind drift and scouring. All the meteorological input variables used in this work were taken from
the SB site. Field surveyed SWE values were quite variable. The field surveyed SWE values for each vegetation class presented here are from locations selected to have their first SWE value most closely matching the SWE value for that vegetation class calculated from snow depth sensors and used to initialize the model. The snow melt and SWE values in the open area and beneath the deciduous and conifer forest canopy were reasonably predicted by the model.

2.6. Discussion

The radiation transmission model we developed is based on a simple two stream approximation that uses leaf area index as a key parameter and provides solution similar to Beer’s law but adjusted for multi scattering. The model is not intended to replace detailed multilayer radiation transfer models that consider the leaf orientation, inclination and distribution for each layer separately, but is suggested as a parsimonious approach when detailed information for each canopy layer is not available.

Overall, in examining the results, we see that the model simulated radiation values were in general agreement with the observed radiation values below different forest canopies. We found that there was a tendency to over predict early season net radiation (Figure 2.10) and overall slightly over predict the fraction of open net radiation found beneath a canopy (Figure 2.11). These effects were generally small and may be due to many factors. The radiation transmission model has a number of simplifications and does not represent canopy architecture, leaf orientation and layering effects. The model calculates average radiation beneath the canopy ignoring vertical and horizontal forest heterogeneity that result in spatial variability of radiation beneath the canopy. Also, the radiation sensor may not have been ideally placed to measure average radiation.
There are also uncertainties associated with the leaf level reflectances that were taken from the literature and estimates of leaf area index. There might also be measurement errors. During the early winter the upper part of the net radiometer had a tendency to catch snow which may result in bias in the measurements. There could also be uncertainty in the partitioning of incoming solar radiation. As direct and diffuse radiation attenuates differently in the canopy, the uncertainty in partitioning may also lead to errors in canopy radiation transmission processes. Small errors in predicting the canopy or surface temperature may cause errors in representing the longwave radiation that has a large contribution to net radiation. Also, the albedo of snow beneath the forest canopy is influenced by the forest litter.

The radiation transfer processes in the conifer canopy was better represented by the model in comparison to that in deciduous canopy (Figure 2.10). The problem in the deciduous site could be the poor representation of canopy structure. In our simulation we assumed similar leaf structures and reflectivity for both deciduous and conifer trees. However the emissivity and scattering characteristics of these two species can be different, as one is leaved and the other is leafless tree during the winter.

Given all the uncertainties and assumptions in the model, the model seems to be successful in terms of predicting the net radiation for snowmelt (Figure 2.10). The model’s generally good prediction of net radiation is reflected in SWE and snowmelt comparisons for open, beneath deciduous and conifer forest canopy (Figure 2.12). Slower ablation as forest density increases (open to deciduous to conifer) is evident in the observations and captured by the model, reflecting the model's capability to, in aggregate,
represent the processes driving snow melt in open and forested areas, with appropriate sensitivity to forest type.

In the SWE comparisons using the full new UEB model, there are model changes in terms of the representation of other canopy processes such as snow interception/sublimation and turbulent fluxes of sensible heat and latent heat that have not been fully described or evaluated in this paper and that do, to some extent impact the results in Figure 2.12, and even to some extent the net radiation comparisons since they impact surface temperature. It is simply not possible to isolate and evaluate only one set of processes in a system such as snow under a canopy where there are many interacting processes. Our focus on the melt period where radiation dominates the beneath canopy latent heat and sensible heat fluxes which totaled about 2% of beneath canopy net radiation serves as the best possible validation of the new radiation components added. Future work will more comprehensively evaluate the other new model components.

This model has parameterized radiation at the scale of a few trees in terms of leaf area index and canopy coverage. It could be used to simulate distributed snowmelt in heterogeneous settings with different grid values of slope and aspect (to account for topography) and leaf area index and canopy coverage to quantify the vegetation, taking advantage of advancing capability for remote sensing of these quantities [e.g. Fassnacht et al., 1997; Running et al., 1989; Zheng and Moskal, 2009]. Our model does not address transition effects such as solar radiation penetration to snow beneath a forest canopy near an opening, or shading of open areas by nearby forests. Further study to understand and quantify the impacts and importance of transitions on snow accumulation and melt is warranted.
2.7. Conclusions

We developed a simple canopy radiation transfer model that looks similar to Beer's law but considers the multiple scattering and reflection of radiation in the canopy based on two radiation streams, upward and downward. The model estimates the radiation beneath the canopy, which is important to predict the snowmelt responsible for water supply, using leaf area index as the key canopy parameter. The model results agreed well with observed surface temperature and radiation at an open site and net radiation and SWE values beneath coniferous and deciduous forest canopies. The model had a weakness in predicting the radiation beneath the canopy during the early winter; however the prediction of radiation for the late winter and spring period was better. The model was able to capture the differences in ablation between open and forested areas and in coniferous and deciduous forest.

The canopy radiation transmission model in this work is an advance over Beer’s law which does not account for multiple scattering of radiation. The solution for multiple scattering in a canopy with finite depth using the two stream approximation is, to our knowledge, new. The findings from this work may be of interest not only to people who want to use the improved UEB model but also to the wider snow modeling community who want to better predict the beneath canopy radiation and energy balance with a parsimonious parameterization of the penetration of radiation through canopy in a forested environment.

References


Nijssen, B., and D. P. Lettenmaier (1999), A simplified approach for predicting shortwave radiation transfer through boreal forest canopies, *J. Geophys. Res.*, **104**(D22), 27859-27868


Storck, P., D. P. Lettenmaier, and S. M. Bolton (2002), Measurement of snow interception and canopy effects on snow accumulation and melt in a mountainous


U.S. Army Corps of Engineers (1956), Snow hydrology, summary report of the snow investigations, U.S. Army Corps of Engineers, North Pacific Division, Portland, Ore.


Table 2.1. Site variables

<table>
<thead>
<tr>
<th>Sites/Variables</th>
<th>Open</th>
<th>Deciduous</th>
<th>Conifer</th>
</tr>
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<tbody>
<tr>
<td>Leaf area index</td>
<td>0.0</td>
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<td>4.5</td>
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<td>Canopy cover fraction</td>
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<td>0.7</td>
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<td>15.0</td>
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<td>Air temperature above which precipitation is all rain ($T_r$)</td>
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<td>20 m hr$^{-1}$</td>
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<td>Interception unloading rate ($U_i$)</td>
<td>0.00346 hr$^{-1}$</td>
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Table 2.3. Comparison of new and original UEB model radiation components with some measurements

<table>
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<th>Mean energy fluxes (W m(^{-2})) averaged for April 1 to June 30 melt period 2009 and 2010</th>
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<th>Deciduous</th>
<th>Conifer</th>
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<td>Surface/sub-canopy solar radiation</td>
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<td>(Q_{ss}) ↓</td>
<td>231.1</td>
<td>-</td>
<td>147.1</td>
</tr>
<tr>
<td>(Q_{ss}) ↑</td>
<td>-</td>
<td>-</td>
<td>82.3</td>
</tr>
<tr>
<td>Surface/sub-canopy longwave radiation</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(Q_{sl}) ↓</td>
<td>284.2</td>
<td>-</td>
<td>306.2</td>
</tr>
<tr>
<td>(Q_{sl}) ↑</td>
<td>-</td>
<td>-</td>
<td>315.9</td>
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<tr>
<td>Surface/sub-canopy net solar radiation</td>
<td></td>
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<td></td>
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<tr>
<td>(Q_{snsi}) ↓</td>
<td>-</td>
<td>-</td>
<td>64.8</td>
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<tr>
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<td></td>
</tr>
<tr>
<td>(Q_{sni}) ↓</td>
<td>-</td>
<td>-</td>
<td>-9.7</td>
</tr>
<tr>
<td>Surface/sub-canopy net radiation</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>(Q_{sn}) ↓</td>
<td>70.3</td>
<td>39.3</td>
<td>55.1</td>
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Figure 2.1. Site map of the TW Daniels Experimental Forest showing weather station towers, vegetation, survey points, pits and SNOTEL site.
Figure 2.2. Partitioning of atmospheric transmission factor, $AT$, into direct and diffuse components, $AT_b$ and $AT_d$. 
Figure 2.3. Illustration of radiation attenuation through a canopy that results in Beer's law.

\[ Q = Q_0 e^{-\rho G y / \cos \theta} \]

\[ dQ = -QG\rho dy / \cos \theta \]
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Figure 2.9. Scatterplot of observed and modeled hourly surface temperature for the year 2008 and 2009 at central tower (snow/shrub).
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CHAPTER 3

TESTING ABOVE AND BELOW CANOPY REPRESENTATIONS
OF TURBULENT ENERGY FLUXES IN AN ENERGY
BALANCE SNOWMELT MODEL

Abstract

Turbulent fluxes of sensible and latent heat are important processes in the surface energy balance that drives snowmelt. Modeling these fluxes in a forested environment is complicated because of the canopy effects on the wind field. In this paper we present and test a turbulent flux model developed to represent these processes in an energy balance snowmelt model. In developing this model our goal is to parsimoniously represent these processes using readily available inputs of canopy height and leaf area index. Wind, and eddy covariance measurements made below and above the canopy were used to evaluate effectiveness of this simplified modeling approach. The model was able to capture above canopy sensible and latent heat fluxes very well. Correlation of up to 80% was observed between hourly observed and simulated flux values. The modeled values of below canopy latent heat fluxes were also close to the EC-measured values. However the model under predicted the below canopy sensible heat flux during the day time. The night time observed and modeled sensible heat fluxes were relatively close. Despite some uncertainties, the results we obtained were encouraging and suggest that reasonable predictions of components of turbulence and subsequent vapor losses from forested environments can be obtained with the parsimonious single layer representation of the canopy and a single surface snow layer in a snowmelt model.

Coauthored by Vinod Mahat, David G. Tarboton, and Noah P. Molotch
3.1. Introduction

Turbulent fluxes of heat and water vapor play an important role in the surface energy and mass balance. The forest canopy strongly influences these fluxes and impacts the energy balance that drives snowmelt and the partitioning of the snow between sublimation and runoff in a forested environment. A number of previous studies focused on snow-vegetation interactions have indicated the importance of radiation and turbulent fluxes in snow cover mass and energy balances [e.g. Bartlett et al., 2006; Ellis and Pomeroy, 2007; Ellis et al., 2010; Essery et al., 2003; Koivusalo, 2002; Link and Marks, 1999; Pearson et al., 1999; Tribbeck et al., 2004].

Snowmelt models need to represent how the available radiative energy incident to the vegetation canopy or snow surface is partitioned into sensible heat (heat), latent heat (water vapor) flux or is stored. This partitioning drives the mass and energy balance in forested areas. In comparison to an open area, forest increases the overall aerodynamic roughness altering the wind profile above the canopy. Within the canopy wind speed is reduced impacting turbulent fluxes originating from the canopy and surface beneath the canopy. Commonly used snowmelt models have been built to operate primarily in open areas where no forest canopy is present [e.g. Anderson, 1976; Jordan, 1991; Marks, 1988; Price, 1988; Price et al., 1976; Tarboton and Luce, 1996]. These models treat the canopy and the underlying surface as a single interface ignoring the separate turbulent exchange of heat and water vapor between the vegetation canopy and the atmosphere. To model snow melt processes in forested environments accurate characterization of turbulent fluxes above, within, and beneath the canopy is important.
A number of land surface models, not specific to a snow environment have been developed to estimate the turbulent transport of heat and water vapor between the surface, vegetation canopy and the atmosphere. These include two layer [e.g. Choudhury and Monteith, 1988; Shuttleworth and Wallace, 1985] or multiple layer [e.g. Bonan, 1991] models. In two–layer models, the model is described by a ground surface and a single vegetation canopy layer [Inclan and Forkel, 1995] while in multiple-layer models vegetation is split into several layers and the energy balance is solved for each layer in order to describe the heat and water vapor transport processes among different canopy components: roots, stems, leaves; and air around [Demarty et al., 2002].

To partition the energy fluxes between the surface and the canopy, representation of at least two layers, i.e., surface and the canopy; and calculation of both canopy and surface temperatures are required. Some snow studies [e.g. Ellis et al., 2010; Gelfan et al., 2004; Hellstrom, 2000; Koivusalo, 2002] have used the canopy and the surface as two separate layers to separate the snow mass in the canopy from the snow at the surface. These studies assumed canopy temperature is equal to air temperature. The assumption of canopy temperature equal to air temperature may lead to uncertainty in the energy balance as canopy temperature controls canopy longwave radiation emission. Recent work by Andreadis et al. [2009] and Essery et al. [2003] solved the canopy energy balance for canopy temperature to separately estimate the turbulent sensible and latent heat fluxes from the canopy, but this work did not include an evaluation of the sensible and latent heat fluxes that emerge from the surface below the canopy.

One direct way to measure the turbulent transfer of heat and mass over the snow is eddy covariance (EC) [Kaimal and Finnigan, 1994; Marks et al., 2008]. Eddy
covariance techniques have been successfully applied to estimate turbulent energy fluxes and the subsequent vapor losses from the snow pack in open areas [e.g. Culle et al., 2007], from snow beneath the canopy [e.g. Marks et al., 2008], from intercepted snow in the canopy [e.g. Nakai et al., 1999; Suzuki and Nakai, 2008], and from both the canopy and beneath canopy snow [e.g. Jarosz et al., 2008; Molotch et al., 2007]. These measurement techniques can be used to validate the fluxes and vapor losses estimated by snowmelt models. Below-canopy turbulence and subsequence vapor fluxes have been modeled and validated in comparison with EC measurements [e.g. Marks et al., 2008]. Similarly, above canopy turbulence and vapor fluxes have been modeled and validated in comparison with EC measurements [e.g. Parviainen and Pomeroy, 2000]. These studies have shown the importance of both below and above canopy turbulent energy fluxes to the energy and mass balance in snow modeling.

This paper focuses on how to calculate forest environment turbulent fluxes of sensible and latent heat in an energy balance snowmelt model. Its objective is to improve snowmelt modeling capability in heterogeneous watersheds. The forest canopy is modeled as a single layer with parameters, leaf area index, canopy cover and canopy height that determine turbulent fluxes given time varying inputs of wind, temperature and humidity above the canopy. This single forest canopy layer was added to the single layer Utah Energy Balance (UEB) snowmelt model [Tarboton and Luce, 1996; Tarboton et al., 1995] resulting in a two layer enhanced model that represents snow energy and mass balances within and beneath the canopy driven by inputs of radiation and weather from above the canopy. The new canopy layer includes representations for radiation, turbulent fluxes, and interception. The turbulent transfer component presented in this paper
combines logarithmic boundary layer wind speed profiles above the canopy and at the
ground/snow surface with an exponential wind speed profile within the canopy. This
paper examines the turbulent flux parts of this model using comparisons of model
simulated values with within and above canopy eddy covariance flux measurements and
within canopy wind speed measurements. The radiation component of the model
enhancements was evaluated by Mahat and Tarboton [Chapter 2]. Evaluation of the
interception component will be the subject of a future paper.

3.2. Study Sites and Measurements

The model was tested using data from the Niwot Ridge AmeriFlux study site,
Colorado and TW Daniels Experimental Forest (TWDEF) site, Utah. At Niwot Ridge
EC measurements and wind speeds were compared to model values while the data from
TWDEF were used to provide an additional evaluation of beneath canopy wind.

3.2.1. Niwot Ridge Ameriflux Study site

The Niwot Ridge Ameriflux site is located at an elevation of 3050 m in the Rocky
Mountains of Colorado (40°1'58"N; 105°32'47"W) approximately 50 km northwest of
Denver (Figure 3.1). The forest surrounding the flux tower is dominated by subalpine fir
(Abies lasiocarpa), Engelmann spruce (Picea engelmannii) and lodgepole pine (Pinus
contorta) [Molotch et al., 2007]. The growing season’s maximum leaf area index is 4.2
m² m⁻² with a canopy height averaging 11.4 m, and gap fraction 17%. The forest slopes
gently (6-7%) and uniformly with elevation increasing from east to west. Average annual
precipitation is about 800 mm of which about 80% is snow [Caine, 1995]. Winds at the
site tend to originate from the west, particularly in the winter when periods of high wind speed and neutral atmospheric stability conditions are frequent [Turnipseed et al., 2002].

Above and below-canopy EC systems were mounted at heights of 21.5 and 1.7 m above-ground, respectively, on two different towers approximately 20 m apart. The EC systems consisted of three-dimensional (3-D) sonic anemometer (Campbell CSAT-3) to measure the wind vector \((u, v, w)\) and air temperature, and an infrared gas analyzer (IRGA-6260, Licor LI-7500) to measure water vapor and carbon flux at 10 Hz. Precipitation was measured with a 385-L, Met One rain and snow gage at a height of 12 m on the above-canopy EC tower. To improve the precipitation gauge catch efficiency; an Alter gauge shield was used. The meteorological measurements taken at this site include air temperature (CSAT-3, Campbell Scientific) and wind (Propvane-09101, RM Young Inc.) at heights of 21.5 and 1.7 m, and relative humidity (HMP-35D, Vaisala, Inc) at a height of 21.5 m above the ground surface. Ground heat flux (HFT-1, REBS) was measured at about 10 cm below the ground surface.

3.2.2. TW Daniels Experimental Forest Site (TWDEF)

The TW Daniels Experimental Forest (TWDEF) site is located at an elevation of 2700 m, about 30 miles North–East of Logan, Utah (41.86° N; 111.50° W). Average annual precipitation is about 950 mm of which about 80% is snow. Vegetation is classified into deciduous forest (Aspen), coniferous forest (Engelmann spruce and subalpine fir), open meadows consisting of a mixture of grasses and forbs, and shrub areas dominated by sagebrush. This site continuously monitors temperature and humidity (Vaisala HMP50), wind (Met One, 014A), and net radiation (Kipp & Zonen NR-Lite) in each of the four vegetation classes. These instruments were placed at heights above the
ground of about 2.5 m in conifer, 4.5 m in deciduous and 4 m in shrub and grass classes.

Additional details about this and other instrumentation and further site information are given by Mahat and Tarboton [see Chapter 2].

3.3. Model Description

The UEB snowmelt model [Tarboton and Luce, 1996] is a physically-based point energy and mass balance model for snow accumulation and melt. Snowpack is characterized using three state variables, namely, snow water equivalent, $W_s$, (m), the internal energy of the snowpack and top layer of soil, $U_s$, (kJ m$^{-2}$), and the dimensionless age of the snow surface used for albedo calculations. The UEB model is a single layer model. $U_s$ and $W_s$ are predicted at each time step based on the energy balance. Details of original and other aspects of enhanced UEB model are given by Tarboton et al. [1995], Tarboton and Luce [1996], You [2004] and Mahat and Tarboton [see Chapter 2].

In this paper the enhancements to UEB for the representation and evaluation of below canopy wind, and atmospheric transport of heat and water vapor in the forested environment are presented. The canopy component is modeled as a single layer, which added to the original single layer UEB model results in a two-layer model that represents the surface and the canopy snow separately. Energy balances are solved iteratively for each layer to provide outputs of surface temperature and canopy temperature based on which the above and below canopy turbulent fluxes are computed. The quantity and state of snow in the canopy is represented by a new state variable, canopy snow water equivalent ($W_c$). We assume that the energy content of intercepted snow in the canopy is negligible so canopy temperature, including snow in the canopy, is assumed to adjust to
maintain energy equilibrium, except when this requires canopy temperature to be greater than freezing when snow is present in the canopy, in which case the canopy temperature is set to freezing and the extra energy drives the melting of snow in the canopy.

In the enhanced UEB the changes with time of state variables, $U_s$, $W_s$ and $W_c$, are determined by following three equations

\[
\frac{dU_s}{dt} = Q_{si} + Q_g - Q_{ms} \quad (3.1)
\]

\[
\frac{dW_s}{dt} = P_r + P_s - i + R_m + M_c - E_s - M_s \quad (3.2)
\]

\[
\frac{dW_c}{dt} = i - R_m - M_c - E_c \quad (3.3)
\]

where energy fluxes are: combined surface energy input, $Q_{si}$, ground heat flux, $Q_g$, and advected heat removed by melt water $Q_{ms}$. Mass fluxes are rainfall, $P_r$, snowfall, $P_s$, canopy interception, $i$, mass release from the canopy, $R_m$, melt water drip from the canopy snow, $M_c$, melt from the surface snow, $M_s$, sublimation from the canopy snow, $E_c$, and sublimation from the surface snow, $E_s$. Terms in the energy balance equation are expressed per unit of horizontal area in kJ m$^{-2}$ hr$^{-1}$. Terms in the mass balance equations are expressed in m hr$^{-1}$.

The combined surface energy input is given by

\[
Q_{si} = Q_{sat} + Q_{at} + Q_{ps} + Q_{hs} + Q_{ct} \quad (3.4)
\]
where $Q_{sns}$ is sub-canopy net shortwave radiation, $Q_{snl}$ sub-canopy net longwave radiation, $Q_{ps}$ advected heat from precipitation, $Q_{ahs}$ sensible heat flux, and $Q_{ae}$ latent heat flux due to sublimation/condensation.

Energy content of the intercepted snow is considered negligible. To reflect this, the canopy energy balance is written

$$Q_{cns} + Q_{cnl} + Q_{pc} + Q_{hc} + Q_{ec} - Q_{mc} = 0$$

(3.5)

where $Q_{cns}$ is canopy net shortwave radiation, $Q_{cnl}$ is canopy net longwave radiation, $Q_{pc}$ is net advected heat from precipitation to the canopy, $Q_{hc}$ is sensible heat to the canopy, $Q_{ec}$ is latent heat to the canopy, and $Q_{mc}$ is advected heat removed by melt water from the canopy.

In equations (3.4) and (3.5) $Q_{sns}$, $Q_{ahs}$, $Q_{es}$, $Q_{cnl}$, $Q_{hc}$ and $Q_{ec}$ are functions of the surface and canopy temperatures $T_s$ and $T_c$, as well as inputs that drive the model. In the case of longwave radiation terms $Q_{snl}$ and $Q_{cnl}$ this functionality is based on Stefan-Boltzmann equations for longwave radiation and the model for transmission of radiation through and emission of radiation by the canopy described by Mahat and Tarboton [see Chapter 2]. Expressions for the turbulent flux quantities $Q_{hs}$, $Q_{es}$, $Q_{hc}$ and $Q_{ec}$ are developed below. UEB uses the modified Force-Restore approach [Luce and Tarboton, 2010] to equate $Q_{si}$ in equation (3.4) to conduction into the snow as a function of $T_s$, its past values, and the average temperature of the snow. We initially set $Q_{mc}$ to 0. The result is two nonlinear equations, (3.4) and (3.5) that depend only on the unknowns $T_c$ and $T_s$. We solve these iteratively, using a Newton-Rhapson approach. In the event that
\( T_c \) is above freezing and there is intercepted snow in the canopy \((W_c > 0)\) we set \( T_c \) to 0 and equation (3.5) is used to evaluate \( Q_{mc} \). In the event that \( T_s \) is above freezing and there is snow on the ground \((W_s > 0)\), we evaluate \( Q_{si} \) with \( T_s \) set to freezing. The resulting extra surface energy above that conducted into the snow calculated using the Force-Restore approach is used to calculate the amount of melt generated at the surface. This melt infiltrates into the snowpack and the energy is added to \( U_s \) during the solution of equations (3.1) to (3.3) which are advanced through time using a Predictor-Corrector approach [Gerald, 1978; Tarboton et al., 1995].

3.3.1. Turbulent Energy Fluxes

The main structure of the flux model and the basic equations used here are similar to those originally devised by Norman et al. [1995] in their series network two-layer flux model. The approach uses the temperature and vapor pressure differences between the snow surface, canopy air and the snow in the canopy to calculate the turbulent flux exchanges between snow on the ground, canopy and the atmosphere (Figure 3.2).

\[
Q_h = \frac{\rho a C_p(T_a - T_{ac})}{R_a} \quad (3.6)
\]

\[
Q_{hs} = \frac{\rho a C_p(T_{ac} - T_s)}{R_c} \quad (3.7)
\]

\[
Q_{hc} = \frac{\rho a C_p(T_{ac} - T_c)}{R_c} \quad (3.8)
\]

\[
Q_e = \frac{1}{R_l} \frac{h_v 0.622}{R_d T_{ac}} (e_a - e_{ac}) \quad (3.9)
\]

\[
Q_{es} = \frac{1}{R_c} \frac{h_v 0.622}{R_d T_{ac}} (e_{ac} - e_s(T_s)) \quad (3.10)
\]
where $Q_h$ and $Q_e$ are total fluxes of sensible heat and latent heat from the atmosphere which is partitioned into $Q_{hc}$ and $Q_{ec}$, the sensible and latent heat flux contributions to the forest canopy, and $Q_{hs}$ and $Q_{es}$, are sensible and latent heat flux contributions to the surface. In these equations $R_a$ and $R_c$ are the aerodynamic resistances to heat and vapor transport between the air in the canopy and air above, and between the snow surface and the air in the canopy, respectively (hr m$^{-1}$), $R_l$ is the bulk leaf boundary layer resistance (hr m$^{-1}$), $T_{ac}$ is canopy air temperature ($^\circ$C), $T_s$ is surface temperature ($^\circ$C), $T_c$ is canopy temperature ($^\circ$C), $T_a$ is above canopy air temperature ($^\circ$C), $e_{ac}$ is canopy air vapor pressure (Pa), $e_a$ is above canopy air vapor pressure (Pa) and $e_s(T_s)$ and $e_s(T_c)$ are surface and canopy saturated vapor pressures calculated as functions of snow and canopy temperatures, respectively. Standard formulae for saturation vapor pressure over ice are used when the temperature is below freezing [Lowe, 1977]. $\rho_a$ is air density, $C_p$ is air specific heat capacity (1.005 kJ kg$^{-1}$ $^\circ$C$^{-1}$), $h_v$ the latent heat of sublimation (2834 kJ kg$^{-1}$), $R_d$ is the dry gas constant (287 J kg$^{-1}$ K$^{-1}$). The partitioning of $Q_h$ and $Q_e$ into $Q_{hc}$, $Q_{ec}$, $Q_{hs}$ and $Q_{es}$, is given by

$$Q_h = Q_{hc} + Q_{hs}$$

(3.12)

$$Q_e = Q_{ec} + Q_{es}$$

(3.13)

These equations facilitate the evaluation of $T_{ac}$ and $e_{ac}$ as functions of $T_a$, $T_s$, $T_c$, $e_a$, and the saturated vapor pressure function $e_s(.)$.  

$$T_{ac} = \frac{T_c}{R_l} + \frac{T_s}{R_c} + \frac{T_a}{R_a} \cdot \left( \frac{1}{R_l} + \frac{1}{R_c} + \frac{1}{R_a} \right)$$

(3.14)
Substitution of equations (3.14) and (3.15) in equations (3.7), (3.8), (3.10) and (3.11) results in expressions for \( q_{hc}, q_{ec}, q_{hs} \) and \( q_{es} \), that are functions only of the inputs and two unknowns \( T_c \) and \( T_s \). Substituting these into equations (3.4) and (3.5) results in all terms on the right of equation (3.4) and all terms except for \( Q_{mc} \) in equation (3.5) being expressed in terms of \( T_c \) and \( T_s \) as required for their solution.

3.3.1.1. Wind Profile and Aerodynamic Resistances

The wind profile is assumed to be logarithmic above the canopy \((z > h)\), exponential within the canopy \((z_{ms} < z < h)\) and again logarithmic over the snow surface on the ground \((z < z_{ms})\) as is typical in the literature [e.g. Bonan, 1991; Cionco, 1972; Dolman, 1993; Koivusalo, 2002] (Figure 3.3). Here \( z \) is height above the ground or snow surface, \( z_{ms} \) a reference height above the surface (2 m) and \( h \) is canopy height. The logarithmic and exponential wind profiles [Bonan, 1991; Brutsaert, 1982] are

\[
u(z) = \frac{1}{k} u^* \ln \left( \frac{z-d}{z_0} \right) \quad \text{for } z > h \text{ and } z < z_{ms} \quad (3.16)
\]

\[
u(z) = u_h \exp[-n(1 - z/h)] \quad \text{for } z_{ms} < z < h \quad (3.17)
\]

where \( u(z) \) is wind speed at height \( z \), \( u_h \) is wind speed at canopy height \( h \), \( n \) is an exponential decay coefficient, \( d \) is zero-plane displacement height and \( z_0 \) is the roughness length. \( z_0 \) is equal to \( z_{oc} \) (roughness length for the top of the canopy boundary layer) for \( z > h \) and \( z_{os} \) (snow surface roughness length) for \( z < z_{ms} \). \( u^* \) is shear velocity and \( k \) is von Karman's constant \((k = 0.4)\).
For the surface logarithmic profile layer \((z<z_{ms})\) we take \(d=0\). For the canopy logarithmic profile layer \((z>h)\), \(d\) and \(z_{oc}\) are estimated as functions of tree height \((h)\), tree profile shape and leaf area index \((L)\) following Shaw and Pereira [1982] as

\[
d = h \left( 0.05 + \frac{y^{0.2}}{2} + \frac{(y-1)^{20}}{20} \right)
\]

(3.18)

\[
z_{oc} = h \left( 0.23 - \frac{0.25}{10} - \frac{(y-1)^{67}}{67} \right)
\]

(3.19)

where \(y\) is an integer indicating one of the three basic forest profiles [e.g. Massman, 1982; Meyers et al., 1998]. \(y = 1\) for young pine, \(y = 2\) for leafed deciduous tree and \(y = 3\) for old pine with long stems and clumping at the top.

The model calculates the wind profile working downwards from the top. Above canopy wind speed \(u_m\) at height \(z_m\) \(>(h)\) above the ground is taken as input and used in (16) to solve for the above canopy \(u^*\). With this \(u^*\), and \(z=h\), (16) gives \(u_h\). Then from (17) \(u_{ms}\) is determined at height \(z_{ms}\). This \(u_{ms}\) is used in (16) to quantify \(u(z)\) below \(z_{ms}\).

The canopy and the surface interact with the atmosphere through above and below canopy aerodynamic resistances, \(R_a\) and \(R_c\) and one bulk leaf resistance, \(R_l\) (Figure 3.3). The aerodynamic resistance \(R_a\) is imposed on the transfer of heat and water vapor between the reference height \(z_m\) and the canopy source height \(d + z_{oc}\) which is taken to be the center of aerodynamic drag (effective momentum sink and source) in the canopy [Choudhury and Monteith, 1988; Inclan and Forkel, 1995; Sellers et al., 1986].

Aerodynamic resistances, \(R_a\) and \(R_c\) are calculated based on K-theory, which states aerodynamic resistance is the integral of the inverse of the eddy diffusion, \(K\) [Choudhury and Monteith, 1988; Demarty et al., 2002; Dolman, 1993; Koivusalo, 2002; Shuttleworth...
and Gurney, 1990] which is assumed to be proportional to the local wind speed [Inclan and Forkel, 1995]. Although there are many questions as to the validity of $K$-theory for within-canopy transfer it is the only practical alternative at present for the information available [Dolman and Wallace, 1991].

Using $K$-theory, above canopy aerodynamic resistance, $R_a$ (for $d + z_{oc} < z < z_m$) and below canopy aerodynamic resistance, $R_c$ (for $z_{os} < z < d + z_{oc}$) are calculated as

$$R_a = \int_h^{z_m} \frac{dz}{K} + \int_{d + z_{oc}}^{h} \frac{dz}{K_c}$$  \hspace{1cm} (3.20)

$$R_c = \int_{z_{ms}}^{d + z_{oc}} \frac{dz}{K_c} + \int_{z_{os}}^{z_{ms}} \frac{dz}{K}$$  \hspace{1cm} (3.21)

Here $K$ and $K_c$ are eddy diffusion coefficients that correspond to logarithmic and exponential wind profiles, respectively. These coefficients are given by [Dolman, 1993]

$$K = k u^* (z - d)$$  \hspace{1cm} (3.22)

$$K_c = K_h \exp[-n(1 - z/h)]$$  \hspace{1cm} (3.23)

where $K_h$ is the eddy diffusion coefficient, $K$, for the canopy evaluated at height $h$. Its value for the input $u = u_m$ at height $z_m$ is determined as

$$K_h = \frac{k^2 u_m (h - d)}{\ln \left( \frac{z_m - d}{z_{oc}} \right)}$$  \hspace{1cm} (3.24)

Substituting values of $K$ and $K_c$ and integrating equations (3.20) and (3.21) yields

$$R_a = \frac{1}{k^2 u_m} \ln \left( \frac{z_m - d}{z_{oc}} \right) \ln \left( \frac{z_m - d}{h - d} \right) + \frac{h}{K_h n} \left[ \exp \left\{ n - n \left( \frac{d + z_{oc}}{h} \right) \right\} - 1 \right]$$  \hspace{1cm} (3.25)

$$R_c = \frac{h \exp(n)}{K_h n} \left[ \exp \left\{ -n \left( \frac{z_{ms}}{h} \right) \right\} - \exp \left\{ -n \left( \frac{d + z_{oc}}{h} \right) \right\} \right] + \frac{1}{k^2 u_{ms}} \ln \left( \frac{z_{ms}}{z_{os}} \right)^2$$  \hspace{1cm} (3.26)
The below canopy resistance, $R_c$ is corrected for atmospheric stability to improve the simulation of snowmelt [Essery et al., 2003; Storck, 2000]. To make the correction for stable and unstable conditions we use the expression suggested by Choudhary and Monteith [1988].

$$R_c = \frac{R_c}{(1 - 5R_i)^2} \quad \text{stable} \quad 0 < R_i \leq R_{imax} \quad (3.27)$$

$$R_c = \frac{R_c}{(1 - 5R_i)^{3/4}} \quad \text{unstable} \quad R_i < 0 \quad (3.28)$$

$$R_i = \frac{g(T_a - T_s)z_{ms}}{u_m^2[0.5(T_a + T_s) + 273.15]} \quad (3.29)$$

where $R_i$ is an estimate of the Richardson number, $R_{imax}$ is the upper limit of the Richardson number, and $g$ is the acceleration due to gravity.

### 3.3.1.2. Leaf boundary layer resistance

Bulk leaf boundary layer resistance is calculated based on wind, leaf dimension and leaf area distribution. Following Jones [1992] leaf boundary layer conductance, which is the reciprocal of the leaf boundary layer resistance is

$$G_b(z) \approx 0.01 \sqrt{u(z)/w} \quad (3.30)$$

where $G_b(z)$ is the boundary layer conductance (m s$^{-1}$) for a unit projected area of a leaf, $w$ is leaf width (m), and $u(z)$ is the wind speed (m s$^{-1}$) at height $z$.

Assuming leaf area is uniformly distributed with the tree height, the mean leaf conductance is obtained [Choudhury and Monteith, 1988] as

$$\overline{G_b} = \int_0^h 0.01 \sqrt{u(z)/w} \, dz/h \quad (3.31)$$
Substituting the height dependent value of $u(z)$ from equation (3.17) and integrating yields

$$
\overline{G_b} = \frac{0.02}{n} \sqrt{u_h/w} \left[1 - \exp\left(-\frac{n}{2}\right)\right]
$$

(3.32)

The mean canopy conductance per unit ground area is obtained multiplying mean leaf conductance ($\overline{G_b}$) with effective leaf area index $LF$, where $L$ is leaf area index and $F$ is canopy coverage fraction. Thus mean canopy resistance is

$$
R_l = 1/ (\overline{G_b} \times LF)
$$

(3.33)

In evaluation of $R_l$, we used a fixed value of 0.04 m for $w$ [Bonan, 1991; Dickinson et al., 1986].

3.3.2. Snow Interception and Water Vapor Flux

We use Hedstrom and Pomeroy’s [1998] event based snowfall interception model which is equivalent to Aston’s [1979] rainfall interception model to develop a continuous interception component for inclusion in the UEB. This model also estimates the snow mass unloading from interception using a mass unloading rate calculated based on the empirical relationship provided by Hedstrom and Pomeroy’s [1998]. Details of the snow interception and unloading processes will be presented in subsequent paper.

Sublimation terms $E_s$ and $E_c$ in equations (3.2) and (3.3) are determined from the corresponding latent heat fluxes using

$$
E_c = - \frac{Q_{ec}}{\rho_w h_v} \text{ and } E_s = - \frac{Q_{es}}{\rho_w h_v}
$$

(3.34)
where \( \rho_w \) is the density of water (kg m\(^{-3}\)). The negative sign reflects our convention that \( Q_{ec} \) and \( Q_{es} \) are energy additions to the canopy and surface, while \( E_c \) and \( E_s \) are losses. Total sublimation is obtained summing up \( E_c \) and \( E_s \).

### 3.3.3. Radiation

Radiation is the main energy source that drives turbulent energy fluxes, sublimation and snowmelt. To estimate the net radiation beneath, in, and above the canopy, we use an approach that describes the penetration of radiation through the canopy based on a two stream approximation accounting for multiple scattering. In the two stream approximation, upward and downward radiation is expressed using two differential equations quantifying the change in downward and upward radiation due to interception, absorption and scattering. This method assumes, as an approximation following Monteith and Unsworth [1990], that multiple scattering occurs along a single path, thereby avoiding the intractable complexity of scattering in multiple directions. It also considers multiple reflections between the canopy and surface treating direct and diffuse radiation separately. A detailed description of how net radiation is calculated beneath, in, and above the canopy in given by Mahat and Tarboton [see Chapter 2].

### 3.4. Eddy Covariance Flux Measurements

The eddy covariance method calculates turbulent fluxes of sensible heat, \( Q_h \), and latent heat, \( Q_e \), based on the covariances between the respective scalars (temperature and water vapor density) and vertical wind measured at high frequency.

\[
Q_h = \rho a c_p \bar{T}' \bar{W}' \quad (3.35)
\]
where $T', w'$ and $\rho_v'$ are deviations from the time average of temperature, vertical wind speed and water vapor density. An averaging time of 30 minutes of 10-Hz measurements was used. Post-processing corrections to the EC data included mathematical coordinate rotation of the mean lateral and vertical wind velocities to zero. Only the lateral component was corrected for in processing the sub-canopy data [Baldocchi and Hutchison, 1987]. The sonic anemometers’ virtual air temperatures were corrected, accounting for wind speed normal to the sonic path and humidity effects [Schotanus et al., 1983]. Energy balance closure was evaluated to derive linear regression coefficients (slope and intercept) using a linear regression between the half hourly estimates of the turbulent flux ($Q_h + Q_e$) and the difference between the radiation and ground heat flux ($Q_{net} - Q_g$). The relationship between above canopy combined turbulent flux ($Q_h + Q_e$) and ($Q_{net} - Q_g$) in Wm$^{-2}$ was $y=0.77x+13$ (R$^2 =0.89$; P$<0.01$). See Molotch et al. [2007] and Turnipseed et al. [2002, 2003] for complete details.

Sub-canopy and total (canopy and sub-canopy) sublimation losses were calculated by dividing below and above canopy EC-measured latent heat fluxes with $\rho_v$ and $h_v$ as in equation (3.34). The net canopy sublimation loss was calculated subtracting the sub-canopy sublimation from the total sublimation.

### 3.5. Model Application

The model was initialized with measured below canopy snow water equivalent and energy content state variables and run for the period between March 1 to April 30, 2002 using above canopy meteorological measurements of precipitation, air temperature,
humidity and wind at the Niwot Ridge Ameriflux site. The canopy snow water 
equivalent state variable was initialized at zero. The vegetation parameters used in the 
model are leaf area index, canopy height and canopy cover. Roughness length for the 
canopy was estimated using equation (3.19). Snow surface roughness length, $z_o$, beneath 
the canopy, and exponential decay coefficient $n$ were adjusted to have model results 
match measurements. The same $z_o$ value was used at all sites while $n$ was estimated at 
each site separately because effective leaf area indices vary. Other snowmelt model 
parameters used in this work follow the original UEB model [Tarboton and Luce, 1996; 
You, 2004] (Table 3.1) with site variables specific to this setting (Table 3.2).

The model is able to predict the snow surface temperature, snow average 
temperature, snow interception, below canopy wind speed, above and below canopy 
turbulent energy fluxes (sublimation) and radiation, and SWE. In this work we mainly 
focus on the prediction of below canopy wind, and turbulent energy fluxes and 
subsequent vapor losses from the snow above and below the canopy and evaluate them in 
comparison to observations. The wind model that calculates the beneath canopy wind 
was also tested using three years of wind data from TWDEF site. Vegetation parameters 
and other site variables used in the wind model for TWDEF are presented in Table 3.2. 
The bias (BIAS) and root mean square error (RMSE) were used as evaluation criteria.

3.6. Simulation Results

3.6.1. Wind

Simulated values of beneath canopy winds compared favorably with the 
observations made inside the forest at Niwot Ridge and inside the deciduous and conifer
forest at TWDEF (Figure 3.4). Inputs to the modeled wind calculation comprise the above canopy measured wind speed for Niwot Ridge and measured wind speed in the open area assumed to be equivalent to the above canopy wind for TWDEF. A higher correlation between modeled and observed values was found for Niwot Ridge in comparison to the TWDEF.

In comparison to the conifer forest wind, modeled deciduous forest wind was more correlated with observations at TWDEF. The mean observed wind value beneath the conifer canopy was about half of that observed beneath the deciduous canopy. The greater density of coniferous forest canopies, reflected by their higher leaf area index, has resulted in the lower beneath canopy wind speeds. Larger $n$ values in the model for coniferous canopies capture this effect.

### 3.6.2. Energy Fluxes

The model predictions of total (canopy plus sub-canopy) sensible heat flux, $Q_h$, followed the above canopy EC-measured sensible heat flux reasonably closely with correlation of 0.8 (Figure 3.5). While there is scatter in the modeled verses measured hourly values, the match of the cumulative plot of sensible heat flux (Figure 3.5c) indicates that hourly errors offset each other when aggregated over time.

The modeled total (canopy plus sub-canopy) latent heat flux, $Q_e$, tended to have a underprediction BIAS that increased towards the end of the simulation (Figure 3.6). There was more scatter indicated by the lower correlation of 0.47 and BIAS of 11.7 Wm$^{-2}$. The underprediction may be because of the model not accounting for the energy loss due to transpiration from the canopy which is not represented in this snow melt model. When there is no intercepted snow, the canopy latent heat is modeled as zero
while EC-measured latent heat is not zero due to transpiration. The top of Figure (3.6a) includes input precipitation and modeled interception storage that was used to separate out the times when there is intercepted snow present in the canopy. To examine the possible transpiration bias we compared measured and modeled latent heat flux, $Q_e$, only for the times when there was intercepted snow modeled to be present in the canopy (Figure 3.7). In Figure 3.7, the correlation increased and the BIAS decreased, compared to Figure 3.6, providing some support for this hypothesis.

The EC measurements below the canopy showed upward (away from the snow surface) sensible heat, $Q_{es}$, flux during the daytime and downward (towards the snow surface) sensible heat flux during the nighttime (Figure 3.8). The model captured the nighttime, downward sensible heat flux quite well but tended to underestimate the daytime, upward sensible heat flux.

Time series of hourly sub-canopy latent heat flux showed a general agreement between the observed and modeled values with BIAS 0.42 Wm$^{-2}$ and RMSE 22.54 Wm$^{-2}$ (Figure 3.9). Though the correlation value was not large, the overall cumulative latent heat flux from the snow surface for both measured and modeled values was similar.

### 3.6.3. Sublimation

A total of 54 mm of precipitation was recorded during the measurement period between March 1 to April 30, 2002 (Figure 3.10). The EC measurement above the canopy showed about 63 mm of sublimation over the same period. More sublimation than recorded precipitation could be due to sublimation of snow already on the ground at the start of this period and due to the transpiration from the snow free canopy being included in the measured sublimation. The model showed about 47 mm of total
sublimation from the canopy and sub-canopy during this two month period. The measured and modeled total sublimation comparison is effectively the same as the latent heat flux comparison of Figure 3.6c with end of period bias likely due to transpiration. Similarly the cumulative EC-measured and modeled sub-canopy sublimation are equivalent to the sub-canopy latent heat flux compared in Figure 3.9c.

The EC-measured net canopy sublimation was calculated by subtracting below canopy EC measurement from above canopy EC measurement, and compared with modeled canopy sublimation for the period between March 1 to April 6, 2002 (Figure 3.10). Extension to dates beyond April 6 was not possible due to missing EC-measured sub-canopy data after April 6. The cumulative modeled values tracked well with the EC measurements up to this date.

3.7. Discussion

This work has developed a flux model and compared the model simulated above and below canopy turbulent fluxes of sensible and latent heat with the EC-measured fluxes to evaluate the model performance. The model was generally able to represent the above canopy sensible heat flux, latent heat flux and also snow mass loss through sublimation. The model exhibited little or no BIAS in the simulation of these energy and mass fluxes above the canopy. The model’s predictions of below canopy wind, latent heat flux and subsequent snow mass loss were also good but the model did not do well in predicting below canopy sensible heat flux especially during the daytime. The night time sensible heat flux was reasonably captured by the model.

Within canopy wind speed was calculated as an exponential function of canopy height following [Bonan, 1991; Dolman, 1993; Koivusalo, 2002]. The exponential decay
coefficient, $n$, that is required to calculate the beneath canopy wind speed has been reported to have a value between 2 and 4 [Bonan, 1991; Brutsaert, 1982]. In testing the model with beneath canopy wind measurements at Niwot Ridge and TWDEF sites, we found that the $n$ values that resulted in beneath canopy wind speed best matching measurements ranged from 0.6 to 1.5 for leaf area index ranging from 1 to 4.5, with increasing $n$ values for increasing leaf area index. This range from 0.6 to 1.5, though determined based on limited data, has captured the sensitivity of within canopy wind speed to forest type and density.

We adjusted the snow surface roughness length beneath the canopy to match the measurements of beneath canopy latent heat flux and vapor loss. This adjustment to $z_o=0.1$ m from the previous open area value of 0.01 m may represent the increased turbulence in wind within a canopy, but may also express the limitations of assuming logarithmic wind and diffusivity profiles near the surface beneath a canopy. Above canopy solutions were found to be less sensitive to this snow surface roughness length and changed insignificantly with this adjustment.

Even after the surface roughness adjustment, the model did not capture the EC-measured beneath canopy sensible heat flux (Figure 3.8c). Most of the time the modeled temperature gradient was downward ($T_{ac} > T_s$) and that downward gradient resulted in downward sensible heat flux. The EC measurements showed downward sensible heat flux ($T_{ac} > T_s$) during the night time and upward sensible heat flux ($T_s > T_{ac}$) during the day time. Modeling a daytime upward sensible heat flux requires surface temperature greater than air temperature. The model holds temperature of the snow covered surface
no greater than freezing so is unable to represent these upward measurements of sensible heat flux by eddy covariance when $T_{ac} > 0$.

There could be many reasons why the modeled and measured beneath canopy sensible heat flux did not match. Although EC is treated as a measurement technique, there are many assumptions in its calculation that can be sensitive to a complex suite of site conditions [Marks et al., 2008]. Especially, the below canopy eddy-covariance flux measurements could be problematic because the underlying assumptions for this method are not expected to be generally valid in the conditions prevailing there: low wind speed, strong heterogeneity and intermittent turbulence [Baldocchi et al., 2000; Blanken et al., 1998; Constantin et al., 1999]. There may be modeling problems too. Uncertainties exist in assumptions of surface roughness length, resistances and the derivation of below canopy wind in the model. Assumed equal resistances for transfer of heat and water vapor used in the turbulence modeling may not be strictly valid below the canopy. It is difficult to say how well the assumptions were satisfied in both modeling and EC measurements in the prevailing conditions.

Despite some uncertainties in the beneath canopy solutions, the model results of latent heat fluxes and vapor losses are encouraging and the method used here to calculate these can be used in any snow model for the calculation of snow losses through sublimation in forested environments. Model results showed good agreement between the modeled and observed sublimation losses from beneath the canopy throughout simulation period and from above the canopy during the early half of the simulation period. Above canopy sublimation during the second half of the simulation was impacted by the canopy transpiration. Measurements of $\mathrm{CO}_2$ fluxes also obtained using EC were
used to confirm this. CO$_2$ measurements during the early half period showed a large positive values of CO$_2$ flux indicating large canopy respiration with little carbon uptake (transpiration). Significant snowfall and interception during this period limit canopy carbon uptake or transpiration. During the latter half of the simulation period, CO$_2$ measurements showed a significant carbon uptake implying transpiration. Therefore the second half above-canopy $Q_e$ or sublimation observations combine latent heat losses due to snow sublimation with transpiration that is not represented by the model.

The modeled canopy and beneath canopy snow sublimation losses were found to be 0.45 mm day$^{-1}$ and 0.3 mm day$^{-1}$, respectively over the study period. The canopy sublimation loss is similar to the canopy sublimation value (0.5 mm day$^{-1}$) reported by Parviainen and Pomeroy [2000] for Canadian boreal forest and slightly less than the value (0.65 mm day$^{-1}$, about 100 mm in 5 months) reported by Storck et al. [2002] for Pacific North West forest. The beneath canopy sublimation is similar to the beneath canopy sublimation value (0.25 mm day$^{-1}$) reported by Marks et al. [2008] for the Fraser Experimental Forest, Colorado.

3.8. Conclusions

This study has tested the turbulent flux components of a parsimonious canopy parameterization for an energy balance snowmelt model. The combined examination of both above and below the canopy turbulent flux representations represents an important check on this model parameterization. We found that the model was able to simulate fluxes generally in agreement with eddy covariance measurements at least when there was snow present in the canopy. Some divergence was noted due to transpiration, a
process not represented in the energy balance snowmelt model. There were also some unresolved discrepancies with beneath canopy sensible heat fluxes where measurements counter to the temperature gradient were not modeled.

Further work is needed to better quantify some of the model parameterizations and their relationship to vegetation properties. Specifically the dependence of wind exponential decay parameter on leaf area index or other canopy properties, and surface roughness beneath the canopy needs further evaluation. More broadly it is important to also evaluate the model over larger areas and examine the effects of vegetation on larger watershed scale response to snowmelt.

References


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Storck, P. (2000), Trees, snow and flooding: An investigation of forest canopy effects on snow accumulation and melt at the plot and watershed scales in the Pacific Northwest, 176 pp., Dept. of Civil and Environmental Engineering, Univ. of Washington, Seattle, Wa.


U.S. Army Corps of Engineers (1956), Snow hydrology, summary report of the snow investigations, U.S. Army Corps of Engineers, North Pacific Division, Portland, Ore.

Table 3.1. Model parameters

<table>
<thead>
<tr>
<th>Name</th>
<th>Values</th>
<th>Basis</th>
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</thead>
<tbody>
<tr>
<td>Air temperature above which precipitation is all rain ($T_r$)</td>
<td>3 °C</td>
<td>Tarboton et al. [1995], U.S. Army Corps of Engineers [1956]</td>
</tr>
<tr>
<td>Air temperature below which precipitation is all snow ($T_{sn}$)</td>
<td>-1 °C</td>
<td>Tarboton et al. [1995], U.S. Army Corps of Engineers [1956]</td>
</tr>
<tr>
<td>Emissivity of snow ($\varepsilon_s$)</td>
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<td>Tarboton et al. [1995]</td>
</tr>
<tr>
<td>Ground heat capacity ($C_g$)</td>
<td>2.09 kJ kg⁻¹°C⁻¹</td>
<td>Tarboton et al. [1995]</td>
</tr>
<tr>
<td>Nominal measurement of height for air temperature and humidity ($z_{ms}$)</td>
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<td>Tarboton et al. [1995]</td>
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<tr>
<td>Surface aerodynamic roughness ($z_{os}$)</td>
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<td>Adjusted from previous open area value of 0.01 m</td>
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<td>Soil density ($\rho_g$)</td>
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<td>Snow saturated hydraulic conductivity ($K_s$)</td>
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<td>Richardson number upper bound for stability correction ($Ri_{max}$)</td>
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<td>Koivusalo [2002]</td>
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<td>Emissivity of canopy (conifer/deciduous) ($\varepsilon_c$)</td>
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<td>Bonan [1991]</td>
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<td>Interception unloading rate ($U_j$)</td>
<td>0.00346 hr⁻¹</td>
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### Table 3.2. Site variables

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<th>Site Variables</th>
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</tr>
<tr>
<td>Latitude (Degree)</td>
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</tr>
<tr>
<td>Longitude (Degree)</td>
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</tr>
<tr>
<td>Branch interception capacity (kg m(^{-2}))</td>
<td>6.6</td>
</tr>
<tr>
<td>Average atmospheric Pressure (Pa)</td>
<td>70623</td>
</tr>
</tbody>
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Figure 3.1. Hillshade relief of the Niwot Ridge, Long Term Ecological Research Site, Colorado. The terrain is illuminated from the northwest and is overlain with a 200-m contour interval. The Niwot Ridge Forest site is shown at lower right of image.
Figure 3.2. Energy exchange between snow at ground, in the canopy and the atmosphere.
Figure 3.3. Schematic of wind/eddy diffusion profiles, within and above canopy aerodynamic resistances, and canopy boundary layer resistance.
Figure 3.4. a) Niwot Ridge Ameriflux site: time series of measured above canopy wind, measured and modeled beneath canopy wind, aggregated hourly, and scatterplot of measured and modeled beneath canopy wind, b) TWDEF site: time series of measured above canopy wind, measured and modeled beneath canopy wind inside conifer and deciduous forest, aggregated hourly, and scatterplots of measured and modeled beneath canopy wind.
Figure 3.5. Total above canopy EC-measured and modeled sensible heat flux ($Q_h$), aggregated hourly and plotted with upward fluxes positive: a) time series, b) scatterplot and c) cumulative plot.
Figure 3.6. Time series of total above canopy EC-measured and modeled latent heat flux ($Q_e$), aggregated hourly and plotted with upward fluxes positive. Precipitation and interception plotted downwards following right axis. b) Scatterplot of EC-measured and modeled latent heat flux ($Q_e$), and c) cumulative plot.
Figure 3.7. Same as Figure 3.6, but for the time when the canopy is modeled to hold snow and canopy sublimation is possible in the model.
Figure 3.8. Beneath canopy EC-measured and modeled sensible heat flux ($Q_{hs}$), aggregated hourly and plotted with upward fluxes positive: a) time series, b) scatterplot and c) cumulative plot.
Figure 3.9. Beneath canopy EC-measured and modeled latent heat flux ($Q_{es}$), aggregated hourly and plotted with upward fluxes positive: a) time series, b) scatterplot and c) cumulative plot.
Figure 3.10. Cumulative precipitation, and measured and modeled sublimation from above the canopy, below the canopy and the total.
CHAPTER 4

REPRESENTATION OF CANOPY SNOW INTERCEPTION,
UNLOADING AND MELT IN A PHYSICALLY
BASED SNOW MELT MODEL

Abstract

To better model the accumulation and melt of snow in forested areas we have
developed physically based algorithms that describe the processes of canopy snow
interception, sublimation, mass unloading and melt. These algorithms have been added
to the Utah Energy Balance snowmelt model, UEB, where they have been evaluated in
comparison to observations made in three different vegetation classes (open, deciduous
forest, coniferous forest) at a forest study area in the Rocky Mountains in Utah, USA.
The model was able to represent the accumulation of snow water equivalent in the open
and beneath the deciduous forest quite well but without accounting for redistribution
tended to overestimate the snow water equivalent beneath the conifer forest. Evidence of
redistribution of the intercepted snow from the dense forest (i.e. conifer forest) to the
adjacent area was inferred from observations. Including a simple representation of
redistribution in the model gave satisfactory prediction of snow water equivalent beneath
the coniferous forest. The simulated values of interception, sublimation and unloading
were also compared with previous studies and found in agreement.

4.1. Introduction

Canopy interception has been reported to be a significant component of the
overall water balance in high altitude and latitude forested mountainous region. Up to
sixty percent of total annual snowfall (up to 40 mm of snow water equivalent) can be intercepted by the forest canopy; and associated sublimation losses may exceed 50% of this [Andreadis et al., 2009; Hedstrom and Pomeroy, 1998; Molotch et al., 2007; Storck et al., 2002]. The presence of vegetation partitions snowfall and rainfall into interception by the canopy and throughfall to the ground [Hedstrom and Pomeroy, 1998]. The intercepted snow may sublimate [Lundberg and Halldin, 1994; Lundberg et al., 1998], redistribute to adjacent clearings or open areas [Hoover and Leaf, 1967], unload as mass [Mackay and Barlett, 2006] or melt within the canopy. Sublimation and redistribution reduce the amount of snow available beneath the canopy, and mass unloading and melt from the canopy affects the depth and density of snow beneath the canopy. All these processes affect the snow accumulation or snow water equivalent (SWE) beneath the canopy in heterogeneous watersheds of Western US. Therefore the study of combined canopy snow processes that include snow interception, redistribution, sublimation, mass unloading and melt from the canopy are important in snow modeling, but numerical models that explicitly represent these processes are rare [Andreadis et al., 2009].

Over the last two decades, a number of field based techniques have been developed to estimate precipitation interception and sublimation from a forest canopy. These include tree weighing [e.g. Hedstrom and Pomeroy, 1998; Lundberg, 1993; Pomeroy and Dion, 1996; Satterlund and Haupt, 1967; Schmidt, 1991], gamma ray attenuation [e.g. Bouten et al., 1991; Calder, 1990], strain gauges [e.g. Friesen et al., 2008; Huang et al., 2005], cantilever deflection [e.g. Hancock and Crowther, 1979] and weighing lysimeters [e.g. Storck et al., 2002]. Most snow modeling approaches that estimate canopy snow interception and sublimation have been developed based on these
field studies. A relationship provided by *Hedstrom and Pomeroy* [1998] to estimate canopy snow interception has been used in snowmelt models by *Bartlett et al.* [2006], *Essery et al.* [2003], *Pomeroy et al.* [1998], *Hellstrom* [2000] and *Gelfan et al.* [2004] while relationships provided by *Storck et al.* [2002] to estimate canopy snow interception and unloading have been used in snowmelt model by *Andreadis et al.* [2009].

A number of past studies [e.g. *Andreadis et al.*, 2009; *Ellis et al.*, 2010; *Essery et al.*, 2003; *Gelfan et al.*, 2004; *Hellstrom*, 2000; *Koivusalo*, 2002] have modeled different aspects of the effects of canopy processes on snow accumulations and melts, however we are not aware of work that has combined canopy energy exchanges with the modeling of canopy snow melt and separated canopy snow melt from the canopy mass unloading. *Hedstrom and Pomeroy* [1998] and *Gelfan et al.* [2004] modeled unloading of intercepted snow for boreal forests ignoring snowmelt from the canopy, while *Hellstrom* [2000] and *Koivusalo* [2002] modeled snow melt from the canopy based on the air temperature but neglected solid phase snow unloading from the canopy. *Andreadis et al.* [2009] and *Essery et al.* [2003] estimated and evaluated canopy snow sublimation and melt based on canopy temperature solved using canopy energy balance. They calculated mass release as a function of snowmelt but ignored solid phase snow mass unloading when there was no canopy snowmelt.

In this paper our objective is to develop and evaluate a canopy interception model that parsimoniously represents the canopy interception processes, and partitions the mass unloading and melt water drip from canopy sublimation. In developing this, our goal is to represent these processes using readily available inputs of canopy height and leaf area index. This canopy interception model was added to the Utah Energy Balance (UEB)
111 snowmelt model [Tarboton and Luce, 1996; Tarboton et al., 1995]. UEB also includes representations for other aspects of the snow-vegetation interactions (i.e. canopy radiation transmission and canopy and sub-canopy turbulent flux exchanges). This paper focuses on evaluation of the interception and canopy unloading components of the model by comparing model outputs to measurements of SWE beneath different canopy types. The radiation component of the model enhancements was evaluated by in Chapter 2 and the flux component of the model enhancements was evaluated in Chapter 3.

4.2. Study Site and Data Collection

Field measurements were carried out at the TW Daniels Experimental Forest (TWDEF) located about 30 miles North–East of Logan, Utah (Figure 4.1). TWDEF comprises an area of 0.78 km² at an elevation of approximately 2700 m. It lies at 41.86° North and 111.50° West. The TW Daniels Experimental Forest is on the divide of the watershed that contributes to the Logan River and Bear Lake. Average annual precipitation is about 950 mm of which about 80% is snow. The maximum snow depth can reach 5 m in the area where snow drifts occur. Vegetation is comprised of deciduous forest (Aspen), coniferous forest (Engelmann spruce and subalpine fir), open meadows consisting of a mixture of grasses and forbs, and shrub areas dominated by sagebrush.

Instrumentation was installed starting 2006 to monitor weather and snow within four different vegetation classes: grass, shrubs, coniferous forest, and deciduous forest; and includes twelve weather station towers (three replicates in each vegetation class), one central tower (in shrub area) with more comprehensive radiation instrumentation and one
SNOTEL station in a clearing within the coniferous forest. The following automated data were collected:

- Continuous measurements of snow depth (Judd communications depth sensor) at each of the twelve stations.
- Continuous measurements of weather: temperature and humidity (Vaisala HMP50); wind (Met One, 014A); net radiation, (Kipp & Zonen NR-Lite) at one station in each vegetation class. These instruments were placed at heights above the ground of about 2.5 m in conifer, 4.5 m in deciduous and 4 m in shrub sites so as to remain above the deep snow that accumulates in the deciduous and shrubs areas.
- Four separate radiation components: downward and upward shortwave and long wave (Hukseflux, NR01 4-way radiometer) and snow surface temperature (Apogee Instrument, IRR-PN) at the centralized weather station.
- The standard suite of SNOTEL observations at the adjacent SNOTEL site, from which we used precipitation. This SNOTEL site was installed in summer 2007, so its data are first available for the 2007/8 winter.

Slope and aspect were determined from a 1 m resolution digital elevation model constructed from bare earth points classified from an airborne LiDAR survey of the site. Table 4.1 lists the site information used in the model.

Field observations roughly every two weeks for four winters (2006/7 - 2009/10) comprised two snow pits: one in the shrub area (Pit 1, Figure 4.1) and the other in a conifer clearing (Pit 2, Figure 4.1), and probing snow depth at multiple locations in all four vegetation classes. Within each snow pit samples were taken at 10 cm vertical
intercepts over the entire snow pit depth using a 250 cm³ stainless steel cutter to measure the snow density. The density measured at the pit in the shrub area was used to represent both shrub and grass areas. Both shrub and grass are regarded as open because during the winter snow season snow completely covers the shrubs. Snow density measured in the conifer clearing was used to represent forested areas (both conifer and deciduous). These density values were used with the depth measurements at multiple locations to derive the snow water equivalent (SWE). Temperature was also measured at the surface and at 10 cm vertical intervals over the entire snow pit depth. These temperature measurements were used to derive the energy content of the snow. Numbered snow survey points (Figure 4.1) show locations where the depth measurements were made across the four vegetation classes. This paper used data from the later 3 years of this data collection (2007/8 to 2009/10) when there was SNOTEL data available to have an input precipitation for the model.

4.3. Model Description

The UEB snowmelt model [Tarboton and Luce, 1996] is a physically-based point energy and mass balance model for snow accumulation and melt. Snowpack is characterized using three state variables, namely, snow water equivalent, $W_s$ (m), the internal energy of the snowpack and top layer of soil, $U_s$ (kJ m⁻²), and the dimensionless age of the snow surface used for albedo calculations. The original UEB model is a single layer model. $U_s$ and $W_s$ are predicted at each time step based on the energy balance. Details of the original UEB model formulation are given by Tarboton et al. [1995], Tarboton and Luce [1996] with enhancements for the calculation of surface temperature
using a modified Force-Restore approach given by Luce and Tarboton [2010] and You [2004].

In this paper we present the canopy snow interception component of an enhanced UEB model that includes representation of canopy processes. The canopy component is modeled as a single layer, which added to the original single layer UEB model results in a two-layer model that represents the surface and the canopy snow separately. Energy balances are solved iteratively for each layer to provide outputs of surface temperature, canopy temperature and the other energy fluxes that are based on canopy or surface temperature. The quantity and state of snow in the canopy is represented by a new state variable, canopy snow water equivalent ($W_c$). We assume that the energy content of intercepted snow in the canopy is negligible so canopy temperature, including snow in the canopy, is assumed to adjust to maintain energy equilibrium, except when this requires canopy temperature to be greater than freezing when snow is present in the canopy, in which case the extra energy drives the melting of snow in the canopy.

In the enhanced UEB the changes with time of state variables, $U_s$, $W_s$ and $W_c$, are determined by the following three equations

\[
\frac{dU_s}{dt} = Q_{ss} + Q_{sm} + Q_{ps} + Q_{g} + Q_{ns} + Q_{es} - Q_{ms} \quad (4.1)
\]

\[
\frac{dW_s}{dt} = p_r + p_s - i + R_m + M_c - E_s - M_s \quad (4.2)
\]

\[
\frac{dW_c}{dt} = i - R_m - M_c - E_c \quad (4.3)
\]
In these equations, energy fluxes are: sub-canopy net shortwave radiation, \( Q_{sns} \), sub-canopy net longwave radiation, \( Q_{sln} \), advected heat from precipitation, \( Q_{ps} \), ground heat flux, \( Q_g \), sensible heat flux, \( Q_{hs} \), latent heat flux due to sublimation/condensation, \( Q_{es} \), and advected heat removed by melt water \( Q_{ms} \). Mass fluxes are rainfall, \( p_r \), snowfall, \( p_s \), canopy interception, \( i \), mass release from the canopy, \( R_m \), melt water drip from the canopy snow, \( M_c \), melt from the surface snow, \( M_s \), sublimation from the canopy snow, \( E_c \), and sublimation from the surface snow, \( E_s \). All terms in the energy balance equation are expressed per unit of horizontal area in W m\(^{-2}\). All terms in the mass balance equations are expressed in m hr\(^{-1}\).

Energy content of the intercepted snow is considered negligible. To reflect this, the canopy energy balance is written

\[
Q_{sns} + Q_{sln} + Q_{pc} + Q_{hs} + Q_{es} - Q_{ms} = 0
\]  (4.4)

where \( Q_{sns} \) is canopy net shortwave radiation, \( Q_{sln} \) is canopy net longwave radiation, \( Q_{pc} \) is net advected heat from precipitation to the canopy, \( Q_{hs} \) is sensible heat to the canopy, \( Q_{es} \) is latent heat to the canopy, and \( Q_{ms} \) is advected heat removed by melt water from the canopy.
4.3.1. Canopy Snow Interception in UEB

The focus in this paper is on the parameterizations for \( i, R_m \) and \( M_c \) that describe the canopy interception, mass unloading and melting of intercepted snow in the canopy, respectively.

4.3.1.1. Canopy Precipitation Interception Models Background

There have been a number of studies on precipitation interception in the past where results have frequently been expressed in the form of empirical regression equations between interception loss and precipitation [e.g. Zinke, 1967]. In contrast to regression equations, Rutter et al. [1972], Merriam [1960], Linsley et al. [1949] and many others developed process based models to evaluate the interception loss from precipitation. Rutter et al.’s [1972] model estimates the interception loss based on a running water balance for the wetted canopy. This model uses an empirical relationship to describe drainage from the canopy. Calder [1990] remarks that Rutter's type model is difficult to use as it requires a precise drainage-leaf drip relationship. Merriam [1960] and Linsley et al.’s [1949] interception models relate the interception storage for a storm event to the maximum interception storage capacity as an exponential function of accumulated precipitation. These models assume a closed canopy in which all raindrops contact the canopy and get intercepted. Extending these ideas to an open canopy in which not all raindrops contact the canopy, Aston [1979] wrote

\[
W_c = I_{\text{max}}[1 - \exp(-kP/I_{\text{max}})]
\] (4.5)
where \( W_c (m) \) is the water equivalent interception load in the canopy at the end of an event, \( P (m) \) is the cumulative precipitation for a storm event, \( I_{max} (m) \) is the maximum interception storage capacity and \( k \) is canopy coverage fraction.

Aston (1979) tested a range of models in a *Eucalyptus pauciflora* tree and found better agreement between the open canopy exponential model’s result and measured rainfall interception in comparison to Rutter et al. ’s [1972] or Linsley et al. ’s [1949] model result. Calder [1990] also noted that the performance of Aston’s [1979] model is better than other interception models. Koivusalo [2002] used Aston’s [1979] rainfall interception model to simulate the snowfall interception for conifer forest.

Satterlund and Haupt [1967], Hedstrom and Pomeroy [1998] and Storck and Lettenmaier [1999] performed extensive field measurements to investigate snowfall interception. Satterlund and Haupt [1967] developed and evaluated a growth curve function that relates the interception storage, \( W_c \), to the maximum interception storage capacity, \( I_{max} \), by an exponential function of accumulated storm snowfall, \( P \).

\[
W_c = \frac{I_{max}}{1 + \exp(-S(P - P_o))}
\]  

(4.6)

where \( S \) and \( P_o \) are fitted parameters.

Schmidt and Gluns [1991] and Nakai [1996] simulated the interception of snow on a single branch using this relationship. This relationship requires calibration of parameters \( S \) and \( P_o \) for different snow event and for different tree species. In addition, this relationship gives a non-zero \( W_c \) value even when snowfall (\( P \)) is zero, which is counter intuitive.
Hedstrom and Pomeroy [1998] suggested a snowfall interception model that calculates weekly interception storage based on forest canopy parameters and other meteorological inputs. This model uses the same formulation as the rainfall interception model given by Aston [1979] but with maximum interception storage capacity an order of magnitude higher for snow. This higher maximum interception storage capacity for snow leads to higher snowfall interception in comparison to rainfall interception. In our evaluation Hedstrom and Pomeroy’s [1998] snowfall interception model which is equivalent to Aston’s [1979] rainfall interception model is the most suitable for adding interception simulation capability to UEB. This model is relatively simple and has parameter \( I_{\text{max}} \) that can be empirically related to leaf area, tree species and snow density, while \( k \) is forest canopy cover fraction.

4.3.1.2. Parameterization of the Interception Model for the UEB

The interception parameterization that we have developed is based on the Hedstrom and Pomeroy [1998] interception model adapted from its event based formulation for use in UEB's continuous simulation approach. In the event based formulation, equation (4.5) represents total interception storage, \( W_c \) (m), as a function of total precipitation, \( P \) (m). To convert this to continuous simulation, these total values are interpreted as cumulative values and the interception rate, \( i \) (m h\(^{-1}\)), is calculated as the derivative of cumulative interception storage

\[
i = \frac{dW_c}{dt} = \frac{dW_c}{dP} \frac{dP}{dt} = \frac{dW_c}{dP} P \quad (4.7)
\]
where the last equivalence recognizes precipitation rate, \( p \) (m h\(^{-1}\)), as the derivative of cumulative precipitation. \( dW_c/dP \) quantifies interception efficiency and is evaluated from equation (4.5) recognizing that \( k \) and \( I_{\text{max}} \) are constants as

\[
\frac{dW_c}{dP} = k \exp(-kP/I_{\text{max}}) \tag{4.8}
\]

Inverting equation (4.5) gives

\[
P = \frac{-I_{\text{max}}}{k} \ln \left(1 - \frac{W_c}{I_{\text{max}}} \right) \tag{4.9}
\]

Substituting this in equation (4.8) and (4.7) gives

\[
i = k \left(1 - \frac{W_c}{I_{\text{max}}} \right) p \tag{4.10}
\]

This is the continuous equivalent to (4.5) and shows a linearly decreasing interception efficiency as a function of interception storage, \( W_c \). This relationship is exactly same as the snow storage function developed from measured attenuation of gamma radiation by snow on canopies of sitka spruce given by Calder (1990) for a closed canopy, \( k =1 \), but generalized for an open canopy. Equation (4.10) is used in equation (4.3) to calculate \( i \) as a fraction of total precipitation from both rain and snow \((p=p_r+p_s)\).

To estimate the interception using equation (4.10), \( I_{\text{max}} \) is calculated as

\[
I_{\text{max}} = I_b L \tag{4.11}
\]

where \( I_b \) (kg m\(^{-2}\)) is the per unit of leaf area interception capacity and \( L \) is leaf area index. Schmidt and Gluns [1991] made extensive field measurements and reported reference \( I_b \) values of 6.6 and 5.9 kg m\(^{-2}\) for pine and spruce trees, respectively. The reference \( I_b \) is adjusted for snow density to give \( I_b \) as
where, \( \rho_s \) is the fresh snow density (kg m\(^{-3}\)) estimated based on air temperature, \( T_a \) (°C) [e.g. Hedstrom and Pomeroy, 1998; Schmidt and Gluns, 1991; U.S. Army Corps of Engineers, 1956] as
\[
\rho_s = 67.92 + 51.25 \exp \left( \frac{T_a}{2.59} \right) \tag{4.13}
\]

4.3.1.3. Canopy Mass Unloading \((R_m)\)

Intercepted snow remains on the canopy until it is removed by means of sublimation, mass unloading or melt. Sublimation is a loss to the atmosphere while mass unloading and melt from the canopy increase the water equivalent of snow under the canopy.

Unloading from canopy may occur as a result of failure in the strength of a leaf or branch (it bends and unloads its snow) or the strength of intercepted snow (it collapses and falls off the branch/leaf). Wind, temperature, radiation, snow density, and wetness affect unloading. Individual effects of these to unloading are difficult to quantify. 

Hedstrom and Pomeroy [1998] used an empirical relationship to estimate unloading from the canopy, ignoring sublimation and melt from the interception, in a cold boreal forest. We used their approach to estimate the mass unloading from the canopy in our temperate semi-arid climate.

Hedstrom and Pomeroy’s [1998] relationship provides an exponential decay of intercepted snow over a time. If \( W_c \) is the intercepted snow load at the start of unloading, the intercepted snow load, \( W_{cT} \) after time \( T \) is given by

\[
W_{cT} = W_c \left( 0.27 + \frac{46}{\rho_s} \right) \tag{4.12}
\]
\[ W_{ct} = W_c \exp(-U_m T) \quad (4.14) \]

where \( U_m \) is mass unloading rate coefficient with the unit time\(^{-1}\).

Equation (4.14) requires estimation of \( U_m \) and \( T \) over which \( W_c \) is calculated from equation (4.3) to estimate \( W_{ct} \). Instead of separate values of \( U_m \) and \( T \), Hedstrom and Pomeroy [1998] estimated the value of a dimensionless unloading coefficient

\[ c = \exp(-U_m T) = 0.678 \]

from measurements of interception for time after snowfall between zero and seven days. Assuming an average time after snowfall of 3.5 days we evaluated \( U_m \) (hr\(^{-1}\)) putting \( T = 3.5 \times 24 \) hours as

\[ U_m = -\frac{1}{T} \ln 0.678 \]

\[ = 0.00463 \text{ hr}^{-1} \quad (4.15) \]

Using this coefficient, mass unloading (m hr\(^{-1}\)) from the intercepted snow is calculated as

\[ R_m = U_m W_c \quad (4.16) \]

### 4.3.1.4. Canopy sublimation \((E_c)\) and Melt \((M_c)\)

Sublimation loss from the intercepted snow is calculated based on the latent heat flux exchange between the canopy snow and surrounding air as

\[ E_c = -\frac{Q_{ec}}{\rho_w h_v} \quad (4.17) \]

where \( Q_{ec} \) (W m\(^{-2}\)) is latent heat flux exchange between canopy snow and the atmosphere, \( h_v \) the latent heat of sublimation (2834 kJ kg\(^{-1}\)), and \( \rho_w \) is density of water (kg m\(^{-3}\)).
Melt water drip from the canopy is quantified based on canopy energy balance, using energy fluxes (radiation, sensible and latent heat) (Equation 4.4). In this equation \( Q_{mc} \) is first set to 0 and the other terms used to solve for canopy temperature, \( T_c \). If the result is above freezing (0 °C) and there is snow present in the canopy, then \( T_c \) is set to freezing and equation (4.4) is solved for \( Q_{mc} \). The resulting \( Q_{mc} \) is used to determine \( M_c \) as

\[
M_c = \frac{Q_{mc}}{\rho_w h_f}
\]

(4.18)

where \( \rho_w \) (kg m\(^{-3}\)) is density of water and \( h_f \) is the heat of fusion (333.5 kJ kg\(^{-1}\)).

4.3.2. Above and Below Canopy Turbulent Energy Fluxes

Turbulent fluxes were modeled using the concept of flux proportional to temperature and vapor pressure gradients and inversely proportional to resistances. With above canopy inputs of air temperature \( T_a \), vapor pressure \( e_a \) and other meteorological variables the model solves two energy balance equations (one at the canopy and the other at the surface) to calculate the canopy temperature, \( T_c \) and snow surface temperature, \( T_s \). Saturated vapor pressure at the snow surface, \( e_s(T_s) \), and canopy, \( e_s(T_c) \), are calculated as functions of respective snow surface temperatures. Aerodynamic and boundary layer resistances are estimated using a wind profile within the canopy based on leaf area index following Choudhury and Monteith [1988] and Dolman [1993]. These resistances are used with the temperature and vapor pressure differences between the snow surface, canopy air and the snow in the canopy to calculate the turbulent flux exchanges between snow on the ground, canopy and the atmosphere. A detailed description of how turbulent fluxes are calculated is given in Chapter 3.
4.3.3. Radiation

Radiation is the main energy source that drives the snow energy balance and snowmelt. To estimate the net radiation beneath, in, and above the canopy, we use an approach that describes the penetration of radiation through the canopy based on a two stream approximation accounting for multiple scattering. In the two stream approximation, upward and downward radiation is expressed using two differential equations quantifying the change in downward and upward radiation due to interception of radiation, absorption and scattering. This method assumes, as an approximation following Monteith and Unsworth [1990], that multiple scattering occurs along a single path, thereby avoiding the intractable complexity of scattering in multiple directions. It also considers multiple reflections between the canopy and surface treating direct and diffuse radiation separately. A detailed description of how net radiation is calculated beneath, in, and above the canopy is given in Chapter 2.

4.3.4. Implementation

The implementation of the interception term $i$ in the overall mass balance equations 4.2 and 4.3 uses equations 4.10 to 4.12. $\overline{T_b}$ is an input model parameter with default value of 6.6 kg m$^{-2}$. Leaf area index $L$ is a site variable that depends on the vegetation. The implementation mass unloading uses equation (4.16) with $U_m$, a model parameter with default value of 0.00463 hr$^{-1}$. The calculations of canopy sublimation and melt do not require additional parameters as they are obtained from the solution of the canopy energy balance described previously (Chapters 2 and 3).
4.4. Results and Discussion

4.4.1. Observations

Snow water equivalent measured in the open, deciduous and conifer forest for our study years is shown in Figure 4.2. These show generally smaller accumulations beneath the forest canopies, in comparison to the open areas, a difference that is attributed to interception and subsequent sublimation and redistribution of intercepted snow by wind, much of it into surrounding open areas. This comparison with open areas is complicated by the considerable variability in open area SWE that is believed to be primarily due to drifting and scour. On the other hand the observations in the forest canopies are much less variable. Peak accumulation averaged at towers SB and SC, and at survey points 1, 2, 13, 19, 20 sites (chosen because we believe these sites are less affected by wind drifting and scouring) in the grass/shrubs (i.e. open) areas was about 10 to 20% higher than the peak accumulation averaged over different locations in the conifer forest areas. There was not much difference in observed accumulation between deciduous forest and open areas. Even though accumulations in forest areas were less, they tended to ablate more slowly and persist longer than the snow in the open areas. More rapid ablation was observed in deciduous forest in comparison to coniferous forest.

4.4.2. Model Application

The model was initialized at the beginning of the water year each year from 2007/8 to 2009/10 and run for the whole water year for the open, deciduous and coniferous forest areas using hourly meteorological inputs of precipitation, temperature, wind speed and relative humidity measured at the open site, shrubs B (SB). Input
precipitation data were taken from the SNOTEL site located in a small opening in the conifer forest. For forested areas, the open site meteorological variables were assumed representative of conditions at a height of 2 m above the forest canopy. The vegetation parameters used in the model are leaf area index, canopy height and canopy cover, interception unloading rate and reference unit interception capacity. The thermal conductivity of snow and soil, and surface roughness length beneath the canopy were obtained from calibration in Chapters 2 and 3. Other model parameters used in this work follow the original UEB model. All parameters are listed in Table 4.2.

The model is able to predict the SWE, snow surface temperature, snow average temperature, canopy wind speed, radiation, energy fluxes and interception for both open and forested areas. In this work we focus on the model’s representation of the canopy processes, i.e. snow interception, sublimation, mass unloading and melt and prediction of SWE beneath the canopies. The interception and redistribution of snow that affects the SWE beneath the canopy is also discussed though this process is not explicitly represented in the model. Canopy interception processes were evaluated by comparing the model simulation of SWE beneath the deciduous and forest canopies with observations, as direct measurements of interception, sublimation, mass unloading and melt were not available at our site to compare to the model results. Model simulations of these were compared with the results of previous studies performed in the similar environments.
4.4.3. Simulation Results

4.4.3.1. Snow Interception, Sublimation, Mass Unloading and Melt

Snow interception, sublimation, mass unloading and melt from the canopy simulated by the model are reported in Figures 4.3 and 4.4. Figure 4.3 shows the cumulative water equivalent of snow interception, mass unloading and melt from interception while Figure 4.4 shows the cumulative sublimation from the snow in the canopies and at the surface below the canopies, and also the cumulative sublimation from the open areas. As expected interception, sublimation and unloading were higher for denser canopy i.e. for conifer canopy in comparison to less dense, deciduous canopy. Canopy intercepted snow was reduced mainly by melt water release during the early and latter half of the simulation period, and mainly by sublimation and mass unloading during the middle of the simulation period.

The average precipitation observed during 2007/8, 2008/9 and 2009/10 simulation periods was 950 mm. For these periods, the model simulated that about 50% of the precipitation was intercepted in the coniferous forest, out of which 17% sublimated, 20% unloaded as mass and 63% melted from the canopy. In the deciduous forest about 27% of precipitation was simulated as interception, out of which 19% sublimated, 13% unloaded as mass and 68% melted. The model results showed that about 8% of the precipitation that falls (75 mm out of 950 mm) was sublimated from the snow intercepted in the coniferous canopy while only about 5% of the precipitation that falls (50 mm out of 950 mm) was sublimated from the intercepted snow. Below canopy sublimation simulated by the model was about 1% of precipitation for coniferous forest and about 4%
of precipitation for deciduous forest. About 7% of the precipitation that falls (69 mm out of 950 mm) was sublimated from the open surface.

The model simulated conifer canopy snow sublimation here is the same as the canopy snow sublimation value (8% of the precipitation) reported by Essery et al. [2003] for douglas fir at Oregon and for boreal forest of central Canadian in their land surface model and slightly lower than the value (about 10% or 100mm) reported by Storck et al. [2002] from field observation at Pacific North West forests in Oregon. Pomeroy et al. [1998] and Bartlett et al. [2006] reported up to about 30 to 40 mm sublimation loss from the canopy in the boreal forest. However, while the total sublimation they report was less the fractional sublimation they reported was greater, about 38 to 45%. The model simulated below canopy sublimation here is also similar to the below canopy sublimation value reported by Bartlett et al. [2006] for different forest sites in central Saskatchewan.

Our model showed that of intercepted snow reaching the surface, 20 to 30% was solid mass unloading, while 70 to 80% was melt. These values are similar to the values, i.e., 28% solid mass and 72% melt water release reported by Storck et al. [2002].

The model simulated slightly greater total sublimation (canopy and surface) from the forest in middle of the simulation period in comparison to open areas. This seems to be responsible for the lower accumulations in the forest. However during the latter part of the simulations there was little intercepted snow in the canopy (and hence little canopy sublimation) suppressing sublimation from the forested area. These effects offset each other so that the net loss to sublimation was, in this specific simulation, roughly equivalent between open and forest areas.
4.4.3.2. SWE Beneath the Canopy

The model’s predictions of SWE beneath the deciduous forest canopy compared well with the observations for all three years that were simulated (Figure 4.5). However the model slightly over predicted the SWE beneath the conifer forest canopy (Figure 4.6). One possible explanation for this overestimation of below canopy SWE is that some snow mass unloading from the canopy is redistributed (blown) by wind into open areas, or underestimation of interception and sublimation from the canopy. Another possible explanation is that the overestimation of below coniferous canopy SWE is due to overestimation of snow mass unloading and accompanying underestimation of interception and sublimation. Parameters could be adjusted to alter this partitioning between loss to sublimation and unloading, but this would offset the agreement between measured and modeled sublimation discussed in Chapter 3, albeit for a different site. Further sublimation losses as calculated here agree reasonably with results of previous studies performed in similar environments, hence we take the model’s predictions of interception and sublimation loss, to be reasonable and infer that the most likely cause of the slight underestimates is some degree of interception redistribution.

Interception redistribution does not affect the net accumulation at watershed scale [Gary, 1974; Hoover and Leaf, 1967], but in a micro scale study it may result in a significant amount of snow loss from forested areas. The redistributed snow is likely to be received by the open areas. In our study, evidence of redistribution of the intercepted snow from the surrounding canopy to the opening where the SNOTEL pillow was located is that SNOTEL pillow data showed about 50 to 100 mm of snow accumulation while there was almost no or little accumulation in other locations during early winter.
(November-December) each year (see Figure 4.5 and 4.6). Also, there was an offset of about 20 to 50 mm between recorded precipitation and SNOTEL pillow measurement during November-December each year and that offset decreased towards the time of SWE peak, indicating that during the main buildup of snowpack the pillow is receiving more snow than the precipitation that is being measured. This additional snow is likely from interception redistribution.

The model does not include a component that calculates intercepted snow redistribution, since there is no established equation to represent this. To, in a simple way attempt to simulate this redistribution, we reduced the beneath canopy precipitation input \( (p_r + p_c) \) by 10% (Figure 4.7). This results in a better match between modeled and measured beneath conifer canopy SWE, supporting this redistribution explanation. The deciduous forest model and measured comparisons are good as they are suggesting that redistribution is likely an insignificant factor in deciduous forest.

## 4.5. Conclusions

Model comparisons were made to snow surveys and field measurements carried out in deciduous and coniferous forest at the TW Daniels Experimental Forest during the three winters from 2007/8 to 2009/10. Forest canopy effects on snow processes and sensitivity of snow accumulation and melt to forest type and canopy density (leaf area) were assessed by adding a vegetation component that includes the processes of snow interception, sublimation, mass unloading and melt from the canopy to an energy balance model.

The parameterization of interception processes in the model resulted in lower accumulation and slower melt rate in forest areas than in open areas consistent with
observations, and reproduced the observed SWE in open and beneath the deciduous canopy. However the model over predicted the SWE beneath the conifer canopy. The observed lower accumulation inside the conifer forest was inferred to be mainly due to the redistribution of intercepted snow. Including this in the model through a simple adjustment to precipitation reaching the surface resulted in a satisfactory prediction of SWE beneath the coniferous forest.

The addition of interception processes to the energy balance model has provided improved and more explicit representations of snow processes governing the accumulation and melt in forested areas. Further evaluation is required in different settings. Extension of the model over larger areas to evaluate the watershed scale impacts of vegetation distribution should also be pursued.

References


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### Table 4.1. Site variables

<table>
<thead>
<tr>
<th>Sites/Variables</th>
<th>Open</th>
<th>Deciduous</th>
<th>Conifer</th>
</tr>
</thead>
<tbody>
<tr>
<td>Leaf area index</td>
<td>0.0</td>
<td>1</td>
<td>4.5</td>
</tr>
<tr>
<td>Canopy cover fraction</td>
<td>0.0</td>
<td>0.7</td>
<td>0.7</td>
</tr>
<tr>
<td>Canopy height (m)</td>
<td>0.0</td>
<td>10.0</td>
<td>10.0</td>
</tr>
<tr>
<td>Slope (degrees)</td>
<td>3.6</td>
<td>5.0</td>
<td>2.0</td>
</tr>
<tr>
<td>Aspect (degrees clockwise from N)</td>
<td>150</td>
<td>0.0</td>
<td>300</td>
</tr>
<tr>
<td>Latitude (degrees)</td>
<td>41.86</td>
<td>41.86</td>
<td>41.86</td>
</tr>
<tr>
<td>Average atmospheric pressure (Pa)</td>
<td>74000</td>
<td>74000</td>
<td>74000</td>
</tr>
</tbody>
</table>
### Table 4.2. Model parameters

<table>
<thead>
<tr>
<th>Name</th>
<th>Values</th>
<th>Basis</th>
</tr>
</thead>
<tbody>
<tr>
<td>Air temperature above which precipitation is all rain ((T_r))</td>
<td>3 °C</td>
<td>Tarboton et al. [1995], U.S. Army Corps of Engineers [1956]</td>
</tr>
<tr>
<td>Air temperature below which precipitation is all snow ((T_{sn}))</td>
<td>-1 °C</td>
<td>Tarboton et al. [1995], U.S. Army Corps of Engineers [1956]</td>
</tr>
<tr>
<td>Emissivity of snow ((\varepsilon))</td>
<td>0.98</td>
<td>Tarboton et al. [1995]</td>
</tr>
<tr>
<td>Ground heat capacity ((C_g))</td>
<td>2.09 kJ kg(^{-1})C(^{-1})</td>
<td>Tarboton et al. [1995]</td>
</tr>
<tr>
<td>Nominal measurement of height for air temperature and humidity ((z_{ml}))</td>
<td>2.0 m</td>
<td>Tarboton et al. [1995]</td>
</tr>
<tr>
<td>Surface aerodynamic roughness ((z_{os}))</td>
<td>0.1 m</td>
<td>Chapter 3</td>
</tr>
<tr>
<td>Soil density ((\rho_g))</td>
<td>1700 kg m(^{-3})</td>
<td>Tarboton et al. [1995]</td>
</tr>
<tr>
<td>Liquid holding capacity of snow ((L_e))</td>
<td>0.05</td>
<td>Tarboton et al. [1995]</td>
</tr>
<tr>
<td>Snow saturated hydraulic conductivity ((K_e))</td>
<td>20 m hr(^{-1})</td>
<td>Tarboton et al. [1995]</td>
</tr>
<tr>
<td>Visual new snow albedo ((\alpha_{vo}))</td>
<td>0.85</td>
<td>Tarboton et al. [1995]</td>
</tr>
<tr>
<td>Near infrared new snow albedo ((\alpha_{vo}))</td>
<td>0.65</td>
<td>Tarboton et al. [1995]</td>
</tr>
<tr>
<td>Bare ground albedo ((\alpha_{bg}))</td>
<td>0.25</td>
<td>Tarboton et al. [1995]</td>
</tr>
<tr>
<td>Thermally active depth of soil ((d))</td>
<td>0.1 m</td>
<td>You [2004]</td>
</tr>
<tr>
<td>Thermal conductivity of snow ((\lambda))</td>
<td>3.6 Wm(^{-1})K(^{-1})</td>
<td>Chapter 2</td>
</tr>
<tr>
<td>Thermal conductivity of soil ((\lambda_g))</td>
<td>14.4 Wm(^{-1})K(^{-1})</td>
<td>Chapter 2</td>
</tr>
<tr>
<td>Atmospheric transmittivity for cloudy conditions ((\alpha_a))</td>
<td>0.25</td>
<td>Shuttleworth [1993]</td>
</tr>
<tr>
<td>Atmospheric transmittivity for clear conditions ((\alpha_a + \beta_j))</td>
<td>0.75</td>
<td>Shuttleworth [1993]</td>
</tr>
<tr>
<td>Ratio of direct to total radiation for clear sky ((\lambda))</td>
<td>6/7</td>
<td>Chapter 2</td>
</tr>
<tr>
<td>Richardson number upper bound for stability correction ((Ri_{max}))</td>
<td>0.16</td>
<td>Koivusalo [2002]</td>
</tr>
<tr>
<td>Emissivity of canopy (conifer/deciduous) ((\varepsilon))</td>
<td>0.98</td>
<td>Bonan [1991]</td>
</tr>
<tr>
<td>Interception unloading rate ((U_m))</td>
<td>0.00346 hr(^{-1})</td>
<td>Hedstrom and Pomeroy [1998]</td>
</tr>
<tr>
<td>Reference unit interception capacity ((I_b)) (kg m(^{-2}))</td>
<td>6.6 kg m(^{-2})</td>
<td>Hedstrom and Pomeroy [1998]</td>
</tr>
</tbody>
</table>
Figure 4.1. Site map of the TW Daniels Experimental Forest showing weather station towers, vegetation, survey points, pits and SNOTEL site.
Figure 4.2. Snow accumulation and melt observed at different locations in shrubs, grass, beneath coniferous forest and deciduous forest during three winters.
Figure 4.3. Model simulated cumulative snow interception, mass unloading and melt from the conifer and deciduous forest canopy.
Figure 4.4. Model simulated cumulative sublimation from the snow in canopy, from the surface snow below the canopy and the total sublimation from the deciduous and coniferous forest; and cumulative sublimation from the snowpack from open areas.
Figure 4.5. Measured SWE values at different locations inside the deciduous forest, modeled SWE values, SWE measured by snow pillow at SNOTEL site and measured cumulative precipitation. Dots are SWE values derived from manual depth measurements and continuous lines are SWE values derived from depth measured by snow depth sensors.
Figure 4.6. Measured SWE values at different locations inside the conifer forest, modeled SWE values, SWE measured by snow pillow at SNOTEL site and measured cumulative precipitation. Dots are SWE values derived from manual depth measurements and continuous lines are SWE values derived from depth measured by snow depth sensors.
Figure 4.7. Modeled conifer forest SE with surface precipitation inputs \((p_r + p_s)\) reduced by 10% to approximate the redistribution of intercepted snow, compared to measured SWE values at different locations inside the conifer forest, SWE measured by snow pillow at SNOTEL site and measured cumulative precipitation. Dots are SWE values derived from manual depth measurements and continuous lines are SWE values derived from depth measured by snow depth sensors.
This dissertation has investigated the forest canopy impacts on snow processes examining the variability of snow accumulation and differences in the timing of melt and sublimation between open (grass/shrubs) and forest (conifer/deciduous) locations at Rocky mountain study sites in the Western US using field observations and developing and evaluating a physically based snowmelt model. Chapters 2 through 4 present the main scientific results of this dissertation. In this chapter I summarize the important conclusions from these chapters and provide suggestions regarding the future directions and opportunities for research in this field.

5.1. Summary and Conclusions

The objective of this research was to understand how vegetation and snow interact and how to better model the snow accumulation and melt over heterogeneous watershed to improve runoff and water supply prediction in mountainous regions of the Western US. To understand the snow-vegetation interactions, field measurements roughly every two weeks for four winters (2006/7 - 2009/10) were made at the TW Daniels Experimental Forest (TWDEF) located about 30 miles North–East of Logan, Utah. These observations consisted of digging two snow pits: one in the shrub area and the other in a conifer clearing, and probing snow depth at multiple locations in all four vegetation classes. Snow density and temperature were measured at 10 cm vertical intervals over the entire snow pit depth. The density values measured at pits were used with the depth measurements at multiple locations to derive the snow water equivalent
(SWE). To improve snowmelt modeling capability in heterogeneous watersheds this
dissertation has added and evaluated three important snow canopy processes, i.e. i) canopy radiation transmission, ii) above and below canopy turbulent fluxes, and iii) snow interception, sublimation, mass unloading, and melt, to the UEB snowmelt model [Tarboton and Luce, 1996; Tarboton et al., 1995; You, 2004]. UEB, prior to this work, was a single layer energy and mass balance snowmelt model with limited ability to physically simulate the effects of vegetation on snow accumulation and melt.

By incorporating these three important canopy processes in a snowmelt model this dissertation developed a complete physically based model that more comprehensively describes the processes of snow accumulation and melting for open and forested areas. I used the model to investigate the effects of forest canopy on snow accumulation and melt processes. The resulting enhanced model represents the system parsimoniously using a single canopy and a single bulk snow layer, and can be driven by modest canopy (leaf area index, canopy height, and canopy cover) and meteorological data (precipitation, air temperature, relative humidity and wind speed).

The enhanced model was mainly evaluated by comparing the model results with observations made at the TW Daniels Experimental Forest in Northern Utah. Data from Niwot Ridge AmeriFlux site, Colorado were used to evaluate flux parts of the model.

Chapter 2 of this dissertation described the enhancement to the UEB model to better represent the penetration of radiation through a forest canopy. A two stream canopy radiation transmission model that explicitly accounts for radiation scattering, absorption and reflection by leaves was added to the model. This model is an advance over Beer’s law which does not account for multiple scattering of radiation, while still
using only a single leaf area index as a key parameter to provide radiation beneath the
canopy. Using the two stream approach we found a significant increase in transmittance
of radiation through the canopy over the Beer's law attenuation.

This work first compared the model simulated four radiation components
(incoming and outgoing, longwave and shortwave) with observations made at an open
site. These comparisons helped to parameterize atmospheric transmittivity, cloud cover,
snow surface albedo and, surface emissivity used by the model to estimate longwave or
shortwave radiation. The radiation transmission model was then tested against the three
years of winter data 2007/8, 2008/9 and 2009/10, measured in the open, beneath the
conifer and deciduous forest canopies. Less net radiation and slower ablation as forest
density increases (open to deciduous to conifer) were evident in the observations and
captured by the model. The model results for open, beneath the deciduous and conifer
forest suggest that reasonable predictions of snowmelt from a heterogeneous watershed,
and the sensitivity of snowmelt to forest type and canopy density (leaf area) can be
obtained with the parsimonious (one layer two stream) representation of canopy radiation
transmission given here.

Marks et al. [2008] state that modeled turbulent fluxes over the snow cover in
forested environments are difficult to evaluate since very few measurements exist for
model comparisons and evaluation. In Chapter 3, we evaluated modeled turbulent fluxes
over the snow cover beneath the canopy, and also over the snow in the canopy by
comparing model results with Eddy Covariance (EC) measurements. Turbulent fluxes
were modeled using the concept of flux proportional to temperature and vapor pressure
gradient and inversely proportional to resistances. Required temperatures were obtained
solving the energy balances for the surface and canopy, and resistances were calculated as a function of wind and leaf area index.

Above canopy sensible heat flux dominated the other above and below canopy turbulent fluxes throughout the simulation period. Latent heat flux transfers and subsequent snow mass losses from the canopy snow were almost double of that from beneath the canopy. Both model and EC measurement showed this. Some discrepancies between modeled and EC-measured below canopy sensible heat fluxes were observed. The below canopy eddy-covariance flux measurements could be problematic because the processes and practical aspects of converting raw instrumental data streams into high quality fluxes may not be valid beneath the canopy where low wind speed, strong heterogeneity and intermittent turbulence prevail [Baldocchi et al., 2000; Blanken et al., 1998; Constantin et al., 1999]. The diffusion theory used by the model to estimate canopy turbulences may also have limited validity within the canopy space. Because of these uncertainties associated with both the measurements and the modeling this research could not resolve all the differences observed between beneath canopy modeled and EC-measured sensible heat fluxes.

Despite some uncertainties, the model evaluation results of turbulence and vapor flux (above and below the canopy) presented in this work were found encouraging. With the inclusion of the above and below canopy turbulent fluxes in a snowmelt model, snowmelt modeling capability in the heterogeneous watershed could be improved.

Snow interception is an important hydrological process, producing complex mass and energy exchanges with the surrounding atmosphere and the snowpack below [Strasser et al., 2011]. To improve our understanding of snow-canopy interception
processes and the associated influences on the snowpack below, we developed and evaluated a canopy model that describes the canopy processes, i.e. processes of canopy snow interception, sublimation, mass unloading and melt within UEB in Chapter 4.

Based on Hedstrom and Pomeroy’s [1998] event based snow interception model, we added a continuous snow interception model to the UEB. The approach used here accounts for forest canopy cover, leaf area and tree species, and snow density while calculating interception storage. We distinguished the snow in the canopy from the snow at the surface, and partitioned the canopy mass unloading and melt from the interception to estimate the net snow mass loss from the canopy.

The model reasonably reproduced the SWE values that declined with increasing leaf area index because of the snow interception and sublimation in the canopy. The model performance further improved when redistribution of intercepted snow was approximated through adjustments to the below canopy precipitation input.

In the evaluation of overall model results in the prediction of surface water input and runoff from snowmelt in heterogeneous watershed, we found greater total sublimation (canopy and surface) from the forested areas in the first half and middle of the simulation period. This seems to be responsible for the lower accumulations in the forest. However during the latter part of the simulations there was little intercepted snow in the canopy (and hence little canopy sublimation) suppressing sublimation from the forested area. This was part of an overall reduction in energy fluxes beneath the forest. These effects offset each other so that the net loss to sublimation was, in this specific simulation, roughly equivalent between open and forested areas. Hence the surface water input was also roughly equivalent. However the more rapid surface water input from
open areas provides greater opportunity for surface runoff and occurs earlier, while the slower surface water input from forested areas persists later into the season, sustaining streamflow longer, but is also more subject to infiltration into the soil and uptake and transpiration by vegetation which may result in reduced stream flow.

5.2. Recommendations

Generally, this dissertation has successfully presented and evaluated some new approaches that advance the understanding of snow-vegetation interactions. However, there are several significant ways to contribute to the better assessments of the model results. The goal of designing the research described in this dissertation is to develop a physically based model so that the model is transportable and applicable without calibration at different locations. The model was successfully validated for one location and some aspects of it tested at one other location, but the generality and transferability of this model to further other locations remains to be tested. As a result, the foremost task among many is to test the model in other regions.

Chapter 2 introduced a new model for canopy radiation transfer for an energy balance snowmelt model. The new model has been validated against the net radiation measurements available at single location treating the forest as a horizontally homogeneous vegetation canopy cover. Since forest structure is highly heterogeneous, it causes variation of radiation below the canopies. Measurements at more than one location and model validation against these measurements are recommended in the future to research ways to mitigate the impacts of forest heterogeneity on below canopy radiation. Also, our model does not address transition effects such as solar radiation penetration to snow beneath a forest canopy near an opening, or shading of open areas by
nearby forests. Further study to understand and quantify the impacts and importance of transitions on snow accumulation and melt is warranted in the future.

This chapter compared the model simulated below canopy net radiation with observed below canopy net radiation. It would have been better to have direct measurements of downward and upward shortwave and longwave radiation (four radiation components) beneath the canopy, to compare to. Observations of these are recommended in the future. The below canopy incoming shortwave and longwave radiation comparisons would evaluate the methods used by the model to simulate the shortwave and longwave radiation transmitted through the canopy, respectively. The below canopy outgoing shortwave and longwave radiation comparisons would serve to check the model albedo, surface temperature and emissivity representations beneath the canopy. The effects of forest litter on the beneath canopy snow albedo could also be modeled with the help of the outgoing shortwave radiation measurements.

Chapter 3 exposed biases in the model prediction of beneath canopy sensible heat flux. Modeled and EC-measured beneath canopy sensible heat fluxes were in opposite directions: model, showing downward and EC measurements, showing upward during the snowmelt period. There is an opportunity for future research to investigate why the EC measurements showed upward sensible heat flux during the snowmelt period when the temperature gradient is downward? The UEB model results can also be compared with those of other snow models under the same or similar environment conditions to evaluate the methods used by the model to simulate the beneath canopy sensible heat flux.

Radiation energy (incoming and outgoing solar and longwave radiation) and turbulent flux energy (sensible heat and latent heat) are the two main components of
energy balance that drive snowmelt and sublimation in an energy balance model. This work validates the radiation component at one location and flux component at other location. Validation of both of these components at a single location would better serve to check the model parameterizations and contributes to the better assessments of the model results. We recommend testing the model at location where all energy flux measurements are available to compare the model results.

Chapter 4 brought out the importance of snow interception, sublimation, mass unloading and melt in the snow mass and energy balance. Very few works have combined these processes in snow modeling [Andreadis et al., 2009]. In this chapter, the model obtained the interception and sublimation loss for the prediction of SWE beneath the canopy that we compared to observations for model evaluation. This comparison could be insufficient for the assessments of the methods used by the model to calculate interception, sublimation, unloading and melt processes. Direct evaluations of the interception and the sublimation are recommended in the future.

Canopy snow sublimation is not the only mechanism for lowering the SWE beneath the canopies. We found evidence that a fraction of the snow intercepted by the canopy is redistributed by the blowing wind. Modeling and evaluating the snow redistribution from the canopy should be included in the snow-vegetation interactions processes. We recommend applying snow redistribution in the snow modeling in the future.

The three papers presented in this dissertation extend an open area energy balance snowmelt model to include a physically based representation of canopy processes that uses leaf area index as a key parameter to predict energy balance and snowmelt beneath
the canopy. In our work we adjusted the leaf area index. The adjustment was done keeping in mind that the adjusted value falls within the range of leaf area index values found in the literature. Leaf area index can be measured directly or indirectly. Direct method involves measuring the area of the leaves within a delimited area or sampling of litter below the canopy during leaf fall while indirect method uses statistical or probabilistic approach [Bréda, 2003]. Though the leaf area index values we used are close to the values we found in the literatures, we recommend measurement of leaf area index and modeling of the snow processes using the measured values in the future.

Finally, climate change is weakening the statistical basis for river forecasts that use empirical relationships, making it important to understand and be able to model snowmelt from a physical basis. The addition of physical processes of snow-vegetation interactions in the energy balance model offered improved and more explicit representations of snow processes governing the accumulation and melt in forested areas using a parsimonious approach that can be used with practically available information. There are already many detailed models that have vegetation component and can be used to study snow processes in forested environments [e.g. Bartlett et al., 2006; Liang et al., 1997; Niu and Yang, 2004; Verseghy, 1991; Verseghy et al., 1993; Wigmosta et al., 1994]. We worked with the UEB because more detailed models require parameters that are difficult or infeasible to obtain in many practical settings. However, comparisons of results from this simpler model with the results of other detailed models may help to assess the effectiveness of the model and is recommended in the future.

Future study could be planned for the extension of the model over larger areas to evaluate the combined effects of vegetation distribution (i.e., impacts of altered radiation,
turbulent fluxes and sublimation losses, and precipitation and through fall due to presence of vegetation in the watershed) on snow accumulation and melt processes. Impacts of land use or vegetation changes due to forest disturbances, fire or beetle kill on hydrological regimes and water resources management could also be pursued.

References


Marks, D., M. Reba, J. Pomeroy, T. Link, A. Winstral, G. Flerchinger, and K. Elder (2008), Comparing simulated and measured sensible and latent heat fluxes over
snow under a pine canopy to improve an energy balance snowmelt model, *J. Hydrometeorol.*, 9(6), 1505-1522, [http://dx.doi.org/10.1175/2008JHM874.1](http://dx.doi.org/10.1175/2008JHM874.1).


APPENDICES
Appendix A

Solution to equations 2.8 and 2.9 in Chapter 2
Equations 2.8 and 2.9 in Chapter 2 are

\[-dU = -UK_h \rho dy + UK_h \rho \frac{\alpha}{2} dy + QK_h \rho \frac{\alpha}{2} dy \quad (A1)\]

\[dQ = -QK_h \rho dy + UK_h \rho \frac{\alpha}{2} dy + QK_h \rho \frac{\alpha}{2} dy \quad (A2)\]

Subtracting equation (A1) from (A2) and dividing by \(dy\) gives

\[\frac{d}{dy} (Q + U) = -\rho K_h (Q - U) \quad (A3)\]

Similarly, adding equations (A1) and (A2) and dividing by \(dy\) gives

\[\frac{d}{dy} (Q - U) = -\rho K_h (Q + U) + \rho K_h \alpha (Q + U) \quad (A4)\]

Let

\[R = Q + U \quad (A5)\]

\[T = Q - U \quad (A6)\]

Substituting \(R\) and \(T\) in equations (A3) and (A4) yields

\[\frac{d}{dy} R = -T \rho K_h \quad (A7)\]

\[\frac{d}{dy} T = -\rho K_h R + \rho K_h \alpha R = -\rho K_h (1 - \alpha) R \quad (A8)\]

Differentiating equation (A8)

\[\frac{d^2}{dy^2} T = -\rho K_h (1 - \alpha) \frac{dR}{dy} \quad (A9)\]

Putting \(dR / dy\) from equation (A7) and rearranging equation (A9) yields
\[
\frac{d^2}{dz^2}y - (\rho K_\rho)^2(1 - \alpha)T = 0
\]  
(A10)

Equation (A10) is a second order linear ordinary differential equation that may be written in operational form as
\[
(D - r)(D + r)T = 0
\]  
(A11)

where
\[
D = \frac{d}{dy} \quad \text{and} \quad r = \rho K_\rho \sqrt{1 - \alpha}
\]  
(A12)

Denoting
\[
(D + r)T = T_i
\]  
(A13)

equation (A11) becomes
\[
(D - r)T_i = 0 \quad \text{or equivalently} \quad \frac{dT_i}{dy} - rT_i = 0
\]  
(A14)

Equation (A14) is a first order linear differential equation with solution
\[
T_i = c_i \exp(ry)
\]  
(A15)

Putting \(T_i\) in equation (A13) yields
\[
(D + r)T = c_i \exp(ry) \quad \text{or equivalently} \quad \frac{dT}{dy} + rT = f_i(y)
\]  
(A16)

where,
\[
f_i(y) = c_i \exp(ry)
\]

The solution to first order linear differential equation (A16) is
\[
T = \exp(-ry)\int \exp(ry)f_i(y)dy + c_2 \exp(-ry)
\]
This is the solution to equation (A10).

Calculating $R$ from equation (A8)

$$R = -\frac{1}{(1-\alpha)\rho K_b} \frac{dT}{dy} \quad \text{(A18)}$$

Differentiating equation (A17), we get

$$\frac{dT}{dy} = C_1 r \exp(ry) - C_2 r \exp(-ry) \quad \text{(A19)}$$

Putting $dT/\ dy$ in equation (A18)

$$R = -\frac{1}{(1-\alpha)\rho K_b} [C_1 r \exp(ry) - C_2 r \exp(-ry)]$$

$$= C_2 \frac{r \exp(-ry)}{(1-\alpha)\rho K_b} - C_1 \frac{r \exp(ry)}{(1-\alpha)\rho K_b} \quad \text{(A20)}$$

From equation (A5) and (A6), we have

$$Q = \frac{R + T}{2} \quad \text{(A21)}$$

and
Putting $R$ and $T$ in (A21) and (A22) gives $Q(y)$ and $U(y)$ as functions of depth $y$

$$U = \frac{R - T}{2}$$

(A22)

$$Q(y) = \frac{1}{2} \left[ C_1 \left( 1 - \frac{r}{(1 - \alpha) \rho K_b} \right) \exp(ry) + C_2 \left( 1 + \frac{r}{(1 - \alpha) \rho K_b} \right) \exp(-ry) \right]$$

(A23)

$$U(y) = \frac{1}{2} \left[ -C_1 \left( \frac{r}{(1 - \alpha) \rho K_b} + 1 \right) \exp(ry) + C_2 \left( \frac{r}{(1 - \alpha) \rho K_b} - 1 \right) \exp(-ry) \right]$$

(A24)

Substituting the value of $r$ from equation (A12) in (A23) and (A24) yields

$$Q(y) = \frac{1}{2} \left[ C_1 \left( 1 - \frac{1}{\sqrt{(1 - \alpha) \rho K_b}} \right) \exp(\sqrt{(1 - \alpha) \rho K_b} y) + C_2 \left( 1 + \frac{1}{\sqrt{(1 - \alpha) \rho K_b}} \right) \exp(-\sqrt{(1 - \alpha) \rho K_b} y) \right]$$

(A25)

$$U(y) = \frac{1}{2} \left[ -C_1 \left( \frac{1}{\sqrt{(1 - \alpha) \rho K_b}} + 1 \right) \exp(\sqrt{(1 - \alpha) \rho K_b} y) + C_2 \left( \frac{1}{\sqrt{(1 - \alpha) \rho K_b}} - 1 \right) \exp(-\sqrt{(1 - \alpha) \rho K_b} y) \right]$$

(A26)

Denoting $k' = \sqrt{(1 - \alpha)}$ gives equations (2.10) and (2.11) in the body of the paper.
Appendix B

Coauthor approval letter
11-08-2011

Noah P. Molotch  
Assistant Professor of Geography  
Fellow, Institute of Arctic and Alpine Research  
University of Colorado at Boulder  
Email: Noah.Molotch@colorado.edu

Dear Noah,

I am in the process of preparing my dissertation in the Civil and Environmental Engineering Department at Utah State University. I hope to complete my degree in November 2011.

I am requesting your permission to include the attached paper, of which you are coauthor, as a chapter in my dissertation. I will include acknowledgements to your contributions as indicated. Please advise me of any changes you require.

Please indicate your approval of this request by signing in the space provided, attaching any other form or instruction necessary to confirm permission. If you have any questions, please contact me.

Thank you,

Vinod Mahat

I hereby give permission to Vinod Mahat to use and reprint all of the material that I have contributed to Chapter 3 of this dissertation.

Noah P. Molotch
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36 Aggie Village Apt L, Logan UT 84341, Ph:1-435-797-6656, Email: mahatvinod@hotmail.com

**Education**

**Doctor of philosophy (Ph.D.),** Civil & Environmental Engineering, Utah State University, Logan, UT, USA (expected: November 2011)
- Dissertation: "Effect of vegetation on the accumulation and melting of snow at TW Daniels Experimental Forest"
- Advisor: Professor Dr. David G. Tarboton

**Master of Science (M.S.),** Hydropower Development, Norwegian University of Science & Technology (NTNU), Trondheim, Norway (2006)
- Dissertation: "Flood forecasting model for Telemark water course"
- Advisor: Dr. Knut Alfredsen

**Bachelor of Engineering (B.E),** Civil Engineering, Institute of Engineering, Pulchowk Campus, T. U., Nepal (1997)
- Project: "Pre-feasibility study of Indrawati hydro-electric Project"
- Advisor: Sailendra Basnet

**Research Interests**
- Interactions of land/vegetation, atmosphere and water
- Modeling of snowmelt
- Hydrology, Mountain hydrology
- Rainfall Runoff & Flood forecast Modeling
- Atmospheric radiation transmission processes
- Processes controlling hydrologic fluxes in cold region

**Working Skills**

Hydro & Hydraulic Software: HEC-1/HEC-HMS, HEC-RAS, DAMBRK/FLDwav, SWMM, HEC-ResSim, UEB, HBV, Hydrus-1D

Spatial Data Analysis: ArcGIS, ArcInfo, TauDEM

Programming Languages: Fortran, C++

Statistical and Mathematical Tools: R, MATLAB

Database Management: Microsoft SQL Server.

Remote Sensing Image Analysis: ERDAS IMAGINE

Other Tools: AutoCAD, Photoshop etc.
Publication & Manuscript in Preparation


Mahat, V. and D. G. Tarboton, "Canopy radiation transmission for an energy balance snowmelt model", Water Resources Research (in-review).


Mahat, V. and D. G. Tarboton, "Representation of canopy snow interception, unloading and melt in a physically based snow melt model ". This manuscript is in preparation.

Conference Presentations


Mahat, V. and D. G. Tarboton, (2009), "Effect of vegetation on the accumulation and melting of snow," Poster presentation, American Geophysical Union, Fall Meeting, San Francisco, December 14-18

Mahat, V. and D. G. Tarboton, (2008), "Snow processes as a function of vegetation at the TW Daniels Experimental Forest," Oral presentation, Spring Runoff Conference, Utah State University, Logan, UT, March 31-April 1.
Mahat, V. and D. G. Tarboton, (2007), "Preliminary study of how vegetation impacts snow and water supply at the TW Daniels Experimental forest," Oral presentation, Spring Runoff Conference, Utah State University, Logan, UT, April 5-6.

Field Experience
Field survey at Rocky Mountain (TW Daniels Experimental Forest), Utah for four winters (2006/7-2009/10) to measure the snow depth, density and temperature and snow water equivalent; discharge measurements; topographic survey; construction supervisions of water works etc.

Employment History
Graduate Research Fellow August 2006 - present
Utah State University, USA Development of new parameterization in a physically based energy balance snowmelt model, Snow survey, Monte Carlo flow simulation, Stochastic analysis of stream flow, Flow frequency analysis, Reservoir sizing etc.


Civil Engineer/Hydrologist Nov 1997 - June 2004
(Nepal Electricity Authority), Construction supervision of buildings, dam; tunnel etc, Topography survey, (Kaligandaki Hydro project) and Stream flow discharge measurement, (NEPAL Consult Pvt Ltd) Hydrologic analysis of Phewa hydroelectric project, Spur design etc.

Licensure
Engineer In Training Certification (aka Fundamental of Engineering)

Memberships
Member: American Geophysical Union (AGU)
Member: Nepal Engineers’ Association
Member: Nepal Engineering Council
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