QUANTIFYING THE HYDROLOGIC IMPACT OF MOUNTAIN PINE BEETLE DISTURBANCE IN A WESTERN MONTANA ECOSYSTEM

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QUANTIFYING THE HYDROLOGIC IMPACT OF MOUNTAIN PINE BEETLE DISTURBANCE IN A WESTERN MONTANA ECOSYSTEM

By

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Thesis

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Mountain pine beetle (Dendroctonus ponderosae, MPB) is a forest pest endemic to the Rocky Mountain West. Since the late 1990s, millions of hectares of lodgepole pine forest have experienced extensive tree mortality due to MPB disturbance and this may have significant implications for forested mountain water supplies. MPB disturbance may affect the amount of moisture that enters and leaves the forest hydrologic system, through changes in snowpack accumulation, snowmelt timing, transpiration and subsequently soil water content. The cumulative effect of these changes is that soil moisture is expected to be higher in disturbed forests as the hydrologic system responds to increased inputs and the cessation of canopy transpiration that accompanies tree mortality. This research examined how MPB-disturbance affects the forest water balance in three plots in western Montana using direct observation and modeling methods. Peak SWE, snowmelt and post-snowmelt water balance parameters were measured in three study plots: a non-disturbed lodgepole pine plot, a plot consisting of lodgepole pine trees in the advanced stage of MPB disturbance, and a nearby clear cut. No significant differences in peak SWE and snowmelt timing were measured between the MPB-disturbed and non-disturbed due to the higher stand density and basal area. However, post-snowmelt measurements of soil moisture, rainfall, understory evapotranspiration and canopy transpiration indicated higher net precipitation and understory evapotranspiration in the MPB-disturbed plot. Additionally, soil moisture was higher in the MPB-disturbed plot, which was likely explained by the absence of canopy transpiration fluxes. Additionally, beyond the factors quantified in this initial study, it is likely that topography and variability in stand characteristics played an important role for observed differences in soil water content. This study provides first steps towards assessing the implications of MPB for changes in mountain water supplies in forested catchments. Future work should seek to use additional study plots with more similar stand characteristics and local topography.
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List of Symbols

$C_{at}$ Atmospheric conductance

$C_{can}$ Canopy conductance

$c_a$ Heat capacity of air

$c_i$ Heat capacity of ice

DBH Diameter at breast height

$e_a^*$ Air saturation vapor pressure

$e_a$ Atmospheric vapor pressure

$e_s$ Snow surface vapor pressure

$ET_u$ Understory evapotranspiration

$ET_0$ Reference understory evapotranspiration

$H$ Sensible heatflux

$k$ von Karman’s constant

$\hat{L}$ Proportion of live trees within a study plot

$LAI'$ Effective leaf area index

$P$ Air pressure

$P_n$ Net precipitation

$RH$ Relative humidity

$R_n$ Net radiation

$S$ Sum of energy available for snowmelt

$S_A$ Sapwood area

$\overline{S_A}$ Average sapwood area
\( S_L \)  Sapwood length

\( S_v \)  Sapflux velocity

SAI  Sapwood area index

\( SWE \)  Snow water equivalent

\( T_c \)  Canopy transpiration

\( T_a \)  Air temperature

\( T_s \)  Snowpack temperature

\( v_a \)  Windspeed

\( z_0 \)  Surface roughness height

\( z_a \)  Sensor height (windspeed and air temperature)

\( z_d \)  Zero-plane displacement height

\( z_m \)  Sensor height

\( \Delta \)  Slope of relationship of saturation vapor pressure and temperature

\( \Delta SWS \)  Change in soil water storage

\( \Delta T \)  Difference in thermal dissipation temperature

\( \alpha \)  Upslope contributing area

\( \beta \)  Slope

\( \lambda E \)  Latent heat flux

\( \lambda_v \)  Latent heat of vaporization

\( \lambda_f \)  Latent heat of fusion

\( \rho_a \)  Density of air

\( \rho_w \)  Density of water
\( \theta \quad \text{Soil volumetric moisture content} \)

\( \gamma \quad \text{Psychometric constant} \)
Introduction

Background

The mountain pine beetle (MPB; *Dendroctonus ponderosae*) is an aggressive forest pest that attacks and kills lodgepole pine (*Pinus contorta*) trees by burrowing through layers of outer bark into the phloem and introducing a blue staining fungus (Gibson et al., 2009). The fungus introduced by the beetles inhibits the transport of water from roots to canopy and is considered the primary cause of tree mortality (Hubbard et al., 2013). Since the early 1990s, warmer winters and higher summer temperatures, combined with the long-term effects of fire exclusion, have promoted a dramatic increase in MPB activity and associated tree mortality across most of the Rocky Mountain West (Bentz et al., 2009). In Montana alone, more than six million acres of forest have been impacted by MPB from 1999 to 2012 (Hayes, 2013).

Tree death and the loss of canopy cover due to MPB disturbance reduces interception losses and alters the stand-level energy budget, which together change the stand-level water balance (Boon 2007 & 2008; Pugh and Small 2011). The changes to the stand-level water balance are driven by the loss of needle foliage following infestation and the cessation of transpiration that accompanies tree death (Adams et al., 2011). Following initial infestation, tree needles begin turning red and begin to fall off (“red phase”) (Pugh and Gordon, 2013). Complete loss of needle foliage
typically follows within 3 – 5 years and the tree is said to enter the “gray phase” (Lewis and Huggard, 2010; Pugh and Gordon, 2013).

As the forest hydrologic system responds to potential increases in net precipitation and the cessation of canopy transpiration, the cumulative effect of these changes is expected to be an increase in soil moisture after MPB disturbance (Winkler et al., 2008). The potential impact of these hydrologic changes at the stand level may become important when scaled to the watershed or regional level as MPB disturbance may result in increased water yield and earlier and larger peak flows (Hélie et al., 2005). Potential implications for resource managers include altering the timing of timber harvest due to wetter soils and replacing infrastructure (culverts, bridges) to accommodate possible increased streamflows. Furthermore, these effects can be significant and long lived. An MPB outbreak on Jack Creek in southwestern Montana during the 1970s led to a 15% increase in discharge and a two-week advancement of peak stream flow (Potts 1984). A spruce beetle (Dendroctonus rufipennis Kirby) infestation in the White River watershed of Colorado led to a 15% -18 % increase in average annual water yield (Mitchell and Love, 1973). These high post-disturbance water yields may persist for up to 25 years (Bethalmy, 1974).

Therefore, it is important to learn more about how MPB-disturbance affects hydrologic processes at the stand scale in order to predict the effect at the watershed or regional scales. The research presented here is unique in that it examined the effect of MPB disturbance
on snow accumulation and ablation, and on post-snowmelt hydrologic processes at the stand scale with a combination of observation-based and model-based analyses.

Hydrologic impacts of MPB disturbance

In snow-dominated regions, the interception and sublimation of snow can constitute a sizeable component of the forest water balance, with interception losses greater than 30% of annual snowfall possible (Pomeroy and Schmidt, 1993). Canopy interception is the process by which precipitation falls on plant surfaces, such as foliage and branches, and returns to the atmosphere via evaporation or sublimation (Dingman, 2002). Disturbances to the forest canopy, such as MPB attack, tree harvest or fire, have been shown to result in increased snow accumulation due to the decrease in canopy interception (Adams et al., 2011, Moore and Wondzell, 2005).

Several recent studies have suggested that peak snow water equivalent (SWE, cm) may increase after MPB disturbance. Boon (2012) compared peak SWE between an MPB-disturbed lodgepole pine stand and a non-disturbed stand. Peak SWE was 0.6 cm and 2.3 cm greater in the MPB plot during years of low and high snowfall, respectively (Boon 2012). In a comparison of eight pairs of disturbed and non-disturbed lodgepole pine stands, Pugh and Small (2012) observed higher snow accumulation in plots in the grey stage compared to paired non-disturbed
stands. The effect of MPB disturbance on changes to net precipitation (rainfall) has remained mostly unexamined, with Pina Poujol (2013) a notable exception. Pina Poujol (2013) measured net precipitation in a lodgepole pine stand that was treated with herbicide to replicate MPB disturbance and did not observe a strong treatment effect on precipitation.

Disturbances to the forest canopy also affect the forest energy budget, which may, in turn, alter the rate of snowmelt. MPB-disturbed stands may experience an increase in subcanopy net radiation as canopy coverage decreases. Pugh and Small (2011) observed that gray stage stands transmitted 6.2% more solar radiation than living stands. This increase in energy inputs to the snowpack drives faster snowmelt in MPB disturbed stands than non-infested forests (Boon 2009, Pugh and Small 2012). Boon (2009) observed differences of 0.14 cm d\(^{-1}\) and 0.12 cm d\(^{-1}\) in ablation rates between live and dead stands in 2007 and 2008, respectively. In another study (Winkler et al., 2014), melt rates in an MPB-disturbed stand were 0.08 to 0.32 cm d\(^{-1}\) higher than a non-disturbed mixed stand. Pugh and Small (2011) found significant and higher ablation rates in gray stands compared to living stands. Changes to snow surface albedos during the MPB disturbance cycle also contributes to differences in ablation rates. For instance, although net radiation in red stage stands do not differ significantly from live stands, ablation rates are higher in red stands, and this increase is attributed to the decrease in snow surface albedo due to increased litter fall (Pugh and Small, 2011). Faster ablation
in gray stage stands, though, is attributed to increases in net radiation passing through the thinned canopy (Pugh and Small, 2011).

MPB-induced tree death also eliminates canopy transpiration. Canopy transpiration \( T_c \) is the process where water molecules absorbed by tree roots are translocated vertically through the tree’s vascular system to stomatal cavities on leaves where they are evaporated to the atmosphere (Dingman 2002). Canopy transpiration can remove a significant amount of water from the forest hydrologic system. For instance, Silins et al., (2007) estimated that canopy transpiration in lodgepole pine forests averages 30% of annual precipitation; with total daily rates between 1.5 - 2.0 mm d\(^{-1}\). Knight et al., (1981) found daily transpiration rates in a 100-year-old lodgepole pine stand to be 3.3 – 3.4 mm d\(^{-1}\), and the maximum 24-hour transpiration for the largest trees (20-26 cm DBH) observed in their study was 40 – 44 L. The effect of MPB on individual tree transpiration can be surprisingly quick as declines have been observed 10 days following MPB-infestation and transpiration rates reaching zero within the year after the initial infestation (Hubbard et al., 2013).

Although MPB disturbance reduces water lost through \( T_c \), understory evapotranspiration \( ET_u \), may actually increase. Understory evapotranspiration is controlled by the amount of solar radiation that is transmitted through the canopy (Boon, 2008). Canopy loss increases the wind and solar radiation reaching the forest floor (Adams et al., 2011), which may increase understory evapotranspiration losses. No study to date
has examined how MPB disturbance might affect $ET_u$, but research on timber harvesting, which often serves as a surrogate for MPB disturbance, suggests canopy removal may increase $ET_u$, (Bethalmy 1963). For instance, Simonin et al., (2007) measured higher rates of understory evapotranspiration in a thinned ponderosa stand compared to a non-thinned stand, which the authors attributed to the response of understory vegetation to the increased light intensity and precipitation following thinning.

Ultimately, MPB disturbance may indirectly lead to an increase in soil moisture by altering the water and energy fluxes described above (Adams et al., 2011). It is well understood that soil moisture increases after thinning and harvesting treatments (Spittlehouse 2007; Simonin et al., 2007) but there have been few studies on the effect of MPB disturbance specifically. Clow et al., (2010) measured soil moisture under living and MPB-killed lodgepole pine trees and observed that soil moisture content was 50% higher under the dead trees. In a recent study, where MPB-disturbance was replicated with herbicide application, soil moisture in the top 0-20 cm of soil was up to 31% greater in plots with simulated MPB mortality than non-disturbed plots, respectively (Pina Poujol 2013). Both studies suggested that the higher soil moisture observed after MPB-disturbance was due to the reduction in canopy transpiration in disturbed plots (Clow et al., 2010, Pina Poujol, 2013).
Project Objective

The goal of this research was to determine how MPB disturbance affects hydrologic processes in lodgepole pine forests in western Montana. Specifically, the objectives of this study were to 1) examine the effect of MPB disturbance on snow accumulation and ablation processes, and 2) examine how MPB disturbance may affect the post-snowmelt water balance.

Methods

Experimental design

The experimental design of this project was based on the space-for-time (SFT) model. The SFT design extrapolates a temporal trend from a series of distinct and different aged stands (Pickett, 1989). The underlying assumption of SFT is that the differences between two experimental units, that differ in successional or disturbance states, represent what would be found over time if a single stand experienced the disturbance event. In addition to making it feasible to track temporal changes within a relatively short period of time, this approach also reduced the effect of year-to-year climatic variability. Space-for-time experiments are also especially useful when general or qualitative trends are desired (Pickett, 1988). Within this study, the pre-disturbance hydrologic state is represented by a non-disturbed live stand plot (LS) consisting of lodgepole pine. The LS plot was compared to a plot in the
grey phase (MPB). Observed differences between the LS and MPB plots were assumed to reflect differences that would have been seen if a single stand were measured before and after MPB disturbance. Hydrologic processes in a clear cut plot (CC) were also measured. The CC plot was used to represent hydrologic fluxes in the complete absence of overstory canopy.

**Study area**

The study was conducted within Lubrecht Experimental Forest (LEF) and Bureau of Land Management (BLM) holdings 53 km east of Missoula, Montana. This region was selected for two reasons: 1) non-infested lodgepole pine stands were found within close proximity to MPB-disturbed stands that had similar topography and climate; 2) the close proximity to Missoula, enabled frequent site visits throughout the study period to collect data and to maintain equipment.

The study area was within a continuous forest consisting of mostly mature lodgepole pine and young Douglas-fir (*Pseudotsuga menziesii*). The dominant soil series within this area were Evaro gravelly loam: a loamy skeletal, mixed superactive Lamellic Haplocryepts formed from colluvium derived mainly from argillite and alluvium (USDA NRCS, 2014). Long-term climatic observations (1990-2013) were available from the North Fork Elk Creek SNOTEL site, which was located approximately 4 km away to the southeast at an elevation of 1905 m. Average annual
precipitation at the SNOTEL site was 663 mm. Average April 1 snow water equivalent (SWE, cm) was 28.4 cm. Mean air temperature was 3.1 °C, with monthly averages ranging from – 5.4 to 14.1 °C for January and August, respectively.

Three 50 x 50 m (2500 m²) study plots were established: one (MPB) was in the grey phase of MPB disturbance, the second within a non-disturbed lodgepole pine live stand (LS), and the third within a nearby clearcut stand (CC). The LS and MPB plots represented two distinct stages of the MPB-disturbance cycle (pre-disturbance and post-disturbance), while the CC plot represented the hydrologic responses in the complete absence of an overstory canopy.

The study plots were located within 0.5 km of each other across a shallow ridge. The LS plot was located on the crest of a ridge with an elevation of 1898 m and a slope of 7 %. The MPB and CC plots were situated to the southeast and west at elevations of 1862 m and 1857 m, respectively. The MPB and CC plots were slightly steeper than the LS plot, with slopes of 10% and 17%, respectively.

A 36-point sample grid, with 10-m spacing intervals, was established within each plot for direct measurements of snowpack, soil moisture and canopy characteristics. Weather stations were also installed within each plot to measure the air temperature, relative humidity, wind speed and net radiation¹ at 10-minute intervals during the course of the

¹ See Appendix A for equations used to calculate long-wave radiation in the CC plot.
study (Table 1). These values were used to compute daily averages. A tipping-bucket rain gauge was also located within each plot to measure total daily precipitation.

**Study Period**

Study plots were instrumented in November 2011 but equipment failures and the installation of additional sensors in 2012 resulted in an inconsistent dataset. The results presented here represent the most comprehensive suite of measurements. Data on snow accumulation and ablation processes were collected between March 24 and May 15, 2013. Post-snowmelt water balance measurements began on May 22, 2013, and lasted through July 7, 2013.

**Data Collection**

*Landscape metrics*

Landscape metrics were calculated from a 1-m digital elevation model (DEM) derived from aerial LiDAR data to characterize topographic variability among study plots. The System for Automated Geoscientific Analyses (SAGA) was used to calculate potential incoming solar radiation (PISR, kwh m$^{-2}$ day$^{-1}$) over the study period, March 1, 2013 to July 15, 2013 (Oke, 1998; Wilson and Gallant, 2000; Boehner and Antonic, 2009). Potential incoming solar radiation reflected the variability of slope, elevation and aspect among the study plots and was modeled between
March 1 and July 15 to calculate total bare-ground insolation during the study period. SAGA was also used to calculate the topographic wetness index (TWI) for the study plots (Beven and Kirkby, 1979). TWI is an index of the relative water availability in the landscape, with low values representing dry location in the landscape and high values corresponding to wet locations. TWI was calculated as:

\[
TWI = \ln \left( \frac{\alpha}{\tan(\beta)} \right) \tag{1}
\]

where \( \alpha \) was the upslope accumulated area, and \( \beta \) was local slope. TWI can explain a significant proportion of soil moisture variability across a landscape (Western et al., 2002), and it was used in this study to quantify differences in relative water availability among plots.

**Stand characteristics**

In order to measure stand characteristics, five fixed radius (4 m) circle subplots were established within the LS and MPB plots. The center of each circle subplot corresponded to a randomly chosen grid point within the LS and MPB plots. The circle subplots were used to survey the plots to determine average tree height, diameter at breast height (DBH) and mortality class within the LS and MPB plots. No stand characteristics were measured in the CC plot due to the lack of overstory canopy.
A LI-COR LAI-2000 Plant Canopy Analyzer (PCA) was used to measure canopy characteristics in the LS and MPB plots. The PCA is a commonly used instrument to indirectly measure leaf area index (Keane et al., 2005). Leaf area index (LAI) is the ratio of leaf surface area present in a forest canopy over a given unit of ground surface area (White et al., 1997). The PCA measures differences in solar radiation between sub-canopy measurements and measurements made simultaneously in a clearing (LI-COR 1992). The ratio of the two values gives the amount of solar radiation transmitted through the canopy (Jonkheere et al., 2004), and LAI is then calculated from these measurements (LI-COR 1992). A characteristic of indirect measurements of LAI, such as via the PCA, is that they do not distinguish between photosynthetically active leaf matter and other canopy elements such as trunks, branches and mosses (Jonkheere et al., 2004). As such, indirect estimates of LAI are often described as “effective LAI” to distinguish them from estimates that measure only the photosynthetically active leaf matter (Jonkheere 2005). This report used effective “leaf area index” (LAI’) to describe the canopy characteristics measured with the PCA.

LAI’ measurements (n=36) were collected along the grid points within each plot in pre-dawn conditions. The PCA unit was oriented west and held level at a height of 1.3 m above the forest floor and a 270º lens mask was used to limit direct solar illumination. Above canopy readings were made with a separate PCA unit located in the CC plot. This unit
recorded above canopy readings at 5-minute intervals while a separate unit made sub-canopy measurements in the LS and MPB plots. Readings from the two units were downloaded and analyzed using the FVP-2200 software (LI-COR 1992). LAI’ measurements were averaged to estimate mean LAI’ values for each plot.

To characterize soil structure within each study plot, bulk density samples were collected in June 2013. Four random sampling points were selected in each plot for sampling. Soil pits were dug at these points and samples were taken from the A and E horizons (approximately 0 - 15 cm and 15 – 30 cm depths, respectively) by hammering a brass cylinder (5.08 cm x 5.08 cm) horizontally into the side of each pit. Bulk density and soil porosity were calculated following Dingman (2002).

**Peak SWE**

Snow water equivalent (SWE, cm) was measured three times during 2013. Sample dates were chosen to coincide with the timing of peak SWE in the Elk Creek watershed and to observe changes in SWE during the snowmelt period. A Federal Snow Sampler was used to obtain snow depth and SWE measurements. Measurements were taken according to USDA Soil Conservation Service guidelines (1984) within 0.5 meters of the sample grid points in each plot (n=36). To measure snowpack temperature, iButton temperature data recorders (Maxim Integrated Products, Inc., 2011) were installed in snowpits in each plot. Sensors were
embedded in the side of each pit at depths of 12.5 cm, 37.5 cm, and 62.5 cm from the snowpack surface, and each pit was backfilled after installation (Pugh and Small 2011). The iButtons recorded 2-hour averages of snow pack temperature at each depth to provide a continuous record of snowpack temperature. As the snowpack melted and exposed the buried sensors, the exposed sensors began to track the ambient air temperature. The timing of complete snowmelt in each plot was indirectly inferred from the iButton sensor output: Complete snowmelt was assumed to occur when the temperature measurements converged and approximated the air temperature.

*Snowmelt model*

In the absence of continuous measurements of snowmelt, an energy balance model was used to estimate snowmelt rates and the timing of complete snowpack removal in the study plots. The model was calculated using daily average meteorological data as model inputs for each plot. The snowmelt model was initialized with the peak SWE measurements and snowpack temperature in each plot, and the model was terminated when SWE equaled 0 cm. The following equations were written and analyzed with R (R Development Core Team, 2013) and derived from Dingman (2002).

The energy balance model simulated energy fluxes between the atmosphere and the snow surface:
\[ S = R_n + H + \lambda E \]  \[2\]

Where \( S \) was the energy available for snow ablation (\( \text{W m}^{-2} \) converted to \( \text{MJ m}^{-2} \text{day}^{-1} \)), \( R_n \) was sub-canopy net radiation, \( H \) was turbulent exchange of sensible heat and \( \lambda E \) was latent heat flux (Dingman, 2002). Heat flux from the ground was not included in this model, as its contribution was usually negligible compared with the other energy balance terms. Advective heat input from rainfall was also not included.

In all study plots, the sensible heat flux was a function of the temperature gradient above the snowpack:

\[ H = \rho_a \times C_a \times \left[ \frac{k^2}{\ln(\frac{z_a-z_d}{z_0})} \right] \times v_a \times (T_a - T_s) \]  \[3\]

where \( \rho_a \) was the density of air (1.29 \( \text{kg m}^{-3} \)), \( C_a \) was the heat capacity of air (0.00101 \( \text{MJ kg}^{-1} \text{K}^{-1} \)), \( k \) was von Karman’s constant (0.40), \( z_a \) was the wind speed and air temperature sensor heights, \( z_d \) was the zero-plane displacement height, \( z_0 \) was the surface roughness height, \( v_a \) was windspeed (\( \text{m day}^{-1} \)), \( T_a \) was air temperature (\( ^\circ C \)) and \( T_s \) was the modeled snowpack temperature (\( ^\circ C \)).

The surface roughness height \( z_0 \) described the irregularity of the snow surface and vegetation projecting above the snow surface (Dingman, 2002). Values of \( z_0 \) typically range between 0.0001 m and 0.038 m, however in forested environments \( z_0 \) may be considerably higher.
(Dingman, 2002). For the LS and MPB plots, $z_0$ was parameterized by selecting values that minimized the difference between the modeled melt date and the melt date inferred from the snow temperature measurements described above. In the CC plot, the minimum value of $z_0$ found in the literature (0.0001 m) (Dingman, 2002) was used to parameterize the model as the $z_0$ needed to meet the melt date approximated from the iButtons was outside the range of published values.

Daily latent heat flux in all plots was a function of vapor pressure gradient above the snowpack:

$$\lambda E = \lambda_v \frac{0.622 + \rho_a}{P} \left[ \frac{k^2}{\ln \left( \frac{e_s - e_a}{z_0} \right)} \right] * v_a * (e_a - e_s)$$  \[4\]

where $\lambda_v$ was the latent heat of vaporization (2.47 MJ kg$^{-1}$), $\rho_a$ was the density of air, $P$ was the atmospheric pressure, and $e_a$ and $e_s$ are atmospheric and snow surface vapor pressures (kPa), respectively. The vapor pressure gradient controls whether latent heat flux removes energy from the snow pack through sublimation ($e_s > e_a$) or whether the snowpack gains energy through condensation ($e_a > e_s$).

Atmospheric vapor pressure was calculated as

$$e_a = RH * 0.661 * \exp \left( \frac{17.3 + T_a}{T_a + 237.3} \right)$$
where $RH$ was the measured air relative humidity, and the other terms have been previously described. Snow vapor pressure ($e_s$) was calculated with the same equation but with air temperature replaced with snowpack temperature. The change in snowpack temperature ($\Delta T_s$) was calculated as:

$$\Delta T_s = \frac{S}{c_i \rho_w \Delta SWE}$$ [6]

where $c_i$ was the heat capacity of ice (0.002102 MJ kg$^{-1}$ K$^{-1}$), $\rho_w$ is the density of water (1000 kg m$^{-3}$), and SWE is the previous day’s value. The modeled snow surface temperature was constrained at 0 C, and no melt occurred when $T_s < 0$ C.

To calculate daily change in SWE($\Delta$SWE), S was converted to depth of water:

$$\Delta$SWE = $-\frac{S}{\rho_w \lambda_f}$$ [7]

The $\Delta$SWE was added to the previous day’s value to estimate daily SWE.

Model performance was assessed by comparing modeled SWE to snow survey measurements made during the snowmelt period. Additionally, the model’s ability to predict the timing of complete snow removal in each plot was qualitatively assessed by comparing the models’
snow-free date to the timing of peak soil moisture (0-30 cm) in each plot. The timing of peak soil moisture has been observed to coincide with the date of snowpack disappearance (Molotoch et al., 2009) and provided an indirect assessment of the model performance.

**Soil Moisture**

Volumetric soil water content ($\theta$, m$^3$/m$^3$) within each stand was measured in two ways. First, two CS616 water content reflectometers (Campbell Scientific, Utah, USA) were installed within each plot to measure the average $\theta$ in the 0 – 30 cm depth. One sensor was installed vertically and measured $\theta$ in the 0 – 30 cm depth. The second sensor was installed at a 30° angle to measure $\theta$ in the 0 – 15 cm depth (Campbell Scientific, Utah, USA). These values were averaged together to estimate average $\theta$ in the 0 – 30 cm depth. The reflectometers were set to standard factory calibration settings and soil water content values were converted to a depth of water (mm) by multiplying the $\theta$ by the probe depth (Sun et al., 2010). These were point measurements and did not necessarily represent soil moisture across the entire plot. Therefore, these values were mainly used to qualitatively observe the timing and magnitude of changes to soil moisture.

Spatially distributed measurements of $\theta$ were also collected. These measurements provided snapshots of how soil moisture changed through time and quantified the spatial variability in soil moisture within each plot.
Measurements of spatially distributed soil water content began in May 22, of 2013 and continued until July 7, 2013. A Hydrosense II portable soil moisture sensor (Campbell Scientific, Inc., Utah, USA) was used to measure θ in the top 12 cm of the soil profile at all 36 sample points within each plots. Soil water content values were converted to a depth of water (mm) by multiplying θ by the probe depth (Sun et al., 2010). Measurements were taken within a 0.5-meter radius of the sample point at 7- to 10-day intervals.

Forest water balance

The forest water balance can be described with the following equation:

\[ \Delta SWS = P_n - T_c - ET_u \]  

where \( \Delta SWS \) was the change in θ (mm) measured at the depths described previously, \( P_n \) was measured precipitation (mm), \( T_c \) was overstory evapotranspiration (mm), and \( ET_u \) was understory evapotranspiration (mm). Understory evapotranspiration included transpiration from understory plants, such as shrubs and grasses, and evaporation from the soil surface. This study did not measure water losses due to overland flow, drainage to the water table, or lateral redistribution of soil moisture (interflow). Values for precipitation and \( \Delta SWS \) were obtained from
methods previously listed, while values for $T_C$ and $ET_u$ were estimated via methods described below.

**Understory Evapotranspiration**

Understory evapotranspiration ($ET_u$) was directly estimated by measuring changes in soil moisture within a volume of soil that was isolated from tree roots and, therefore, from the effect of canopy transpiration flux (Simonin et al., 2007). In each plot, a soil profile and associated understory vegetation were removed by shovel and placed into a plastic 5-gallon (18.9 L) bucket. Care was taken to minimally disturb the soil profile and vegetation during this process. The bucket was then placed in the pit from which the soil profile was removed. A CS616 water content reflectometer was installed vertically in the center of each control volume to measure $\theta$ in the first 30 cm of the soil profile. By excluding tree roots, $\Delta SWS$ in the control volumes was equal to the difference between net precipitation and understory evapotranspiration. Therefore, $ET_u$ (mm) in the 0 – 30 cm depth was estimated according to Simonin et al., (2007):

$$ET_u = \theta_i + P_n - \theta_f$$  \[9\]

where $\theta_i$ and $\theta_f$ were the initial and final $\theta$ values (converted to units of depth, mm) within the control volume, respectively, and $P_n$ was net precipitation (mm).
The Penman-Monteith Equation (Dingman, 2002; Allen et al., 1998) was used to estimate reference understory evapotranspiration ($ET_0$). Reference evapotranspiration was defined as the hypothetical evapotranspiration from a reference grass, not limited by soil moisture, with a height of 120 mm, an albedo of 0.23 and a canopy conductance of 14.5 mm s$^{-1}$ (Dingman 2002, Allen et al., 1998). Calculating $ET_0$ provided a way to quantify the evaporative demand of the atmosphere independently of understory vegetation and soil moisture variability among the study plots (Allen et al., 1998).

The Penman-Monteith model combined mass-transfer and energy-balance equations and a conductance term to estimate $ET_0$ (Dingman 2002):

$$ET_0 = \frac{(\Delta R_n + \rho_a c_a C_{at} e^*_a (1 - RH))}{\rho_w \lambda_v \Delta + \gamma^* \left(1 + \frac{C_{at}}{C_{can}}\right)}$$  \hspace{1cm} [10]

where $\Delta$ is the slope of the relation between saturation vapor pressure and temperature, $R_n$ is net radiation (MJ d$^{-1}$), $\rho_a$ is the density of air (1.29 kg m$^{-3}$), $c_a$ is the heat capacity of air (0.00101 MJ kg$^{-1}$ K$^{-1}$), $C_{at}$ is atmospheric conductance (m d$^{-1}$), $C_{can}$ is canopy conductance (14.5 mm s$^{-1}$), $e^*_a$ was the air saturation vapor pressure (kPa), RH is relative humidity (as a ratio), $\rho_w$ was the density of water (1000 kg m$^{-3}$), $\lambda_v$ was the latent
heat of vaporization (2.47 MJ kg\(^{-1}\)), and \(\gamma\) was the psychrometric constant (kPa K\(^{-1}\)). The equation for calculating \(\Delta\) was

\[
\Delta = \left[\frac{2508.3}{(T+237.3)^2}\right] \times \exp\left[\frac{17.3+T}{T+237.3}\right]
\]

[11]

where \(T\) was air temperature (°C).

Atmospheric conductance for water vapor (m d\(^{-1}\)) was calculated as

\[
C_{at} = \frac{v_a}{6.25 \times \left[\ln\left(\frac{z_m - z_d}{z_0}\right)\right]^2}
\]

[12]

where \(v_a\) was wind speed (m d\(^{-1}\)), \(z_m\) was the height of wind speed sensor (2.5 m), \(z_d\) was the zero-plane displacement (m), and \(z_0\) was the roughness height (m). The air saturation vapor pressure (\(e_a^*\), kPa) was calculated as:

\[
e_a^* = 0.611 \times \exp\left[\frac{17.3+T}{T+237.3}\right]
\]

[13]

where \(T\) is in degrees Celsius. The psychrometric constant (\(\gamma\)) is calculated with

\[
\gamma = \frac{c_a + p}{0.622 + \lambda_v}
\]

[14]
Estimates of $ET_0$ were calculated on a daily time step and summed across the post-snowmelt study period to calculate total $ET_0$ within each plot. Long-wave radiation was estimated in the CC plot, so several modification and assumptions were required to calculate $ET_0$ with the Penman–Monteith equation (see Appendix A for the equations and assumptions).

**Canopy transpiration**

The canopy transpiration ($T_c$) within the LS plot was estimated from sap flux velocity measured with thermal dissipation probes (TDP-30, Dynamax Inc., Houston, Texas; Granier, 1985 & 1987). Canopy transpiration was only measured within the LS plot, as $T_c$ was non-existent within the CC plot and assumed to be zero within the MPB plot due to high tree mortality. Six trees within the LS plot were selected for the TDP probes based on their proximity to the data logger. Tree cores taken from the six instrumented trees were used to calculate sapwood area ($S_A$, cm$^2$).

Bromocerol green stain was applied to the tree cores to differentiate the sapwood from the heartwood (Simonin et al., 2007), which was then measured to estimate sapwood length ($S_L$, cm). The following equation was used to calculate sapwood area from DBH (cm) and $S_L$:

$$S_A = \pi \left( \frac{DBH}{2} \right)^2 - \pi \left( \frac{DBH}{2} - S_L \right)^2$$  \[15\]
In addition to collecting tree cores from the six instrumented trees, tree cores samples were taken from an additional 29 trees in the LS plot. Sapwood areas for these trees were calculated to estimate sapwood area index (SAI, $m^2 m^{-2}$) for the LS plot. SAI was the ratio of the total sapwood to ground area (Quinoñez-Piñón, 2007), and was calculating by

$$ SAI = \frac{TPHA \cdot \hat{L} \cdot \overline{SA}}{10000 \ m^2} \ \ \ \ [16] $$

where TPHA was trees per hectare, $\hat{L}$ was the proportion of live trees in the plot and $\overline{SA}$ was the average sapwood area of the 29 trees.

The thermal dissipation probes were installed on the south side of the six sampled trees at approximately 1.3 meters above the forest floor. After installation, the tree trunks and sensors were wrapped in reflective insulating material to protect the sensors from the effect of solar and thermal heating. Each sensor consisted of two 3.0 cm long thermocouple needles, which measured the temperature of the surrounding sapwood. The needles were inserted into two vertically oriented holes drilled 4.0 cm apart. The sensors recorded the temperature difference between the upper needle, which contained a heating element, and the lower needle, which measured the ambient sapwood temperature. The temperature difference ($\Delta T$) was related to the sap flux velocity $S_p$ ($cm \ s^{-1}$) by an empirical relationship (Granier, 1985; 1987):
where

\[ K = \frac{\Delta T_m - \Delta T}{\Delta T} \]  
[18]

and \( \Delta T_m \) was the maximum temperature difference between the needles, which usually occurred during the overnight hours when sapflow was minimal. Measurements of \( S_v \) from the six instrumented trees were scaled up to the stand scale with the following equation (Kume et al., 2010):

\[ T_c = J_s \times SAI \]  
[19]

where \( T_c \) was the stand-scale canopy transpiration (mm h\(^{-1}\)), \( J_s \) was the average sap flux of the six gauged trees (cm h\(^{-1}\)) and SAI was the sapwood area index (m\(^2\) m\(^{-2}\)). The average hourly sapflux (\( J_s \)) was calculated as:

\[ J_s = \frac{1}{n} \times \sum_{i=1}^{n} S_{vi} \times S_{Ai} \times 36000 \text{ s h}^{-1} \]  
[20]

where \( S_{vi} \) was the sapflux velocity of the \( i \)-th tree, and \( S_{Ai} \) was the sapwood area of the corresponding tree (m\(^2\)). Hourly rates of \( T_c \) were summed each day to calculate total daily canopy transpiration (mm d\(^{-1}\)).
Statistical analysis

Statistical analyses of the study results consisted of direct comparisons of peak SWE, the rate and timing of snowmelt, and the post-snowmelt water balance parameters. The relative differences in these measurements among the study plots were assumed to reflect the effect of tree death and canopy loss associated with MPB disturbance. Additionally, a multiple linear regression analysis was used to quantify the influence of topographic and canopy characteristics on soil moisture within each stand. The independent variables of interest in this analysis were TWI, LAI’ and PISR. Sampling date was included as a factor variable in the analysis to account for precipitation events in a non-parametric manner. Within each stand, every sample point (n=36) had a unique value of LAI’ (the exception being the CC plot, where LAI’ was zero), TWI and PISR. Assumptions regarding linearity of the relationships, constant variance and temporal autocorrelation were validated graphically and quantitatively. All statistical analyses were completed using R.

Results

Stand and Topographic Characteristics

The LS and MPB plots did not have identical stand characteristics (Table 2). The LS plot had nearly 2.5 times more trees per hectare than the MPB plot. Additionally, the MPB plot contained taller and larger trees than the LS plot. Although the LS plot had more trees, stand basal area in
the MPB plot was more than twice that in the LS plot due to larger tree size. Tree mortality within the MPB plot was 83%, whereas tree mortality in the LS plot was 16%.

Average soil bulk density ranged from 0.99 g cc\(^{-1}\) within the LS plot to 1.26 g cc\(^{-1}\) within the CC plot. Soil porosity ranged from 0.53 in the CC plot to 0.63 in the LS plot. The understory vegetation heights in the LS and MPB plots were roughly 0.3 m and 1 m respectively. The CC plot had only sparsely distributed vegetation, so understory vegetation mean height was assumed to be zero.

The mean topographic wetness index (TWI) for the MPB and LS plots were almost identical, 3.6 and 3.7, respectively, with the CC plot slightly higher (4.6). Total potential incoming solar radiation was highest in the MPB plot, while CC and LS plots were nearly equal (Table 3). The average LAI’ for the MPB plot was 1.6 and 1.3 for the LS plot. The CC plot did not contain overstory canopy therefore LAI’ was zero.

**Snow Accumulation and Ablation Results**

During the snowmelt period (March 24 – May 15), mean daily air temperatures were similar in the MPB and LS plots (1.13 °C and 1.47 °C, respectively), and the CC plot was considerably warmer (2.21 °C). Relative humidity was nearly identical among all the plots (Table 4). Net radiation in the CC plot was more than twice the sub-canopy measurements, and much more variable (Fig. 7), in the LS and MPB plots,
respectively. Appreciably higher winds also passed through the CC plot, both in terms of average and extreme windspeeds (Fig. 2).

The March 24 sampling event closely coincided with the date of peak SWE at the North Fork Elk Creek SNOTEL site, which was recorded on March 26, 2013 (21.1 cm). The long-term average (29 year) peak SWE at the SNOTEL site is 28.7 cm. Peak SWE was highest in the MPB plot (plot mean of 17.0 cm), followed by the LS plot (plot mean 16.0 cm), and lowest in the CC plot (plot mean of 15.0 cm). Figure 3 shows the distribution of SWE measured during the snow surveys. The CC plot consistently had the least snowpack during the snowmelt period, and, based on snow survey observations, it was snowfree the earliest (May 3). Median snowpack in the MPB plot was consistently higher than in the LS plot (Fig. 3). The greatest difference in SWE among the plots was observed on April 17. In the CC plot, there were several sample points with trace snow cover and one point that was completely bare. By the May 3 sampling event, the CC plot was completely snow-free while snow cover was still continuous in the LS and MPB plots. May 3 was the final SWE sampling event so the exact date of complete snowmelt in the MPB and LS plots were not directly observed.

Snowpack temperatures in all plots became isothermal at 0° C on March 28 (Fig. 4). Fluctuations in snowpack temperature and the timing of isothermal conditions were similar in the MPB and LS plots. Complete snowmelt was inferred when all iButton sensor measurements were
positive and converging on the ambient air temperature. In the LS and MPB plots this occurred on May 5 and May 7, respectively. In the CC plot, the temperature sensor buried at the 12 cm depth began to fluctuate with average air temperature on March 30, which suggested that enough snow had melted to expose it to ambient air temperatures and solar radiation. However, the remaining sensors appeared to be snow-covered until approximately April 20.

The snowmelt model predicted complete snowmelt in the CC plot on April 14, followed by the LS plot (May 8), and lastly, the MPB plot (May 10) (Fig. 5). Snowmelt rates calculated from model estimates and snow survey measurements (change in SWE over change in time) were highest in the CC plot (Table 5) except for the period between April 17 and May 5, when the observed melt rate was highest in the MPB plot. There was little difference in observed snowmelt rates ( < 0.1 cm d\(^{-1}\)) between the LS and MPB plots.

Modeled SWE agreed reasonably well with direct observations made on April 17 and May 3 in the MPB and LS plots (Table 6, Fig. 6). In the MPB plot, the modeled SWE was more than 4 cm higher than observed SWE on both snow survey dates. In the LS plot, modeled SWE was also approximately 4 cm higher than direct observations on April 17, and 2.5 cm higher on May 3. The model performed poorly in the CC plot throughout the snowmelt period, consistently overpredicting snowmelt.
For instance, the model predicted complete snowmelt on April 14, whereas the 3.7 cm of SWE was measured three days later on April 17, although the snow cover was not continuous.

The model’s ability to predict the timing of complete snowmelt was indirectly assessed by comparing the predicted date of complete snowmelt to the date of peak θ (0 – 30 cm) in each plot (Fig. 7). Peak soil moisture in the LS and MPB plots occurred on May 5 and May 7, respectively. The model predicted complete snowmelt in the LS on May 8, which lagged peak θ by three days. In the MPB plot, the lag between the modeled melt date (May 10) and peak θ (May 7) was also three days. In the CC plot, peak θ (April 5) preceded the modeled snowfree date (April 14) by more than a week, and preceded the observed snow-free date (May 3) by almost a month.

**Post-snowmelt Hydrologic Processes and Plot Water Balances Results**

Mean daily air temperature and relative humidity were similar across all stands during the post-snowmelt study period (Table 7). Net radiation and windspeed were both higher, and more variable, in the CC plot than measurements in the LS and MPB plots, respectively (Fig. 8).

**Net Precipitation and Canopy Transpiration**

Net precipitation was highest in the CC plot (107.6 mm) followed by the MPB plot (99.6 mm) and lowest in the LS plot (89.0 mm) (Table
Four days of rainfall data (June 7 – June 11) in the MPB plot were missing due to power failure in the rain gauge. Values for the two rainfall events during this period were estimated by via the linear relationship between rainfall in the CC and MPB plots ($R^2 = 0.73, p < 0.001$). Total canopy transpiration ($T_c$) in the LS plot during the study period was 28.6 mm. The average daily rate was 0.6 mm d$^{-1}$. Canopy transpiration in the MPB and CC plots was assumed to be zero (Table 8).

**Understory Evapotranspiration**

Understory evapotranspiration ($ET_u$) calculated within the control volumes was highest in the MPB plot (129.6 mm) and lowest in the LS plot (110 mm) (Table 9). $ET_u$ in the CC plot was 126.6 mm. Reference understory evapotranspiration ($ET_0$) estimated with the Penman-Monteith model was highest in the CC plot (88.9 mm). Estimates in the LS and MPB plots were relatively similar (52.4 mm and 46.9 mm, respectively).

**Soil moisture and plot-scale water balance**

Soil moisture was consistently higher in the MPB plot than in the LS and CC plots. The higher $\theta$ in the MPB plot was observed in the 0 – 30 cm measurements (Fig. 9) and as well in the spatially distributed measurements at 0-12 cm depth (Fig. 10). For both measurement depths, the LS plot had the lowest $\theta$, with the CC plot only slightly wetter. The time series of $\theta$ measured in the 0 – 30 cm depths showed similar patterns.
of wetting and drying in response to rainfall events and evapotranspiration, although the live stand did not exhibit a peak in soil moisture in late May (Fig. 9, bottom panel). Average θ in the 0 – 12 cm depths were generally higher than the deeper measurements, although the latter were usually captured within the variability of the spatially distributed measurements. Table 8 presents the post-snowmelt water balances for the study plots. Water balances for all plots were negative, with the LS plot experiencing the largest absolute ΔSWS. The ΔSWS of -49.6 mm in the LS, which included losses due to canopy transpiration, was 20 and 30 mm below ΔSWS in the MPB and CC plots, respectively.

Soil Moisture Multiple Linear Regression Analysis

The multiple regression model explained 60 percent of the variability in soil moisture measurements ($R^2=0.60$, $p < 0.001$) (Table 10). Increases in TWI were positively associated with soil moisture while an increase in potential solar radiation was negatively associated. The interaction effect between stand and PISR was assessed and found to not improve the explanatory power of the model. LAI′ was positively associated with soil moisture in the MPB stand and negatively associated in the LS stand. The sum of squares value described the error explained by each model term after all others have been accounted for (Table 10). The interaction between stand and LAI′ and the sample date factor accounted for most of the variance in the soil moisture measurements. TWI and
potential incoming solar radiation accounted for similarly minimal components of the variability, although both were statistically significant \((p < 0.05)\)

**Discussion**

The objective of this research was to examine how canopy loss and tree death caused by MPB disturbance affected hydrologic processes in lodgepole pine forests. Minor differences in snow accumulation and ablation processes were observed between the disturbed (MPB) and non-disturbed (LS) plots. Results of the post-snowmelt water balance measurements indicated higher soil moisture in the MPB plot compared to the LS plot. Water balances for all plots were negative, which suggested that evapotranspiration fluxes were greater than precipitation inputs during the post-snowmelt study period. Net precipitation was 10 mm higher in the MPB plot than the LS plot, but water lost through canopy transpiration in the LS plot accounted for the greatest difference between plot water balances. The measurements in the clear cut plot (CC) were generally consistent with the expected effect of complete canopy removal. The results of the multiple linear regression analysis suggested that canopy structure was the most important factor influencing the spatial variability of soil moisture within each plot. In summary, although this study did not observe a large difference in snow accumulation and ablation between the
MPB and LS plots, the results of the water balance measurements were generally consistent with the expected outcome of MPB disturbance.

Stand Characteristics

Canopy loss and tree mortality induced by MPB disturbance are the primary drivers behind changes to the forest water balance. Thus, it is important to comment on how the stand and canopy characteristics of the study plots may have influenced the results. The study plots were selected to represent two points in a chronosequence of the MBP disturbance cycle—from pre-disturbance (LS) through advanced disturbance (MPB). The CC plot represented hydrologic processes in the complete absence of any canopy effect. The canopy conditions and tree mortality in the LS and MPB plots appropriately represented this chronosequence (i.e. the trees in the LS plot were alive and non-infested and the trees in the MPB plot were nearly all dead and had lost most needles). However, stand density, tree height and tree diameter were not identical between stands. For instance, trees in the MPB plot were fewer in number but 2.5 times broader and 2.4 times taller than trees in the LS plot. LAI′ was also slightly higher in the MPB plot than the LS plot (1.6 versus 1.3, respectively). Such differences in stand structure might seem extreme but these discrepancies are comparable with previous studies that have examined stand-level effects of MPB disturbance. Pugh and Small (2011) compared snow accumulation and melt in eight pairs of infested and non-infested lodgepole pine stands.
In that study, absolute differences in stem density between their plot pairs ranged from 87.4% to 0.9%, and absolute differences in DBH ranged from 143.7% to 13.5% (Pugh and Small 2011). Boon (2009) compared snow hydrologic processes in two lodgepole stands with dissimilar characteristics: Tree height was 2 times taller and DBH was 2.7 times greater in the MPB-disturbed plot than the non-disturbed plot.

The relationship observed between LAI' in the LS and MPB plots disagreed with measurements made in other studies with similar methods (Pugh and Gordon 2013, Winkler et al., 2014). For instance, although the MPB plot had lost most of its needle foliage, the average effective LAI (LAI') was actually higher in the MPB plot than the LS plot (1.6 and 1.3, respectively). This was counter to expectations since MPB disturbance will reduce canopy cover once needles begin to fall (Pugh and Gordon 2013). One explanation of the higher LAI' in the MPB plot was that LAI' measures total plant area, which includes woody materials such as trunks and branches, not just needle foliage. Therefore, the larger, taller trees in the MBP plot may have contributed to the higher LAI' measurement. Additionally, substantial dark mosses were observed clinging to tree crowns throughout the MPB plot, which would have generated higher LAI' despite the loss of needle foliage. Comparing stands with different characteristics introduces considerable uncertainty but it was almost unavoidable when studying MPB disturbance. MPB preferentially select larger trees for hosts and do
not generally infest trees smaller than 10 cm (Cole and Amman 1969). By default, this disturbance pattern creates a forest where most of the large trees are dead and only the smaller trees remain living. Although the disturbed and non-disturbed stands used for this study appropriately represented pre- and post-disturbance conditions in terms of tree mortality and canopy foliage conditions, the dissimilarity in stand structure and lack of replication made it difficult to attribute a particular effect to MPB disturbance and limits the range of inferences that can be drawn from these results.

**Topographic Variability**

Local topography has been indicated as a major source of variability for subsurface flow of water (Beven and Kirkby, 1979; Jencso et al., 2009) and the persistence of soil water (Western and Grayson; Varhola et al., 2010) in many hydrologic studies ( ). This study area was selected because it provided pre- and post-disturbance stands and a clearing in close proximity to one another. However, the plots were arrayed across a relatively planar ridgeline so topography was consistent across the study plots. The CC plot was the most topographically dissimilar of the three study plots. The dominant aspect in the CC plot was westerly, while the MPB and LS plots were southeast and slightly southeast, respectively. The CC plot was also considerably steeper than
the LS and MPB plots. Topographic differences between the LS and MPB plots, however, were substantially less severe.

These topographic differences may have influenced the results, but the landscape metrics calculated from the bare ground digital elevation model (DEM) suggested that topography did not have a strong effect on insolation or relative water availability among the plots. The average TWI was nearly identical in the LS and MPB plots (3.7 and 3.6, respectively) and average TWI in the CC plot was 4.6. TWI typically ranges from 1 (drier) to 20 (wetter) (Lin et al., 2006), and the low TWI in the study plots likely reflected their upland positions within the watershed. Potential incoming solar radiation (PISR), which represented the effect of slope and aspect, was also fairly consistent across the study plots. Total PISR during the study period was nearly identical in the CC and LS plots (1087 kwh m$^{-2}$ and 1096 kwh m$^{-2}$, respectively), while the MPB experienced slightly more insolation (1113 kwh m$^{-2}$). Thus, the results of the landscape analysis suggested that underlying topographic differences did not have a strong effect on soil moisture or solar radiation.

Peak SWE

The results of the peak SWE measurements did not show a strong effect of MPB disturbance on snow accumulation. Although peak SWE was higher in the MPB plot than in the LS plot, the difference was small (1 cm) and likely not meaningful given the potentially large error
associated with measurements of peak SWE (Varhola et al., 2010). Federal snow samplers are widely used across North America but measurement errors of up to 12% are possible with this instrument (Varhola et al., 2010). Other sources of error in SWE measurements can include uncalibrated springs or the contamination of the snow sample with soil or debris (Winkler et al., 2005). Therefore, an unknown but likely non-zero error was associated with the measurements of peak SWE. Furthermore, given the dissimilarity in stand characteristics and canopy structure it was impossible to know whether these results accurately reflected the effect of MPB disturbance on snow accumulation. For this reason, clearings are often used in studies of snow accumulation to serve as a reference condition for maximum accumulation. However, peak SWE in the CC plot was lower than in the MPB or LS plot. This was unexpected as numerous studies have shown that snow accumulation is greater in clearings than in forested plots (Moore and Wondzell, 2005). The low SWE in the CC plot was likely explained by the plot’s westerly aspect, which exposed it to strong, prevailing winds. Average windspeed in the CC plot was 5 and 3 times faster than in the MPB and LS plots, respectively, and was coupled with much higher peak windspeeds. The high winds can reduce peak SWE by redistributing snow to the plot’s margins and by increasing snowpack lost via sublimation (Golding and Swanson 1986), and these increased losses may offset the effect of increased accumulation due to reduced interception (Woods et al., 2006).
Snow Ablation

The results of both the model and direct measurements showed that snowmelt occurred the quickest in the CC plot. The rapid snowmelt measured in the CC plot was consistent with previous studies that observed faster snowmelt in clearings than in forested locations (Boon 2009; Winkler et al., 2014). The relatively high melt rate observed in the CC plot can be explained by the complete lack of forest canopy in this plot. Canopy removal exposed the snowpack to greater incident solar radiation and higher wind speeds, which can increase energy inputs to drive ablation processes (Moore and Wondzell, 2005). This was supported by the measurements of net radiation and windspeed in the CC plot during the snowmelt period. Net radiation in the CC plot was more than twice the values in the MPB or LS plots, and windspeed was also considerably higher in the CC plot.

Snowmelt in the MPB and LS plots occurred after the May 3 snow survey, so the exact date of complete snow removal in these plots was not observed and had to be estimated with the energy balance model. The model predicted that complete snowmelt in the LS plot occurred on May 8, with complete snowmelt in the MPB plot occurring two days later on May 10. These results were inconsistent with previous studies that observed faster snowmelt in MPB-disturbed stands (Boon 2009, Winkler et al., 2014). Furthermore, average daily snowmelt (calculated from the
change in SWE between March 24 and the modeled melt date) was identical in the MPB and LS plots.

The lack of difference between melt rates in the MPB and LS plots was likely explained by the similar energy balances measured in these plots. The average daily energy available for snowmelt calculated with the snowmelt model was only slightly higher in the LS plot (19.11 W m\(^{-2}\) d\(^{-1}\)) compared to the MPB plot (18.65 W m\(^{-2}\) d\(^{-1}\)). The sub-canopy energy balance is a function of forest canopy (Boon 2009), and, as previously mentioned, LAI’ in the MPB and LS plots were not that dissimilar (1.6 and 1.3, respectively). Therefore, it was not unexpected that snowmelt rates were also similar.

The model performed relatively well in predicting snowmelt in the LS and MPB plots. In the MPB plot, modeled SWE was 4 mm (29% and 43%) higher than the average SWE measured during snow surveys on April 17 and May 3. In the LS plot, modeled SWE was 4 mm (32%) and 2 mm (36%) greater than average SWE measured during the April 17 and May 3 snow surveys. In the CC plot, however, the model did a poor job of predicting snowmelt. The model predicted complete snowmelt in that plot on April 14, but the average SWE measured on April 17 was 3.7 cm. However, the April 17 snow survey found several sample points with trace snow and one sample point that was bare ground.

Snowmelt model performance was assessed indirectly by comparing the modeled date of snow removal to the timing of peak \(\theta\) (0-
30 cm) in each plot. The timing of peak soil moisture has been observed to coincide with the date of snowpack disappearance (Moltoch et al., 2009), so this comparison provided an indirect and coarse check of the model’s ability to predict when each plot became snowfree. In the LS and MBP plots, maximum θ occurred three days before the models predicted complete snowmelt in these plots, which suggested a reasonable level of model performance. In the CC plot however, peak θ (April 5) occurred more than a week before the modeled snowfree date (April 14). The lack of agreement between the soil measurement and the modeled and observed SWE measurements, suggested that there may have been isolated snowmelt in the vicinity of the soil moisture sensors that caused the early spike in θ. The lack of agreement between the model and observed ablation rates in the CC plot suggested that the factors that drive snowmelt in this plot were perhaps too complex to be accurately modeled (Varhola et al., 2010). For instance, an evaluation of multiple snowpack models (Rutter et al., 2009) concluded that no model best fits all locations and that a model that perform well in forested plots may not perform as well in clearings. Additionally, the assumptions required to estimate net long-wave radiation introduced considerable uncertainty that was propagated through the model. For instance, atmospheric emissivity was estimated under clear sky conditions and not adjusted for cloudiness (Dingman, 2002), which may not have reflected actual conditions. However, the close agreement between the modeled and measured snowmelt rates in the LS
and MPB plots suggested that the model did a reasonable job of predicting the timing of snowmelt in those plots.

Overall, large differences between the MPB and LS plots in terms of snow accumulation and ablation were not observed. The quantification of any real effect of MPB disturbance was confounded by the dissimilarity in stand characteristics between the MPB and LS plots. However, canopy removal had a clear effect on peak SWE and snowmelt in the CC plot. These results suggested that the effect of MPB disturbance on snow accumulation and ablation will be variable depending on stand characteristics and local topography.

Net precipitation and canopy transpiration

Net precipitation is expected to increase as canopy cover decreases following MPB disturbance (Winkler et al., 2014). However, because the LS and MPB plots had different underlying stand characteristics, this study was unable to definitively quantify the magnitude of change to net precipitation caused by MPB-induced canopy loss. Furthermore, although the MPB plot recorded 10 mm more of rainfall than the LS plot, the error associated with estimating rainfall for the four days of missing data, along with the uncertainty of the rain gauge, suggested that this difference may not be meaningful. Measurements of net precipitation in the CC plot, however, showed a clear effect of canopy removal on net precipitation. Net precipitation in the CC plot was 8 mm
and 18 mm higher than rainfall in the forested MPB and LS plots, respectively. This was consistent with studies that observed higher net precipitation after thinning or clear cut treatments (Spittlehouse, 2007, Simonin et al., 2007).

The results of the transpiration measurements generally agreed with reported values of canopy transpiration rates in lodgepole pine forests. The median daily canopy transpiration in the LS plot was 0.57 mm d\(^{-1}\), which agreed with the median canopy transpiration rates of 0.48 and 0.71 mm d\(^{-1}\) observed in a lodgepole pine stand in Alberta (Pina Poujol, 2013). However, these measurements were lower than the 2.6 mm d\(^{-1}\) Pataki et al. (2000) observed in a lodgepole pine stand with larger trees. Knight et al. (1981) measured transpiration in a lodgepole pine stand using whole tree potometers and found a strong linear relationship between maximum daily transpiration and basal area. Using their model and the average per-tree basal area of the six sampled trees (164.5 cm\(^2\)), maximum daily transpiration was predicted to be 13.5 L, which was slightly higher than the measured average maximum daily transpiration of 9.9 L.

Canopy transpiration in the MPB plot was assumed to be non-existent given the high tree mortality observed there (83%) but sapflux was not measured in this plot and the validity of this assumption was not tested. However, this assumption was supported by a recent study (Hubbard et al., 2012) that monitored the decline in transpiration in 17 trees attacked by MPB and observed a rapid decrease in transpiration
within 10 days of MPB infestation. Transpiration had completely ceased by the following year (Hubbard et al., 2012). Given the dissimilar stand characteristics between the MPB and LS plots, it was unlikely that the canopy transpiration estimated in the LS plot reflected the pre-disturbance canopy transpiration in the MPB plot. Using 621 cm² as the average per-tree basal area in the MPB plot and the linear relationship between basal area and transpiration described previously (Knight et al., 1980), the maximum daily transpiration was estimated to be 52.8 L, which was almost 5 times greater than what was observed in the LS plot. Therefore, the change in canopy transpiration in the MPB plot after tree die off was possibly greater than the 28.6 mm measured in the LS plot.

Canopy transpiration $T_c$ in the LS plot was estimated using thermal dissipation probes and the empirical relationship developed by Granier (1985, 1987). This method is widely used for its relative simplicity and agreement with other methods (Granier et al., 1996; Saugier et al., 1997), but error may be introduced in expanding the tree-scale measurements to the plot scale (Granier et al., 1996; Kume et al., 2009; Kumagai et al., 2005a). Sapwood area is often not uniform around most trees stems which causes variability in sapflux around the tree trunk (Clearwater et al., 1999). However, this radial variability in sapflux is thought to be less than inter-tree variability (Kumagai et al., 2005a), and some researchers have suggested allocating sap flux sensors across as many trees as possible (Vertessy et al., 1997). Kumagai et al. (2005a) recommends monitoring a
minimum of six trees to capture tree-to-tree variability. Six trees were monitored in the LS plot, a number that was likely more than sufficient to capture the tree-to-tree variability within the stand given the relative homogeneity of tree sizes and height measured in the LS plot. However, the steps involved with scaling these measurements to the plot level potentially introduced additional error into the results (Oishi, 2008). For instance, proper scaling requires an accurate estimate of sapwood area index, which is the ratio of total sapwood area and the research area footprint. Besides the six monitored trees, sapwood depths were taken from an additional 29 trees within the LS plot to better estimate the SAI. However, sampling error was still likely present in the estimates of sap flux. For instance, estimates of canopy transpiration in a study that monitored sapflux in 15 trees found potential errors of up to 21% (Kume et al., 2009).

**Understory Evapotranspiration**

The results of measurements of understory evapotranspiration ($ET_u$) were broadly consistent with the expected effect of canopy disturbance and tree die off. $ET_u$ was highest in the MPB and CC plots, the two sites that experienced canopy disturbance and tree removal/tree death. The relatively high $ET_u$ in the CC plot was consistent with previous studies that predicted higher $ET_u$ rates following canopy removal.
(Simonin et al., 2007; Bethalmy, 1967). Canopy removal in the CC plot allowed increased wind and solar radiation to reach the forest floor to drive evaporative processes (Adams et al., 2011; Baldoci and Ryu, 2011). This inference was supported by the significantly greater average daily net radiation observed in the CC plot (57.1 W m$^{-2}$) compared to the LS and MPB plots (13.3 and 15.5 W m$^{-2}$, respectively) between May 22 and July 7. Average wind speed in the CC plot was twice and nearly twice the averages in the LS and MPB plots, respectively. Understory vegetation was beginning to reestablish itself in the CC plot during the study period but was generally discontinuous, which suggested that bareground evaporation comprised the majority of $ET_u$ in this plot.

Understory evapotranspiration in the MPB plot was 20 mm greater than the LS plot. Although this relationship was consistent with the expected effect of MPB-associated tree mortality, the higher temperature, wind speed and lower humidity in the LS plot suggested that evaporative demand would be greater in the non-disturbed stand. Soil moisture, however, was higher in the MPB plot, whereas the lower soil moisture in the LS plot may have constrained $ET_u$. The fact that $ET_u$ was actually higher in the MPB plot than in the CC plot was surprising given the higher net radiation and windspeed measured in the CC plot. However, this may be explained by the fact that the evaporative flux observed in the CC plot was likely limited to just bareground evapotranspiration, whereas the MPB experienced the combined fluxes of evaporation and transpiration.
processes. It should be noted that the values of $ET_u$ presented here were calculated with Equation 9, and are affected by the accumulated error of the instruments used to measure net precipitation and soil moisture. Additionally, two days of net precipitation measurements had to be estimated in the MPB plot, which likely introduced additional errors.

The Penman-Monteith model was used to estimate reference evapotranspiration ($ET_0$) in the study plots. The $ET_0$ estimates provided a way to quantify the subcanopy evaporative demand that was independent of understory vegetation and soil moisture variability among the plots. $ET_0$ was highest in the CC plot, which was to be expected given higher net radiation and other factors previously described. Interestingly, $ET_0$ was higher in the LS plot than in the MPB, which was opposite of what was observed in the control volumes. This suggested that the understory microclimatology conditions (e.g. net radiation, humidity and windspeed) in the LS plot generated higher evaporative demand than in the MPB plot. Although net radiation was higher in the MPB plot, the LS plot was warmer and experienced slightly higher winds which may account for the higher evaporative demand. However, the 7 mm difference between the two estimates was likely within the errors associated with the climate data that was used to parameterize the model.

The lack of agreement in the terms of magnitude between the $ET_0$ and the $ET_u$ measured within the control volumes was not unexpected and can be explained by the fact that the reference crop parameters did not
represent the understory vegetation characteristics, or the soil moisture conditions, found in the study plots.

Soil moisture

The results of the soil moisture measurements were consistent with previous studies that have documented increased soil moisture following canopy disturbance and tree death (Spittlehouse 2007; Adams et al., 1991; Clow et al., 2011). Soil moisture was highest in the MPB plot and lowest in the LS plot at both the 0 – 30 cm and 0 – 12 cm depth intervals. Point measurements were used to calculate average volumetric water content (θ) in the 0-30 cm soil depths in each plot. These measurements provided continuous soil moisture data but they only captured θ at the point scale and soil moisture can have extreme variability across a landscape (Western et al., 2002). To better estimate average soil moisture θ given this high spatial variability, soil moisture (0-12 cm depth) was measured at 36 sample points within each plot. In all plots, the average θ in the 0-12 cm depth was generally higher than average θ in the 0-30 cm but displayed similar temporal trends. The difference in θ between the two depths potentially reflected inherent spatial variability rather than real difference in soil moisture. For instance, the range of the spatially distributed samples generally captured θ measured in the 0 – 30 cm depth.

A multiple linear regression model was used to further examine how canopy structure and topographic factors influenced soil moisture
within each plot. The model explained 60% of the variability in soil moisture at the 0-12 cm depth and provided insight into how soil moisture was influenced by canopy characteristics and topography. Topographic factors were represented by the TWI, and the joint effect of slope and aspect was represented in the model-generated potential incoming solar radiation (PISR). Neither of these terms accounted for much variance in the model (sum of squares: 204.4 and 293.9, respectively), which suggested that topography had little influence on soil moisture in the study plots. The interaction of sample date and LAI′ in the MPB and LS plots was statistically significant and accounted for most of the variance in soil moisture measurements (sum of squares: 22451.5). The sample date was included in the model to remove the effect of climatic changes (e.g., increasing net radiation and air temperature) and rainfall events during the post-snowmelt sampling period. The results of the model suggested that LAI′ was a significant predictor of soil moisture in the LS and MPB plots but the effect of LAI′ was different in each stand. For instance, LAI′ was negatively associated with soil moisture in the LS plot, while the association was positive in the MPB plot. This may be attributed to the fact that the LAI′ in the LS plot represented a live, photosynthetically active canopy, whereas the LAI′ in the MPB plot represented a dead, non-transpiring canopy. In the LS plot, soil moisture was dominated by canopy transpiration so an increase in the amount of active canopy would also increase transpiration losses and potentially lead to drier soils (Clow et al.,
Soil moisture in the MPB plot, however, was dominated by $ET_u$, which is driven by radiation inputs. LAI’ in the MPB plot was not photosynthetically active, so an increase in LAI’ in this plot would provide additional shade that would attenuate $ET_u$ losses. Although it is contradictory to imagine increasing canopy cover in a disturbance cycle characterized by canopy loss, this model provided insight into the contradictory effect that tree cover has in pre- and post-disturbance conditions.

**Plot-Scale Water Balances**

Direct measurements of net precipitation, canopy transpiration and $ET_u$ were used to calculate post-snowmelt water balances (Equation 8) for each plot. Water balances were negative in all plots, which suggested that evapotranspirative fluxes were greater than precipitation inputs.

Canopy transpiration comprised the largest proportion of the differences in water fluxes between the MPB and LS plots. The MPB plot experienced 10 mm more $P_n$, and $ET_u$ was 20 mm higher than in the LS plot. However, $T_C$ in the LS plot accounted for nearly 50% of the total relative difference in water fluxes between the LS and MPB plots. Compared to the CC plot, which experienced 18.6 mm more $P_n$ and 16.6 mm more $ET_u$ flux than the LS plot, $T_C$ comprised approximately 45% of the total relative difference between these plots. Therefore, these results
suggested that the cessation of $T_C$ that followed MPB disturbance and harvesting the in the MPB and CC plots, respectively, likely explained the wetter soils observed in these plots compared to LS plot.

Although these results were consistent with the response expected following MPB disturbance or tree harvesting, the water balance model used in this study (Equation 8) was quite simple and relied on several assumptions that were not tested. Thus, the results presented here may not have captured all water fluxes in the forest hydrologic system. For instance, Equation 8 did not account for vertical drainage or lateral movement of soil moisture. Additionally, the water balance also assumed that overland flow did not deliver or remove moisture to and from the plots. As mentioned previously, $T_C$ in the MPB plot was assumed to be zero and was not monitored. It was likely, however, that the few live lodgepole pine trees remaining in the canopy would have removed an unknown but non-zero volume of water.

Overall, the results of the post-snowmelt water balance measurements suggested that soil moisture may increase after MPB disturbance primarily due to the cessation of $T_C$ that accompanies MPB infestation.

**Conclusion**

Contrary to the expected effect of MPB disturbance, this study did not observe large differences in peak SWE accumulation or snow ablation
rates between the MPB and LS plots. Peak SWE was lowest in the CC plot and it experienced the quickest snowmelt of all plots. Measurements of the post-snowmelt water balances, however, were generally consistent with the expected effect of MPB disturbance. Specifically, net precipitation and soil moisture were higher in the MPB plot than the LS plot. Although the CC and MPB plots experienced increased net precipitation and understory evapotranspiration compared to the LS plot, the lack of canopy transpiration represented the greatest difference in plot-scale water balances among the plots. Cumulatively, these results suggested that even in the absence of any impact to peak SWE, snow ablation or rainfall, MPB-disturbed stands may likely experience increased soil moisture due to the cessation of $T_C$. However, any broad inferences that can be drawn from this research are limited by the variability in stand characteristics among the study plots. Therefore, this research suggests that any future efforts to investigate the hydrologic impacts of MPB disturbance should incorporate multiple study plots with similar stand characteristics to ensure that the non-disturbed plots accurately represent pre-disturbance conditions.
Tables

Table 1: Parameters measured in each study site, sensor model, manufacturer and accuracy. Sensor profiles were identical in the LS and MPB plots, but the CC plot had a different sensor profile.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Sensor Model and Manufacturer</th>
<th>Accuracy</th>
</tr>
</thead>
<tbody>
<tr>
<td>Soil moisture</td>
<td>CS616 (Campbell Scientific)</td>
<td>3%</td>
</tr>
<tr>
<td>Temperature &amp; Relative Humidity</td>
<td>CS215 (Campbell Scientific)</td>
<td>±0.01° C; ±2% RH (0-90%); ±3% RH (&gt;90%)</td>
</tr>
<tr>
<td></td>
<td>THB-M002 (ONSET)(^1)</td>
<td>±0.130° C; ±2.5% RH (0-90%)</td>
</tr>
<tr>
<td>Wind Speed</td>
<td>014A (Campbell Scientific)</td>
<td>±0.01 m s(^{-1})</td>
</tr>
<tr>
<td></td>
<td>WSA-M003 (ONSET)(^1)</td>
<td>±1.1 m s(^{-1})</td>
</tr>
<tr>
<td>Rainfall</td>
<td>RGB-M002 (ONSET)</td>
<td>±0.2 mm</td>
</tr>
<tr>
<td>Net Radiation</td>
<td>CNR2 (Campbell Scientific)</td>
<td>&lt;10% of daily total</td>
</tr>
<tr>
<td></td>
<td>S-LIB-M003 (ONSET)(^1)</td>
<td>±5%</td>
</tr>
<tr>
<td>Soil moisture</td>
<td>CS616 (Campbell Scientific)</td>
<td>±3%</td>
</tr>
<tr>
<td></td>
<td>HydroSense 2 (Campbell Scientific)</td>
<td>±3%</td>
</tr>
<tr>
<td>Snowpack Temperature</td>
<td>DS1921G iButton (Embedded Systems)</td>
<td>±1° (-30° C to +70° C)</td>
</tr>
</tbody>
</table>

\(^1\): Sensors used in the CC plots to measure net short-wave radiation. For calculations of long-wave radiation, see Appendix A.

Table 2: Stand characteristics of live and dead trees in the MPB and LS plots.

<table>
<thead>
<tr>
<th>Study Plot</th>
<th>Trees per hectare</th>
<th>Basal Area (m²/ha)</th>
<th>Mean DBH(^1) (cm)</th>
<th>Mean Height(^1) (m)</th>
<th>LAI(^1)</th>
<th>Mortality(^2) (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MPB</td>
<td>956</td>
<td>63.6</td>
<td>28.1 (6.1)</td>
<td>22.2 (7.3)</td>
<td>1.6 (0.15)</td>
<td>83</td>
</tr>
<tr>
<td>LS</td>
<td>2,430</td>
<td>26.3</td>
<td>11.3 (3.0)</td>
<td>9.1 (1.5)</td>
<td>1.3 (0.19)</td>
<td>16</td>
</tr>
</tbody>
</table>

\(^1\): Standard deviation in parentheses  \(^2\): Ratio of dead to total.

Table 3: Landscape metrics (and standard deviations) observed & modeled with SAGA GIS software using a 1-m digital elevation model.

<table>
<thead>
<tr>
<th>Study Plot</th>
<th>Elevation (m)</th>
<th>Aspect</th>
<th>Slope (%)</th>
<th>TWI</th>
<th>PISR (kwh m(^{-2}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>MPB</td>
<td>1857</td>
<td>SE</td>
<td>10</td>
<td>3.6 (2.1)</td>
<td>1115.4 (47)</td>
</tr>
<tr>
<td>LS</td>
<td>1898</td>
<td>SE</td>
<td>7</td>
<td>3.7 (1.5)</td>
<td>1097.1 (30)</td>
</tr>
<tr>
<td>CC</td>
<td>1862</td>
<td>W</td>
<td>17</td>
<td>4.6 (2.5)</td>
<td>1094.9 (42)</td>
</tr>
</tbody>
</table>

\(^1\): Standard deviation in parentheses
Table 4: Daily averages (and standard deviations) during the snow accumulation and ablation period (March 24, 2013 – May 15, 2013).

<table>
<thead>
<tr>
<th>Study Plot</th>
<th>Air Temperature (°C)</th>
<th>Relative Humidity (%)</th>
<th>Wind Speed (m s(^{-1}))</th>
<th>Net Radiation (W m(^{-2}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>MPB</td>
<td>1.13 (6.42)</td>
<td>68.39 (11.31)</td>
<td>0.92 (0.24)</td>
<td>8.26 (8.04)</td>
</tr>
<tr>
<td>LS</td>
<td>1.47 (6.55)</td>
<td>68.05 (13.31)</td>
<td>1.04 (0.37)</td>
<td>6.74 (7.06)</td>
</tr>
<tr>
<td>CC</td>
<td>2.21 (6.51)</td>
<td>67.50 (15.71)</td>
<td>1.93 (1.22)</td>
<td>17.88 (58.23)</td>
</tr>
</tbody>
</table>

Table 5: Average daily modeled and measured snow ablation rates (cm d\(^{-1}\)). Ablation rates were calculated as the change in SWE over the change in time.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Mod</td>
<td>Measured</td>
<td>Mod</td>
</tr>
<tr>
<td>MPB</td>
<td>0.00</td>
<td>0.19</td>
<td>0.31</td>
</tr>
<tr>
<td>LS</td>
<td>0.13</td>
<td>0.27</td>
<td>0.31</td>
</tr>
<tr>
<td>CC</td>
<td>0.63</td>
<td>0.48</td>
<td>NA</td>
</tr>
</tbody>
</table>

Table 6: Modeled and measured snow water equivalent (SWE, cm).

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Model</td>
<td>Measured(^1)</td>
<td>Model</td>
</tr>
<tr>
<td>MPB</td>
<td>17.00</td>
<td>17.00</td>
<td>16.90</td>
</tr>
<tr>
<td>LS</td>
<td>16.00</td>
<td>16.00</td>
<td>13.00</td>
</tr>
<tr>
<td>CC</td>
<td>15.00</td>
<td>15.00</td>
<td>0.00</td>
</tr>
</tbody>
</table>

\(^1\)Plot average of SWE measurements (n=36)

Table 7: Daily averages (and standard deviations) during the post-snowmelt study period (May 22, 2013 – July 7, 2013).

<table>
<thead>
<tr>
<th>Study Plot</th>
<th>Air Temperature (°C)</th>
<th>Relative Humidity (%)</th>
<th>Wind Speed (m s(^{-1}))</th>
<th>Net Radiation (W m(^{-2}))</th>
<th>Precipitation (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MPB</td>
<td>10.26 (5.75)</td>
<td>67.90 (12.81)</td>
<td>0.81 (0.17)</td>
<td>18.46 (8.08)</td>
<td>99.6</td>
</tr>
<tr>
<td>LS</td>
<td>10.79 (6.09)</td>
<td>64.82 (12.36)</td>
<td>0.93 (0.32)</td>
<td>15.76 (5.84)</td>
<td>89</td>
</tr>
<tr>
<td>CC</td>
<td>11.12 (5.96)</td>
<td>65.18 (17.28)</td>
<td>1.43 (0.79)</td>
<td>57.08 (52.40)</td>
<td>107.6</td>
</tr>
</tbody>
</table>
Table 8: Water balances for study plots between May 22, 2013 and July 7, 2013. \( \Delta \text{SWS} \) for the LS plot included water lost through canopy transpiration.

<table>
<thead>
<tr>
<th>Stand</th>
<th>Precipitation (mm)</th>
<th>Understory ET (mm)</th>
<th>Canopy Transpiration (mm)</th>
<th>( \Delta \text{SWS} ) (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MPB</td>
<td>99.6</td>
<td>129.6</td>
<td>0</td>
<td>-30</td>
</tr>
<tr>
<td>LS</td>
<td>89.0</td>
<td>110.0</td>
<td>28.6</td>
<td>-49.6</td>
</tr>
<tr>
<td>CC</td>
<td>107.6</td>
<td>126.6</td>
<td>0</td>
<td>-19</td>
</tr>
</tbody>
</table>

Table 9: Sum of reference evapotranspiration (mm) and measured understory evapotranspiration (mm) during the post-snowmelt study period (May 22, 2013 – July 7, 2013).

<table>
<thead>
<tr>
<th></th>
<th>Penman-Monteith ((ET_0))</th>
<th>Measured (ET_u)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MPB</td>
<td>46.9</td>
<td>129.6</td>
</tr>
<tr>
<td>LS</td>
<td>52.4</td>
<td>110.0</td>
</tr>
<tr>
<td>CC</td>
<td>88.9</td>
<td>126.6</td>
</tr>
</tbody>
</table>

Table 10: ANOVA table for factors influencing the spatial distribution of soil moisture between May 22, 2013 and July 7, 2013.

<table>
<thead>
<tr>
<th>Factors</th>
<th>DF</th>
<th>Sum of Squares</th>
<th>F-value</th>
<th>P-value</th>
</tr>
</thead>
<tbody>
<tr>
<td>TWI</td>
<td>1</td>
<td>204.4</td>
<td>7.2</td>
<td>0.008</td>
</tr>
<tr>
<td>Sample Date</td>
<td>6</td>
<td>9156.7</td>
<td>53.7</td>
<td>&lt; 0.001</td>
</tr>
<tr>
<td>PIR</td>
<td>1</td>
<td>293.9</td>
<td>10.3</td>
<td>0.001</td>
</tr>
<tr>
<td>LAI x Stand</td>
<td>2</td>
<td>22451.5</td>
<td>394.7</td>
<td>&lt; 0.001</td>
</tr>
<tr>
<td>Residuals</td>
<td>744</td>
<td>21160</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figures

Figure 1: Site map of study plot. Inset map shows location of study plots in reference to Lubrecht Experimental Forest.
Figure 2: Micrometeorology (daily averages) during the snow accumulation and melt period (March 24, 2013 to May 15, 2013).
Figure 3: Distribution of SWE (cm) measured during snow surveys (n=36 for each plot). Bars and whiskers represent minimuma and mixima value, black bars represents median values.
Figure 4: Snowpack temperature and ambient air temperature. The vertical dotted lines represent the estimated timing of complete snowmelt based upon temperature sensor readings.
Figure 5: Modeled melt of snowpack between March 24, 2013 and May 15, 2013. Snowpack was converted to snow water equivalent (SWE, cm). The dotted blue line is SWE recorded at the North Fork Elk Creek SNOTEL station. Triangles correspond date of complete snowmelt estimated from the iButton sensors. The circle denotes the date of peak θ in the CC plot. In the LS and MPB plots, peak θ occurred on the same dates as estimated by the iButton sensors.
Figure 6: Modeled and measured SWE. Boxplots show distribution of SWE measured during snow surveys on April 17, 2013 and May 3, 2013.
Figure 7: Modeled SWE and volumetric soil moisture ($\theta$, m$^3$ m$^{-3}$). Orange vertical lines represent the timing of peak $\theta$ during the snowmelt period.
Figure 8: Micrometeorology (daily averages) during the post-snowmelt period (May 22, 2013 – July 7, 2013).

Post-Snowmelt Daily Averages (May 22 - July 07, 2013)

- Temp °C
- Relative Humidity %
- Wind Speed (m s⁻¹)
- Net Radiation (W m⁻²)
Figure 9: Post-snowmelt water balances for the LS plot (a), MPB plot (b) and CC plot (c). Canopy transpiration was not measured in the MPB and CC plots so no transpiration is presented for these plots.
Figure 10: Spatially distributed soil moisture measurements (0-12 cm)
Literature Cited


Appendix A: Estimating Long-wave Radiation and Reference Understory Evapotranspiration in the CC plot

In 2012 an Onset weather station was installed in the CC plot to replace a suite of older Campbell Scientific instruments. Two Onset pyranometers were installed to measure incoming and outgoing solar radiation. However, these instruments only measured short-wave radiation, not net radiation. As a result, long-wave radiation had to be calculated empirically (Dingman 2002). Additionally, the Penman-Monteith equation (equation 10) had to be modified to accommodate the long-wave radiation estimates.

Incoming long-wave radiation was calculated as

\[ LW_{in} = \varepsilon_{at} \times \sigma \times (T_a + 273.2)^4 \]

where \( \varepsilon_{at} \) was the emissivity of the atmosphere, \( \sigma \) was the Stefan-Boltzmann constant \( (4.90 \times 10^{-9} \text{ MJ m}^{-2} \text{ day}^{-1} \text{ K}^{-1}) \), and \( T_a \) was air temperature \( (^\circ \text{C}) \).

Atmospheric emissivity was estimated using clear sky conditions with the following equation:

\[ \varepsilon_{at} = 1.72 \times \left( \frac{e_a}{T_a+273.2} \right)^{\frac{4}{7}} \]

where \( e_a \) was the atmospheric vapor pressure \( (\text{kPa}) \) and \( T_a \) was the air temperature.

Outgoing long-wave radiation was calculated two different ways depending on whether the ground was expected to be snow covered or bare ground. The presences or absence of snow cover affects the emissivity and
temperature of ground surface, which affects the magnitude of long-wave
radiation emitted. The CC plot was expected to be snow covered for the
snowmelt period (March 24 – May 15) and the following equations were used to
estimate snow surface temperature and $LW_{\text{out}}$:

$$LW_{\text{out}} = \varepsilon_{\text{ss}} \cdot \sigma \cdot (T_{\text{ss}} + 273.2)^4$$

where $\varepsilon_{\text{ss}}$ was the emissivity of the snow surface, which is very near 1.0
(Dingman 2002). $T_{\text{ss}}$ was the snow surface temperature and estimated with the
following equation:

$$T_{\text{ss}} = \min[T_a - 2.5, 0]$$

Soil surface temperature was not measured, so the following equation was used
to estimate $LW_{\text{out}}$ when the ground was not snow covered:

$$LW_{\text{out}} = \varepsilon_f \cdot \sigma \cdot (T_a + 273.2)^4$$

where $\varepsilon_f$ is the emissivity of a typical field (0.95) (Dingman 2002), and the rest
of the terms are as previously described.

To calculate $ET_0$ in the CC plot, net long-wave radiation was calculated as

$$LW_{\text{net}} = \varepsilon_f \cdot LW_{\text{in}} - LW_{\text{out}}$$

The psychrometric constant ($\gamma$) in the Penman-Monteith equation (equation 8)
was replaced with the following term

$$\gamma' = \gamma + \left( \frac{4 \cdot \varepsilon_f \cdot \sigma \cdot (T_a + 273.2)^3}{K_E \cdot \rho_w \cdot \lambda \cdot \sigma \cdot v_a} \right)$$
where $K_E$ is the coefficient of turbulent transfer efficiency and all the other terms have been previously described. $K_E$ was calculated as

$$K_E = \left(\frac{0.622 \cdot \rho_a}{\rho_w \cdot P_*}\right) \cdot \frac{k^2}{\ln\left(\frac{z_u-z_d}{z_0}\right)}$$

These terms were previously defined: $k$ is von Karman’s constant (0.40), $z_u$ is the wind speed sensor height, $z_d$ is the zero-plane displacement height, and $z_0$ is the surface roughness height. These terms replaced net long-wave radiation and the psychrometric constant values in the Penman-Monteith equation (equation 8) to calculate ET$_0$ for the CC plot.