12-2009

A Comparison of Soil Moisture and Hillslope-Stream Connectivity Between Aspen and Conifer-Dominated Hillslopes of a First Order Catchment in Northern Utah

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Recommended Citation

A COMPARISON OF SOIL MOISTURE AND HILLSLOPE-STREAM

CONNECTIVITY BETWEEN ASPEN AND CONIFER DOMINATED HILLSLOPES

OF A FIRST ORDER CATCHMENT IN NORTHERN UTAH

by

Amy Burke

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

of

MASTER OF SCIENCE

in

Watershed Science

Approved

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UTAH STATE UNIVERSITY
Logan, Utah

2009

ABSTRACT

A Comparison of Soil Moisture and Hillslope-Stream Connectivity Between Aspen and

Conifer-Dominated Hillslopes of a First Order Catchment in Northern Utah

by

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Utah State University, 2009

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Mountain headwater catchments in the semi-arid Intermountain West are important sources of surface water because these high elevations receive more precipitation than neighboring lowlands. The hydrology of these mountain catchments is especially important as the region faces water shortages and conflicts. Conifer encroachment on aspen stands has been observed across the western US and can result in a decline in water yield. The overall objective of this study was to further our understanding of hillslope-stream connectivity in a headwater catchment of Northern Utah and any observable differences in this connection between aspen and conifer hillslopes. Hillslopes are the fundamental unit of a watershed. Therefore understanding processes at the hillslope scale is pertinent to managing valuable water resources. However, hillslope hydrology is understudied in the snow-driven, semi-arid west, leaving a gap in our knowledge of how watersheds function. This thesis focuses on how and

when hillslope water contributes to stream water: hillslope-stream connectivity. Its specific objectives are (1) to compare peak snow accumulation under aspen and conifer stands, (2) to determine if shallow soil moisture shows organized patterns, indicating hillslope-connectivity and compare these patterns between vegetation types, (3) to examine hillslope-stream connectivity within deep layers of the soil profile and compare times of connectivity between vegetation types and (4) to find any thresholds past which hillslope-stream connectivity begins.

(125 pages)

ACKNOWLEDGMENTS

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CHAPTER 1

INTRODUCTION

Vegetation and the water balance are undeniably linked, especially in forests dominated by snow, since forest canopies affect snow deposition and melt (Marsh, 1999). Woody vegetation also affects soil water movement, through transpiration and by influencing infiltration, percolation, retention and water repellency properties (Williams et al., 2003). The mechanisms of streamflow generation must be known for the role of woody plants to be understood (Huxman et al., 2005). The field of ecohydrology was born out of the recognition that to solve real world problems, a clear understanding of biological and hydrological systems interact is needed (Bond et al., 2007).

Understanding how biologic and hydrologic components affect one another will help forecast the nature and magnitude of environmental changes and natural resources (Newman et al., 2006). Just as vegetation affects the hydrological cycle, changes in vegetation cover will alter the interaction between biological and hydrological components of an ecosystem. In Northern Utah there is a concern that a shift from aspen to conifer-dominated forests is causing a decline in surface water yield. While the ecological importance of aspen is quite clear (Shepperd et al., 2006), there is a lack of concrete supporting evidence from the field that conifer invasion into aspen stands decreases surface water yield (LaMalfa and Ryel, 2008). In order to fully understand vegetation and water yield, we must remember that annual streamflow and evapotranspiration do not tell the whole story because of seasonal interactions of factors affecting the water balance such as soil moisture (Brown et al., 2006).

Hillslopes are the fundamental unit of landscapes and the building blocks for catchment models (Lin et al., 2006; Tromp-van Meerveld and McDonnell, 2006b). Working at the hillslope scale guarantees relatively uniform climate and soil type (Tromp-van Meerveld and McDonnell, 2006b). Hillslope hydrology helps connect point observations to hillslope phenomena and then to the whole catchment processes (Lin et al., 2006). Ideally hillslope analysis should include integrated (trenchflow) and point scale (soil moisture) measurements coupled with an understanding of surface and subsurface features. However, field hydrology will always be limited by measurement techniques and site choices (Freer et al., 1997). Blume et al. (2007) found that high temporal resolution data is a cost effective means of measuring hydrology in mountain catchments.

This study examined subsurface lateral flow from two hillslopes, one aspen the other conifer, in a first order catchment of Northern Utah. Each vegetation type sits on a distinct soil profile, aspen over a deep mollic epipedon and conifer over a shallow rocky entisol. Because vegetation effects cannot be uncoupled from soil effects, the two hillslopes were considered as two endpoint communities. The study focused on lateral flow through shallow soil layers because this pathway flows through the root zone is therefore affected by vegetation. I measured snow water equivalents, trenchflow, and streamflow and soil moisture in both space and time. The first objective, addressed in chapter 1, was to determine whether shallow soil moisture reflects hillslope connectivity to the stream and if there is any difference between vegetation types. Differences in water inputs between the vegetation types are also considered. The second objective, addressed in chapter 2, was to look deeper into the soil profiles, using soil moisture

profiles as well as hillslope trenches, to decide when hillslope-stream connectivity is

established and if it differs between the aspen and conifer hillsides.

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CHAPTER 2

SHALLOW SOIL MOISTURE PATTERNS AND HILLSLOPE-STREAM

CONNECTIVITY IN A MONTAIN CATCHMENT OF NORTHERN UTAH

ABSTRACT

Conifer invasion has been considered a significant factor in the declining extent of aspen cover in Utah. A consequence of this change in vegetation type may be lower water yields from mountain headwater catchments in the region. This study examined two hillslopes, one aspen dominated, and the other conifer, adjacent to a stream in a headwater catchment of the Ogden River basin. Snow accumulation, streamflow and shallow soil moisture patterns were compared between the conifer and aspen dominated hillslopes to understand the differences in hillslope-stream connectivity in the top 12 cm of soil. Snow water equivalent was measured in the catchment, at peak snow accumulation in 2007 and 2008, under both vegetation types. Below average precipitation was observed in 2007 and an average amount in 2008. Shallow soil moisture was measured in a grid of 90 points (10 by 20 m) three times during the spring of 2008. Geostatistical analysis was used to determine if soil moisture showed spatial organization indicating hillslope connectivity. Streamflow was monitored at three points in the catchment, in 2008, in order to isolate the contribution of streamflow from the aspen and conifer dominated hillslopes. To capture temporal patterns in soil moisture, water content reflectometry probes were installed on both hillslopes. Transects of eight sensors, buried at 5 cm, extended uphill from the stream on each hillslope. Snow surveys revealed an average 11 and 23% greater snow water equivalents under aspen stands

compared to conifer in 2007 and 2008, respectively. Overland flow was observed but not quantified during the spring snow melt of 2008. Semi-variograms of soil moisture showed hillslope scale organization in the aspen hillslope only in June. No hillslope-stream connectivity was detected in the top 12 cm of soil in either hillslope. However, plots of soil moisture and streamflow show a threshold relationship between soil moisture and streamflow, indicating hillslope-stream connectivity, perhaps at deeper layers within the soil profile.

INTRODUCTION

Mountain headwater catchments in the semi-arid Intermountain West are important sources of surface water because these high elevations receive more precipitation than neighboring lowlands (Grant et al., 2004). The hydrology of these catchments is especially important as the region faces water shortages and conflicts. Snow melt typically makes up 80% or more of streamflow in the semi-arid West and snow deposition and melt in montane catchments are largely a function of topography and vegetation (Marks et al., 2002). The snow-driven system of the semi-arid Intermountain West is understudied, leaving a gap in our knowledge of how watersheds function and how runoff is produced (Grant et al., 2004). There is a lack of information on soil moisture patterns in semi-arid catchments, yet the consequences of runoff and potential erosion are even greater in these regions (Williams et al., 2003).

In locations across the Intermountain West conifer species are encroaching on aspen stands (Bartos and Campbell, 1998). Possible reasons for this encroachment include human alterations in disturbance regimes, such as fire suppression and grazing, as

well as climate change (Bartos and Campbell, 1998; Brown et al., 2006; Shepperd et al., 2006). Utah holds 25% of the aspen in the western United States but it was reported that, since European settlement, there has been at least a 59% decline in aspen domination in Utah and a 66% decline in the Wasatch Cache National Forest (Bartos, 2001). Though the exact mechanisms are not known and appear to be site specific, conifer encroachment into aspen stands has been blamed for declining water yield and water quality (Shepperd et al., 2006). Bartos and Campbell (1998) claim that for every 1,000 acres of aspen that are replaced by conifers, an estimated 250 to 500 acre feet of water is lost as transpiration.

Soil moisture plays an important role in various hydrologic processes at multiple scales, ranging from event to long term scale (Grayson et al., 1997). It affects the partitioning of precipitation into infiltration and runoff (Grayson et al., 1997; Hupet and Vanclooster, 2002). It has control on runoff generation (O'Loughlin, 1986; Fitzjohn et al., 1998; Grant et al., 2004; Blume et al., 2007), erosion (Grayson et al., 1997; Fitzjohn et al., 1998; Western et al., 2004), flooding (Western et al., 2004), solute transport (Grayson et al., 1997; Western et al., 2004), water availability for plants (Grayson et al., 1997; Kelleners et al., 2005) and groundwater recharge (Grant et al., 2004). Soil moisture also influences the partitioning of net radiation into sensible and latent heat, linking the surface energy balance to the water balance (Grayson et al., 1997; Hupet and Vanclooster, 2002; Western et al., 2004). An understanding of small scale soil moisture patterns can lead to a greater understanding of how integrated catchments will behave and how to best manage water resources (Western et al., 1999; Williams et al., 2003). Despite its importance, there is a lack of field data on soil moisture patterns (Grayson et

al., 1997). This study focuses on spatial and temporal soil moisture patterns under aspen and conifer stands in a montane catchment in the Intermountain West and links these patterns to streamflow fluctuations.

Objectives and hypotheses

The objectives of this study were to compare surface soil moisture patterns between aspen and conifer hillslope and to determine if soil moisture shows an organized pattern, in relation to vegetation type, hillslope position, and surface topography. More specifically to:

1. Examine and compare hillslope soil moisture patterns between vegetation types.

2. Find any thresholds after which hillslope-stream connectivity begins and compare between soil/vegetation types.

3. Find any seasonal patterns of hillslope-stream connectivity and compare between vegetation/soil types.

I hypothesize that:

1. Soil moisture patterns will remain in a wet, organized state longer in the aspen hillslope due to greater snow melt inputs.

2. Hillslope-stream connectivity will occur after a soil moisture content threshold. This connection will last longer in the aspen hillslope due to greater snow melt inputs.

3. Subsurface lateral flow will connect the stream and hillslope only during the wettest periods of the spring snow melt while for the remainder of the year the hillslope and stream will remain disconnected and streamflow will not react to surface water inputs.

The influence of vegetation on soil moisture

Vegetation influences the hydrologic cycle in multiple ways: by affecting soil properties and ground cover and therefore influencing infiltration capacity (Schume et al., 2003; Williams et al., 2003), by creating macro-pores and altering subsurface flowpaths (Williams et al., 2003), through transpiration (Williams et al., 2003; Huxman et al., 2005) and by holding precipitation in the canopy and altering throughfall patterns (Schume et al., 2003; Western et al., 2004; Huxman et al., 2005). In water limited systems, the type and pattern of woody vegetation affects stream and groundwater recharge, nutrient cycles and biodiversity (Newman et al., 2006). Vegetation is, in turn, affected by soil properties and climate. By measuring sap flow, soil moisture, surface and subsurface topography at the Panola Mountain Research Watershed in southeast Georgia, Tromp-van Meerveld and McDonnell (2006a) noted that vegetation had a large impact on soil moisture and that early season transpiration dictated patterns of soil water depletion, which controlled patterns of transpiration later in the season; a feedback loop between soil moisture and vegetation.

Soils under aspen stands are often significantly different than those under adjacent forests that have succeeded to conifer species. One notable difference found in the Sierra Nevada and Intermountain West was that soils under aspen stands had higher water holding capacity due to increased organic matter inputs from the annual leaf litter. Where conifers invaded aspen stands, needle litter altered soil properties over time and that soil was ultimately reflective of the vegetation that had occupied the site longest (Shepperd et al., 2006). In Western Colorado, Cryer and Murray (1992) monitored soils under aspen stands for 10 to 12 years and found that heavy leaf fall in aspen stands

maintained a mollic horizon until the trees deteriorated due to a lack of disturbance. Mollic horizons were then replaced with albic (leached) horizons and pH was reduced. In sites of conifer invasion into aspen stands the speed of this transition to a leached acidic profile was greatly increased.

Vegetation also affects hydrologic processes by altering soil properties. Schume et al. (2003) describe a feedback loop between tree species, in which trees affect soil properties including litter, structure and infiltration. Soil properties determine rooting depth and therefore the soil volume from which water is extracted for transpiration. In California and Utah, aspen stands had nearly 100% cover (no bare soil) between the canopy, thick under-story and leaf litter, greatly reducing chances of overland flow (Shepperd et al., 2006), which can be an important contributor to streamflow in water-limited systems (Newman et al., 2006).

Vegetation can also greatly alter water movement through soil by creating macropores, usually formed by dead or live roots. Bypass flow is the vertical movement of water by way of continuous pores through otherwise unsaturated or partially saturated soil (McDonnell, 1990). At three catchments in Japan, there was significant flow on steep wet hillslopes through discrete soil pipes (soil macropores that conduct water at a higher rate than the surrounding soil matrix), most prevalent at the soil bedrock interface or at impeding layers (Uchida et al., 2004). Field investigations in the Hitachi Ohta, Japan, indicated that the junctions, or nodes, where macropores connect to one another are the true limiting factors in preferential flow systems (Sidle and Noguchi, 2001). Flow of water through macropores, bypassing unsaturated soil, also occurred through root

derived macropores in the upper soil horizons of a semi-arid ponderosa pine hillslope

in New Mexico (Newman et al., 2004).

Many factors control snow retention and loss by sublimation in snow-driven

systems, including weather patterns, terrain and vegetation cover. Forest canopy

characteristics are dominant factors affecting throughfall patterns and snow accumulation

in mountain headwater catchments. Snow in exposed regions of the semi-arid Owyhee

mountains was more vulnerable to turbulent air currents (the primary source of melt

energy during the winter) as well as solar radiation (the primary source of melt energy in

the spring), melted sooner and provided less surface water inputs overall. Drift and

shelter areas provided much more water inputs and snow there lasted later in the season

(Marks et al., 2002). Hiemstra et al. (2002) also found that trees helped retain snow in

the high elevation tree line zone (3100-3300m) of the Medicine Bow Mountains in

Wyoming, where snow makes up 50 to 80% of precipitation. However different canopy

types, as in the case of comparing aspen to conifer stands, may also differ in their ability

to retain precipitation. In a rain-dominated, mixed stand in Southern Austria, interception

loss by the coniferous spruce canopy was much greater than that of the deciduous beech

canopy, causing uneven recharge patterns (Schume et al., 2003).

Previous investigations in Northern Utah revealed that snow accumulation was

greater under aspen than conifer stands (LaMalfa and Ryel, 2008). Conifer species retain

their needles year round and thus are able to accumulate more snow in their canopy,

lessening the snowpack on the ground around them. Snow held in the canopy is more

susceptible to sublimation; a water loss to the system, or simply redistribution, possibly

to nearby aspen stands. Experiments using simulated conifer branches showed that

sublimation rates from sunlit branches were double those from shaded branches and the ground surface and that sublimation could explain most or all of the differences in SWE under aspen and conifer stands (LaMalfa et al., 2007). LaMalfa and Ryel (2008) also reported that aspen stands had a greater potential for generating runoff and groundwater recharge than conifer in Northern Utah.

The majority of water uptake by vegetation is performed by fine, young roots in the upper soil horizons where nutrients are more abundant, however, roots can also be found in deep layers, even in bedrock, especially in water limited areas (Bond et al., 2007). Patterns of water depletion from the soil profile indicate water uptake and root activity (Caldwell and Virginia, 1989). Schume et al. (2003) found that under vegetation, species-specific rooting depth and transpiration were the main source of soil moisture variation and that rooting depth decides the volume of soil from which trees access for water uptake. Reproduction of aspen is primarily asexual, by means of root sprouts (Bartos, 2001). These lateral roots are normally found within a foot of the soil surface and can be inhibited by subsurface rocks (Shepperd et al., 2006). In contrast, conifer species have deeper roots, able to reach lower sources of groundwater compared to the relatively shallow rooted aspen. Average maximum depth for temperate conifer forests was 3.9 m and only 2.9 m for temperate deciduous forests (Canadell et al., 1996).

It is common for evapotranspiration to have direct impacts on streamflow or groundwater levels. In a humid, Pacific Northwest fir catchment, Bond et al. (2002) demonstrated the link between streamflow and vegetation by comparing diel patterns of transpiration and streamflow. Along the arid Rio Grande River, diel fluctuations in evapotranspiration from riparian vegetation resulted in daily responses in groundwater

levels during the growing season (Dahm et al., 2002). According to Huxman et al. (2005), the effect of woody plants on streamflow generation depends on whether the woody vegetation is found in riparian or hillslope areas, whether surface water inputs are supplied during the growing season and whether or not there is a potential for bypass flow such as deep sandy soils, or fractured or Karst bedrock. The more dominant subsurface runoff paths, the greater the expected impacts on streamflow by woody plant encroachment.

Aspen and conifer have differences in water use through transpiration; therefore their effects on streamflow may be different. Part of the competitive strategy of evergreen trees is to start significant water uptake as soon as climatic conditions allow (Schume et al., 2003). This coincides with LaMalfa's (2007) findings that conifer stands in Northern Utah had longer transpiration activity periods relative to aspen, starting 5 days before aspen and ending 24 to 31 days after aspen.

Though Bartos and Campbell (1998) estimate great water yield losses associated with conifer replacement of aspens, the longer transpiration period in conifers seems to be offset by higher transpiration rates during leaf expansion of aspen or deciduous species. Shume et al. (2003) found that beech trees used more water than nearby spruce, which were able to conserve water by closing their stomata during dry times. Aspen showed less of an ability to regulate water loss through stomatal closure than neighboring pine and fir species in Wyoming, where aspen sap flux increased during the early season leaf-out period (Pataki et al., 2000). In Northern Utah, aspen evapotranspiration was 25% higher than conifer during leaf out, most likely due to more water in aspen soil profile (LaMalfa, 2007). Both Pataki et al. (2000) and LaMalfa (2007) found that soil

moisture and water movements to both species regulate transpiration. By the end of the water year both aspen and conifer had used all the profile water available in their soil profiles (LaMalfa, 2007).

Soil moisture grids

Working in the temperate, semi-arid, rain dominated Wagga Wagga and Tarrawarra catchments of Australia, Grayson et al. (1997) identified two states in soil moisture, a dry state and a wet state. In the dry state evapotranspiration exceeded surface water inputs and soil moisture was controlled by local factors, such as vegetation, local topography, and infiltration capacity. No hillslope scale pattern of soil moisture was seen while soil was in the dry state. In a wet state, water inputs exceeded evapotranspiration and saturated zones expanded to meet one another, hydraulically connecting the hillslope. In a wet state, non-local factors controlled soil moisture, such as hillslope scale topography and water inputs. Patterns of soil moisture form as soil moisture content increased with upslope contributing area (Grayson et al., 1997). Fitzjohn et al. (1998) noted that the dry state in soil moisture dominates in semi-arid regions and local controls create random patterns of soil moisture. The mosaic of different levels of soil moisture demonstrates a mosaic of the variables that control soil moisture including: organic matter, topography, soil texture and vegetation. In a snow dominated, semi-arid catchment in Idaho, McNamara et al. (2005) used data from soil moisture profiles, stream and hillslope modeling to show that, in the transition from summer to fall, a balance of evapotranspiration and precipitation controlled the switch between soil moisture states.

Semi-variograms describe spatial organization, indicating a wet soil moisture state or even hillslope-stream connectivity. As soil moisture increases, saturated zones grow to meet one another connecting the hillslope hydrologically. This is usually reflected in an increase in range (or correlation length) of the semi-variogram. In an arid catchment of central Spain, an increase (double or more) in correlation length in soil moisture indicated spatial continuity and greater hillslope connectivity (Fitzjohn et al., 1998). However, in the humid Panola catchment of Georgia, wet and dry periods saw no spatial organization of soil moisture, only transitional states, between 45 and 70% saturation, showed organization (longer correlation lengths) (Tromp-van Meerveld and McDonnell, 2006a). Shume et al. (2003) used distinct changes in spatial correlation to decide when spatial continuity of soil moisture was found and what factors controlled it, in a beech and spruce forested catchment in Southern Austria.

The transition between the dry and wet states of soil moisture has been described as abrupt, indicating a threshold being exceeded. When enough isolated areas have saturated to the point of producing runoff, a wetness threshold value is reached and widespread runoff can occur (Fitzjohn et al., 1998). There was a precipitation threshold of 55 mm to start hillslope-stream connectivity in the Panola Mountain Watershed in Georgia (Tromp-van Meerveld and McDonnell, 2006b). Research at Reynolds Creek Experimental Watershed in the Owyhee Mountains of Idaho showed a soil moisture content threshold (field capacity) at which soil switched to a wet state, the hillslope water reached the stream and stream discharge became highly reactive to soil water inputs (Grant et al., 2004). Finding a threshold that dictates at what point lateral flow begins, allows us to compare between hillslopes rather than gathering information on a single

hillslope whose behavior may not be representative of the area (Tromp-van Meerveld and McDonnell, 2006b).

Shallow measurements of soil moisture in a grid have been used to capture spatial patterns, including hydraulic connectivity within a hillslope or catchment. By focusing on shallow layers of soil, we are more likely to detect vegetation influences on soil water, since plant water uptake occurs mainly in the top layers where more water and nutrient rich organic matter is found. According to Western et al. (2004), topography determines surface and to a lesser extent subsurface flowpaths, as well as hydraulic gradients driving flow. Measuring shallow soil moisture is helpful because in many situations, runoff is often generated due to saturation of near surface horizons (Fitzjohn et al., 1998).

METHODS

Soil moisture patterns were compared between aspen and conifer hillslopes in a first order, montane catchment in Northern Utah. Grids of surface soil moisture measurements were made on each hillslope and analyzed using semi-variograms. Soil moisture at 6 fixed points and streamflow were continuously monitored to detect thresholds after which hillslope-stream connectivity begins. Water inputs to the catchment were quantified using precipitation data from the Natural Resource Conservation Service (NRCS) Snow Telemetry (SNOTEL) sites as well as snow surveys at peak snow accumulation. Infiltration and other soil physical properties were measured as possible explanatory variables for surface soil moisture patterns.

Site description

The study site is located in Northern Utah in the headwaters of the Ogden River along the eastern edge of Weber County (Figure 2.1). The study catchment lies roughly at UTMs 0461515, 4579126, inside a larger paired watershed study on the Deseret Land and Livestock Ranch. These mountain headwaters, around 2500 meters in elevation, feed Pineview and Causey reservoirs, which are used for recreation, water storage and irrigation. According to the SNOTEL data, the average annual temperature is 4.3 °C, with an average annual maximum temperature of 29.3 °C and an average annual minimum temperature of -21.0 °C. Average annual precipitation is 861 mm with approximately 74% falling as snow in the winter months.

The study catchment is located in Frost Canyon and contains both aspen and conifer hillslopes adjacent to a first order perennial stream (Figure 2.2). The catchment drains to the north with an area of approximately 1 km^2 and elevations ranging from 2464 to 2691 m. The upper hillslopes are dominated by mature Quaking aspen, (*Populus tremuloides*), with a lush under story made up primarily of Snowberry (*Symphoricarpos albus),* Serviceberry *(Amelanchier alnifolia),* Ragwort *(Senecio serra),* Wild Geranium *(Geranium richardsonii),* Mountain Bluebells *(Mertensia ciliata)* and Horse Mint *(Agastache urticifolia*). Average slope on the aspen hillside is 39%, with a confined V-shaped riparian area and a northwest aspect. Sagebrush grassland is found in a few patches of the upper hillslopes and on the lower eastern hillslope of the catchment. The lower western hillslopes are dominated by mature conifer species consisting mainly of White fir (*Albeis concolor*), Douglas fir (*Psuedodotsuga menziesii*), and Engelmann Spruce (*Picea, engelmannii*). The conifer under story is mainly sparse Goose Berry

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(*Ribes montigenum*), Snowberry (*Symphoricarpos albus)* and grasses. Average slope

on the conifer hillslope is 40%, with a northern aspect. The riparian zone is U-shaped

and less confined than the upstream aspen section. The aspen's under story is much

denser than the conifer's and varies yearly depending on water availability. It is common

to find different soil characteristics under different tree species, in part because of the

species ability to alter soil properties. Since the aspen and conifer stands at the study site

have very different soil profiles, this study addresses the combined influence of

vegetation and soil on subsurface flow. The study area lies between the Rich and Weber

County Soil Surveys and no soil survey information is available for the exact site.

According to the nearby Rich County Soil Survey of 1982, soils in similar areas range

from mollisols to entisols and inceptisols depending on slope and landscape position (Soil

Conservation Service, 1982).

Previous studies

The study catchment is contained within a larger paired watershed study on the

Deseret Land and Livestock Ranch. Bear and Frost canyons are the two small, adjacent,

paired watersheds with a combined area of 26.7 km^2. In 2005 and 2006, snow water

equivalent (SWE), snow density, sublimation, condensation, soil moisture and sap flux in

aspen and conifer stands on north facing slopes of Frost watershed were monitored

(LaMalfa, 2007). Though sublimation from the underlying snowpack was not

significantly different between the vegetation types, aspen plots had 34 to 44% higher
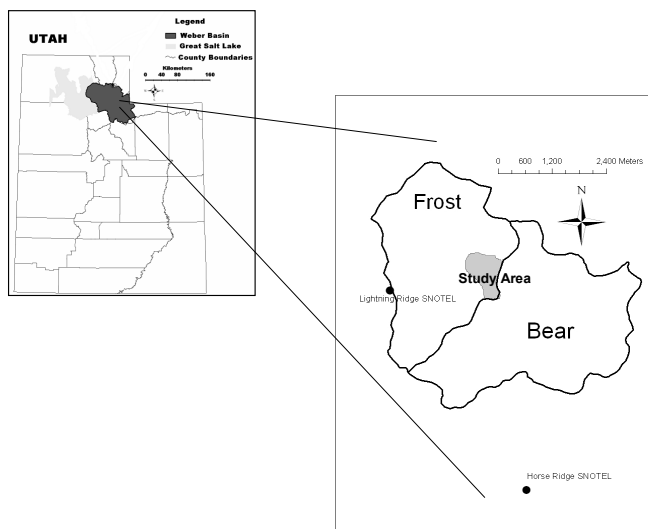
Figure 2.1. The study site within the paired watersheds in the Weber River Basin in Northern Utah.
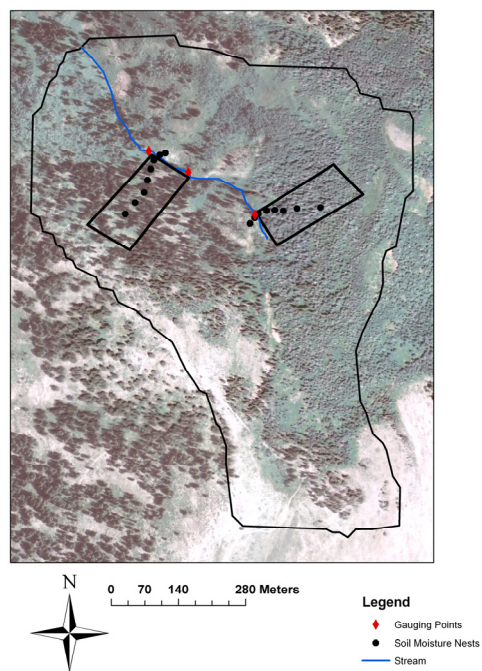
Figure 2.2. Map of the study site. Dots indicate transects of soil moisture sensor nests rectangles the area of soil moisture grid measurements, diamonds the gauging points, and the blue line the stream.

SWE at the time of peak snow accumulation than adjacent conifer plots. The greater

SWE in aspen stands, combined with longer transpiration periods in conifers, lead to 42

to 83% more potential runoff/groundwater recharge from aspen stands than from

neighboring conifer stands (LaMalfa, 2007). In Frost canyon, hillslope gradient was the

most important factor for flow gains and was negatively correlated. Steeper hillsides,

with more exposed bedrock and less ability to retain water, yielded less water during low

flows. Water chemistry suggested that the primary source of baseflow in Frost canyon is

deep groundwater and that Karst geology may contribute to heterogeneity of streamflow.

Flow at the confluence of Bear and Frost canyons showed diel patterns after snow melt

was complete, indicating vegetation influence on baseflow (Shakepeare, 2006). Other

complementary research is ongoing in the watersheds, including a study to estimate

residence times and flowpaths in Bear canyon. To better understand the effects of conifer

encroachment on water yield, Bear canyon will be treated by removal of conifers while

Frost canyon will be left as a control.

Instrumentation

Bear and Frost catchments have been instrumented with weather and stream

stations since 2004 and measurements are recorded each hour onto a data logger (CR10X,

Campbell Scientific, Logan, UT, USA). The streams are monitored just before their

confluence for stage (PDCR 1830-8388, Druck, Melrose, MA, USA), temperature and

electrical conductivity (CS547A, Campbell Scientific, Logan, UT, USA). Weather

stations at the top of Bear canyon and at the confluence of Bear and Frost monitor

relative humidity, air temperature, snow depth, precipitation, solar radiation, wind speed,

and soil temperature at 10, 20 cm. Data is also available from three nearby SNOTEL

sites. These three SNOTEL sites were chosen because of their proximity to the study

catchment and their similar elevation. Lightening ridge SNOTEL sits on the west ridge

of Frost canyon, at 2504 m above sea level and has been installed since 2004. Horse

ridge SNOTEL is located 2 km south of Bear canyon, at 2487 m above sea level and has

been collecting data since 1978. Dry Bread Pond SNOTEL site is 7.4 km southwest of

the study catchment, and 2545 m above sea level and has been recording data since 1978.

All three stations measure air temperature, snow depth, and precipitation every hour.

Water inputs

To quantify snow accumulation within the study catchment and its variance with

hillslope position and vegetation type, snow water equivalent (SWE) was measured in

three transects starting at the stream and moving uphill in four meter increments. Snow

surveys data were supplemented with recent surveys in nearby aspen, conifer and open

plots. SWE was measured as close to peak snow accumulation as possible, on March

16th, 2007 and April 6th through 12th, 2008. Standard snow sampling tubes (Carpenter

Machine and Supply, Seattle, WA, USA) were used to measure SWE and snow depth.

Tared snow tubes were inserted vertically into the snowpack until soil was reached.

Snow depth was recorded and the snow-filled tubes were weighed after any soil on the

bottom of the sample was removed. Daily snow pillow and rain gauge data were used

from the three nearby SNOTEL sites with similar elevation to the study site. Rain events

on days with an average temperature below 0 °C were considered invalid and

disregarded. Since snow is not an immediate water input to the soil, snow ablation and rain were combined to make surface water inputs (SWI).

Infiltration

Infiltration capacity in both the aspen and conifer forest floor was measured using a double ring infiltrometer on June 27th and 28th, 2007 and again on June 14th, 2008. The infiltrometer was inserted into the ground until water added inside the rings could not leak out and was forced to infiltrate into the soil (a depth of approximately 5 to 10 cm). Water levels in the outer ring were kept constant so that water infiltrating in the inner ring was maintained as a one-dimensional infiltration process. Water was added to the inner ring and the time the water took to drop 1 cm was recorded repeatedly until a constant rate was reached. Measurement locations were chosen as randomly as possible though relatively flat areas were required for the infiltrometers to be used. Two readings were taken in 2007 (four in aspen, two in conifer) and 10 were taken in 2008, five in each vegetation type. Cumulative infiltration was graphed against time. Infiltration capacities are considered the rate of infiltration once a constant rate had been reached.

Soil physical properties

Soil texture from 44 randomly chosen grid points (22 from each hillslope) was determined using hydrometers. Samples were sieved to 2 mm and 50 g of each was suspended in deionized water. The solutions were shaken with 50 ml of 10% Calgon for 12 hours. Deionized water was then added to each sample until there was a total volume of 1 liter in each graduated cylinder. A hydrometer was placed in the cylinder and

readings were taken after the sand has settled at 70 seconds and after the silt had settled after 7 hours.

To determine bulk density, between 150 and 600 cm^3 of soil were collected, dried and weighed. The subsequent holes were then filled with a known volume of sand. Rock content by volume was measured from 21 bulk density samples from multiple depths. To determine rock volume, particles greater than 2 mm from each sample were submerged in a partially filled graduated cylinder and the subsequent increase in volume was recorded.

Stream gauging

Stage was monitored continuously for the 2008 water year using WT-HR water level loggers (TruTrack, Christchurch, New Zealand), which were installed at three points in the stream (Figure 2.2). The first level logger measured the stage of the stream just below the contributing area dominated by aspen. The contributing area between the first and second level loggers is aspen/conifer mixed and encompassed a seep and a small tributary to the stream. The portion of the catchment between the second and third level logger is dominated by conifer species. The level loggers recorded hourly readings of water height in mm. Contributing areas to each gauging point were calculated using Arc GIS and Image J software.

The stream was gauged using slug injection dilution gauging. This approach is well suited for small streams that cannot be otherwise accurately gauged with flow meters (Moore, 2005). In this method, NaCl is used as a tracer and dumped as a slug injection into the stream. Downstream, an electrical conductivity (EC) probe measures

the pulse of salt as an increase in EC. The area under the break-through curve was calculated to estimate the flow in the stream.

An injectate solution was mixed using 300 to 500 g NaCl dissolved into 2 to 3 L stream water. More salt was used during high flows and less during baseflows. Small amounts of the injectate were added to 1 L of stream water and its electrical conductivity (EC) was measured at each known concentration (YSI 85, Yellow Springs, OH, USA). In this way a calibration curve was made to calculate concentration of NaCl from EC using a calibration constant K. The background EC was recorded both at the upstream and downstream reaches to be gauged before the slug of injectate was dumped. Each of the reaches gauged was long enough for adequate mixing of the salt-water slug with the stream. The slug was measured every two seconds as it passed at the downstream end of the reach by an EC probe (CS547A conductivity and temperature probe Campbell Scientific, Logan, UT, USA) and logged onto a data logger (CR 510, Campbell Scientific, Logan, UT, USA). The resulting curve of EC was translated into concentration at each time step using the calibration previously made.

$$C(t) = EC(t)*K \qquad (g\ L^{-1})$$

where C(t) is the concentration at time t, EC(t) is the electrical conductivity at time t and K is the calibration constant obtained from the calibration curve. The background concentration was accounted for and the mass of salt passing at each time step was calculated:

$$\text{Mass NaCl recovered (t)} = (C(t)-Cb)*d(t) \qquad (g\ L^{-1}s^{-1})$$

where Cb is the background concentration and d(t) is the time step (in this case two seconds). The total mass of NaCl recovered from all time steps was calculated:

Mass NaCl recovered = $\Sigma(C(t)-Cb)*d(t)$ $(g\ L^{-1}s^{-1})$

Flow (Q) was calculated by:

$Q = (L_{injectate}*Concentration_{injectate})/$Mass of NaCl recovered $(L\ s^{-1})$

Using the stage measurements from the times of stream gauging, a rating curve was made for each of the three stream gauging sites. A hydrograph was made using this rating curve and continuous stream stage measurements.

Continuous soil moisture

Soil moisture was continuously monitored at 16 locations in the study catchment (Figure 2.2). Nested soil moisture probes were installed in the summer of 2006. One transect of eight nests lies perpendicular to the stream in the aspen dominated stand and another in the conifer stand. Nests consist of three probes at 5, 20 and 100 cm or as deep as the probe could be installed (45 to 110 cm).

Grid measurements

During the spring snow melts of 2007 and 2008 volumetric soil water content was measured to a depth of 12 cm in grids, one on the conifer and aspen hillslope (Figure 2.2). All measurements were taken between 10 am and 8 pm, while soil temperatures were above freezing. Grid spacing was 20 meters in the uphill direction and 10 meters in the upstream direction, with a total of 90-100 points measured on each hillslope. Grids were laid out using a meter tape and compass with each point being marked and surveyed with a total station (TopCon GTS-226, Monsen Engineering, Salt Lake City, UT, USA). On May 15th and 16th of 2007, three measurements were taken at each point with Time Domain Reflectometry (TDR) (TDR100, Campbell Scientific, Logan, UT, USA). The

average of the three measurements was used as the soil water content at each point. The grids were measured again June 23rd through Jun 26th, 2007 but the dry soil made inserting the probe very difficult and only one measurement was taken per point. In the rockier conifer soil, in June the probe could not be inserted and a soil sample was taken for gravimetric water content. Volumetric water content was later calculated from gravimetric water content by multiplying by the bulk density of the soil.

In the spring of 2008 new grids were lain since the previous points were under snow and too difficult to find. The grids were measured on May 19th and 20th, June 14th and July 10th. Each grid point was measured three times with a two-pronged, 12 cm time reflectometer (CS616 Campbell Scientific, Logan, UT, USA) instead of the three pronged TDR used in 2007. CS616 sensors are transmission line oscillators, which operate similarly to time domain reflectometers but do not have a separate pulse and sampling units. Instead, consecutive voltage pulses are generated inside the probe head, with each pulse being triggered by the arrival of the reflected previous pulse. CS616 sensors function best with calibration curve for each soil being measured (Kelleners et al. 2005). In this study, a calibration curve was made for each of the two soil types, aspen and conifer, by packing a flower pot with soil to approximately the same bulk density as was found in the field (Appendix Figure A1). The soil was then saturated and the CS616 inserted. As the soil dried it was repeatedly weighed and measured with the CS616. The CS616, with a length of 12 cm as used here, has an accuracy of $\pm$ 3%. Insertion error is possible if the prongs are inserted in such a way that void space is created around them. Because CS616 probes have exhibited some temperature dependence (Campbell Scientific, 2004), 14 temperature sensors (temperature pendants, HOBO, Pocasset, MA,

USA) were buried at a variety of hillslope positions. Seven were buried in each

vegetation type during the 2008 water year, four at 5 cm depth and two at 20 cm. These

sensors log temperature hourly and were installed to study freezing pattern of soils during

fall-winter-spring months. Other factors measured at each grid point were litter cover,

litter depth, rock cover (%), time of day, vegetation cover (%), position as related to the

canopy and snow depth (when applicable). Litter cover, rock cover vegetation cover and

each grid point's relation to canopy (under, outside, or on the edge of) were estimated

through observation.

Geostatistical analysis

Geostatistics can be used to describe spatial organization of soil moisture. Semi-

variograms depict the semi-variance between points as a function of the distance between

them (Western et al., 2004). This shows the range, or lag distance, over which soil

moisture is spatially correlated (Fitzjohn et al., 1998). The main features of a semi-

variogram are the sill, range and nugget. The sill is the semi-variance at which the semi-

variogram levels off: the semi-variance of the more distantly separated measurement

points. The correlation length, is the lag distance at which the semi-variogram reaches its

sill (Western et al., 2004). The effective range is considered 3 times the correlation

length. With spatial soil moisture data, effective ranges should reflect the correlation

length of a variable (or variables) controlling soil moisture (Fitzjohn et al., 1998). Soil

moisture correlations lengths typically vary between 1 and 600 m, with most small

catchment studies having lengths between 20 and 300 m. The nugget is the semi-variance

at which the semi-variograms begin: the semi-variance between closely spaced points. It

reflects a combination of small scale variability and measurement error (Western et al., 2004).

Semi-variograms were made from the soil moisture grid data using R software for statistical computing version 2.5.1. Exponential models were fit to the empirical semi-variograms using weighted least squares and Cressie weights. Nuggets were determined by the semi-variance of closely spaced measurements and were specified for each vegetation type. Contour maps were made using Golden Software Surfer, which uses kriging to interpolate between measurements.

RESULTS

Water inputs

Data from nearby SNOTEL sites showed that the area received more snow in the 2008 water year than in 2007 (Figure 2.3). The average precipitation since 1990 as measured by three nearby SNOTEL sites was 887 mm a year, with 74% falling as snow. In 2007 the catchment received only 84% of average precipitation: 743 mm, with 66% falling as snow. In 2008 it received 889 mm, 100% of average precipitation, with 86% falling as snow. Peak snow accumulation came on March 12th in 2007 and April 13th in 2008. The average day of peak SWE since 1979 is April 4th. Figure 2.3 also shows peak SWE from snow survey measurements at the study site under aspen and conifer canopies.

In both years the average SWE surveyed under forest canopies (Figure 2.3) was less than SWE recorded at the SNOTEL sites, which are in clearings. SWE measurements in upslope transects are shown in Figure 2.4. Though SWE varies, there is no accumulation of snowpack in the lower portions of the hillslopes; as of peak snow

accumulation, snowpack did not vary with hillslope position. A comparison of peak

SWE under aspen and conifer canopies (Figure 2.5) shows that on average, aspen

accumulated more SWE than conifer and this difference was more pronounced in the

wetter year (2008). A two-sample t-test revealed that in 2007 the aspen stand had an

average of 315 mm of peak SWE on and the conifer stand had 280 mm, an 11%

difference (t = 2.22, df = 72.78, p-value = 0.029). In 2008 aspen had 509 mm and conifer

390 mm, a 23% difference (t = 5.55, df = 153.97, p-value = 1.18e-07).

Under the conifer canopy there was a notable pattern of drifts forming between

trees or groups of trees. While areas under the densest part of the canopy were free of

snow by May 20th, snow in drift areas lasted well into June (Figure 2.6). Figure 2.7

shows SWE measurements made under (n = 20), outside (n = 51) or on the edge (n = 19)

of the conifer canopy at peak accumulation in 2008. An analysis of variance (ANOVA)

was used to assess the differences in accumulation among the locations relative to conifer

canopy cover. At the edge of the conifer canopy there was an average of 431.8 mm (SE

= 25.18), which was not significantly different (p = 0.38) from the average outside of the

canopy (401.2 mm, SE = 19.92). The average snow accumulation under the conifer

canopy was 323.85 mm, significantly less than both the edge (p = 0.010) and outside (p =

0.025).

Snow melt (considered SWE ablation) and rain measured by snow pillows and

collection cans at the three SNOTEL sites were combined to estimate surface water

inputs, shown in Figure 2.8. These values should be taken as maximum possible surface

water inputs since snow ablation could represent sublimation, evaporation and melt. The

spring pulse of inputs was greater and later in the 2008 water year than in 2007. The

maximum daily surface water input in 2007 was 46 mm on April 25th and 85 mm on

May 20th of 2008.
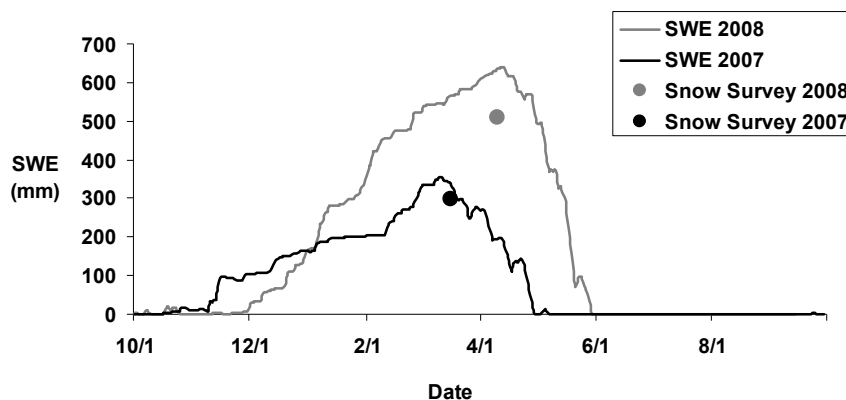
Figure 2.3. Average snow survey measurements for each year (circles) and daily average
SWE since 1990, in the 2007 and 2008 water years, as measured by Lightning Ridge,
Horse Ridge and Dry Bread Pond SNOTEL stations (lines).

Infiltration

Infiltration capacities were measured at 16 points, 9 on the aspen hillslope and 7

on the conifer. The minimum infiltration rate found in the aspen stand was 0.9 mm min^{-1}

in the aspen stand and 0.42 mm min^{-1} in the conifer stand. Infiltration rates are shown in

Table 2.1 and a comparison of aspen to conifer average rates is shown in Figure 2.4. A t-

test showed that there was no significant difference between average infiltration rates in

the aspen (3.32 mm hour^{-1}) and conifer (4.54 mm hour^{-1}) sites (t = -0.49, df = 11.46, p-

value = 0.63).

(a)

(b)
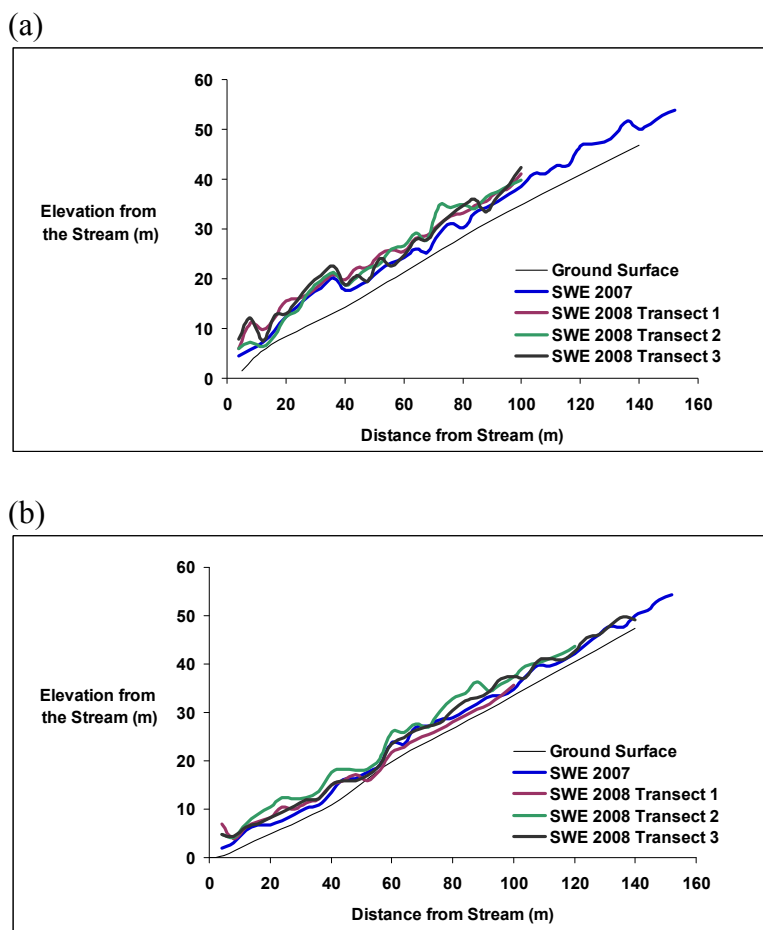
Figure 2.4. SWE measured in upslope transects during peak snow accumulation in 2007 and 2008 on (a) aspen and (b) conifer hillslopes.
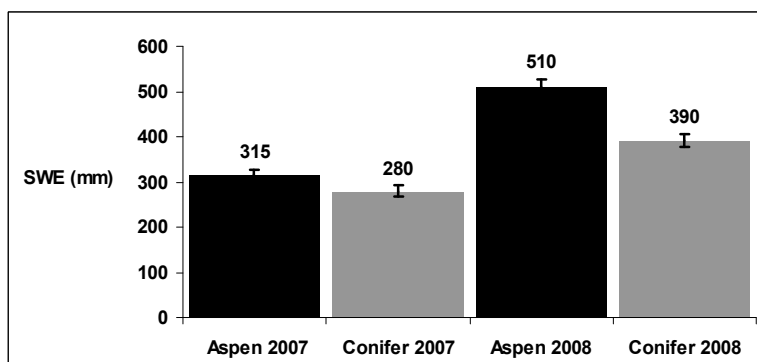
Figure 2.5. SWE under aspen and conifer canopies for the two water years studied. Error bars represent one standard error.

a)   b)
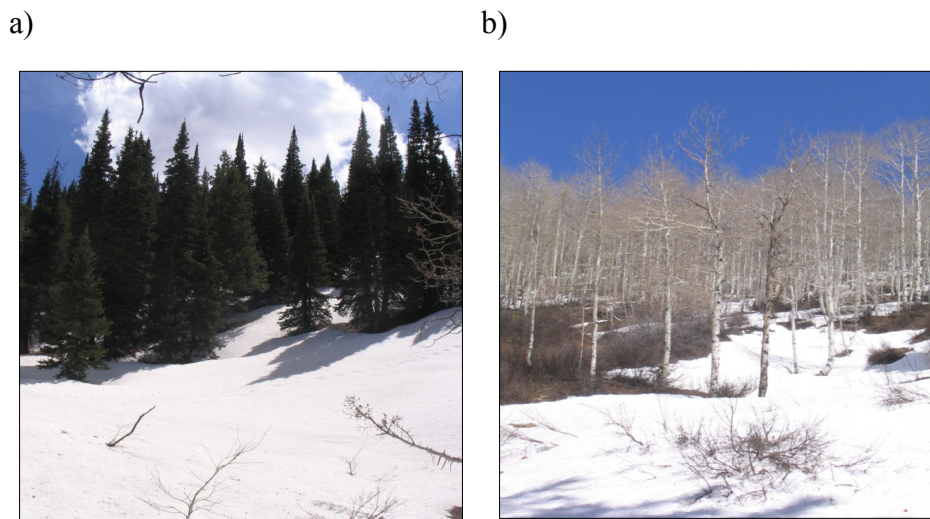
Figure 2.6. On May 28th snow drifts in the conifer site (a) were deeper than in the aspen site (b).
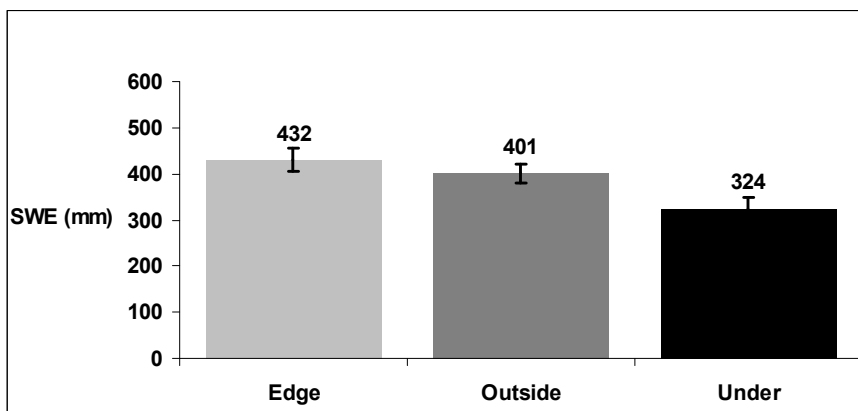
Figure 2.7. SWE measured under, outside and at the edge of the conifer canopy during peak snow accumulation in 2008. Error bars represent one standard error.

(a)

(b)
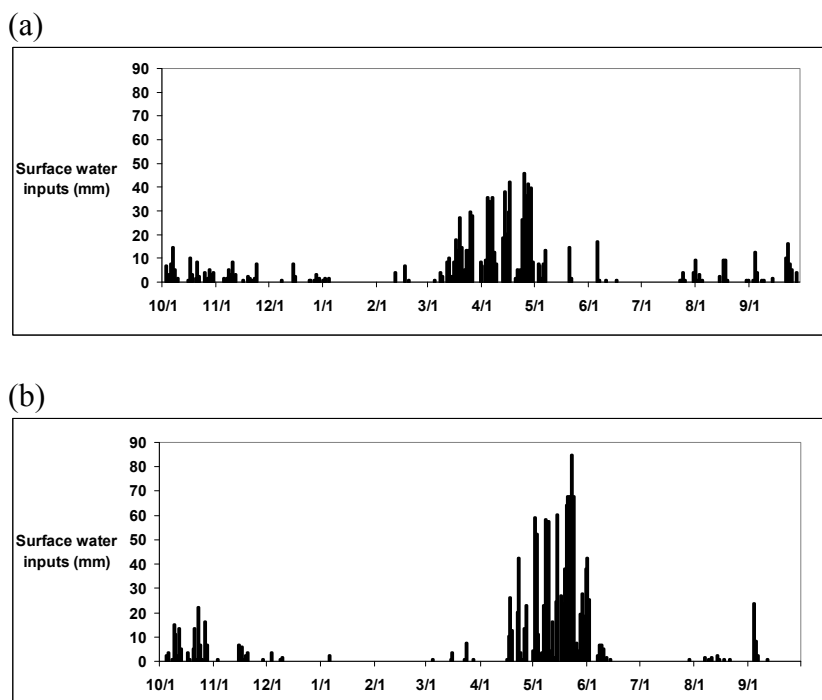
Figure 2.8.   Averaged surface water inputs measured at three SNOTEL sites in 2007 (a) and 2008 (b).
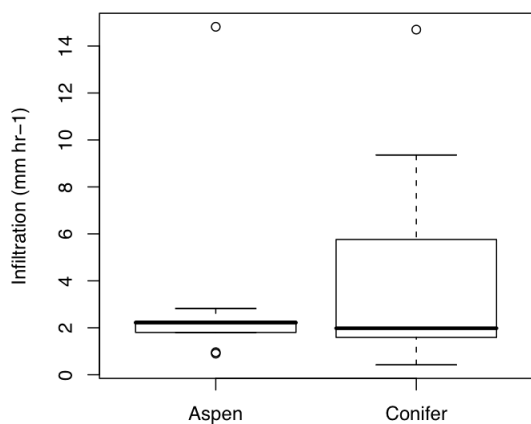
Figure 2.9. Boxplots of infiltration rates in conifer and aspen, median values (solid line), interquartile range (inside box), whiskers 1.5 times the interquartile range, circles represent outliers.

Table 2.1 Infiltration rates measured in each vegetation type.

	Infiltration (mm min^{-1})	R^2
Aspen	0.90	1.00
	0.96	1.00
	1.80	1.00
	1.80	1.00
	2.22	0.99
	2.28	1.00
	2.28	0.99
	2.82	1.00
	14.82	1.00
Conifer	0.42	0.99
	1.50	1.00
	1.68	1.00
	1.98	0.99
	2.16	1.00
	9.36	0.97
	14.70	1.00

Soil physical properties

Soil textures at varying depths for both vegetation types are shown in Figure

2.10 and Table 2.3. For details on locations where the samples were taken, see

Appendix, Table A2. The average textural class for soils under both vegetation types was

loam, though some lower horizons of the conifer soil were sandy loam. There was a

notable difference between the rock content and soil depth of the profiles. The aspen soil

had an average rock content by volume of 1% (n = 12)) and its depth was greater than 1

m. The conifer soil was very gravelly, with an average rock content of 46% by volume

(n = 9). The profile became increasingly rocky with depth, fading into fractured bedrock

between 45 and 60 cm. Average bulk densities were higher in the conifer site, due to the greater rock content. However if both the volume and weight of particles greater than 2 mm were removed from the samples, conifer soil bulk density was considerably lower (Table 2.2).

Average temperatures in the top 5 cm soils was just above freezing while the soil moisture grids were being measured under snow on May 19th and 20th of 2008. The average temperature was 4.8°C in the aspen soil and 0.23°C in the conifer. Soil temperatures were well above freezing during the other grid measurements. Periods of subfreezing temperatures in the 2008 water year began as early as November 4th and lasted until April 14th in the conifer and to May 3rd in the aspen. Soil temperatures from the three nearby SNOTEL sites show only one site briefly dropping below freezing in the fall of 2006. While in the fall of 2007, all sites fell below freezing due to a drop in air temperatures before an insulating snowpack could accumulate. Our study site experienced more shade than the SNOTEL sites and therefore had subfreezing temperatures in the soil for longer.

A t-test showed that percent cover of leaf litter was not significantly different between the conifer and aspen sites (t = 0.043, df = 170.29, p-value = 0.97). Aspen had an average of 51.0% and conifer 50.9%. Depth of leaf litter was significantly greater at the conifer site (average depth was 1.06 cm) than at the aspen site (average depth was 0.10 cm) (t = -4.07, df = 89.97, p-value = 0.0001).

Table 2.2. Average soil textures at multiple depths in both vegetation communities.

Depth (cm)		n	Sand %	Silt %	Clay %	Texture Class
0-5	**Aspen**	14	44.8	37.7	17.5	Loam
	Conifer	13	41.5	41.6	16.9	Loam
5-20	**Aspen**	4	38.9	39.5	21.6	Loam
	Conifer	5	35.4	44.8	19.8	Loam
20+	**Aspen**	4	40.3	38.0	21.7	Loam
	Conifer	4	54.8	28.9	16.4	Sandy Loam

Table 2.3. Average bulk densities at multiple depths. Bulk densities including all particle sizes (with rocks) are highlighted in grey. Bulk densities with both the volume and weight of particles greater than 2 mm removed (without rocks) are also included.

	Depth (cm)	Volume of rocks (%)	n	Average Bulk Density with Rocks (g cm^{-3})	n	Average Bulk Density without Rocks (g cm^{-3})	n	Average Porosity with Rocks %	n	Average Porosity without Rocks %	n
Conifer	0-5	35	3	1.19	52	0.76	3	55	52	71	3
	5-20	36	3	1.32	3	0.61	3	50	3	77	3
	20 +	68	3	1.91	3	0.67	3	27	3	74	3
Aspen	0-5	0.9	3	1.13	26	1.13	3	57	26	63	3
	5-20	1.1	4	1.13	4	1.13	4	57	4	57	4
	20 +	0.7	4	1.17	4	1.17	4	55	4	56	4
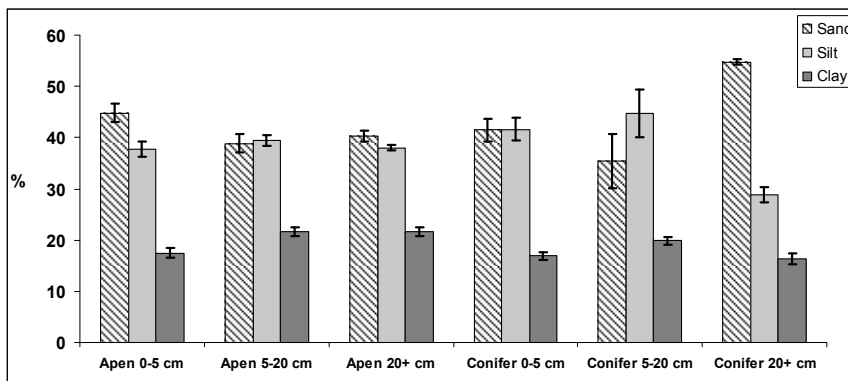
Figure 2.10. Percent sand silt and clay for soils (< 2 mm fraction) under aspen and conifer stands, error bars represent plus or minus 1 standard error.

(a)

(b)
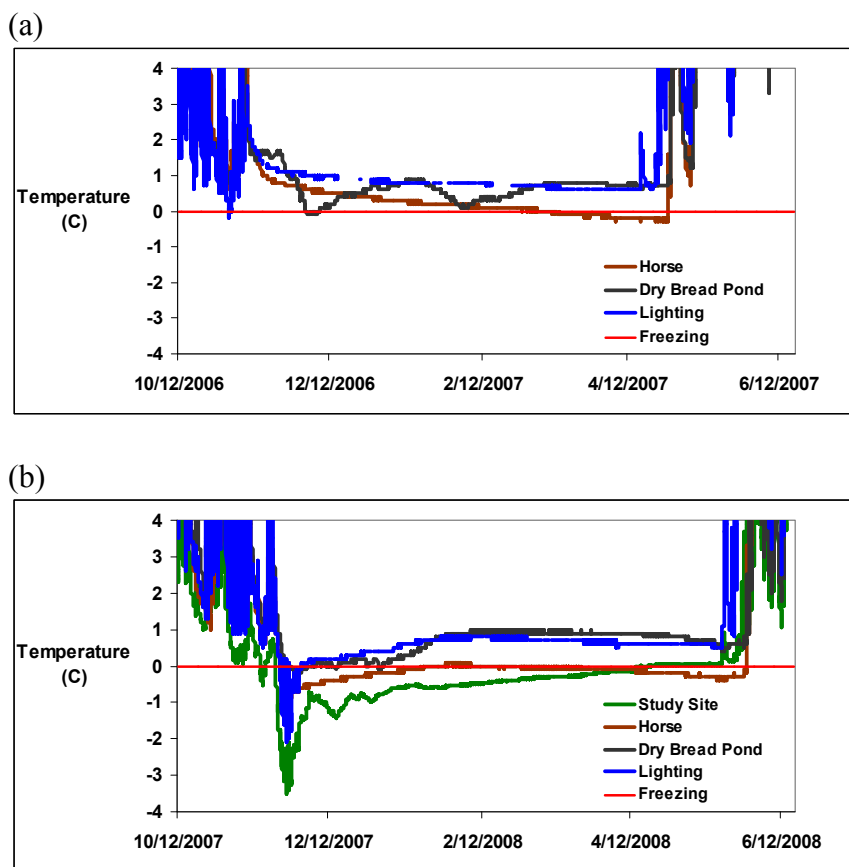
Figure 2.11. Soil temperatures at 5 cm for (a) water year 2007 measured three SNOTEL sites and (b) water year 2008 at three SNOTEL stations and at the study site.

Streamflow

Streamflow, gauged at three points along the stream (Figure 2.2), showed two distinct hydrograph peaks during the spring snow melt. The first peak below the aspen hillslope was 30.8 L s^{-1} on May 18th. The conifer-gauging point peaked at 69.7 L s^{-1} on May 20th (Figure 2.12a). Figure 2.12 also shows surface water inputs (SWI) measured at the three nearby SNOTEL sites. Here SWI was a combination of rain and snow ablation (to represent snow melt). The lull in streamflow between the two peaks can be attributed to a storm passing, which dropped temperatures from a daily average of 14 to 1°C and left 29 mm of SWE between May 21st and 23rd. The second hydrograph peaks were greater for the middle and conifer gauging points, which peaked at 31.7 L s^{-1} on June 1st, and 82.3 L s^{-1} on June 2nd, respectively. Streamflow below the aspen hillslope did not peak at this time, but sustained slightly raised flows (~12 L s^{-1}) until sensors at all gauged points dropped off to baseflows in mid June (~12th).

A hydrograph from the conifer dominated contributing area was isolated by subtracting flow from the middle section from flows at the conifer gauge (Figure 2.12b). The total estimated water yield for the 2008 water year from the conifer contributing area was 0.73 m m^{-2}, with approximately 60% coming with the spring snow melt pulse (April 6th to July 4th). The 2008 annual water yield below the aspen hillslope was 0.64 m m^{-2}. With a smaller and more upland contributing area, baseflow below the aspen hillslope was a mere trickle and 87% of all water flowing there comes during the spring pulse (February 13th to June 17th). This demonstrates the importance of the spring snowmelt for water yield in this snow driven system.

Plots of soil moisture in the top 5 cm of soil measured at eight points along transects in each vegetation type, versus streamflow show a threshold behavior at both gauging points (Figure 2.13). In this case, flow gains from the conifer site are isolated by subtracting the flow at the middle gauging point from that measured below the conifer hillslope. The plots (Figure 2.13) show a break in slope, indicating a soil moisture threshold for increased streamflow. Soil moisture thresholds, past which streamflow increased, were 0.34 cm^3 cm^{-3} in the conifer and 0.22 cm^3 cm^{-3} in the aspen.

(a)
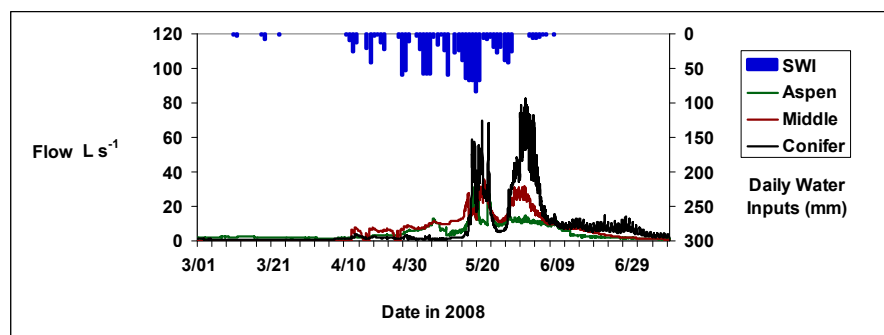
(b)
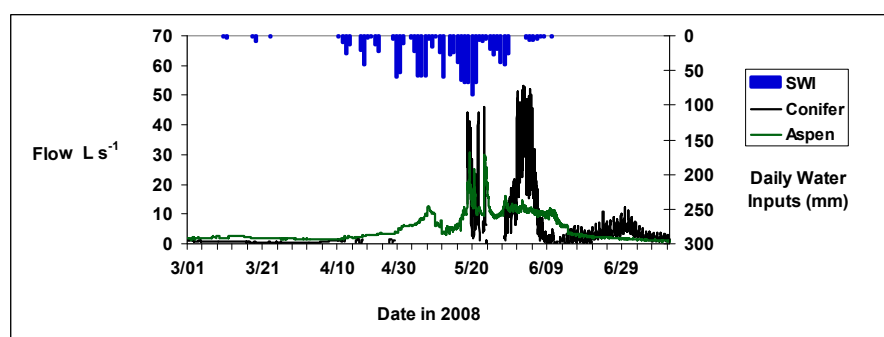
Figure 2.12. (a) Surface water inputs (SWI) and streamflow at three points in the stream, below the aspen hillslope, in the middle, dividing the aspen from the conifer source areas and below the conifer dominated area, in the 2008 water year. (b) SWI and hydrographs from the aspen and conifer contributing areas.

Continuous soil moisture

Continuous volumetric water content, measured in the top 5 cm of the soil profile, was averaged among all points in both the aspen and conifer plots. Peak soil moisture in 2008 came later and was lower than in 2007 (Figure 2.14).   Both sites peaked on April 29th, 2007, with a water content of 0.57 in aspen and 0.29 in conifer. In 2008, they

(a)

(b)
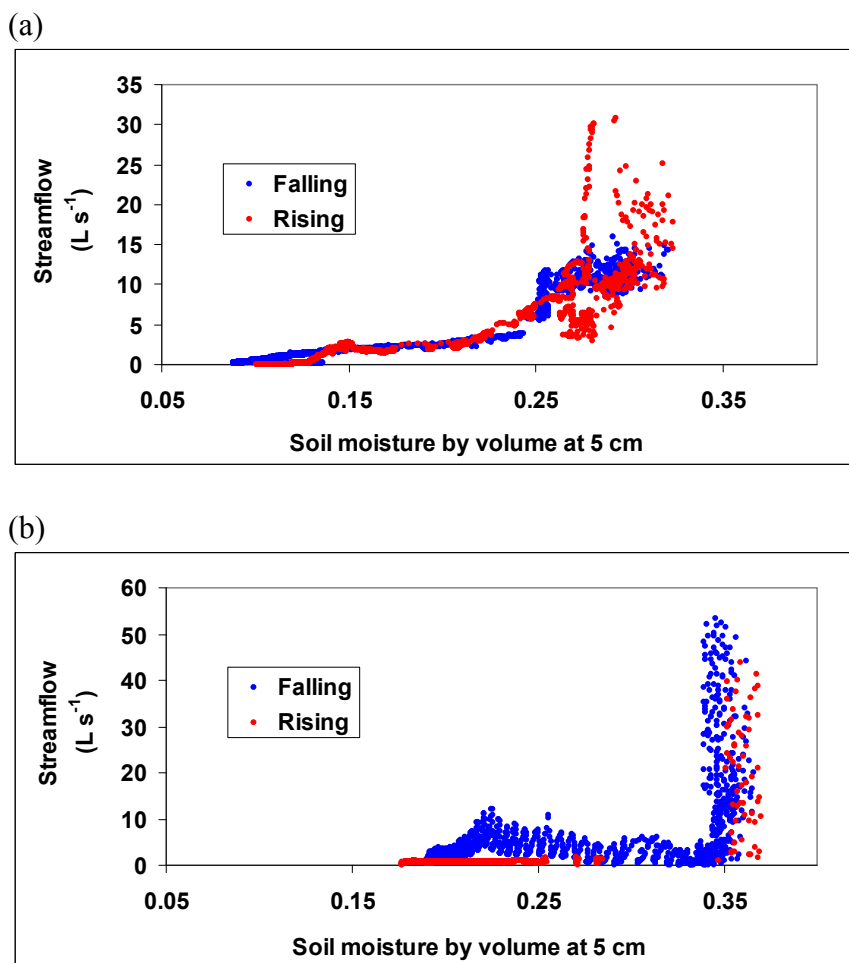
Figure 2.13. Streamflow plotted with averaged soil moisture measured at 5 cm depth at (a) eight points in the aspen hillslope and (b) eight points in the conifer hillslope. Red points represent measurements on the rising limb of the hydrograph, blue represent those on the falling limb.

peaked on May 19th, with a water content of 0.51 in the aspen and 0.21 in conifer. It is worth noting that the 2008 water year received 20% more snow than the 2007 water year, yet soil moisture started 35% lower in the fall and peaked 10% lower in the spring. This could have been an artifact of soil settling around the probes (reducing porosity) after the wetting and drying cycle of their first year of installation. Another possible explanation is that in 2008 average soil temperatures in the top 5 cm of soil fell below 0°C, as early as November 4th before significant snow cover fell in early December (Figure 2.14b). Frozen soils may have impeded water from infiltrating into the soil or frozen water within the soil could reduce its permittivity. In the fall of 2007, neither weather stations showed subfreezing temperatures (Figure 2.14a).

Soil moisture grids

Soil moisture grid measurements in 2007 showed no spatial organization in either hillslope. It was a dry year, peak soil moisture had already passed by the time of measurement, and the soil moisture was too low to show any spatial organization. Figure 2.15 shows the relative timing of peak soil moisture, streamflow, surface water inputs and soil moisture grid measurements. A hydrograph from the middle gauging station in 2008 shows that coincidentally, both streamflow and soil moisture peaked on the same day as the first soil moisture grid measurements on May 19th and 20th. The second time the grids were measured, the hydrograph peak had already receded and soil moisture was beginning to drop. The third grid measurements were made at baseflow in the stream and after the soil moisture peak had passed.

(a)

(b)
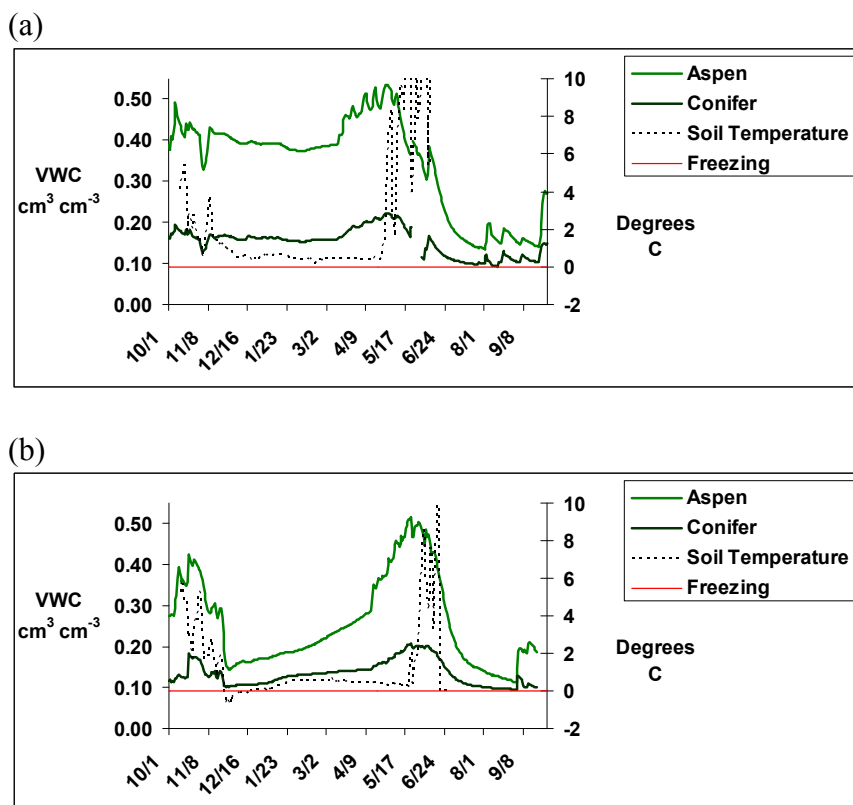
Figure 2.14. Soil moisture in the top 5 cm of soil, averaged over eight points on transects in each vegetation types for (a) the 2007 water year and (b) the 2008 water year. Plots include soil temperatures from three nearby SNOTEL sites.

Surveyed points of the grids from both vegetation sites show slopes and elevation gains within the grids to be relatively equal (Figure 2.16). Soil moisture measurements on the aspen grid during the dry down of 2008 show some spatial organization during the wettest measurements of the aspen grid (Figure 2.17). Semi-variograms of the residuals are shown in Figure 2.18. Exponential models fit to the empirical semi-variograms estimate sills and effective ranges shown in Table 2.4. The semi-variogram of the aspen soil moisture in July had no logical sill; the average variance between points was smaller that the measurement error or nugget (Figure 2.18 and Table 2.4).

As both aspen and conifer hillslopes dried down, shallow soil moisture became more uniform (Figures 2.17 and 2.19). The average aspen soil moisture dropped 7% between May 19th and June 14th and 11% by July 10th. In the conifer site average soil moisture dropped 27% between May 20th and June 14th and 88% by July 10th. The decreases in soil moisture variability accompanied by dry down are reflected in the sills of the semi-variograms in Figure 2.18 and Table 2.4. The sills, or the semi-variance between distant points, decreased from 2.39 to 0.014 on the aspen hillslope. The drastic decrease in soil moisture (88%) from May to July in the conifer hillslope coincides with the drastic decrease in sills (74 to 2.6).

The aspen soil moisture measurements show organization in May and June, but by July soil moisture appears to have no spatial correlation. The conifer hillslope showed some spatial correlation in May but did not show any in June or July. The effective range or correlation length was the longest on the aspen hillslope in June (114 m).

Other factors measured on grid as possible explanatory variables included litter cover, litter depth, rock cover (%), time of day, vegetation cover (%), position as related to the canopy and snow depth (when applicable). While none of the variables measured appeared to have strong control on soil moisture, a few were found to be significant. Litter cover had a mild, positive correlation to soil water content in May ($R^2 = 0.14$, F = 14.34 on 1 and 88 df, p-value = 0.00028) and June on the aspen hillslope ($R^2 = 0.10$, F= 10.11 on 1, 88 df, p-value = 0.0020). On the conifer hillslope litter cover and snow depth (measured in May) together were positively correlated with soil moisture in June ($R^2 =$ 0.20, F = 11.25 on 2 and 87 df, p-value = 4.53e-05) and July ($R^2 = 0.20$, F = 11.06 on 2 and 87 df, p-value = 5.27e-05). The time of day each measurement was taken had no

significant effect on soil moisture, indicating that our measurements were taken close

enough together so as not to pick up diurnal melting patterns.
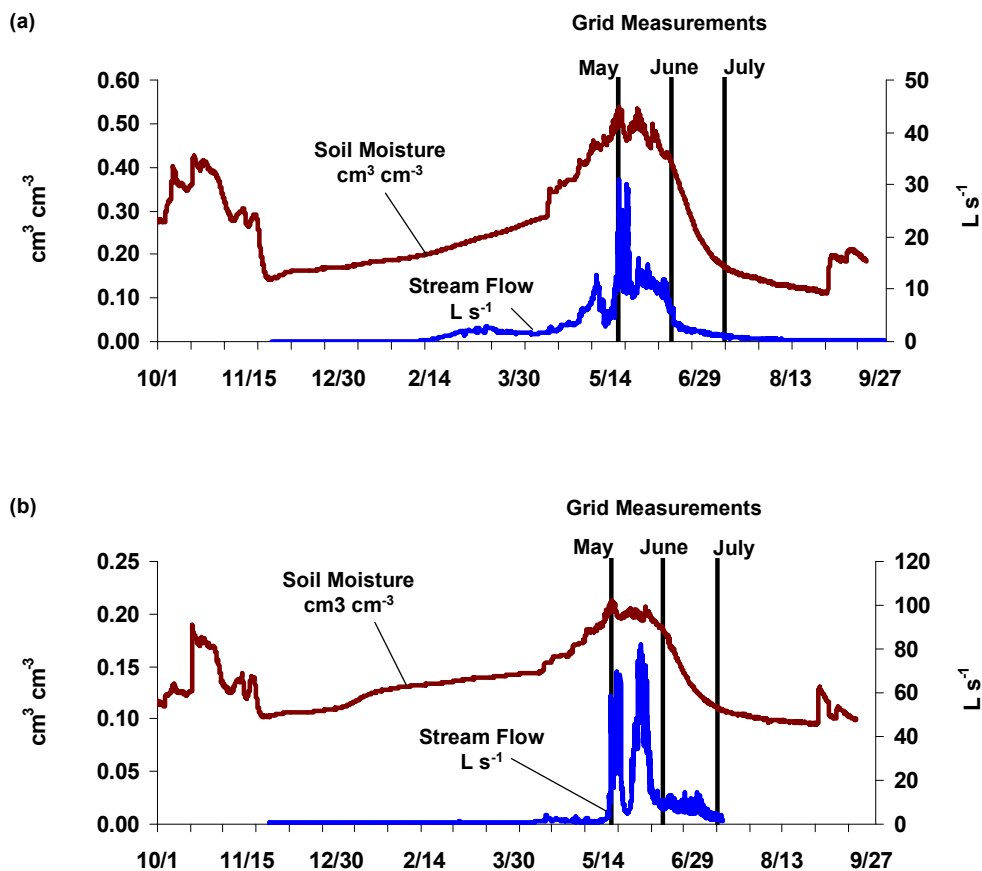
Figure 2.15. Timing of three soil moisture grid measurements, and average soil moisture at 5 cm depth (red line) and streamflow (blue line) below (a) the aspen hillslope and (b) the conifer hillslope during spring snowmelt 2008.
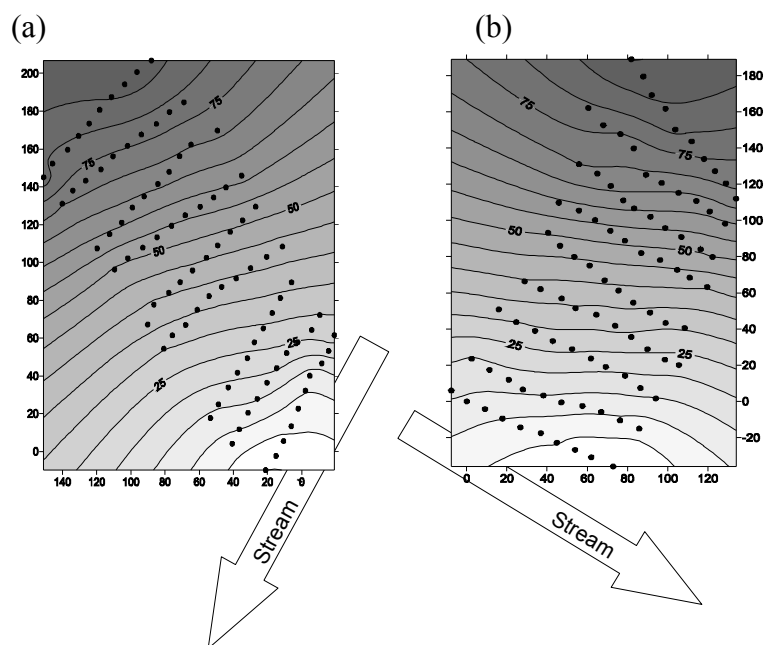
Figure 2.16. Topography of the aspen (a) and conifer (b) hillslopes. Contour lines are at 5 m intervals. Dots indicate soil moisture grid points in 2008.

Table 2.4. Effective range, sill, and nuggets for all soil moisture grids measured in 2008.

		Effective Range (m)	Sill	Nugget	Mean Soil Moisture (%)	Minimised Weighted Sum of Squares
Aspen	May	17.58	2.59	0.1983	35.47	67.67
	June	114.06	1.02	0.1983	20.67	112.72
	July	3.00	0.01	0.1983	5.85	778.44
Conifer	May	12.93	74.47	9.00E-05	24.23	110.42
	June	9.21	65.60	9.00E-05	17.66	23.71
	July	0.93	2.58	9.00E-05	2.94	41.06

(a)

(b)
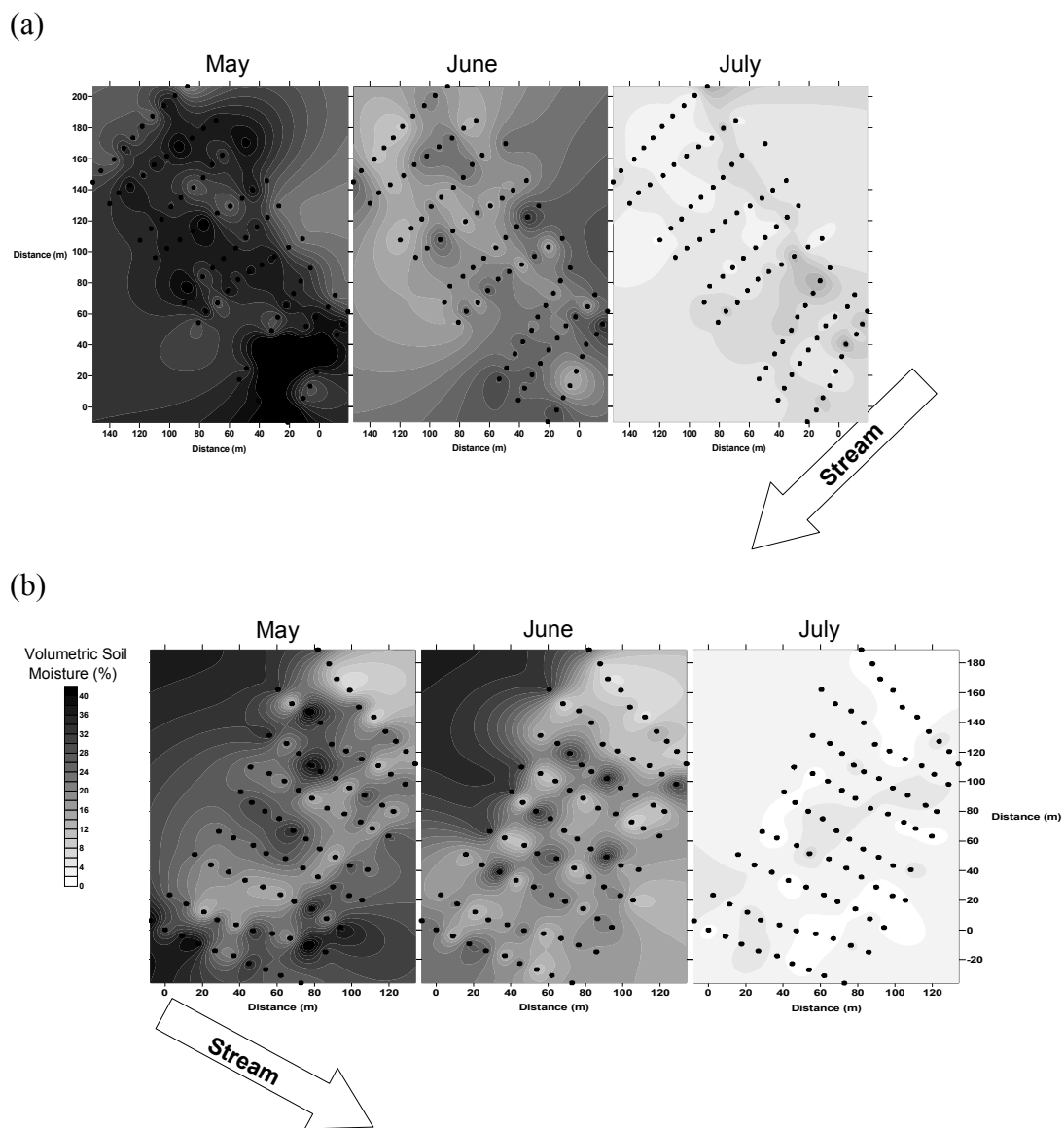
Figure 2.17. Soil moisture measurements made during dry down in 2008 on (a) aspen and (b) conifer hillslopes.
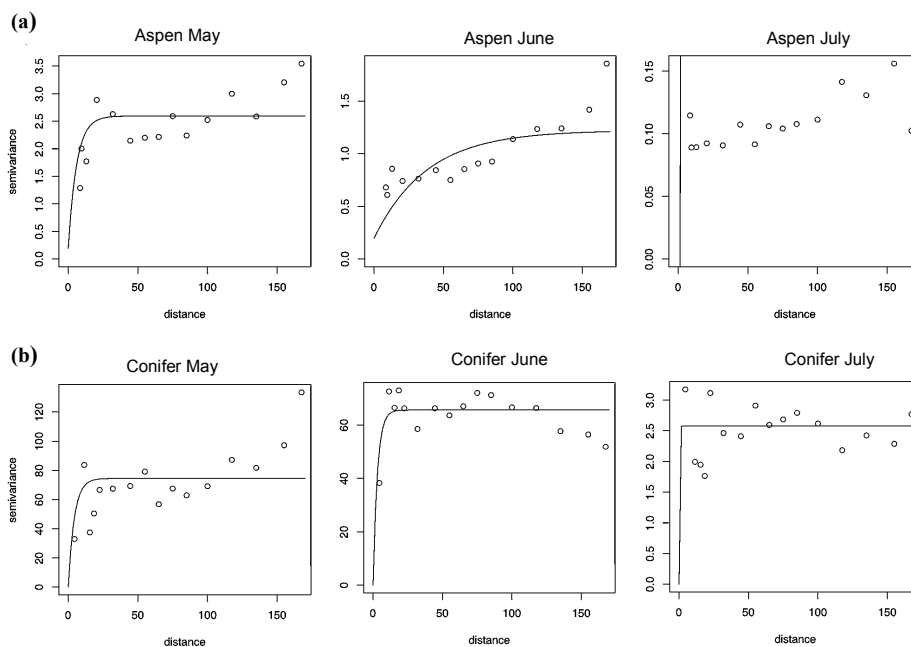
Figure 2.18. Semi-variograms of residuals from soil moisture grid measurements for (a) aspen and (b) conifer sites on May 20th, June 14th and July 10th.
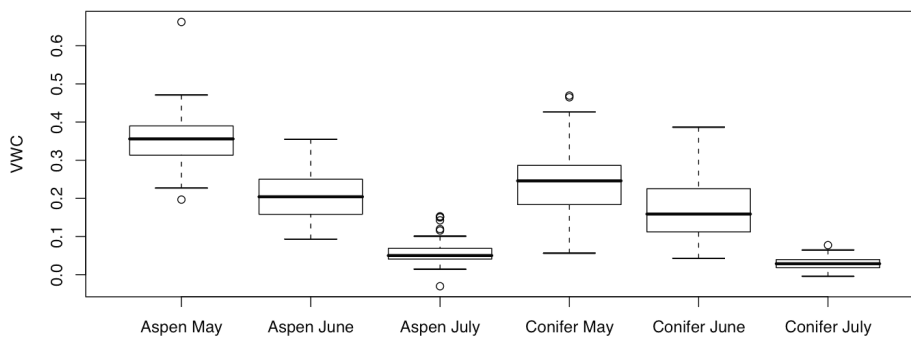
Figure 2.19. Boxplots of volumetric soil moisture from grid measurements on 3 dates during spring snowmelt in 2008

Overland flow

Upon first arrival to the catchment in the spring, saturation overland flow was observed (but not quantified) in the lower portions of the catchment, on May 20th. Thin films of surface water were observed at the aspen site on May 19th but had grown patchy

by May 28th. Overland flow and return flow were observed at the conifer site on May

20th, 28th and 31st but had subsided by June 7th (Figure 2.20). Since snow still covered the

soils surface, observations of overland flow on middle and upper portions of the

hillslopes were limited to points that were dug to for soil moisture measurements.

However, near the stream, sheet flow was seen on areas of exposed soil near the stream,

routing water from the base of the snowpack to the stream. Overland flow was observed

on the rising limb of the hydrograph, after May 20th, the day of highest surface water

input during for the 2008 water year. However, since no one was there before May 20th,

it is not known exactly when overland flow began.

(a)

(b)
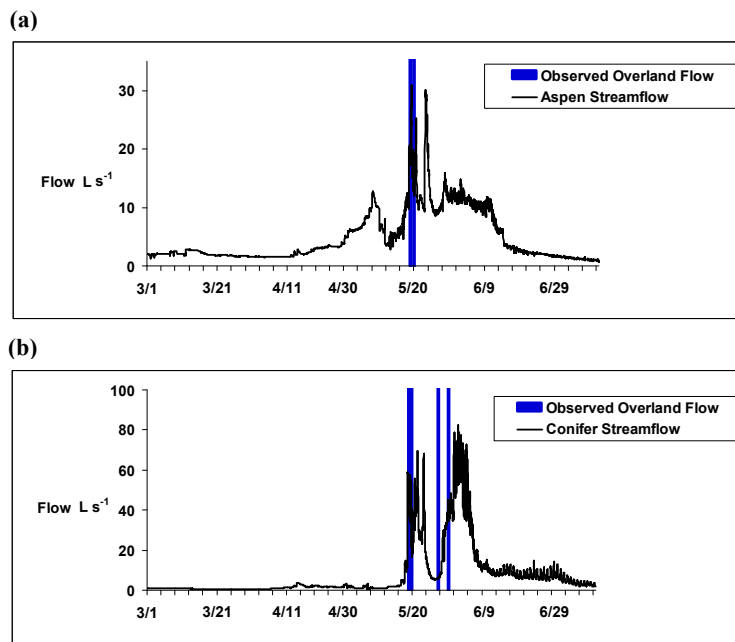
Figure 2.20. The dates of overland flow observed in each vegetation type plotted with
the 2008 hydrograph gauged below (a) the aspen hillslope and (b) the conifer hillslope.

DISCUSSION

This project focused on possible differences in water yield between aspen and conifer dominated stands, by concentrating on the soil moisture patterns on hillsides adjacent to a first order stream in Northern Utah. The goal of the study was to gain a clear understanding of soil moisture patterns and their relationship to vegetation types and runoff production.

Canopy effects on snow accumulation

Canopy effects on snow accumulation are important because they affect surface water input and soil moisture. Student t-tests revealed that the aspen stand had an average 11 and 23% greater peak SWE than the conifer stand in 2007 and 2008 respectively (Figure 2.5). These differences reinforce conclusions from a study in a neighboring catchment, with similar slopes and aspect, in 2005 and 2006 (LaMalfa and Ryel, 2008). They found that SWE under conifer plots was 34 to 44% less than in nearby aspen stands. Differences in snow accumulation under aspen and conifer in the study area showed a linear increase with precipitation ($R^2 = 0.9876$) (Figure 2.21). The four years of SWE surveys SWE at the site span the interquartile range of annual precipitation measured at the three nearby SNOTEL sites measured since 1990 (the last 18 years).

The significant difference between SWE accumulation in aspen and conifer stands indicates that the conifer canopy lessened the surrounding snowpack. Whether or not this
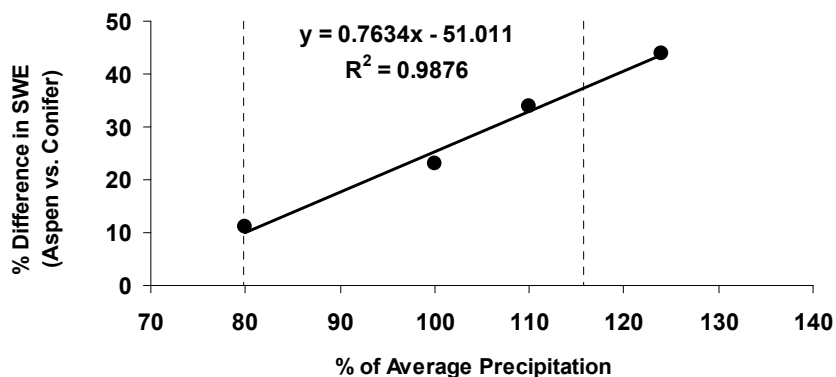
Figure 2.21. Percent less SWE in conifer stands than in aspen in the four years (2005-2008). Dashed vertical lines represent the interquartile range of all annual precipitation since 1990 measured at the three SNOTEL sites.

difference in SWE translates into less hillslope runoff from the conifer site depends on if

snow caught in the canopy is being re-distributed or sublimated (a true loss to the

system). Redistribution of snow within a hillslope has implications for snowpack

duration since larger drifts will last longer into the year as seen in the conifer site but

does not mean a water loss from the catchment. Canopy shelter decreases redistribution

and sublimation from the snowpack under the canopy, providing more soil water inputs

(Hiemstra et al., 2002; Marks et al., 2002). However, the shelter of a conifer canopy also

comes with greater canopy interception. Snow held in the canopy has greater surface

area and more exposure to wind, leading to more sublimation than the underlying

snowpack. While sublimation from the snowpack in the Intermountain West has been

shown to be relatively small in some cases (McNamara et al., 2005; LaMalfa and Ryel,

2008), LaMalfa et al. (2007) used artificial branches in experiments to show that

sublimation could explain most or all of the differences in SWE accumulation under

aspen and conifer canopies. Sublimation losses from intercepted snow in conifer canopies warrant further investigation.

Snow deposition and melt, which are largely a function of topography and vegetation, are typically responsible 80% or more of streamflow in the semi-arid West (Marks et al., 2002). Runoff production in these systems is driven by infrequent and short spring melt events. In order to understand runoff production in these systems we must understand their snow hydrology, including the timing, intensity and volume of snow melt and precipitation (Wilcox et al., 1997). In a semi-arid catchment of New Mexico, receiving half of its precipitation as snow, Wilcox et al. (1997) list snow accumulation patterns as a major factor affecting fall and winter runoff. Our catchment receives even a greater percentage of it precipitation as snow (74%), so snow accumulation and melt patterns are likely to be even more important. A more detailed examination of snow hydrology, including redistribution of melt water within the snowpack and sublimation from conifer canopies, will help elucidate mechanisms of runoff production in the study catchment.

Soil moisture and water inputs

Grayson et al. (1997) defined a wet state of soil moisture as when water inputs exceeded evapotranspiration and saturated zones expanded to meet one another, hydraulically connecting the hillslope. In a wet state, non-local factors controlled soil moisture, such as hillslope scale topography and water inputs. Patterns of soil moisture form as soil moisture content increases with upslope contributing area. In this study longer periods of a wet soil moisture state were found in the aspen hillslope than the

conifer, most likely due to greater water inputs in the aspen stand from its greater SWE

accumulation (Figure 2.5). However, providing a clear link between snow melt inputs

and soil moisture patterns proved difficult. Differences in peak SWE between the two

vegetation types can be attributed to canopy effects in the conifer stand. Even within the

conifer stand there was greater peak SWE outside the canopy than at the edge or under

the canopy (Figure 2.7). Varied snow accumulation explained 20% of the variation in

soil moisture measured on the conifer grid. No other variable measured (litter depth, rock

cover, and under story vegetation) had major control on surface soil moisture, there are

other possible controlling factors that were not measured at each grid point. Bulk

density, soil texture and soil depth were not measured at each point and have potential for

dictating soil moisture patterns. Grant et al. (2004) measured landscape features, soil

depth, soil physical properties, vegetation and SWE and found that only the clay and

coarse fragment fractions of the soil were correlated to soil water content. Soil texture at

my site did not vary greatly between the vegetation types. However, bulk density was

higher in the conifer soils, where the profile was shallower. Rock content in particular

could be very significant in controlling soil moisture patterns on the conifer hillslope.

The volume of subsurface occupied by rocks is essentially space that cannot be occupied

by moisture; therefore it behaves and is measured as dry volume. Root distribution was

not examined but is assumed to differ between aspen and conifer. Conifer root systems

are deeper than deciduous and control the volume of soil from which plants harvest their

water (Schume et al., 2003). Bond et al. (2007) also pointed out that roots in dry climates

may go as deep as bedrock in search of available water. An investigation of root depths

and densities in each of the vegetation types would be helpful in understanding vegetation effects on soil moisture.

The timing of surface water inputs and transpiration were different between the vegetation types. Snow had disappeared from the aspen hillslope six days sooner than from the conifer site, making surface water inputs for the aspen shorter in duration and higher in intensity. Previous studies at this site showed that conifers have longer annual transpiration activity periods, beginning earlier in the spring with the advantage of retaining their needles throughout the winter (LaMalfa, 2007). Several studies found vegetation to be a leading control on soil moisture (Gomez-Plaza et al., 2001; Chaplot and Walter, 2003; Tromp-van Meerveld and McDonnell, 2006a), but McNamara et al. (2005) showed that the energy balance, evapotranspiration and water inputs control the switch between soil moisture states in a snow-driven system.

Our findings support hypothesis (1), that soil moisture patterns will remain organized longer in the aspen hillslope. This is seen in Figures 2.17 and 2.18 which show a longer duration of a wet state in the aspen soil. Whether this can be attributed to greater snow melt inputs on the aspen slope alone remains unclear. It is most likely due to synergistic relationships between snow-accumulation, melt rate, and soil physical properties.

Soil moisture variability

As both aspen and conifer hillslopes dried down, shallow soil moisture became more uniform as seen in Figures 2.17 and 2.19. A similar positive relationship between mean soil moisture and soil moisture variability was found by Western et al. (2004) in a

dry catchment of Australia with well drained soils. Seasonal evolution of soil moisture variability, as found in our catchment, has implications for erosional processes as demonstrated by Fitzjohn et al. (1998). They showed that in a semi-arid catchment, increasing soil moisture variability across the landscape increased its ecological function as a buffer for erosion. Drier, unsaturated areas acted as sinks by absorbing overland flow produced from neighboring patches of saturated surface soil. In our study, the variability of soil moisture decreased as the soil dried out through the season (Figure 2.19).

Soil moisture correlation lengths

Correlation lengths in the aspen hillslope increased from 17 m in May, to 114 m in June, and then decreased to 3 m when soil moisture was lowest in July. The soil moisture grid measured in May, at peak soil moisture when hillslope connectivity is most probable, does not exhibit the longest correlation length. These results are similar to those found in the humid Panola catchment by Tromp-van Meerveld and McDonnell (2006a), who saw no spatial organization during wet periods, when vertical percolation through bedrock presumably dominated, nor during dry periods, when local factors controlled soil moisture, but only in a brief transitional state between wet and dry. Our study also implies that, because water inputs were highly concentrated during the spring melt pulse, soil moisture contents were at or near saturation across the hillslope and no hillslope scale spatial variation in soil moisture was detected. Thus, during the wettest time, when subsurface lateral flow was most probable, no down-slope accumulation of soil moisture was seen.

In the conifer site there was spatial structure in May, with an effective range of 13 m. By June 10th, the conifer soil moisture grid showed no spatial organization. None of these correlation lengths are great enough to imply hillslope scale connectivity, though there is a positive relationship between mean soil moisture and the correlation length of spatial structure.

In Australia and New Zealand, Western et al. (2004) found that the relationship between soil moisture and correlation length differed between sites and that correlation lengths of organized soil moisture reflect the scale of processes controlling soil moisture at a given time. Shume et al. (2003) warned against over-interpreting semi-variograms, stating that conclusions should be drawn only from distinct changes in sills or ranges. Fitzjohn et al. (1998) state that an increase (double) in correlation length in soil moisture indicated spatial continuity and greater hillslope connectivity. In this case, the long effective range of soil moisture patterns on the aspen hillslope in June shows that the aspen hillslope remained in a wet state longer than the conifer hillslope. The conifer hillslope never exhibited hillslope scale correlation lengths and therefore did not show hillslope scale connectivity in shallow soil layers. The findings of this study agree with Fitzjohn et al. (1998), who showed that the dry state in soil moisture dominates in semi-arid regions and local controls create random patterns of soil moisture with only rare events leading to hillslope scale organization.

Hillslope-stream connectivity

Since spatial organization of shallow soil moisture in the conifer hillslope never exhibited a correlation length representative of hillslope scale connectivity, I presume

that hillslope-stream connectivity was never made in the shallow soil horizons (12 cm) measured. In the aspen hillslope the grid measurement in June had the longest correlation length and was the most likely to represent hillslope-stream connectivity. However, the stream below the aspen hillslope fell to baseflow at the same time as the June measurement. Therefore spatial organization observed in the aspen hillslope did not indicate hillslope-stream connectivity.

Although semi-variograms of soil moisture in the top 12 cm of soil do not appear to have detected hillslope-stream connectivity in either vegetation type, this is not to say connection did not happen in deeper layers. Figure 2.13 shows a threshold response of streamflow to soil moisture. Stream gauging below both hillslopes shows a sharp increase in flow after soil moisture thresholds of 0.34 cm^3 cm^{-3} in the conifer and 0.22 cm^3 cm^{-3} in the aspen. However, an increase in streamflow cannot be used as a surrogate for hillslope water contributions to the stream since other flowpaths have been observed in the catchment. So, although Figure 2.13 shows streamflow reacting to a threshold in soil moisture, it is not practical to assume that the top layers of soil are providing stream water. Though there is a clear relation between soil moisture and streamflow, most likely the stream is also receiving water from deeper layers within the soil profiles or along or within the underlying fractured bedrock. Tromp-van Meerveld and McDonnell (2006a) note that many studies using shallow soil moisture patterns have not explained how these shallow patters represent the entire soil profile. Several studies also recognize that subsurface lateral flow is more dependent on bedrock topography than surface topography (Freer et al., 1997; Chaplot and Walter, 2003; McDonnell, 2003; Tromp-van Meerveld and McDonnell, 2006a). Though there is no bedrock topography available for

our site, it was clear that aspen soils are much deeper than their shallow and rocky

conifer counterparts, leading to less storage volume in conifer soils. While considering

only the top 12 cm of soil, it quickly becomes clear that deeper soil horizons need to be

investigated if we are to draw conclusions about water movement through the hillslope.

Digital elevation models (DEMs) are a tempting source of hydrological data because they

have become so readily available in recent years. However, it is imperative to remember

that soil moisture variability should be studied at all depths and not only the topmost

horizon captured in remote sensing and often represented in DEMs (Hupet and

Vanclooster, 2002).

Hypothesis (2), that hillslope-stream connectivity will occur after a threshold of

soil moisture content is reached and that this connection will last longer in the aspen

hillslope due to greater snow melt inputs, cannot be upheld since no hillslope-stream

connectivity was detected in the top 12 cm of soil. However, it is clear that some

relationship between soil moisture and streamflow is present. Hypothesis (3) was upheld;

subsurface lateral flow connected the stream and hillslope only during the wettest spring

snow melt periods, while for the remainder of the year the hillslope and stream were

disconnected and streamflow did not react to surface water inputs. But one must look

deeper within the soil horizons in order to understand the timing of hillslope-stream

connectivity. Our study site is consistent with others in dry climates, where shallow

subsurface flow connects hillslope water to stream water for very short period of extreme

wet conditions (Weiler et al., 2005; Newman et al., 2006).

Other pathways to the stream

In the study catchment I observed soil moisture in both hillslopes reacting to water inputs after the stream has already begun to rise. In the both hillslopes the soil moisture response lagged streamflow response by roughly 65 days. In a Chilean mountain catchment, Blume et al. (2007) also found that soil moisture response lagged streamflow, noting that fast streamflow response indicates preferential flows. This means that in our catchment, other pathways were contributing to streamflow before subsurface lateral flow within the top 5 cm of the soil matrix. To understand woody vegetation effects on hillslope scale processes, subsurface flow through the root zone, a possible dominant pathway to the stream, must be understood (Huxman et al., 2005). We must also understand what other flowpaths to the stream are possible, or significant. Water arriving at the stream via overland flow is essentially ineligible for uptake by plants since it travels relatively quickly and does not pass through the root zone. So changes in woody vegetation will have little influence on overland flow inputs to the stream. Overland flow also has implications for water quality and erosion (Fitzjohn et al., 1998). Though many researchers working in humid forested systems do not observe overland flow and consider it negligible (Hewlett and Hibbert, 1963; Freer et al., 2002; Uchida et al., 2005), it has been found to be a major contributor to runoff in semi-arid systems (Newman et al., 1998).

Overland flow was observed (but not quantified) in the lower portions of the catchment on the rising limb of the hydrograph, May 20th through 31st, the period of highest surface water input during for the 2008 water year, but had subsided completely by June 7th (Figure 2.20). By comparing the maximum rate of snow ablation to the

minimum infiltration rate for each soil type in this study, it is clear that infiltration

excess overland flow is not likely. Even combining the maximum daily surface water

input of 85 mm (May 20th of 2008) with the minimum infiltration rate of 0.42 mm/min in

the conifer stand, complete infiltration of a day's worth of water inputs would only take

3.4 hours. This is contrary to surface runoff at a snow-influenced, New Mexican

Ponderosa Pine site where runoff was produced through infiltration (not saturation)

excess overland flow, when surface water inputs quickly exceeded infiltration capacities

of the soil. Frozen soil runoff at the New Mexico site was also produced when

temperatures dropped and soils froze before the snowpack could insulate them. Frozen

soils did not allow water inputs to infiltrate and surface flow was produced (Wilcox et al.,

1997). In our catchment, periods of subfreezing temperatures in the top 5 cm of soil

began as early as November 4th while significant snow cover was not present until early

December (Figure 2.3). This leaves overland flow due to frozen soils as a possibility but

further investigations would be needed at our site before conclusions can be drawn about

overland flow contributions to the stream.

Overland flow is not the only other possible path for the stream to receive water

from the landscape. Lateral flow is also possible through the snowpack, leading to

moisture accumulation down slope. Using artificial snow profiles, dye tests, and

moisture sensors, irregular melt water inputs (for example from the result of a tree

crown), changed crystal sizes and formed ice layers, transformed the snowpack from

homogeneous to heterogeneous, leading to paths of preferential flow (Waldner et al.,

2004). These irregularities within the snowpack can lead to horizontal flow, an example

of which is pipes of preferential flow near the base of the snowpack (Marsh, 1999).

Water chemistry suggests that deep groundwater sources are also contributing to baseflows in the study catchment (Shakespeare, 2006). This is expected in dolomite/limestone geology where leaky bedrock will not confine flows within the soil profile but rather allow for percolation past the root zone, leading to groundwater recharge. Bedrock permeability has the capacity to negate lateral flow on the time scale of a precipitation events (Weiler et al., 2005). The effects of vegetation on water yield would then depend of how much water the plants can use before it has percolated out of the bottom of the root zone.

CONCLUSIONS

This study showed that soil moisture variability decreased as average soil moisture became drier. Soil moisture patterns showed spatial organization lasting longer in the aspen hillslope. This spatial organization was linked to higher water inputs in the aspen stand due to less canopy interception. Snow surveys showed that SWE accumulation was greater in the aspen stand than in the conifer and supported previous observations that the difference in SWE accumulation between the vegetation types is greater in years of greater annual snowfall.

Hypothesis (1), that soil moisture patterns will remain organized longer in the aspen hillslope due to greater snow melt inputs, was supported. However, shallow soil layers from neither hillslope appeared to be contributing to the stream, since baseflows had resumed by of the spatial organization seen in the aspen hillslope. Hillslope-stream connectivity was inferred from soil moisture thresholds, past which streamflow greatly increased. Hypothesis (2), that hillslope-stream connectivity will occur after a soil

moisture threshold of some soil moisture content/soil water potential above and that

this connection will last longer in the aspen hillslope due to greater snow melt inputs, was

not supported because no hillslope-stream connectivity was detected in the top 12 cm of

soil. Hypothesis (3), that subsurface lateral flow will connect the stream and hillslope

only during the wettest periods of the spring snow melt while for the remainder of the

year the hillslope and stream will remain disconnected and streamflow will not react to

surface water inputs, was confirmed. Hydrograph peaks provided an average of 68% of

total annual runoff.

A full understanding of how this system cycles water is needed before more

definitive conclusions can be drawn. Overland flow was observed contributing to the

hydrograph peak but needs to be quantified. We also need to understand how much snow

is sublimated from the conifer canopy, how much water is translocated downhill through

the snowpack before the spring melt and how much water reaches the stream via other

pathways such as overland flow and flow from deep groundwater sources. To understand

subsurface flow as a pathway, we cannot assume that shallow soil horizons are reflective

of the entire soil profile and that vegetation effects on hydrological processed will be

represented there. Investigations into deeper layers of the soil are called for.

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CHAPTER 2

SUBSURFACE LATERAL FLOW AND HILLSLOPE-STREAM CONNECTIVITY IN

ASPEN AND CONIFER-DOMINATED HILLSLOPES

ABSTRACT

Conifer invasion is considered a significant factor in the declining extent of aspen cover in Utah. A consequence of this change in vegetation may be lower water yields from headwater catchments. This study was an effort to understand and compare subsurface flow through two hillslopes representing the end point communities of aspen and conifer. Snow accumulation, temporal patterns of soil moisture profiles, trenchflow, and streamflow were compared between aspen and conifer stands on hillslopes feeding a first order stream in Northern Utah. Peak snow water equivalent was measured along uphill transects in 2007 and 2008, under both vegetation types. To capture temporal patterns in soil moisture profiles, transects of water content sensors were installed in each vegetation type. Transects of eight nests extend uphill from the stream on each hillslope. Nests consist of three probes at 5, 20 and 100 cm or as deep as the probe could be installed (45 to 110 cm). Trenches were dug at the base of each hillslope to quantify subsurface lateral flow leaving the hillslope. Streamflow was monitored at three points in the stream to isolate contributing areas of each vegetation type. Soil profiles under the two vegetation types were found to vary, with aspen having a deeper profile and conifer having a shallow and extremely rocky profile. While snow accumulation was greater in 2008 than 2007, surveys from both years revealed more snow on the aspen hillslope. Two years of hourly soil moisture showed that fall precipitation did not penetrate the

deeper layers of the aspen soil profile (60 to 100 cm). Spring snow melt saturated profiles in both vegetation types and sensors at all depths reacted to melt water inputs. Soil moisture thresholds, beyond which streamflow greatly increased, were exceeded for longer duration in the aspen hillslope. However, trenchflow revealed that subsurface flow at ~50 cm was sustained longer in the conifer hillslope. This study showed that for most of the year, water inputs remain in the soil profile and only during the spring melt is there a chance for percolation to deeper groundwater storage or for lateral fluxes downslope.

INTRODUCTION

Mountain headwater catchments in the semi-arid Intermountain West are important sources of surface water because these high elevations receive more precipitation than neighboring lowlands (Grant et al., 2004). The hydrology of these catchments is especially important as the region faces water shortages and conflicts. Snow melt typically makes up 80% or more of streamflow in the semi-arid West. Snow deposition and melt in montane catchments are largely functions of topography and vegetation (Marks et al., 2002). The snow-driven system of the semi-arid Intermountain West is understudied, leaving a gap in our knowledge of how watersheds function and how runoff is produced (Grant et al., 2004).

Understanding how biotic and hydrologic components of an ecosystem affect one another will help with forecasting the nature and magnitude of environmental changes that affect natural resources (Newman et al., 2006). In locations across the Intermountain West conifer species are encroaching on aspen stands (Bartos and Campbell, 1998).

Possible reasons for this encroachment include human alterations in disturbance regimes, such as fire suppression and grazing, as well as climate change (Bartos and Campbell, 1998; Brown et al., 2006; Shepperd et al., 2006). Utah holds 25% of the aspen in the western United States but it was reported that, since European settlement, there has been at least a 59% decline in aspen domination in Utah and a 66% decline in the Wasatch Cache National Forest (Bartos, 2001). Conifer encroachment into aspen stands results not only in a decline in the species diversity that accompanies aspen stands but also in alterations in water yield and nutrient cycling. Though the exact mechanisms are not known and appear to be site specific, conifer encroachment is blamed for declining water yields and water quality (Shepperd et al., 2006). Bartos and Campbell (1998) claim that for every 1,000 acres of aspen that are replaced by conifers, 250 to 500 acre feet of water is lost as transpiration. The details of streamflow generation must be understood in order to comprehend the role of woody vegetation encroachment (Huxman et al., 2005).

Subsurface flow is lateral movement of water through near surface soils, regolith and bedrock (Newman et al., 1998). It is also called interflow, lateral flow and soil water flow. In dry systems subsurface lateral flow occurs under only extremely wet conditions, when matric potential is above zero (Weiler et al., 2005). Subsurface lateral flow is commonly initiated when water, percolating through a soil profile, meets an impeding layer of less conductive soil or bedrock and is diverted laterally, downslope. Soil hydraulic conductivity decreases with depth, often shifting water fluxes from vertical to lateral (O'Loughlin, 1986). Hillslope scale hydraulic connectivity is needed for the switch from vertical to lateral flow to happen at this scale, causing widespread lateral

flow (McNamara et al., 2005). Commonly observed subsurface flow on steep, wet hillslopes has been through discrete soil pipes or macropores, most prevalent at the soil bedrock interface or above impeding layers (Sidle and Noguchi, 2001; Uchida et al., 2005; Tromp-van Meerveld and McDonnell, 2006b). Understanding water storage and the rate of water movement underlies all efforts to manage water yield (Hewlett and Hibbert, 1963). Hillslopes are the fundamental units of landscape and their high degree of hydrologic complexity is due to many processes interacting at different scales, leading to a diversity of flowpaths and residence times (Lin et al., 2006). Subsurface flow has water quality implications because flowpaths determine how long water remains in the subsurface and consequently what its chemical make up will be. Subsurface lateral flow can also cause positive pore pressure within the soil, leading to an increased risk of landslides (Weiler et al., 2005). When considering vegetation effects on streamflow generation, it is important to focus on subsurface lateral flow because this flowpath leads water through the root zone, where it is vulnerable to plant uptake. The more dominant subsurface runoff paths are in a system, the greater the expected impact of woody plants on streamflow (Huxman et al., 2005). Especially in water limited landscapes, the type and pattern of woody vegetation affects stream-water and groundwater recharge (Newman et al., 2006). Subsurface storm flow controls runoff generation in many part of the world (Uchida et al., 2004). Despite its importance, lateral subsurface flow in semi-arid regions has been under studied (Newman et al., 2004).

Objectives and hypotheses

The objective of this study was to determine the timing and quantity of subsurface flow within hillslopes or between hillslopes and their adjacent stream using hourly soil water contents measured in transects across conifer and aspen hillslopes, combined with streamflow, trenchflow and surface water inputs. More specifically the objectives were to:

1. Compare soil moisture and trenchflow to streamflow to assess when the hillslope is connected to the stream and compare the timing between soil/vegetation types.

2. Find any thresholds in the hillslope past which subsurface flow contributes to the stream and compare these between soil/vegetation types.

Hypotheses:

1. Hydraulic connection within the hillslopes during the spring snow-melt will have longer duration in the aspen than the conifers hillslope because of greater surface water inputs.

2. A soil water content threshold for subsurface flow will be exceeded for greater time periods in the aspen stand due to greater surface water inputs.

The influence of vegetation on the hydrologic cycle

Vegetation influences the hydrologic cycle in multiple ways: by affecting soil properties and ground cover and therefore influencing infiltration capacity (Schume et al., 2003; Williams et al., 2003), by creating macro-pores and altering subsurface flowpaths (Schume et al., 2003; Williams et al., 2003), by water use through transpiration (Williams et al., 2003; Huxman et al., 2005) and by holding precipitation in the canopy and altering

throughfall patterns (Schume et al., 2003; Western et al., 2004; Huxman et al., 2005). In water limited systems, the type and pattern of woody vegetation influences stream and groundwater recharge, nutrient cycles and biodiversity (Newman et al., 2006). Vegetation is, in turn, affected by hydrologic regimes, soil properties and climate (Tromp-van Meerveld and McDonnell, 2006a).

Many factors control snow retention and loss by sublimation in snow-driven systems, including weather patterns and terrain. Forest canopy characteristics are dominant factors affecting throughfall patterns and snow accumulation in mountain headwater catchments. Snow in exposed regions of the semi-arid Owyhee mountains was more vulnerable to turbulent air currents (the primary source of melt energy during the winter) as well as solar radiation (the primary source of melt energy in the spring), melted sooner and provided less surface water inputs overall. Drift and shelter areas provided much more water inputs and the snow lasted later in the season (Marks et al., 2002). Hiemstra et al. (2002) also found that trees helped retain snow in the high elevation tree line zone (3100-3300 m) of the Medicine Bow Mountains in Wyoming, where snow makes up 50 to 80% of precipitation. However different canopy types, as in the case of comparing aspen to conifer stands, may also differ in their ability to retain precipitation. In a rain-dominated, mixed stand in Southern Austria, interception loss by the coniferous spruce canopy was much greater than that of the deciduous beech canopy, causing uneven recharge patterns (Schume et al., 2003).

Previous investigations in Northern Utah indicate that snow accumulation was greater under aspen stands compared to conifer stands (LaMalfa and Ryel, 2008). Conifer species retain their needles year round and thus are able to accumulate more

snow in their canopies, lessening the snowpack on the ground around them. Snow held

in the canopy is more susceptible to sublimation; a water loss to the system, or simply

redistribution, possibly to nearby aspen stands. LaMalfa (2007) also reported that aspen

stands had a 40-70% greater potential for generating runoff and groundwater recharge.

Vegetation changes are expected to have a greater impact on streamflow

generated by subsurface lateral flow, which can only reach the stream if it is not first

taken up by roots (Huxman et al., 2005). The majority of water uptake by vegetation is

performed by fine, young roots in the upper soil horizons where nutrients are more

abundant. However, roots can also be found in deep layers, even in bedrock, especially

in water limited areas (Bond et al., 2007). Patterns of water depletion from the soil

profile indicate water uptake and root activity (Caldwell and Virginia, 1989). Rooting

depth determines the volume of soil from which trees access for water uptake (Schume et

al., 2003). Reproduction of aspen is primarily asexual, by means of root sprouts (Bartos,

2001). These lateral roots are normally found within a foot of the soil surface and can be

inhibited by subsurface rocks (Shepperd et al., 2006). In contrast, conifer species have

deeper roots, able to reach lower sources of groundwater compared to the relatively

shallow rooted aspen. Average maximum depth for temperate conifer roots was 3.9 m

and only 2.9 m for temperate deciduous roots (Canadell et al., 1996).

Evapotranspiration can have direct impacts on streamflow. In a humid, Pacific

Northwest fir catchment, Bond et al. (2002) demonstrated the link between streamflow

and vegetation by comparing diel patterns of transpiration and streamflow. According to

Huxman et al. (2005), the effect of woody plants on streamflow generation will depend

on whether the woody vegetation is found in riparian or hillslope areas, whether surface

water inputs are supplied during the growing season and whether or not there is a potential for bypass flow such as with fractured or Karst bedrock. The more dominant subsurface runoff paths are, the greater the expected impacts on streamflow by woody plant encroachment.

Aspen and conifer have differences in water use through transpiration, therefore their effects on streamflow may be different. Part of the competitive strategy of evergreen trees is to start water uptake as soon as climatic conditions allow (Schume et al., 2003). This coincides with LaMalfa's (2007) findings that conifer stands in Northern Utah had longer transpiration activity periods relative to nearby aspen, starting 5 days before aspen and ending 24 to 31 days after aspen. However, in terms of annual water use, this longer transpiration period in conifer seems to be offset by higher transpiration rates during leaf expansion of aspen or deciduous species. Shume et al. (2003) found that when beech used water, they use more than nearby spruce, which were able to conserve water by closing their stomata during dry times. Aspen showed less ability to regulate water loss through stomatal closure than neighboring pine and fir species in Wyoming, where aspen sap flux increased during the early season leaf-out period, leaf production continued well into the growing season (Pataki et al., 2000). In Northern Utah, aspen evapotranspiration was 25% higher than conifer during leaf out, most likely due to more water in the aspen soil profile (LaMalfa, 2007). Both Pataki et al. (2000) and LaMalfa (2007) found that soil moisture and water movements to both species regulate transpiration. By the end of the water year, both aspen and conifer had used all the profile water available to them (LaMalfa, 2007).

Mechanisms of subsurface lateral flow

How and when lateral subsurface flow occurs is a product of catchment characteristics and will affect the timing and quantity of water delivery to a stream via this pathway. This flowpath is of interest because it shares space with the root zone, therefore a shift in vegetation type may affect how much water is contributed to the stream. Factors controlling subsurface lateral flow include macropores and other preferential flowpaths (Newman et al., 1998; Sidle and Noguchi, 2001; Tromp-van Meerveld and McDonnell, 2006b; Blume et al., 2007), hillslope permeability (Uchida et al., 2004), bedrock topography (Freer et al., 1997; Chaplot and Walter, 2003; Tromp-van Meerveld and McDonnell, 2006b) and the amount and intensity of water inputs (Newman et al., 1998; McNamara et al., 2005; Tromp-van Meerveld and McDonnell, 2006b).

Macropores, formed from decayed and live roots, subsurface erosion, bedrock fractures, and animal burrows are often the main sources of lateral preferential flow (Sidle and Noguchi, 2001). At the Panola trench, subsurface flow was delivered through pipes and macropores sooner than it was through the soil matrix (Tromp-van Meerveld and McDonnell, 2006b). In a humid Chilean catchment with high porosity, dye tracers showed preferential flowpaths leading from the surface to an approximate depth of 1 m. Heterogeneous throughfall, along with hydrophobic soils, caused fingers of saturation that acted as preferential flowpaths. Soil moisture probes throughout the catchment confirmed that this point scale preferential flow was of importance for hillslope or catchment scale processes (Blume et al., 2007). Though the hydrology, subsurface lateral flow in particular, of semi-arid systems is expected to be different from that in humid

regions, Newman et al. (1998) argued that macropores are important in both regimes. They collected lateral flow with PVC pipe collectors and a trench on a semi-arid Ponderosa pine hillslope in New Mexico to determine that lateral subsurface flow from the hillslope was episodic, with the largest events (90% of all flow) in the spring snow melt. The majority of this flow (80%) came through the macropore ridden B-horizon.

In two zero-order humid catchments in Japan, hillslope discharge was examined as a function of upslope pore pressure and upslope subsurface saturated area using tensiometers and a V-notch weir. During stormflow periods, hillslope discharge at both sites was strongly related to both upslope pore pressure and upslope saturated area. However, between stormflow periods, this relationship only held true for the catchment with high hillslope permeability and not for the catchment with lower permeability (Uchida et al., 2004).

After examining digital terrain maps, trenchflow and point scale soil moisture measurements at humid catchments in Georgia and New Zealand (Panola and Maimai), Freer et al. (1997) comment that, although many models assume that hillslope flow is controlled by surface topography, it is really subsurface topography, or bedrock topography that controls this flowpath. In a first order catchment of Western France, digital elevation models of both surface and subsurface topography were considered and lateral subsurface flow was found to be more affected by subsurface than surface topography (Chaplot and Walter, 2003). Subsurface stormflow, in the form of transient saturation at soil-bedrock interface, was studied using a hillslope trench and shallow groundwater wells at the humid Panola catchment in Georgia. Bedrock depressions had to be filled before spilling over, causing subsurface saturation to become connected to the

trench face, where flow was observed to increase more than five-fold during these times of hillslope connectivity. Subsurface lateral flow was therefore controlled by the amount of precipitation as well as the relatively shallow depth to bedrock (~1 m) (Tromp-van Meerveld and McDonnell, 2006b).

Water inputs must be great enough to satisfy soil water deficit, plants needs and overcome capillary forces holding the water in place before preferential flow is initiated, and/or subsurface saturation accumulates enough to cause lateral fluxes of hillslope water to the stream. This is most common in areas with large or intense periods of rain or snow melt (Newman et al., 1998). In a snow-driven watershed in Idaho, McNamara et al. (2005) found that zones of hydrophobicity, along the soil/bedrock interface, prevented hillslope connectivity until energy inputs were great enough to provide sufficient snow melt to overcome the dry zones, which only happened for a short time out of the year. Subsurface lateral flow was therefore driven by a combination of surface water inputs and bedrock topography. In the humid Panola catchment, subsurface lateral flow was controlled by the relatively shallow depth to bedrock (~1 m) as well as by the amount of precipitation the hillslope received (Tromp-van Meerveld and McDonnell, 2006b).

Subsurface lateral flow thresholds

Rather than the initial variable source area concept, developed in the 1960's and 1970's, in which a stream's source area gradually expands and contracts, it has been recognized that there is often a threshold, after which hillslopes "turn on" and subsurface lateral flow contributes hillslope water to the stream. This hillslope input to the stream, or hillslope-stream connectivity, is generally greater on the receding limb of the hydrograph

(McDonnell, 2003). At Panola, Tromp-van Meerveld and McDonnell (2006b) found that 55 mm of rain were needed to fill bedrock depressions before significant subsurface storm flow began. After this precipitation threshold, subsurface flow at the trench face increased more than 75 times that when the hillslope was not connected. At Reynolds Mountain East in the Owyhee Mountains of Idaho, Grant et al. (2004) found a soil moisture threshold, around field capacity, over which streamflow became highly reactive to soil water inputs. At soil water contents higher than 33%, major increases in lateral flow were seen in a semi-arid Ponderosa Pine hillslope in New Mexico studied by Newman et al. (1998). Ninety percent of lateral flow, measured at a trench, happened beyond this threshold, mostly via macropores, which shunted water from the surface to the B-horizon, often bypassing the A-horizon. Thresholds provide a simplified way of describing hillslopes and means for comparing them to one another, rather than describing a single hillslope, which may be an anomaly (Tromp-van Meerveld and McDonnell, 2006b).

Individually describing the different mechanisms and thresholds leading to subsurface lateral flow is not meant to imply that they are mutually exclusive. Multiple factors, on multiple scales, simultaneously affect subsurface lateral flow. For example in Panola, Georgia, Freer et al. (2002) described how macropores pipe water quickly to the soil bedrock (or impeding layer) interface, where it forms a perched zone of saturation and is transported quickly to the stream. Soil properties and vegetation also help dictate when subsurface lateral flow will happen and after what thresholds.

METHODS

To compare hillslope connectivity and subsurface lateral flow between the aspen and conifer hillslopes, streamflow, soil moisture and trenchflow were continuously monitored. Water inputs to the catchment were quantified using precipitation data from the Natural Resource Conservation Service (NRCS) Snow Telemetry (SNOTEL) sites as well as snow surveys at peak snow accumulation. Infiltration and other soil physical properties were measured as possible explanatory variables for hillslope-stream connectivity timing and quantity.

Site description

The study site is located in Northern Utah in the headwaters of the Ogden River along the eastern edge of Weber County (Figure 2.1). The study catchment lies roughly at UTMs 0461515, 4579126, inside a larger paired watershed study on the Deseret Land and Livestock Ranch. These mountain headwaters, around 2500 meters in elevation, feed Pineview and Causey reservoirs, which are used for recreation, water storage and irrigation. According to the SNOTEL data, the average annual temperature is 4.3°C, with an average annual maximum temperature of 29.3°C and an average annual minimum temperature of -21.0°C. Average annual precipitation is 861 mm with approximately 74% falling as snow in the winter months.

Previous studies

The study catchment in contained within a larger paired watershed study on the Deseret Land and Livestock Ranch. Bear and Frost canyons are the two small, adjacent,

paired watersheds with a combined area of 26.7 km^2. In 2005 and 2006, snow water equivalents (SWE), snow density, sublimation, condensation, soil moisture and sap flux in aspen and conifer stands on north facing slopes of Frost watershed were monitored (LaMalfa, 2007). Though sublimation was not significantly different from the ground under the two vegetation types, aspen plots had 34-44% higher SWE at the time of peak snow accumulation than adjacent conifer plots. The greater SWE in aspen stands, combined with longer transpiration periods in Conifers, lead to 42-83% more potential runoff/groundwater recharge from aspen stands than from neighboring conifer stands (LaMalfa, 2007). In Frost canyon, hillslope gradient was the most important factor for flow gains and was negatively correlated. Steeper hillsides, with more exposed bedrock and less ability to retain water, yielded less water during low flows. Water chemistry suggested that the primary source of baseflow in Frost canyon is deep groundwater and that Karst geology may contribute to heterogeneity of streamflow. Flow at the confluence of Bear and Frost canyons showed diel patterns after snow melt was complete, indicating vegetation influence on baseflow (Shakespeare, 2006). Other complementary research is ongoing in the watersheds including a study to estimate residence times and flowpaths in Bear canyon. To better understand the effects of conifer encroachment on water yield, Bear canyon will be treated by removal of conifers while Frost canyon will be left as a control.

Instrumentation

Bear and Frost catchments have been instrumented with weather and stream stations since 2004 and measurements are recorded each hour onto a datalogger (CR10X,

Campbell Scientific, Logan, UT, USA). The streams are monitored just before their

confluence for stage (PDCR 1830-8388, Druck, Melrose, MA, USA), temperature and

electrical conductivity (CS547A, Campbell Scientific, Logan, UT, USA). Weather

stations at the top of Bear canyon and at the confluence monitor relative humidity, air

temperature, snow depth, precipitation, solar radiation, wind speed and soil temperature

at 10, 20 cm. Data is also available from nearby SNOTEL sites (Figure 2.1). These three

SNOTEL sites were chosen because of their proximity to the study catchment and their

similar elevation. Lightening ridge SNOTEL sits on the west ridge of Frost canyon, at

2504 m above sea level and has been installed since 2004. Horse ridge SNOTEL is

located 2 km south of Bear canyon, at 2487 m and has been collecting data since 1978.

Dry Bread Pond SNOTEL site is 7.4 km southwest of the study catchment, and 2545 m

and has been recording data since 1978. All three stations measure hourly air

temperature, snow depth, and precipitation.

Catchment/hillslope specifics

      The study catchment is located in Frost Canyon and contains both aspen and

conifer hillslopes adjacent to a first order perennial stream (Figure 3.1). The catchment

drains to the north with an area of approximately 1 km^2 and elevations ranging from 2464

to 2691 m. The upper hillslopes are dominated by mature Quaking aspen, (*Populus*

tremuloides), with a lush under story made up primarily of Snowberry (*Symphoricarpos*

albus), Serviceberry *(Amelanchier alnifolia),* Ragwort *(Senecio serra),* Wild Geranium

(Geranium richardsonii), Mountain Bluebells *(Mertensia ciliata)* and Horse Mint

(Agastache urticifolia). Average slope on the aspen hillside is 39%, with a confined V-

shaped riparian area and a northwest aspect. Sagebrush grassland is found in patches

of the upper hillslopes and on the lower eastern hillslope of the catchment. The lower

western hillslopes are dominated by mature conifer species consisting mainly of White fir

(*Albeis concolor*), Douglas Fir (*Psuedodotsuga menziesii*), and Engelmann Spruce

(*Picea, engelmannii*). The conifer under story is mainly sparse Goose Berry (*Ribes

montigenum*), Snowberry (*Symphoricarpos albus)* and grasses. Average slope on the

conifer hillslope is 40% with a northern aspect. The riparian zone is U-shaped and less

confined than the upstream aspen section. The aspen's under story is much denser than

the conifer's and varies yearly depending on water availability. It is common to find

different soil characteristics under different tree species, in part because of the species'

ability to alter soil properties. Since the aspen and conifer stands at the study site have

very different soil profiles, this study addresses the combined influence of vegetation and

soil on subsurface flow. The study area lies between the Rich and Weber County Soil

Surveys and no soil survey information is available for the exact site. According to the

nearby Rich County Soil Survey of 1982, soils in similar areas range from mollisols to

entisols and inceptisols depending on slope and landscape position (Soil Conservation

Service, 1982).

Water inputs

To understand snow accumulation within the study catchment and its variance

with hillslope position and vegetation type, snow water equivalents (SWE) and snow

depth were measured in transects starting at the stream and moving uphill from the

stream in four meter increments. Snow surveys were supplemented with recent surveys

in nearby aspen, conifer and open plots. SWE was measured as close to peak snow
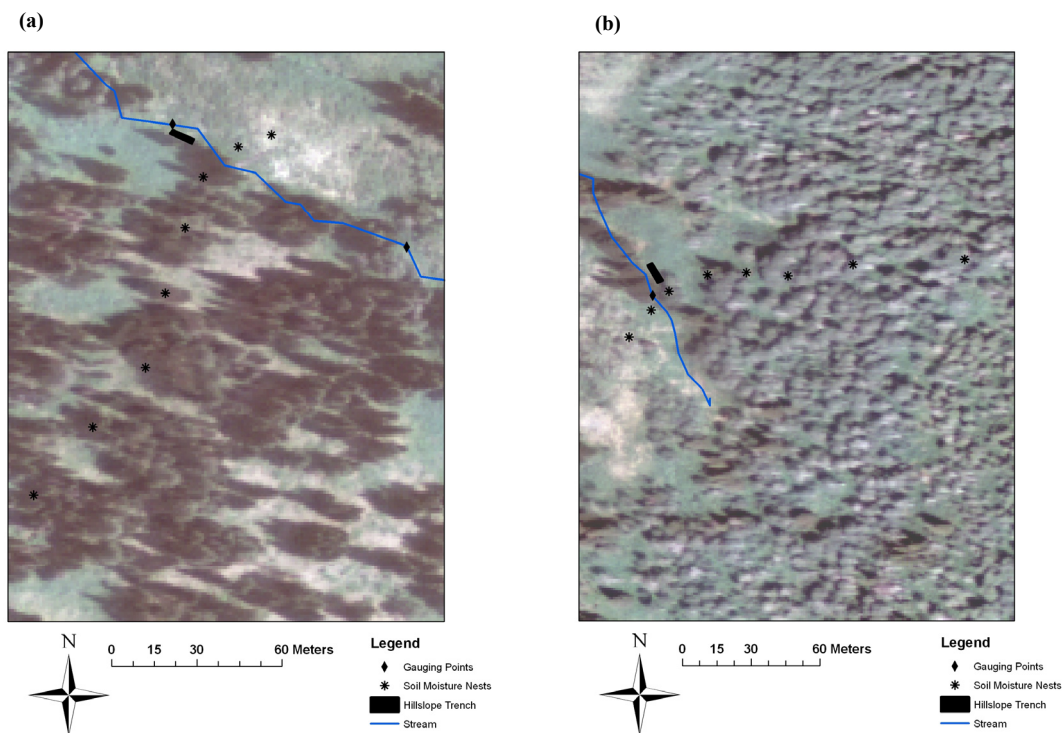
Figure 3.1. Arial photographs of the study hillslopes (a) conifer, (b) aspen. Dots indicate location of transects of soil moisture sensor nests. The blue line indicates the stream.

accumulation as possible, on March 16th, 2007 and April 6-12th, 2008. Standard snow sampling tubes (Carpenter Machine and Supply, Seattle, WA, USA) were used to measure SWE and snow depth. Tared snow tubes were inserted vertically into the snowpack until soil was reached. Snow depth was recorded and the snow-filled tubes were weighed after any soil on the bottom of the sample was removed.

Daily snow pillow and rain gauge data were used from the three nearby SNOTEL sites, chosen for their proximity and similar elevation to the study site. Rain events on days with an average temperature below 0°C were considered invalid and disregarded.

Since snow is not an immediate water input to the soil, snow ablation (measured by the SNOTEL snow pillows) and rain were combined to make surface water inputs (SWI).

Infiltration

Infiltration capacity in both the aspen and conifer forest floor was measured using a double ring infiltrometer on June 27th and 28th, 2007 and again on June 14th, 2008. The infiltrometer was inserted into the ground until water added inside the rings could not leak out and was forced to infiltrate into the soil (a depth of approximately 5 to 10 cm). Water levels in the outer ring were kept constant so that water infiltrating in the inner ring was maintained as a one-dimensional infiltration process. Water was added to the inner ring and the time the water took to drop 1 cm was recorded repeatedly until a constant rate was reached. Measurement locations were chosen as randomly as possible though relatively flat areas were required for the infiltrometers to be used. Two readings were taken in 2007 (four in aspen, two in conifer) and 10 were taken in 2008, five in each vegetation type. Cumulative infiltration was graphed against time. Infiltration capacities are considered the rate of infiltration once a constant rate had been reached.

Soil physical properties

Soil texture from 44 randomly chosen grid points (22 from each hillslope) was determined using the hydrometer method. Samples were sieved to 2 mm and 50 g of each was suspended in deionized water. The solutions were shaken with 50 ml of 10% Calgon (a dispersing agent) for 12 hours. Deionized water was then added to each sample until there was a total volume of 1 L in each graduated cylinder. A hydrometer

was placed in the cylinder and readings were taken after the sand has settled at 70 seconds and after the silt had settled after 7 hours.

To determine bulk density, between 150 and 600 ml of soil were collected, dried and weighed. The subsequent holes were then filled with a known volume of sand. Rock content by volume was measured from 21 bulk density samples from multiple depths. To determine rock volume, particles greater than 2 mm from each sample were submerged in a partially filled graduated cylinder and the subsequent increase in volume was recorded.

Soil moisture

Soil moisture was continuously monitored at 16 locations in the study catchment (Figure 3.1). Nests of 30 cm soil water content reflectometer probes (CS616, Campbell Scientific, Logan, UT, USA) were installed in the fall of 2006. Two transects of 8 nests were installed, one transect across an aspen dominated portion of the catchment and the other across a conifer dominated portion as mapped in Figure 3.1. Each nest consists of 3 probes at depths of 5, 20 and 100 cm in aspen transect, and at 5, 20, 45-60 cm in the conifer transect. A fractured rock layer underlying the conifer site prevented probe installation at greater depths.

CS616 sensors are transmission line oscillators, which operate similarly to time domain reflectometers but do not have a separate pulse and sampling units. Instead, consecutive voltage pulses are generated inside the probe head, with each pulse being triggered by the arrival of the reflected previous pulse. CS616 sensors function best with calibration curve for each soil being measured (Kelleners et al., 2005). In this study, a calibration curve was made for each of the two soil types, i.e. aspen and conifer, by

packing a soil column to approximately the same bulk density as was found in the field. The soil was then saturated and the CS616 inserted. As the soil dried, it was repeatedly weighed and measured with the CS616 (Figure A1, Appendix). The 30 cm CS616 has an accuracy of ±3%, although insertion error is possible if the prongs are inserted in such a way that void space is created around them.

Because CS616 probes have exhibited some temperature dependence (Campbell Scientific, 2004), 14 temperature sensors (temperature pendants, HOBO, Pocasset, MA) were buried at a variety of hillslope positions. Seven were buried in each vegetation type, four at a depth of 5 cm and two at 20 cm. These sensors log temperature hourly and were installed to find any freezing pattern of soils.

Trenches

Two hillslope trenches were dug in the summer of 2007, one in each the aspen and conifer stands (Figure 3.1). The trenches were dug 7 m long, 1.5 m wide and 1 m deep. The conifer trench reached fractured bedrock at a depth of ~50 cm. The aspen trench did not reach bedrock even after a depth of 1.5 m. In the aspen trench collection systems were inserted into the face of the trench at depths of 20 and 50 cm. Sections of sheet metal and pipe, 15 cm long and inserted at least 5 cm into the upslope wall of the trench. The upper layer collection system collected water from a total cross-sectional area of ~6200 cm^2. While the lower collection system collected flow from a cross-sectional area of ~ 2100 cm^2. Water from each of the collection layers was routed into a tipping bucket (Rain Wise Inc., Bar Harbor, ME, USA) and logged onto a datalogger (HOBO Event Data Logger, Bourne, MA, USA). The tipping bucket recorded each time that 10

ml had accumulated in the bucket. In this way water flowing down-slope through the subsurface was collected, quantified and the timing of its flow recorded.

A similar system was installed in the conifer trench. However, due to the extremely rocky soil, only one layer of collection system was installed at ~20 cm depth with a total cross-sectional area of ~900 cm^2. Each trench was covered with a sheet metal roof to keep snow and rain out. Sheet metal was also inserted into the upslope soil surface so that any potential overland flow would be diverted over the roof and not down the trench face. In the spring, the conifer trench was found full of water and the tipping bucket floating. Once the water had receded, the underground flow was exposed and flow from the conifer trench face was measured by hand on May 31st. A 6 cm wide, concave strip of sheet metal was pushed into trench wall, routing flow into a container. The volume of water collected in 15 seconds was used to calculate flow (L s^{-1}) every 50 cm for the length of the trench (12 points total). Flow was collected at depths ranging from 45 to 70 cm, depending on where the highest water yield was found. A water height sensor (Levelogger, Solinst, Georgetown, Ontario, Canada) was installed in the bottom of the conifer trench to monitor water height from June 14th to June 29th. An atmospheric pressure logger (HOBO Pressure, Bourne, MA, USA) was also deployed at the trench to adjust the pressure logger for atmospheric pressure changes.

Stream gauging

Stage was monitored continuously for the 2008 water year using WT-HR water level loggers (TruTrack, Christchurch, New Zealand), which were installed at three points in the stream. The first level logger measured the stage of the stream just below

the contributing area dominated by aspen. The contributing area between the first and second level loggers was aspen/conifer mixed and encompassed a seep and a small tributary to the stream. The portion of the catchment between the second and third level logger was dominated by conifer species. The level loggers recorded hourly readings of water height in mm.

The stream was gauged using slug injection dilution gauging. This approach is well suited for small streams that cannot be otherwise gauged with flow meters (Moore, 2005). In this method, NaCl is used as a tracer and dumped as a slug injection into the stream. Downstream, an electrical conductivity (EC) probe measures the pulse of salt as an increase in EC. The area under the break-through curve represents was calculated to estimate the flow in the stream.

An injectate solution was mixed using 300 to 500 g of NaCl dissolved into 2 to 3 liters of stream water. More salt was used during high flows and less during baseflows. Small amounts of the injectate were added to 1 L of stream water and its electrical conductivity (EC) was measured at each known concentration (YSI 85, Yellow Springs, OH, USA). In this way a calibration curve was made to calculate concentration of NaCl from EC using a calibration constant K. The background EC was recorded both at the upstream and downstream reaches to be gauged before the slug of injectate was dumped. Each of the reaches gauged was long enough for adequate mixing of the salt-water slug with the stream. The slug was measured every two seconds as it passed at the downstream end of the reach by an EC probe (CS547A conductivity and temperature probe Campbell Scientific, Logan, UT, USA) and logged onto a datalogger (CR 510,

Campbell Scientific, Logan, UT, USA). The resulting curve of EC was translated into concentration at each time step using the calibration previously made.

$$C(t)=EC(t)*K \qquad (g\ L^{-1})$$

where C(t) is the concentration at time t, EC(t) is the electrical conductivity at time t and K is the calibration constant obtained from the calibration curve. The background concentration was accounted for and the mass of salt passing at each time step was calculated:

$$\text{Mass NaCl recovered } (t) = (C(t)\text{-}Cb)*d(t) \qquad (g\ L^{-1}s^{-1})$$

where Cb is the background concentration and d(t) is the time step (in this case two seconds). The total mass of NaCl recovered from all time steps was calculated:

$$\text{Mass NaCl recovered} = \Sigma(C(t)\text{-}Cb)*d(t) \qquad (g\ L^{-1}s^{-1})$$

Flow (Q) was calculated by:

$$Q = (L_{injectate}*\text{Concentration}_{injectate})/\text{Mass of NaCl recovered} \qquad (L\ s^{-1})$$

Using the stage measurements from the times of stream gauging, a rating curve was made for each of the three stream gauging sites. Using this rating curve, a hydrograph was made from the continuous stream stage measurements.

<div align="center">RESULTS</div>

Water inputs

Data from nearby SNOTEL sites show that the area received more snow in the 2008 water year than in 2007 (Figure 2.3). The average precipitation since 1990 as measured by three nearby SNOTEL sites was 887 mm a year, with 74% falling as snow. In 2007 the catchment received only 84% of average precipitation: 743 mm, with 66%

falling as snow. In 2008 it received 889 mm, 100% of average precipitation, with 86%

falling as snow. Peak snow accumulation came on March 12th in 2007 and April 13th in

2008. The average day of peak SWE since 1979 is April 4th. Figure 2.3 also shows peak

SWE from snow survey measurements at the study site under or near aspen and conifer

canopies.

In both years the average SWE surveyed under forest canopies was less than SWE

recorded at the SNOTEL sites, which are in clearings (Figure 2.3). A comparison of peak

SWE under aspen and conifer canopies (Figure 2.5) shows that on average, aspen

accumulated more SWE than conifer, and this difference was more pronounced in the

wetter year (2008). A two-sample t-test revealed that in 2007 the aspen stand had an

average of 315 mm of peak SWE on and the conifer stand had 280 mm, an 11%

difference ($t = 2.22$, $df = 72.78$, p-value = 0.0294). In 2008 aspen had 509 mm and

conifer 390 mm, a 23% difference ($t = 5.55$, $df = 153.97$, p-value = 1.178e-07).

Under the conifer canopy there was a notable pattern of drifts forming between

trees or groups of trees. While areas under the densest part of the canopy were free of

snow by May 20th, snow in drift areas lasted well into June (Figure 2.6). Figure 2.7

shows SWE measurements made under ($n = 20$), outside ($n = 51$) or on the edge ($n = 19$)

of the conifer canopy at peak accumulation in 2008.   An analysis of variance (ANOVA)

was used to assess the differences in accumulation among the locations relative to conifer

canopy cover. At the edge of the conifer canopy there was an average of 431.8 mm (SE

= 25.18), which was not significantly different ($p = 0.38$) from the average outside of the

canopy (401.2 mm, SE = 19.92). The average snow accumulation under the conifer

canopy was 323.85, significantly less than both the edge (p = 0.010) and outside (p = 0.025).

Since snow is not a direct water input to the catchment, snow melt (considered snow ablation) and rain measured at the SNOTEL sites were combined to make surface water inputs (SWI), shown in Figure 2.8. These values should be taken as maximum possible surface water inputs since snow ablation could represent sublimation, evaporation as well as melt. The spring pulse of inputs was greater and later in the 2008 water year than in 2007. The maximum daily surface water input in 2007 was 46 mm on April 25th and 85 mm on May 20th of 2008.

Infiltration

Infiltration rates were measured at 16 points, 9 on the aspen hillslope and 7 on the conifer. The minimum infiltration rate found in the aspen stand was 0.9 mm min^{-1} in the aspen stand and 0.42 mm min^{-1} in the conifer stand. Infiltration rates are shown in table 2.1 and a comparison of aspen to conifer average rates is shown in Figure 2.9. A t-test showed that there was no significant difference between average infiltration rates in the aspen (3.32 mm hour^{-1}) and conifer (4.54 mm hour^{-1}) sites (t = -0.49, df = 11.46, p-value = 0.63).

Soil physical properties

Soil textures at varying depths for both vegetation types are shown in Figure 2.10 and Table 2.3. For details on locations where the samples were taken, see Appendix Table A2. The average textural class for soils under both vegetation types was loam, though some lower horizons of the conifer soil were sandy loam. There was a notable

difference between the rock content and soil depth of the profiles. The aspen soil had

an average rock content by volume of 1% (n = 12)) and its depth was greater than 1 m

(bedrock was not reached at this site). The conifer soil was very gravelly, with an

average rock content of 46% by volume (n = 9). The profile became increasingly rocky

with depth, fading into fractured bedrock between 45 and 60 cm. Average bulk densities

show higher bulk densities in the conifer site, most likely due to the greater rock content

(Table 2.3).

A t-test showed that percent cover of leaf litter was not significantly different

between the conifer and aspen sites (t = 0.043, df = 170.29, p-value = 0.97). Aspen had

an average of 51.0% and conifer 50.9%. Depth of leaf litter was significantly greater at

the conifer site (average depth was 1.06 cm) than at the aspen site (average depth was

0.10 cm) (t = -4.068, df = 89.97, p-value = 0.0001).

Continuous soil moisture

Continuous soil moisture at three depths was averaged among all points on

transects in both the aspen and conifer plots (Figure 3.2). Shallower horizons were the

most reactive while lower layers, especially in the aspen profile reacted slower to spring

and fall wetting events. Figure 3.3 shows shallow soil in the profiles reacting to fall

precipitation, while lower horizons remain dry. Peak soil moisture in 2007 came on April

29th on both hillsides. Maximum soil moisture measured by the deepest sensors was 0.35

cm^3 cm^{-3} under aspen and 0.38 cm^3 cm^{-3} under conifer. Peak soil moisture came later in

2008, on May 19th. Lower horizons showed roughly the same water contents (0.36 cm^3

cm^{-3} under aspen and 0.38 cm^3 cm^{-3} under conifer).

Streamflow is plotted with the soil moisture records in Figure 3.2. In the spring

of both 2007 and 2008 water years soil moisture shows an abrupt increase. These sudden

soil moisture increases happen at soil water contents between 18 and 25%. Streamflow

recorded in the 2008 water year shows the stream beginning to peak shortly after these

critical soil moisture levels are reached.

There is only a brief window of time when high water contents are seen

throughout all profiles on a given hillslope (Figure 3.3). These brief periods represent the

wettest and most likely times for soil water to contribute to streamflow or groundwater

recharge. Soils in the winter of 2008 were recorded as notably drier, especially under the

aspen stand. This is contrary to our observation that 2008 was a wetter year. However,

there is a chance that subzero temperatures in 2008 caused the sensors to record

artificially lower moisture contents. Or that frozen soil prevented water from infiltrating

or percolating through the profiles. Periods of subfreezing temperatures in the 2008

water year began as early as November 4th and lasted until April 14th in the conifer and

until May 3rd in the aspen (Figure 2.11). Soil temperatures from the three nearby

SNOTEL sites show only one site briefly dropping below freezing in the fall of 2006.

While in the fall of 2007, all sites fell below freezing due to a drop in air temperatures

before an insulating snowpack could accumulate. Our study site experienced more shade

than the SNOTEL sites and therefore had subfreezing temperatures in the soil for longer.

Trenchflow

The upper level collection system (20 cm deep) in the aspen trench recorded a

very small amount of flow (Figure 3.4). Tipping bucket records from the aspen trench

(a)

(b)
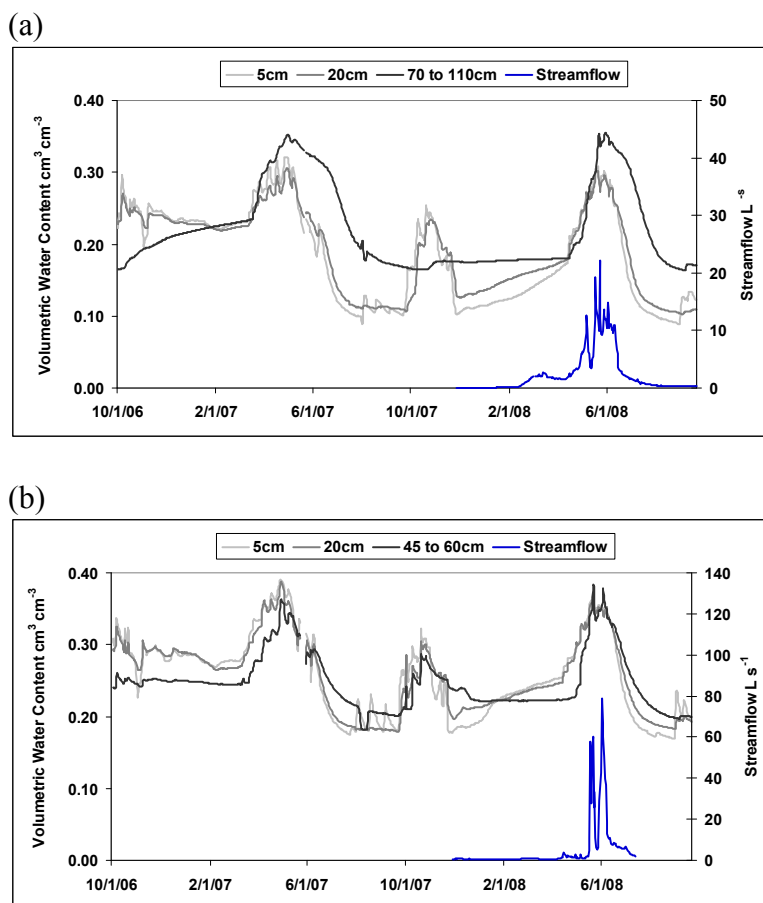
Figure 3.2. Average soil moisture for two water years at three depths in the (a) aspen profile and (b) conifer profile. Blue lines indicate streamflow which was only recorded during the 2008 water year.

collection system at 50 cm show significant flow from the trench face May 15th through May 21st. Clear diel pulses were seen each day. Flow started each day between 2:00 and 3:00pm, peaked between 5:00 and 7:00pm and subsided between 1:00 and 4:00 the next morning (Figure 3.4).

The conifer trench was uncovered on May 28th and found to contain ~1 m of standing water. Judging by when the tipping bucket started recording movement, I inferthat the trench contained a substantial amount of water by April 25th. Once the water

receded, it was clear that the flow was coming from the lower portions of the soil profile, 45 to 70 cm and below. Using flow data measured by hand from the trench face on May 31st, flow from the entire conifer hillslope was estimated at 3.5 L s^{-1}. A water height recorded placed in the bottom of the trench starting June 14th, shows the trench had drained completely by June 28th.

By considering trenchflow initiation as an indicator of subsurface lateral flow, the soil water content threshold for hillslope connectivity can be identified from the continuous soil moisture measurements in the hillslope. Trenchflow began with an average water content of 0.29 cm^3 cm^{-3} in the deepest sensors of the aspen hillslope and 0.22 cm^3 cm^{-3} in the conifer hillslope.

Streamflow

Streamflow, gauged at three points along the stream (Figure 2.2), showed two distinct hydrograph peaks during the spring snow melt. The first peak below the aspen hillslope was 30.8 L s^{-1} on May 18th. The conifer-gauging point peaked at 69.7 L s^{-1} on May 20th (Figure 2.12a). Figure 2.12 also shows surface water inputs (SWI) measured at the three nearby SNOTEL sites. Here SWI was a combination of rain and snow ablation (to represent snow melt). The lull in streamflow between the two peaks can be attributed to a storm, which lowered temperatures from a daily average of 14 to 1°C and left 29 mm of SWE between May 21st and 23rd. The second hydrograph peaks were greater for the middle and conifer gauging points, which peaked at 31.7 L s^{-1} on June 1st, and 82.3 L s^{-1} on June 2nd, respectively. Streamflow below the aspen hillslope did not peak at this time,

(a)

Aspen Profiles, 2007 Water Year

Distance from
Stream (m)

140
90
60
15
5

Oct ⟶ Sept

Time

Volumetric Water
Content
cm3 cm^{-3}

0.3
0.28
0.26
0.24
0.22
0.2
0.18
0.16
0.14
0.12
0.1
0.08

(b)

Aspen Profiles, 2008 Water Year

Distance from
Stream (m)

140
90
60
15
5

Oct ⟶ Sept

Time

(c)

Conifer Profiles, 2007 Water Year

Distance from
Stream (m)

140
90
60
40
15
5

Oct ⟶ Sept

Time

Volumetric Water
Content
cm3 cm^{-3}

0.3
0.28
0.26
0.24
0.22
0.2
0.18
0.16
0.14
0.12
0.1
0.08

(d)

Conifer Profiles, 2008 Water Year

Distance from
Stream (m)

140
90
60
40
15
5

Oct ⟶ Sept

Time
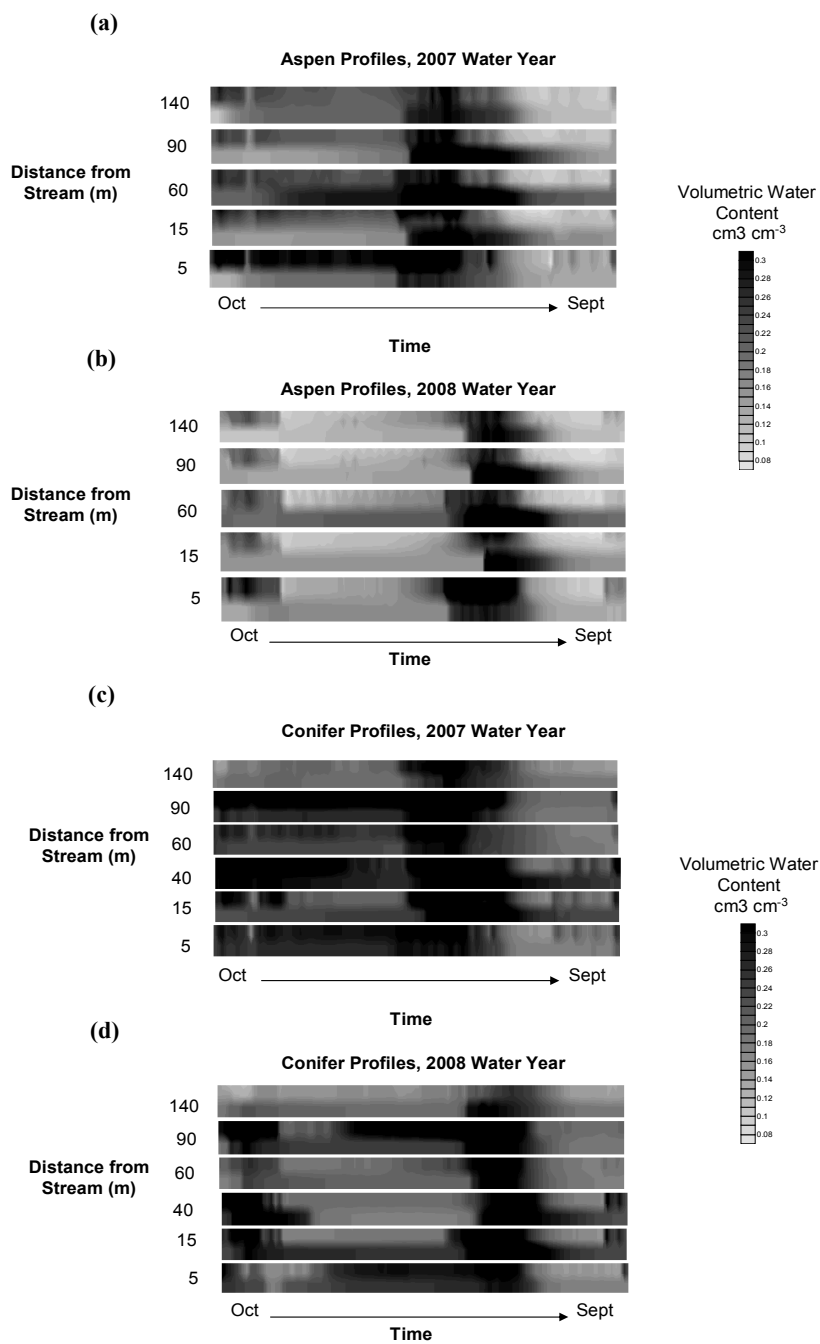
Figure 3.3. Soil volumetric water contents (cm^3 cm^{-3}) from profiles on each of the hillslopes for both water years 2007 and 2008.

(a)

(b)

(c)
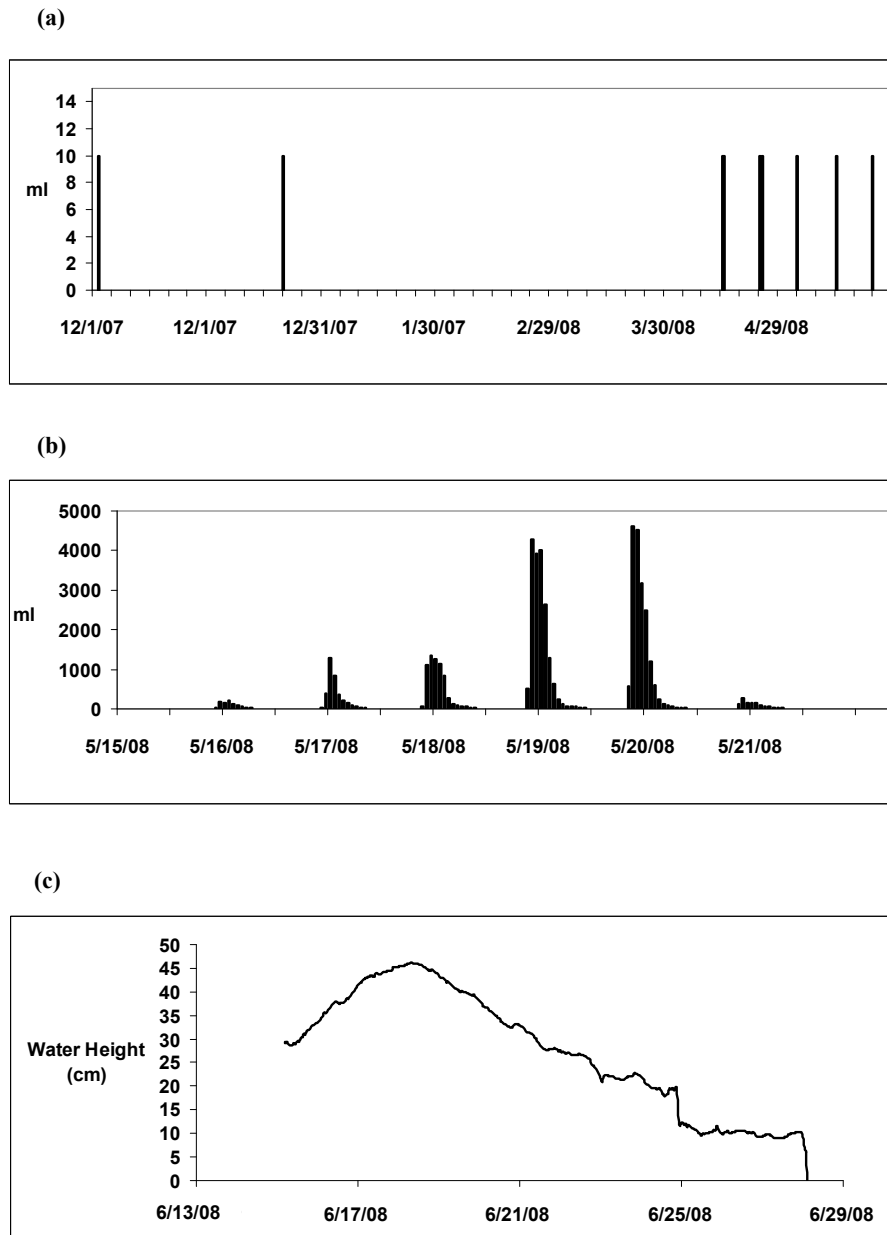
Figure 3.4. Trenchflow from (a) the aspen trench at 20 cm depth, (b) the aspen trench at 50 cm depth and (c) the level logger in the bottom of the conifer trench. Note that time scales differ for each plot.

but sustained slightly raised flows (~12 L s^{-1}) until sensors at all gauged points dropped

off to baseflows in mid June (~12th).

A hydrograph from the conifer dominated contributing area was isolated by subtracting flow from the middle section from flows at the conifer gauge (Figure 2.12b). The total water yield for the 2008 water year from the conifer contributing area was 1.46 m m^{-2}, with approximately 53% coming with the spring snow melt pulse (April 6th to July 4th). The total estimated water yield for the 2008 water year from the conifer contributing area was 0.73 m m^{-2}, with approximately 60% coming with the spring snow melt pulse. The 2008 annual water yield from the aspen dominated contributing area was 0.64 m m^{-2}. With a smaller and more upland contributing area, baseflow below the aspen hillslope was a mere trickle and 87% of all water flowing there comes during the spring pulse (February 13th to June 17th). This demonstrates the importance of the spring snowmelt for water yield in this snow driven system.

Snow had disappeared from the aspen hillslope by June 14th and from the conifer hillslope by June 20th. Therefore the diel fluctuations in streamflow after these dates (Figure 3.5) can be attributed to evapotranspiration, not snow melt. On June 14th, the stream gauged below the aspen hillslope showed a weak diurnal signal on June 14th, becoming slightly more pronounced from June 20th to 25th, during the beginning of leaf out (Figure 3.5). But after June 25th the streamflow was too low for consistent detection by the sensor, which requires at least 5 cm of water height. Diurnal fluctuations in streamflow at the middle gauging point were faint during the steep descend of the falling limb of the hydrograph but became apparent by June 28th and continued until the sensor was removed on August 28th. At the gauging point below the conifer hillslope diel fluxes were clearly seen throughout the spring melt and summer until the sensor malfunctioned in late June.

Plots of soil moisture measured at the deepest probes, versus streamflow show a threshold behavior at both gauging points (Figure 3.7). Flow gains from below the conifer contributing area are isolated by subtracting the flow at the middle gauging point from flow below the conifer hillslope. The soil moisture thresholds, past which streamflow increased, were 0.28 cm^3 cm^{-3} in the aspen and 0.32 cm^3 cm^{-3} in the conifer.
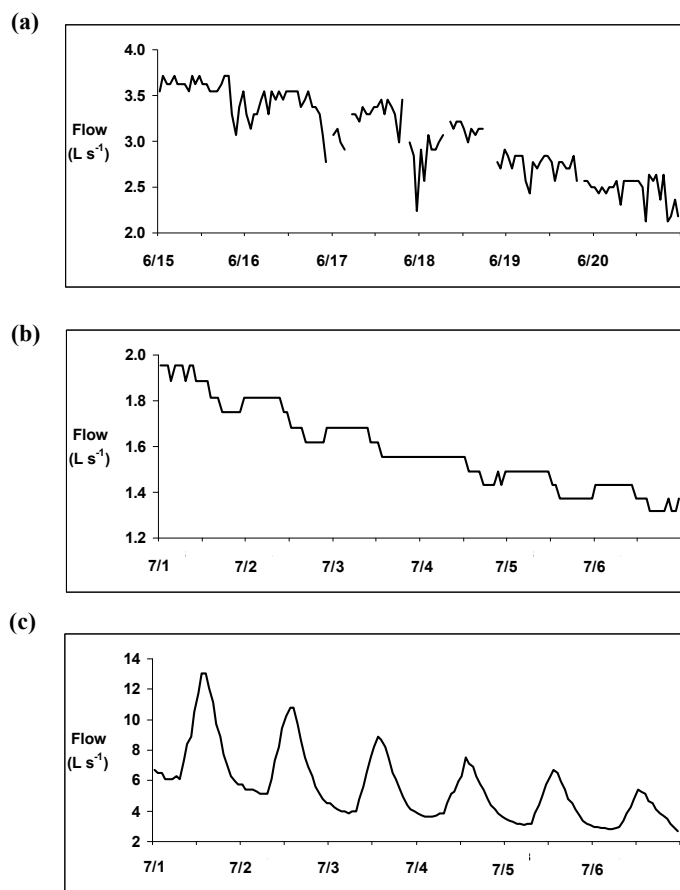
Figure 3.5. Examples of diel fluctuation in streamflow after snow disappeared from the catchment in the 2008 water year. (a) Streamflow gauged below the aspen hillslope, (b) streamflow gauged between the aspen and conifer contributing areas and (c) streamflow gauged below the conifer hillslope.

(a)

(b)

(c)
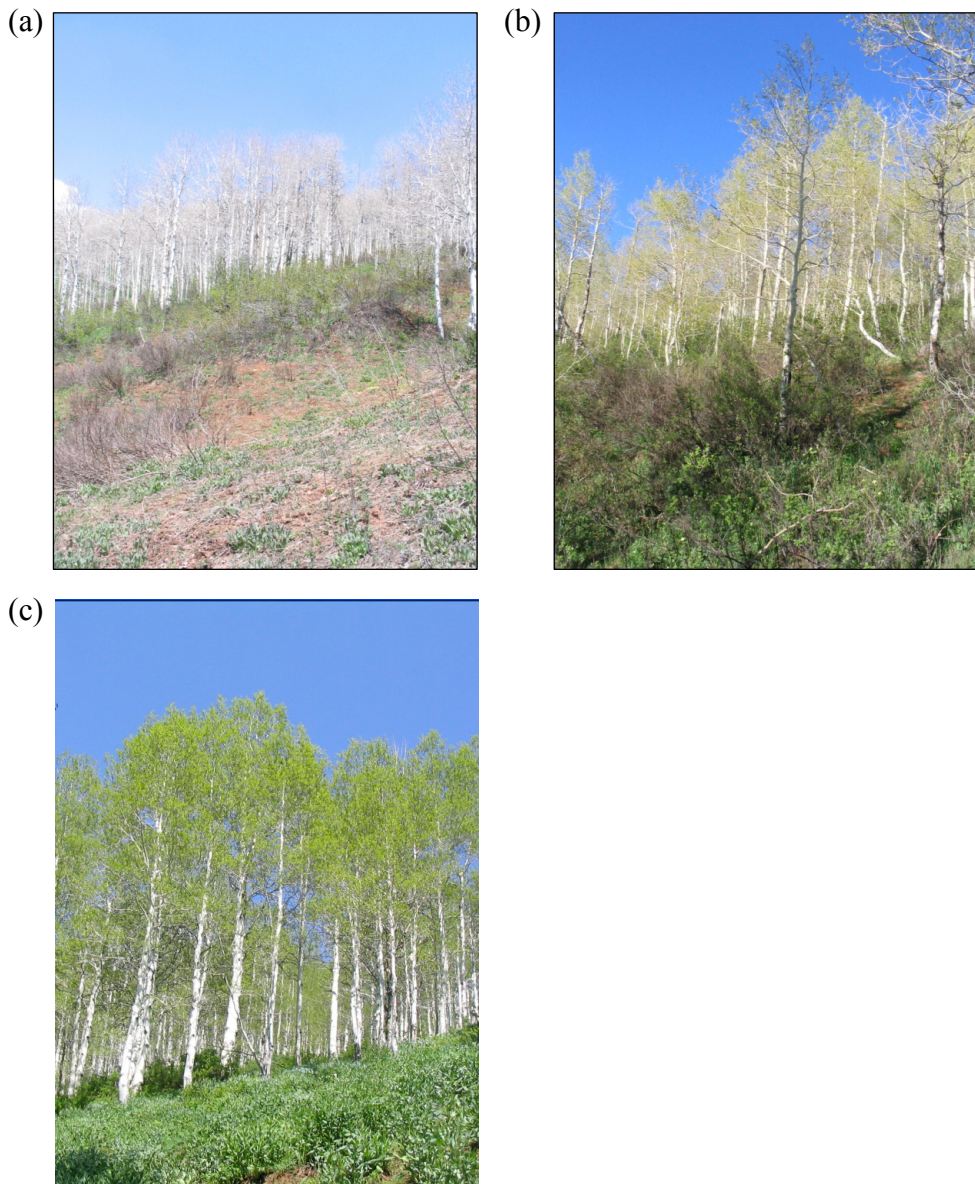
Figure 3.6. Pictures of aspen leaf out (a) no leaves on June 15th, (b) the beginning of leaf out, June 20th, and (c) full leaf out on June 24th.

(a)

(b)
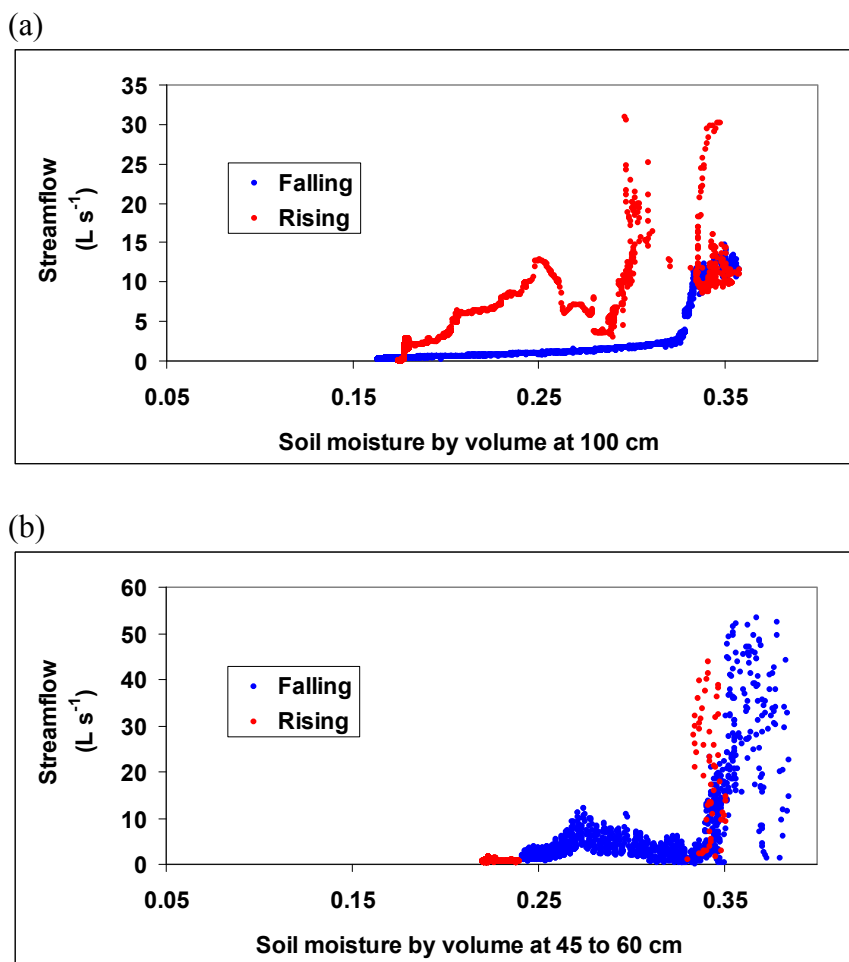
Figure 3.7. Soil moisture at the deepest sensors versus streamflow for (a) the aspen contributing area and (b) the conifer contributing area. The aspen hillslope shows hysteresis.

Overland flow

Upon first arrival to the catchment in the spring, saturation excess overland flow was observed (but not quantified) in the lower portions of the catchment, on May 20th. Thin films of surface water were observed at the aspen site on May 19th but had grown patchy by May 28th. Overland flow and return flow were observed at the conifer site on May 20th, 28th and 31st but had subsided by June 7th. Since snow still covered the soils

surface, observations of overland flow on middle and upper portions of the hillslopes were limited to points that were dug to for soil moisture measurements. Near the stream, sheet flow was seen on areas of exposed soil near the stream, routing water from the base of the snowpack to the stream. Overland flow was observed after May 20th, the day of highest surface water input during for the 2008 water year. However, since no one was there before May 20th, it is not known exactly when overland flow began.

DISCUSSION

To understand woody vegetation effects on hillslope scale hydrologic processes one needs to know if subsurface flow through the root zone is a dominant pathway to the stream (Huxman et al., 2005). One must also understand what other flowpaths to the stream are possible, or significant. In this study surface water inputs, soil moisture, trenchflow and streamflow were assessed to infer when the hillslopes of aspen and conifer were major contributors to streamflow. The goal of the study was to detect differences in subsurface flow between aspen and conifer hillslopes.

Soil moisture

No downslope accumulation of moisture was seen on either hillslope; soil moisture was not greater, nor did it last longer in the profiles closest to the stream (Figure 2.3). This was no surprise because subsurface lateral flow is highly heterogeneous and may not be detected at a series of singe point measurements. Using transect data to characterize an entire hillslope is difficult because flowpaths are spatially variable (Tromp-van Meerveld and McDonnell, 2006b). There was a brief window of widespread

high soil moisture throughout all profiles in both the 2007 and 2008 water years. This

would be the most likely time for subsurface lateral flow to be contributing to the stream.

By plotting soil moisture with streamflow, a soil moisture threshold beyond which

streamflow greatly increased was estimated (Figure 2.7). Comparing the number of days

in 2008 that soil moisture in each hillslope was above its threshold, demonstrates that the

aspen hillslope was above its threshold for longer than the conifer hillslope (57 versus 20

days). This result appears to support hypothesis (1), that hydraulic connection within the

aspen hillslope will have longer duration in the aspen than the conifers hillslope because

of greater surface water inputs.

Trenchflow

Trenchflow in the aspen stand showed the 20 cm deep collection system receiving

very little, sporadic flow as early as December 31st, before any large amounts of surface

water input (Figure 3.4). Condensation was also observed on the roof of the trench and

was considered the only cause of registered inputs at the upper collection system of the

aspen trench. The 50 cm deep collection system in the aspen trench flowed from May

15th to 21st (six days). Maximum flow from the seven-meter-long aspen trench was 0.38

L s^{-1} on May 20th when streamflow below the hillslope was 15 L s^{-1}. The time of

trenchflow in the aspen hillslope coincided with the highest rate of snow melt and clear

diel pulses of trenchflow are attributed to daily fluctuations in melt rate. By piecing

together tipping bucket data and water height data from the conifer trench I estimate that

the conifer trench flowed for approximately two months, April 25th to June 28th. Flow

from the conifer trench, measured by hand and extrapolated to represent flow from all

seven meters of the trench, was 4.79 L s^{-1} on May 31st. Given that streamflow below

the conifer hillslope was 60 L s^{-1} on May 31st, I assume that hillslope contributions made

up a significant portion of streamflow at this time. Trenchflow indicated the conifer

hillslope connection was sustained much longer than the aspen. Not only did flow in the

conifer trench exceed the aspen trench in duration but in quantity as well. Longer and

greater subsurface flow from the conifer hillslope would lead me to reject both

hypothesis (1) and (2), that hydraulic connection within the aspen hillslope will have

longer duration in the aspen than the conifers hillslope because of greater surface water

inputs.

Preferential flow through macropores is often important in generating lateral flow

(Lin et al., 2006). Macropore flow was initiated before matrix flow on a trench face in

the humid Panola catchment (Tromp-van Meerveld and McDonnell, 2006b). Preferential

flow on our conifer hillslope was observed between the shards of fracture bedrock

observed in the conifer profiles. Uchida et al. (2005) cite the most prevalent preferential

flow as that at the soil bedrock interface or at impeding layers (Uchida et al., 2005) which

appears to be the case at this study site as well.

Hillslope-stream connectivity

Using the soil moisture content at which streamflow becomes reactive (Figure

3.7), we detect longer connectivity in the soil profile of the aspen hillslope. This method

of defining hillslope-stream connection was used by McNamara et al. (2005) but may be

skewed at this study site by snow melt directly entering the stream via overland flow,

causing a rise in the hydrograph unrelated to subsurface lateral flow. Saturation excess

overland flow was observed (but not quantified) in the lower portions of the catchment, on May 20th, through 31st, the period of highest surface water input during for the 2008 water year, but had subsided completely by June 7th.

Using information from the trenches, longer connectivity was detected at the soil bedrock interface of the conifer hillslope (Figure 3.4). This method of defining hillslope-stream connectivity was used by Tromp-van Meerveld and McDonnell (2006b) and by Newman et al. (1998). While the two methods of identifying hillslope-stream connectivity appear to contradict each other, one must note that each are measuring hillslope processes at different depths and scales. The soil moisture probes supplied information about moisture conditions at the point scale, within the soil profile where probe installation was possible. The trenches collected water from a greater cross-section of hillslope that included macropores and the soil bedrock interface in the conifer trench.

Uncertainties are associated with both methods. In the soil moisture-streamflow relationship approach, soil moisture is used as a surrogate for soil water potential, which dictates water flow in soils. Soil texture was relatively homogeneous at the study site, and thus the use of soil moisture was reasonable. Aspen soils may have held more water but it could have been held with greater force. Aspen soils did not have greater clay content or bulk density but the high rock content in the conifer soils most likely decreased its relative ability to retain water. Uncertainty involved in trenchflow measurements may include the formation of a saturated wedge uphill from the trench face (Atkinson, 1978) and saturated flow around the trench (Tromp-van Meerveld et al., 2007). A saturated wedge is likely to have less of an effect on the conifer hillslope,

where rocky soils have higher hydraulic conductivities and offer less resistance at the soil-air interface of the trench face.

While the existence of a soil moisture-streamflow threshold relationship (Figure 3.7) does not necessarily signify subsurface lateral flow, there is an obvious relationship between hillslope soil moisture and streamflow. Figure 3.2 also shows both soil moisture and streamflow begin their spring pulses in an abrupt, threshold-like manner. By combining the information from these two figures (3.2 and 3.7), I infer that these thresholds of soil water contents lie between 18 and 28% in aspen and 23 and 32% in conifer. These thresholds most likely are due to soil pores reaching a moisture-content above field capacity, at which point hydraulic conductivity increased and soil water became drastically more mobile within the hillslope and more easily contributed to streamflow.

Soil depth

Soil depth or depth to bedrock was found to be a controlling factor for subsurface lateral flow in several study hillslopes, including the humid Panola catchment (Tromp-van Meerveld and McDonnell, 2006b), a humid catchment in Pennsylvania (Lin et al., 2006), and an agricultural catchment in Western France (Freer et al., 1997; Chaplot and Walter, 2003; Tromp-van Meerveld and McDonnell, 2006b). In this study, the soil profile under conifers was shallow and fractured bedrock was reached within the top meter. The aspen profile was very deep and though the deepest part of the aspen trench was nearly 1.5 m deep, true bedrock was never reached. The approximate storage capacity of each profile was estimated using physical properties of the soil profiles. Soil

depths were observed at eight soil pits dug for sensor installation as well as at the

trenches. For simplicity's sake I have rounded soil depths off to conservative estimates

of 1 m in the aspen and 0.5 m in the conifer. Porosity was considered the maximum soil

moisture contents observed (100% saturation). Values were averaged from all sensors on

each hillslope and found to be 35% is aspen and 38% in conifer (Figure 3.2). Rock

contents were averaged from samples on each hillslope (Table 2.3), at 1% in aspen and

46% in conifer. The remaining profile volume was assumed to be soil, including mineral

and organic particles smaller that 2 mm. Comparisons of each profile water storage

capacity (Figure 3.8) show that maximum storage capacities would be overcome after 35

cm of water inputs in the aspen and after 19 cm water inputs in the conifer. Therefore

excess water would have been -4 and 16 cm from the aspen profile, in 2007 and 2008

respectively and 7 and 20 cm from the conifer profile, in 2007 and 2008 water years

respectively.

Essentially the deeper aspen profile, with more void space, was able to absorb the

majority of water inputs, leaving little water for possible subsurface flow (or other

pathways to the stream or deep groundwater storage). Whereas the shallow, rocky

conifer profile, with less storage capacity, produced more water per area for subsurface

lateral flow, overland flow or deep groundwater percolation. This corresponds with

calculated water yield from each of the contributing areas, the conifer yielding a greater

amount (0.73m m^{-2} versus 0.64 m m^{-2}).

Plots of soil moisture with streamflow appears to support hypothesis (1), that

hydraulic connection within the aspen hillslope will have longer duration in the aspen

than the conifers hillslope because of greater surface water inputs. However, a longer
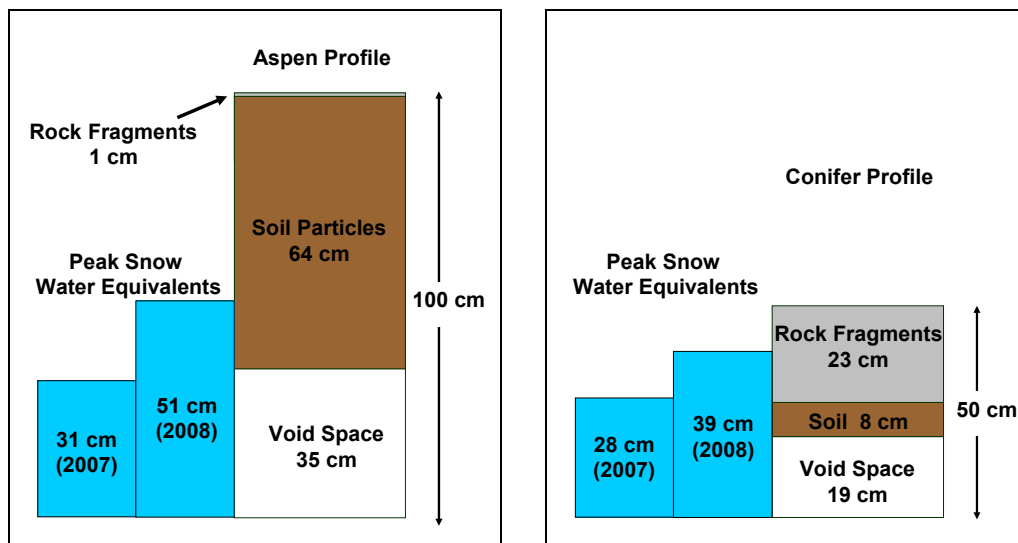
Figure 3.8. Scaled representations of accumulated snow water equivalent and estimated storage capacity for each profile.

observed period of trenchflow in the conifer hillslope, together with greater storage capability in the aspen profile, lead me to reject hypothesis (1). Though the conifer canopy lessened its surrounding snow-pack, the shallow soil profile yielded more water via subsurface lateral flow in the top meter of soil.

Canopy effects on snow duration, soil moisture, and streamflow

Differences in aspen and conifer canopy characteristics lead to differences in quantity and timing of input water inputs to the catchment   Figure 2.5 shows that the aspen hillslope accumulated 11 and 23% more SWE than the conifer hillslope in the 2007 and 2008 water years. Observations shown in Figure 2.6, demonstrate that snowdrifts lasted longer under the conifer stand. Other studies in the snow-driven Intermountain

West have noted that tree canopies alter the delivery of surface water inputs
(Hiemstra et al., 2002; Marks et al., 2002; LaMalfa and Ryel, 2008). In this study
catchment, the conifer canopy appears to attenuate spring snow melt water inputs,
lessening the snowpack and storing drifts for longer. However, soil moisture peaks in the
hillslopes do not differ in their duration and streamflow contributions from the conifer
dominated area appear to peak more intensely than those from the aspen area. The
attenuation of water inputs does not translate into attenuation of soil moisture or
streamflow in the conifer hillslope.

In this study, water yield appears to be affected by a combination of soil depth
and canopy type. The conifer canopy decreased the accumulated SWE available for run
off yet the conifer soil profile held less water, therefore contributing more to the stream.
The aspen hillslope received more SWE but held melt-water in the soil more efficiently,
yielding less stream-water. Changes in vegetation will bring about changes in the soil
profile. However, soil characteristics are much less dynamic than vegetation and
significant changes in profile water storage capacity will lag changes in vegetation type
(Shepperd et al., 2006). Surface water yield might then be decreased most in an invasion
scenario, where a conifer canopy would occupy an aspen soil profile. This study
therefore suggests that restoration efforts aimed at increases water yield through
vegetation management should focus on conifer invasion into deep (aspen) soils.

Deep groundwater sources

The aspen contributing area includes a head-cut containing a spring from which
the stream originates. A previous study in the watershed found that stream water

chemistry suggests that groundwater is the source of baseflow in this system

(Shakespeare, 2006). While water surfacing at the head-cut is most likely a mix of

subsurface flow and deeper groundwater, this is evidence that other flowpaths besides

subsurface lateral are playing a role in runoff generation from the catchment. It is known

that deep groundwater sources supply the stream's baseflow but the magnitude of their

contribution to peak flow in the spring snow melt is unknown. Mechanisms of surface

water generation must be known for the effects of woody plants on streamflow to be

understood (Huxman et al., 2005; Weiler et al., 2005). The more dominant subsurface

flow is as a pathway, the more streamflow will be affected by a shift in woody vegetation

types (Huxman et al., 2005). However, even if deep groundwater is the primary source

of streamflow, water inputs must still percolate past the root zone in order to recharge

aquifers. Vegetation affects streamflow in the study catchment, as seen through the diel

fluctuations in stream discharge seen after snow was gone from the catchment (Figure

3.5). Using the same concept, Bond et al. (2002) linked streamflow and vegetation by

comparing diel patterns of transpiration and streamflow after all snow had melted from

the watershed. Further investigations are needed in the study catchment to assess how

much streamflow in the spring peak is supplied by deep groundwater sources and to what

extent vegetation affects the recharge of these sources.

CONCLUSIONS

Using the soil-moisture threshold past which streamflow greatly increased, a

comparison between the vegetation types reveals that the aspen hillslope was connected

to the stream for longer than the conifer hillslope. However, upon comparing trenchflow

between the sites, it becomes clear that the conifer hillslope was connected for longer and experienced more flow. Hypothesis (1), that hydraulic connection within the hillslopes during the spring snow melt will have longer duration in the aspen than the conifers hillslope because of greater surface water inputs, is therefore rejected. Thresholds past which trenchflow began were exceeded for longer in the conifer hillslope. This lead to also reject hypothesis (2), that a soil water content threshold for subsurface flow will be exceeded for longer time periods in the aspen stand due to greater surface water inputs.

Soil depth was found to be a controlling factor on subsurface lateral flow since calculations indicate that conifer soils have much less capacity for storing water. Differences in snowpack duration were observed between the canopy types, but this did not translated into short but intense pulses of soil moisture or streamflow in the aspen stand where more SWE accumulated and melt time was less.

Further investigations into other flowpaths in the catchment and their relative contributions are needed. Deep groundwater sources and overland flow are contributing to streamflow but understanding their quantity and timing would help estimate vegetation effects on streamflow. Soil depth is a likely control on subsurface lateral flow in this catchment and mapping bedrock topography would surely elucidate how the two hillslopes function in comparison to one another.

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CHAPTER 4

CONCLUSIONS

This study examined hillslope-stream connectivity in two end point communities of aspen and conifer. Surface soil moisture is important for many components of ecohydrology and understanding how it differs between these vegetation types leads to how these different forest types will cycle water. Subsurface lateral flow is greatly affected by vegetation therefore comparing this flowpath between vegetation types will help our understanding of vegetation effects on water yield.

Canopy type at this site clearly affects water inputs. But further research into sublimation of snow retained in the canopy is needed to clarify the extent to which this changes water yield. Overland flow and deep groundwater have both been observed contributing streamflow. Quantifying these inputs would greatly improve our knowledge of how streamflow generation mechanisms in this catchment are influenced by vegetation. Soil depth or depth to bedrock was also found to control subsurface lateral flow and warrants further investigation.

The complexity of hillslope hydrology is due to a multitude of processes interacting at different scales. Because of diel fluctuations in streamflow after the disappearance of snow in our catchment, we know that transpiration is affecting streamflow below both aspen and conifer hillslopes. Shallow soil moisture shows connectivity lasting longer in the aspen stand, but examination of the deeper trenchflow shows greater volumes and longer duration of subsurface lateral flow in the conifer site.

Similar to other semi-arid systems, subsurface lateral flow connected the stream and

hillslope of this catchment only during the wettest periods of the spring snow melt. For

the remainder of the year the hillslope and stream were disconnected.

APPENDIX

(a)
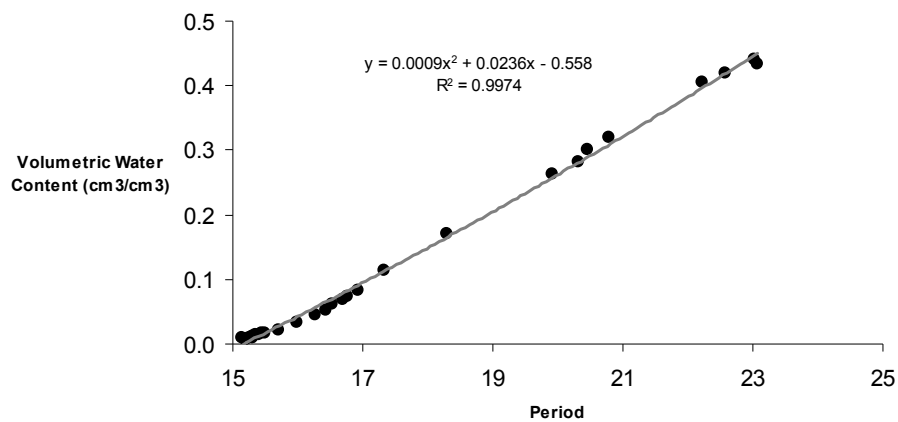
(b)
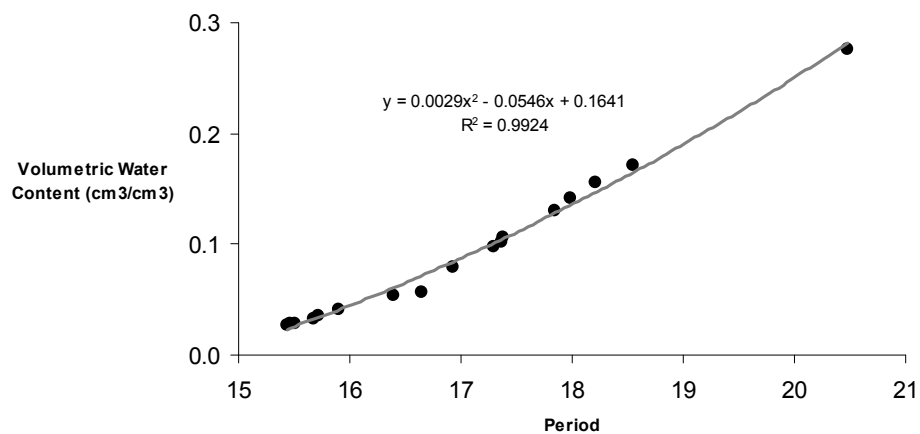
Figure A1. Calibration Curves for soil water content probes (CS616) (a) 12 cm probe in aspen soils, (b) 12 cm probe in conifer soils, (c) 30 cm probe in aspen soils and (d) 30 cm probe in conifer soils.

(c)

$y = 0.0004x^2 + 0.025x - 0.4732$
$R^2 = 0.9608$

Volumetric
Water Content
(cm3/cm3)

period

(d)

$y = 0.0007x^2 - 0.0152x + 0.1495$
$R^2 = 0.9916$

Volumetric
Water Content
(cm3/cm3)
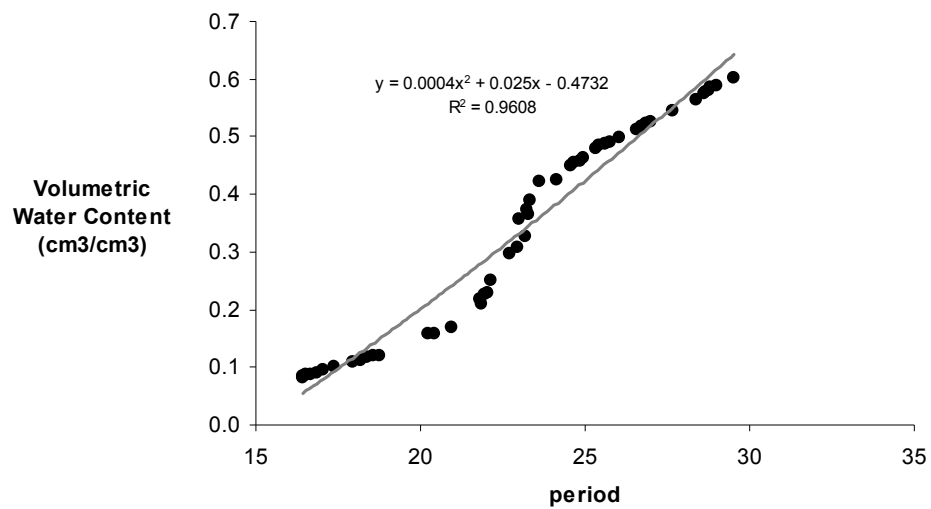
period
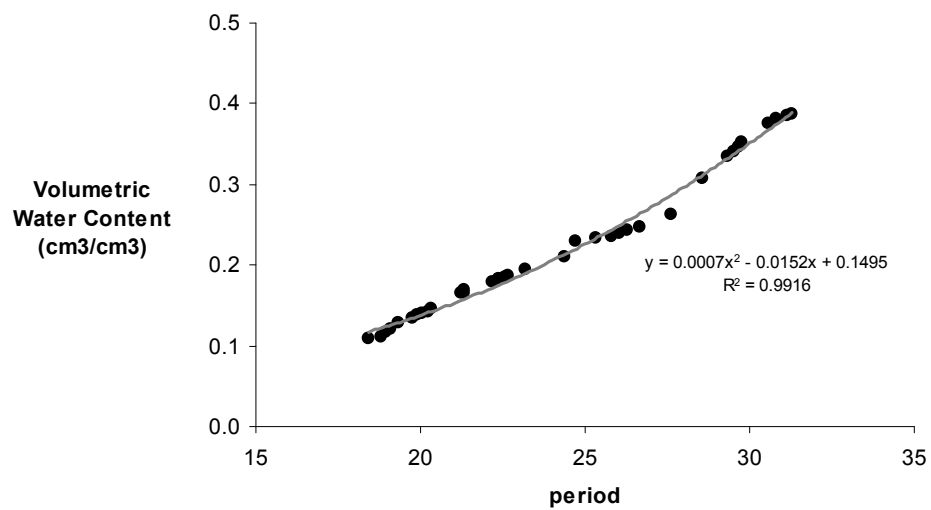
Table A1. Soil texture for each point measured.

Point on Apen Hillslope	Depth (cm)	Sand %	Silt %	Clay %	Texture Class
17	0-5	31.89	46.31	21.80	Loam
24	0-5	46.22	33.47	20.32	Loam
35	0-5	42.02	41.47	16.51	Loam
47	0-5	43.33	46.13	10.55	Loam
5	0-5	40.39	38.46	21.14	Loam
54	0-5	52.70	37.95	9.35	Sandy Loam
63	0-5	52.13	30.63	17.23	Sandy Loam
76	0-5	53.54	30.24	16.21	Sandy Loam
81	0-5	49.47	33.40	17.13	Loam
96	0-5	53.28	29.18	17.55	Sandy Loam
A1	0-5	36.90	42.32	20.78	Loam
A4	0-5	41.11	41.42	17.47	Loam
A3	0-5	41.44	38.40	20.16	Loam
A7	0-5	43.23	38.27	18.50	Loam
A7	5-20	42.43	37.62	19.96	Loam
A4	5-20	38.40	40.55	21.06	Loam
A3	5-20	40.92	37.95	21.13	Loam
A1	5-20	33.84	42.00	24.16	Loam
A1	20-88	37.53	38.38	24.09	Loam
A3	20-96	41.85	36.65	21.50	Loam
A4	20-126	39.57	39.02	21.41	Loam
A7	20-80	42.34	38.02	19.64	Loam
Average Aspen		42.93	38.08	18.98	Loam
Standard Deviation		6.33	4.79	3.81	

Table A1. (Continued)

Point on Conifer Hillslope	Depth (cm)	Sand %	Silt %	Clay %	Texture Class
19	5	38.49	43.86	17.65	Loam
25	5	32.29	48.19	19.52	Loam
4	5	50.23	35.33	14.44	Loam
47	5	41.21	48.93	9.86	Loam
50	5	53.21	30.57	16.21	Sandy Loam
63	5	42.33	39.16	18.51	Loam
78	5	41.00	41.56	17.44	Loam
87	5	33.56	50.00	16.44	Silt Loam
95	5	35.51	44.93	19.56	Loam
B3	5	43.82	40.90	15.29	Loam
B3	5	44.48	38.05	17.47	Loam
B5	5	27.96	54.85	17.20	Silt Loam
B1	5	54.92	24.96	20.12	Sandy Clay Loam
B1	5-20	54.56	28.00	17.44	Sandy Loam
B3	5-20	37.23	43.33	19.44	Loam
B4	5-20	25.89	52.67	21.44	Silt Loam
B5	5-20	25.93	53.55	20.52	Silt Loam
B7	5-20	33.29	46.39	20.32	Loam
B1	20-75	55.23	26.79	17.98	Sandy Loam
B3	20-93	56.02	25.90	18.09	Sandy Loam
B5	20-95	53.87	32.49	13.64	Sandy Loam
B7	20-85	53.96	30.34	15.71	Sandy Loam
Average Conifer		41.46	41.64	16.90	Loam
Standard Deviation		8.09	8.25	2.69	

Table A2. Bulk densities and location of soil measurement points.

Aspen Point	UTM N	UTM E	Depth (cm)	Volume (cm3)	Dry weight (g)	Rock Volume (%)	Bulk Density (g/cc)	Porostity (%)
5	461398	4579137	5	280	441.1		1.58	40.55
6	461401	4579131	5	400	446		1.12	57.92
10	461411	4579105	5	350	442.8		1.27	52.26
16	461413	4579134	5	560	607.6		1.09	59.06
17	461412	4579135	5	410	455.3		1.11	58.09
24	461431	4579146	5	400	439		1.10	58.58
35	461448	4579140	5	260	439.5		1.69	36.21
36	461454	4579126	5	230	255.5		1.11	58.08
47	461407	4579125	5	200	243.1		1.22	54.13
52	461481	4579166	5	310	436.6		1.41	46.85
54	461483	4579152	5	320	256		0.80	69.81
55	461493	4579153	5	210	229.7		1.09	58.72
63	461499	4579162	5	420	677.2		1.61	39.16
66	461500	4579139	5	480	494.1		1.03	61.16
70	461502	4579100	5	300	310.9		1.04	60.89
73	461513	4579167	5	320	433.5		1.35	48.88
76	461518	4579140	5	530	436		0.82	68.96
79	461528	4579186	5	450	499.1		1.11	58.15
81	461539	4579169	5	250	232.3		0.93	64.94
85	461523	4579229	5	190	239.5		1.26	52.43
90	461550	4579188	5	380	438.2		1.15	56.48
91	461555	4579179	5	540	493.4		0.91	65.52
93	461528	4579255	5	450	405.5		0.90	66.00
96	461543	4579230	5	370	295.4		0.80	69.87
nest 6	461436	4579152	5	150	146.8	1.33	0.98	63.07
nest 6	461436	4579152	20	275	306.7	0.73	1.12	57.91
nest 6	461436	4579152	65	195	223.5	0.51	1.15	56.75
nest 5	461400	4579168	5	235	233.1	0.85	0.99	62.57
nest 5	461400	4579168	20	265	308.5	0.38	1.16	56.07
nest 5	461400	4579168	90	255	299.5	0.39	1.17	55.68
nest 7	461466	4579147	5	330	330.5	0.61	1.00	62.21
nest 7	461466	4579147	20	220	245.7	0.91	1.12	57.86
Trench	461395	4579190	5	245	273.3	2.45	1.12	57.91
Trench	461395	4579190	20	225	217.1	0.44	0.96	63.59
Trench	461395	4579190	60	145	202.2	1.38	1.39	47.38
nest 8	461501	4579132	20	275	298.7	0.55	1.09	59.01
Average Aspen						0.88	1.13	57.30
Standard Deviation						0.60	0.21	7.95

Table A2. (Continued)

Conifer Point	UTM N	UTM E	Depth (cm)	Volume (cm3)	Dry weight (g)	Rock Volume (%)	Bulk Density (g/cc)	Porostity (%)
1	461157	4579299	5	350	285.2		0.81	69.25
3	461170	4579289	5	270	363.2		1.35	49.24
4	461181	4579289	5	350	400.6		1.14	56.81
8	461217	4579270	5	280	311.5		1.11	58.02
9	461236	4579260	5	210	213.2		1.02	61.69
14	461173	4579275	5	510	660.5		1.30	51.13
15	461191	4579251	5	920	1636		1.78	32.90
16	461203	4579273	5	310	446.6		1.44	45.64
18	461202	4579302	5	570	119.8		0.21	92.07
19	461213	4579262	5	330	370		1.12	57.69
21	461159	4579275	5	390	610		1.56	40.98
22	461159	4579268	5	490	415.6		0.85	67.99
23	461172	4579254	5	400	393		0.98	62.92
24	461175	4579253	5	340	529.7		1.56	41.21
31	461144	4579259	5	400	549		1.37	48.21
32	461160	4579239	5	370	477.2		1.29	51.33
33	461151	4579232	5	440	736.9		1.67	36.80
34	461175	4579226	5	390	641.3		1.64	37.95
35	461166	4579236	5	510	494.1		0.97	63.44
43	461154	4579222	5	400	315.2		0.79	70.26
44	461164	4579219	5	520	417		0.80	69.74
45	461178	4579211	5	460	562.8		1.22	53.83
46	461178	4579210	5	260	298.7		1.15	56.65
47	461184	4579207	5	210	379.4		1.81	31.82
48	461193	4579204	5	210	305.5		1.45	45.10
49	461209	4579192	5	250	399		1.60	39.77
54	461176	4579190	5	360	441		1.23	53.77
55	461180	4579186	5	430	138.4		0.32	87.85
56	461188	4579182	5	320	458.2		1.43	45.97
57	461199	4579177	5	500	531.7		1.06	59.87
59	461217	4579168	5	330	323.5		0.98	63.01
60	461217	4579158	5	450	630.1		1.40	47.16
63	461151	4579176	5	360	325		0.90	65.93
64	461157	4579162	5	290	350.9		1.21	54.34
65	461171	4579172	5	300	511.9		1.71	35.61
67	461185	4579154	5	350	552.8		1.58	40.40
68	461200	4579148	5	570	645.8		1.13	57.25
69	461204	4579147	5	400	588.2		1.47	44.51
76	461177	4579156	5	220	208.4		0.95	64.25
78	461198	4579141	5	330	343.1		1.04	60.77
82	461147	4579150	5	280	57.2		0.20	92.29
83	461153	4579146	5	400	661.5		1.65	37.59
87	461184	4579119	5	350	594.8		1.70	35.87
95	461174	4579125	5	270	337.7		1.25	52.80

Table A2. (Continued)

96	461182	4579110	5	210	318.5		1.52	42.77
nest 5	461193	4579267	5	235	357.9	46.81	1.52	42.53
nest 5	461193	4579267	20	600	1139.5	61.67	1.90	28.33
nest 5	461193	4579267	45	325	604.8	73.85	1.86	29.78
nest 6	461188	4579276	5	240	253.4	25.00	1.06	60.16
nest 6	461188	4579276	20	175	181.3	20.00	1.04	60.91
nest 6	461188	4579276	45	315	623.7	60.32	1.98	25.28
Trench	461157	4579299	20	270	273.1	25.93	1.01	61.83
Trench	461157	4579299	40	380	721	69.74	1.90	28.40
Trench	461157	4579299	5	275	308.3	32.73	1.12	57.69
Average Aspen						46.23	1.28	51.88
Standard Deviation						20.87	0.42	15.78