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An evaluation of the hydrologic relevance of lateral flow in snow at hillslope and catchment scales

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Abstract:

Lateral downslope flow in snow during snowmelt and rain-on-snow (ROS) events is a well-known phenomenon, yet its relevance to water redistribution at hillslope and catchment scales is not well understood. We used dye tracers, geophysical methods, and hydrometric measurements to describe the snow properties that promote lateral flow, assess the relative velocities of lateral flow in snow and soil, and estimate volumes of downslope flow. Results demonstrate that rain and melt water can travel tens of metres downslope along layers within the snowpack or at the snowpack base within tens of hours. Lateral flow within the snowpack becomes less likely as the snowpack becomes saturated and stratigraphic boundaries are destroyed. Flow along the base can be prevalent in all snowpack conditions. The net result of lateral flow in snow can be the deposition of water on the soil surface in advanced downslope positions relative to its point of origin, or direct discharge to a stream. Although both melt and ROS events can redistribute water to downslope positions, ROS events produced the most significant volumes of downslope flow. Direct stream contributions through the snowpack during one ROS event produced up to 12% of streamflow during the event. This can help explain rapid delivery of water to streams during ROS events, as well as anomalously high contributions of event water during snowmelt hydrographs. In catchments with a persistent snowpack, lateral redistribution of water within the snowpack should be considered a relevant moisture redistribution mechanism. Copyright © 2012 John Wiley & Sons, Ltd.

KEY WORDS lateral flow; snow; rain on snow; snowmelt; runoff generation

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INTRODUCTION

Catchments convert spatially distributed, vertical inputs of precipitation into lateral flow towards a common outlet.† The transformation from vertical flow to lateral, or slope-parallel flow is commonly assumed to occur in and on soils. Lateral flow, however, can also occur in sloping snowpacks (Wankiewicz, 1979; Singh *et al*., 1997; Kattelmann and Dozier, 1999; Peitzsch, 2009; Whitson, 2009). Despite an early recognition by hydrologists of the occurrence of lateral flow through snow (Horton, 1915; Horton, 1938), its hydrologic significance at hillslope and catchment scales has received little attention. Hydrologic models often assume that melt water or rain water exits the base of the snowpack directly beneath its point of origin (Kelleners *et al*., 2009; Seyfried *et al*., 2009). Lateral flow in snow is often viewed as an impedance to vertical percolation, rather than as an important mechanism in itself (Colbeck, 1975; Jordan, 1983) however this may be due to the majority of studies occuring at more easily monitored flat research sites. The occurrence of lateral flow in snow, however, may help explain some important problems in snow hydrology. For example, rain-on-snow (ROS) can deliver large volumes of water to streams much more rapidly than soil-based runoff generation mechanisms can explain (Marks *et al*., 1998; Stratton *et al*., 2009). Further, streamflow in response to water input tends to be dominated by pre-event water (water in the catchment prior to the storm), as opposed to event water (new rain or snowmelt), but less so when event water comes from snowmelt (Buttle, 1994; Sueker *et al*., 2000; Shanley *et al*., 2002). Finally, lateral flow can be an important mechanism for wet snow avalanches. Lateral flow causes local concentrations of liquid water, which decreases shear strength and rapidly increases strain rate, and is likely the cause of widespread avalanching during ROS events† (Conway and Raymond, 1993; McClung *et al*., 2006; Peitzsch, 2009).

A snowpack on a hillslope is an additional porous media layer that receives, stores, and transmits water according to physical principles similar to those that govern flow in soil. Flow within the porous matrix is driven primarily by gradients in fluid potential energy and moderated by conductance as described by Darcy's law (Colbeck, 1975). In the absence of impeding layers or strong lateral pressure differences, vertical energy gradients in unsaturated media will dominate over lateral gradients. If, however, flow-impeding layers cause saturated or near-saturated zones, vertical energy gradients within those zones become negligible, enabling lateral energy gradients to drive flow.

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An impeding layer occurs where a contrast in hydraulic conductivity renders flow through the layer slower than the flow parallel to the layer. Snowpacks are typically composed of layers of snow from multiple storms with spatially and temporally variable properties (Gustafsson *et al*., 2004). Layer boundaries can create the hydraulic conductivity contrasts necessary for impeding vertical flow (Seligman, 1936; Gerdel, 1954; Wankiewicz, 1979; Woo *et al*., 1982; Marsh and Woo, 1985; Schneebeli, 1995; Singh *et al*., 1997; Kattelmann and Dozier, 1999; Waldner *et al*., 2004; Peitzsch, 2009; Whitson, 2009; Williams *et al*., 2010).

Solid ice layers are considered to be an important type of flow-impeding layer in a snowpack (Seligman, 1936; Conway and Benedict, 1994). Indeed, ice layers can have permeability five to ten times lower than the surrounding snowpack (Albert and Perron, 2000). However, lateral flow has also been observed in snow with no prominent ice layers (Wankiewicz, 1979; Kattelmann and Dozier, 1999; Waldner *et al*., 2004; Whitson, 2009). Kattelmann and Dozier (1999) noted that some ice layers have the appearance of 'swiss cheese' with 50–75% of total ice area composed of holes. Such ice layers tend to route water laterally only for short distances before concentrating water into vertical channels (Gerdel, 1948; Gerdel, 1954; Jordan, 1983; Marsh and Woo, 1985; Furbish, 1988). Other examples of potential impeding layers include transitions where density (Illangasekare *et al*., 1990), pore size (Waldner *et al*., 2004), and grain size (Pfeffer and Humphrey, 1996) change abruptly between layers. In particular, a fine to coarse grain transition can cause a capillary barrier wherein water may be held in the overlying fine material because of capillary forces (Oldenburg and Pruess, 1993). Although the capillary barrier effect is not well documented in snow, the conditions that cause capillary barriers can occur between stratigraphic layers and at the snow–soil interface (Waldner *et al*., 2004; Ohara *et al*., 2011).

Although the occurrence of lateral flow in snow is well known, few studies have investigated its potential velocities, flow rates, and impacts on hillslope and catchment-scale hydrologic processes. Notable exceptions are Higuchi and Tanaka (1982), English *et al*. (1986), and Ohara *et al*. (2011). Higuchi and Tanaka (1982) measured outflow rates from rivulets in the surface of a snowpack but did not place the results in the context of catchment hydrological processes. English *et al*. (1986) attributed increased downslope outflow from five sequential runoff plots to lateral water redistribution on ice layers but did not document the existence of ice layers. Ohara *et al*. (2011) measured overland flow during snowmelt in a study site that was characterized by highly conductive, unfrozen soils (i.e. conditions that would not typically promote classic Hortonian overland flow). They concluded that overland flow was the result of melt water movement held in the base of the snowpack by surface tension.

Here, we evaluate the importance of lateral flow in snow in the context of hillslope and catchment scale water redistribution in the semi-arid snow-dominated mountains of southwest Idaho, USA. In this region, improved understanding of snow hydrology is essential as climate warming imposes increased frequency of ROS events and mid-winter melting. Specifically, this study assesses the lengths scales, velocities, and quantities of lateral flow in snow to determine its potential impact on hillslope hydrologic pathways and streamflow generation. Accordingly, we conducted a series of field experiments with the following objectives:

1. Describe the conditions that promote lateral flow of melt water and rain using dye tracers.
2. Compare relative velocities of flow in snow and soil using tracers and geophysical methods.
3. Measure quantities of downslope water movement within snow using a paired lysimeter experiment.

The following two sections describe the methods and results for each objective, respectively. The Discussion section integrates the objective-specific results to evaluate the hydrologic significance of our results at hillslope and catchment scales.

METHODS

Study area

All experiments in 2007, 2008, and 2011 were conducted within the southern mountains of the Boise National Forest (Figure 1). Although intermittent snow occurs in Boise, Idaho (~850 m), the elevation of the persistent snowpack in this region tends to be around 1500 m. The next subsection describes melt water tracer experiments performed near Mores Creek Summit and ROS experiments conducted in the upper reaches of the Dry Creek Experimental Watershed (DCEW), located just outside of Boise, Idaho. Tracer velocity and lysimeter experiments (as discussed in the succeeding sections) were conducted in, and adjacent to, the Treeline Catchment ($0.02\ km^2$) within the DCEW.

The Treeline Catchment is oriented northwest–southeast and located at a mean elevation of 1620 m with 70 m of total relief. It has two small tributaries contributing to one main ephemeral channel that typically begins flowing in late autumn and ceases in late spring or early summer. The total stream network length is approximately 250 m and is gauged with a v-notch weir. The catchment contains standard meteorological instrumentation (air temperature, wind speed and direction, and relative humidity), a four component radiometer, several ultrasonic snow depth sensors, and numerous soil moisture measurements. See McNamara *et al*. (2005) and Williams *et al*. (2009) for descriptions of the instrumentation at the Treeline Catchment. Precipitation is measured in a shielded weighing bucket gauge and corrected for wind using the standard World Meteorological Organization gauge catch correction equations for rain and snow (Dingman, 1994). Precipitation phase is determined with a 0 °C dew point temperature threshold (Marks and Winstral, 2007). Two runoff plots

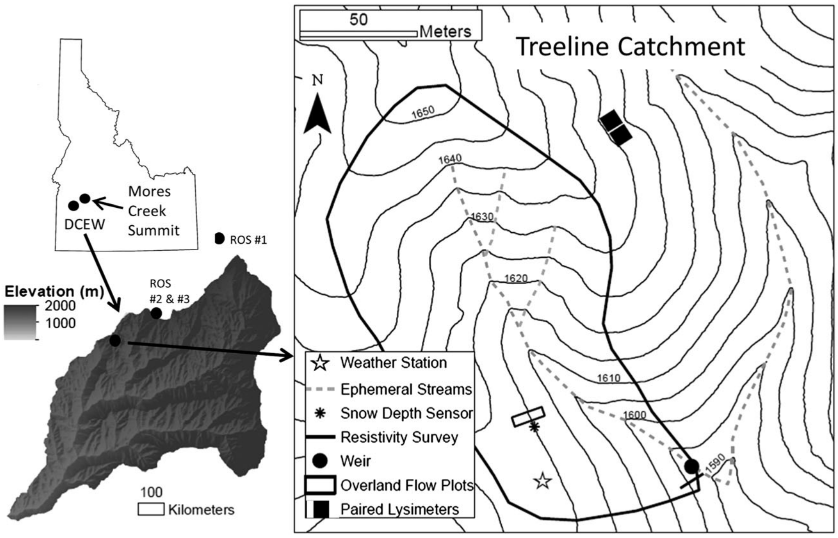

Figure 1. Map of study area illustrating locations of tracer tests, lysimeter studies, and instrumentation

collect overland flow from isolated 3 × 10-m areas routed to 1893-L tanks on the northeast-facing slope.

During the 2010–2011 snow season, the Treeline Catchment received 54 mm of precipitation, 37 mm of which fell as snow. A snowpack persisted on the northeast-facing slope of the Treeline Catchment from 18 November 2010 to 12 April 2011 with an average depth of 30–50 cm and a maximum depth of about 80 cm recorded during mid-December (Figure 2). In contrast to the northeast-facing slope, the increased solar radiation on the southwest-facing slope promoted the development of a transient snowpack that accumulated and melted several times throughout the winter months. Stream discharge from the Treeline Catchment ranged from 0 to a peak of 505 L min^{-1}. Peak flow was associated with a ROS event that yielded 53 mm of rain over a 27-h period on 15–17 January 2011. Near-surface, volumetric soil moisture content was in the 'wet, low-flux' state as described by McNamara *et al.* (2005) for the majority of the winter with an average value of ~0.15. In early March, the soil moisture state transitioned to the 'wet, high-flux' condition with peak volumetric moisture of ~0.22. At no point during the 2010–2011 snow season were the frozen soil conditions observed in the Treeline Catchment. Because of the shallow and transitional nature of the snowpack, limited snowpack stratigraphy was present. For the duration of this study, the snowpack was typically dominated by 0.25–1 mm rounds with the occasional weak melt–freeze crust and faceted layer.

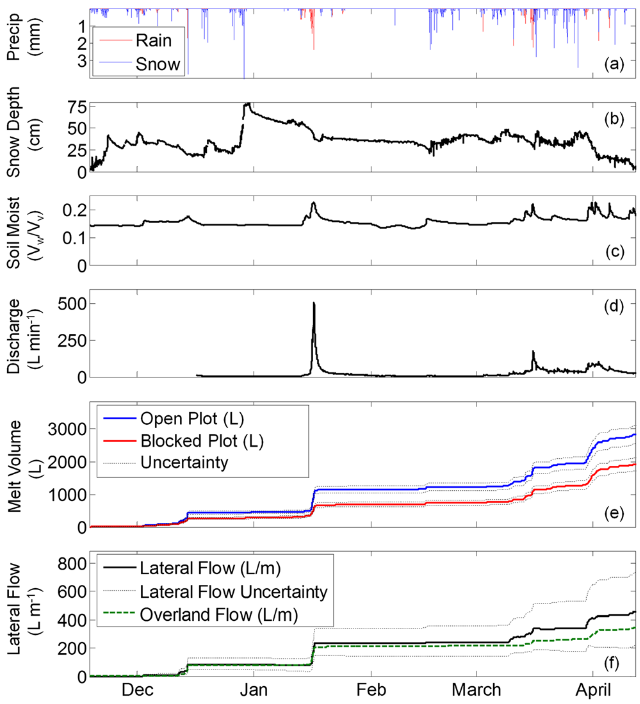

Figure 2. Summary of Treeline Catchment hydrology (a–d) and paired lysimeter experiment (e–f) results. (a) Precipitation intensity and phase. (b) Discharge from the Treeline Catchment measured at the Treeline weir. (c) Near-surface volumetric moisture content. (d) Snow depth. (e) Cumulative melt for each collection plot. (f) Difference between the two plots (interpreted to be the lateral flux through the snowpack) and overland flow measurements. The uncertainty in both plots was calculated assuming a tipping bucket error of ±0.97 mL per tip

Objective 1: describe the conditions that promote lateral flow of melt water and rain using dye tracers

We conducted tracer experiments to observe lateral flow by melt water generation and by ROS events in several snowpacks. All experiments were performed on slopes between 10° and 30° (Table I). In each melt water tracer experiment, we applied FD&C blue dye #1. The dye was emplaced with a 3-m length of 2″ diameter plastic pipe, cut in half lengthwise. This half-pipe was placed on the snow surface and rotated upslope to inject a line source. Immediately after each application, the dye line was covered with 6–8 cm of snow to minimize preferential melt because of the altered albedo. The dye was allowed to move with melt water for several hours, after which snowpits were systematically excavated beginning at the furthest downslope extent of dye occurrence to document snowpack properties that promote lateral flow.

Table I. Snowpack conditions and noteworthy results of passive and active tracer tests

Experiment	Slope angle	Snow depth (cm)	Snow temp. (°C)	Air temp. (°C)	Noteworthy snowpack conditions	Key observations
Melt1: Mores Creek Summit 12 March 2007	31	120	0 isothermal	3	Homogenous grains, no ice layers	Dye infiltrated vertically to an undocumented stratigraphic layer 20 cm below snow surface, moved 24.3 m in 21.5 h (1.14 m h^{-1}). Melt water readily flowed from layer in pit face. Minimal infiltration below this layer.
Melt2: Mores Creek Summit 21 March 2007	31	103	0 isothermal	5	5 cm new snow, wet to very wet pack, several saturated layers on stratigraphic boundaries, ice layer at 33 cm depth, average density 367 kg m^{-3}, homogeneous grains	Dye infiltrated vertically to new snow boundary then moved 2.78 m in 2 h (1.58 m h^{-1}), some infiltration to base of snowpack, no dye movement along ice layer
Melt3: Mores Creek Summit 23 March 2007	31	55	0 isothermal	8	wet to very wet, saturated layers, average density 398 kg m^{-3}, highly rounded homogeneous grains, no ice layers	Vertical infiltration to base of snowpack, dye moved 13.6 m in 2 h (6.9 m h^{-1}) along base of snowpack at approximately 1.6 m h^{-1}
Melt4: Treeline Catchment 15 April 2008	28	47	0 isothermal	14	Wet to very wet, saturated layers, average density 398 kg m^{-3}, highly rounded homogeneous grains, no ice layers	Vertical infiltration to base of snowpack, dye moved 13.8 m in 2 h (1.58 m h^{-1}) along base of snowpack
ROS1: Shafer Creek 11 January 2008	10	114	−4 at surface, 0 at base	−2	Dry snowpack, 5 cm fresh snow, no distinct ice layers or density contrasts, average density 209 kg m^{-3}	Lateral flow from pit wall at base of new snow, minimal vertical infiltration
ROS2: Sinker Creek 8 February 2008	13	138	0 at surface, −3 mid pack, 0 at base	2	Dry snowpack, no distinct ice layers or density contrasts, average density 223 kg m^{-3}	Lateral flow in runnels on snow surface and at base of a new graupel layer
ROS3: Sinker Creek 4 March 2008	14	112	−6 at surface, 0 at base	−3	Many ice layers, average density 324 kg m^{-3}	Minimal lateral movement, infiltration through tubes in ice layers to base of snowpack
ROS4: Shafer Creek 14 April 2008	16	120	0 isothermal	10	Many saturated layers within snowpack, average density 357 kg m^{-3}	Substantial lateral movement in many saturated layers

Rain-on-snow tracer experiments were performed with a rainfall simulator consisting of a 4-m mast arm, equipped with eight Spraying Systems Co. H-VV nozzles mounted in 1″ diameter PVC pipe (Figure 3). The water pressure produced by a 12-V pump produced a fine mist from the spraying nozzles, generating raindrops <2 mm in diameter. The rainfall simulator mast was connected to an actuator arm that moved the nozzles in a path parallel to the snowpack surface. Water was pumped from a barrel containing water dyed with Rhodamine WT. For all ROS experiments, the simulator was mounted 1 m above the snowpack, which allowed for an experimental plot area of approximately 4 m^2. Rain was applied for a minimum of 30 min at a rate of 19 mm h^{-1} for each experiment. Upon completion of a rainfall experiment, the entire plot area was incrementally excavated and photographed to document layers producing lateral flow, as well as the occurrence of vertical flow channels between layers. Snow pits at each study site were described in accordance with procedures outlined in the International Classification for Seasonal Snow on the Ground (Fierz *et al*., 2009). Descriptors included density, hardness, liquid water content, and depth of distinct stratigraphic layers. Additionally, 4-m^2 plots were excavated by sequential removal of layer increments to assess the abundance of vertical flow tubes highlighted by the dye.

In the fourth ROS experiment (ROS4), lateral flow collectors constructed of Plexiglas sheets were installed in the snowpit face just below distinct stratigraphic horizons, as these represented the zones within the snowpack where lateral water movement was most likely to occur. Water was drained from the lateral flow collectors to 1-L polyethylene amber sample bottles through gravity drainage. Sample bottles were manually swapped out during the experiment as they filled with runoff water from the pit face (Figure 3c). All liquid water outflow samples collected from the pit face were analysed for Rhodamine WT concentration using standard laboratory procedures with a Turner Designs 10 AU Digital Fluorometer. Blank and duplicate samples were also analysed concurrently as quality controls during laboratory analysis.

Objective 2: compare relative velocities of flow in snow and soil using tracers and geophysical methods

To evaluate the relative velocities of flow in snow and flow in soil, we applied separate tracers to the snow surface and the soil surface, approximately 11.5 m upslope from the Treeline Catchment stream channel (Figure 1). We applied 0.74 L of ~2100 ppm Rhodamine WT to a 1.5 × 0.5-m patch of the snow surface with a misting sprayer. Immediately adjacent to the Rhodamine WT application location, we dug to the base of the snowpack and applied 3.2 L of ~61 000 ppm NaCl to a 1.5 × 0.5-m patch of soil, and then backfilled the excavation with snow. Concentrations of Rhodamine WT and chloride were logged every 5 min in stream using a Hach Hydroprobe.

To document the behaviour of the NaCl tracer on the hillslope, we imaged the plume bi-weekly with two dimensional resistivity surveys. Resistivity surveys are used

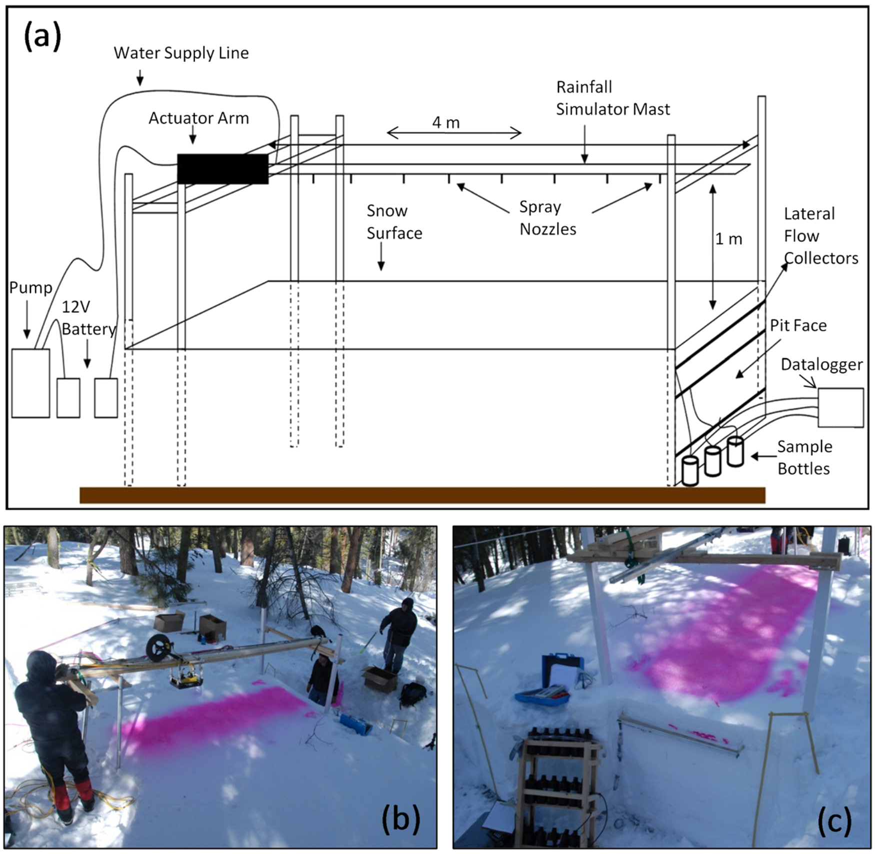

Figure 3. (a) Schematic of rainfall simulator. (b) Photograph of simulator and dyed snow, including a ground-penetrating radar mounted on the actuator (data not presented). (c) Snow pit face with a flow collection trough

to approximate two-dimensional and three-dimensional profiles of resistivity values in the subsurface (Miller *et al*., 2008; Graeff *et al*., 2009). This is accomplished by distributing either a line or grid of electrodes in the soil either in a borehole or along the surface. Data are acquired by using an electrode pair to induce a current into the ground and, at the same instance, measuring the potential difference at one or more additional electrode pairs. The known current and measurement potentials are used to compute apparent resistivity values that are then inverted to estimate resistivity profiles of the subsurface (Ward, 1990; Daily *et al*., 2005). In this study, we used an IRIS Syscal Pro multi-electrode resistivity instrument. The electrode array consisted of 18 electrodes along a line with 1-m spacing between electrodes. The electrode line followed the slope gradient with the first electrode placed at the stream bank. Electrodes consisted of 1-m-long galvanized rods that were pushed through the snowpack and approximately 0.3 m into the soil. The NaCl tracer was placed on the soil surface between electrodes 12 and 13 (Figure 4). The electrodes were left in place throughout the study to minimize differences between surveys. Data were acquired in a dipole–dipole geometry with up to 16 different potential electrode pair measurements for each current injection. We used the commercial software package RES2DINV to invert the data (Loke and Barker, 1996). The programme utilizes a least-squares inversion algorithm coupled with time-lapse constraints to analyse changes over time.

Resistivity profiles were acquired both before and immediately after the NaCl application on 12 March. For purposes of this analysis, the resistivity profile collected immediately after tracer placement is defined as the initial profile, and all subsequent profiles show changes relative to this initial measurement. Although we expect resistivity to change generally as a function of soil moisture, resistivity profiles taken on 16 March, 15 April, and 10 June show resistivity decreases of up to 50% in the vicinity of the NaCl tracer application. This substantial localized change indicates the high sensitivity of the method to the NaCl tracer and provides a solid basis for mapping the bulk tracer movement in the soil.

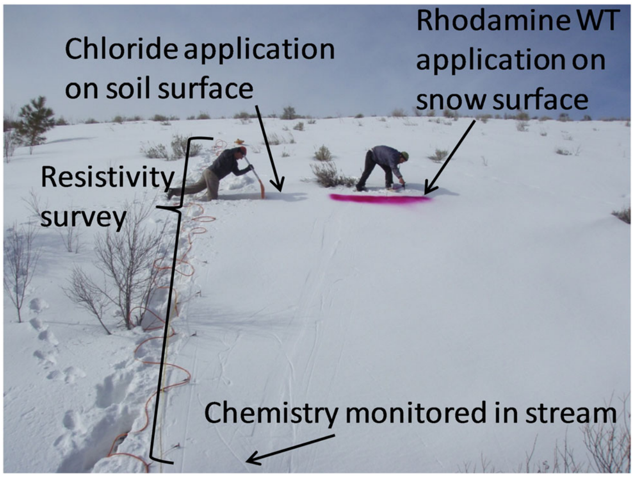

Figure 4. Illustration of relative tracer velocity experimental design

Objective 3: measure quantities of downslope water movement within snow in a paired lysimeter experiment

We installed two adjacent 4-m^2 snowmelt lysimeters to collect water exiting the base of the snowpack on a northeast-facing, 20° hillslope adjacent to the Treeline Catchment (Figure 5). One lysimeter was blocked from lateral upslope contributions, whereas the second lysimeter was open to inputs from ~20 m of upslope contributing area. Both plots were blocked on the downslope side to prevent melt water from exiting the collection area. Implicit in this experimental design is the hypothesis that the open and blocked lysimeter volumes should be equal if lateral flow is *not* hydrologically relevant. If the open lysimeter collects more water than the blocked lysimeter, that difference is attributed to lateral flow from upslope. The wooden lysimeter sides were ~20 cm high and lined with

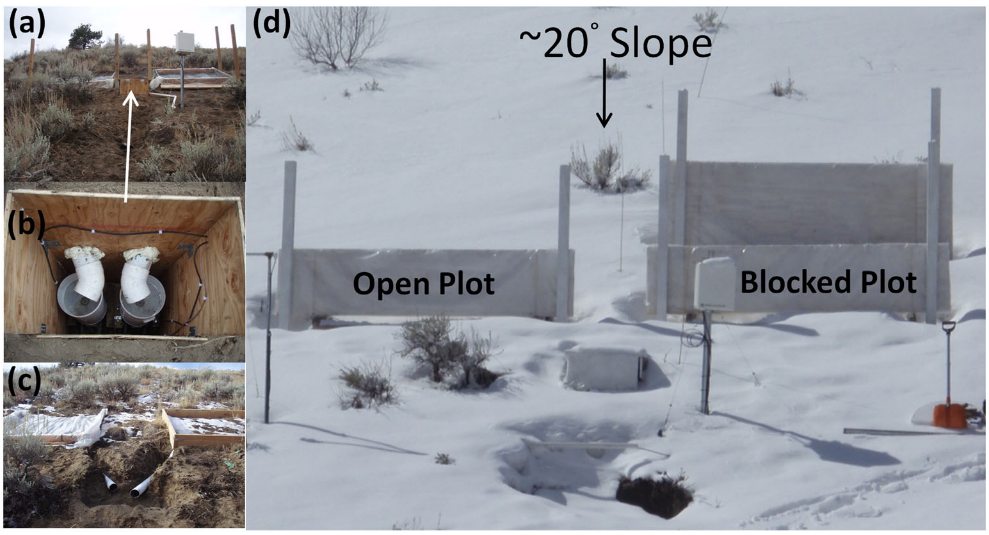

Figure 5. Paired lysimeter experiment photos. (a) Paired lysimeter experiment prior to snowfall. (b) Tipping buckets and plywood housing. (c) Buried melt water pipes (taken during construction). (d) Completed experiment during winter

polyethylene plastic sheeting (Figure 5). Lysimeter barriers were installed in two stages. In the early snow accumulation season, the 20-cm sides served as barriers. The blocked lysimeter was enclosed on all four sides, whereas the open lysimeter was only enclosed on three sides with the upslope edge completely open. We waited until approximate maximum snow accumulation (6 March) before inserting 1-m tall barriers in order to minimize snowpack disturbance and artificial melt and also to ensure natural snowpack accumulation. During the time between when snow accumulated above the 20-cm sides and when the tall barriers were inserted, lateral flow in the upper snowpack was not collected. Consequently, measured lateral flow volumes during this period are minimum values. Melt water was routed to a downslope corner of each plot from where it was routed 2 ft underground through 4-in drain pipes to tipping bucket gauges housed in plywood boxes and buried 4 ft below ground surface. Cumulative tips were logged every 15 min on a Campbell Scientific CR800 data logger. Minimal blockages due to ice buildup occurred during the snow season.

The tipping bucket gauges used in this experiment were originally designed to accurately measure small precipitation volumes, not large melt water volumes sourced from a 4-m^2 collection plot. During high flow rates, the buckets cannot tip fast enough, and some water is consequently spilled and not measured. To correct for this, we developed a calibration curve that relates the tip rate to the volume of water per tip entering the tipping bucket instrument (Figure 6). We determined the relationship to be linear over the range of our data ($r^2 = 0.87$) and calculated an RMSE of 0.97 mL per tip. We applied this linear relationship and calculated uncertainty to all snowmelt data presented herein.

RESULTS

Objective 1: describe the conditions that promote lateral flow of melt water and rain using dye tracers

Lateral flow occurred to some extent in all melt water and ROS tracer experiments. Complete snowpack descriptions can be found in Whitson (2009). Here, we summarize general results (Table I).

In all melt water experiments, dye moved within the snowpack tens of metres downslope with minimum velocities ranging from 1 to 7 m h^{-1} (Table I). The experimental snowpacks can be classified as those with ice layers and those without. Two mid-winter snowpacks exhibited no discernible ice layers and were less dense and less wet and contained more snow layers than two ice-rich snowpacks. In both ice-free cases, lateral dye movement occurred along multiple layers within the snowpack, including at the base of a layers of new snow that fell the previous night (Figure 7a, b). In one case, lateral movement occurred along a nearly imperceptible layer within the snowpack characterized by very minor grain size differences that were only detected because of the presence of dye. In this case, dyed water flowed from the pit wall during excavation indicating that dye was moving by advection rather than by diffusion through the saturated layer (Figure 8). The two late spring ice-rich snowpacks were otherwise homogeneous with no perceptible stratigraphy. In both ice-rich snowpacks, lateral movement within the snowpack was minimal. Melt water moved vertically to the ice layers through vertical conduits and then to the base of the snowpack, where substantial lateral movement occurred.

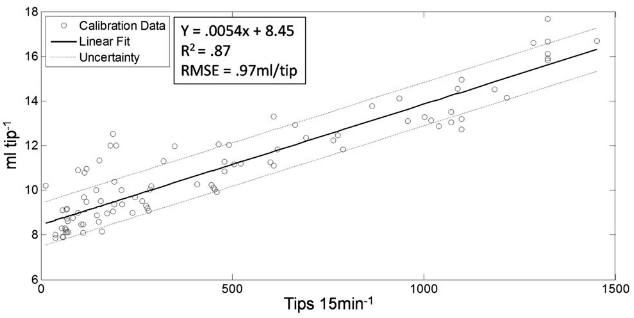

Figure 6. Tipping bucket calibration data

The four ROS experiments were performed over a greater diversity of snowpack conditions than were covered by the melt water tracer experiments (Table I). Lateral flow was observed from the observation pit wall at the downhill side of the rain event during three of the four experiments (ROS1,2,4). In the experiment that did not exhibit flow from the pit wall (ROS3), some flow was observed at the base of the snowpack. However, most water exited the base of the ROS3 snowpack prior to reaching the downhill observation pit. As was true for the melt experiments, the ROS experiment on the snowpack with considerable ice (ROS3) exhibited the least lateral flow.

In the ROS2 experiment, lateral flow was observed at and near the snow surface, with the majority of flow occurring just below the snow surface. This flow occurred in a zone where there were no ice layers and no discernible discontinuities in hardness, grain size, or grain shape. No water infiltrated through the snowpack to the ground surface during post-experiment excavation of the experimental plot (Figure 9a). Lateral flow was also observed during the second ROS experiment as rivulets at the snow surface, as well as concentrated flow above and within a graupel layer just below the snow surface. Again, lateral flow occurred within this snowpack without the presence of ice layers in the flow zones, and no water infiltrated to the base of the snowpack.

The ROS4 experiment occurred on a snowpack containing several thin saturated layers. There were numerous abrupt changes in hardness but no distinct or continuous ice layers within the profile. Lateral flow collectors were installed in the snow pit face below saturated layers at heights above ground surface of 89 and 100 cm, as well as at the snow–ground interface (Figure 9b). Water began draining from the pit face into the lateral flow collector trays installed at 89- and 100-cm heights almost immediately after they were inserted and continued to drain from these layers for the duration of experiment. The majority of observed lateral flow occurred at the lower layer (89 cm). No

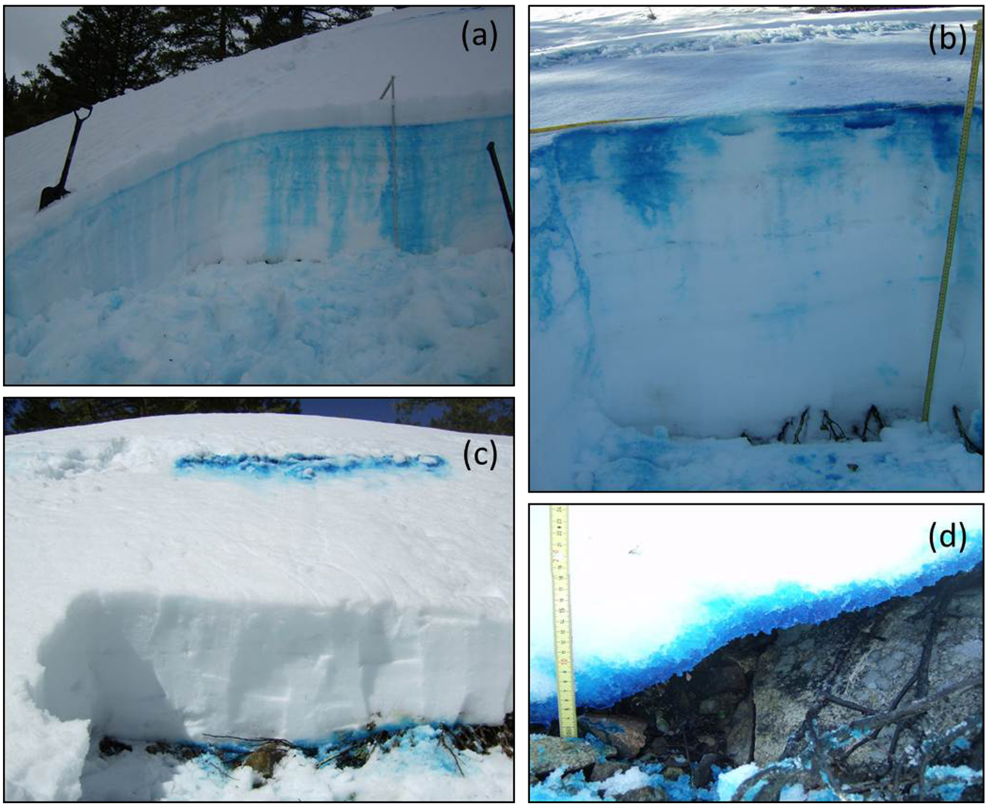

Figure 7. (a) Excavated snow pit at MCS Site 1. Injection site is visible in the top-right corner of the photograph approximately 2.5 m upslope. (b) Snow pit face at MCS Site 2. Extensive down slope dye movement occurred very near the snow surface, through layer of fresh snow. (c) Snow pit face at MCS Site 3. Lateral water movement occurred at the base of snowpack. (d) Close-up view of the base of MCS Site 3 snowpack. Dyed liquid water is visible and moving laterally down slope at the snow–ground interface

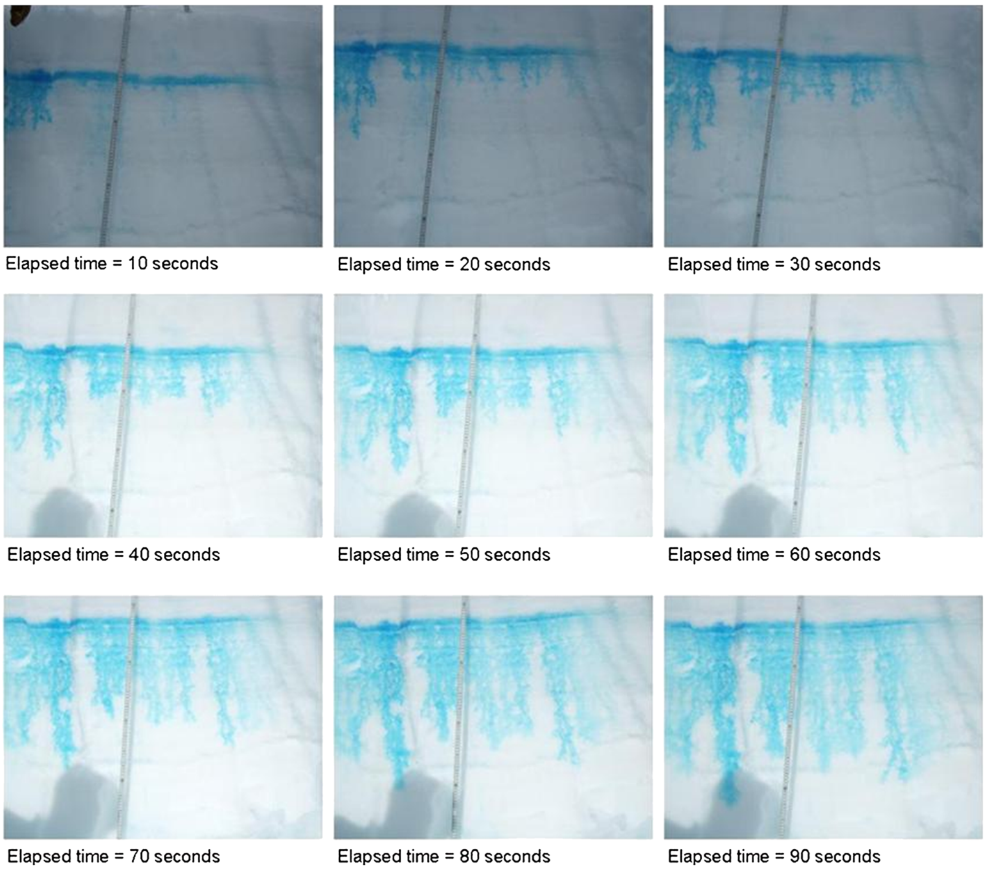

Figure 8. Time-lapse photography of snow pit face at MCS Site 1. Liquid water is draining from a saturated layer near the surface of the snowpack

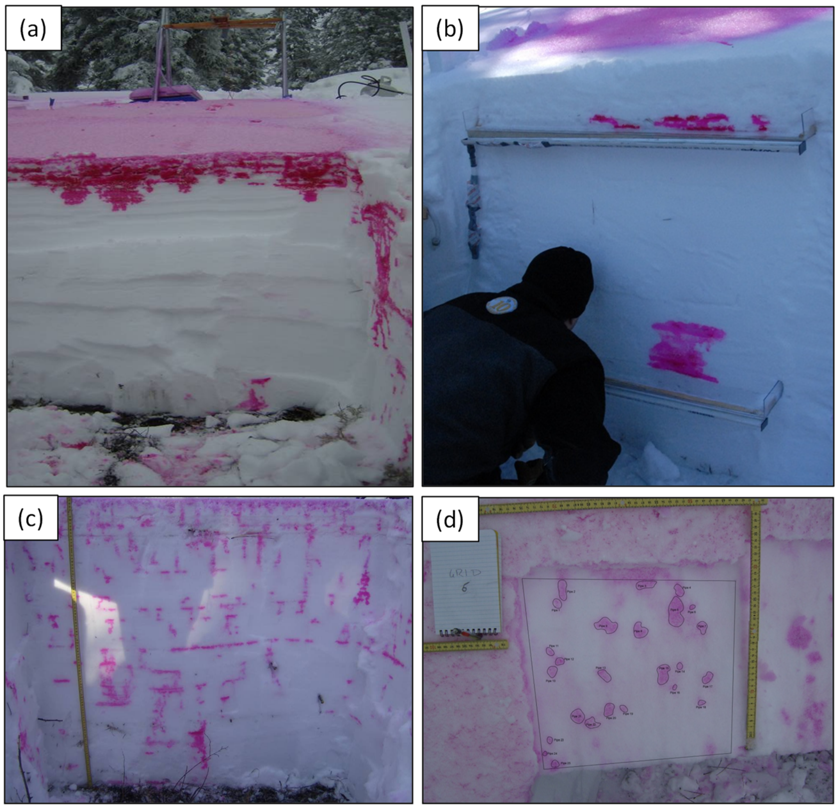

Figure 9. Photographs illustrating select results from ROS experiments. (a) ROS 1 snow pit face post-experiment. Note that there is no vertical infiltration of rainwater to the base of snowpack; water movement was confined to upper 20 cm of snowpack. (b) Collection troughs placed below seeping layers during ROS4. (c) ROS 3 snow pit face. Note the presence of numerous vertical flow channels. Most channels occurred along vertical ice columns or vegetation branches. (d) Photo of vertical flow channels after photo has been dimensionalized

lateral flow from the pit face was observed within layers below this lateral flow collector. The peak flow rate for this collector was 9.67 mL s^{-1}, with an average flow rate of 7.18 mL s^{-1}. This represents a flow rate from this layer of nearly half the input rate (21.9 mL s^{-1}). The peak discharge from the upper layer (100 cm) was 1.94 mL s^{-1}, with an average flow rate of 1.2 mL s^{-1}. Rhodamine WT concentration values (not shown) for samples collected at these layers were near the input concentration, indicating little snow melt within the snowpack. Most of the dilution of the runoff water was likely due to natural snowmelt, considering the warm ambient air temperature.

In all ROS snowpacks, vertical conduits were present between lateral flow layers. Approximately 2–6% of the each 4-m^2 plot area was composed of vertical flow channels between layers (Figure 9c, d). Some vertical channels occurred around previously formed ice columns, whereas others occurred around vegetation.

Objective 2: compare velocities of flow in snow and soil using tracers and geophysical methods

Rhodamine WT was applied to the snowpack surface simultaneously with the application of NaCl at the soil surface to evaluate the relative travel time to the stream. Following the application of Rhodamine WT to the snow surface on 12 March, we observed downslope water movement through the near-surface layers of the snowpack. By 14 March, the Rhodamine WT had travelled at least 5–6 m downslope from the application point, and by 17 March, the Rhodamine WT infiltrated into the snowpack and was no longer visible on the surface. The first measurable increase of Rhodamine WT in the stream occurred 4 days after application (16 March, Figure 10). The next major pulse of Rhodamine WT was recorded approximately 2 months after initial application on 17 May. Note that this peak coincided with a precipitation event that resulted in the rapid accumulation and melt of about 20 cm of snow. Prior to this snowfall event, the entire snowpack had melted (Figure 10). Therefore, these later peaks in Rhodamine WT concentrations can be attributed to the flow through the soil. Rhodamine WT concentrations in the stream increased again during two events at the end of May and the beginning of June.

We observed no significant change in chloride concentrations in the stream during the study period (Figure 10). Although it is possible that we simply missed the arrival of NaCl in the stream or that the tracer became too diluted to measure, the resistivity survey supports the conclusion that the NaCl tracer did not reach the stream (Figure 11). Two days after placement of the tracer, it was apparent that the majority of the tracer moved up to 1 m vertically into the soil column, and a small amount of tracer had moved approximately 1 m downslope near the soil surface. By 15 April, after all snow had melted, little tracer movement was observed, and the majority of the tracer appeared to remain immediately below the injection point, whereas some tracer had moved downslope by

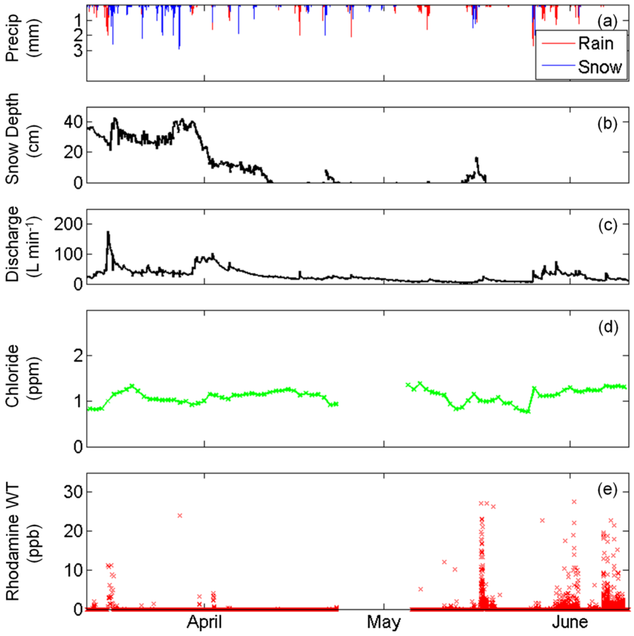

Figure 10. Summary of 2011 relative tracer velocity data. (a) Precipitation phase/intensity for the duration of the study. (b) Snow depth in the Treeline Catchment. (c) Discharge (LPM) at the Treeline weir. (d) Daily median chloride concentrations (ppm) measured in the stream channel. (e) In-stream Rhodamine WT concentrations. Rhodamine WT and chloride concentrations were measured using a continually deployed Hach Hydroprobe that recorded measurements every 5 min. The late April–early May data gap was due to instrument malfunction

approximately 3 m. By 10 June, a lobe of the tracer had moved downslope approximately 3–4 m near the surface, still roughly 8 m from the stream. The majority of the tracer appeared to remain just below the injection point. Note that the resistivity within the tracer zone changed little over most of the observation period indicating that there was little change in concentration. We can conclude that although some movement is observed, the rate of lateral transport through the soil is substantially slower than that through the snow. Water delivery over this 12-m transect occurred in just 4 days by lateral flow through the snow. Delivery by lateral flow through the soil was not observed in this study, but our results suggest that residence time in the soil is at least several months for water that enters the soil column just 12 m from the stream.

Objective 3: measure quantities of downslope water movement within snow in a paired lysimeter experiment

A blocked lysimeter collected melt water from directly above its surface area, whereas an open lysimeter collected melt water from an upslope contributing area. During the 2010–2011 snow season, the blocked lysimeter collected 1925 ± 202 L, and the open lysimeter collected 2884 ± 283 L (47% more than the blocked lysimeter). Therefore, we estimate that 959 ± 485 L or 480 ± 283 L m^{-1} of hillslope width of water was routed laterally through the snowpack to the open lysimeter. Most of the lateral flow recorded by the lysimeters occurred during four ROS events (Table II). During these events, the open lysimeter collected between 31% and 136% more water than the blocked lysimeter. Measureable volumes of lateral flow were observed during every measurable melt event throughout the season. All of the individual melt/lateral flow events recorded by the lysimeters coincided with some degree of rain. Conversely, every instance of measurable infiltration to the soil (as evidenced by increases in soil moisture) also produced a measurable amount of lateral flow (compare

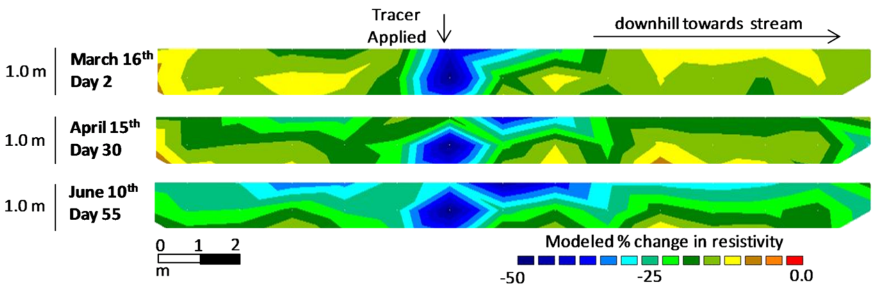

Figure 11. Electrical resistivity tomography images of a NaCl tracer applied to a hillslope. Each image displays the change in resistivity since the date of application

Table II. Summary of lateral flow events

Lateral flow start	Open plot runoff (L)	Blocked plot runoff (L)	Lateral flow (L)	Lateral flow increase (%)	Event rain (mm)	Blocked plot runoff/rain ratio
11 Dec, 22:45	85 ± 9	36 ± 4	49 ± 13	136	2	5.8
14 Dec, 6:00	159 ± 13	68 ± 7	91 ± 20	134	15	1.2
16 Jan, 0:45	605 ± 51	293 ± 29	312 ± 80	106	53	1.4
9 Mar, 13:00	135 ± 14	65 ± 7	70 ± 21	108	5	3.2
13 Mar, 16:00	118 ± 12	78 ± 8	40 ± 20	51	7	1.8
15 Mar, 11:30	229 ± 21	156 ± 15	73 ± 36	47	26	1.5
30 Mar, 2:30	441 ± 43	290 ± 29	151 ± 72	52	10	7.5
1 Apr, 12:00	159 ± 16	121 ± 13	38 ± 29	31	4	7.6

Figure 2c and 2e). The most significant melt event of the season occurred on 15–17 January, during which approximately 4.9 cm of rain fell on the snowpack over a 27-h period. During this storm, the open lysimeter collected 106% more water than the blocked lysimeter (605 ± 51 and 293 ± 29 L, respectively). This single event accounted for approximately 33% of the total lateral flow observed during the snow season.

Because our interpretation of the lysimeter, results are predicated on the assumption that the open and blocked lysimeters will collect the same amount of water in the absence of lateral flow, we used a variety of methods to verify that the two plots accumulated the same amount of snow and that the relative collection efficiency of the two plots was the same. Snow surveys performed in the immediate vicinity of each lysimeter plot show that between 12 March and 15 April neither plot preferentially accumulated or melted snow (Eiriksson, 2012). These results also show that the ultrasonic snow depth sensor located in the basin adjacent to the lysimeters on a slope of similar aspect provides an accurate representation of snow depth at the lysimeters (Figure 12). As an additional check on the accuracy of the lysimeters, we compared precipitation gauge data to total lysimeter outflow. During the 2010–2011 snow season, the Treeline Catchment precipitation gauge collected a wind corrected total of 54 cm of water (rain and snow combined). This converts to an expected total of 2031 L of precipitation on each plot, which is within the uncertainty of the measured snowmelt outflow from the blocked plot (1925 ± 202 L) (Table III). Further, the expected melt volume was considerably less than the total open lysimeter snowmelt outflow volume (2884 ± 283 L). Because the open plot collected more than the expected volume derived from precipitation data, we interpret that the additional water volume was sourced upslope and delivered to the collection plot laterally through the snowpack.

To assess the relative collection efficiencies of the two plots in a controlled environment, after the snow was gone, we applied ~6 L of water on each plot and collected it at the outlet, just before passing through the tipping buckets (to eliminate the error associated with tipping bucket volume measurements). We repeated this experiment three times on both plots, and in all cases, each plot recovered at least 95% of the applied water (Table IV), indicating that the plots at the paired lysimeters have high and comparable collection efficiencies.

Table III. Analysis of lysimeter performance

Total rain (mm)[a]	168
Total snow water (mm)[a]	372
Total precipitation (mm)[a]	540
Total precip. on blocked plot (L)[a]	2031
Total control (L)[b]	1925 ± 202
Total experimental (L)[b]	2884 ± 283

[a] Rain gauge data.
[b] Lysimeter data.

DISCUSSION

Occurrence of lateral flow in snow

Our tracer experiments and many others (Schneebeli, 1995) suggest that lateral flow of melt water and rain in a sloping snowpack is a common occurrence, rather than simply an impediment to vertical percolation. Water can travel substantial lateral distances in near-surface rivulets, in thin saturated or near saturated zones on stratigraphic boundaries, and along the basal snow–soil or snow–air interfaces. Each boundary presumably forms a hydraulic conductivity contrast that promotes ponding and subsequent slope-parallel flow. Regardless of the water source, ice layers can form when percolating water refreezes at

Table IV. Collection efficiencies of block and open plots

Plot	Applied volume (L)	Recovered volume (L)	% Recovered
Open	6.3	6	95
Open	6.4	6.1	95
Open	6.5	6.2	95
Blocked	5.8	5.6	97
Blocked	6.3	6	95
Blocked	6.3	6	95

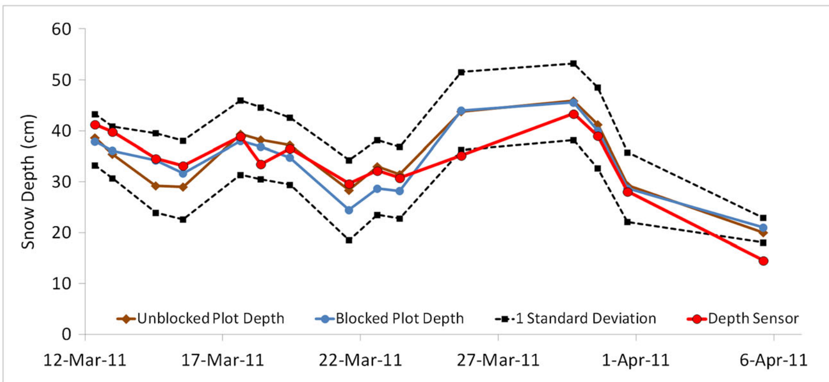

Figure 12. Comparison of manually measured snow depth on the ground adjacent to runoff plots with automated snow depth measurements near the meteorological station

stratigraphic boundaries (Wolken *et al.*, 2009). However, lateral flow on ice layers was not important in this study. Lateral flow along stratigraphic boundaries within the snowpack, whether generated by rain or melt, is most prevalent in snowpacks that have not experienced considerable melt–freeze metamorphism. As water and vapour migrate through a snowpack, grains become rounded, and stratigraphic boundaries are destroyed. When lateral flow does occur, occasional vertical breakthroughs in stratigraphic boundaries can route water to deeper boundaries, which can further route water laterally such that water takes a stepped path from its source to the base of the snowpack (Langham, 1974). At a snowpack base, water can travel as apparent overland flow until it exits the snow matrix at some point downslope from its point of origin. The net result is that there is a downslope migration of water prior to entering the land component of the catchment hydrologic system, the implications of which are discussed in subsequent sections.

The fact that we were able to induce steady-state flow through the ROS4 snowpack suggests during ROS events a snowpack can transmit considerable quantities of water downslope. Although it is possible that outflow during ROS events is in part due to snowmelt, rain is often a minor component of the energy balance during a ROS event (Marks *et al.*, 1998). Additionally, Rhodamine WT concentrations in the ROS4 outflow suggest that melt water contributions were negligible. We suggest that lateral flow in snow can help explain the difficulties that hydrologic modellers face when simulating hydrographs resulting from ROS events (Stratton *et al.*, 2009). In the case of melt-generated lateral flow, the snowpack is simultaneously the source of water and the porous matrix; generation of melt water consumes the matrix through which it flows, thereby limiting the water available for lateral transport. Unfortunately, we could not measure the volume of downslope flow during major melt events because most events were accompanied by rain.

Lateral flow in snow as a hillslope hydrologic pathway

Lateral flow in snow adds supra-surface hydrologic pathways to the suite of commonly known overland and subsurface pathways. Although flow can occur throughout a snowpack, the similarity between lateral flow in snow and overland flow measured on a nearby hillslope (Figure 2f) suggests that in the Treeline Catchment flow occurred primarily in the base of the snowpack (Figure 7d). As a consequence of its design, the overland flow collector will collect water flowing in the basal 11 cm of the snowpack, in addition to the classical overland flow. Given the high hydraulic conductivity at the Treeline Catchment of $288\,\mathrm{mm\,h^{-1}}$ in the near-surface soils (Gribb *et al.*, 2009), it is unlikely that overland flow by infiltration excess, called Hortonian overland flow, is ever a dominant runoff generation process. However, small volumes of Hortonian overland flow have been measured in the absence of a snowpack in the Treeline Catchment. For example, 41 mm of rain fell onto bare soil over a 75-h period from 4 October 2011 to 7 October 2011 yielding $2.5\,\mathrm{L\,m^{-1}}$ of overland flow. It is noteworthy that the magnitude of overland flow measured during this event is considerably less than the magnitude of overland flow measured in the presence of a snowpack (Figure 2e, f) confirming that snowpacks enhance lateral redistribution of water in this site by facilitating apparent overland flow.

The hydrologic importance of lateral flow in snow depends on the length of lateral flowpaths and the quantities of transmitted water. Quantities of downslope flow are discussed in the following section. The contributing length of lateral flow in snow can be estimated by dividing the volume of lateral flow for an event by the product of plot width and water input depth (Ohara *et al.*, 2011). This is only possible for ROS events in which net water storage in the snowpack is negligible. Equilibrium discharge is difficult to achieve in a snowpack as melt is often accompanied by rain, which can further generate melt, all of which can be stored within the snowpack. However, the 14 December and 16 January ROS events produced near-equilibrium flow as the ratios of blocked lysimeter outflow depth over rain depth were close to 1 (Table II). Accordingly, contributing lengths for the 14 December and 16 January ROS events were 3.1 and 2.9 m, respectively. These are minimum estimates because the calculation assumes that all water raining on a 2-m-wide by 3-m-long area upslope of the open lysimeter is transported laterally to the lysimeter. In reality, some water generated in the area may leave the base of the snowpack, and the actual contributing lengths may be longer.

The contributing lengths calculated above demonstrate that lateral flow in snow can deposit water to the soil surface at positions downslope from its melt or rain source. In the days immediately following the application of Rhodamine WT to the snow surface, the tracer appeared in the stream, likely due to direct contributions from lateral flow in snow (Figure 10). Beginning in the middle of May, after the snow had completely melted, Rhodamine WT again was observed in the stream. Meanwhile, chloride from the NaCl addition to the soil surface never reached the stream (Figure 11). It is possible that Rhodamine WT was transported several metres downslope prior to complete snowpack ablation, placing it in an advanced downslope position for later transport to the stream. Because the chloride tracer did not have the benefit of downslope transport through the snowpack, its soil travel distance was much greater, and it failed to reach the stream channel during the duration of our study. This interpretation may explain the observation by Williams *et al.* (2009), wherein it was observed that near-surface soil moisture content tends to increase downslope in the Treeline Catchment despite the lack of evidence supporting lateral flow in near-surface soils at the same site (Makram-Morgos, 2006).

Contributions of lateral flow in snow to runoff generation

The coincident timing of peaks in lateral flow in snow and streamflow (Figure 13) suggests that a certain amount of water is routed downslope and delivered directly to the

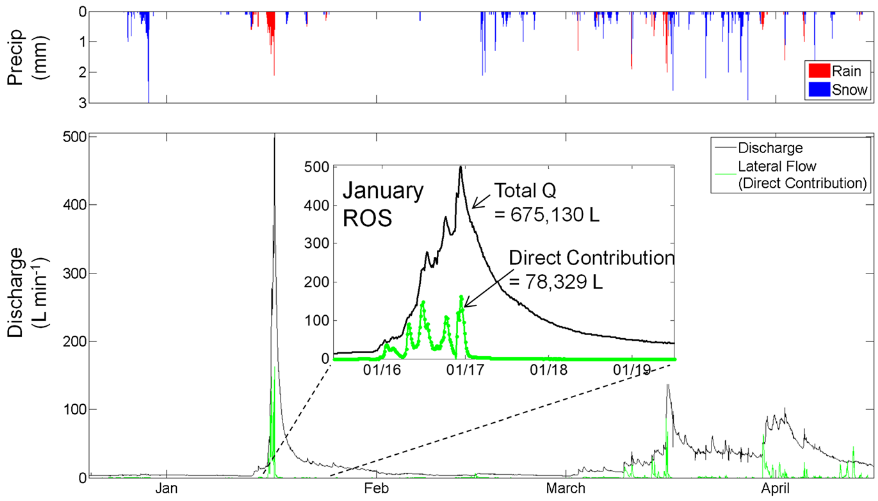

Figure 13. Calculated direct contribution of lateral flow in snow to Treeline Catchment discharge

stream channel, bypassing the soil. This results in an effective area of direct contributions of event water that is greater than the channel area. If we assume that the lateral flow per unit hillslope width is constant throughout the catchment and that the effective contributing hillslope length of lateral flow in snow is less than the actual hillslope length, the volume of direct contributions from lateral flow in snow can be estimated as the lateral flow per unit hillslope width times twice the channel length (~250 m). With these assumptions, lateral flow in snow contributed as much as 12% of total discharge for the 17 January ROS event (Figure 13). If vertical percolation from the snowpack were the only source of direct event contribution, the contributions over the 0.3-m-wide channel would only account for 1% of the total storm discharge. Although the estimated 12% direct contribution during the January ROS event is a large component of total streamflow, this percentage does not account for event water that travels through the soil to the stream during the event. This rough analysis demonstrates that lateral flow in snow can elevate event water contributions to streamflow. Although this analysis was for a ROS event, it is reasonable to assume that melt events can also contribute snowmelt directly to the stream from areas extending beyond the channel area.

Direct contributions of melt water to a stream by lateral flow in snow may help explain high proportions of event water that has been reported in literature. Several decades of research have demonstrated that rain-generated storm hydrographs tend to be dominated by pre-event water, or water that was in the catchment prior to the storm. A review of the topic nearly two decades prior to this research demonstrates the pervasiveness of this concept (Buttle, 1994). Although a rigorous analysis has not been conducted, Buttle (1994) suggested that storms generated by snowmelt tend to have lower pre-event water contributions, or higher event water contributions. Indeed, Yenko (2003) reported event water contributions to a snowmelt hydrograph of 59–65% in the Treeline Catchment. This includes event water that travels overland or through soils, mixing with pre-event water. It is difficult to explain such high proportions of event water using soil-based runoff generation mechanisms given that overland flow is a rare occurrence in this catchment. Lateral flow in snow offers a pathway of rapid, high volume delivery of water that can increase event water contributions above those which soils are capable of transmitting.

CONCLUSIONS

This study demonstrates that lateral flow of rain and melt water in snow can redistribute water in quantities sufficient to impact hillslope and catchment scale hydrologic processes. Near-surface runnels, saturated or near saturated zones on stratigraphic boundaries within the snowpack, and the snow–soil or snow–air interface at the snowpack base can transform vertically percolating water into lateral flow. In the locations studied here, cold, midwinter conditions tended to promote flow within the snowpacks, whereas highly metamorphosed ripe snowpacks displayed water movement in the basal layers. Numerous vertical conduits in ice layers, when present, tended to minimize their effectiveness for routing water laterally. Along all boundaries, occasional features allow water to break through and percolate to deeper boundaries, which can further route water laterally, such that water is deposited on the soil surface in advanced downslope positions relative to its point of origin, or direct discharge to a stream by extending the effective channel area. In one ROS event in the Treeline Catchment, where previous studies have documented anomalously high proportions of new water in streamflow during snowmelt, direct contributions of lateral flow in snow contributed approximately 12% of total runoff. This result suggests that lateral flow in snow may be responsible, in part, for the rapid delivery of water to stream during ROS events, as well as anomalously high contributions of event water during snowmelt hydrographs. Our study suggests that hydrologic predictions for the timing and magnitude of ROS events and perhaps snowmelt events could be significantly improved if models incorporate lateral flow in snow as a water routing mechanism.

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