Aspect and soil textural controls on snowmelt runoff on forested Boreal Plain hillslopes

Todd Redding and Kevin Devito

ABSTRACT

Plot studies were conducted on a jack pine forest with sandy soil and aspen forests with sandy and loam soils to examine the controls of slope aspect, soil texture and fall soil moisture content on near-surface snowmelt runoff and infiltration. It was hypothesized that near-surface runoff would be greater from north-facing slopes on loam soils with increased fall soil moisture content. Fall soil moisture had no measurable effect on spring snowmelt runoff. Infiltration of snowmelt dominated (drainage coefficients 53–100%, median 87%) over near-surface runoff (runoff coefficients 1–65%, median 7%) for most plots. Runoff was related to concrete frost at the mineral soil surface. In contrast to the processes hypothesized, south-facing hillslopes with sandy soils generated greater runoff than north-facing slopes or sites with finer-textured soils. These results were due to greater concrete frost development resulting from periodic spring snowmelt and re-freezing in the upper soil. South-facing hillslopes with sandy soils featured lower canopy cover, allowing greater solar radiation to reach the snow surface which led to the formation of concrete frost and faster melt rates resulting in near-surface runoff. Where hillslopes are connected to receiving surface waters by continuous concrete frost, snowmelt runoff at the watershed scale may be enhanced.

Key words | boreal plain, concrete frost, frozen soil infiltration, snowmelt lysimeter, snowmelt runoff

Todd Redding* (corresponding author)
Kevin Devito
Department of Biological Science,
University of Alberta,
CW 405 Biological Sciences Centre,
Edmonton, Alberta,
Canada, T6G 2E9
E-mail: *todd.redding@gmail.com*
*Present address: Department of Geography, Earth & Environmental Science, Okanagan College,
583 Duncan Street West, Penticton, BC,
Canada, V2A 8E1

INTRODUCTION

In the Boreal Plains ecozone, snowmelt produces a large quantity of water during a relatively short period at a time of year when evapotranspiration is very low and moisture surpluses are possible. This is in contrast to the growing season where, in the subhumid climate of this region, precipitation (rain) is less than potential evapotranspiration (Devito *et al.* 2005a). Infiltration of snowmelt and early spring rain is a major source of annual groundwater recharge on some landforms in the Boreal Plains (Smerdon 2006). Climate change has the potential to decrease snow accumulation and accelerate the timing and rate of melt in this region (Barnett *et al.* 2005). In the western boreal forest, regional groundwater recharge and streamflow volumes are low; it is therefore important to understand the controls on snowmelt runoff and infiltration from hillslopes on the Boreal Plain to understand and model the potential effects of climate change on hydrological processes and landscape-scale runoff. Rapid changes in land use on the Boreal Plains have the potential to alter snow accumulation and melt dynamics by removing forest cover, which will result in decreased interception of snow and increased melt rates. To accurately predict the effects of resource development, it is critical to understand the natural controls on snowmelt infiltration and runoff.

In the boreal forest, infiltration often dominates over surface runoff at the hillslope scale during the snowmelt period (Mace 1968; Kachanoski & de Jong 1982; Elliott *et al.* 1998; Kalef 2002; Whitson *et al.* 2004) due to the high infiltration capacity of most forest soils, even under frozen conditions (Pierce *et al.* 1958). A large available soil moisture storage capacity prior to snowmelt plays an important role in reducing snowmelt runoff in soils that have high infiltration rates (Kalef 2002; Devito *et al.* 2005b). It has been suggested that under certain conditions, near surface

doi: 10.2166/nh.2011.162

runoff (over the mineral soil surface, but through the forest floor layers) may occur due to the presence of concrete frost that impedes infiltration (e.g. Stein *et al*. 1994). Given the large heterogeneity in texture of the deep, dry soils on the Boreal Plains (Whitson *et al*. 2004; Devito *et al*. 2005a) and the potentially high infiltration rates during frozen conditions (Kalef 2002), there is considerable uncertainty about the importance and magnitude of snowmelt runoff contributions from upland forested hillslopes to wetlands.

Surface runoff occurs when the snowmelt rate is greater than the surface infiltration rate. The dominant controls on snowmelt runoff have been shown to be the amount of snow, autumn soil moisture content and snowmelt rate (Zhao & Gray 1999). Soils with high autumn soil moisture content can have increased runoff as wetter soils have a greater potential to develop frost, which may impede infiltration rates. In addition, soil that is wetter in the autumn has less available soil moisture storage during snowmelt (Zhao & Gray 1999). Soil frost may also develop due to meteorological conditions during winter and snowmelt periods that involve periodic melting and refreezing events (Mace 1968; Proulx & Stein 1997). Slope aspect can have a large influence on snowmelt rate and therefore runoff potential. Greater snowmelt rates typically occur on south-facing than north-facing slopes (Hart & Lomas 1979; Harms & Chanasyk 1998; Woo 2005) due to greater incoming solar radiation (Bonan 2002). However, greater snowmelt runoff may occur on north-facing slopes related to the presence of frozen soils (Carey & Woo 1999) or to decreased available storage due to reduced evaporation and transpiration (Cline *et al*. 1977; Hart & Lomas 1979). These trends may be enhanced or moderated by soil texture, which is a control over water storage and transmission properties, and by forest canopy vegetation (deciduous and coniferous) which influences snow interception and radiation transmission.

On the western Boreal Plains, high spatial variability in soil texture is related to the spatial distribution of glacial landforms. Vegetation tends to vary with landforms. Aspen grows on uplands with soils that have greater water-holding capacity, while jack pine is the dominant forest canopy species on coarse-textured soils with less water-holding capacity. The variation in forest vegetation communities may result in high spatial variability in snow accumulation and melt rates, and potentially in runoff. Therefore, understanding where and when upland snowmelt runoff contributes to down-slope wetlands or surface waters is critical to understanding the potential impacts of resource development or climate change. The objective of this research was to examine the controls on snowmelt runoff generation on forested hillslopes on the western Boreal Plains. Near-surface snowmelt runoff, infiltration and snowmelt season water balances were measured on two landforms and within two vegetation types (outwash-jack pine, outwash-aspen and moraine-aspen) to assess the relative importance of aspect (north- and south-facing slopes), soil texture (sand, loam) and autumn soil moisture content (autumn irrigation, non-irrigated) on snowmelt runoff. Specifically, it was hypothesized that near-surface runoff would be greater from north-facing slopes on loam soils with increased autumn soil moisture content.

METHODS

The snowmelt runoff experiments were carried out at the Utikuma Region Study Area (URSA) near Utikuma Lake (lat. 56° N, long. 115°30′ W) in north-central Alberta (Devito *et al*. 2005a). The site is located within the Boreal Plains ecozone (EcoRegions Working Group 1989), and has a mean annual temperature of 1.2 °C and mean monthly temperatures over the snowmelt period of –12.1, –5.5 and 3.4 °C in February, March and April, respectively (Marshall *et al*. 1999). The mean annual precipitation is 481 mm with 137 mm falling as snow, and annual potential evapotranspiration is 518 mm (Marshall *et al*. 1999). Mean month-end depth of snow on the ground at Slave Lake (ca. 100 km south of URSA) is 0.25, 0.16 and 0.01 m for February, March, and April, respectively (Environment Canada 2004). The long-term (1987–2001) mean annual runoff for the Red Earth Creek watershed (619 km^2, ca. 70 km north of URSA) is 67 mm a^{-1}, and ranges between 4 and 246 mm a^{-1} (median 58 mm a^{-1}), which corresponds to runoff coefficients between 1 and 53% (median 14%) (Environment Canada 2006a).

Three sites with a range of surfical geology, soil texture and forest cover were selected for this study: sand-jack pine (SP), sand-aspen (SA) and loam-aspen (LA). In the study region, glacial till deposits range from 20 to 240 m in thickness over Upper Cretaceous Smoky Group shale bedrock (Vogwill

1978). The SP site is located in an area of glacial outwash (Fenton *et al.* 2003) with soils developed from sand and classified as Dystric Brunisols (Soil Classification Working Group 1998). The SA site is also located in an area of glacial outwash (Fenton *et al.* 2003); however, there are bands of finer-textured materials through the soil profile which are classified as Eutric Brunisols (Soil Classification Working Group 1998). The LA site is located on a disintegration moraine (Fenton *et al.* 2003) featuring heterogeneous soils with a dominant loam texture near the surface and which are classified as Gray Luvisolic (Soil Classification Working Group 1998). The overstory at the SP site is comprised of jack pine (*Pinus banksiana*), while the overstory at the SA and LA sites is dominated by trembling aspen (*Populus tremuloides*). There is very little shrubby understory vegetation at the SP site, but the ground is covered with a layer of feathermoss on north-facing slopes and by lichens on south-facing slopes. The SA site has a shrubby understory comprised largely of rose (*Rosa acicularis*), whereas the LA site has a thick understory of rose and high-bush cranberry (*Viburnum edule*).

At each of the three sites (SP, SA and LA), four runoff frames, each with an area of 1 m^2, were constructed at plots on a south-facing (Sf) and north-facing (Nf) slope. In total, eight runoff frames were constructed per site; 24 were established for the full study. Aluminium flashing was installed around all sides of each frame and was inserted 0.05 m into the mineral soil (e.g. Harms & Chanasyk 1998) below the forest floor (FF) horizons. At the down-slope end of the plot, a section of trough was inserted 0.06 m into the soil to collect flow from the top 0.05 m of mineral soil and the overlying forest floor. Water collected in the troughs was routed to a bucket in a pit down-slope from the frame to allow for measurement of runoff volumes. Within each frame, vegetation was clipped to a height of 0.05 m above the ground surface to eliminate differences in understory interception between plots. Given that the runoff collection trough extended under approximately 6% of the frame area and may have collected only vertical infiltration, the detection limit for runoff collection was set at a runoff coefficient (RC) of 6%.

To manipulate the autumn soil moisture content and assess the potential for surface runoff under unfrozen conditions, half of the runoff frames were randomly selected to receive irrigation treatments (e.g. Kane & Stein 1984). On the irrigated frames, 40 mm of water was applied at an intensity of 80 mm hr^{-1} on 21 October 2003 (DOY 294), and included a Br^- tracer for a related experiment. The water was applied using a hand-held watering can. The plots that did not receive 40 mm of water were irrigated with the equivalent of 4 mm of precipitation to apply a Br^- tracer for a related experiment. The plots irrigated with 4 mm are hereafter referred to as non-irrigated.

To measure the amount of snowmelt water available for runoff or infiltration at each study plot, a snowmelt lysimeter was installed at each plot. The lysimeters were located adjacent to runoff frames with similar aspect, slope and canopy cover. The lysimeters were constructed of whitish, opaque plastic pans that had an area of 0.47 m^2 (0.52×0.9 m) and a depth of 0.11 m. The pans were installed so that all water would drain to one corner where a hole was cut, and outflow was routed into a collection bucket buried down slope. Water volumes in the bucket were measured at the same time as runoff frame outflow.

During each site visit, snow depth was measured with a ruler at five locations within each frame and lysimeter. Snow depths were converted to snow water equivalent (SWE) using density values obtained approximately weekly from snow tube measurements adjacent to the frames at each study plot.

The presence of near-surface soil frost was measured periodically within each frame and across the study hillslopes. Frost was measured in the runoff frames using a metal pin (0.002 m diameter) once the snow depth was less than 0.3 m to minimize potential disturbance to the snow surface. The pin was inserted into the soil at five locations per plot. If the pin met resistance and was not able to penetrate into the mineral soil, the measurement was considered to be concrete frost. At the hillslope scale, frost measurements were made along transects across various slope positions using a pointed metal rod 0.008 m diameter, and a subsample of probing locations was visually inspected for ice content (e.g. Young *et al.* 1997).

The method of Proulx & Stein (1997) was used to examine the long-term potential for concrete frost development during winter and spring based on meteorological conditions. The method examines the potential of weather conditions to melt some amount of snow and deliver it to the ground surface where it can re-freeze (Proulx & Stein 1997). These weather conditions consist of mean daily temperature greater than 0 °C to melt snow followed by a period where the maximum daily temperature is below 0 °C to freeze liquid water at the soil surface. For this analysis, daily air temperature data (maximum, minimum and mean) from Red Earth Creek was used for the period 1995–2006 (Environment Canada 2006b).

The soils at all plots were described in the field, and samples were collected for laboratory analysis of bulk density and particle size. Soil bulk density was sampled using 0.06 m diameter and 0.05 m long cores. Mineral soil samples were dried at 105 °C for 24 hrs; forest floor (FF) samples were dried at 70 °C for 48 hours. Porosity was calculated using the measured bulk density data and particle density values of 1.54 Mg m^{-3} for FF and 2.65 Mg m^{-3} for mineral soil (Redding *et al.* 2005). Particle size was analyzed in 63 soil samples per plot using the hydrometer method with carbonate pre-treatment (Kalra & Maynard 1991).

Infiltration rates were measured at the slope crest above all plots in both frozen and unfrozen conditions. Four metal rings of 0.15 m diameter were inserted 0.05 m into the mineral soil, and the forest floor was left intact within the rings. Since rings were installed on relatively flat ground to ensure accuracy of measurement, statistical tests for aspect effects were not possible; results are therefore presented by site. During autumn 2003, four of the eight infiltration rings at each site were measured prior to soil freezing. During spring 2004, all eight rings were measured at each site. Spring measurements were made once the snow had melted from the rings but while the soils were still frozen. All spring measurements were made using water that had been allowed to cool outdoors overnight during subfreezing temperatures so that water temperatures were 1–2 °C at the time of infiltration measurement. Infiltration was measured and calculated following the method of Reynolds & Elrick (2002).

Soil moisture storage in the top 1 m was also monitored manually using the Profile Probe (PR-1, Delta-T Designs, UK). The Profile Probe (similar to time domain reflectometry (TDR)) does not sense ice due to its low dielectric content; measurements were therefore not made during the period of frozen soils. The Profile Probe access tubes were installed and initial measurements were made in autumn 2003 prior to soil freezing. Measurements were also made following soil thaw in June 2004. The use of factory calibrations for the profile probe could lead to water contents greater than the porosity of the soil; a site-specific calibration was therefore developed for the soils at SP, SA and LA sites using the formula:

$$\mathrm{VWC} = (0.00004 \times \mathrm{V}^{1.4364}) \tag{1}$$

where VWC is the volumetric water content ($m^3\ m^{-3}$) and V is the voltage reading from the profile probe. The relationship has $r^2 = 0.62$ and constrains the VWC estimates to realistic values at higher V, which has been shown to be a problem with the factory calibration equation (Evett *et al.* 2006).

To test the effects of aspect and irrigation on snowmelt runoff, a randomized complete block analysis of variance (ANOVA) was used to compare RCs between aspect (Sf and Nf) and autumn irrigation (irrigated and non-irrigated). In this design, the site was used as the block and the ANOVA tested the effects of aspect, autumn irrigation and the interaction between the two treatments. To test for differences between time to steady-state infiltration among the autumn, spring irrigated and spring not-irrigated trials, a one-way ANOVA was used followed by a Tukey multiple comparison test (Zar 1984). All analyses were assessed for significance at the $P = 0.05$ level, and analysis was completed using PopTools (CSIRO, Canberra, Australia).

RESULTS

Site and soil characteristics

The vegetation cover and soil texture characteristics of the study plots are presented in Table 1. The sand-jack pine (SP) site had an overstory of jack pine with a coniferous canopy that persisted through the winter period. Winter canopy cover was greatest at the SP-Nf plot relative to all other plots. At the SA and LA sites, canopy cover increased greatly during the leafed-out growing season. Slope gradients ranged from 14 to 23% across all study plots with the SA site having the steepest slopes and the SP site the gentlest slopes (Table 1).

The clay content of the soils in the LA site was much higher than in the sand-rich SA and SP sites and showed greater variability between sites, especially in the LA-Sf plot (Table 2). Forest floor thickness and composition also

Table 1 | Site characteristics of snowmelt plots

Site[1]	Plot[2]	Canopy species	Slope orientation (°)	Slope gradient (%)	Canopy cover (%)[1]	Unfrozen infiltration rate[2] K_s (mm hr^{-1})
SP	Sf	Jack Pine	200	16	5	787
	Nf	Jack Pine	35	14	35	(226)
SA	Sf	Aspen	206	23	5 (30)	824
	Nf	Aspen	70	20	5 (30)	(411)
LA	Sf	Aspen	160	18	10 (60)	644
	Nf	Aspen	350	20	15 (60)	(656)

[1] Canopy cover in leafless and full-leaf (in parentheses) conditions.
[2] K_s is saturated hydraulic conductivity, data presented are means of Nf and Sf plots ($n = 4$) with standard deviation in parentheses.

Table 2 | Forest floor (FF) and upper mineral soil properties (0.05 m) in runoff frames. Texture data (S: sand % and C: clay %) and initial volumetric water content (VWC) were measured in the upper 0.05 m of the mineral soil. Initial moisture content was measured 8 days after the initial irrigation of plots, immediately prior to soil freezing in November 2003. Relative saturation was calculated by dividing the VWC by the soil

Site	Plot	Frame No.	Irrigation depth (mm)	FF thickness (m)	S (%)	C (%)	Initial VWC (m^3 m^{-3})	Rel. Sat.
SP	Sf	3	40	0.05	92	6	0.05	0.09
		4	40	0.05	93	5	0.09	0.16
		1		0.03	93	5	0.06	0.11
		2		0.05	92	6	0.08	0.14
	Nf	3	40	0.09	93	4	0.07	0.12
		4	40	0.09	93	3	0.05	0.08
		1		0.09	93	5	0.07	0.12
		2		0.12	95	4	0.03	0.05
SA	Sf	1	40	0.06	88	7	0.06	0.11
		3	40	0.06	86	6	0.06	0.11
		2		0.06	88	6	0.05	0.09
		4		0.07	86	8	0.08	0.15
	Nf	1	40	0.07	87	6	0.09	0.17
		2	40	0.06	88	7	0.09	0.17
		3		0.06	88	5	0.14	0.26
		4		0.06	88	5	0.06	0.12
LA	Sf	1	40	0.10	39	22	0.17	0.31
		4	40	0.08	80	6	0.07	0.13
		2		0.10	41	19	0.16	0.30
		3		0.08	62	13	0.06	0.11
	Nf	3	40	0.13	60	17	0.13	0.22
		4	40	0.13	30	23	0.12	0.20
		1		0.13	36	20	0.22	0.37
		2		0.12	54	16	0.26	0.43

varied between sites and plots. The SP-Sf plot had the thinnest forest floor, which was comprised of lichens and needles. It was less than 50% of the thickness of the feathermoss/needle FF on the SP-Nf plot (Table 2). The SA and LA sites both had an aspen/litter FF, but it was slightly thicker on the LA site (Table 2). Steady-state infiltration rates, measured in autumn 2003, were slightly higher on the SP and SA sites than the LA site, which also had a greater spatial variability (Table 1).

To test the influence of soil moisture increases during autumn rains, 40 mm of simulated precipitation was added to half of the frames during late October. A 40 mm event is roughly double the average October rainfall and has a probability of <1%. Although the irrigations represent high-intensity storm events, no surface runoff was measured from any of the frames during the initial 40 mm irrigation. In addition, there was no statistically significant difference ($P = 0.66$) in soil moisture at 0–0.05 m depth in the mineral soil between irrigated and non-irrigated frames 8 days after the irrigation applications and immediately prior to soil freezing.

Over and above the irrigations, there were site differences in the initial soil moisture conditions. There was a statistically significant negative relationship between initial soil moisture content and sand content ($P < 0.001$). Initial soil moisture content was greatest in the LA frames (range 0.06–0.26 $m^3\ m^{-3}$) (Table 2). Initial moisture content at the sand-rich SP and SA sites were similar and generally less than 0.10 $m^3\ m^{-3}$ (Table 2).

Snow accumulation

During the winter, prior to the start of intensive field measurements (1 November 2003 to 14 March 2004, DOY 74), total precipitation recorded by Belfort gauges near the SP and LA sites was 85 and 62 mm, respectively (Table 3). Total precipitation over the same period at the closest Environment Canada stations at Red Earth Creek and Slave Lake was 60 and 62 mm, respectively (Table 3). At the start of the intensive measurements, average SWE on the study sites ranged from 9 mm on the SA-Sf plot to 37 mm on the SA-Nf plot (Table 3).

The differences between north- and south-facing slopes resulted from melt that occurred between 4 February and 3 March 2004, when the lysimeters at the SP-Sf and SA-Sf plots had collected significant meltwater (26 and 36 mm, respectively). At the start of the intensive measurements there was therefore less snow on the Sf than on the Nf plots at the SP and SA sites; however, there were no differences in snow accumulation between aspects at the LA site. During the period of 15 March to 19 April 2004 (DOY 75–110), measured precipitation was 70 mm at the LA site and 86 mm at the SP site (Table 3, Figure 1) compared to 25 mm and 42 mm at Red Earth Creek and Slave Lake, respectively. More than 50% of the annual snowfall came in the last 6 weeks of winter. Small but almost daily snowfalls occurred during the first 2 weeks of the intensive measurement period, and a large snowfall event occurred in mid-April (Figure 1). There was considerable variability in measured precipitation among sites; the ca. 30 mm event recorded on DOY 88 near the SP site deposited only about 8 mm at the LA site (Figure 1).

Table 3 | Precipitation (P) measured in the open, below canopy lysimeter outflow (Lys) and snow accumulation (SWE) measurements from the beginning of the experiment until the end of the snowmelt period (1 November 2003 to 19 April 2004)

Site	Plot	01/09–31/10/03 P (mm)	01/11/03–14/03/04 P (mm)	01/11/03–14/03/04 Lys (mm)	SWE[1] (mm)	15/03–19/04/04 P (mm)	15/03–19/04/04 Lys (mm)	Total: 01/11/03–19/04/04 P (mm)	Total: 01/11/03–19/04/04 Lys (mm)
SP	Sf	42	85	26	26	86	86	171	112
	Nf	42	85	8	36	86	60	171	68
SA	Sf			36	9		57		93
	Nf			4	37		83		87
LA	Sf	32	62	3	33	70	131	132	134
	Nf	32	62	0	32	70	92	132	92
Red Earth Creek		56	60			25		85	
Slave Lake		40	62			42		104	

[1]Average plot SWE measured on 13 or 14 March 2004 (DOY 73/74).

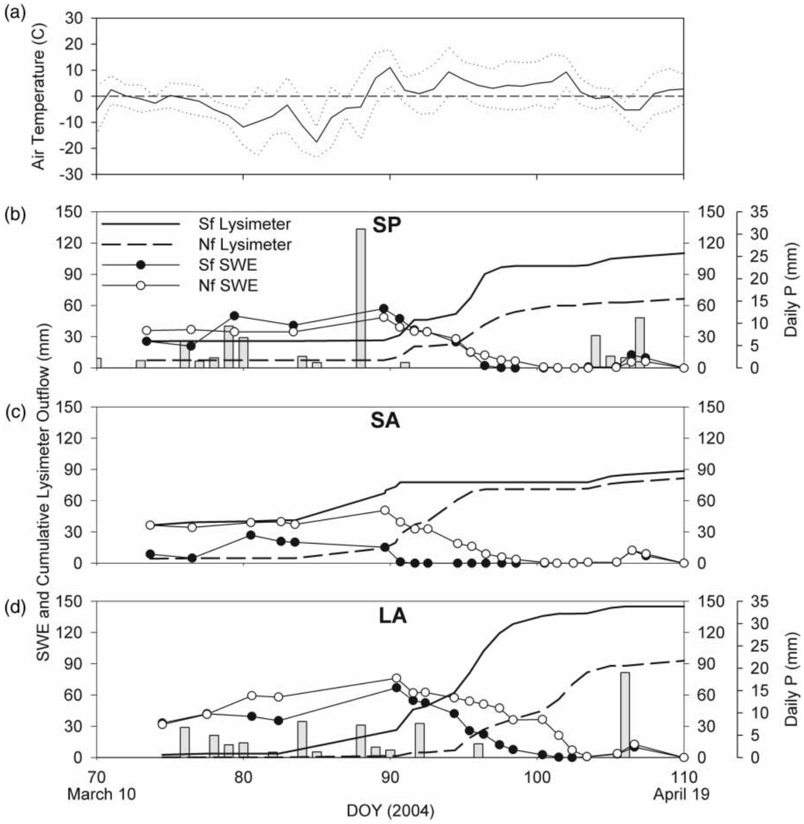

Figure 1 | (a) Time series of daily air temperature (solid line is mean, dashed lines are maximum and minimum); (b) snow water equivalent (SWE); (c) cumulative snowmelt outflow for lysimeters and daily precipitation at the Sand-Pine (SP) and Sand-Aspen (SA); and (d) Loam-Aspen (LA) sites. Precipitation plotted for the SP and LA sites was measured with Belfort gauges, while the intermediate site did not have a Belfort gauge.

Maximum snow accumulation occurred on most plots by DOY90, with the exception of the SP-Nf and SA-Sf plots, where only slight increases in SWE were observed from DOY81 to 90 (Figure 1). Accumulation was much more pronounced at the LA site between DOY82 and 90. The peak accumulation at SA-Sf likely occurred during the large snowfall in mid-April and was missed by the field measurements. Peak accumulation amounts varied considerably with aspect and among sites. In general, they were less than 60 mm at the SP and SA sites and greater than 70 mm at the LA site (Figure 1).

Snowmelt

High variability in snow depth and total melt occurred among sites and aspects. Prior to the start of intensive measurements at the three sites, there was significant melt at the SP and SA Sf plots (Table 3). Melt occurred at some sites from DOY 82 to 90 when the air temperature rose above 0°C for several days which reduced SWE accumulation, especially at the SA-Sf plot (Figure 1). Intense and complete snowmelt occurred after the DOY88 snowstorm, when average air temperature rose well above 0 °C for more than 10 days (Figure 1, Table 4). Snow disappeared earlier on Sf plots than Nf plots at all sites and on the sand outwash sites (SP and SA) relative to the loam site, which was a reflection of low SWE at the start of the study (Tables 3 and 4, Figure 1). During the main melt period, the timing and rate of snowmelt differed among sites and aspects (Table 4). The Sf plots had shorter melt periods than Nf plots and almost double the average and maximum melt rates of Nf plots (Table 4). The maximum daily snowmelt rate ranged from 9.8 to 23.7 mm d^{-1} (Table 4). The maximum hourly snowmelt rate measured was 2.5 mm hr^{-1} over a 2 hour period at the SA-Sf plot on DOY 90. The SP-Sf and -Nf plots and the LA-Sf plot had

Table 4 | Snowmelt characteristics by plot. Melt depths and rates were calculated using lysimeter outflows. SWE (mm) and depth (m) values are from the lysimeters

		Primary melt period						
Site	Plot	Start (DOY)	End (DOY)	Start SWE (mm) and depth (m)	Length (days)	Mean melt rate (mm d^{-1})	Max melt rate (mm d^{-1})	DOY of maximum melt rate
SP	Sf	89	98	49 (0.26)	9	9.5	23.7	95
	Nf	89	101	23 (0.15)	12	5.7	9.8	95
SA[1]	Sf	82	90	19 (0.10)	8	5.1	13.0	90
	Nf	80	96	24 (0.20)	16	4.1	12.0	90
LA	Sf	86	98	106 (0.37)	12	12.5	21.8	95
	Nf	90	105	74 (0.40)	15	5.9	14.5	102

[1] The primary melt event occurred at the SA-Sf plot when staff were not on site to make frequent measurements.

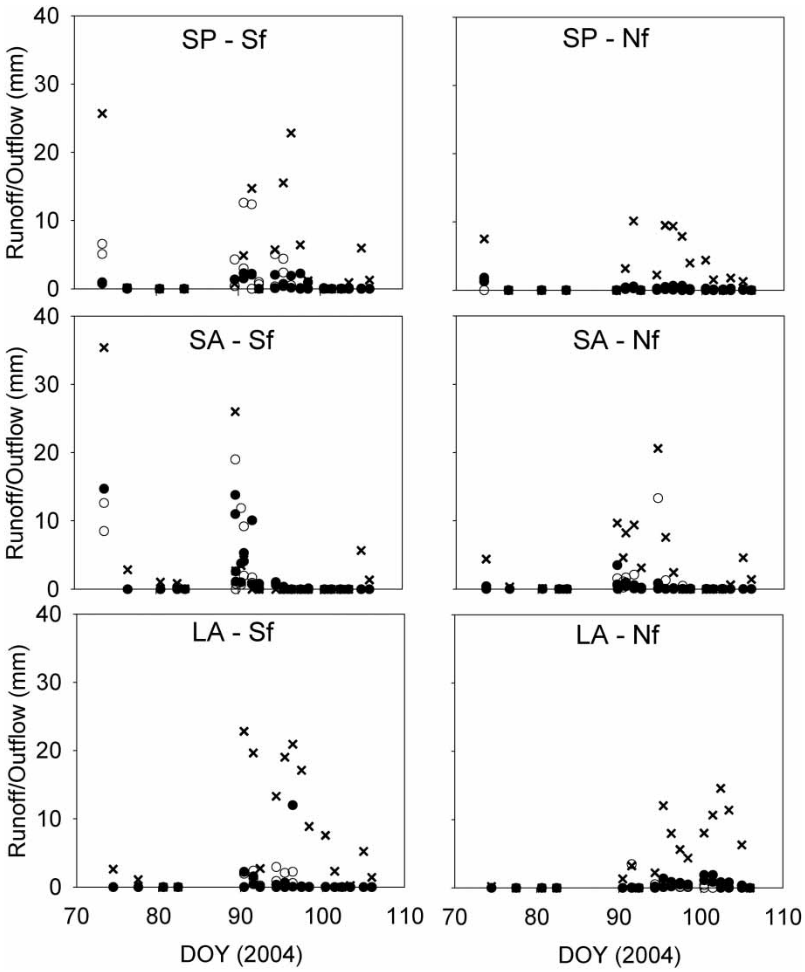

Figure 2 | Time series of outflow from lysimeters (crosses) and runoff frames (circles) by site and plot. Panels on the left side are for south-facing plots and panels on the right side are for north-facing plots. Frames that were irrigated in autumn 2003 are indicated by filled circles; non-irrigated frames are indicated by open circles. Instances where frame runoff was greater than lysimeters outflow corresponds to times when the lysimeters were snow limited relative to the frames, due to spatial variability in snow accumulation and melt.

maximum melt rates on DOY 95, which had a daily maximum temperature of near 20 °C and was preceded by five days with mean temperatures greater than 0 °C (Figure 2).

Snowmelt runoff

Prior to 13 March (DOY 73), runoff was restricted to a few south-facing SP and SA frames where significant snowmelt had been measured (Figure 2). Significant runoff (RC > 6%) was observed only during DOY 90–97, when air temperatures increased and the greatest snowmelt rates occurred. Runoff was largely restricted to SP-Sf and SA-Sf frames, with greatest amounts at SA-Sf frames. Despite high snowmelt rates in LA-Sf plots, runoff was measured only during one day from one frame (Figure 2). Few of the Nf frames had RCs greater than 6%, and thus cannot be confidently interpreted as having runoff.

Cumulative runoff to DOY 110 varied from 1 to 60 mm, corresponding to RCs of 1–65% (Table 5). Analysis of the entire dataset indicates there was no statistically significant difference ($P = 0.59$, $F = 0.32$, $df = 1$) in RCs between irrigated and non-irrigated frames (Figure 3). There was a modest significant difference with greater runoff from Sf than Nf plots ($P = 0.08$, $F = 3.99$, $df = 1$). Increased runoff was also observed from sandy soils compared to loam (Figure 3); however, statistical testing was not possible due to the lack of replicate loam sites.

Table 5 | Frame water balances from autumn 2003 (DOY 305) to the end of snowmelt (DOY 110, 2004) and soil thaw (DOY 165, 2004). RC is runoff coefficient and ΔS is change in soil moisture storage in the top 1 m of the soil. Total inputs include snowmelt, rain and spring 2004 irrigation amounts

Site	Plot	Frame No.	Autumn 2003 irrigation depth (mm)	Snowmelt inputs to DOY 110 (mm)	Snowmelt runoff to DOY 110 (mm)	Snowmelt RC to DOY 110 (%)	Spring 2004 irrigation depth (mm)	Total inputs to DOY 165 (mm)	ΔS to DOY 165 (mm)	Drainage below 1 m to DOY 165 (mm)	Drainage coefficient to DOY 165 (%)
SP	Sf	3	40	112	14	13	0	153	–3	142	93
		4	40	112	8	7	20	173	–1	166	96
		1	4	112	45	40	2	155	0	110	71
		2	4	112	14	13	0	153	2	137	90
	Nf	3	40	68	5	7	0	105	–5	105	100
		4	40	68	4	6	20	125	–4	125	100
		1	4	68	3	4	2	107	–2	106	99
		2	4	68	2	3	2	107	0	105	98
SA	Sf	1	40	93	33	35	0	128	–1	96	75
		3	40	93	38	41	20	148	–1	111	75
		2	4	93	33	35	0	128	14	81	63
		4	4	93	60	65	2	130	1	69	53
	Nf	1	40	87	4	5	0	116	0	112	97
		2	40	87	5	6	20	136	–3	134	99
		3	4	87	2	2	0	116	0	114	98
		4	4	87	22	25	2	118	–2	98	83
LA	Sf	1	40	134	12	9	0	205	13	180	88
		4	40	134	5	4	20	225	16	204	91
		2	4	134	9	7	0	205	39	157	77
		3	4	134	9	7	2	207	23	175	85
	Nf	3	40	92	5	5	20	160	64	91	57
		4	40	92	7	8	0	140	49	84	60
		1	4	92	3	3	2	142	20	119	84
		2	4	92	1	1	0	140	46	93	66

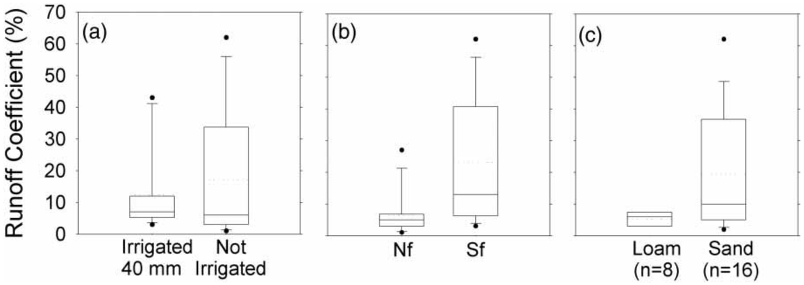

Figure 3 | Boxplots of snowmelt runoff coefficients by site for (a) irrigated versus non-irrigated frames; (b) aspect; and (c) soil texture. Irrigated frames had 40 mm equivalent precipitation added in late October of 2003. For panel (c), the boxplots were developed based on RCs from the 8 frames on loam soils (LA site) and 16 frames on sand soils (SP + SA sites).The boxplot components are: centre of box is median, upper and lower ends of box are the 25 and 75th percentiles, the ends of the whiskers are the 10 and 90th percentiles, dots are outlying points and the dotted line is the mean.

For each vegetation type, greater runoff was observed in the more southern facing plots with large runoff occurring on the south-facing sandy soils. The one exception is the SA plot which has a strong easterly exposure, may receive extended solar radiation relatively early in the day and continues to receive solar radiation inputs into the afternoon due to open canopy conditions. Analyses of individual frames indicate that the highest RCs corresponded to frames with the greatest sand content (Figure 4) which also correspond to low autumn soil moisture content (Figure 4). This was a function of greater runoff being generated from the SP and SA sites, which have sandy soils with low water-holding capacity. However, the sandiest soils generated a wide range of runoff from the highest to the lowest recorded (Figure 3), and there was no discernable pattern except for aspect. The LA site, which featured soils with a greater water-holding capacity, did not generate significant lateral flow (Figure 3).

There was a positive relationship between runoff and the extent of concrete frost within the frames (Figure 4). As the extent of concrete frost at the date of maximum melt rate increased, so did runoff once the area of concrete frost in the plot reached 80% or greater (Figure 4). The extent of concrete frost was not related to autumn moisture content at these study sites. During the study period, surface runoff in response to snowmelt was visually observed only on road surfaces and in ditches at the SA and LA sites.

Concrete frost occurrence

As illustrated earlier, concrete frost development was not directly related to autumn irrigation treatment. The spatial extent of concrete frost within runoff frames at the time of maximum melt rate was associated with the extent of surface runoff during the intense melt period (Table 6, Figure 4). At all three sites, the area of frames occupied by concrete frost was greatest for Sf frames; little to no concrete frost was noted in Nf frames at any of the sites (Table 6). Concrete frost covered 100% of all runoff frames in SA-Sf plots at the start of the intensive snowmelt period, when consistent and high runoff rates were observed. While concrete frost persisted to DOY 111–112 in some frames, runoff did not occur due to the lack of snow. For upland locations, concrete frost formed at the FF mineral soil interface and ranged in thickness from 1 to 4 cm (data not shown).

A transect survey on April 11 (DOY 102) indicated that the proportion of concrete frost occurrence in the runoff frames was similar to that at the midslope position for each study hillslope (plot) (Table 7). This suggests that construction of the runoff frames did not influence the formation of concrete frost.

Frozen soil infiltration

Infiltration rates measured in the frozen soils were orders of magnitude greater than the maximum measured snowmelt rate of 2.5 mm hr^{-1} (Figure 5). At the time of the infiltration measurements in April 2004 the soils were frozen, but there was no concrete frost. The time required to reach steady-state infiltration increased significantly ($P = 0.02$, $F = 9.02$, $df = 2$) from unfrozen conditions (autumn 2003; range 1–5 minutes; median 2 minutes) to frozen soil conditions in spring (range 1–32 minutes; median 9 minutes) (Figure 5).

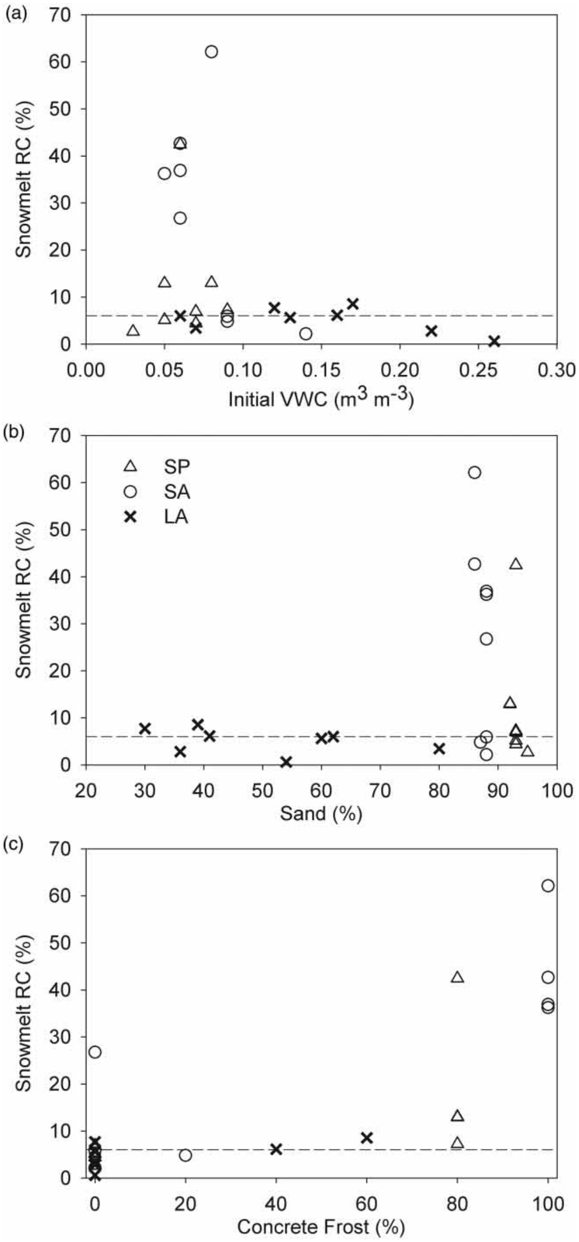

Figure 4 | Relationship of snowmelt runoff coefficient (RC) to (a) initial (autumn) volumetric water content (VWC); (b) sand content; and (c) concrete frost occurrence by site. Concrete frost occurrence is the proportion of sampled locations per frame as measured on (or closest date to) the date of maximum snowmelt. The dashed lines at an RC of 6% in each panel represent the runoff detection limit for the frames.

The time required to reach steady-state infiltration under frozen conditions at sites not previously measured in autumn 2003 ranged from 1 to 33 minutes, with a median of 5 minutes (Figure 5). This was not statistically significantly different ($P > 0.05$) from either the autumn measurements or spring re-measurements. The infiltration measurements were carried out after all the snow had melted, so direct comparison between these measurements and the runoff measurements is not possible. The infiltration measurements demonstrate that under frozen but not concrete frost conditions, these soils can drain water at rates that exceed the maximum measured melt rates.

Snowmelt infiltration and frame water balances

Infiltration was greater than runoff for all frames except one (SA-Sf-4) (Table 5). The drainage coefficients (the proportion of the total snowmelt + rainfall inputs to DOY 165 that were residual after runoff and changes in soil moisture storage) were lowest at the SA-Sf site due to large amounts of snowmelt runoff. In addition, the LA-Sf site (and the LA-Nf site, to a certain extent) had lower drainage coefficients due to storage in the upper metre of the silt-clay-rich soil. Based on the proportion of total snowmelt, Nf plots had greater infiltration than Sf plots; however, based on water depth equivalent, Sf plots on the SP and LA sites had greater infiltration due to greater snowmelt inputs (Table 5). At the SA site, due to the large amounts of runoff from the Sf plot, both the relative proportion and total depth of infiltration was greater for the Nf plot.

Increases in soil moisture storage from autumn 2003 to spring 2004 were greater at the LA site than at the other sites (Table 5). Changes in soil moisture storage reported in Table 5 included some spring rainfall. The very small changes in storage at the SP and SA sites (Table 5) are reasonable given the low water-holding capacities of the coarse-textured soils on those sites. The soil moisture content data collected using the profile probes indicate either considerable (minimum 75 mm) vertical drainage below 1 m or lateral flow in the B or C horizons (Table 5).

DISCUSSION

Controls on snowmelt runoff from forested hillslopes

The dominance of snowmelt infiltration over runoff on hillslopes in the boreal forest is well documented (Price &

Table 6 | Occurrence of concrete frost in runoff frames. Values are the percentage of locations tested ($n = 5$ per frame) that were underlain by concrete frost as determined by probing with a metal pin. Empty cells indicate that no measurements were made at that frame on that date

Site	Plot	Frame No.	Irrigation depth (mm)	April 1 DOY 91	94	95	96	97	100	111	April 22 112	May 6 126
SP	Sf	3	40	60	80	80		100	80	60		0
		4	40	60	80	80		60	80	40		0
		1	4	60	80	80		60	80	20		0
		2	4	60	80	80		100	80	40		0
	Nf	3	40	0	0	0		0	40	20		0
		4	40	0	0	0		0	60	20		0
		1	4	0	0	0		0	0	0		0
		2	4	0	0	0		0	0	0		0
SA	Sf	1	40	100	100	100	100	80	80	60		0
		3	40	100	100	100	100	80	80	60		0
		2	4	100	100	100	100	80	80	60		0
		4	4	100	100	100	100	80	80	60		0
	Nf	1	40	0	0	20	20	20	20	0		0
		2	40	0	0	0	0	20	20	0		0
		3	4	0	0	0	0	20	20	0		0
		4	4	0	0	0	0	20	20	0		0
LA	Sf	1	40		0		40	60	20		50	0
		4	40		0		20	0	0		50	0
		2	4		0		40	40	20		50	0
		3	4		0		20	0	0		50	0
	Nf	3	40		0		0	0	0		40	0
		4	40		0		0	0	0		40	0
		1	4		0		0	0	0		40	0
		2	4		0		0	0	0		40	0

Hendrie 1983; Kachanoski & de Jong 1982; Kane & Stein 1984; Whitson *et al.* 2004; Devito *et al.* 2005b). At a site approximately 200 km southeast of this study site, with similar soils to the LA site, Whitson *et al.* (2004) measured RCs of less than 1% on a clearcut harvested hillslope with an easterly aspect. The RCs in this study are similar to those of Kane & Stein (1984) who found RCs from snow plots in central Alaska ranged from 0 to 54% for a south-facing aspen-birch forested hillslope with silt loam soils. Similarly, Dunne & Black (1971) recorded hillslope scale surface RCs of 32–47% on agricultural soils in the north-eastern US.

Although runoff amounts were low in this study, this research (along with that of others e.g. Kachanoski & de Jong 1982; Kane & Stein 1984) clearly illustrates that runoff can occur in the boreal forest. There are some landform-vegetation combinations that show potential for snowmelt runoff; however, the controls in the western Boreal Plains (at least for 2003–2004) were not necessarily what was expected. The presence of concrete frost played a critical role in the generation of near-surface runoff from the study plots by providing a nearly impermeable surface over which runoff could occur (Bayard *et al.* 2005). The largest amounts of concrete frost, highest melt rates and corresponding RCs were measured on the Sf slopes at the sand-rich SP and SA sites.

The influence of autumn soil moisture content on snowmelt runoff was minimal in this study. Previous research indicated that if the relative saturation of the surface soils is

Table 7 | Concrete frost occurrence by topographic position and wetland type at the study hillslopes on 11 April 2004 (DOY 102). Values are the percentage (%) of surveyed locations at each topographic position with concrete frost, as determined by penetration by metal pole (n = 7–10 locations per topographic position). The values in parentheses are the plot mean concrete frost coverage for the runoff frames at the same slope positions on DOY 100 (calculated from data in Table 6)

Site	Aspect	Topographic position	% Locations with concrete frost
SP		Crest	70
	Sf	Mid-slope	60 (80)
		Toe-slope	85
		Sedge fen	100
	Nf	Mid-slope	60 (25)
		Toe-slope	60
		Peatland	100
SA		Crest	100
	Sf	Mid-slope	85 (80)
		Open fen	100
	Nf	Mid-slope	30 (20)
		Toe-slope	60
LA		Crest	30
	Sf	Mid-slope	0 (10)
		Toe-slope	40
		Peatland	100
	Nf	Mid-slope	0 (0)
		Toe-slope	15
		Peatland	100

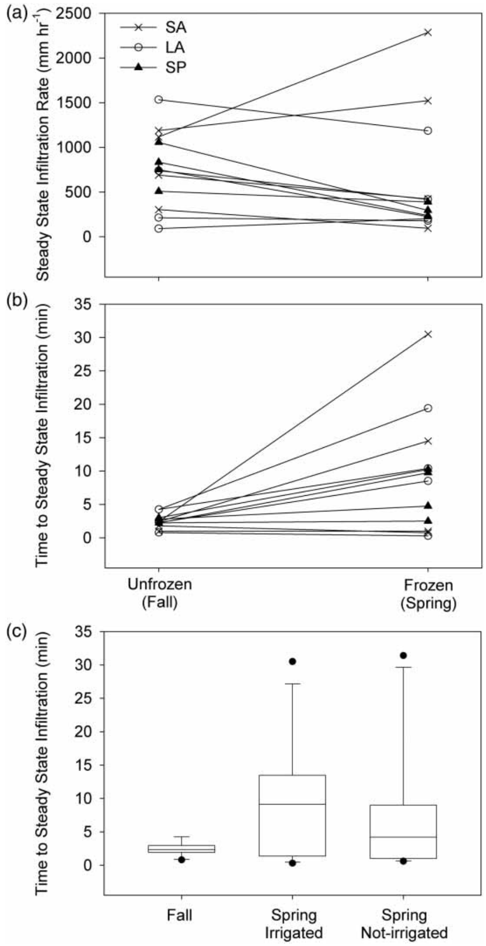

Figure 5 | (a) Frozen soil infiltration rates; (b) infiltration time to steady state; and (c) boxplots comparing infiltration rates between autumn (measured 5–7 October 2003), spring previously irrigated (measured 9–22 April 2004) and spring not-irrigated (measured 9–22 April 2004). For each boxplot $n = 12$. The lines between the points in panels (a) and (b) indicate the change in infiltration rate or time to steady state between autumn and spring measurements for the same locations.

greater than 0.3, concrete frost development and spring snowmelt runoff will be enhanced (Granger *et al.* 1984; Pomeroy *et al.* 1997). In contrast, Kane & Stein (1984) found no clear relationship between autumn irrigation amounts and spring runoff. They found that spring surface runoff increased with increasing total snow accumulation and increasing melt rate. In general, the initial soil moisture conditions in the western boreal forest are often below the threshold observed in agricultural settings on the Canadian Prairies (Granger *et al.* 1984; Pomeroy *et al.* 1997) due to moisture uptake by forest vegetation and soil drainage during drier autumn periods (Devito *et al.* 2005b). Further, concrete frost was observed only in sandier soils with relative saturation (measured in the autumn) that was half that of the threshold value hypothesized for runoff generation on prairie soils (Zhao & Gray 1999). In this study, the development of concrete frost on the silt-clay-rich moraine soils with high autumn soil moisture content was less extensive than on the drier soils at the SP and SA sites. This indicates that soil frost development occurred during the winter rather than as a result of autumn moisture content, and runoff is controlled by snowmelt processes.

In this study, runoff was not positively correlated with the fine grain size fraction as expected. The low water-holding capacity of the coarse-textured soils was expected to hinder the development of concrete frost, and hence runoff. The results presented herein on the influence of soil texture may apply only to regions or years with dry autumn conditions leading to low autumn soil moisture content.

The interaction of aspect with vegetation cover controls inputs of solar radiation, and proved to be the strongest influence on the generation of near-surface runoff in this study. Due to greater incoming solar radiation and lower canopy cover (which results in reduced interception and shading of the snowpack surface), SWE and melt rates on south-facing slopes are often greater than on north-facing slopes, which results in greater snowmelt runoff (Harms & Chanasyk 1998; Carey & Woo 1999; Bonan 2002; Murray & Buttle 2003; Woo 2005). The maximum melt rate measured in this study was 2.5 mm hr^{-1}, much lower than the measured infiltration rates. However, because runoff was collected from the frames, the melt rate must have exceeded the infiltration rate where concrete frost occurred. In this study, higher melt rates may have been required to generate surface runoff. However, infiltration rates exceeded snowmelt rates in the LA plots, indicating that both concrete frost development and high rates of snowmelt may be required to generate surface runoff. These results contradict those of Harms & Chanasyk (1998), who measured greater surface runoff from north-facing than south-facing slopes associated with reclaimed mine soils in central Alberta. There was very high spatial variability among runoff frames within that site, however. Carey & Woo (1999) found that south-facing slopes with loam soils and aspen vegetation did not contribute any runoff, even when soils were frozen, because they were sufficiently permeable to allow infiltration of all snowmelt. Significant runoff occurred only from north-facing slopes with permafrost forming in the thick organic soil layers (Carey & Woo 1999). These studies indicate that concrete frost development and high melt rates on south-facing slopes may be driving surface runoff processes in the present study. Vegetation density, which seems to co-vary with soil texture at these sites, may control the occurrence of concrete frost on these sites.

The Sf plots at the SP and SA sites had the lowest canopy cover, which has been shown to correspond to increased development of concrete frost (Pierce *et al.* 1958). Similar to this study, Kachanoski & de Jong (1982) measured surface runoff from clearcut harvested slopes but not from adjacent forested areas, possibly due to reduced shading and increased energy inputs on the clearcut sites. It has also been hypothesized that concrete frost development in forest soils is limited by the presence of the forest floor layer (Hardy *et al.* 2001). In this study, sites with thinner forest floor layers (SP-Sf and SA) had a greater spatial extent of concrete frost than those with thicker forest floor layers (SP-Nf, LA).

The mechanism of concrete frost formation on the SP and SA sites appears to be as follows: warm temperatures for 1 or 2 days cause melting of snow, which infiltrates into the frozen upper soil and refreezes as temperatures decrease. Typically, this occurs either during mid-winter or at the beginning of the melt period (Mace 1968; Stadler *et al.* 1996; Proulx & Stein 1997; Bayard *et al.* 2005). However, the probability of occurrence increases with increased radiation inputs at sites with southern exposures and low canopy cover.

The characteristics of concrete frost development make it difficult to apply snowmelt infiltration/runoff equations that rely on autumn moisture content (e.g. Granger *et al.* 1984) due to the multiple melt and freeze events that alter soil moisture content during winter (Harms & Chanasyk 1998). The analysis of daily climate data using the method of Proulx & Stein (1997) indicated that in the study area, the meteorological conditions necessary for concrete frost development are common and occurred in 10 out of 12 years. The mean number of warming and cooling events per year was 1, with a range of 0–3 events per year. The requisite meteorological conditions were recorded three times in 1995, twice in 2005, once in each of 1996, 1997, 2000–2004 and 2006, but not in 1998 and 1999. Most climate change scenarios predict reduced snow depths and warmer winter temperatures (Barnett *et al.* 2005), which may result in greater potential for concrete frost formation during mid-winter and early melt periods (Hardy *et al.* 2001). Whether the effects of climate change will result in increased hillslope runoff reaching streams is difficult to predict, given the poor hydrologic connectivity between uplands and surface water bodies on the Boreal Plains (Devito *et al.* 2005b).

The fact that runoff was greater on sandy soils than loamy soils in the study area is initially counterintuitive; however, in light of the effect slope aspect has on runoff, it is perhaps not surprising. The following conceptual model explains why greater runoff occurred on south-facing slopes with sandy soils.

1. South-facing sandy slopes have the lowest soil water-holding capacity and greatest solar radiation inputs; therefore, they cannot support dense vegetation cover (Grier & Running 1977) resulting in a more open canopy.
2. The open canopy and south aspect allows greater incoming radiation to melt the snow earlier and faster, increasing the potential for late-winter melt and development of concrete frost.
3. The presence of concrete frost combined with the higher melt rates increases the potential for surface runoff from south-facing slopes during peak snow melt.

This conceptual model combines aspect and soil texture to explain the observed runoff data and accounts for the lower runoff from the LA-Sf plot, even though it had wetter soils and the greatest total snowmelt. The low runoff was due to the limited extent of concrete frost (less than 80% frame area at the time of maximum snowmelt rate, Figure 4), which allowed infiltration at the time of maximum melt rate. These differences are related to higher canopy cover on the LA-Sf plot than on the SP and SA-Sf plots, and may also be due to site topography. The late April snowmelt did not result in any runoff, likely due to the amount of concrete frost being below the 80% threshold.

Snowmelt lysimeter water balance

The snowmelt lysimeter data were crucial to determining runoff coefficients and net infiltration at the soil surface. The lysimeter outflows corresponded relatively closely to the measurements of accumulated precipitation at the SP and LA sites, which were located near existing winter precipitation measurement stations (Table 3). The lysimeter outflows were less than the accumulated precipitation, except at the LA-Sf plot, which may indicate there was some leakage of water into the LA-Sf collection system. At the other plots, the reduced outflows relative to accumulated precipitation are likely a reflection of spatial variability in accumulation due to canopy interception (especially for the SP-Nf plot) and sublimation from the ground surface. Arain *et al.* (2003) measured winter evaporation rates (above canopy eddy covariance measurements) of 0.1–0.25 mm d^{-1} in a boreal black spruce stand. Pomeroy *et al.* (1998) found sublimation of canopy-intercepted snow was approximately 13% of annual snowfall in a mixed aspen (70%) and spruce (30%) stand in Saskatchewan. Ground surface sublimation was not measured in this study, and values for the Boreal Plains are not reported in the scientific literature. However, Molotch *et al.* (2007) reported approximately 37% sublimation from the ground surface snowpack in sub-alpine conifer forests over the winter. These estimates of sublimation losses indicate that the lysimeter outflow values are reasonable when compared to the total measured precipitation from the SP and LA Belfort gauge records. A greater understanding of sublimation processes under different stand types and canopy conditions is necessary to improve the understanding of winter water balances for the Boreal Plains.

At all plots, the total outflow from the lysimeters was greater than the maximum measured SWE. Thus, without the lysimeters, the RCs would have been overestimated due to the thin snowpack and intermittent periods of melting and accumulation during the late winter and early spring. During the study period, there were three melt periods and two accumulation periods. If manual snow surveys for estimating SWE are not frequent enough (e.g. daily), the data do not capture the snowmelt dynamics. For locations with deeper snowpacks, regular snow surveys (e.g. weekly or bi-weekly) may be adequate for capturing the melt dynamics; however, lysimeter data are valuable for developing and testing snowmelt models (Winkler 2001). The use of simple snowmelt lysimeters can improve estimates of water balance components that are critical to understanding and modelling hydrological processes in this landscape.

Hillslope and landscape scale implications

At the hillslope scale, the connectivity of areas underlain by concrete frost will determine whether runoff water reaches receiving surface waters (Woo 2005). Given the patchy spatial distribution of concrete frost in uplands, it appears that only peatlands or ephemeral draws have consistent and extensive concrete frost formation, and therefore regularly act as an impervious surface over which runoff may occur. This agrees with hydrometric (Kalef 2002; Devito *et al.* 2005b) and isotopic studies (McEachern *et al.* 2006) of runoff on the Boreal Plains, which have shown that streamflow originates mainly from wetland areas and only infrequently from uplands. The patchy distribution of concrete frost on hillslopes results in only periodic high runoff responses; in most years, vertical flow will dominate (Price & Hendrie 1983).

Annual peak streamflows for Red Earth Creek (Figure 6) show that the timing of peak flows are not driven by snowmelt runoff, as compared to many montane catchments (e.g. Moore & Scott 2005). The distribution of annual peak flows indicates that over the period of record, half occur before DOY 175 (June 24). Given that snow at the URSA has typically melted by early May, only two of the measured annual peak flows can likely be directly attributed to snowmelt (Figure 6). Peak flows that occur prior to June have a mean flow of 0.63 mm d^{-1}, while those occurring after have a mean flow of 1.55 mm d^{-1}. The larger peak flows occurring later in the year therefore indicate that rain events and antecedent storage drive peak flow generation in this landscape (Devito *et al.* 2005b). As a result, snowmelt on the hillslopes will only infrequently influence regional runoff responses during the snowmelt period even though thaw and re-freezing events frequently occur to generate concrete soil frost (typically discontinuous) on uplands (Devito *et al.* 2005b, McEachern *et al.* 2006). The primary role of snowmelt in the generation of runoff at the watershed scale is to satisfy storage prior to spring or summer rain events (Devito *et al.* 2005b). In 1997, there was a large streamflow peak in the early spring, and an even larger peak (largest daily flows on record) later in the growing season. It is also important to note that rain-on-snow events are uncommon in this region, so they are not expected to be a major driver of runoff at the hillslope or watershed scale as they are in other regions (e.g. Harr 1981; Marks *et al.* 2001). However, the occurrence of rain-on-snow events may increase as the climate warms (Barnett *et al.* 2005).

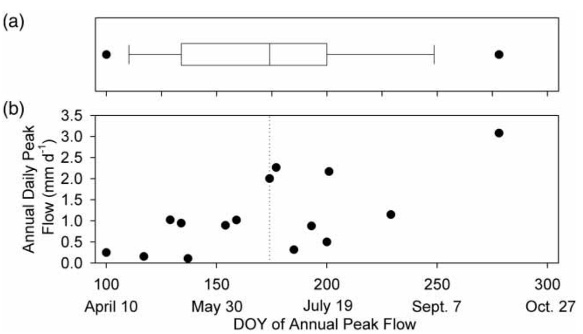

Figure 6 | Annual daily peak stream flow for REDEARTH CREEK NEAR RED EARTH CREEK (07JC002), Alberta, 1987–2001: (a) temporal distribution of peak flow date and (b) the relationship between peak flow date and peak flow magnitude. The dotted line in panel (b) indicates the median date of the annual peak flows for the available record.

Removal of the forest canopy by forest harvesting or oil and gas development can lead to increased snowmelt runoff at the hillslope and watershed scales. When the forest canopy is removed, snow accumulation may increase due to decreased interception, and melt rates increase due to greater incoming solar radiation (Winkler *et al.* 2005). Similarly, the potential for concrete frost formation increases with increased solar radiation. In addition, increased surface runoff can occur where soils have been compacted by machinery (e.g. on skid trails, roads) due to the generation of concrete frost and the presence of highly compacted non-conductive materials. Surface runoff from roads can potentially impact water quality of downstream surface water bodies.

CONCLUSIONS

The results of this research indicate that infiltration dominates near-surface runoff over a range of hillslopes with varying soil types and vegetation covers during the snowmelt period on the Boreal Plains. Slope aspect plays a crucial role in local runoff generation through its relationship with concrete frost development. Runoff frames on south-facing hillslopes generated more runoff than frames on north-facing slopes due to more extensive concrete frost development near the soil surface. Concrete frost development is postulated to be a result of low canopy cover reducing snowfall interception and increasing solar radiation to the ground surface, which results in greater thaw and refreeze cycles. Runoff at the hillslope scale is controlled by the connectivity of concrete frost. Where much of the hillslope is connected to receiving surface waters or wetlands, snowmelt runoff at the watershed scale may be enhanced. Infiltration rates under frozen soil conditions were greater than the maximum melt rates over the portions of the hillslope in this study that lacked concrete frost, indicating a dominance of vertical flow and little opportunity for runoff generation.

Future research needs to examine the potential for snowmelt-driven subsurface flow from similar hillslopes, as this has been noted in previous studies on the Boreal Plains (Kachanoski & de Jong 1982) and other areas that have similar soil properties as the LA site (Newman *et al.* 2004). Climate change may have a large effect on near-surface runoff from hillslopes, as predicted increases in temperatures and decreases in snowpack thickness (Barnett *et al.* 2005) will result in conditions that are amenable to concrete frost formation and more rapid melt rates or rain-on-snow events. Resource development that reduces forest canopy cover may also result in greater potential for near-surface runoff due to increased snow accumulation, faster melt and increased concrete frost development, and to an increase in compacted road surfaces.

ACKNOWLEDGEMENTS

This work was supported by an NSERC-CRD Grant (HEAD2) and Industry Partners (Alberta Pacific Forest Industries, Forest Products Association of Canada, Ducks Unlimited Canada, TOLKO Forest Industries, Syncrude Canada Ltd., Suncor, Canadian Natural Resources Ltd., PetroCanada, Albian Sands, Total EP Canada, Imperial Oil) to KJD (PI), equipment funding from Syncrude Canada Ltd to KJD and TER, a field work grant from C-BAR to TER and scholarships from NSERC and Alberta Ingenuity Fund to TER. We thank Joanne Livingstone, Jeremy Hrynkiw, Brian Smerdon, Wayne Bell and Assar Grinde for field assistance. Field accommodation was provided by Carolyn Forsyth at the Artis Inn.

REFERENCES

Arain, M. A., Black, T. A., Barr, A. G., Griffis, T. J., Morgenstern, K. & Nesic, Z. 2003 Year-round observations of the energy and water vapour fluxes above a boreal black spruce forest. *Hydrol. Process.* **17**, 3581–3600.

Barnett, T. P., Adam, J. C. & Lettenmaier, D. P. 2005 Potential impacts of a warming climate on water availability in snow-dominated regions. *Nature* **438**, 303–309.

Bayard, D., Stahli, M., Parriaux, A. & Fluhler, H. 2005 The influence of seasonally frozen soil on the snowmelt runoff at two alpine sites in southern Switzerland. *J. Hydrol.* **309**, 66–84.

Bonan, G. B. 2002. *Ecological Climatology*. Cambridge University Press, Cambridge, UK.

Carey, S. K. & Woo, M. K. 1999 Hydrology of two slopes in subarctic Yukon, Canada. *Hydrol. Process.* **13**, 2549–2562.

Cline, R. G., Haupt, H. F. & Campbell, G. S. 1977 *Potential Water Yield Response Following Clearcut Harvesting on North and South Slopes In Northern Idaho.* USDA Forest Service, Intermontane Research Station, Research Paper INT-191.

Devito, K. J., Creed, I. F. & Fraser, C. J. D. 2005a Controls on runoff from a partially harvested aspen-forested headwater catchment, Boreal Plain, Canada. *Hydrol. Process.* **19**, 3–25.

Devito, K., Creed, I., Gan, T., Mendoza, C., Petrone, R., Silins, U. & Smerdon, B. 2005b A framework for broad-scale classification of hydrologic response units on the Boreal Plain: is topography the last thing you should consider? *Hydrol. Process.* **19**, 1705–1714.

Dunne, T. & Black, R.D. 1971 Runoff processes during snowmelt. *Water Resour. Res.* **7**, 1160–1172.

Ecoregions Working Group. 1989 *Ecoclimatic regions of Canada*. Government of Canada, Ottawa. Ecological Land Classification Series No. 23.

Elliot, J. A., Toth, B. M., Granger, R. J. & Pomeroy, J. W. 1998 Soil moisture storage in mature and replanted sub-humid boreal forest stands. *Can. J. Soil Sci.* **78**, 17–27.

Environment Canada. 2004 *Canadian Climate Normals 1971–2000: Slave Lake, Alberta*. http://climate.weatheroffice.ec.gc.ca/climate_normals/index_e.html (accessed November 8, 2009).

Environment Canada. 2006a *Archived Hydrometric Data: REDEARTH CREEK NEAR RED EARTH CREEK (07JC002)*. http://www.wsc.ec.gc.ca/hydat/H2O/index_e.cfm?cname=main_e.cfm (accessed April 15, 2010).

Environment Canada. 2006b *Canadian Climate Data Online: Red Earth, Alberta*. http://climate.weatheroffice.ec.gc.ca/climateData/canada_e.html (accessed November 8, 2009).

Evett, S. R., Tolk, J. A. & Howell, T. A. 2006 Sensors for soil profile moisture measurement: Accuracy, axial response, calibration, precision and temperature dependence. *Vadose Zone J.* **5**, 894–907.

Fenton, M. M., Paulen, R. C. & Pawlowicz, J. G. 2003 *Surficial geology of Lubicon Lake area, Alberta (NTS 84B/SW).* Alberta Geological Survey, Edmonton.

Granger, R. J., Gray, D. M. & Dyck, G. E. 1984 Snowmelt infiltration to frozen prairie soils. *Can. J. Earth Sci.* **21**, 669–677.

Grier, C. G. & Running, S. W. 1977 Leaf area of mature northwestern coniferous forests: relation to site water balance. *Ecology*, **58**, 893–899.

Hardy, J. P., Groffman, P. M., Fitzhugh, R. D., Henry, K. S., Welman, A. T., Demers, J. D., Fahey, T. J., Driscoll, C. T., Tierney, G. L. & Nolan, S. 2001 Snow depth manipulation and its influence on soil frost and water dynamics in a northern hardwood forest. *Biogeochem.* **56**, 151–174.

Harms, T. E. & Chanasyk, D. S. 1998 Variability of snowmelt runoff and soil moisture recharge. *Nord. Hydrol.* **29**, 179–198.

Harr, R. D. 1981 Some characteristics and consequences of snowmelt during rainfall in western Oregon. *J. Hydrol.* **53**, 277–304.

Hart, G. E. & Lomas, D. A. 1979 Effects of clearcutting on soil water depletion in an Engelmann spruce stand. *Water Resour. Res.* **15**, 1598–1602.

Kachanoski, R. G. & de Jong, E. 1982 Comparison of the soil water cycle in clear-cut and forested sites. *J. Env. Qual.* **11**, 545–549.

Kalef, N. 2002 *Interlinking Hydrological Behaviour and Inorganic Nitrogen Cycling a Forested Boreal Wetland.* M.Sc. Thesis, University of Alberta, Edmonton, Alberta, 83pp

Kalra, Y. P. & Maynard, D. G. 1991 *Methods Manual for Forest Soil and Plant Analysis.* Forestry Canada, Northern Forestry Centre. Information Report NOR-X-319, Edmonton, AB.

Kane, D. L. & Stein, J. 1984 Plot measurements of snowmelt runoff for varying soil conditions. *Geophysica* **20**, 123–135.

Mace, A.C. 1968 *Effects of Soil Freezing on Water Yields.* USDA Forest Service Rocky Mountain Research Station, Research Note RM-121.

Marks, D., Link, T., Winstral, A. & Garen, D. 2001 Simulating snowmelt processes during rain-on-snow over a semi-arid mountain basin. *Ann. Glaciol.* **32**, 195–202.

Marshall, I. B., Schut, P. & Ballard, M. (compilers). 1999 *A National Ecological Framework for Canada: Attribute Data*. Environmental Quality Branch, Agriculture and Agri-Food Canada, Ottawa/Hull. http://sis.agr.gc.ca/cansis/nsdb/ecostrat/data_files.html (November 8, 2009).

McEachern, P., Prepas, E. E. & Chanasyk, D. S. 2006 Landscape control of water chemistry in northern boreal streams of Alberta. *J. Hydrol.* **323**, 303–324.

Molotch, N. P., Blanken, P. D., Williams, M. W., Turnipseed, A. W., Monson, R. K. & Margulis, S. A. 2007 Estimating sublimation of intercepted and sub-canopy snow using eddy covariance systems. *Hydrol. Process.* **21**, 1567–1575.

Moore, R. D. & Scott, D. F. 2005 Camp Creek revisited: Streamflow changes following salvage harvesting in a medium-sized, snowmelt dominated catchment. *Can. Water Resour. J.* **30**, 331–344.

Murray, C. D. & Buttle, J. M. 2003 Impacts of clearcut harvesting on snow accumulation and melt in a northern hardwood forest. *J. Hydrol.* **271**, 197–212.

Newman, B. D., Wilcox, B. P. & Graham, R. C. 2004 Snowmelt driven macropore flow and soil saturation in a semarid forest. *Hydrol. Process.* **18**, 1035–1042.

Pierce, R. S., Lull, H. W. & Storey, H. C. 1958 Influence of land use and forest condition on soil freezing and snow depth. *For. Sci.* **4**, 247–263.

Pomeroy, J. W., Granger, R. J., Pietroniro, A., Elliott, J.E., Toth, B. & Hedstrom, N. 1997 *Hydrological Pathways in the Prince Albert Model Forest.* The Prince Albert Model Forest Association.

Pomeroy, J. W., Parviainen, J., Hedstrom, N. & Gray, D. M. 1998 Coupled modeling of forest snow interception and sublimation. *Hydrol. Process.* **12**, 2317–2337.

Price, A. G. & Hendrie, L. K. 1983 Water motion in a deciduous forest during snowmelt. *J. Hydrol.* **64**, 339–356.

Proulx, S. & Stein, J. 1997 Classification of meteorological conditions to assess the potential for concrete frost formation in boreal forest floors. *Can. J. For. Res.* **27**, 953–958.

Redding, T. E., Hannam, K. D., Quideau, S. A. & Devito, K. J. 2005 Particle density of aspen, spruce and pine forest floors in Alberta, Canada. *Soil Sci. Soc. Am. J.* **69**, 1503–1506.

Reynolds, W. D. & Elrick, D. E. 2002 Sing-ring and double-or concentric-ring infiltrometers. In: *Methods of Soil Analysis: Part 4, Physical Methods* (ed. J. H. Dane & G. C. Topp), Soil Science Society of America, Madison, WI, pp. 821–826.

Smerdon, B. C. 2006 *The Influence of Climate on Water Cycling and Lake-groundwater Interaction In An Outwash Landscape on the Boreal Plains of Canada*. PhD Thesis, University of Alberta, Edmonton, AB.

Soil Classification Working Group. 1998 *The Canadian System of Soil Classification (Third Edition)*. Ottawa, ON, Agriculture and Agri-Food Canada.

Stadler, D., Wunderli, H., Auckenthaler, A., Fluhler, H. & Brundle, M. 1996 Measurement of frost-induced snowmelt runoff in a forest soil. *Hydrol. Process.* **10**, 1293–1304.

Stein, J., Proulx, S. & Levesque, D. 1994 Forest floor frost dynamics during spring snowmelt in a boreal forested basin. *Water Resour. Res.* **30**, 995–1007.

Vogwill, R. 1978 *Hydrogeology of the Lesser Slave Lake area, Alberta*. Alberta Research Council, Edmonton.

Whitson, I. R., Chanasyk, D. S. & Prepas, E. E. 2004 Patterns of water movement on a logged Gray Luvisolic hillslope during the snowmelt period. *Can. J. Soil Sci.* **84**, 71–82.

Winkler, R. D. 2001 *The Effects of Forest Structure on Snow Accumulation and Melt in South-central British Columbia.* PhD Thesis, University of British Columbia, Vancouver, BC.

Winkler, R. D., Spittlehouse, D. L. & Golding, D. L. 2005 Measured differences in snow accumulation and melt among clearcut, juvenile, and mature forests in southern British Columbia. *Hydrol. Process.* **19**, 51–62.

Woo, M. K. 2005 *Snowmelt runoff generation (Ch. 114)*. In: *Encyclopedia of Hydrological Sciences* (ed. M. G. Anderson), John Wiley and Sons Ltd, Chichester, UK, pp. 1741–1749.

Young, K. L., Woo, M. K. & Edlund, S. A. 1997 Influence of local topography, soils and vegetation on microclimate and hydrology at a high Arctic site, Ellesemere Island, Canada. *Arctic Alpine Res.* **29**, 270–284.

Zar, J. H. 1984 *Biostatistical Analysis.* 2nd ed. Prentice-Hall, Inc., Englewood Cliffs, NJ.

Zhao, L. & Gray, D. M. 1999 Estimating snowmelt infiltration into frozen soils. *Hydrol. Process.* **13**, 1827–1842.

First received 4 December 2009; accepted in revised form 5 July 2010