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The role of catchment scale and landscape characteristics for runoff generation of boreal streams

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Summary The effect of catchment scale and the influence of landscape characteristics on runoff generation were investigated during snow melt in 15 nested boreal streams within the Krycklan catchment in northern Sweden. We used detailed oxygen-18 analyses of soils from two characteristic landscape types, snow melt samples and water samples from 15 streams with subcatchments ranging in size from 0.03 to 67 km^2. The detailed process understanding that was derived from isotopic and hydrometric measurements at a wetland and a forest site, in combination with the stream monitoring, enabled the development of a conceptual framework that could explain the variability in hydrological pathways over a range of catchment scales. While the proportion of new or event water was over 50% in wetland dominated catchments, the event water contribution in forested catchments was between 10% and 30%. The results suggest a large degree of scale-independence of hydrological flow pathways during the snow melt period, controlled by the proportion of wetland and median subcatchment area, across three orders of magnitude in spatial scale. The results from this study highlighted the importance of different runoff generation processes in different landscape elements, an understanding that can be useful in disentangling the temporal dynamics in hydrology and biogeochemistry during snow melt episodes when moving from small headwater streams to catchment outlets.

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Introduction

The effect of catchment scale and the influence of landscape characteristics on runoff generation are still not fully understood because of a complex multi-scale dynamics with numerous processes operating concurrently (Blöschl, 2001; Gergel et al., 1999; McGlynn et al., 2004). Evidence from hydrometric as well as isotopic and chemical tracer studies has been used to infer how the partitioning of event and pre-event water during episodes (Brown et al., 1999; Shanley et al., 2002) and mean transit time (McGuire et al., 2005; Shaman et al., 2004) are affected by catchment scale. Although some evidence of self-similarity of hydrological pathways and transit times across scales has been presented, the construction of hydrological models and river management tools that operate at different spatial and temporal scales remains a challenging task.

One of the most widely recognized methodologies for understanding and quantifying hydrological pathways in catchments is the use of natural stable isotopes as environmental tracers for isotopic hydrograph separation (IHS). The use of IHS has provided important understanding for questions related to water resource management, transport of contaminants and biogeochemical cycling. Its use dates back to the pioneering work by Dinçer et al. (1970) and has since been used in various environments around the world (Laudon and Slaymaker, 1997; Rodgers et al., 2005; Sklash et al., 1986; Stadnyk et al., 2005; Uhlenbrook et al., 2002). The general finding from these studies has been that so-called old, or pre-event, water dominates the hydrograph during events, whereas contribution from new, or event, water during rain storms and snow melt events remains generally small.

As valuable as IHS is, it only provides an answer to the question of the relative contribution of two sources, event and pre-event water, leaving many questions about the specific flow pathways and runoff mechanisms unanswered. Most previous IHS work has been based on isotopic information of inputs and outputs (i.e., precipitation and stream water) only, relying on assumptions about what occurs in the catchment soils. Using internal isotopic information and by combining isotopic and hydrometric information, new process understanding can be acquired that can help decipher the dominant runoff generation mechanisms during hydrological episodes in ways that are otherwise not possible (Burns, 2002).

As with most other process-oriented hydrological investigations, previous IHS studies are mainly based on small individual catchments or hillslopes. The few multi-scale IHS studies that have been conducted provide inconclusive results (Buttle, 2005). For example, Rhode (1987) compared the pre-event fraction for a number of runoff events from different small catchments. He found a decrease of pre-event water fraction for larger events, but no relation to catchment area. Brown et al. (1999) showed that catchment size was negatively correlated with event water contribution during heavy summer rain storms, whereas Shanley et al. (2002) found a positive correlation during snow melt episodes (i.e., the amount of event water increased with catchment size).

Although there is an unavoidable loss of mechanistic understanding when moving from hillslopes or small catchments to larger watersheds, this can be compensated by a more integrated understanding of catchment processes (Soulsby et al., 2006a). Large scale catchment investigations are also needed to improve our ability to understand and predict hydrologic and biogeochemical responses to natural disturbance and human activity over a wide range of climatic and geographic conditions. Furthermore, as there is an urgent need to support decision making at scales where water resource management most often occurs, a more advanced scientific understanding of the hydrological functioning of larger catchments is required (Kirchner, 2006).

In the boreal region, stream networks drain landscapes comprised of a mosaic of forest, wetlands and lakes. This varying landscape organization results in a complex and dynamic hydrology that can vary with stream size, flow and season (Spence and Woo, 2006). Another defining feature of the boreal landscapes is that the hydrology is dominated by snow melt during spring and early summer, often making up 50% of the total annual water yield (Barnett et al., 2005).

At present, little is known about how the pattern in event/pre-event water contribution is affected by major landscape elements or catchment scale in the boreal region. The purpose of this study was therefore to investigate (1) whether similar processes are important for runoff generation in forested and wetland catchments during the spring flood and (2) whether scale (e.g., catchment size) influences the relative contribution of event and pre-event water to spring runoff. Stream water isotopic data from 15 nested streams draining catchments ranging in size from 0.03 to 67 km^2 were used in combination with detailed soil isotopic data from two contrasting catchments (representing peat wetlands and coniferous forest underlain by till, respectively) to answer these questions.

Study area

The study was performed as a part of the multidisciplinary Krycklan Catchment Study at Vindeln Experimental Forests (64°14′N, 10°46′E), approximately 50 km northwest of Umeå, Sweden (Fig. 1). Within this catchment 15 partly nested subcatchments were instrumented for continuous discharge measurements and stream water sampling. The upper part of the catchment, including the two experimental catchments used in this study (Västrabäcken, catchment 2 (C2) and Kallkälsmyren, catchment 4 (C4)), is well documented, as both climatic and hydrological studies have been performed in the area for nearly three decades (Bishop et al., 1990; Folster et al., 2003). Short summers and long winters characterize the climate in the region. Snow covers the ground on average for 171 days, from the end of October to the beginning of May (Ottosson-Löfvenius et al., 2003). The mean annual precipitation and temperature are 600 mm and 0 °C, respectively. Approximately 50% of the annual precipitation falls as snow and the average January temperature is −10 °C. The upland parts of the catchment are mainly forested with Norway spruce (*Picea abies*) in low-lying areas and Scots pine (*Pinus sylvestris*) in upslope areas. There are also large patches of mires predominantly in the upper part of catchment. Further downstream, Nor-

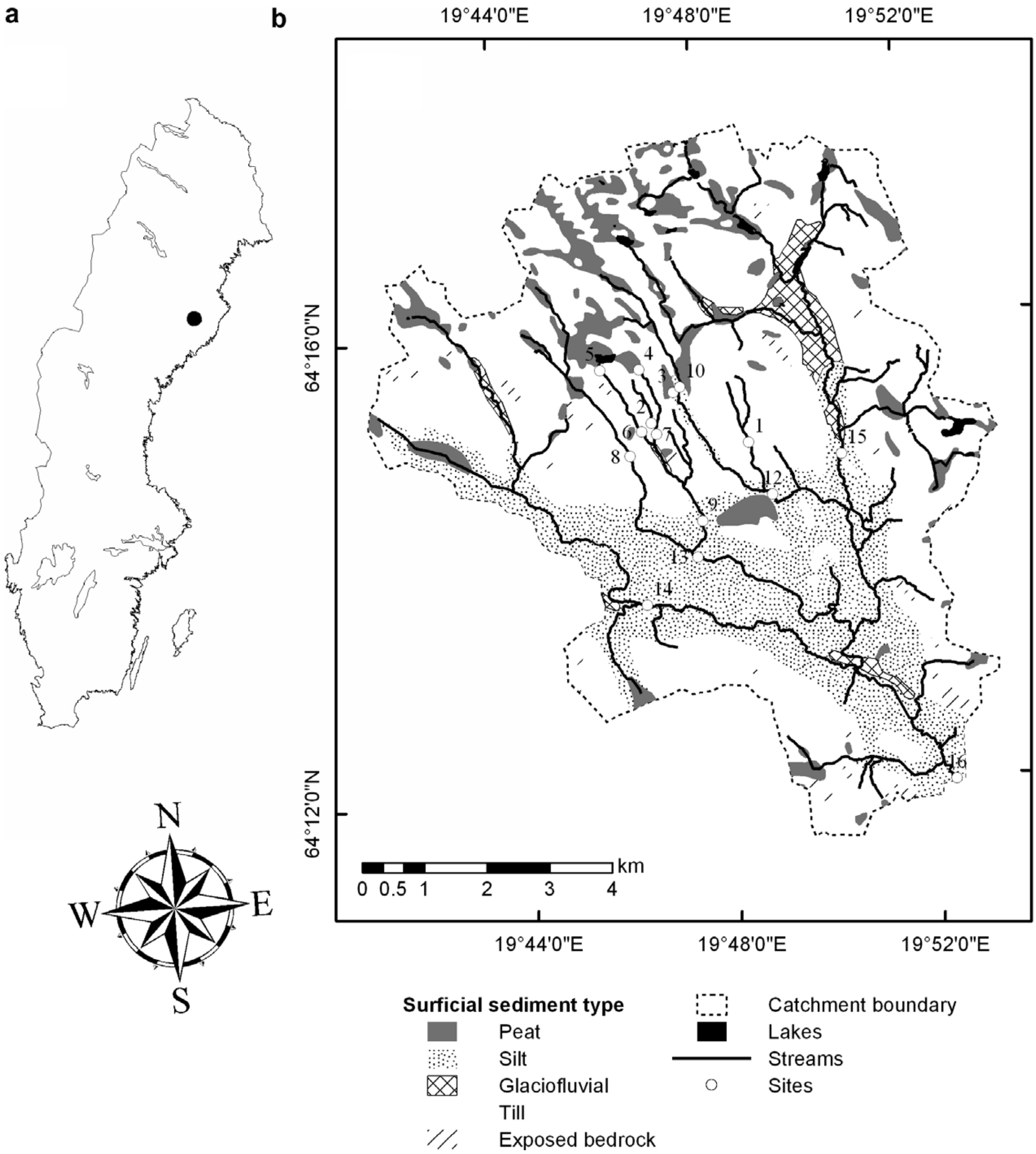

Figure 1 Location of the Krycklan catchment in northern Sweden (a), and distribution of major surficial sediment types and location of stream study sites within the catchment (b).

way spruce and Scots pine are still the dominant tree species, but deciduous trees and shrubs become more common along the stream channels. The glacial till in the upper portion of the catchment gives way to sorted sediments consisting mainly of sand and silt towards the catchment outlet (Fig. 1). Small areas of agricultural fields are found in the lower part of the catchment. More details of the 15 study catchments are presented in Tables 1 and 2 as well as by Cory et al. (2006) and Buffam et al. (2007).

Methods

Continuous measurements of discharge in all 15 catchments (C1—C16, Table 1) have been conducted during the snow free period. Discharge was computed on an hourly basis from water level measurements (WT-HR capacitive water height data loggers, Trutrack Inc., New Zealand) behind thin plate, 90° V-notch weir in smaller streams and at road culverts or well defined natural stream sections in larger

Table 1 Catchment characteristics (scale descriptors and land cover) of 15 streams included in this study

Catchment number	Stream name	Stream order	A_C (km^2)	A_{MSC} (km^2)	Lake (%)	Forest (%)	Wetland (%)	Arable land (%)
C1	Risbäcken	1	0.60	0.16	0.0	87.2	2.7	0.0
C2	Västrabäcken	1	0.13	0.08	0.0	98.7	1.3	0.0
C3	Lillmyrbäcken	1	0.04	0.03	0.0	54.7	45.3	0.0
C4	Kallkälsmyren	1	0.16	0.16	0.0	49.6	50.4	0.0
C5	Stortjärnen Outlet	1	0.81	0.78	5.8	57.5	36.7	0.0
C6	Stortjärnbäcken	1	1.28	0.97	3.7	70.7	25.6	0.0
C7	Kallkälsbäcken	2	0.50	0.21	0.0	82.7	17.3	0.0
C8	Fulbäcken	2	2.51	0.67	0.0	87.8	12.2	0.0
C9	Nyängesbäcken	2	3.11	0.52	1.5	83.8	14.6	0.0
C10	Stormyrbäcken	3	3.25	2.13	0.0	72.7	25.9	0.0
C12	Nymyrbäcken	3	5.71	1.96	0.0	79.3	17.4	0.2
C13	Långbäcken	3	7.26	0.95	0.8	87.4	11.1	0.3
C14	Åhedbäcken	3	12.6	2.11	0.1	86.4	5.7	3.0
C15	Övre Krycklan	4	19.7	0.86	2.0	77.6	13.7	0.2
C16	Krycklan	4	66.8	1.49	0.7	84.0	8.6	1.9

A_C is the entire catchment area, whereas A_{MSC} denotes the median subcatchment area.

Table 2 Catchment surficial sediment coverage for 15 streams used in this study

Catchment number	Peat (%)	Till (%)	Silt (%)	Sand (%)	Gravel (%)	Glaciofluvial (%)	Bare rock (%)
C1	0.0	100	0.0	0.0	0.0	0.0	0.0
C2	0.0	100	0.0	0.0	0.0	0.0	0.0
C3	51.9	41.2	0.0	6.9	0.0	0.0	0.0
C4	55.4	44.6	0.0	0.0	0.0	0.0	0.0
C5	41.0	54.2	0.0	0.0	0.0	0.0	0.0
C6	27.1	67.7	0.0	0.0	0.0	0.0	2.0
C7	17.9	82.2	0.0	0.0	0.0	0.0	0.0
C8	17.3	81.3	0.0	0.0	0.0	0.0	1.4
C9	14.7	75	5.9	0.0	1.5	0.0	1.6
C10	29.8	69.4	0.0	0.8	0.0	0.0	0.0
C12	20.5	72	3.4	4.1	0.0	0.0	0.0
C13	12.3	69.6	15.7	0.0	0.7	0.0	1.2
C14	7.3	58.3	30.8	0.0	0.0	1.8	1.8
C15	14.1	73.3	2.0	0.0	0.0	7.9	0.8
C16	9.5	59.4	25.7	0.4	0.1	3.0	1.3

Sediment types as described in text.

streams. Rating curves were derived based on discharge measurements using salt dilution or the time-volume (bucket) method. For winter periods reliable discharge measurement were available from only one stream (C7) where a V-notch weir was located inside a heated housing. Data from this stream, scaled for differences in catchment area, were used to compile continuous discharge records for both winter baseflow and the snow melt period for all 15 streams used in this study. The stream sampling program was based on weekly samples of baseflow prior to the onset of the snow melt, and then every second to third day during the spring until the flow receded to levels close to baseflow.

Snow lysimeters with an area of 1.44 m^2 were placed in the catchment in three types of vegetation: closed canopy spruce forest, open canopy pine forest and open field. For each vegetation type there were three replicates and, thus, in total nine snow lysimeters were used. The snow lysimeters were used for measuring snow melt volumes, melt intensity and isotopic composition of melt water. Snow melt water was sampled every 1–3 days and then bulked into 4–6-day intervals for $\delta^{18}O$ analyses. Vegetation type classification based on satellite images (Reese et al., 2003) was used to compute weighted areal mean values from the snow melt volume and $\delta^{18}O$ contribution from spruce, pine and open canopy for each catchment. Snow melt was measured every 1–3 days. These data were linearly interpolated to an hourly melt rate following Laudon et al. (2002). A snow survey of snow cores (N = 40) distributed throughout the 67 km^2 catchment was conducted prior to the onset of snow melt to estimate the spatial variability of bulk snow $\delta^{18}O$.

Detailed soil water measurements were conducted in two of the catchments: in the forested dominated catchment (C2) and in the wetland dominated catchment (C4). In C2 soil water samples were collected from three soil profiles located

4, 12 and 22 m from the stream along a transect during the spring of 1999 and 2004 (Laudon et al., 2004b; Petrone et al., 2007). The transect was aligned, based on the topography, to follow the assumed lateral flow paths of the groundwater towards the stream. The riparian soil profile closest to the stream, S04, was dominated by organic material. The upslope location, S22, was located in a typical podzolic soil with a 10–15 cm organic layer overlying the mineral soil. Soil profile S12 was between the riparian and the upslope location in an organic rich mineral soil. Each profile consisted of ceramic suction lysimeters (P100), as well as time domain reflectrometry probes (TDR) and thermistors connected to a Campbell Scientific data logger (CR10) to measure soil water content and soil temperatures at six soil depths between 5 and 90 cm. Soil frost was assumed when soil temperature was below 0 °C at a specific depth. Groundwater levels were also measured continuously using pressure transducers connected to a CR10 in shallow groundwater wells perforated along the full length and extending approximately 1 m below the soil surface.

Soil measurements in the wetland dominated catchment (C4) were conducted using 12 nested wells extending to different depths in the wetland, ranging from 25 to 350 cm below the ground surface. The wells have a closed bottom and were perforated over the lowest 10 cm. Samples from each well (if the water at the sampling depth was unfrozen) were collected at pre-flood and peak flood during both 1999 and 2004. The soil frost depth at the wetland of C4 was estimated indirectly using the ice cores that had developed over winter in the wells and by using an ice drill on an adjacent wetland during the winter of 2002.

All samples for $\delta^{18}O$ were stored in 25 ml glass bottles free of headspace and refrigerated before analysis. The $\delta^{18}O$ analysis was based on a CO_2–H_2O equilibration technique using a GAS Bench II (Finnigan MAT) connected to a Delta plus mass spectrometer (Finnigan MAT). Analyses reproducibility was better than 0.2 ‰ for $\delta^{18}O$.

A two-component hydrograph separation (Eq. (1)) was used to separate event and pre-event water in the stream and soil. The fraction of pre-event water (f_p) was calculated as:

$$f_p = \frac{\delta^{18}O_s - \delta^{18}O_e}{\delta^{18}O_p - \delta^{18}O_e} \tag{1}$$

$\delta^{18}O_s$, $\delta^{18}O_e$ and $\delta^{18}O_p$ are isotopic compositions and the subscripts s, e and p refer to stream water (sampled runoff water), event water (new melt or rain water) and pre-event water (water in the catchment prior to the event), respectively.

The $\delta^{18}O_e$ was defined using the runCE method proposed by Laudon et al. (2002) and tested by Laudon et al. (2004b) (Eq. (2)). The runCE method accounts for both the timing and the amount of melt water entering the soil water reservoir, as well as the runoff of previously melted and subsequently stored water in the soil at every time step during the episode. The isotopic composition of this event water is based on a comparison between cumulative snow melt (and rain water contributions) from the snow lysimeters and the cumulative volume (depth) of melt water that has left the snow pack but has not yet been discharged to the stream during the event. Assuming that the soil reservoir is fully mixed, the time lag between the melting of snow and its arrival at the stream is taken into account (Laudon et al., 2002):

$$\delta^{18}O_e(t) = \left(\sum_{i=1}^{t} M(i)\delta^{18}O_m(i) - \sum_{i=1}^{t} E(i)\delta^{18}O_e(i)\right) \Bigg/ \left(\sum_{i=1}^{t} M(i) - \sum_{i=1}^{t} E(i)\right) \tag{2}$$

$M(i)$ is the incrementally collected melt water depth, and $E(i)$ is the incrementally calculated event water discharge at time t (Eq. (2)). $\delta^{18}O_e(i)$ and $\delta^{18}O_m(i)$ are the event and melt water isotopic compositions, respectively.

We performed an uncertainty analysis to quantify the effects of different sources of errors on the pre-event calculation at the 15 catchments. The uncertainty of the IHS was calculated using the method proposed by Laudon et al. (2002) which is based on both the analytical uncertainty and error propagation from the event water calculation. A standard error of 0.2‰ for the $\delta^{18}O$ analysis was used in the uncertainty analyses. Furthermore, the uncertainty in spatial variability in the use of snow lysimeters in both snow melt volumes and $\delta^{18}O$ was tested. A 30% error in snow melt volumes and 1‰ error in snow isotopic composition was used to estimate the maximum uncertainty that could occur with the available data. This was based on maximum melt water volumes and $\delta^{18}O$ variability between the snow lysimeters on any given day. The uncertainty associated with the assumption of a spatially uniform specific discharge across all 15 sites was also tested. A 30% maximum error in specific discharge was used for all sites. This was based on a statistical analysis of measured instantaneous discharge values at the 15 streams at different occasions. The mean error (error here being the difference between instantaneous discharge at a given site and discharge at C7, the reference site) was 25% for occasions with medium flows (0.5–2.5 mm day^{-1}, $N = 107$) and 24% for high flows (>2.5 mm day^{-1}, $N = 89$).

All landscape calculations were based on a gridded digital elevation model (DEM) with a grid resolution of 50 m complemented with field surveys for the smallest subcatchments. A gridded stream network was computed from the DEM after 'burning in' the mapped streams into the DEM. For each stream cell the catchment area (A_C) based on topography was computed. In a second step, for each stream cell the median subcatchment area (A_{MSC}) was computed as the median catchment area of all stream cells upstream of the cell in question. McGlynn et al. (2003) computed this measure for a catchment in New Zealand and found a correlation between A_{MSC} and mean transit times of water in the runoff. The idea behind this measure is that the subcatchment area distribution quantified by its median might be a more suitable measure of catchment organization and function than total catchment area. As an example one can consider two subcatchments of the same size, the first with one long stream, the other with two similar streams joining just above the outlet. In the first case A_{MSC} is the local catchment area accumulated in the stream cell halfway along the one long stream, in the second case A_{MSC} is the local catchment area halfway along the smaller streams. Thus, while A_C is the same for both cases, A_{MSC} is smaller in the second case. Generally, A_{MSC}

is smaller if the upstream stream network has a more tree-like structure. While A_C increases continuously along the stream, A_{MSC} tends to increase with A_C only in the headwaters. At a certain scale where larger catchments are mainly the sum of small catchments, A_{MSC} remains rather constant while A_C still increases. This is manifested by the strong correlation between A_{MSC} and A_C for the smaller catchments (<3 km^2) in Krycklan ($R^2 = 0.61$, $p < 0.01$, $n = 8$) while the correlation for the larger watersheds (>3 km^2) is lacking entirely ($R^2 = 0.01$, $p > 0.1$, $n = 7$).

For land cover data the 'property' map (1:12,500, Lantmäteriet, Gävle, Sweden) was used to define the coverage of lakes, forest, wetlands and agricultural areas (Table 1). A Quaternary deposits coverage map (1:100,000, Geological Survey of Sweden, Uppsala, Sweden) was used to define the surficial coverage of peat, till, silt, sand, gravel, glaciofluvial sediment, and bare rock (Table 2). Till coverage included the sum of two categories from the original map, ''till'' and ''thin sediment deposits'' which referred to regions with particularly shallow till deposits over bedrock. Multiple linear regression (MLR) analysis (SPSS, version 14) was used to test which catchment characteristics best explained the pre-event fraction during the rising limb, peak flood and falling limb, respectively. For the MLR analyses, the backward selection method was used (with $P_{out} = 0.10$), with the proportion of the major sediment types ($p_{peat}, p_{silt}, p_{till}$) and scale measures ($\log(A_C)$ and A_{MSC}) as potential predictor variables. Other, minor sediment types were excluded from analysis due to their rarity (<2% total coverage in Krycklan, <10% in all subcatchments, and 0 coverage in over half of the subcatchments). Land cover variables were also excluded as they are highly co-correlated with surficial sediment coverage variables. Peat and wetland coverages, for instance, were almost identical and highly correlated ($R^2 = 0.98$, $p < 0.0001$).

Results

Snow melt started in the middle of April, following more than 5 months of permanent snow cover. Total snow melt during the spring was 126 mm ($\sigma = 5$ mm) from open field, 135 mm ($\sigma = 12$ mm) from open canopy and 116 mm ($\sigma = 12$ mm) from closed canopy. Maximum snow melt intensity also varied spatially, with the highest melt rate of 15 mm day^{-1} on April 19 in open field, 20 mm day^{-1} on April 26 in open canopy and 17 mm day^{-1} on May 3 in closed canopy. Average runoff during late winter (March to mid-April), prior to any snow melt was 0.15 mm day^{-1}. Runoff peaked at 8 mm day^{-1} on May 4. Between April 18 and May 17, the hydrograph exceeded 1 mm day^{-1} and between April 30 and May 5 flow exceeded 5 mm day^{-1}. The period at the onset of spring flood with flow between 1 and 5 mm day^{-1} is denoted the *rising limb*. The period exceeding 5 mm day^{-1} is denoted the *peak flood* and the period at the receding end of the hydrograph with a flow below 5 but above 1 mm day^{-1} is the *falling limb* (Fig. 2). The total runoff during spring flood was 93 mm, which is approximately 30% of the long-term annual average runoff in the catchment.

Isotopic composition of melt water displayed a large temporal, but relatively small spatial, variability. The $\delta^{18}O$ in the first melt water leaving the snow pack was −18.1‰, −18.3‰ and −17.3‰ in open field, open canopy and closed canopy, respectively. At the end of the melt season the corresponding values were −15.7‰, −15.8‰ and −15.2‰, suggesting a large fractionation during snow melt. The volume-weighted average snow melt water from the snow lysimeters was −16.1‰, −16.0‰ and −15.9‰ for open field, open canopy and closed canopy, respectively. The spatial variability of $\delta^{18}O$ was also relatively small for the snow sampled during the snow survey in open fields distributed over the far south, east, west and north corners of the 67 km^2 catchment. The average value for $\delta^{18}O$ was −17.0‰ with a standard deviation of 0.40‰ ($N = 40$), indicating a relatively small spatial variability of $\delta^{18}O$ in the catchment. The difference in isotopic composition of open field snow lysimeter and snow survey samples is attributed to fractionation during evaporation and was hence not included in the IHS.

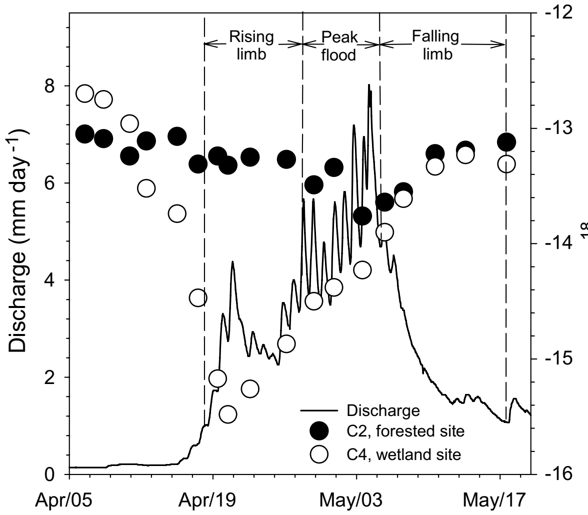

Figure 2 Spring flood discharge at catchment, C7, defining the rising limb, peak flood and falling limb periods as used in this study. Oxygen-18 ($\delta^{18}O$) values for catchments C2 and C4 are shown.

Isotopic composition of the saturated and near saturated soil water in the riparian zone of the forested catchment, C2, prior to snow melt ($\delta^{18}O$ −13.2‰, $\sigma = 0.27$‰, $N = 4$ depths, Fig. 3a) was close to baseflow stream water $\delta^{18}O$ composition (−13.1‰, $\sigma = 0.11$‰, $N = 6$) at the 13 ha catchment outlet. At peak flow a change in both soil water isotopic composition to an average saturated value in the riparian soil of −13.8‰ ($\sigma = 0.20$‰, $N = 4$) and in the stream outlet (−13.8‰, $\sigma = 0.10$‰, $N = 3$) had occurred. Soil water samples collected prior to snow melt in the wetland wells demonstrated a $\delta^{18}O$ gradient with depth, from −10.8‰ at 50 cm depth, which was just below the depth of soil frost, to −13.6‰ at 300 cm depth. The outlet of the wetland, C4, had a baseflow $\delta^{18}O$ signature of −12.8‰ ($\sigma = 0.12$‰, $N = 4$). During peak flow major deviations from baseflow values could be seen for two depth intervals, 0–50 cm and 200–250 cm, whereas other soil depths remained constant during the spring flood (Fig. 4a). Water standing at the surface had a $\delta^{18}O$ value of −16.0‰. At the catchment outlet at C4 the rising limb $\delta^{18}O$ was on average −15.2‰

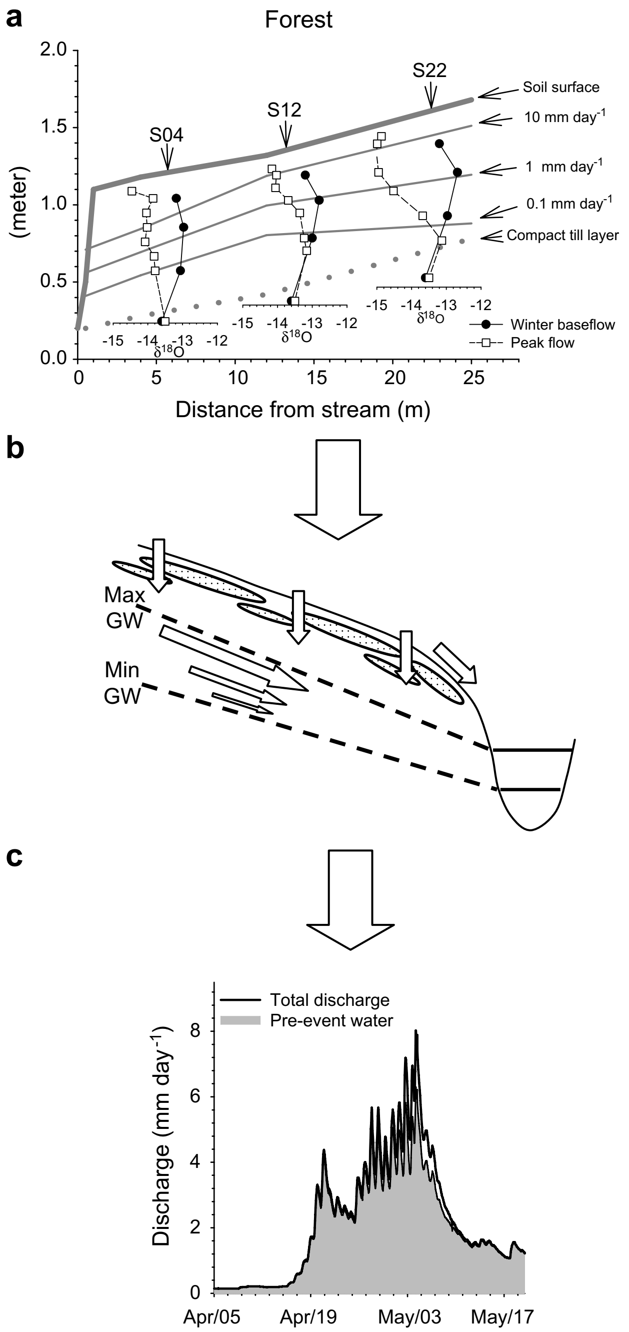

Figure 3 Evolution of event and pre-event water during the spring flood through the forested catchment soils, in C2 (a). A schematic view of the soil hydrological flow pathways during snow melt, where snow melt waters infiltrate through the topsoil because of permeable soil frost (grey ovals), raising the groundwater level into soil horizons of higher hydrologic conductivity which results in more rapid lateral flow (b). The resulting stream spring hydrograph in the forested catchment is dominated by pre-event water (c).

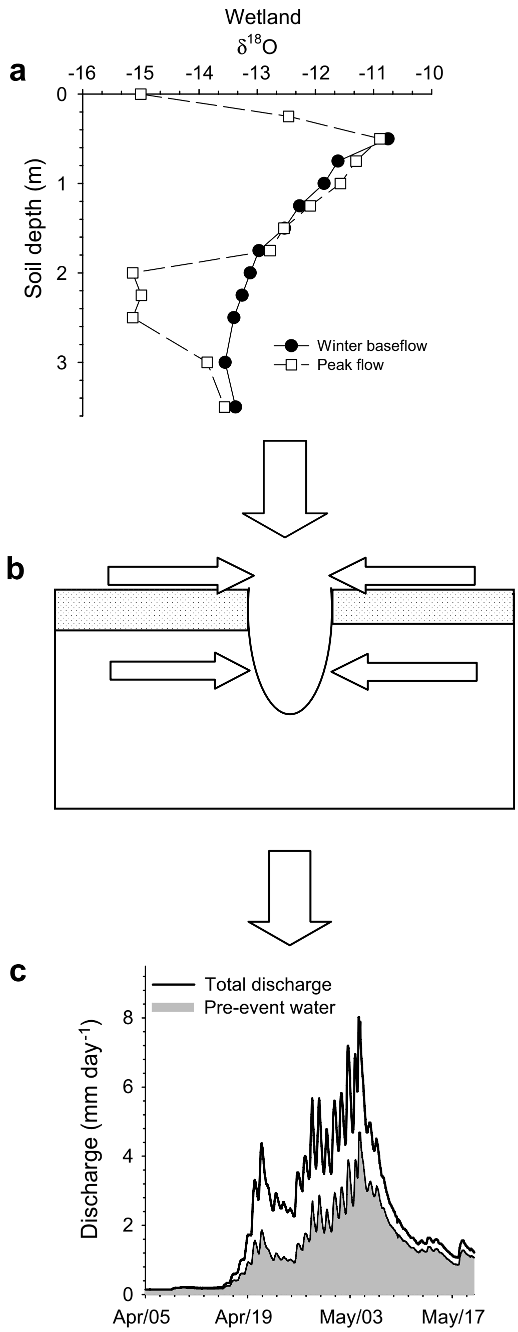

Figure 4 Evolution of event and pre-event water during the spring flood through the wetland peat, in catchment C4 (a). A schematic view of the subsurface hydrological flow pathways where snow melt waters cannot infiltrate the soil because of impermeable soil frost (grey rectangles) and instead runoff as overland flow or through preferential flow pathways (b). The resulting stream spring hydrograph in the wetland catchment has similar proportions of event and pre-event water (c).

(σ = 0.25‰, N = 4; Fig. 2) and at peak flow −14.2‰ (σ = 0.16‰, N = 3; Fig. 5).

Both base flow and peak flow δ^{18}O composition displayed a large variability between the 15 streams (Fig. 5). Baseflow δ^{18}O was heaviest in the two streams with >3% lakes in the catchment (C5 and C6) followed by C9 and C15 with 1% to 3% lakes (Fig. 5). For all streams δ^{18}O values decreased during snow melt towards the isotopic composition of snow

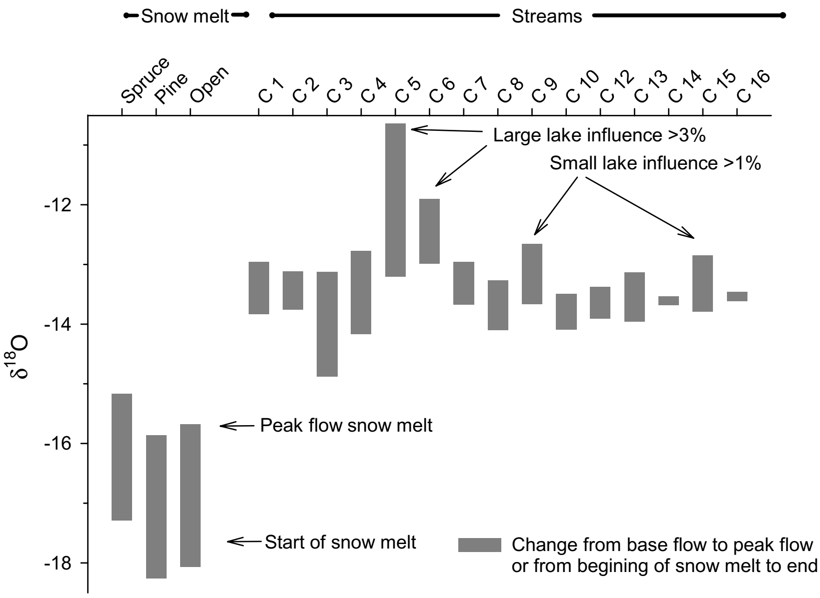

Figure 5 Change in $\delta^{18}O$ signature during the spring flood, from baseflow to peak flow in the 15 streams and from beginning of snow melt to the termination of snow melt from lysimeters in three types of vegetation: closed canopy spruce forest, open canopy pine forest and open field. Note that the $\delta^{18}O$ signal in snow melt water goes from light to heavy, whereas the opposite is true for the stream water.

melt water, with the largest change in C3, C4, C5 and C6. The smallest $\delta^{18}O$ change was found in two of the largest sites (C14 and C16).

The pre-event water fraction varied between 43% and 93% during the rising limb, between 36% and 90% during peak flow and between 61% and 98% during the falling limb (Table 3). During all three phases, the smallest pre-event water contribution was found for the small wetland-dominated catchments (C3, C4 and C5). The largest pre-event water fraction was found for two of the largest catchments, C14 and C16. Using simple linear regression as an exploratory tool, percent coverage of peat was the single factor that best explained the pre-event water contribution (Table 4). Peat coverage alone explained 79%, 63% and 33% of the variation in the rising limb, peak flow and falling limb pre-event water fraction, respectively (Fig. 6).

MLR analyses can be difficult to interpret when potential predictor variables covary strongly. The only significant ($p < 0.05$) bivariate correlations in the predictor dataset were a negative correlation between coverage of peat and till ($r = 0.79$), a weak positive correlation between A_{MSC} and silt coverage ($r = 0.54$), a positive correlation between A_C and A_{MSC} ($r = 0.65$), and a positive correlation between A_C and silt coverage ($r = 0.72$). Exploratory multiple linear regression (MLR) analysis demonstrated that the combination of the two most common

Table 3 Summary of the pre-event water fraction during spring flood and peak flood, including uncertainty estimates in parentheses

Catchment number	Rising limb (%)	Peak flood (%)	Falling limb (%)
C1	83 (8)	71 (10)	75 (12)
C2	77 (7)	70 (10)	84 (11)
C3	59 (7)	36 (10)	61 (14)
C4	43 (7)	54 (9)	74 (11)
C5	67 (4)	52 (5)	65 (6)
C6	74 (5)	70 (7)	79 (8)
C7	73 (7)	75 (10)	87 (11)
C8	81 (7)	69 (10)	80 (12)
C9	73 (7)	67 (9)	82 (10)
C10	70 (8)	73 (11)	91 (14)
C12	76 (8)	77 (11)	90 (13)
C13	77 (7)	70 (11)	83 (12)
C14	75 (8)	90 (12)	99 (14)
C15	77 (7)	70 (9)	84 (10)
C16	84 (8)	90 (11)	98 (13)

Table 4 Strength of regression (adjusted R^2) between pre-event fraction and different combinations of catchment predictor variables

Response period	Univariate predictor variables					MLR combinations of predictor variables					
	A_C	A_{MSC}	p_{till}	p_{peat}	p_{silt}	p_{till}, p_{silt}	p_{seds} [a]	A_C, A_{MSC}	A_C, p_{seds}	A_{MSC}, p_{seds}	A_C, A_{MSC}, p_{seds}
Rising limb	n.s.	n.s.	0.56	**0.79**	n.s.	0.74	0.79	n.s.	0.77	0.77	0.77
Peak flood	0.35	0.21	0.20	0.63	0.27	0.72	0.72	0.31	0.74	**0.81**	0.79
Falling limb	0.42	0.41	n.s.	0.33	0.37	0.53	0.53	0.46	0.56	**0.71**	0.69

The best model for each period as selected by MLR analysis is indicated with boldface type.
n.s. = not significant at the $p = 0.10$ level.
[a] p_{seds} = proportional coverage of sediments, including only those combinations which are best correlated with pre-event water fraction in the MLR analysis. For the rising limb period, this is p_{peat}. For the peak flood and falling limb periods, this includes both p_{till} and p_{silt}.

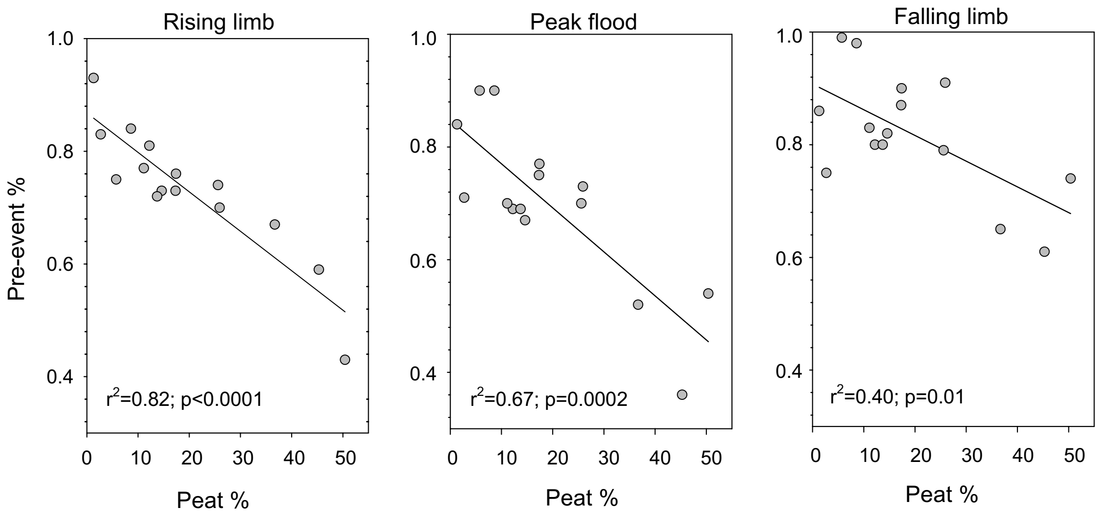

Figure 6 Correlation between peat percent coverage and pre-event water fraction during rising limb, peak flood, and falling limb.

surficial mineral sediment types (silt in the lower portion of the catchment and till primarily in the upper portion) was equally or more highly correlated than peat coverage with pre-event fraction, depending upon the period (Table 4). A_C and A_{MSC} were uncorrelated with pre-event fraction during the rising limb, but were weakly to moderately positively correlated with pre-event fraction during peak flood and the falling limb. When combined with sediment coverage characteristics, the inclusion of A_C did not further substantially improve the correlation with pre-event fraction, while the inclusion of A_{MSC} gave substantial additional explanatory power during peak flood and particularly for the falling limb period (Table 4). For the rising limb only peat coverage was significant in the best MLR model, explaining 79% of the variation in pre-event water during the rising limb (Tables 4 and 5). During peak flood A_{MSC} together with till and silt coverage explained 81% of the variability in pre-event water in the best MLR model. During the falling limb the same three variables were again selected, explaining 71% of the variation in pre-event water (Tables 4 and 5).

Soil frost depth (measured as temperature <0 °C) in C2 extended down to approximately 30 cm soil depth, but due to the low water content of the upper soil horizons (>10%) the soil frost was likely porous, allowing melt water to infiltrate into the soil. Furthermore, no pooling of water

Table 5 Best significant MLR model results for each time period

Time period	MLR equation	Adjusted R^2
Rising limb	$f_{pre\text{-}event} = 0.862 - 0.606p_{peat}$	0.79
Peak flood	$f_{pre\text{-}event} = 0.188 + 0.068A_{MSC} + 0.768p_{silt} + 0.584p_{till}$	0.81
Falling limb	$f_{pre\text{-}event} = 0.523 + 0.072A_{MSC} + 0.511p_{silt} + 0.295p_{till}$	0.71

Units of catchment area (A_C) and median subcatchment area (A_{MSC}) are km^2; p_{peat}, p_{till}, p_{silt} are the relative proportions of catchment area covered by the respective surficial sediment types.

during snow melt was ever observed in the forested soils. In the wetland soils of C4, indirect evidence suggests a continuous soil frost covering the waterlogged soils. The ice cores that developed during the winter in the sampling wells were solid ice and extended approximately 30–50 cm below the soil surface. Ice drilling of an adjacent wetland during the winter of 2002 also showed that a solid ice lens developed already in December and was at the time of spring flood more than 30 cm deep (P. Blomkvist, Umeå University, Personal Communication). Furthermore, visual observations at the beginning of the snow melt period at C4 showed large quantities of pooled melt water at the soil surface, which at least partly, drained directly to the stream.

The uncertainty analysis based on both the analytical uncertainty and error propagation from the event water calculation revealed an uncertainty of 4–8% for the pre-event component during the rising limb, 5–12% during peak flood and 6–14% during the falling limb of the hydrograph (Table 3). The largest uncertainty was in general found at the sites with peak flow δ^{18}O values closest to snow melt values and the two largest sites with the least change in δ^{18}O during the spring. Pre-event water calculation uncertainty based on errors in both snow melt volumes and isotopic composition, as well as in specific discharge, was less than 5% for all catchments.

Discussion

Large differences in hydrological flow paths between the forested catchment (C2) and the wetland dominated catchment (C4) were evident from the isotopic composition of both soil and stream water. In C2 the hydrograph was dominated by pre-event water during all three hydrological phases (77%, 70% and 84%, respectively). Despite the large pre-event fraction in the stream, the soil water isotopic composition in the forested catchment indicates that soil water is affected by recent snow melt water, but that the effect is larger at higher flows and further away from the stream (Fig. 3a). This suggests that a majority of the melt water is infiltrating into the soil, raising the groundwater level and resulting in mobilization of pre-event water stored in the hillslope to the stream (Fig. 3b and c). As evident from the increased event water fraction in the soil transect at the later stage of the hydrograph, snow melt water will eventually reach the stream, but not until high flow conditions when the melting snow has replaced the pre-melt water.

In contrast to the forested catchment, the wetland dominated catchment, C4, had a lower pre-event water fraction during all three hydrological phases (43%, 54% and 74%, respectively; Fig. 4a). The soil water isotopic data indicated two hydrological flow pathways: a shallow pathway close to the surface and a preferential pathway at 200–250 cm depth (Fig. 4b and c). The isotopic composition of the water standing on the surface in combination with the rapid input of event water at the rising limb of the hydrograph indicates that overland flow, especially at the onset of the snow melt period, is an important pathway during snow melt (although this pathway has not been quantified relative to the preferential flow pathway). This pathway is probably caused by a continuous soil frost layer inhibiting melt water from infiltrating into the peat. The thick continuous soil frost (>30 cm extent in winter) is a common phenomenon in wetlands in the area and originates due to wet conditions and high groundwater levels in autumn before the winter cold arrives. The preferential pathway at 200–250 cm depths is probably originating from melt water infiltrating soils in the forest surrounding the wetland or at the wetland perimeter and flowing along layers of higher hydrological conductivity (Sirin et al., 1998).

The forested catchment follows the general pattern found elsewhere with a hydrograph dominated by pre-event water during snow melt (Shanley et al., 2002). In boreal till-derived soils the hydrological conductivity typically increases exponentially towards the soil surface (Rhode, 1987), which could explain the larger fraction of event water during the period of highest flow (peak flood). Soil frost in the forested C2 catchment does not seem to substantially affect the partitioning of event and pre-event water. In contrast, the wetland catchment response illustrated the importance of soil frost in generating large event water contributions, especially at the onset of the snow melt season, a mechanism which has rarely been documented in the boreal region. Similar findings have been shown for arctic watersheds with continuous permafrost where the spring flood is often totally dominated by event water (McNamara et al., 1997). In areas with discontinuous permafrost the contribution of pre-event water varies depending on melt intensity, active layer development (Hayashi et al., 2004; Metcalfe and Buttle, 2001; Woo et al., 2000; Quinton and Marsh, 1999) and topography (McEachern et al., 2006). Overland flow due to soil frost has also been shown for agricultural fields, where soil frost may cause overland flow because of reduced infiltration capacity (Dunne and Black, 1971; Stähli et al., 1996). The current study is, as far as we are aware, the first to clearly demonstrate the connection between soil frost and the partitioning of event and pre-event water during snow melt in boreal regions that do not contain permafrost.

The importance of soil frost in generating a large event water fraction from wetland dominated catchments during snow melt episodes is not a unique occurrence of the 2004 spring flood or for this particular catchment. A similar event water contribution has been described in a previous spring flood study by Cory (1999) at the same stream (C4). The strong correlation between event water contribution of the 15 streams and catchment peat coverage (Fig. 6) also supports this finding. The varying dynamics of dissolved organic carbon (DOC) with increasing values in forested catchments and decreasing values in wetland dominated catchments during the spring flood have been attributed to a dilution effect of DOC during spring flood in wetland dominated catchments (Laudon et al., 2004a; Ågren et al., 2007). This agrees with the finding of a large event water component during spring flood.

In the MLR analysis in this study, depending on the time period, either peat (negatively) or the predominant mineral sediment coverage (silt and till, both positively) were significantly correlated with pre-event fraction. Since the peat coverage is essentially 100% minus the sum of the percent coverage of the other two sediment types, these results are complementary rather than contradictory. While we have provided a mechanistic explanation for the importance of both peat and till coverage in the catchment for the event and pre-event partitioning particularly during the rising limb (Figs. 3 and 4), the role of silt during the peak flood and falling limb is less clear. There are different possible explanations for the negative influence of silt coverage and event water contribution during the snow melt period. The hydraulic conductivity in silty sediments is lower compared to soil matrixes found in till and peat catchments. Lower hydraulic conductivity gives rise to longer transit times compared to a more rapid delivery of event water compared to soils with higher hydraulic conductivity. This agrees with findings by Soulsby et al. (2006b) where both flow path partitioning and transit time were correlated to the distribution of hydrologically responsive soils in the catchment. The silt coverage might also indirectly point towards other covarying catchment characteristics that are not included in the analyses such as topography or riparian buffer width.

In addition to the importance of peat (or silt and till) for the separation of event and pre-event water during snow melt, the multiple linear regression analyses also showed a clear influence of median subcatchment area (A_{MSC}) during the peak flood and falling limb periods. In contrast, catchment area (A_C) apparently did not affect the partitioning once the sediment characteristics had been accounted for (Table 4). The fact that A_{MSC} (rather than A_C) together with sediment type is strongly correlated with pre-event fraction suggests that it is primarily stream network organization, rather than size, which influences hydrological pathways in Krycklan. This supports previous research by McGlynn et al. (2004) that demonstrated that A_{MSC} was correlated with transit time of water in different catchments.

Upscaling of hydrological processes understanding gained at the scale of hillslopes and small catchments to understand and predict patterns at larger scales has received increased scientific attention recently (Blöschl, 2006; Shaman et al., 2004; Soulsby et al., 2003; Uchida et al., 2005). The key question is whether runoff from larger basins is simply the sum of runoff from hillslopes and small catchments in the basin. In other words, do the governing runoff generation processes remain the same or do other processes become more important at larger scales? In this study, the partitioning between event and pre-event water depended on median subcatchment area and the proportion of major sediment types, rather than on total catchment area. This indicates that catchment size is less important for runoff generation during the spring flood than the relative coverage of major landscape units and the organization of the stream network at the scale of headwaters. These results suggest some scale-independence with respect to runoff generation processes, in spite of catchment area varying by over three orders of magnitude. This is similar to the findings of McGlynn et al. (2003) and McGuire et al. (2005), which both found catchment size to be poorly correlated with water transit times. However, other studies have shown that catchment size was negatively correlated with event water contribution during heavy summer rain storms (Brown et al., 1999), whereas Shanley et al. (2002) found a positive correlation during snow melt. These different findings warrant further investigation on the role of catchment size during different seasons and climatic conditions.

Concluding remarks

The detailed process understanding that was derived from the isotopic and hydrometric measurements of runoff, combined with soil measurements in wetland and forested sites, enabled the development of a conceptual framework to help understand the variability in hydrological pathways over a range of catchments. The results suggest scale-independence in terms hydrological flow pathways and instead some degrees of self-similarity in the Krycklan catchment, determined largely by the amount of wetland area, during the snow melt period across spatial scales from small (<0.1 km^2) to mesoscale (67 km^2) catchments. The results further highlight the importance of stream network organization (as indicated by median subcatchment area) and different runoff generation processes in different landscape elements (e.g., peat wetlands and silt deposits). These results can be useful when trying to disentangle the temporal dynamics in hydrology and biogeochemistry during snow melt episodes when moving from small headwater streams to the catchment outlet.

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