Hydrological flow paths during snowmelt: Congruence between hydrometric measurements and oxygen 18 in meltwater, soil water, and runoff

Hjalmar Laudon,[1] Jan Seibert,[2] Stephan Köhler,[3] and Kevin Bishop[2]

Received 4 July 2003; revised 31 October 2003; accepted 16 January 2004; published 6 March 2004.

[1] Streamflow generation in boreal catchments remains poorly understood. This is especially true for snowmelt episodes, which are the dominant hydrological event in many seasonally snow covered regions. We examined the spatial and temporal aspects of flow pathways by linking detailed oxygen 18 observations of stream, melt, soil, and groundwater with hydrometric measurements in a small catchment in northern Sweden during the snowmelt period. The results demonstrate that soil horizons below 90 cm were hardly affected by the approximately 200 mm of snowmelt water infiltrating into the soil during the spring. The approximately sixtyfold increase in runoff, from 0.13 mm d^{-1} to 8 mm d^{-1}, was generated by a 30–40 cm rise of the groundwater level. The total runoff during the snowmelt period from late April to late May was 134 mm, of which 75% was preevent water. Mass balance calculations based on hydrometric and isotopic data independently, both using upscaling of a hillslope transect to the entire 13-ha catchment, provided similar results of both water storage changes and the amount of event water that was left in the catchment after the snowmelt. In general, groundwater levels and runoff were strongly correlated, but different functional relationships were observed for frozen and unfrozen soil conditions. Although runoff generation in the catchment generally could be explained by the transmissivity feedback concept, the results suggest that there is a temporal variability in the flow pathways during the spring controlled by soil frost during early snowmelt. *INDEX TERMS:* 1860 Hydrology: Runoff and streamflow; 1836 Hydrology: Hydrologic budget (1655); 1821 Hydrology: Floods; 1823 Hydrology: Frozen ground; *KEYWORDS:* hydrograph separation, oxygen 18, snowmelt, spring flood, boreal, northern Sweden, transmissivity feedback

Citation: Laudon, H., J. Seibert, S. Köhler, and K. Bishop (2004), Hydrological flow paths during snowmelt: Congruence between hydrometric measurements and oxygen 18 in meltwater, soil water, and runoff, *Water Resour. Res.*, *40*, W03102, doi:10.1029/2003WR002455.

1. Introduction

[2] Hydrograph separation using stable isotopes (isotope hydrograph separation, IHS) is a widely recognized method for quantifying runoff components during snowmelt and rain events. The use of IHS dates back to the pioneering work by *Dinçer et al.* [1970] and has since been used in various environments world wide [*Sklash et al.*, 1986; *McDonnell*, 1990; *Pionke and Dewalle*, 1992; *Lakey and Krothe*, 1996; *Laudon and Slaymaker*, 1997; *Uhlenbrook et al.*, 2002]. The most common finding has been that so-called old, or preevent, water dominates the hydrograph during the event whereas there was only a relatively small contribution from new, or event, water during storms and snowmelt events (see *Kendall and McDonnell* [1998] for a review).

[1]Department of Forest Ecology, Swedish University of Agricultural Sciences, Umeå, Sweden.

[2]Department of Environmental Assessment, Swedish University of Agricultural Sciences, Uppsala, Sweden.

[3]Géochimie, Transferts et Mécanismes, CNRS, UMR 5563, University of Toulouse, Toulouse, France.

0043-1397/04/2003WR002455$09.00

[3] Most IHS studies have been based on isotopic information for only the input and output, i.e., precipitation/snowmelt and streamflow. Thus they have relied on assumptions about what occurs in the soil. Such a separation approach is a useful tool for defining sources of water, but it does not provide any information about hydrological flow paths or runoff mechanisms as soils are essentially treated as a black box [*Kendall et al.*, 2001]. Several authors have recently called for the use of more detailed isotopic information from within the catchment or hydrometric data in IHS studies to better understand the hydrological processes in the catchment [*Rodhe*, 1998; *Burns*, 2002]. Although the value of incorporating groundwater or soil-water isotopic information in the separation analyses has been known for some time [*Dewalle et al.*, 1988; *Ogunkoya and Jenkins*, 1993; *Hinton et al.*, 1994; *McGlynn et al.*, 1999] only a few studies have actually used such information, mainly because collecting this data often is time consuming and difficult, especially during winter conditions with deep snowpacks and extensive soil frost.

[4] To our knowledge, a mass balance to test the inferred changes in soil water storage and the amount of new water entering that storage has not previously been published. In this study we combined detailed hydrometric investigations (soil water measured by time domain reflectometry (TDR)

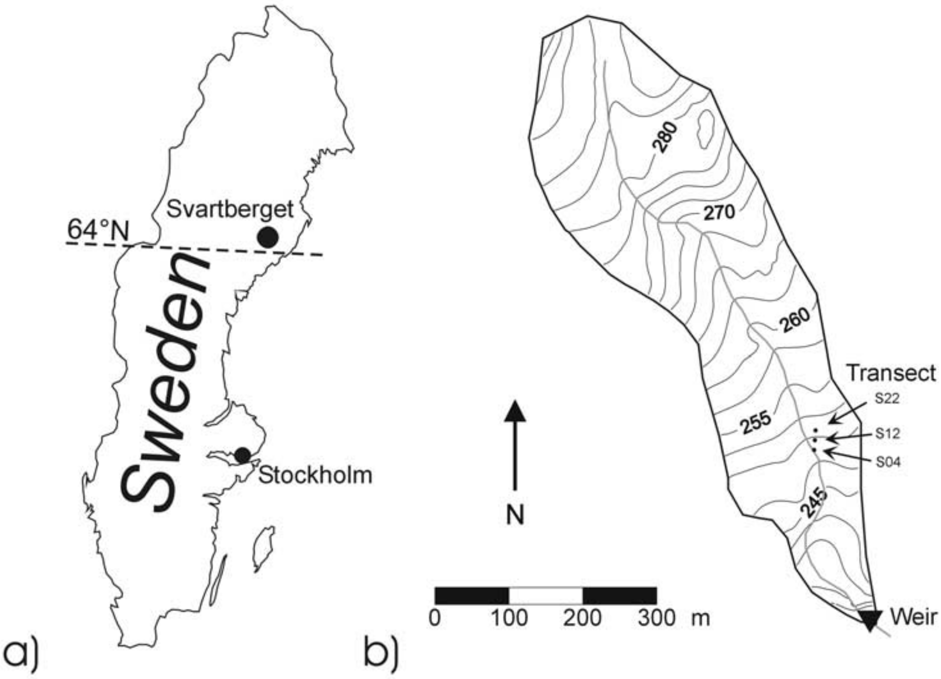

Figure 1. Map of the field site.

and groundwater levels) with isotopic information from stream water and snowmelt as well as soil water and groundwater in a transect following the topography toward the stream. The purpose of this study was to investigate: (1) Do isotopic and hydrometric results agree? (2) Can the transmissivity feedback mechanism explain the rapid transfer of preevent water during the snowmelt period when soils are frozen? and (3) Does soil frost significantly affect the hydrological pathways of water?

[5] In this study we focused on spring snowmelt. This is the dominant hydrological event in many high latitude regions of the world and thus is of major concern for water quality issues in watercourses, lakes and coastal areas. Because of the difficulty in sampling during winter conditions much of the existing knowledge about dominant flow paths and runoff mechanisms are based on stream-based studies or hydrometric methods alone [*Maulè and Stein*, 1990; *Shanley and Chalmers*, 1999; *Nyberg et al.*, 2001].

[6] Many studies have tried to identify the mechanism of the rapid preevent water delivery during hydrological episodes (see *Buttle* [1994] for a review). The hypothesis we tested for the till soils of northern Sweden is based on the transmissivity feedback mechanism [*Bishop*, 1991]. This term refers to a condition commonly observed in glacial till soils, where the lateral saturated hydraulic conductivity increases toward the soil surface. As a result the transmissivity and consequently the rate of lateral water movement increase dramatically as the groundwater level rises into the superficial soil layers.

2. Study Location

[7] The study was carried out during the 1999 snowmelt period on a 13 ha catchment at the Svartberget Research Station (64°14′N, 10°46′E) approximately 50 km inland from Umeå and the Baltic Sea coast (Figure 1). The studied stream, Västrabäcken, drains a forested headwater catchment typical for the boreal region of northern Sweden. The elevations range from 235 to 310 m above sea level. The channel was straightened and deepened in the 1920s, which was a common practice in northern Sweden to improve drainage and thereby forest productivity. A locally derived glacial till with a thickness of up to 10 or 15 m overlays gneissic bedrock. Soils are predominately well developed iron podzols, with organic-rich soils that have an average depth of 0.45 m in the riparian zone within 10 m of the stream channel [*Bishop et al.*, 1994].

[8] The mean annual precipitation at the Svartberget research station (1981–1999) is close to 600 mm, of which approximately 35% falls as snow [*Löfvenius et al.*, 2003]. Detailed descriptions of the physiographical, hydrological and hydrochemical characteristics are provided by *Bishop et al.* [1990] and *Laudon et al.* [1999].

3. Methods

[9] Discharge was computed on an hourly basis from water level measurements (using a pressure transducer connected to a Campbell Scientific data logger) at a thin plate, 90°V notch weir at the outlet of the catchment. The rating curve was derived using manual, instantaneous discharge measurements (39 measurements in the range from 0.1 to 9 mm day^{-1}). The stream sampling program was based on weekly samples of base flow prior to the onset of the snowmelt, and then every second day during the spring until the flow receded to levels close to those of base flow.

[10] Three 1.5 m^2, acid washed, Teflon-coated snow lysimeters located in the closed canopy spruce forest were used for measuring snowmelt volumes, melt intensity and isotopic composition. Snowmelt water was sampled daily to twice daily and then bulked into three to four day intervals

for $\delta^{18}O$ analyses. Data were linearly interpolated to an hourly basis.

[11] For the soil measurements three soil profiles were located along a 25 m transect. The transect was aligned based on the topography to follow the assumed lateral flow paths of the groundwater toward the stream. The riparian soil profile closest to the stream, S04, was dominated by organic material with a transition from organic to organic-rich mineral soil at 30 cm depth. The organic enrichment continues to a depth of 60 cm. The upslope site, S22, was located in a typical podzolic soil with a 10 to 15 cm organic layer overlying the mineral soil. Site S12 was between the riparian and the upslope site in an organic rich mineral soil with a transition from organic to organic-rich mineral soil at 20 cm depth. The organic enrichment continues to a depth of 50 cm. Details about organic content and grain size distribution of the soil profiles are provided by *Nyberg et al.* [2001].

[12] In the three soil profiles, soil water content and soil temperatures were recorded at four-hour intervals using time domain reflectrometry (TDR) and thermistors. The recorded data was stored in a Campbell scientific data logger. The TDR probes (length 30 cm) were installed horizontally at six levels between 5 cm and 90 cm soil depth. For details about the TDR system, see *Nyberg et al.* [2001].

[13] Soil core samples were collected on 23 March, well before any snowmelt had occurred from depths of 25 cm, 45 cm and 65 cm at S04 and S22 as well as on one occasion during peak flow at S22 on 26 April. The soil cores were collected approximately 3 m from other soil installations because of the destructive sampling. The soil core samples were centrifuged at 14000 rpm using a Beckman J/E 21 centrifuge. In addition, soil water solution was sampled using suction lysimeters after snowmelt was over on 27 May. The lysimeters collected water from seven horizons in each soil profile between 5 and 90 cm soil depth. The soil water samples from both centrifuged soil cores and the suction lysimeters were analyzed for $\delta^{18}O$.

[14] Groundwater levels were measured manually (as saturated water level below soil surface). Groundwater samples for $\delta^{18}O$ were collected each third to fifth day at S04, S12 and S22 in shallow groundwater wells. The shallow groundwater wells where perforated at its full length and extending approximately 1 m below the soil surface. Samples for $\delta^{18}O$ were collected by first emptying water in the well and then collecting the water that refilled the well. Deep groundwater samples were collected prior to snowmelt from groundwater wells on a small esker 100 m outside the catchment. These deeper groundwater wells were open only at the bottom, which sampled at depths of 3 to 5 m below ground surface (and 1–2 m below the groundwater level). The sampling followed the same procedure as for the shallower wells. The deep groundwater wells were sampled on three occasions: before (23 March), during (11 May) and after (11 June) snowmelt.

[15] Stream, melt, soil and groundwater were analyzed for $\delta^{18}O$ at the Department of Forest Ecology, SLU Umeå, Sweden using a IRMS with a TG preparation unit (20-20 Stable Isotope Analyser, Europa Scientific Ltd, Crewe UK). All $\delta^{18}O$ values are expressed relative to Vienna-standard mean ocean water (VSMOW). All samples for isotopic analyses were collected in 25 mL, narrow-mouth glass bottles. The samples were stored cold and dark until analyzed.

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

$$f_p = \frac{\delta^{18}O_s - \delta^{18}O_e}{\delta^{18}O_p - \delta^{18}O_e} \qquad (1)$$

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

[17] The $\delta^{18}O_e$ was defined using the runCE method proposed by *Laudon et al.* [2002] (equation (2)). The runCE method accounts for both the timing and the amount of meltwater 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 snowmelt (and rainwater contributions) from snow lysimeters and the cumulative volume (depth) of meltwater that has left the snowpack but has not yet discharged to the stream during the event. This way the lag between the melting of snow and its arrival at the stream is taken into account [*Laudon et al.*, 2002].

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

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

[19] Upscaling from the soil transect to the catchment was based on defining the catchment as a rectangle with a length of approximately 750 m along the stream and with 80 m of catchment area on either side of the stream. The soil profiles S04, S12 and S22 characterized 10%, 10% and 80% of the catchment, respectively. That corresponds to S04 representing 8 m on either side of the stream, S12 representing the next 8-m bands, and S22 the remaining 64 m to the edge of the catchment.

[20] The change in catchment water storage during snowmelt was estimated using two different approaches. The first was subtracting the total spring runoff from the total snowmelt, with evapotranspiration during the snowmelt assumed to be negligible. The other estimate was based on the TDR data for the total water content in the S04, S12, and S22 profiles prior to the spring (1 April) and after the termination of snowmelt (1 June).

[21] The amount of event water remaining in the soil after the termination of snowmelt has been calculated using three different methods. In the first method the total snowmelt was compared to the amount of event water that left the soil

in the stream during the spring as estimated from equation (2). The other two calculations were based on isotopic data from the soil before and after spring snowmelt using isotope data from the shallow groundwater wells and the suction lysimeters respectively.

[22] Event water calculation of soil water composition using shallow groundwater tubes was based on equations (1) and (2), but with (1) the average presnowmelt groundwater isotopic composition ($\delta^{18}O_g$) used in place of the preevent ^{18}O ($\delta^{18}O_p$), and (2) the isotopic composition from the shallow groundwater tubes ($\delta^{18}O_{gt}$) sampled at the time of the separation substituted for the stream water composition ($\delta^{18}O_s$). The event water calculation using suction lysimeters was carried out similarly to the shallow groundwater wells with the difference that the stream water composition ($\delta^{18}O_s$) was exchanged with the isotopic composition from the suction lysimeters ($\delta^{18}O_{sl}$) at the time of separation.

[23] Quantification of water routed as overland flow was based on the difference between the fraction of event water in S04 and the fraction of event water in the stream at the time of separation.

4. Uncertainty Analyses

[24] We performed an uncertainty analysis to quantify the effects of different sources of errors on our results. The standard error of the $\delta^{18}O$ analysis as estimated by the isotope laboratory is 0.15‰. However, in order to include any uncertainty associated with sampling, storage and handling a standard error of 0.2‰ was used in the uncertainty analyses.

[25] 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. The uncertainty in the calculation of the preevent fraction of water in the soil, after snowmelt was over, using both suction lysimeters and shallow groundwater tubes, was carried out the same way as for the stream water separation.

[26] The uncertainty in the total runoff was calculated to be approximately 2% by comparing the manual, instantaneous discharge measurements and the data logger based runoff measurements. A standard deviation of 5% in the specific discharge was however used in order to add possible errors in the catchment area calculation.

[27] A 5% uncertainty expressed as standard deviation in the TDR measurement was assumed to include any errors in those measurements (Lars Nyberg, personal communication) and in the volume estimates of water in the soil profiles. The standard deviation in the measure of total snowmelt was calculated to be 11 mm or approximately 5%. A standard deviation of 10% was however used in the calculations in order to include unaccounted errors in the spatial variability of snowmelt.

[28] The uncertainties introduced in the upscaling from S04, S12 and S22 to the entire catchment were estimated using Monte Carlo simulations with two concurrent uncertainty sources. The first uncertainty sources were the soil-water-volume estimates using the TDR data or the event water calculation for each profile. The estimated standard deviation in the TDR measurements or event water separation in the soil was used to create a normal distribution of the associated errors. The second uncertainty source was the proportions of the catchment subareas that were characterized by S04, S12 and S22. The proportions were varied following a normal distribution using a standard deviation of 5% of the area represented by S04 and S12, as well as a standard deviation of 7% for the area represented by S22 (with the constraint that the sum was 100%). The combined uncertainty for the storage calculations was finally estimated using 10,000 realizations with random parameters generated from the distributions discussed above and calculating the standard deviation (σ) of the 10,000 storage estimates.

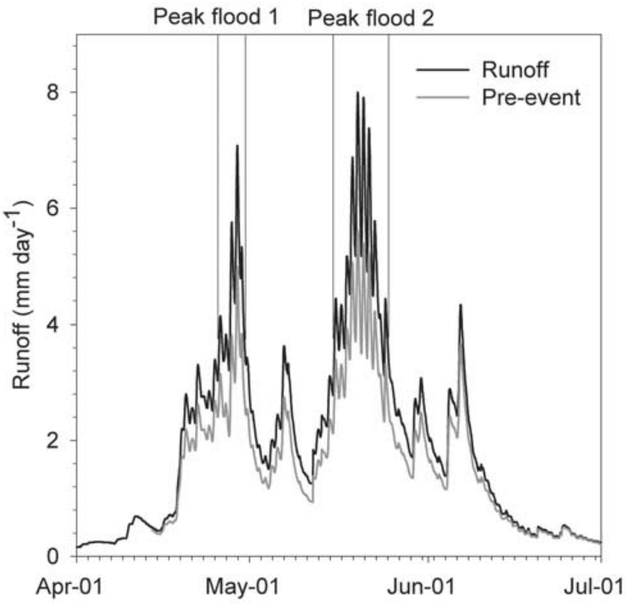

Figure 2. Separation of event and preevent water in Västrabäcken. “Peak flow 1” and “Peak flow 2” are marked for reference in the text.

5. Results

[29] Snowmelt started in the middle of April, following more than 5 months of permanent snow cover. The snowmelt derived hydrograph consisted of two major peak flow events, 29 April (peak flow 1) and 19 May (peak flow 2) (Figure 2). Peak flows 1 and 2 occurred when 26% and 77% of all snow had melted, respectively. The total runoff during the spring was 134 mm (σ = 7 mm). The total snowmelt was 198 mm (σ = 20 mm) (Table 1). Neglecting evaporation, this gives an increase in the subsurface water storage of 64 mm (σ = 21 mm).

[30] Unsaturated soil water ($\delta^{18}O$ −13.01, σ = 0.18) and shallow groundwater ($\delta^{18}O$ −13.10, σ = 0.05) had almost identical $\delta^{18}O$ signatures prior to snowmelt whereas the deep groundwater $\delta^{18}O$ was somewhat lighter ($\delta^{18}O$ −13.73, σ = 0.07) (Figure 3). The $\delta^{18}O$ (−13.31, σ = 0.10) of base flow (0.11 to 0.17 mm d^{-1}) was in between the unsaturated soil water/shallow groundwater and deep groundwater suggesting that the stream prior to snowmelt consisted of water from both of these sources (Figure 3). At the onset of snowmelt, concurrent with a discharge increase from 0.13 to 0.50 mm d^{-1} the $\delta^{18}O$ of the stream became similar to that of the soil water and shallow groundwater. As the rate of snowmelt increased and the runoff rose, the stream $\delta^{18}O$ became lighter due to the influx of snowmelt water.

Table 1. Snowmelt Runoff Results[a]

Stream and Snow Based Calculations	Value			
Total runoff[b]	134 mm (σ = 7 mm)			
Total snowmelt[c]	198 mm (σ = 11 mm)			
Preevent water in stream[d]	100 mm (σ = 9 mm)			
Event water in stream[d]	34 mm (σ = 3 mm)			
	S04	S12	S22	Catchment
	Soil Water Storage Change			
TDR based change[e]	29 mm (σ = 14 mm)	33 mm (σ = 12mm)	62 mm (σ = 13 mm)	56 mm (σ = 15 mm)
Mass balance based change[f]				64 mm (σ = 21 mm)
	Event Water in Soil After Spring			
Shallow groundwater tubes[g]	85 mm (σ = 7 mm)	92 mm (σ = 8 mm)	165 mm (σ = 14 mm)	150 mm (σ = 13 mm)
Suction lysimeters[h]	94 mm (σ = 8 mm)	63 mm (σ = 5 mm)	178 mm (σ = 15 mm)	158 mm (σ = 14 mm)
Mass balance based[i]				164 mm (σ = 20 mm)

[a]Here σ is standard deviation as estimation of error.
[b]Measured at the weir.
[c]Average of measurements from three snowmelt lysimeters.
[d]Preevent and event water in the stream is derived using equations (1) and (2).
[e]Calculation based on water content measured by TDR at several depths in the upper 90 cm of soil.
[f]Difference of snowmelt and total runoff (assuming that evapotranspiration is negligible).
[g]Event water calculation using shallow groundwater tubes was based on equations (1) and (2) substituting preevent ^{18}O ($\delta^{18}O_p$) with an average presnowmelt groundwater isotopic composition ($\delta^{18}O_g$) as well as substituting the stream water composition ($\delta^{18}O_s$) with isotopic composition from the shallow groundwater tubes ($\delta^{18}O_{gt}$).
[h]Event water calculation using suction lysimeters was also based on equations (1) and (2) by substituting preevent ^{18}O ($\delta^{18}O_p$) with an average presnowmelt groundwater isotopic composition ($\delta^{18}O_g$) as well as substituting the stream water composition ($\delta^{18}O_s$) with isotopic composition from the suction lysimeters ($\delta^{18}O_{sl}$).
[i]Difference of snowmelt and event water in the stream.

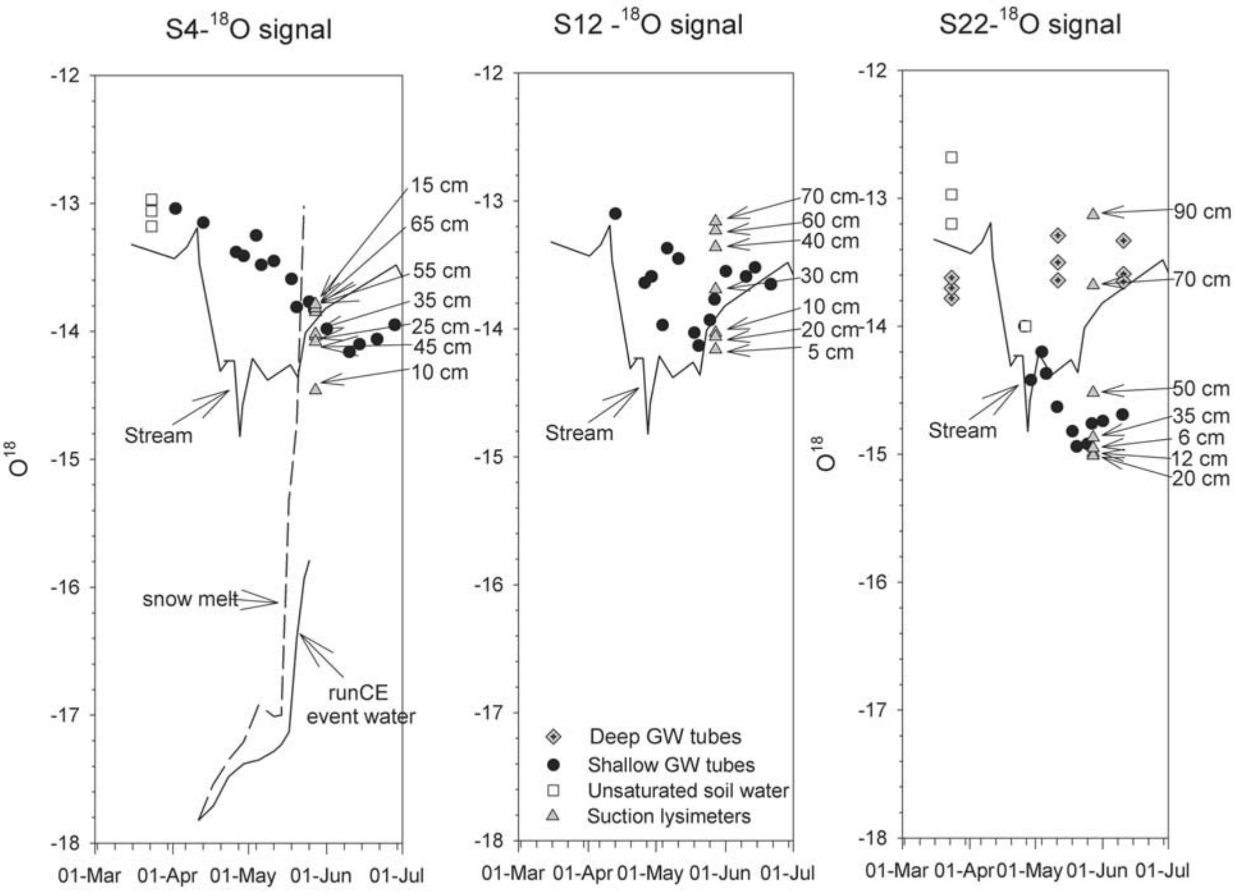

Figure 3. The δ^{18}O data in the three soil profiles. The stream water δ^{18}O is shown in all three profiles for reference. Note that the deep groundwater data shown in S22 actually is sampled outside the catchment. No soil water from the unsaturated zone was collected at S12.

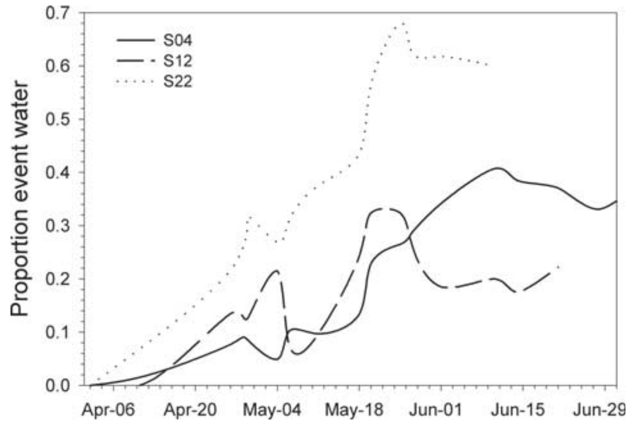

Figure 4. The fraction of event water in each soil profile. The calculations are based on shallow groundwater wells.

[31] There was a large difference between the measured snowmelt $\delta^{18}O$ and the runCE $\delta^{18}O$ signatures (Figure 3). The runCE based event water accounts for both the temporal change in the $\delta^{18}O$ of the snowmelt and the temporary storage of meltwater in the catchment [*Laudon et al.*, 2002].

[32] The preevent fraction of the entire snowmelt period was 75% (σ = 8%). The preevent fractions during peak flows 1 and 2 were 69% (σ = 6%) and 73% (σ = 6%), respectively. The fraction of event water in the soil (as calculated from the shallow groundwater wells; Figure 3) varied both spatially and temporally (Figure 4). The event water fraction in the soil increased relatively systematically from the initiation of snowmelt to the termination. At S04 the event water fraction increased for over two weeks after snowmelt had terminated. The event water fraction at S22 was in general two to three times as large as that in S04 and S12. During peak flow 1, the event water fraction in S04 was approximately 9% (σ = 5%). During peak flow 2 the event water fraction in S04 was 23% (σ = 5%). At S22 the event water fraction during peak flow 1 and 2 was 30% and 55% respectively.

[33] Suction lysimeters sampled after the termination of snowmelt indicated that soil horizons at 70 cm and deeper in S22 and S12 were not affected by snowmelt water as $\delta^{18}O$ values were still similar to premelt soil water values after the snowmelt was over (Figure 3). This suggested that a majority of all lateral transport of event water occurs above 70 cm. Lysimeter water from both S04 and the more superficial soil horizons in S12 and S22 showed isotopic signals similar to those of the shallow groundwater sampled on the same occasion. The similarity of the water from the groundwater wells and the suction lysimeters in the saturated zone indicate that the suction lysimeters sampled mainly mobile water (i.e., water which flowed into the groundwater wells when no suction was applied).

[34] The water content of the upper soil horizons varied considerably during the snowmelt period (Figure 5). Maximum water content in the three soil profiles coincided with peak flow 2. During peak flow 1 the water content was well

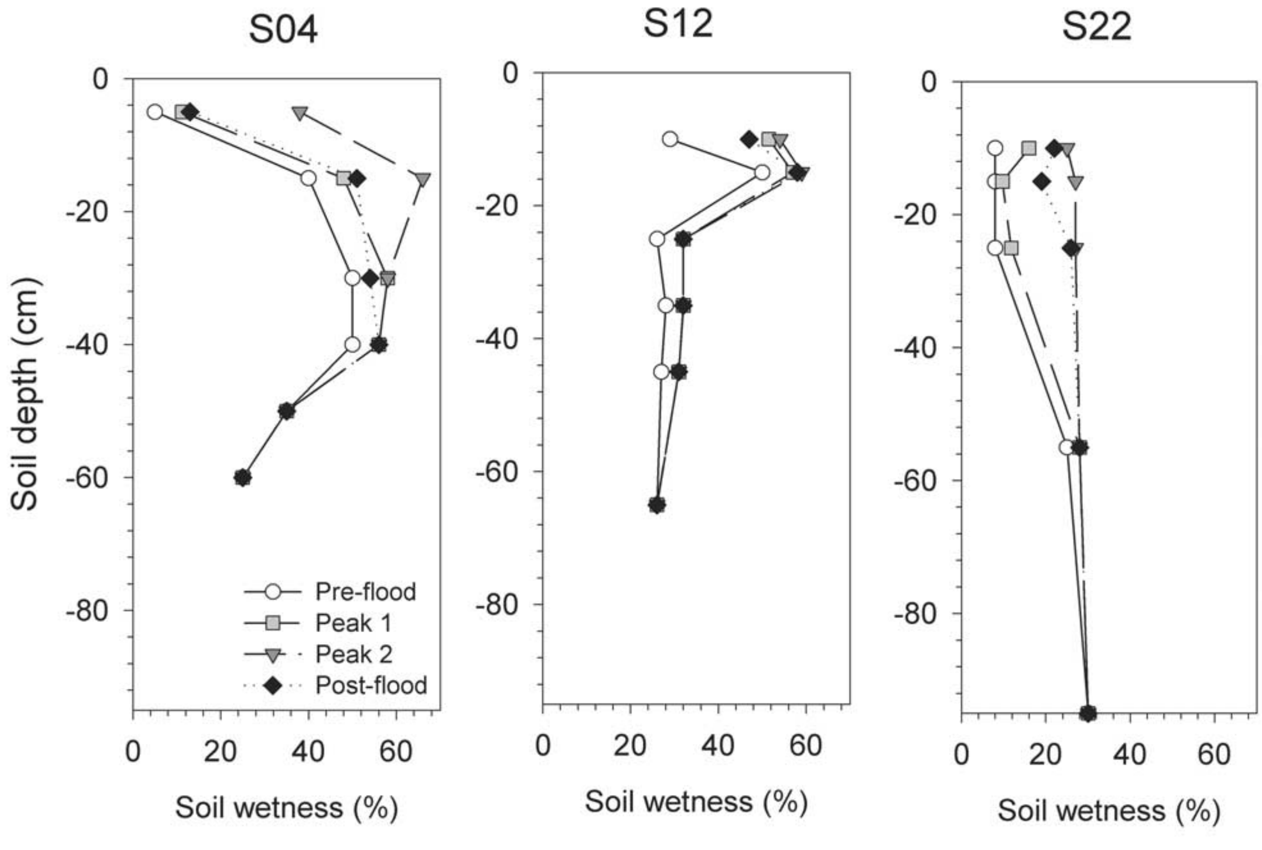

Figure 5. Soil water content in the three soil profiles at different times during the snowmelt period.

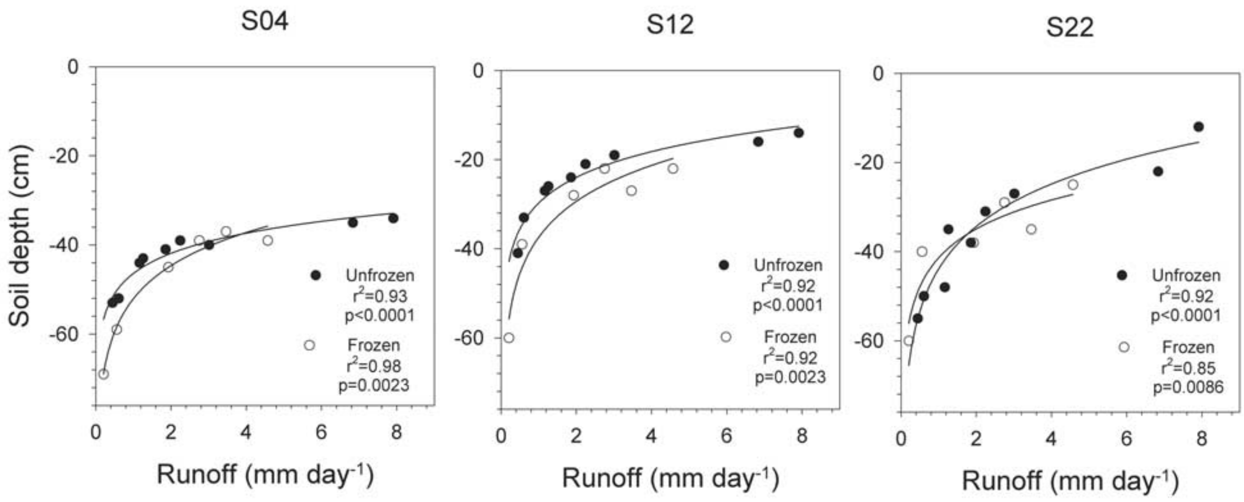

Figure 6. Relationships between runoff and groundwater levels along the soil transect. There was a statistically significant difference between the relationships based on data from periods with and without soil frost for all three profiles.

below the values for peak flow 2 although the runoff was at approximately the same levels. This was especially evident in the most superficial soil horizons in S04 and S22 (Figure 5). Water content in the lower soil profiles, below 40 to 50 cm soil depth, remained relatively constant during the entire study period near the porosity of the soils, suggesting almost saturated conditions during the study period.

[35] The soil water content above 90 cm prior to any snowmelt in the catchment was estimated to be 193 mm (σ = 8 mm), using the TDR measurements and the spatial weighting discussed above. The soil water content after the entire snowmelt had occurred was estimated to be 249 mm (σ = 10 mm). The storage change during the spring estimated from these measurements was 56 mm (σ = 15 mm) (Table 1).

[36] Soil temperature in the three soil profiles remained above 0°C during the entire winter at soil depths below 25 cm. The soil was frozen (temperatures <0°C) at 5 and 10 cm soil depth to around 10 May in S04 and at 5 and 12 cm soil depth in S12 to 15 May. At S22 the soil remained frozen at 11 and 17 cm until 10 and 17 May, respectively.

[37] The groundwater levels at all three soil profiles were strongly correlated to catchment runoff (Figure 6). There was however a marked difference in the relationship for the observations from dates when superficial soils were frozen (temperatures <0°C) and unfrozen in the two soil profiles closest to the stream.

6. Discussion

[38] As in most other IHS investigations, the isotopic results from the snowmelt and stream water demonstrate a large preevent water fraction in the stream during the spring melt. While the skepticism that such results originally met has subsided, this study is the first to attempt (as far as we are aware) to close the catchment δ^{18}O mass balance.

[39] We found that the changes in the preevent reservoir implicit in IHS using runCE were consistent with the soil-water content observations (Table 1). The agreements in both water storage changes and the amount of event water, which was left in the catchment after the termination of the snowmelt runoff period, support the validity of the results. Spring melt is the most important recharge period in many boreal regions, and a correct quantification of storage changes can provide essential information for understanding the hydrology of these systems. Valuable as IHS is, it only provides an answer to the question of the relative contribution of two sources, event and preevent water, leaving many questions about the specific flow pathways and runoff mechanisms unanswered. Combining isotopic and hydrometric information, however, opens possibilities for defining mechanisms of runoff generation that would not be otherwise possible.

[40] The quantification of catchment storage and change in this storage over time also makes it possible to address the question of how preevent water can be mobilized so quickly. In the study transect, there was not only sufficient preevent water to support the 134 mm of snowmelt runoff, but this source of preevent water was confined to the upper 90 cm of the soil. This is evident from the fact that the infiltrating snowmelt water did not affect soil horizons at more than 90 cm soil depth, which suggested that the entire runoff event occurred in the upper meter of soil and groundwater from deeper soil layers was not activated. The soil isotopic data (Figure 3) also showed that the soil water (sampled with soil lysimeters at the termination of the snowmelt period) was isotopically most similar to the stream water in the soil horizons where most water is predicted to be transferred to the stream according to the transmissivity feedback concept. The deep soil water (below 70 cm) is similar in isotopic composition to the groundwater/soil water composition prior to snowmelt while the most superficial soil horizons are isotopically lighter than any groundwater/stream water and hence more directly affected by snowmelt.

[41] In a recent study, *McGlynn et al.* [1999] demonstrated that the riparian zone was fed by two sources of water. These distinct sources were deep riparian groundwater and upland water, which followed discrete flow paths inhibiting vertical mixing. Contrary to that work, our results suggest that the

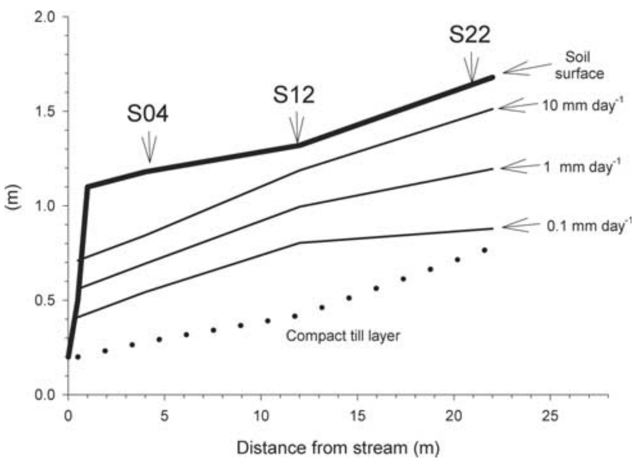

Figure 7. Schematic figure of the hillslope transect with groundwater levels associated with a given runoff.

water in the riparian soil, generating streamflow, is well mixed, compared to the upland soil profiles. This could be due to vertical dispersion as the water moves down slope toward the stream.

[42] The event water fraction of the soil (Figure 4) increased away from the stream during the snowmelt period. In the soil profile closest to the stream the preevent water fraction during both peak flows 1 and 2 is approximately one third of the preevent water fraction in S22. This could partly be explained by the lower water content of the soil further away from the stream (Figure 5), but could also be due to a delay in the lateral displacement of water from the upslope areas. The latter explanation is supported by the observation of declining event water content in S22 as soon as snowmelt is terminated, whereas the event water in S04 continues to increase for another two weeks.

[43] In a Canadian study, *Hinton et al.* [1994] showed that groundwater flow through glacial till made a significant contribution to streamflow despite lower hydraulic conductivity compared to more superficial soils. Although deep groundwater contributes a significant portion of water during base flow, the contribution during the peak flow is believed to be marginal due to low hydraulic conductivity of deep soil layers. This assumption is supported by the fact that neither the deep groundwater $\delta^{18}O$ nor the soil water $\delta^{18}O$ in deeper layers were affected by infiltrating snowmelt.

6.1. Evidence of Transmissivity Feedback

[44] One of the strongest sources of evidence for the transmissivity feedback mechanism is the strong correlation between groundwater levels and stream discharge. The exponential relationship between groundwater and discharge suggest that runoff is substantially more influenced by a given rise of the groundwater table during wet conditions compared to dry conditions. As an example, a 10 cm increase of the groundwater table at a base flow of 0.1 mm d^{-1} will result in an increase in runoff to 0.4 mm d^{-1}. The same groundwater level rise at an intermediate flow of 2 mm d^{-1} will increase the runoff to more than 9 mm d^{-1} (Figure 7).

[45] The transmissivity feedback mechanism is based on water flow through the soil matrix (as opposed to bypass flow). The assumption of matrix flow being, at least partly, responsible for the streamflow generation during the snowmelt period is consistent with three aspects of what has been discussed above; (1) preevent water was mobilized as meltwater was recharging the groundwater storage, (2) lateral flow of preevent water down slope was displaced in front of event water, and (3) a lack of clear layering of event and preevent water in the riparian soil.

[46] *Kendall et al.* [1999] found a hysteresis in the groundwater-discharge relationship and different directions of the hysteresis loops in riparian sites and hillslope sites further away from the stream. In contrast, our results showed no hysteresis as there was no difference in the groundwater-discharge relationship between the rising and falling limbs of the hydrograph. Instead different functional relationships were found for frozen and unfrozen soil conditions (Figure 6). In both of the soil profiles closest to the stream during frozen conditions, a lower groundwater level was associated with a given discharge than during unfrozen conditions. This pattern was not observed at S22. The difference in the functional groundwater-streamflow relationships indicates that the flow pathway in the riparian soil might be affected by the frozen soil.

6.2. Do Frozen Soils Create Overland Flow?

[47] The role of soil frost in runoff generation has long been a subject of discussion [*Dunne and Black*, 1971]. A number of studies, often conducted on arable field soils [*Kane and Stein*, 1983; *Thunholm et al.*, 1989; *Stähli et al.*, 1996], have shown that infiltration may be seriously reduced by soil frost, mainly due to the blocking effect of the ice. Two recent studies of forested catchments in Vermont [*Shanley and Chalmers*, 1999] and at the Nyänget catchment in northern Sweden [*Lindström et al.*, 2002] using long time series conclude that no clear connection between the extent of soil frost and the timing and magnitude of runoff during snowmelt could be found. Furthermore, hydrometric studies by *Nyberg et al.* [2001] could not find conclusive evidence of a soil frost effect on flow paths at Nyänget. In the work presented here, however, the hydrometric and isotopic results demonstrate that soil frost, can in fact, affect the flow paths of water during the early part of the snowmelt period, but that this influence decreases as the spring progresses.

[48] *Lindström et al.* [2002] suggest that one important reason for the soil frost not influencing the runoff is that the soils are often thawed before the start of snowmelt. This was not the case in this study where the soil temperatures remained below 0°C until mid May. The water content data confirmed that the soils remained frozen until between peak flow 1 and peak flow 2 (Figure 5). The more superficial soil horizons were less saturated during peak flow 1, especially in the riparian site, compared to peak flow 2 despite similarity in the maximum runoff. This suggests that the most superficial soil horizons were at least partly frozen during the early part of the spring.

[49] More direct evidence of overland flow comes from the different functional relationships of groundwater level and runoff in frozen and unfrozen soils (Figure 6). The lower groundwater table for a given discharge when the soils were frozen suggest that water found its way to the stream without connecting to the groundwater, and thus that meltwater could become Hortonian overland flow on top of a frozen soil layer. Furthermore, the event water fraction

in the groundwater at S04 during peak flow 1 was only 9%. This could not contribute the 31% event water in the stream during peak flow 1. This suggests that approximately 20% of the stream runoff was derived from snowmelt that did not connect with the groundwater during peak flow 1. During peak flow 2 the event water fraction in the groundwater at S04 was 23%, which is more similar to the 27% event water in the stream at that time. This translates to less than 3% of the snowmelt during peak flow 2 that reached the stream without first connecting to the groundwater.

[50] While none of these sources of evidence might be enough to argue satisfactorily for a contribution to flow via a bypass mechanism as opposed to the matrix flow of the transmissivity feedback mechanism, the convergence of three different types of evidence make a strong case for some Hortonian overland flow during peak flow 1. As a proportion of snowmelt runoff, this bypass amounts to 20 percent of the runoff during peak flow 1, and approximately 6 percent of the entire snowmelt runoff. While not dominant in volume, the existence of this complementary pathway can be of significance for getting some chemical constituents to the stream that would be removed along subsurface pathways.

7. Conclusion

[51] A mixture of possible runoff mechanisms is a factor that makes hydrology complex. IHS has been a valuable tool in understanding runoff generation in its own right. This study, though, has succeeded in using complementary hydrometric and isotopic tracer approaches on a 2D transect that is taken to represent the catchment. The congruence of the results confirmed the importance of the transmissivity feedback mechanism, and revealed that the depth of soil involved in mixing with snowmelt was less than 90 cm. The combined approach was also sufficiently precise to resolve the extent to which soil frost created an overland flow path for a portion of spring flood. While this finding comes from just one site, we argue that it is an example of what can be gained by combining different measurement approaches on a transect cross-sections. This provides more detailed process information than can be achieved by working at the catchment scale alone.

[52] **Acknowledgments.** We wish to thank Ulf Juto and Johan Temnerud for help with the field work. The financial support for this work was provided by the Swedish National Science Foundation as well as an Oscar and Lili Lamms Foundation stipend to the first author.

References

Bishop, K. H. (1991), Episodic increases in stream acidity, catchment flow pathways and hydrograph separation. Ph.D. thesis, 241 pp., Cambridge Univ., Cambridge, U.K.

Bishop, K. H., H. Grip, and A. O'Neill (1990), The origins of acid runoff in a hillslope during storm events, *J. Hydrol.*, *116*, 35–61.

Bishop, K., C. Pettersson, B. Allard, and Y. H. Lee (1994), Identification of the riparian sources of aquatic dissolved organic-carbon, *Environ. Int.*, *20*, 11–19.

Burns, D. A. (2002), Stormflow-hydrograph separation based on isotopes: The thrill is gone—What's next?, *Hydrol. Processes*, *16*, 1515–1517.

Buttle, J. M. (1994), Isotope hydrograph separations and rapid delivery of pre-event water from drainage basins, *Prog. Phys. Geogr.*, *18*, 16–41.

Dewalle, D. R., B. R. Swistock, and W. E. Sharpe (1988), 3-component tracer model for stormflow on a small Appalachian forested catchment, *J. Hydrol.*, *104*, 301–310.

Dinçer, T., B. R. Payne, T. Florkowski, J. Martinec, and E. Tongiorgi (1970), Snowmelt runoff from measurements of tritium and oxygen-18, *Water Resour. Res.*, *6*, 110–124.

Dunne, T., and R. D. Black (1971), Runoff process during snow melt, *Water Resour. Res.*, *7*, 1160–1172.

Hinton, M. J. S., S. L. Schiff, and M. C. English (1994), Examining the contributions of glacial till water to storm runoff using two-component and three-component hydrograph separations, *Water Resour. Res.*, *30*, 983–993.

Kane, D. L., and J. Stein (1983), Water movement into seasonally frozen soils, *Water Resour. Res.*, *19*, 1547–1557.

Kendall, C., and J. J. McDonnell (1998), *Isotope Tracers in Catchment Hydrology*, Elsevier Sci., New York.

Kendall, C., J. J. McDonnell, and W. Z. Gu (2001), A look inside "black box" hydrograph separation models: A study at the Hydrohill catchment, *Hydrol. Processes*, *15*, 1877–1902.

Kendall, K. A., J. B. Shanley, and J. J. McDonnell (1999), A hydrometric and geochemical approach to test the transmissivity feedback hypothesis during snowmelt, *J. Hydrol.*, *219*, 188–205.

Lakey, B., and N. C. Krothe (1996), Stable isotopic variation of storm discharge from a perennial karst spring, Indiana, *Water Resour. Res.*, *32*, 721–731.

Laudon, H., and O. Slaymaker (1997), Hydrograph separation using stable isotopes, silica and electrical conductivity: An alpine example, *J. Hydrol.*, *201*, 82–101.

Laudon, H., S. Köhler, and K. H. Bishop (1999), Natural acidity or anthropogenic acidification in the spring flood of northern Sweden?, *Sci. Total Environ.*, *234*, 63–73.

Laudon, H., H. F. Hemond, R. Krouse, and K. H. Bishop (2002), Oxygen 18 fractionation during snowmelt: Implications for spring flood hydrograph separation, *Water Resour. Res.*, *38*(11), 1258, doi:10.1029/2002WR001510.

Lindström, G., K. Bishop, and M. O. Löfvenius (2002), Soil frost and runoff at Svartberget, northern Sweden—Measurements and model analysis, *Hydrol. Processes*, *16*, 3379–3392.

Löfvenius, M. O., M. Kluge, and T. Lundmark (2003), Snow and soil frost depth in two types of shelterwood and a clear-cut area, *Scand. J. For. Res.*, *18*, 54–63.

Maulè, C., and J. Stein (1990), Hydrologic flow path definition and partitioning of spring meltwater, *Water Resour. Res.*, *26*, 2959–2970.

McDonnell, J. J. (1990), A rationale for old water discharge through macropores in a steep, humid catchment, *Water Resour. Res.*, *26*, 2821–2832.

McGlynn, B. L., J. J. McDonnell, J. B. Shanley, and C. Kendall (1999), Riparian zone flowpath dynamics during snowmelt in a small headwater catchment, *J. Hydrol.*, *222*, 75–92.

Nyberg, L., M. Stähli, P. E. Mellander, and K. H. Bishop (2001), Soil frost effects on soil water and runoff dynamics along a boreal forest transect: 1. Field investigations, *Hydrol. Processes*, *15*, 909–926.

Ogunkoya, O. O., and A. Jenkins (1993), Analysis of storm hydrograph and flow pathways using a 3-component hydrograph separation model, *J. Hydrol.*, *142*, 71–88.

Pionke, H. B., and D. R. Dewalle (1992), Intrastorm and interstorm O-18 trends for selected rainstorms in Pennsylvania, *J. Hydrol.*, *138*, 131–143.

Rodhe, A. (1998), Snowmelt-dominated systems, in *Isotope Tracers in Catchment Hydrology*, edited by C. Kendall and J. J. McDonnell, pp. 391–433, Elsevier Sci., New York.

Shanley, J. B., and A. Chalmers (1999), The effect of frozen soil on snowmelt runoff at Sleepers River, Vermont, *Hydrol. Processes*, *13*, 1843–1857.

Sklash, M. G., M. K. Stewart, and A. J. Pearce (1986), Storm runoff generation in humid headwater catchments: 2. A case-study of hillslope and low-order stream response, *Water Resour. Res.*, *22*, 1273–1282.

Stähli, M., P. E. Jansson, and L. C. Lundin (1996), Preferential water flow in a frozen soil—A two-domain model approach, *Hydrol. Processes*, *10*, 1305–1316.

Thunholm, B., L.-C. Lundin, and S. Lindell (1989), Infiltration into a frozen heavy clay soil, *Nord. Hydrol.*, *20*, 153–166.

Uhlenbrook, S., M. Frey, C. Leibundgut, and P. Maloszewski (2002), Hydrograph separations in a mesoscale mountainous basin at event and seasonal timescales, *Water Resour. Res.*, *38*(6), 1096, doi:10.1029/2001WR000938.

K. Bishop and J. Seibert, Department of Environmental Assessment, Swedish University of Agricultural Sciences, Box 7050, SE-750 07 Uppsala, Sweden. (kevin.bishop@ma.slu.se; jan.seibert@ma.slu.se)

S. Köhler, Géochimie, Transferts et Mécanismes, CNRS, UMR 5563, University of Toulouse, 38 rue des trente-six ponts, F-31400 Toulouse Cedex, France. (skohler@lmtg.ups-tlse.fr)

H. Laudon, Department of Forest Ecology, Swedish University of Agricultural Sciences, 901 83 Umeå, Sweden. (hjalmar.laudon@sek.slu.se)