Runoff production in a forested, shallow soil, Canadian Shield basin

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Abstract. Storm flow in forested basins on the Canadian Shield is largely supplied by subsurface water; however, mechanisms by which this water reaches the stream remain unclear. Side slope contributions to storm flow were studied using throughflow trenches on slopes in a headwater basin near Dorset, Ontario. Discharge, soil water content, and chemical and isotopic signatures of subsurface water were monitored at each site. Four hypotheses were tested: (1) most flow occurs at the soil-bedrock interface on side slopes with thin soil; (2) a significant fraction of event water moves vertically to bedrock via preferential flow pathways and laterally over the bedrock surface; (3) relative preevent water contribution to subsurface flow on shield slopes is a function of soil thickness; and (4) a significant portion of event water flux in storm flow from forested basins with shallow soil cover is supplied from side slopes via subsurface flow along the soil-bedrock interface. Hypothesis 1 was confirmed from hydrologic observations during spring and fall rainstorms. Hypotheses 2 and 3 were supported by temporal trends in dissolved organic carbon and 18O in flow at the soil-bedrock interface and by isotopic hydrograph separations (IHSs) of hillslope runoff. Comparison with the streamflow IHS indicated that event water flux from the basin in excess of that attributable to direct precipitation onto near-channel saturated areas could be supplied by flow along the bedrock surface (hypothesis 4). Flow at the soil-bedrock interface on side slopes also contributed ~25% of preevent water flux from the basin. Much of the event water component of basin storm flow may travel considerable distances via subsurface routes and is not necessarily contributed by surface runoff processes ( Horton flow or saturation overland flow). Therefore the assumption that event water undergoes little interaction with the soil during its passage downslope may be unwarranted here.

Introduction

Hydrometric, isotopic and geochemical studies in forested basins on the Canadian Shield [e.g., Bottomley et al., 1984; Wels et al., 1991a, b; Hinton et al., 1994] have revealed that storm flow is supplied largely by preevent water (water stored in the basin prior to rainfall or snowmelt) moving via subsurface routes to the channel. Despite this apparent consensus the mechanisms by which this water reaches the stream remain unclear.

One concept which may account for rapid subsurface transfer of preevent water to stream channels is the groundwater ridging mechanism [Sklash and Farvolden, 1979]. This process is limited to the near-channel area and results from rapid conversion of the tension saturated zone (capillary fringe) to phreatic water by infiltrating event water. The subsequent rise in the near-stream water table increases the hydraulic gradient to the stream and/or the size of seepage faces, resulting in pronounced groundwater discharge to the channel. This rapid rise in the water table has been noted in laboratory [Abdul and Gillham, 1984] and field investigations [Novakowski and Gillham, 1988; Abdul and Gillham, 1989] and is supported by numerical simulation [Sklash and Farvolden, 1979]. However, studies on the Canadian Shield have questioned the ability of groundwater ridging to serve as the dominant means of supplying storm flow to the stream during snowmelt. Thus Buttle and Sami [1992] found that groundwater ridging did not explain storm flow production in a small wetland basin, while Wels et al. [1991b] postulated that storm runoff delivery to the stream in a shallow soil basin was via widespread preferential flow at the soil-bedrock interface on the channel's side slopes.

Our aim was to investigate hillslope subsurface flow processes and their contributions to storm flow in a headwater basin using an integrated hydrodynamic, isotopic, and geochemical tracing approach. Four hypotheses were tested: (1) on shield slopes with thin soil cover, most flow occurs at the soil-bedrock interface; (2) a significant fraction of event water (water supplied to the basin during rainfall or snowmelt) moves via preferential flow pathways vertically to the bedrock surface and laterally along the soil-bedrock interface; (3) the relative preevent water contribution to subsurface flow on
its, especially on the west slope, the mineral (B) horizon has not developed and bedrock is covered by an A horizon and/or a humus layer ("runker") [Wels et al., 1991b]. Vertical and horizontal saturated hydraulic conductivities for the podzols are high, decreasing from $2 \times 10^{-4}$ (vertical) and $2 \times 10^{-3}$ (horizontal) m s$^{-1}$ near the soil surface to $3 \times 10^{-5}$ m s$^{-1}$ vertical and horizontal above the bedrock [Peters, 1994]. The stream valley contains deeper (>1 m) orthic gleysols [Lozano et al., 1987]. Vegetation is dominated by coniferous species (white pine, eastern hemlock, and white cedar). Deciduous species are not as common and consist largely of white birch, red oak, and striped maple [Lozano and Parton, 1987].

Materials and Methods

Hydrology

A tipping bucket gauge recorded rainfall at 10-min intervals in an open area adjacent to site 2 (Figure 1). Intercception was not monitored; however, Renzetti et al. [1992] found that throughfall in PC1-08 was 90% of open area totals. Therefore open area measurements may overestimate input to the soil surface in vegetated areas. Streamflow was gauged using a 90° V notch weir equipped with a water level recorder.

Deep and shallow trenches were excavated to bedrock in order to monitor subsurface flow at each of two sites considered representative of the basin side slopes (Figure 1). Site 1 was installed outside of PC1-08 so as not to interfere with existing OMEE instrumentation. Characteristics of the microcatchments drained by each trench were determined from surveys of surface and bedrock topography (Figures 2 and 3, Table 1). Bedrock topography was mapped by measuring the soil depth at each surveyed surface elevation point.

Flow at sites with relatively deep soil (trenches 1A and 2A) was monitored at three levels (Figure 4). Stainless steel troughs, inserted ~0.15 m into the soil face, collected flow through and/or over the organic-Ae horizon (Ae) and from an intermediate level (IN1) within the B horizon ~0.10 m above the bedrock surface. The INT trough was installed to monitor flow from the B horizon separate from runoff through a saturated layer over the bedrock surface [Renzetti et al., 1992]. Flow at the soil-bedrock interface (BR) was collected by a polyvinyl trough sealed to the bedrock with silicone. Soil cover at shallow soil sites (trenches 1B and 2B) consisted of an organic layer ~0.1 m thick, and a polyvinyl trough at the bedrock surface collected flow through and over this layer.

Flows were piped to measuring devices. Flow at shallow trenches and at the Ae and INT levels of deep trenches was measured using tipping bucket flowmeters. Flow over the bedrock surface at the deep trenches was measured using 4° V notch weir boxes, equipped with potentiometric water level recorders. Signals from tipping buckets and potentiometers were recorded by Campbell CR21X data loggers and converted to discharge using field and laboratory calibrations. Trenches and flow gauges were covered to prevent entry of precipitation and debris.

Solute and Isotopic Sampling

Bulk samplers beside the rain gauge collected rainfall for chemical and isotopic analyses. PC1-08 streamflow was sampled using an automatic water sampler. Grab samples were taken from pipe inlets to the weir boxes, and bulk flow samples from the tipping buckets were obtained using polyvinyl pails. Two porous cup lysimeter nests sampled soil water in the

Study Site

The study was conducted in Plastic Lake basin 1-08 (PC1-08) located on the Canadian Shield 20 km south of Dorset, Ontario, Canada (45°11′N, 78°50′W) (Figure 1). Basin hydrochemistry has been monitored through the Acid Precipitation In Ontario Study (APIOS) of the Ontario Ministry of the Environment and Energy (OMEE) since 1985. The area receives 1100 mm y$^{-1}$ of precipitation, 73% of which is rain (30-year mean [Environment Canada, 1982]). PC1-08 (3.22 ha) is drained by an ephemeral stream, with yearly high flows occurring during the spring freshet and fall rains. Basin morphology consists of two steep convex side slopes draining to a narrow valley bottom occupying a fault trough [Wels et al., 1991a].

The basin is underlain by granitic gneiss and amphibolite which occupy the topographic highs and lows, respectively [Kirkwood and Nesbit, 1991]. Bedrock outcrops are common and cover ~10% of the basin [Lozano et al., 1987]. Underlying bedrock is impermeable except where faults channel a small fraction of surface runoff underground [Shihabatani, 1988]. Weakly developed orthic humo-ferric and orthic ferro-humic podzols have formed on the thin (<1 m) sandy basal till mantling the east and west slopes, respectively. Soil thickness is highly variable, ranging from 0 to 1.5 m and averaging 0.29 m [Lozano et al., 1987]. In areas of thin or absent surficial deposi-
organic-Ae horizon, at the midpoint of the mineral (B) horizon, and −0.03 m above the bedrock surface upslope of the deep trenches (Figures 2 and 3). Lysimeters were maintained at a tension of −80 kPa, which was sufficient to sample micropore water [Luxmoore, 1981].

Samples were analyzed for dissolved organic carbon (DOC, in milligrams per liter) and oxygen 18 ($\delta^{18}O$). DOC was analyzed by the Laboratory Services Branch, OMEE, Dorset, by persulfate digestion preceded and followed by inorganic carbon analysis [OMEE, 1983]. Oxygen 18 contents were analyzed at the University of Waterloo Environmental Isotope Laboratory by mass spectrometry and reported as the relative deviation of the isotope ratio ($^{18}O/^{16}O$) from that of SMOW (standard mean ocean water). Analytical accuracy of $\delta^{18}O$ results was ±0.12‰.

A two-component mixing model was used to separate hillslope subsurface flow and storm flow into preevent and event water fractions:

$$Q_p = \left( \frac{C_r - C_s}{C_r - C_p} \right) \times Q_e$$

(1)

where $Q_p$ is streamflow; $Q_e$ is the contribution from preevent water (soil water/groundwater, base flow); $Q_r$ is contribution from event water (rainfall); and $C_r$, $C_p$, and $C_s$ are the corresponding $^{18}O$ values in streamflow, preevent, and event waters. $C_r$ was represented by bulk rainfall $^{18}O$, while isotopic hydrograph separations (IHSs) for hillslope subsurface flow and PCI-08 streamflow used prestorm soil water and base flow $^{18}O$ as $C_p$, respectively.

**Figure 2.** Instrumentation and bedrock topography for trench 1A and 1B microcatchments. Instrumentation codes are AT, neutron access tube; L, suction lysimeter nest.

**Figure 3.** Instrumentation and bedrock topography for trench 2A and 2B microcatchments. Instrumentation codes are A11, neutron access tube; L, suction lysimeter nest.

<table>
<thead>
<tr>
<th>Site</th>
<th>Drainage Area, $m^2$</th>
<th>Soil Depth, m</th>
<th>Bedrock Gradient, %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Trench 1A</td>
<td>307</td>
<td>0.43</td>
<td>0.43</td>
</tr>
<tr>
<td>Trench 1B</td>
<td>16</td>
<td>0.11</td>
<td>0.11</td>
</tr>
<tr>
<td>Trench 2A</td>
<td>822</td>
<td>0.31</td>
<td>0.31</td>
</tr>
<tr>
<td>Trench 2B</td>
<td>23</td>
<td>0.14</td>
<td>0.14</td>
</tr>
</tbody>
</table>
where \( r \) is radius of curvature of the interface (meters), \( \sigma \) is surface tension (7.27 x 10^{-2} J m^{-2} at 20°C), and \( \rho_w \) is the density of water (kilograms per cubic meter). The pore size distribution of the soil cores was determined by combining the results of (3) with the SWR curves.

Error Analysis

Fractional errors associated with runoff depth, soil water depth, preevent water depth and stream valley saturated area were approximated by the probable error associated with the sum or difference of the independent variables, or with their product or quotient [Davidson, 1978]. Table 2 gives the assumed fractional error for each component.

Results and Discussion

General Description of 1991 Study Periods

Twenty-nine rainfall events occurred over the periods from May 1 (year day (YD) 121) to June 21 (YD 172) 1991, and from September 22 (YD 265) to November 15 (YD 319) 1991, with depths and peak 10-min intensities ranging from 1.3 to 47.8 mm and 0.3 to 27.6 mm h^{-1}, respectively. Comparison with OMEE’s 6-year 24-hour rainfall frequency distribution shows that the events covered the range of rainfalls experienced in PCI-08 (Figure 5a). Most (74%) events for both distributions fell in the 0- to 10-mm range. Maximum 6-year 24-hour rainfall was 68 mm (OMEE unpublished data, 1991),

### Table 2. Fractional Errors Associated With Components Used in the Determination of Runoff Depth, Soil Water Depth, Preevent Water Depth, and Stream Valley Saturated Area

<table>
<thead>
<tr>
<th>Components</th>
<th>Fractional Error</th>
</tr>
</thead>
<tbody>
<tr>
<td>Runoff volume</td>
<td>0.05^a</td>
</tr>
<tr>
<td>Microcatchment area</td>
<td>0.1^b</td>
</tr>
<tr>
<td>Soil Water Depth</td>
<td></td>
</tr>
<tr>
<td>Volumetric water content</td>
<td>0.05^c</td>
</tr>
<tr>
<td>Soil depth</td>
<td>0.2^d</td>
</tr>
<tr>
<td>Microcatchment area</td>
<td>0.1^b</td>
</tr>
<tr>
<td>Representative area for a given access tube</td>
<td>0.05^b</td>
</tr>
<tr>
<td>Preevent Water Depth</td>
<td></td>
</tr>
<tr>
<td>Runoff depth</td>
<td>0.11</td>
</tr>
<tr>
<td>Preevent water fraction</td>
<td>0.04^c</td>
</tr>
<tr>
<td>Stream Valley Saturated Area</td>
<td></td>
</tr>
<tr>
<td>Area obtained from maps</td>
<td>0.05^b</td>
</tr>
<tr>
<td>Regression (equation 2)</td>
<td>0.1^b</td>
</tr>
</tbody>
</table>

^a From Winter [1981].

^bArbitrary.

^cEstimated from neutron probe calibrations, where fractional error ERR is given by \( ERR = \frac{n}{\sigma \cdot [1 - \frac{1}{n}]} \), where \( n \) is number of observations, \( \sigma \) is predicted volumetric water content, and obs is observed volumetric water content.

^dEstimated as \( z_{1/2} \) contour interval (0.1 m)/mean soil depth (0.5 m).

^eCumulative effect of error in \( 8^{18} \text{O} \) of \( C_s \), \( C_p \), and \( C_o \) (assumed to be 0.12%) which produced the greatest uncertainty in the preevent water fraction.

Estimation of Saturated Area in the Stream Valley

The extent of surface saturation in the PCI-08 stream valley prior to storm flow was estimated by the following regression equation:

\[
A_{sat} = 446.5 + 6010.4 \sqrt{Q} 
\]  

\[n = 4 \quad R^2 = 0.95
\]  

where \( A_{sat} \) (square meters) is the saturated area in the stream valley and \( Q \) (liters per second) is daily mean discharge at PCI-08. This relation was derived from Shihui [1988], where PCI-08 \( A_{sat} \) was estimated from his areal saturation maps of April 30, May 3, and May 5, 1985. \( A_{sat} \) was assumed to be 0 when there was no streamflow.

Soil Water Content

Seven neutron probe access tubes were installed upslope of the deep trenches (Figures 2 and 3). Volumetric soil water content was measured from 0.20 m below ground surface to the bedrock at 0.10-m intervals using a neutron probe, calibrated in the field using soil core samples. The area represented by the total soil water depth (millimeters) measured at each access tube was obtained using Thiessen polygons [Duane and Leopold, 1978], and an area-weighted soil water depth (millimeters) was determined for trench 1A and 2A microcatchments.

Soil Water Release Curve

Soil water release (SWR) curves for soils upslope of the deep trenches were obtained using a pressure plate apparatus. Lateral cores were taken at the midpoint of the B horizon at site 1 (0.25 m above bedrock), and 0.02 m above the bedrock surface at site 2. Equilibrium moisture contents were determined at matric potentials \( \Psi \) of 0, -2, -4, -6, -8, -10, -20, and -40 kPa. The equivalent radius of the largest water-filled pore was computed at the specified \( \Psi \) using the capillary rise equation [Campbell, 1985]:
while the study period maximum was 47.8 mm. Of the monitored storms, 12 yielded no flow, 8 produced only hillslope flow, and 9 produced both hillslope and streamflow responses. Storms in which runoff responses overlapped were combined to simplify analysis.

**Hydrologic Response to Storms**

The relationship between rainfall and subsurface flow or streamflow (Figure 5b) indicated that rainfall thresholds of ~8 and ~17 mm had to be exceeded to produce a hillslope and a stream response, respectively. Thus most rainfalls in PCI-08 do not generate storm flow and simply contribute to soil moisture recharge and/or base flow. Only events producing a basin-wide response were selected for subsequent analysis, given our objective to determine hillslope subsurface flow processes and their contributions to storm flow production. Runoff characteristics for these events are summarized in Table 3.

**Shallow soil cover.** Trenches 1B and 2B were highly responsive to incoming rainfall. They were the first trenches to register flow at the BR depth, with runoff usually commencing within a half hour of initial rainfall. Mean lag-to-peak times (time between center of mass of rainfall and a peak runoff) at 1B and 2B were 2.7 and 2.9 hours, while peak runoffs ranged from 1.4 to 7.4 mm h⁻¹ and from 0.6 to 4.8 mm h⁻¹, respectively. Runoff ratios were high, ranging from 0.41 to 0.89 and 0.20 to 0.89 at trenches 1B and 2D, respectively (Table 3), and were similar to values from lichen-covered shield bedrock [Allan and Routel, 1994].

The event that began on YD 136 (peak 10-min intensity of 18.3 mm h⁻¹, total depth of 44.5 mm) illustrates the control of rainfall on runoff characteristics in areas of thin soil (Figure 6).
Hydrograph and hyetograph shapes were similar, with synchronous peak intensities. High recession rates resulted in runoff ending shortly after rainfall cessation.

The main differences between the two sites were consistently lower peak runoffs and runoff ratios at trench 2B (Table 3). For example, peak runoff rates on YD 136 were 5.8 and 3.3 mm h\(^{-1}\) at trenches 1B and 2B, respectively. The runoff ratio at trench 1B was 0.70 compared to 0.42 at 2D for the same event. This reflects the thinner soil cover and steeper bedrock gradient above trench 1B (Table 1).

The responsive nature of flow at the shallow trenches is due to the small area and thin soil cover of their microcatchments (Table 1). The thin, loosely structured organic soil limited potential water storage and the travel time to bedrock, as supported by high runoff ratios and short lag-to-peak times. A cavity ~0.01 m thick between the soil and the bedrock surface served as a conduit for subsurface flow. Thus rainfall was able to infiltrate rapidly to bedrock, where lateral saturated flow over its surface quickly transmitted the water downslope.

**Deep soil cover.** Areas with deeper soil were not as responsive to rainfall compared with the shallow trenches, with smaller runoff totals, peak runoff rates and runoff ratios, and longer lag-to-peak times (Table 3). For example, peak runoff rates on YD 136 at the soil-bedrock interface of trench 1A were 2 orders of magnitude smaller and were 3 to 6 times smaller at trench 2A than at the shallow trenches (Figures 6b, 6c, 7a, and 7b). Lags to peak were 13 and 3.8 hours longer at trenches 1A and 2A, respectively, than at 1B and 2B for this event, and runoff coefficients were up to an order of magnitude smaller.

A small amount of flow was observed from the Ae trough at the deep trenches for most events (Table 3). Ae runoff usually preceded runoff at deeper levels and was of short duration (Figure 7b). Since this flow usually occurred at the onset of rainfall, lags-to-peak were ~0 hours. Peak runoffs, runoff coefficients, and total runoff from Ae troughs were several orders of magnitude smaller than flow at deeper levels in the soil (Table 3).

Ae runoff was likely the result of rain falling upon leaf litter immediately upslope of the trench. Ae runoff at trench 2A during the YD 136 event had a δ\(^{18}\)O signature (~5.92‰) very similar to that of rainfall (~5.45‰), indicating that this water was Horton flow over the leaf litter and organic mat. This process has been noted during early autumn storms [Renzetti et al., 1992] and was observed over short distances on hardwood forested slopes [Whipkey, 1965]. Rapid cessation of flow over the litter layer and the minute amount of Ae runoff may reflect progressive wetting of the litter, which would promote a shift from flow over the organic mat in the early part of the storm to infiltration through the litter layer into the underlying mineral soil later in the event, thus terminating runoff into the Ae trough.

Subsurface flow over the bedrock surface constituted 97 to 99.9% (trench 1A) and 91 to 99.9% (trench 2A) of total runoff at the sites with deeper soil (Table 3). Runoff at the INT level...
was recorded in four events, and its contribution to total runoff ranged from 0 to 2% at trench 1A and from 0 to 7.9% at trench 2A. Flow at the soil-bedrock interface averaged 2 orders of magnitude greater than INT flow at trench 1A and 1 order greater at trench 2A (Table 3). INT flow occurred only when the saturated layer, initiated at the bedrock surface, rose above the intermediate elevation; thus INT flow consisted of the upper portion of bedrock flow. Several lines of evidence support this: (1) INT flow never occurred in the absence of bedrock flow; (2) it consistently commenced after flow initiation at the soil-bedrock interface; (3) it occurred along with peak bedrock flow, with lags to peak 0.3 and 4.0 hours longer than at the soil-bedrock interface of trenches 1A and 2A, respectively; and (4) it quickly receded along with bedrock flow recession (Figures 7a and 7b).

Comparison of the deep trenches reveals consistent differences. Trench 2A had shorter lag-to-peak times and greater peak runoffs and runoff ratios than trench 1A (Table 3). Both mean runoff ratio and mean bedrock peak runoff rate were an order of magnitude greater at trench 2A, while mean lag to peak was almost 1.5 times greater at trench 1A. These differences in runoff characteristics also result from differing microcatchment properties. A portion of trench 2A’s contributing area consisted of exposed bedrock and thin organic soil cover; therefore all the rainfall did not have to percolate through ~0.5 m of soil before reaching bedrock as at trench 1A. This permitted rapid generation of saturated flow at the bedrock surface, which then quickly passed through the deeper soil of the microcatchment along the soil-bedrock interface. Trench 1A’s contributing area consisted of a thicker, more uniform soil cover, such that infiltrating rainfall had to percolate through a larger potential storage reservoir. This resulted in smaller peak runoffs and runoff coefficients, and the thicker soil cover also imposed a longer vertical travel path for subsurface flow and hence longer lags to peak (Table 3).

Results from the throughflow trenches confirm that hillslope runoff consists almost entirely of subsurface flow over the bedrock surface, thus supporting hypothesis 1: On shield slopes with thin soil cover, most flow occurs at the soil-bedrock interface. This is partially supported by Tsuboyama et al. [1994], who found that flow captured at the bedrock surface on a steep forested slope in Japan supplied 70–95% of total discharge from an instrumented soil pit; however, their experimental design did not indicate if flow was concentrated at the soil-bedrock interface, as in PCI-08.

**Hillslope Contributions to Basin Storm Flow**

Streamflow reacted rapidly to rainfall that generated a basin-wide response, with a mean lag to peak of 9.7 hours. Peak runoff rates ranged from $3 \times 10^{-2}$ to 1.7 mm h$^{-1}$ (Table 3), and both peak runoffs and lags to peak are indicative of basins dominated by subsurface flow [Dunne, 1978], as confirmed by the throughflow trenches. Runoff coefficients ranged from 0.15 to 0.86 and increased with rainfall depth (Figure 5b). Quick flow constituted 9 to 66% of total runoff, and quick flow coefficients ranged from 0.02 to 0.57 (Table 3). These results suggest that a large portion of the basin responded rapidly to rainfall, especially during larger events [cf. Wels et al., 1991b]. Therefore hillslope runoff responses were compared to those of PCI-08 to assess the importance of side slope subsurface contributions to storm flow production.

Roughly 10% of PCI-08 consists of exposed bedrock and thin organic-Ae soil cover. These areas can deliver water rapidly to deeper soils in the stream valley and may be important contributors to streamflow generation on the hydrograph’s rising limb. However, runoff from these zones likely loses its distinctive character during flow through deeper soils on the side slopes and in the near-stream area, as is suggested by dissimilar hydrographs for the shallow trenches relative to that of PCI-08 (Figures 6b, 6c, and 7c).

**Figure 8.** Peak and total runoff at the soil-bedrock interface at trenches 1A and 2A versus (a) peak runoff and (b) quick flow for PCI-08.

Figure 8 compares microcatchment responses for areas with deeper soil cover (trenches 1A and 2A) to those of the basin. Trenches 1A and 2A were assumed to represent their respective side slopes, such that the trench runoffs were averaged to obtain the total side slope contribution to the channel. Flow at the soil-bedrock interface of the side slopes bracketed the basin response and could supply 233 ± 188% of peak storm runoff (number of observations $n = 6$; range, 63–500%) and 114 ± 59% of quick flow ($n = 6$; range, 63–225%) from PCI-08. These estimated contributions to peak storm flow are similar to those obtained by Renzetti et al. [1992] using trench 1A data. Side slope lag-to-peak times were shorter than those of the basin, implying that the slopes were able to supply this runoff in time to contribute to peak storm flow (Table 3). Thus despite their small size, the throughflow trench microcatchments appear to represent flow on the convex slopes draining to the near-stream zone in PCI-08.

Although these comparisons show that subsurface flow at the soil-bedrock interface on the side slopes can account for PCI-08’s peak runoffs and quick flows, they are unable to answer the following questions: (1) are the side slopes necessary for storm flow generation in the basin, and (2) if they are, what are the mechanisms capable of delivering subsurface flow to the channel in time to contribute to storm flow?

**Preferential Flow Pathways: DOC Tracing**

Rainwater picks up DOC in throughfall [Likens et al., 1977] and by dissolution of organic matter as it percolates through
the soil’s organic layer [LaZerte, 1989; Jardine et al., 1990]. A combination of microbial decomposition, adsorption, and precipitation causes percolate to lose DOC as it passes through mineral horizons [Stevenson, 1985]. Thus $-15 \, \text{g DOC m}^{-2} \text{y}^{-1}$ is exported from organic soil horizons in PC1-08; yet $<5\%$ of this passes the lower mineral (B) horizons [LaZerte, 1991].

Peced studies by Jardine et al. [1989a, 1990] used DOC levels in soil water as an indirect means of assessing preferential flow. They argued that large DOC fluxes observed at depth in the soil were indicative of water movement via preferential flow pathways (macropores), since flow through the matrix would lead to immobilization of the reactive DOC. Use of conservative tracers by other investigators [White et al., 1984; 1sboya et al., 1994] has demonstrated the strong control that macropores exert on solute and water flow through soil. This arises from the ability of macropores to conduct water quickly to depth, thus allowing solutes to bypass the matrix and undergo limited interaction with the soil. Conversely, slower matrix flow (i.e., through micropores) increases contact with reactive surfaces and prevents DOC migration [Jardine et al., 1990].

Hillslope DOC transport was examined to assess the potential for vertical and lateral preferential flow pathways through the soil. Rainfall on YD 136 contained 1 mg L$^{-1}$ DOC. As expected, highest lysimeter DOC concentrations were found in the organic-Ae horizon lysimeters, and a preevent sample at trench 2A had 47 mg L$^{-1}$ DOC. Runoff from this Ae trough had 58–70 mg L$^{-1}$ DOC, with the highest level occurring at peak runoff (Figure 9a). These concentrations were similar to those in runoff from the shallow soil covered areas (not shown). Therefore rainwater acquired a high DOC concentration during throughfall and percolation through the I FH (organic layers developed from leaves, twigs, and woody materials) and organic-Ae horizon, as is typically found in podzols [e.g., McDowell and Wood, 1984; LaZerte, 1989].

Mineral (B) horizon lysimeter DOC concentrations were much lower than those in the organic-Ae horizon (Figure 9). Midslope concentrations during this event averaged $3.68 \pm 1.09 \, \text{mg L}^{-1}$ (0.31 m depth) and $6.36 \pm 1.72 \, \text{mg L}^{-1}$ (0.52 m depth); mean concentrations immediately upslope of the trench face were $9.96 \pm 0.68 \, \text{mg L}^{-1}$ (0.28 m depth) and $8.08 \pm 0.88 \, \text{mg L}^{-1}$ (0.57 m depth) (Figures 9b and 9c). Despite infiltration of DOC-rich water from the overlying organic-Ae horizon and temporal variations in soil water content which would induce changes in concentrations of less reactive solutes, mineral (B) horizon lysimeter samples showed no marked change in DOC during or after rainfall (Figures 9b and 9c).

These relatively constant DOC levels are likely not the result of adsorption by the porous ceramic cups, since concentrations reported here are similar to mean DOC levels obtained from polystyrene zero-tension lysimeters in PC1 by the OMEE (1987–1992), which ranged from 1.67 ± 0.02 to 8.1 ± 1.10 mg L$^{-1}$ [Findesh et al., 1993]. Thus mesopore and micropore water sampled by the suction lysimeters appears representative of the soil water solution. Lysimeter results indicate that water which moved uniformly through the soil matrix quickly lost DOC by adsorption [McDowell and Wood, 1984; Jardine et al., 1989b] and acquired a DOC signature similar to that of preevent soil water.

Runoff DOC levels from both deep soil microcatchments were significantly greater than lysimeter concentrations. At trench 2A, runoff DOC concentrations at the soil-bedrock interface ranged from 6.4 to 14.2 mg L$^{-1}$ during the YD 136 event, and peak runoff DOC levels were almost twice those of lysimeter samples. Values declined slowly following peak concentration and eventually converged with lysimeter values during hydrograph recession, implying that matrix water contributions became increasingly important later in the event (Figure 9c). Runoff DOC concentrations from the INT trough were significantly lower than at the bedrock surface, ranging from 4.8 to 6.9 mg L$^{-1}$ (Figures 9b and 9c). These levels were similar to those found in the suction lysimeters, suggesting that flow from the INT trough consisted mostly of matrix water. Similar trends occurred at trench 1A and at both sites during a fall event (YD 274).

Marked differences between lysimeter and soil-bedrock interface runoff DOC levels at both sites indicate that some infiltrating water moved via preferential flow pathways, both vertically to bedrock and laterally over the bedrock surface. Flow solely through the matrix, either vertically and/or laterally, would have been indicated by subsurface runoff DOC concentrations similar to those of lysimeter samples. Thus although preferential flow was not measured directly, DOC results support hypothesis 2: A significant fraction of event water moves via preferential flow pathways vertically to the bedrock surface and laterally along the soil bedrock interface.

Eluviated macropores (up to 30 mm in diameter) were found at the trench faces, and subsurface flow occurred along the surface of live roots emerging into the trench. Independent support for preferential flow is supplied by combining results from (3) with the SWR curves to obtain the pore size distr-

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**Figure 9.** Hillslope runoff and lysimeter DOC concentrations for the (a) Ae, (b) intermediate, and (c) bedrock levels at trench 2A for YD 136 event.
bution of the extracted cores (Figure 10). The distribution shows that ~15% of total porosity consisted of pores ≤0.14 mm in diameter (Ψ = -2 kPa), while pores with diameters from 0.015 to 0.14 mm make up ~30% of soil porosity. Thus ~45% of the porosity of PCI-08 soils consists of macropores and mesopores [Lusumore, 1981], which exert a strong control on subsurface flow and solute transport in forest soils [Jardine et al., 1989a].

Further support for preferential flow pathways comes from a chemical weathering study of PCI-08 soils [Kirkwood and Nesbitt, 1991]. Utilizing a chemical index of alteration (based on the molar proportion of Al₂O₃ and labile oxides found in feldspar soils), they found a highly weathered zone (0–0.03 m thick) overlying the bedrock that resembled the eluviated Ae horizon. Kirkwood and Nesbitt [1991] concluded that the permeable nature of the sandy soils, combined with the smoothly polished bedrock surface, allows flow through this thin zone at the soil-bedrock interface. This highly weathered zone, which may be connected with other lateral and vertical preferential flow pathways (e.g., eluviated macropores), potentially serves as a conduit for preevent and event water fluxes to the near-channel area.

Figure 11. IHSs for the (a) intermediate and (b) bedrock levels at trench 1A for YD 136 event.

Origin of Hillslope Subsurface Flow and Basin Streamflow: Isotopic Hydrograph Separation

Hillslope subsurface flow and PCI-08 storm flow for the YD 136 event were separated into event and preevent contributions using δ¹⁸O content and (1). Key assumptions underlying these IHSs are that storm runoff is from two water sources (event and preevent) and that each source has a spatially and temporally uniform isotopic signature [Buttle, 1994].

Use of the bulk rain sample collected in an open area (~5.45%) as $C_e$ presents two problems when performing IHS for the YD 136 event. This storm lasted 17.5 hours, and a bulk sample masks temporal changes in $C_e$ during the event such that $Q_p$ at any instant is influenced by the δ¹⁸O of rain that has yet to fall [McDonnell et al., 1990]. Rain δ¹⁸O tends to become depleted during large storms of long duration [e.g., Pionke and DeWalle, 1992]. Any depletion of rain δ¹⁸O during the YD 136 event implies that the bulk sample underestimates $Q_p$ at the beginning of runoff and overestimates it near the end of flow by an unknown amount. The second drawback to use of the bulk sample as $C_e$ is that it may not represent throughfall δ¹⁸O in PCI-08. Saxena [1986] found that throughfall was enriched by up to 1.2‰ relative to rainfall over a Swedish pine forest, while DeWalle and Swistock [1994] observed a maximum enrichment of 1.61‰ in throughfall in a spruce forest during spring rains. A 1‰ enrichment of throughfall for YD 136 means that bulk rainfall would underpredict $Q_p/Q_s$ by up to 6%, which does not alter our interpretation of the IHS results.

Preevent soil water δ¹⁸O decreased with depth at site 3. This spatial variability was incorporated in hillslope IHSs by assuming that $C_p$ was represented by soil water sampled ~1.5 hours before rainfall immediately upslope of the trench at the depth corresponding to the trough for which a separation was performed (i.e., intermediate soil water used for INT flow separation; near-bedrock soil water for soil-bedrock interface flow separation). The worst case scenario (10% underestimation and 8% overestimation of $Q_p/Q_s$) occurred when the mean B horizon soil water δ¹⁸O (~11.67 ± 0.60‰, n = 6) was used as $C_p$ instead of the appropriate soil water samples; thus IHS results were not seriously compromised by spatial variability in $C_p$.

Preevent base flow (~11.96‰) was assumed to represent the isotopic signature of near-stream groundwater likely to reach the stream channel during an event [Sklash, 1990]. Base flow δ¹⁸O was not significantly different from the mean preevent soil water signature, and PCI-08 base flow was apparently supplied by soil water contributions from the side slopes. Base flow δ¹⁸O (and hence $C_p$) was assumed to be temporally constant, as in numerous IHSs [Buttle, 1994].

Assumptions underlying the IHS technique were satisfied sufficiently to permit its use in separating event and preevent contributions to hillslope subsurface flow and to storm flow. Nevertheless, Rodhe [1987] suggests that unaccounted spatial and temporal variations in $C_e$ and $C_p$ may introduce uncertainty of the order of ±10% to IHS results.

Deep Soil Trenches 1A and 2A

There was significant enrichment of soil water δ¹⁸O at both sites as a result of infiltration of rainwater (Figures 11 and 12), and soil water δ¹⁸O near the bedrock surface increased from
levels reflected immobilization of DOC in water percolating through micropores.

Relative event water contributions at the soil-bedrock interface of trench 1A peaked at the onset of flow, and preevent soil water contributions became increasingly important with time (Figure 11b). A similar pattern was noted by Wilson et al. [1991] for subsurface flow from a deep forest soil in Tennessee. Large event water contributions during initial flow confirm that a significant fraction of rainfall traveled to and over the bedrock via preferential flow pathways, consistent with DOC tracing. Conversely, 100% matrix flow of event water would have displaced preevent soil water and resulted in minimal event water contributions on the hydrograph’s rising limb.

Only one bulk runoff sample was available for IHS at trench 1A’s INT trough (Figure 11a). Event water constituted 25% of total runoff, which is identical to that estimated at the soil-bedrock interface (Table 4). This suggests that INT and BR runoff originated from the same source, which was dominated by preevent water. Event water contributed 20% of total INT runoff at trench 2A (Table 4). Initial flow consisted almost entirely of preevent soil water, while relative event water contributions increased with time. Given that DOC results indicate predominantly matrix flow in the B horizon, it appears that a saturated layer "backed up" into mineral soil above the bedrock surface and produced the initial displacement of preevent soil water by transitory flow. This soil water was then increasingly replaced by event water. Runoff $\delta^{18}O$ at the INT trough was consistently lighter than that at the soil-bedrock interface throughout the event (Figure 12), which is similar to the trend for DOC. This was reflected in greater relative event water contributions to flow over the bedrock and supports the presence of a thin conductive zone at the soil-bedrock interface capable of translating event water inputs rapidly downstream.

Approximately 14% of incident rainfall left the trench 2A microcatchment as subsurface runoff at the soil-bedrock interface, and event water supplied 48% of total subsurface runoff on YD 136. This compares to the throughputs of <1% of rainfall on the trench 1A microcatchment along the bedrock surface, where event water supplied only 25% of total subsurface flow (Table 4). As with trench 1A, event water contributed to flow at trench 2A throughout the event; however, there was a lag between peak runoff and maximum relative event water contributions. Event water inputs decreased on the hydrograph’s falling limb and then rapidly leveled off at 50% of total

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Table 4. Pre-event and Event Water Contributions to Subsurface Flow and Streamflow for Year Day 136 Event

<table>
<thead>
<tr>
<th>Site</th>
<th>Trough*</th>
<th>Runoff Depth, mm</th>
<th>Percent of Total Runoff</th>
<th>Runoff Depth, mm</th>
<th>Percent of Total Runoff</th>
<th>Percent of Rainfall</th>
</tr>
</thead>
<tbody>
<tr>
<td>Trench 1A</td>
<td>INT</td>
<td>$6 \times 10^{-4}$</td>
<td>75</td>
<td>$2 \times 10^{-4}$</td>
<td>25</td>
<td>&lt;1</td>
</tr>
<tr>
<td></td>
<td>BR</td>
<td>0.6</td>
<td>75</td>
<td>0.2</td>
<td>25</td>
<td>&lt;1</td>
</tr>
<tr>
<td></td>
<td>total</td>
<td>0.6</td>
<td>75</td>
<td>0.2</td>
<td>25</td>
<td>&lt;1</td>
</tr>
<tr>
<td>Trench 2A</td>
<td>INT</td>
<td>0.8</td>
<td>80</td>
<td>0.2</td>
<td>20</td>
<td>&lt;1</td>
</tr>
<tr>
<td></td>
<td>BR</td>
<td>0.2</td>
<td>50</td>
<td>6.2</td>
<td>50</td>
<td>14</td>
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<tr>
<td></td>
<td>total</td>
<td>7.0</td>
<td>52</td>
<td>6.4</td>
<td>48</td>
<td>14</td>
</tr>
<tr>
<td>PCI-08</td>
<td></td>
<td>14.8</td>
<td>77</td>
<td>4.5</td>
<td>23</td>
<td>10</td>
</tr>
</tbody>
</table>

Contributions are expressed as runoff depth and as a percentage of total runoff, while event water runoff is also expressed as a percentage of rainfall. IHS was not performed for flow from the Ae troughs.

*INT, intermediate; BR, bedrock.
flow, indicating contributions from a well-mixed soil reservoir where runoff $^{818}O$ equaled that of soil water (Figure 12b). A similar trend in DOC levels was also observed at trench 2A (Figure 9).

Antecedent soil water depths at both sites were more than sufficient to account for estimated preevent soil water contributions to flow (Table 4). The preevent water fraction of total subsurface flow was greater at trench 1A compared with trench 2A. This corresponds to the thinner soil cover and smaller reservoir of preevent water within the latter’s microcatchment and supports hypothesis 3: The relative preevent water contribution to subsurface flow on shield slopes is a function of soil thickness.

**PC1-08**

Isotopic separation of the PC1-08 hydrograph (Figure 13) indicated that storm flow was dominated by preevent water input on YD 136 (Table 4). This accords with studies on the Canadian Shield and with previous work in PC1 during rain and snowmelt events [e.g., Bottomley et al., 1984; Wels et al., 1991a, b; Hinton et al., 1994]. Potential water storage in PC1-08 is 154 mm, assuming an average soil depth of 0.29 m and porosity of 53% [Lozano et al., 1987], such that only 10% relative saturation can account for the 14.9 mm of preevent water supplied to storm flow (Table 4). Since preevent soil water depths on the side slopes exceeded 80 mm, there was sufficient soil water and/or groundwater to account for the preevent water flux from the basin.

Unlike the side slopes, little event water contributed to initial storm flow (Figure 13). Peak event water input coincided with peak runoff, and the absence of a lag between peak preevent and total runoff discharges suggests that event water delivery by direct precipitation onto saturated areas (DPSA) near the stream was not an important contributor to storm flow [cf. Sklash et al., 1986]. Instead, the minor event water contribution at the start of storm flow, the coincidence of peak total and peak event water runoff, and high relative event water contributions on the falling limb indicate that event water was supplied mainly by subsurface flow. Event water contributions to storm flow via DPSA likely increase with antecedent wetness, and tracing studies by Wels et al. [1991b] during snowmelt in PC1-08 indicated that DPSA accounted for up to 15% of basin runoff.

Although ~10% of the basin consists of exposed bedrock and thin soil, rainfall onto this area must flow through soil before reaching the stream [Wels et al., 1991b]. Trench hydrologic results showed that Horton flow was insignificant. Thus the only zone capable of translating rainfall input directly to storm flow was the near-channel saturated area. The extent of prestorm stream valley saturation was $1869 \pm 205 \text{ m}^2$ from (2). This estimate was derived from Shibata's [1988] saturated area mapping conducted during snowmelt, when antecedent wetness was much greater than for this event. Therefore (2) likely overestimated $A_{sat}$ prior to the YD 136 event, and field observations during the event also suggest that the actual saturated valley area was less than that predicted by (2). Rainfall onto $A_{sat}$ contributed 2.3 \pm 0.3 mm of event water to storm flow. Although this value overestimates DPSA, it accounts for only ~51% of the total event water contribution to storm flow (Table 4). Therefore subsurface event water contributions from other portions of the basin were needed to supply the remaining 2.2 \pm 0.3 mm of event water in storm flow. Event water flux from the unsaturated area of the stream valley is unknown; however, PC1-08 side slopes were capable of rapidly delivering 73% (3.3 mm) of the total event water input to streamflow, and therefore all event water not accounted for by DPSA. This supports hypothesis 4: A significant portion of event water flux in storm flow from forested basins with shallow soil cover is supplied from side slopes via subsurface flow along the soil-bedrock interface.

These results have other implications: (1) in addition to significant event water contributions, side slopes could supply ~25% of total preevent water flux from the basin via subsurface flow over the bedrock surface; (2) a large portion of event and preevent water supplied to storm runoff may have traveled considerable distances via preferential subsurface routes, such that a substantial fraction of PC1-08 contributes to storm flow during large rainfalls; and (3) the groundwater ridging mechanism [Sklash and Farvolden, 1979] is not the only means by which event and preevent water may be rapidly delivered to the channel.

**Conceptual Model of Runoff Production**

Sklash [1990] suggested that groundwater ridging is the dominant mechanism capable of rapidly delivering large volumes of preevent water to storm flow (Figure 14b). This hypothesis posits that preevent water is supplied by groundwater in the near-stream zone and that the event water fraction of storm flow results from direct precipitation onto groundwater discharge zones. Formation of near-stream groundwater ridges is apparently independent of the hydrological response of upslope areas, which have little influence on initial storm flow generation but may become important later in the event [Sklash and Farvolden, 1979].

Our results suggest an alternative conceptual model of storm flow generation in forested shield basins with shallow soil cover (Figure 14b). A significant fraction of event water falling on hillslopes infiltrates to the impermeable bedrock via vertical preferential flow, while the remainder of the event water is routed via slower Darcian matrix flow where storage and mixing with preevent water occurs. A saturated layer consisting of event and preevent water forms above the bedrock over large portions of upslope areas in the basin, initiated by mixing of event water fluxes via macropore flow with soil water immediately above the bedrock and production of phreatic conditions. DOC results suggest that saturation does not occur by mixing of event and preevent water in the upper soil matrix and piston-type displacement of this water to the bedrock, since this would result in DOC levels in BR flow similar to those in the matrix soil water. Instead, high DOC concentrations in initial flow at the soil-bedrock interface provide strong evi-
Figure 14. Potential pathways of event and pre-event water discharging to the channel in PC1-08: (a) the groundwater ridging mechanism and (b) vertical movement of event water via macropores and lateral movement of event and pre-event water to the near-stream zone via flow at the soil-bedrock interface. Hypothesized water table profiles prior to the event ($t = 0$) and at peak runoff are indicated by dashed lines.

dence for mixing of DOC-rich event water with a thin layer of pre-event water and subsequent rapid movement of this mixture along the soil-bedrock interface to the near stream zone. Matrix water contributions become increasingly important later in the event, as indicated by convergence of BR runoff and soil water DOC levels. Pre-event contributions from near-stream groundwater and event water delivery via DPSA are still important storm flow components but are less significant than envisaged by the groundwater ridging mechanism. However, processes by which slope contributions interact with near-stream soil and groundwater prior to discharge to the channel are unknown, and studies linking slope runoff with near stream hydrologic conditions are needed.

This model is consistent with results of previous tracer studies in PC1-08 [Wels et al., 1991a, b], where significant subsurface transport of event water was indicated. The model is also a variant of that proposed by McDonnell [1990] to reconcile the importance of macropore flow in hillslopes to large pre-event
contributions to throughflow and streamflow in the Maimai basin, New Zealand. He proposed a mechanism whereby vertical soil cracks allow event water to infiltrate to an underlying impermeable conglomerate surface, where small amounts of event water convert the nearly saturated soil above this surface into a transient perched aquifer. This saturated layer consists largely of preevent water, which is subsequently drained rapidly downslope via a network of pipes. Such models of coupled vertical and lateral preferential flow in shallow forested soils may describe runoff processes over large regions of the Canadian Shield. For example, the thin till and exposed bedrock that compose PCI-08 makes up >40% of the surficial geology in 17 of the 32 AP1OS basins in central Ontario [Dillon et al., 1991]. However, these models do not apply to the heterogeneous shield landscape studied by Allan and Routel [1994], where Horton flow over bedrock is the primary runoff mechanism and subsurface storm flow from forested soil pockets is through near-surface organic horizons rather than deeper mineral soils. They are also inappropriate in areas with deeper overburden and more extensive groundwater discharge zones, where near-channel groundwater fluxes dominate storm flow production and DPSA regulates event water fluxes from the basin [e.g., Hinton et al., 1994].

Summary and Conclusions

Hydrometric, geochemical, and isotopic results from a shallow soil headwater basin on the Canadian Shield indicate that subsurface flow on the basin’s side slopes plays an important role in storm flow generation. Results from throughflow trenches draining several microcatchments within the basin confirm the hypothesis that ~100% of flow occurs within a thin weathered zone at the soil-bedrock interface. This flow was capable of supplying most quick flow and peak runoff from the basin. Non-Darcian flow appears to dominate water flux in these forest soils, and joint use of reactive (DOC) and nonreactive (18O) tracers indicated preferential movement of rainwater vertically to bedrock and laterally over the bedrock surface. This vertical and lateral preferential flow is coupled, such that rapid infiltration of event water results in soil saturation above the bedrock and in enhanced downslope flux of both event and preevent water.

Runoff characteristics recorded at the throughflow trenches were a function of soil depth within the microcatchments. Greater soil depth was directly related to lag-to-peak times and inversely associated with peak runoff and runoff coefficients. IHSs for a large rainstorm suggested that the relative contribution of preevent water to total runoff from the microcatchments increased with mean soil thickness and prestorm soil water depth.

The IHSs also indicated that side slopes could supply a significant fraction of preevent water flux from the basin, along with the event water fraction of basin storm flow not attributable to DPSA. This event water may travel considerable distances over the bedrock prior to reaching the near-stream zone, and a greater proportion of shallow soil basins may participate in storm flow generation than is envisaged by the groundwater ridging hypothesis. The chemistry of event water moving via macropores to bedrock and along the soil-bedrock interface to the channel may differ from that of DPSA because of an enhanced opportunity for theformer to interact with soil and regolith during passage through the basin. Residence times for this preferential flow may be insufficient to neutralize acidic inputs to hillslope soils, so that a greater fraction of the basin may contribute to total acid export than envisaged by the groundwater ridging model. Nevertheless, flow above the bedrock surface may contain higher concentrations of weathering products than in DPSA. Therefore rapid movement of event water by subsurface routes should be considered when using environmental isotopes to assist in interpreting the hydrochemical behavior of forested basins.

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