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# STREAMFLOW GENERATION IN A HEADWATER BASIN ON THE PRECAMBRIAN SHIELD

CHRISTOPH WELS, COLIN H. TAYLOR, R. JACK CORNETT*
*Watershed Ecosystems Program, Trent University, Peterborough, Ontario K9J 7B8, Canada*

AND

BRUCE D. LAZERTE
*Dorset Research Centre, Water Resources Branch, Ontario Ministry of the Environment, P. O. Box 39, Dorset, Ontario P0A 1E0, Canada*

## ABSTRACT

Current conceptual runoff models hypothesize that stormflow generation on the Canadian Shield is a combination of subsurface stormflow and saturation overland flow. This concept was tested during spring runoff in a small (3·3 ha) headwater basin using: (1) isotopic and chemical hydrograph separation and (2) field mapping and direct tracing of saturated areas. Isotopic and chemical hydrograph separation indicated three runoff components: (1) pre-melt subsurface flow; (2) subsurface flow of new (event) water; and (3) direct precipitation on to saturated areas (DPS). During early thaw–freeze cycles, their relative contributions to total flow remained constant (65 per cent, 30 per cent, and 5 per cent respectively). It is hypothesized that lateral flow along the bedrock/mineral soil interface, possibly through macropores, supplied large volumes of subsurface flow (of both old and new water) rapidly to the stream channel. Much higher contributions of DPS were observed during an intensive rain-on-snow event (15 per cent of total flow). Mapping and direct tracing of saturated areas using lithium bromide, suggested that saturated area size was positively correlated to stream discharge but its response lagged behind that of discharge. These observations suggest that the runoff mechanisms, and hence the sources of stream flow, will vary depending on storm characteristics.

KEY WORDS Runoff processes Macropore flow Saturated areas Isotopic hydrograph separation Chemical tracers

## INTRODUCTION

Numerous isotope studies in humid forest environments have shown that old water is the dominant contributor to stream discharge during both rain storms (e.g. Sklash and Farvolden, 1979; Sklash *et al.*, 1986) and snowmelt events (e.g. Rodhe, 1981; Bottomley *et al.*, 1986). The small but significant fraction of new water delivered directly to the stream in many watersheds on the southern Canadian Shield can be satisfactorily accounted for as direct precipitation on to saturated surfaces (DPS) near stream channels (e.g. McDonnell, 1985; Shibatani, 1988).

The rapid delivery of large volumes of old water is less easily explained. Sklash and Farvolden (1979) proposed groundwater ridging as the dominant mechanism. However this process is limited to the near-channel area and cannot account for the majority of streamflow in watersheds, where the total seasonal runoff in response to snow melt is often more than 80 per cent (McDonnell, 1987; Shibatani, 1988; Wels, 1989). Here the entire watershed must be contributing to stormflow. Processes for the delivery of large volumes of old water from distant upland areas to near-stream areas are not well understood. Piston translatory flow (Hewlett and Hibbert, 1967) through the soil matrix (Darcian flow) is too slow in many watersheds to explain the fast delivery of old water to the stream. Therefore, non-Darcian flow through macropores (root channels, soil cracks, flow on bedrock) has been suggested as a delivery mechanism (e.g. Beven and Germann, 1982). However, flow through macropores has not been adequately described yet (Germann, 1989).

*Also at: Enviromental Research Branch, Chalk River Nuclear Laboratories, Chalk River, Ontario K0J 1J0, Canada

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The objective of this paper is to investigate streamflow generation for headwater catchments with shallow soils on the Canadian Shield during spring runoff. We used isotopic and chemical hydrograph separation, direct tracing and mapping of saturated areas, and hydrometric field observations to test the hypotheses that:

1. Subsurface flow comprising both old and new water is the dominant flow component in spring runoff;
2. New water in the upland areas infiltrates the soils and contributes to streamflow as return flow via subsurface pathways; and
3. All surface runoff (defined as water which retains the chemical signature of precipitation) is generated as direct precipitation on to near-stream saturated surfaces, which vary in extent during spring runoff.

Based on the results of this study, we developed a conceptual model to explain streamflow generation in headwater regions of the Canadian Shield.

## STUDY SITE

The study was conducted in a small headwater catchment (Plastic-108, 3·3 ha) within the Plastic Lake watershed (45°11′N 78°50′W) located on the Precambrian Shield in the Muskoka-Haliburton region, south-central Ontario. The mean annual January and July air temperatures are − 11·0 and 17·7°C, respectively. The mean annual precipitation depth is ≈1000 mm (1941–1980), with approximately 26 per cent falling as snow (Shibatani, 1988). Spring runoff is the dominant hydrological event in this region. The long term annual runoff is 400–600 mm with 50–75 per cent occurring during March–April in response to snowmelt (Scheider *et al.*, 1983).

The topography of the Plastic-108 subbasin is characterized by a relatively narrow valley with steep side slopes (Figure 1). Large, permanent wetlands do not exist in Plastic-108 and only the valley bottom is commonly saturated during periods of high flow. Surficial deposits vary considerably in depth. On the eastern side of the valley, till thickness ranges from 0·3 to 1 m and the podzolic soil sometimes shows the development of a C-horizon. The surficial deposits on the western side are much thinner. Here, the B-horizon of the podzolic soil is often absent and the bedrock is covered only by an A-horizon and/or a humus layer.

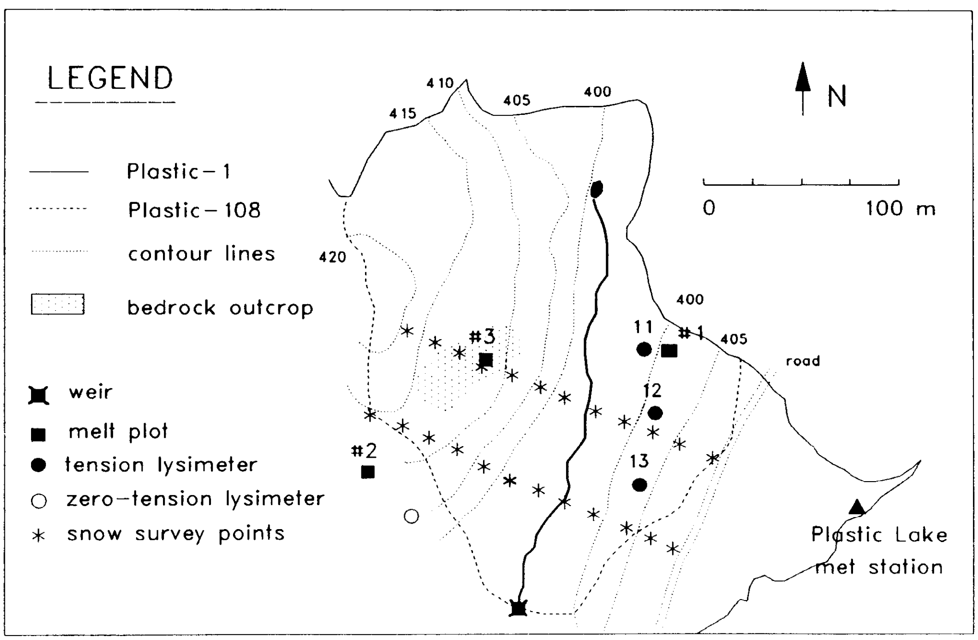



Figure 1. Field instrumentation in the Plastic-108 subbasin and vicinity. The contour interval in 5 m. The boundaries of the Plastic-1 catchment and Plastic-108 subbasin are shown as a thin solid line and a dashed line respectively

Hardly any soil and vegetation cover is found on a large bedrock outcrop on the western side of the valley that comprises $\approx 10$ per cent of the total area of Plastic-108 (Figure 1). The soils are sandy podzols on the side slopes and orthic gleysols in the valley bottom supporting a mature coniferous forest dominated by white pine and hemlock. A detailed description of the physiography, geology, and land use of the subbasin is given elsewhere (Wels, 1989).

## MATERIALS AND METHODS

### *Field methods*

Stream flow was gauged using a 90° V notch weir equipped with a Stevens water level recorder and heated to maintain ice-free conditions throughout the winter and spring months. Hourly discharge values were calculated from a stage-discharge relationship (O.M.E. unpublished data). Stream samples for isotopic and chemical analysis were collected using an ISCO automatic water sampler which was placed in the heated weir hut to prevent the suction lines and samples from freezing. The sampling interval was three hours during periods of high flow and six hours or more during baseflow and following peakflow.

Melt water was sampled at three sites (Figure 1). Two snowmelt lysimeters (#1 and #2) were installed on the west and east facing slope of Plastic-108 (Figure 1). Each lysimeter consisted of a clear (1 $m^2$) polyethylene sheet laid out before the first snow and held by a wooden frame. The combined runoff of melt and rain water from the snow lysimeter was collected into a 20 L carboy and sampled at least once a day. The volume of the water that had accumulated in the carboy was recorded to estimate the runoff from the melt plots. The third site (#3) was located in the centre of the large bedrock outcrop and was much more exposed than the two forested sites #1 and #2 (Figure 1). Here, an eavestrough intercepted the melt water at the base of the snowpack.

A snow survey consisting of 12 survey points along two parallel transects in an east-west direction was conducted prior to the onset of spring using a Canadian M.S.C. (A.E.S. large diameter) shallow snow sampler. The average of all survey points yielded an estimate of the water equivalent of the original snowpack. Temperature and precipitation were continuously recorded at the Plastic Lake meteorological station located in close proximity to the Plastic-108 subbasin and operated by the Ontario Ministry of the Environment (Figure 1).

A tracing experiment, using lithium bromide, was conducted in the near-stream area early in the snowmelt period in an attempt to identify the extent of the surface runoff contributing zone.

### *Analytical methods*

Samples for deuterium analyses were measured by mass spectrometry at Chalk River Nuclear Laboratories (Chalk River, Ontario) and concentration values are expressed in ppm D (I.A.E.A., 1984). Dissolved reactive silicate ($SiO_2$), expressed in mg $L^{-1}$ Si, was determined using colourimetry (blue silicomolybdate complex) (Strickland and Parsons, 1972). Both analyses were very reproducible with mean standard deviations of 0·13 ppm D and 0·02 mg $L^{-1}$ Si, respectively (Wels, 1989). Lithium and bromide were measured at Chalk River Nuclear Laboratories using an IL 251 flame atomic emission spectrophotometer and a Dionex 2000 Series 2020Y ion chromatograph with eluent suppression, respectively (Wels, 1989). The high sensitivity of the lithium analysis (lower limit of detection (LLD) of $\approx 0.05$ $\mu g$ $L^{-1}$ Li) allowed the direct measurements of very low lithium concentrations in the $\mu g$ $L^{-1}$ range (Wels, 1989). The bromide analysis was much less sensitive (LLD $\approx 0.05$ mg $L^{-1}$ Br) and demanded concentration of almost all samples. Similarly, the precision of the lithium analysis was much higher (mean $\sigma = 0.05$ $\mu g$ $L^{-1}$ Li) than that of the bromide analysis (mean $\sigma = 0.05$ mg $L^{-1}$ Br).

### *Hydrograph separation techniques*

The stable isotope deuterium and the weathering product silica were used to determine three runoff components to streamflow: (1) old water transmitted as subsurface flow; (2) new water (melt/rain) transmitted as subsurface flow; and (3) new water transmitted as surface flow. First, stream runoff was separated into old and new (melt/rain) water using the stable isotope deuterium as a non-reactive tracer on the basis of steady state mass balances (Dinçer *et al.*, 1979):

$$X_O = \frac{(C_S - C_N)}{(C_O - C_N)} \tag{1}$$

where $X_O$ = the relative contribution of old soil/groundwater, expressed as a fraction of total streamflow, $C$ = concentration of the isotope, and the subscripts $S$, $O$, and $N$ refer to the stream, old, and new water components respectively. The deuterium concentration of premelt stream baseflow (146·7 ppm, Figure 2) was used as the constant D of all old water $C_O$ (Wels, 1989). This allowed the separation of truly old water that was stored in the basin prior to the onset of melt. Daily values of $C_N$ were chosen as input variable to Equation 1 since it may provide more realistic estimates of the temporal changes in $X_O$ than the volume weighted mean $C_N$. The assumptions of isotopic hydrograph separation are discussed in more detail in Wels *et al.* (1990a, b).

In a second step, the geochemical tracer silica was used to separate streamflow into surface flow (SF) and subsurface flow (SSF) independent of its origin (old versus new) using the same equation (Wels, 1989). Here both input variables ($C_N$ and $C_O$) were assumed constant over time (0·3 and 2·78 mg L$^{-1}$) due to the consistent absence of silica in melt/rainwater and its rapid release from the mineral soil (Wels *et al.*,1990b). All new melt/rain water that contacted the mineral soil matrix (several hours contact time) before discharging to the stream would be considered 'new subsurface flow', and was estimated by subtracting the old water component derived from the D separation from the total subsurface flow estimated from the Si separation.

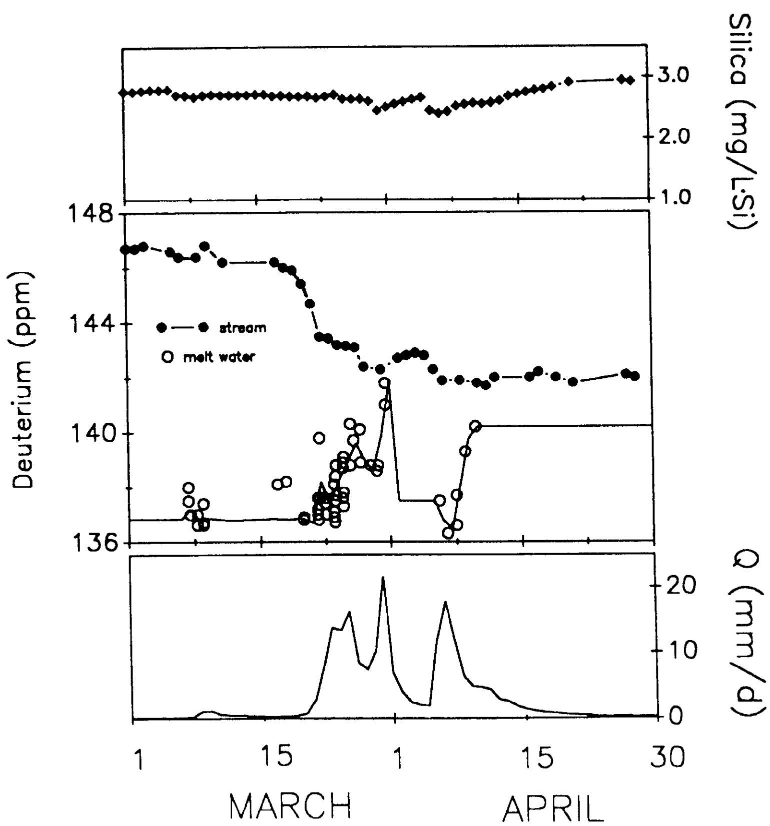



Figure 2. Temporal trends in deuterium and silica of streamflow and deuterium in runoff from three melt plots in Plastic-108. Mean daily discharge ($Q$) is shown for comparison. The solid line shows the splined input function for $C_N$

## RESULTS AND DISCUSSION

### *Characteristics of spring runoff 1987*

The 1987 spring runoff was of unusually low magnitude as a result of the rapid melt and the small amounts of melt/rain water (Wels, 1989). However, runoff response ratios (total runoff as a proportion of total available precipitation) and peak discharge rates were similar to previous years suggesting that the runoff response in 1987 was of average intensity. The runoff mechanisms studied during the 1987 spring runoff season may thus be dominant in years of larger runoff magnitude as well.

The total amount of water available for runoff during the spring melt of 1987 (March 1–April 31) was approximately 224 mm water equivalent. Of this, 134 mm was supplied by the original snowpack (March 1) and the remaining 90 mm water equivalent was supplied by additional snow and rain during the melt period. The total discharge in the first-order stream Plastic-108 during spring melt was 204 mm yielding a response ratio of 91 per cent. This suggests that evaporation was not important during spring runoff and that the headwater basin has only a limited capacity to store water in its shallow overburden.

The 1987 spring runoff can be divided into two melt periods: (1) a first melt period (M1, March 22–April 4) consisting of a period dominated by radiative melt (M1a) and a rain-on-snow event (M1b) (19·0 mm; Figure 3a); and (2) a shorter second melt period (M2) (April 4–14) with fewer rain inputs (8·9 mm; Figure 3b). The first melt period (M1) yielded considerably more runoff than the second melt period (M2); however, the peak daily runoffs were similar (Table I). Only the rain-on-snow event M1b (Figure 3a) produced flow rates ($> 1{\cdot}5$ mmhr$^{-1}$) above those observed during the remainder of the 1987 spring runoff (Table I; Figures 3a, b). The intense rain-on-snow event yielded a faster runoff response (lag-to-peak $\approx 2{\cdot}5$ hr) than observed during the previous melt-induced diurnal hydrographs (lag-to-peak $\approx 4{\cdot}5$ hr). Variations in antecedent soil moisture levels also account for some of the differences in runoff response from event to event.

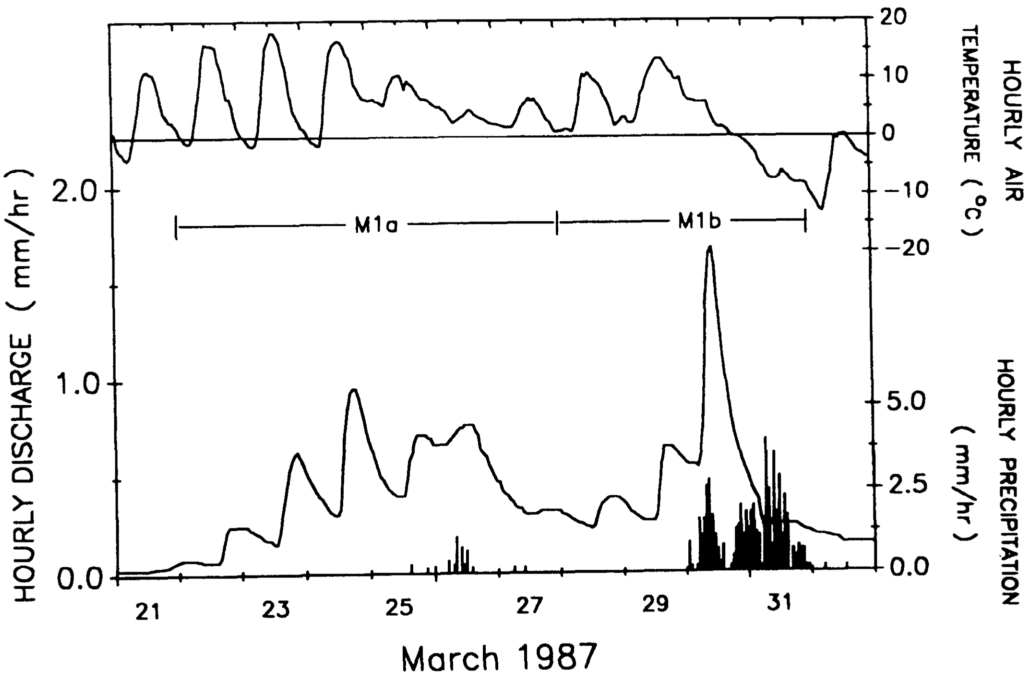
Dunne (1978) investigated the influence of runoff mechanisms and basin size on peak runoff rate and lag-to-peak time of streams in forested basins. He showed that the runoff response is faster and more intense in basins which produce saturation overland flow (SOF) compared to those which produce only subsurface stormflow (SSSF). Peak runoff rates and lag-to-peak times in Plastic-108 were intermediate between those reported for watersheds dominated by SOF and SSSF suggesting that neither one is dominant in Plastic-108 (Wels, 1989). Alternatively, Dunne's set of observations may not represent the entire range of possible runoff responses induced by SSSF and SOF. For example, snowmelt runoff studies, which are largely underrepresented in Dunne's set of data, may induce a slower, less intense runoff response.

### *Three-component hydrograph separation*

Silica concentrations in stream water showed small but consistent depressions on days of peakflow (Figure 2). The largest depressions in stream silica were observed in response to the rain-on-snow event on March 31 (15 per cent) and during peak melt on April 6 (18 per cent) (Figure 2). Apparently, contributions of surface flow to streamflow which would result in a decrease of stream silica concentrations were limited to days of highest flows. Stream silica concentrations showed no pronounced seasonal trend and, following an event, recovered within hours to pre-melt values (Figure 4).

In contrast, a very pronounced change was observed in the deuterium concentrations ($D$) of stream water over the course of spring runoff (Figure 2). The lower $D$ concentration of post-melt baseflow (142·0 ppm) compared to that of the pre-melt baseflow (146·7 ppm), indicated that considerable mixing of infiltrating melt water with old soil/groundwater occurred. The temporal changes in stream water $D$ were very smooth. Apparently, the spatial and temporal variability in the isotopic content of melt/rain water (Figure 2) and soil/groundwater is averaged out on a basin scale. Some short-term variations in stream water $D$ were observed during diurnal fluctuations in flow (Figure 4). However, they were much smaller than the seasonal $D$ trend.

Unfortunately, some days of high flow had considerable inputs of isotopically heavy rain water (e.g. March 30), introducing a larger uncertainty in the estimates of old water discharge rates (Wels, 1989). The temporal trends in old water and subsurface flow contributions to streamflow are best compared during the onset of spring runoff (March 22–March 27; Figure 5) when the isotopic hydrograph separation could be made with more confidence.



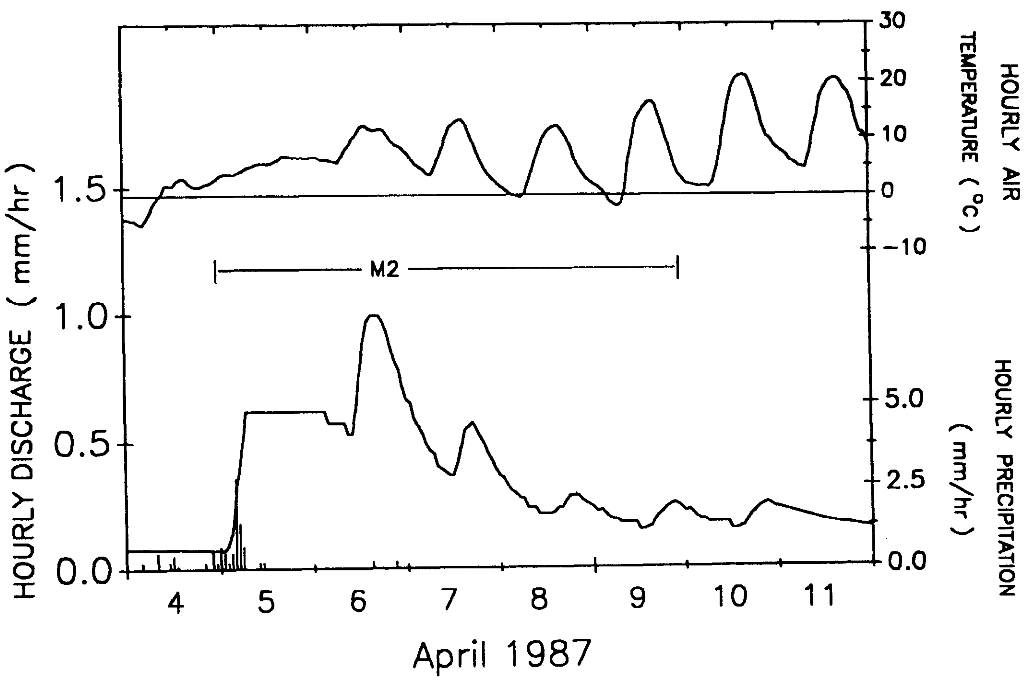




Figure 3. (a) Hydrometeorological conditions in the Plastic Lake watersheds during the first melt period. Hourly specific discharge of Plastic-108 obtained from outflow weirs. Hourly precipitation rate (vertical bar) and hourly air temperature (upper panel) observed at Plastic Lake Met Station; (b) Hydrometeorological conditions in the Plastic Lake watersheds during the second melt period. Hourly specific discharge of Plastic-108 obtained from outflow weirs. Hourly precipitation rate (vertical bar) and hourly air temperature (upper panel) observed at Plastic Lake Met Station. (An ice sheet blocked the inflow to the Plastic-108 weir for much of April 5)

Table I. Hydrological characteristics of the two melt periods M1 and M2 during the 1987 spring runoff for Plastic-108

| Melt period* | Total runoff (mm) | Peak daily/runoff† (mm day$^{-1}$) | Max. runoff rate‡ (mm hr$^{-1}$) |
|---|---|---|---|
| M1 | 118 | 21·5 | 1·68 |
| M2 | 69 | 17·7 | 0·99 |

*M1—1st melt period (March 22-April 4); M2—2nd melt period (April 5-April 14)
†Obtained from mean daily discharge records
‡Obtained from hourly discharge records

As hypothesized (hypothesis 1), the great majority of streamflow was delivered via subsurface pathways. The relative contributions of subsurface flow ($X_{SSF}$) were constant ($\approx$95 per cent) as flow changed during this five-day period (Figure 5). The only small, but significant, depressions in $X_{SSF}$ (down to $\approx$ 90 per cent) were observed on march 23 and 24 in response to pronounced daily discharge peaks (Figure 5). These small depressions in $X_{SSF}$ always occurred between $\approx 15^{00}$ and $18^{00}$ hrs, i.e. shortly before and/or at peakflow, and were likely the result of direct inputs of melt water onto the saturated areas linked to the stream channel (DPS) (see below).

The temporal trends in $X_O$ clearly contrasted those of $X_{SSF}$. The relative contributions of old water ($X_O$) showed a general decrease of 35 per cent over this five-day period. Superimposed on this trend were daily

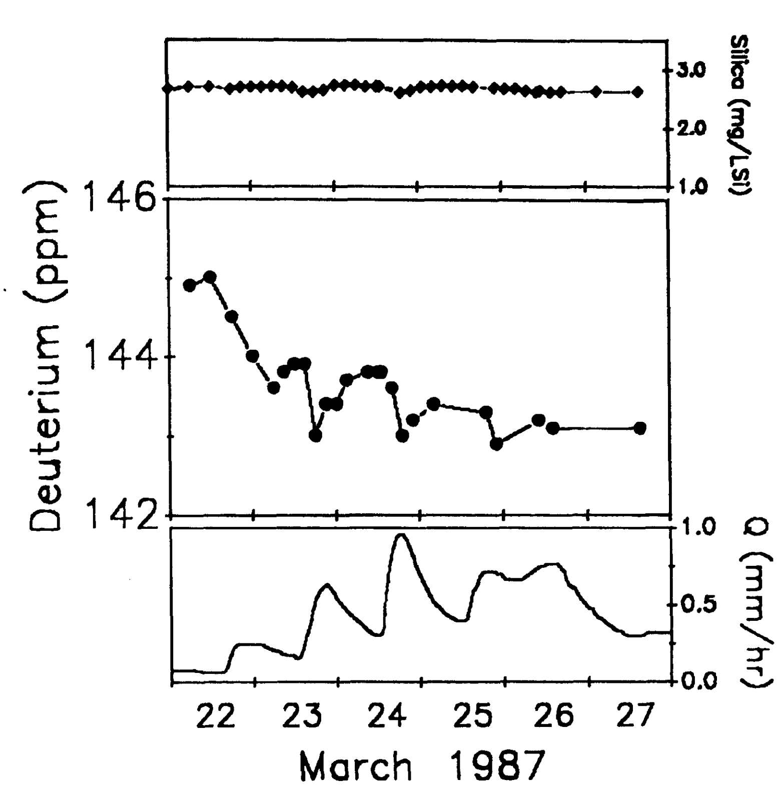



Figure 4. Daily variations in stream water deuterium and silica in Plastic-108 during diurnal fluctuations in flow. The hourly discharge ($Q$) is shown below

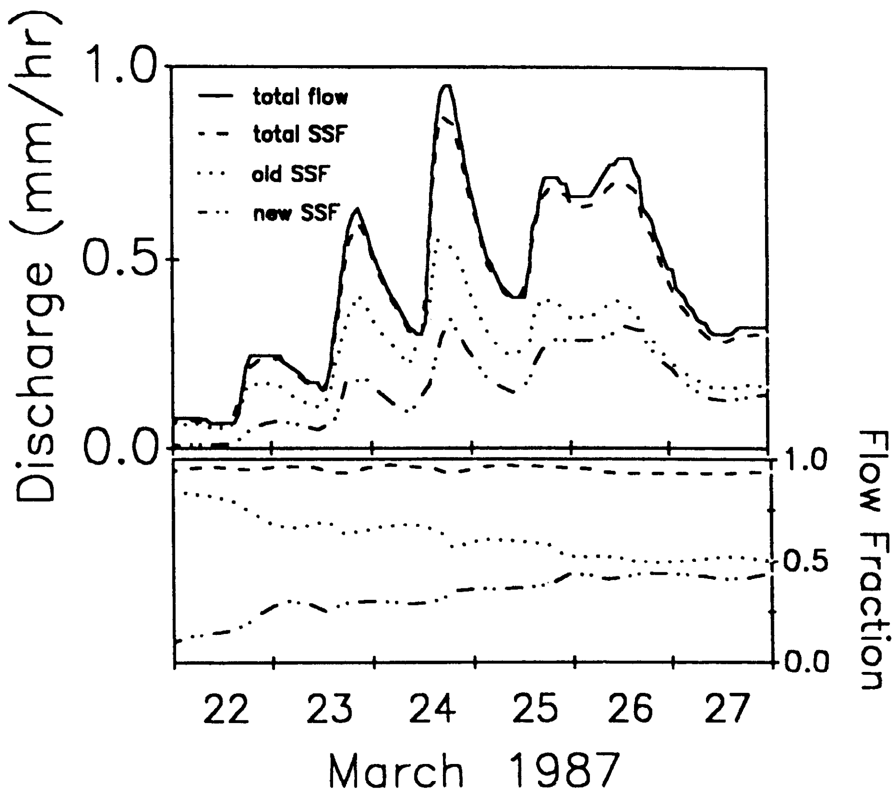



Figure 5. Chemical and isotopic hydrograph separation of diurnal fluctuations in flow, Plastic-108. The contributions of old subsurface flow, new subsurface flow and new surface flow to streamflow are shown in the upper panel. Their relative contributions to streamflow (fractions in %) are shown in the lower panel

fluctuations in $X_O$ in response to changes in flow (Figure 5). The fractions of old water were inversely related to flow with a range in variation in the order of 10 per cent.

The seasonal decrease in relative contributions of old water can be explained by larger contributions of new subsurface flow to stream flow, as outlined in hypothesis 2. The fact that new water became an important flow component even during the early melt period (Figure 5) is an indication of the small (pre-melt) groundwater volume in the study watershed, estimated to be $\sim$200 mm (Wels, 1989). Since this reservoir was even smaller than the total amount of new water available for spring runoff in 1987 (224 mm) significant contributions of new water transmitted as subsurface flow to streamflow have to be expected. This explains why the new water contribution to subsurface flow is larger in this watershed than in many others reported in the literature (e.g. Bottomley *et al.*, 1986; Rodhe, 1987).

*Subsurface flow mechanisms*

Several processes have been suggested to explain the large and rapid response of subsurface waters to an impulse of snowmelt and rainwater inputs at the ground surface. These include:
1. Groundwater ridging due to capillary fringe rise (e.g. Sklash and Farvolden, 1979; Gillham, 1984)
2. Subsurface flow through the soil matrix (e.g. Hewlett and Hibbert, 1967; Rodhe, 1987)
3. Macropore flow (e.g. Beven and Germann, 1982; Roberge and Plamondon, 1987)

All three subsurface flow mechanisms could theoretically explain the observed rapid and large discharge of subsurface waters to the stream. Hydrometric measurements were made to test the possibilities further. The following general field observations are of interest in this context: (1) seasonal runoff response was $\sim$91 per cent of melt/rainwater inputs, suggesting that a large fraction of the watershed was contributing to streamflow; (2) infiltrating meltwater gradually mixed with old soil/groundwater; and (3) subsurface flow must have been, at least for some time, in contact with the mineral soil matrix. None of these observations can rule out any of the subsurface flow mechanisms. Nevertheless, it is instructive to consider uncertainites in reconciling the various subsurface flow mechanisms with the above listed observations.

*Groundwater ridging* The groundwater ridging hypothesis offers a mechanism that can rapidly translate water stored in the stream valley into streamflow. It has been demonstrated by field measurements (Abdul, 1984) as well as computer simulations (Sklash and Farvolden, 1979). It has been used to explain the response of a number of watersheds in Quebec and Ontario (Sklash and Farvolden, 1979) and New Zealand (Sklash *et al.*, 1986). However there is insufficient storage capacity in the near-channel zone in Plastic-108 to account for more than one day of runoff at the peak of the melt season (Wels, 1989) and further, the high seasonal runoff response ratio (close to 100 per cent) indicates that the entire catchment must have been actively contributing to the stormflow.

*Subsurface stormflow through the soil matrix* Hewlett and Hibbert (1967) proposed that translatory (piston) flow through the soil matrix, in combination with an expanding channel network, could deliver water rapidly enough to contribute to stormflow. More recently, some researchers (e.g. Rodhe, 1987) have proposed that relatively rapid lateral flow takes place through the upper soil layers in response to rising water levels from below. However, simple calculations show that matrix (Darcian) flow from the side slopes can account for only a small proportion of observed stream flow in Plastic-108.

By reviewing published data for similar soil types, Shibatani (1988) estimated the maximum saturated hydraulic conductivity ($K_S$) of soils in the study watershed to be 2 m day$^{-1}$ (0·083 m hr$^{-1}$). Assuming soil depths near the base of the side slopes in Plastic-108 average 0·5 m, and the hydraulic gradient corresponds to the average slope of the valley sides (0·25), then Darcian flow per metre width of hillside, using the $K_S$ of 0·083 m hr$^{-1}$, would be 0·0104 m$^3$ hr$^{-1}$. Given that the valley is 275 m in length, the total flow from both side slopes would be 5·72 m$^3$ hr$^{-1}$. This corresponds to a runoff rate of 0·17 mm hr$^{-1}$ for the watershed, which is less than observed flow rates at the weir, which averaged 0·35 mm hr$^{-1}$ over the entire runoff season with a maximum of 1·68 mm hr$^{-1}$.

*Macropore flow* Macropore flow provides much faster flow rates than Darcian flow (Beven and Germann, 1982). Subsurface flow routed through macropores could reach the stream channel within a couple of hours, even from the most distant parts of the Plastic-108 watershed, and explain the rapid and volumetrically large contributions of subsurface flow to stream flow. Shibatani (1988) monitored water levels in shallow wells along two transects up side slopes in the Plastic-1 watershed during spring runoff in 1985. The topography and soils on these slopes are similar to those in Plastic-108. He found that a saturated layer developed at the base of the soil directly on top of the bedrock surface. This saturated zone never extended to the surface on the side slopes but thickened as a wedge toward the base of the slope. Water levels in the wells were at their maximum height at the time of peak runoff and then declined gradually but did respond to subsequent rainfall inputs. Shibatani concluded that the majority of subsurface flow was transmitted downslope through the saturated layer. We envisage this as a form of macropore flow supplied by a mixture of old and new water mixing and infiltrating down through the soil profile. The infiltration of meltwater and mixing of waters in the soil profile has been observed using isotopic analysis (Wels *et al.*, 1990a). The existence of macropore flow also is further supported by descriptions of a thin, highly weathered zone at the bedrock/soil interface in Plastic-108 (Kirkewood, 1988) and our own observations of flow over the bedrock surface in pits which we dug to install equipment in the fall of 1988. A field study using subsurface flow trenches is in progress to investigate the process further.

### *Source areas for surface flow*

In this study, surface runoff is defined as water which has very little contact with the mineral soil and therefore retains the chemical signature of the snow pack or rainfall when it reaches the stream (see Wels *et al.*, 1990b).

The chemical hydrograph separation results suggest that surface runoff accounted for only 17 mm or 8 per cent of all meltwater available for runoff during the entire 1987 spring runoff period. Based on individual separations, the proportion of surface runoff ranged from 5 per cent during low flows to a high of 18 per cent during peak flow on April 30 and March 6 (Wels, 1989). Given the relatively high infiltration capacities of soils in the watershed relative to melt and rainfall intensities, it seems probable that the only runoff pathway that could account for the surface water inputs is direct precipitation on to saturated surfaces alongside the stream channel. Although surface runoff is generated from the bare rock surfaces on the west side of the valley, visual observations suggest that this water infiltrates the soils after running off the rock and reaches the stream as subsurface flow. It is therefore excluded in our definition of surface runoff based on streamwater chemical separation.

To test the hypothesis that direct precipitation on to saturated areas (DPS) accounts for the surface flow inputs, we related the surface flow fractions from the chemical hydrograph separation to the extent of saturated areas mapped by Shibatani (1988) and also conducted a tracing experiment in the near channel zone. Given that the total runoff response for the spring period was $\sim 91$ per cent, indicating that most of the watershed contributed water, the proportion of surface water estimated by chemical separation and the proportion of the total watershed that was saturated should be in agreement.

Shibatani's field surveys during spring runoff in 1985, using the method of Dunne *et al.* (1975), showed the extent of saturated surface in the near-stream zone to range from 8 per cent of the total watershed area at low flow to 20 per cent at high flow (Shibatani and Taylor, 1988). This is quite similar to the range in surface water contributions in 1987 of 5 per cent to 18 per cent, based on chemical separations. Shibatani found a positive correlation between stream discharge and the extent of surface saturation, similar to that described elsewhere (e.g. Dunne *et al.*, 1975; Taylor, 1982; Myrabo, 1986). The variable extent of saturated surfaces thus can account for the varying contributions of surface flow during events and through the entire runoff period, as postulated in hypothesis 3.

The relationship between surface flow contributions and discharge for three melt events is illustrated in Figure 6. The surface flow fractions show a small but significant anticlockwise hysteresis in all three melt events, i.e. a lag in the recovery of $X_{SF}$ towards low pre-storm values during flow recession (Figure 6). This hysteretic pattern of $X_{SF}$ may reflect the slow retreat of saturated areas during flow recession which would maintain high contributions of surface flow (i.e. DPS). The saturated zones would be maintained by continuing drainage from the side slopes after peak discharge. Such a lag in the retreat of saturated areas during flow recession has been observed in many basins in Vermont (Dunne, 1978). Alternatively, the melt of the snowpack on the saturated areas of the sheltered stream valley may lag behind that of the remainder of the catchment. This would result in higher $X_{SF}$ during flow recession even with a constant fraction of saturated areas. A detailed monitoring of the melt rates within the stream valley combined with chemical hydrograph separation would be necessary to determine whether such a hysteresis in the relationship between the extent of near-stream saturated areas and streamflow does occur.

The hysteretic loops are superimposed on a seasonal trend in the surface flow fractions, i.e. a general increase in $X_{SF}$ from the early event on March 23/24 to the final melt period (M2) (Figure 6). Similar to the daily trends, this could either reflect a slow increase in saturated areas during spring runoff or result from higher melt water inputs to saturated areas in subsequent melt events. If the fraction of surface flow represents the fraction of near-stream saturated areas in the watershed, it follows that only $\approx 8$ per cent of the watershed, i.e. less than 50 per cent of the maximum near-stream saturated areas, was saturated on average over the entire spring runoff. During the early melt period, the saturation of the near-stream zone would have been smaller, i.e. only 2 per cent of the watershed or 10 per cent of the maximum near-stream saturated area.

*Tracing direct melt water inputs*

The comparison of saturated area mapping and chemical hydrograph separation suggested that: (1) surface flow traced by silica equals DPS and (2) saturated areas producing DPS are confined to a small near-stream zone of variable size. A tracing experiment which was conducted during the spring runoff of 1987 in the near-stream area of Plastic-108 provides an additional test of these hypotheses.

On March 18, a solution of lithium bromide was sprinkled onto the snowpack along a 6 m strip on both sides of the stream channel along the full length of the stream valley. The total area covered was $\approx 0{\cdot}13$ ha or 4 per cent of the watershed. During the main runoff period, the recovery of lithium bromide in streamflow at

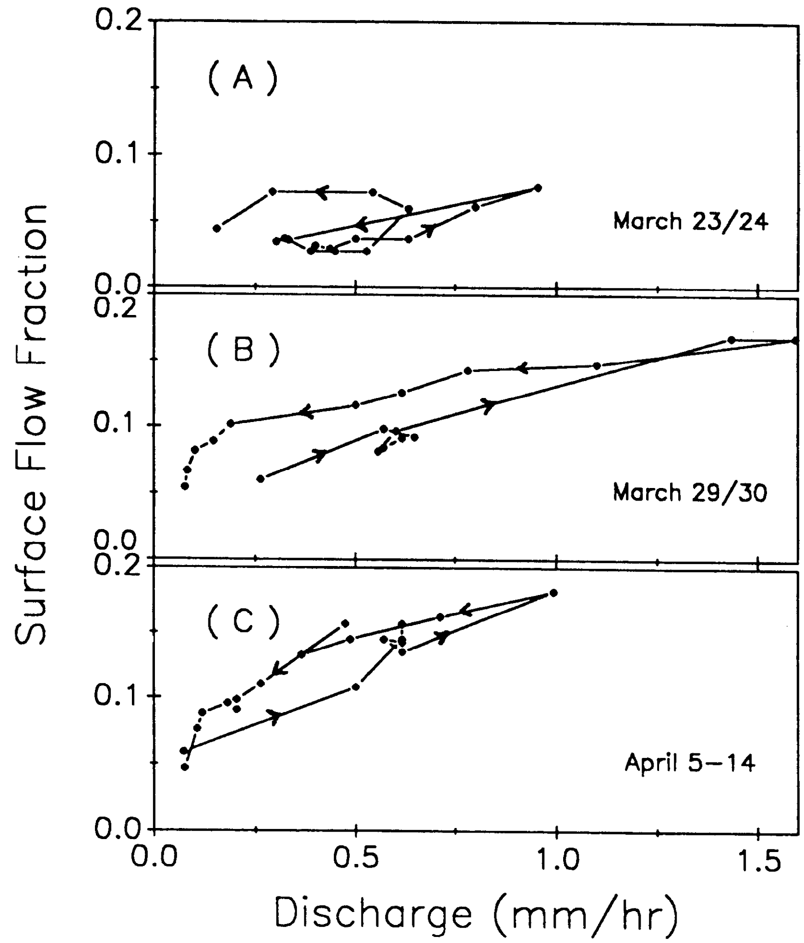



Figure 6. Surface flow fractions versus stream discharge for three melt events in Plastic-108. The upper panel (A) shows an early radiative melt event on March 23/24. The middle panel shows the intense rain-on-snow event on March 30 (B). The lower panel shows the second melt period on April 5–14 (C)

the outflow weir was monitored. Based on the above two hypotheses, it was expected that the tracer mass in streamflow would be proportional to the volumetric contributions of surface flow from chemical hydrograph separation.

The masses of lithium and bromide recovered in the stream showed very similar diurnal fluctuations for the first three daily melt events (Figure 7). The tracer masses quickly increased on the rising limb of the daily hydrographs, peaked close before or at peakflow, and then decreased rapidly to baseflow concentrations. The peak of tracer masses always coincided with peaks in surface flow supporting the hypothesis that surface flow traced by silica represents DPS (Figure 7). However, the magnitudes of the tracer peaks were not proportional to the volumes of surface flow (Figure 7). This discrepancy was probably largely a result of preferential elution of the tracers from the snowpack.

The lithium concentrations in snow cores from the traced near-stream area decreased considerably during this early melt period (Table II), suggesting that the tracer concentrations in the melt water decreased over time. An enrichment of the first portion of melt water leaving the snowpack has been observed in field and laboratory studies for all major ions (Johannessen and Henriksen, 1987). Preferential elution of the tracer was probably pronounced in the present study due to melt-freeze cycles (Colbeck, 1981) and the high tracer concentrations in the top layers of the snowpack which melted first (Table II). Unfortunately, the tracer

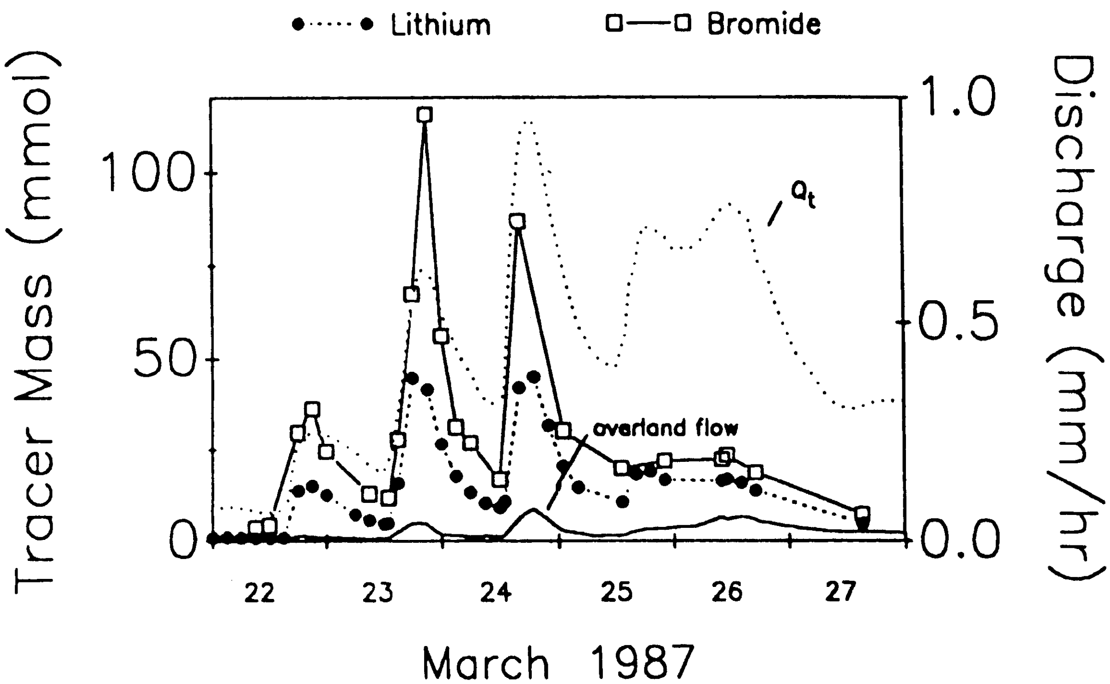



Figure 7. Tracer masses recovered in streamflow during the early melt period. Hourly total discharge (dotted line) and surface flow (solid line) are shown for comparison. The tracer lithium bromide was applied to the near-stream snowpack on March 18. The total recovery of lithium and bromide until March 30 was 45 per cent and 87 per cent, respectively

concentrations in melt water were only measured from changes in the snowpack precluding a rigorous quantitative comparison of traced melt water inputs (DPS) and surface flow traced by silica.

The recovery of lithium and bromide provided a first clue to the extent of saturation in the near-stream zone. The masses of bromide recovered were consistently greater than those of lithium (Figure 7). By March 30, 87 per cent of bromide had been recovered compared to only 45 per cent lithium. The total recoveries over the entire spring runoff were 101 per cent and 59 per cent for bromide and lithium, respectively, This suggests that bromide was essentially conservative whereas lithium was taken up due to contact of the traced melt water with the soils. This was verified by elution tests, using packed columns of soil from the near-stream area, with recoveries of 98 per cent and 68 per cent for bromide and lithium, respectively (Wels, 1989). The

Table II. Vertical concentration profiles of lithium in the snowpack of the near-stream area traced. Lithium concentrations in mmol $L^{-1}$

| Height above ground (cm) | March 19 | March 23 | March 25 |
|---|---|---|---|
| 60–51 | 183 | — | — |
| 50–41 | 30 | 38 | — |
| 40–31 | 18 | 64 | 20 |
| 30–21 | 12 | 34 | 18 |
| 20–19 | 1 | 12 | 8 |
| 10–0 | 1 | 8 | 6 |
| Average* | 49 | 31 | 13 |

*Estimate assumes uniform density in snowpack.

preferential retention of lithium in the field experiment suggests that some of the melt water had contacted the soils, unless lithium was retained within the stream.

Apparently not all of the near-stream area had been saturated. The ratio of lithium recovery (reactive) to bromide recovery (conservative) should give an estimate of the extent of saturation in the near-stream zone. Assuming complete retention of lithium in subsurface flow and no retention of lithium in surface flow, $\approx 50$ per cent of the traced near-stream area (2 per cent of total watershed area) would have been saturated over the period March 18–March 30. If only 66 per cent retention of lithium in soils is assumed, the saturated area estimate reduces to 1·5 per cent of the total watershed area for the entire runoff period.

These values are consistent with estimates of saturated areas based on chemical hydrograph separation. Surface flow traced by silica comprised 2 per cent (or 1·4 mm) of the total runoff during the five day period shown in Figure 7. This suggests an 'event mean' extent of saturated areas of $\approx 2$ per cent of the total watershed area, assuming homogeneous melt rates within the basin. The two independent estimates of the extent of saturated areas for the early melt period (2 per cent) were reasonably close to those surveyed directly by Shibatani (1988) for post-melt baseflow (8 per cent).

Assuming that bromide was conservative and lithium was taken up by the soils at a constant rate, the ratio of lithium to bromide should be directly proportional to direct melt water inputs to the stream. The use of a ratio of the two tracers as opposed to absolute masses has the advantage of being fairly independent of temporal variability in melt water tracer concentrations. Preferential elution should be similar for both ions (e.g. Davies *et al.*, 1987; Johannessen and Henriksen, 1978) leading to a constant molar ratio of the two tracers in melt water over time. The above hypothesis was tested by correlating the Li/Br ratio to contributions of surface flow traced by silica.

The Li/Br ratio showed a highly significant ($p < 0.001$; $R^2 = 0.88$) logarithmic relationship with surface flow (Figure 8). As hypothesized, higher relative contributions of lithium indicated increased routing of melt water over saturated areas into the stream (DPS). A notable exception was observed on peakflow of March 23 when the exceptionally high recovered mass of bromide resulted in a very low Li/Br ratio (Figure 8). This was probably a result of contamination of the sample.

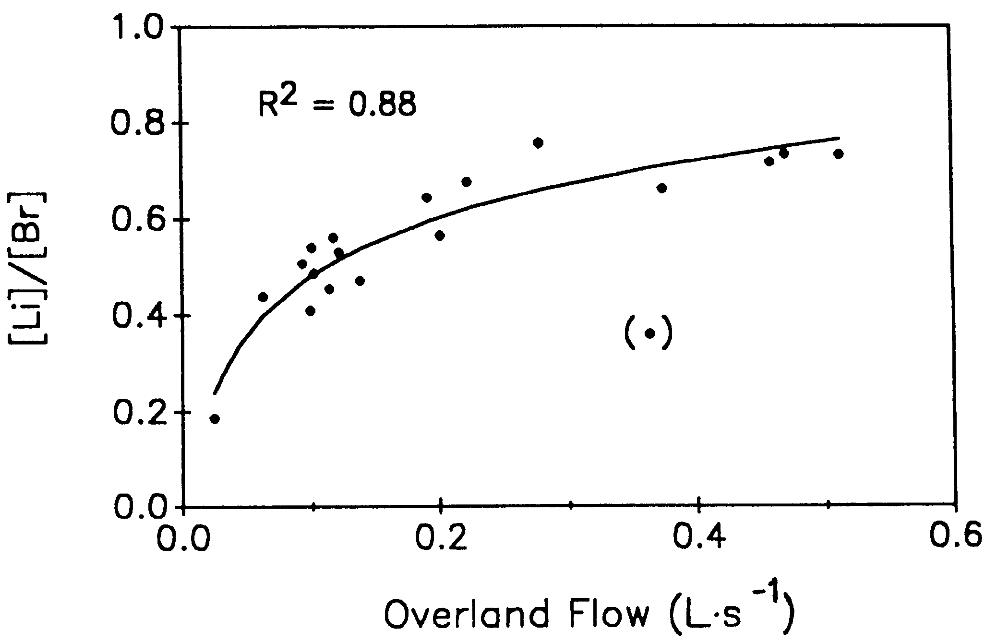



Figure 8. Ratio of [Li]/[Br] versus surface flow for the period March 22–March 27. The sample from peakflow on March 23 (value in brackets) was not used for regression analysis

If the Li/Br ratio is an indicator for direct melt water runoff from saturated areas (DPS) it may also be used to examine temporal trends in the extent of saturated areas. However, at this point the use of a Li/Br ratio for this purpose can only be speculated upon. The range in the Li/Br ratio suggests that the saturation in the near-stream zone changes over time. This is consistent with observations based on chemical hydrograph separation. The almost constant Li/Br ratio $\approx 0{\cdot}7$ for surface flow contributions greater than $\approx 0{\cdot}3$ L $s^{-1}$ (Figure 8) could indicate a steady-state condition where the entire spiked near-stream zone was saturated and produced DPS. Any additional volume of surface flow would originate from saturated areas outside of the traced near-stream zone. For a further quantification of the extent of saturated areas, a calibration of the Li/Br ratio to field mapping would be required.

Preliminary results presented here suggest that the extent of saturated areas may be investigated either directly by using a combined conservative/reactive tracer such as lithium bromide or indirectly by interpretation of chemical hydrograph separation. Both types of measures indicate that: (1) near-stream saturated areas were only in the order of 2 per cent of the total watershed during the early melt period; and (2) near-stream saturated areas showed small, but significant changes over the course of an event, as proposed in our earlier hypothesis 3. The use of a snow lysimeter in the stream valley would greatly aid in the attempts to examine temporal trends in saturated areas by providing (1) melt rates needed to estimate saturated areas from chemical hydrograph separation and (2) tracer concentrations in melt water needed for chemical tracing of DPS and saturated areas.

## CONCLUSIONS

The results of the three-component hydrograph separation and direct mapping and tracer experiments on saturated areas suggest the following combination of runoff mechanisms in headwater watersheds on the southern Canadian Shield. Water is supplied to the near-channel zone by subsurface flow from the side slopes, probably mainly as saturated macropore flow along the impermeable bedrock surface underlying shallow soils. This water consists of a mixture of pre-melt and event water, with the relative proportion of the latter increasing over the course of the melt period. Water in the near-channel zone makes its way to the stream partly as subsurface flow across the channel boundary and partly as saturation overland flow (SOF), described originally by Dunne and Black (1970). SOF consists of return flow from the side slopes (old and new water) combined with direct precipitation onto the saturated surfaces (DPS). Of all constituents, only the DPS has the chemical signature of surface runoff.

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