Snowmelt Runoff Processes in a Headwater Lake Catchment, Subarctic Canadian Shield

CORRINNE MIELKO[1] AND MING-KO WOO[1]

ABSTRACT

During the snowmelt season, surface runoff into northern Canadian Shield lakes are governed by the distribution of snow and its ablation, storage in catchment slopes and valleys, and flow delivery from uplands, bottomlands and from upper lakes. For the lake studied, time lags exist between the commencement of snowmelt and the arrival of runoff to the lake, with evaporation from the uplands and bottomlands consuming some of the meltwater produced. Inflow from an upper lake is delayed. With all these inputs, the lake becomes flooded, first forming a moat, followed by an expansion of the open water areas which are subject to evaporation loss. Lake outflow begins when water level rises above the outlet threshold. For the snowmelt season, outflow constitutes 40 percent of total melt and rainfall. The lake itself is a major storage that continues to maintain outflow until the lake level drops below the threshold in early summer, after which flow ceases. Snowmelt is the main period when runoff is generated from the headwater catchment to replenish storage depleted in the previous summer, and to produce flow along the Shield valley that contains a chain of lakes.

Keywords: snowmelt, Shield hydrology, flow threshold, lake storage, runoff generation

INTRODUCTION

The Precambrian Shield makes up about one-third of the North American landmass (Shilts et al., 1987). A large part of the Shield in central and eastern Canada was scoured by Pleistocene glaciation to expose gneissic and granitoid rocks, with glaciofluvial and lacustrine deposits partially infilling many depressions in the bedrock. The general landscape consists of bedrock outcrop, soil-filled valleys, lakes and wetlands (Spence and Woo, 2003). The Shield has myriads of lakes of different dimensions, from small ones measuring <1 km^2 to the enormous Great Slave and Great Bear lakes. Northern lakes acquire an annual ice cover, while the lands surrounding them are underlain by seasonal frost and/or permafrost. The relatively impervious, sometimes well-fractured crystalline rocks of the uplands often shed runoff effectively and runoff-plots on bedrock slopes have yielded a range of ratios, from under 0.05 to over 0.8 (Landals and Gill, 1972, Spence and Woo, 2002). For areas with lakes, the runoff ratios are not well established.

Runoff generation and groundwater flow in the Canadian Shield has been studied, though very little attention has been placed on the influence of small lakes on the flow in headwater catchments. Spence and Woo (2002) note the importance of surface ponding in delaying runoff and increasing evaporation and infiltration. While crystalline rocks usually shed snowmelt and rainfall efficiently, infiltration has been found to be much larger than expected when a well connected fracture network exists (Spence and Woo, 2002; Thorne et al., 1999). Thorne et al. (1999) found hydraulic conductivities in the Shield bedrock to vary between 7x10^{-10} m/s to 2x10^{-5}

[1] School of Geography and Earth Sciences, McMaster University, Hamilton, Ontario L8S 4K1.

m/s. Runoff from uplands often enters the soil-filled valleys along the interface between bedrock and the soil (Allan and Roulet, 1994; Buttle and Sami, 1992; Peters et al., 1995). The flow is then modified by the valley storage, following the fill and spill mechanism in which runoff from the slopes has to satisfy storage demands of the valley before outflow can commence (Spence and Woo, 2003). Flows along the valley can be intermittent due to seepage loss along the flow path.

Snowmelt period in the subarctic usually produces high flows (Woo, 2000). The conventional role of lakes is to store and retard inflows so that outflows are delayed and modulated. For the Shield environment, Spence (2000) noted that only 7% of snowmelt water was delivered at the outlet of a small lake (area 0.043 km^2) situated in a 0.575 km^2 basin. Such a low runoff ratio was attributed to the need to satisfy the lake storage deficit, possibly due to continuous drawdown in the previous year. FitzGibbon and Dunne (1981) also noted that the snowmelt runoff was delayed by storage in a complex chain of lakes in Schefferville, Quebec.

The processes of runoff generation and delivery to a headwater Shield lake during the snowmelt period have not been analysed in detail. It is the objective of this project to investigate systematically the manner in which runoff is delivered to a lake in a headwater catchment and to understand the role of Shield lakes in streamflow generation during the snowmelt season. Pertinent to this investigation are several important questions. (1) What mechanisms control streamflow generation? (2) When does runoff occur in relation to snowmelt? (3) How much streamflow is generated by snowmelt? Results from this study would enhance our understanding of the manner in which outflow is produced from lakes which has implications for future modeling of water balance and streamflow from Precambrian Shield catchments in a relatively dry northern environment.

STUDY AREA

This study was carried out in a headwater catchment (area 0.56 km^2, 62^{o}35’N; 114^{o}22’W) located 15 km north of Yellowknife, Northwest Territories, Canada. It is a basin in the subarctic Canadian Shield, containing Shadow Lake with a surface area of 0.09 km^2. The catchment can be subdivided into bedrock uplands, vegetated bottomlands and the lake (Fig. 1). The upland is occupied by crystalline bedrock that is fractured to varying degrees. The rock surface has been sculpted into shallow depressions some of which have thin soil covered by lichens and other vegetation including creeping junipers that grow along the cracks, scattered stands of Jack Pine and occasional aspen and birch. The bottomlands are soil filled, usually an organic layer overlying mineral soils, with permafrost being prevalent. Drier parts of the bottomlands have an open canopy of white spruce, birch, black spruce and tamarack, an understory of willow and birch shrubs and a ground cover of mosses, club lichens, bear berry and cranberry. Saturated bottomlands are wetlands with moss, grasses, sedges, Labrador tea, and possible open canopy of tamarack and black spruce.

The area has a subarctic continental climate. Mean January temperature falls to –29^{o}C and mean July temperature reaches 16^{o}C. Annual precipitation is 267 mm for Yellowknife, about 100 mm of which comes as snowfall (Environment Canada, 1993).

METHODS

A snow survey was performed in early April 2004, employing 540 depth measurements and 50 density (using Meteorological Service of Canada snow sampler) determinations throughout the Shadow Lake and its nearby areas. The snow survey points were classified according to terrain types and the average snow water equivalent (SWE) from each land cover was obtained. Snow boards with an area of 0.5x0.5 m^2 were deployed after the snow survey to collect additional snow fall. Shadow Lake basin was subdivided into 100x100 m^2 grids and the fractions of each terrain type within the grids were determined. Mean grid SWE was calculated as the fraction of each terrain type multiplied by its respective mean SWE. Snow ablation lines were installed at each terrain type to obtain the daily ablation rates using the method outlined by Heron and Woo (1978).

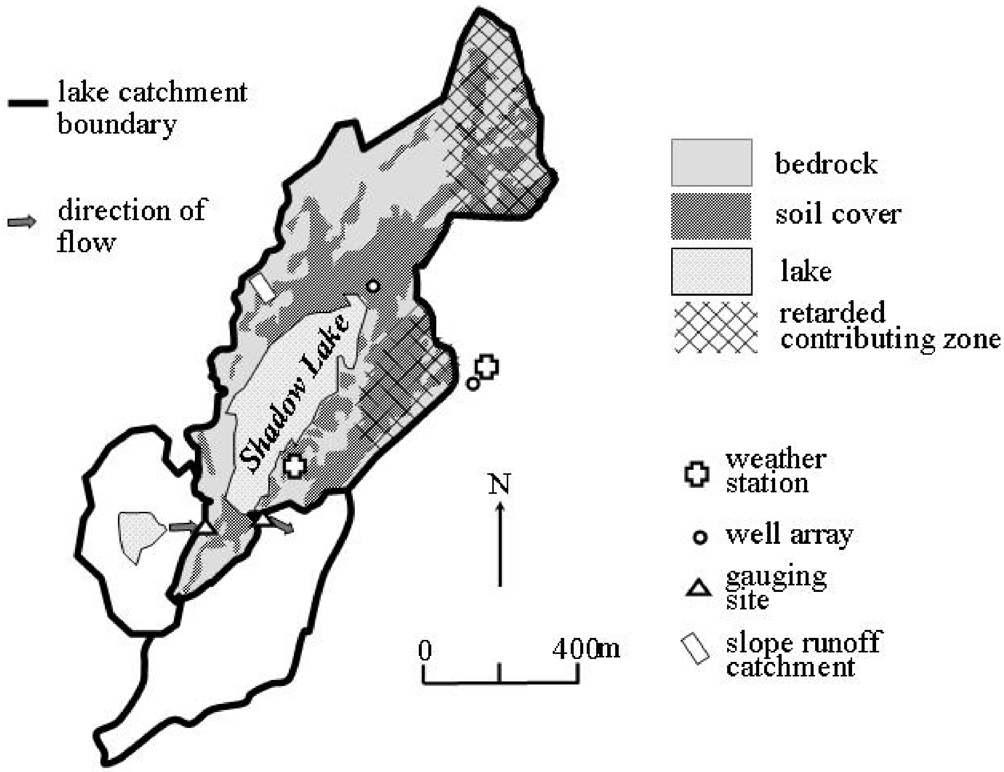

Figure 1: Shadow Lake and its catchment: distribution of land surface types and instrumentation.

The spatial pattern of melt was obtained by subtracting the daily ablation rates for various terrain types from the SWE values in each grid, until it becomes bare. Rainfall was measured using a tipping bucket rain-gauge and supplemented by five manual gauges spread across the catchment to obtain the spatially weighted average.

Meteorological towers set up in and immediately adjacent to the catchment provided data for the calculation of evaporation. Half hourly evaporation from the bedrock and the vegetated bottomland was obtained by the Bowen ratio energy balance approach (Ohmura, 1982), using measured net radiation (NR Lite Net Radiometers), air temperatures and relative humidity at two heights (Vaisala HMP 35c and 35cf, housed in multiplate radiation shields), and ground temperatures (TidbiT temperature sensors) for each environment. Lake and pond evaporation was calculated at half hourly intervals by the Priestley and Taylor (1972) method, using net radiation and air temperature measured over the open water surface of a nearby lake (unpublished data from Bob Reid, 2004), together with Shadow lake and pond water temperatures (TidbiT temperature sensors). Shadow Lake acquired a seasonal ice cover that reached a maximum thickness of 0.6m at end of the 2003–04 winter. Photographs were taken of the decaying lake ice cover during the spring to estimate the changing ice free fraction. Lake and pond water levels and groundwater levels at three wells were measured using Remote Data Systems Ecotone Water Level Monitoring Systems. Channel flow into and out of the lake and out of the experimental plot were gauged periodically by converting the lake stage into discharge using empirically determined rating curves.

An experimental slope of 5300 m^2 was used to study snowmelt runoff delivery from the upland. This slope extends from the bedrock upland to a bowl shaped depression at the base which captured runoff from the entire slope. The depression has a poorly developed mineral soil underlying a thin organic layer. It has an outlet where a weir was installed to gauge the outflow from the depression. For this pond, water level change was monitored half hourly using the above Ecotone Water Level Monitoring System and averaged daily, while the areal extent of the water body was mapped.

Two transects of groundwater wells were installed, one in a wetland and the other in relatively dry bottomland and measured manually with the exception of the three automatic systems systems. Each well was a perforated pipe of either 35 mm or 48 mm diameter. Hydraulic conductivity was obtained by pumping test described by Luthin (1966). Frost table along these transects was measured by pounding a steel rod into the ground until frozen soil was encountered.

Unless otherwise stated, the magnitude of all hydrologic variables will be reported in this paper in volumetric units (i.e. m^3).

RESULTS

Snow survey carried out in early April 2004 enabled mapping of snow distribution and evaluation of total winter accumulation. Additional snowfall that occurred after the survey was measured and added to the survey results. Figure 2 shows that most of the snow was captured in the low lying areas, whereas the uplands had lower but more variable snow accumulation due to exposure to drifting and microtopographic variations. Average snow water equivalent (SWE) for the bedrock, valley, wetland and lake were 92, 105, 102, and 107 mm, respectively. Weighting these values by areas occupied by the terrain types yielded a pre-melt snow storage of 55,000 m^3 SWE.

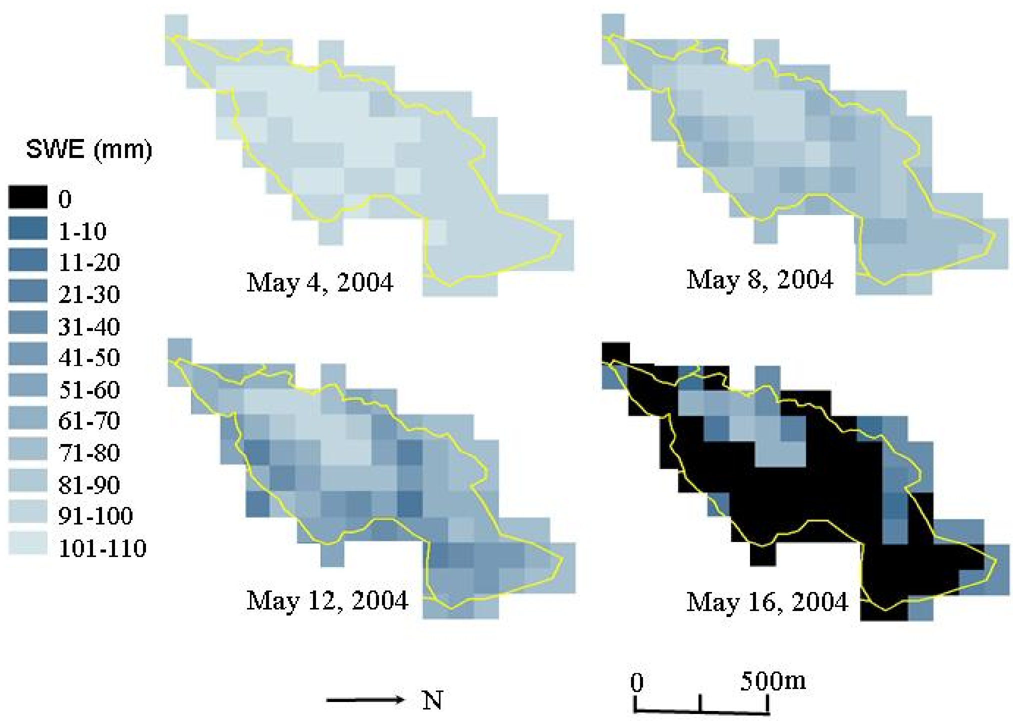

Figure 2: Changing snow distribution pattern: on May 4 (initial snow condition), May 8, May 12 and May 16.

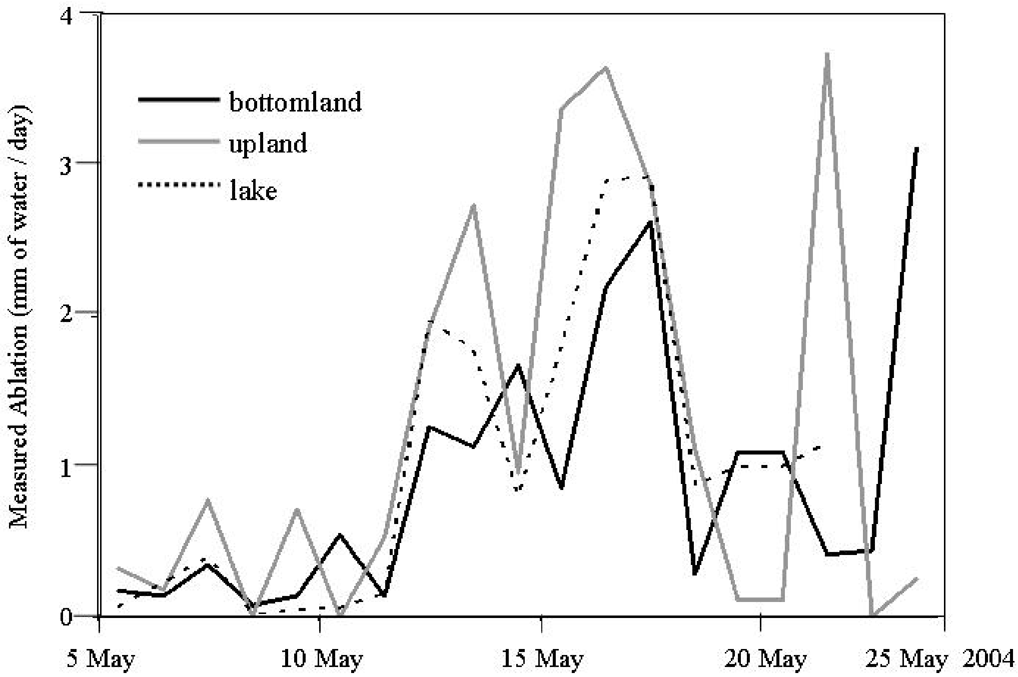

Figure 3: Daily snow ablation on three types of terrain, including upland, bottomland and lake.

Snowmelt commenced on May 5 and daily snow ablation for each of terrain type is presented in figure 3. To verify the efficacy of the ablation measurements, several transects (strips of land) in the basin were repeatedly photographed. There was a good match between the snow free areas as photographed and as calculated using the ablations measurements. The ablation values were used to produce daily gridded maps of the residual snow cover (figure 2). Up to May 12, the snow cover was relatively continuous. Between May 12 and May 14, ablation depleted the snow from most of the eastern catchment. This can also be seen in the changing snow free area in the basin (figure 4). By May 15 only the lake and major valleys retained residual snow while over 90% of the basin snow cover disappeared by May 20. Field observations indicated that the basin became completely snow free on May 24.

The weighted daily snowmelt and the changing fraction of snow free area in the basin are shown in figure 4. Initial snowmelt proceeded at a moderate rate yielding roughly 3000 m^3/day. May 10 had low snowmelt of 2500 m^3 due to a cold spell. Accelerated melt occurred on May 14 reaching almost 14,000 m^3. High melt rate of ~8000 m^3 continued on the next day, rapidly depleting the residual snow cover. A reduction of the daily melt contribution after May 15 is attributed not to a lower snowmelt rate but was due to a rapid increase in the snow free area (figure 3).

Rain fell on May 30, June 2 and 6, after all the snow had disappeared. Total rainfall amounted to 6300 m^3 (figure 4). For the entire study period a combined melt and rainfall of 61600 m^3 reflects the relatively dry shield environment.

RUNOFF TO THE LAKE

Runoff produced by snowmelt on upslope bedrock areas is either conveyed directly to the lake or seeps into the soil of the bottomlands (Spence and Woo, 2003). The soil covered bottomlands themselves also receive meltwater directly from the snow within these areas. This water can reach

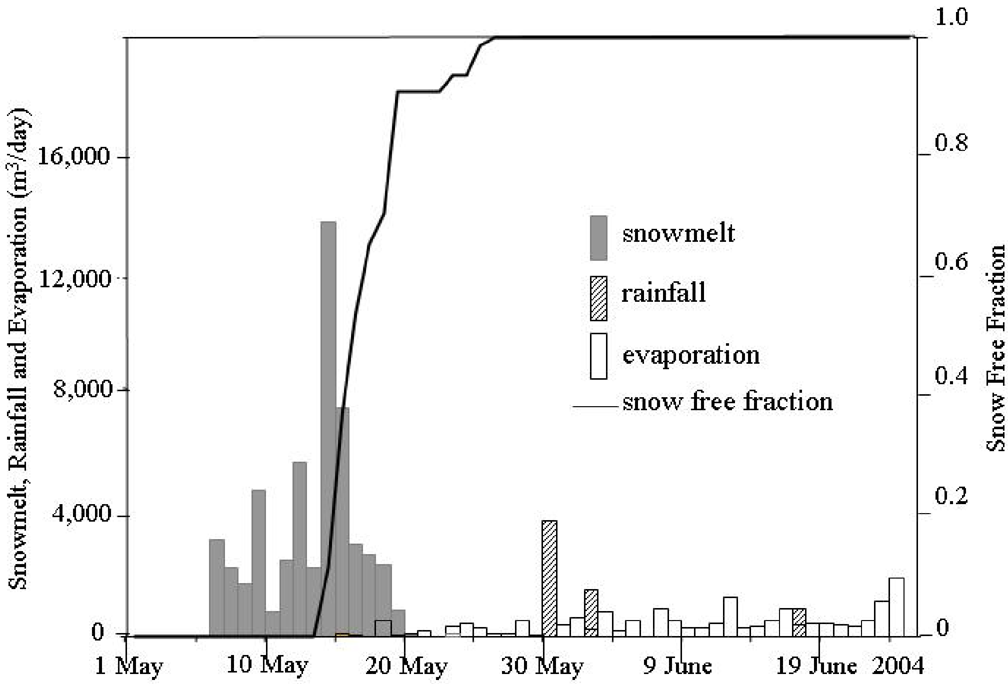

Figure 4: Daily snow melt, snow-free fraction, rainfall and evaporation from land surface.

the lake quickly as overland flow or is discharged gradually to the lake as ground water. Some of the meltwater produced, however, may not run off immediately but is retained in isolated troughs in the upland. Field observations confirmed that the water in these isolated troughs either infiltrated or was lost to evaporation. These zones do not contribute direct runoff to Shadow Lake and are considered to be areas where runoff delivery to the lake is retarded. A final source of water comes from an upper lake which drained for a short period into Shadow Lake through an ill-defined channel.

Runoff from Upland

Result from the experimental slope permits assessment of runoff contribution from a typical upland slope. The bowl shaped depression at the foot slope collected slope runoff. This created a pond with its surface area expanding as the pond level rose. Pond storage increase was pronounced immediately after snow melt. There was an outlet from the depression which commenced flowing six days after the initial pond level rise (on May 25), yielding a maximum outflow of >10m^3/day on the following two days. Outflow declined afterwards but was revived by rainfall before it finally ceased on June 6. Evaporation of the pond water increased as the pond area expanded but diminished when the pond shrank. Using these results, daily upland runoff (Q_{upland}) was evaluated by

$$Q_{upland} = \Delta S_{pond} + Q_{pond} + E_{pond.} \qquad (1)$$

Figure 5 shows the daily values of change in pond storage (ΔS_{pond}), pond outflow (Q_{pond}) and evaporation (E_{pond}). Daily pond storage change from day t to day t+1 was obtained by

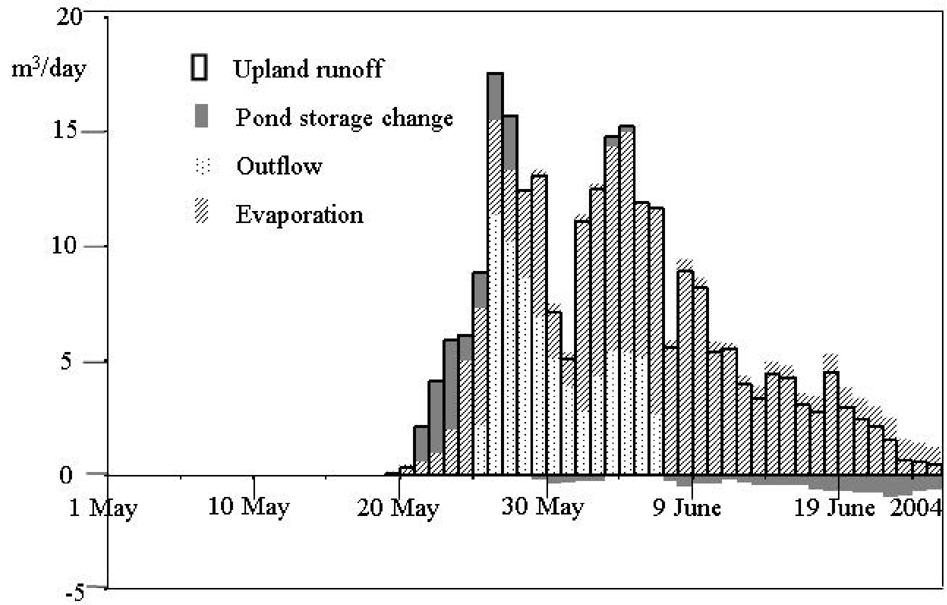

Figure 5: Estimated daily upland runoff from an experimental plot, calculated as the sum of pond storage change, pond outflow and evaporation.

$$\Delta S_{pond} = (H_{t+1} - H_t)\,(A_{t+1} + A_t)/2 \tag{2}$$

in which H and A are the elevation and the surface area of the pond, both of which changed from day to day. Over the entire period between initial infilling and complete drying of the pond, total upland runoff as calculated by eq. 1 was 259 m^3, equivalent to 49 mm per unit slope area.

Runoff from Bottomlands

The presence of seasonal frost and permafrost hindered the infiltration of melt water causing overland flow to appear in parts of the bottomland during the final phase of snowmelt. Ground thaw proceeded rapidly after the snow disappeared, allowing meltwater percolation and a descent of the water table. Measurements of the frost and water tables along two lines of wells showed that the saturated layer was relatively thin (in the order of centimetres) (Figure 6). Darcian flow across a transect line was calculated by

$$Q_g = \sum K\,(\Delta h_i / \Delta L_i)\,HS_i w_i \tag{3}$$

with Q_g being ground water flow obtained by summing the flow from different sections (i) across the transect. K is the hydraulic conductivity which was 0.1 m/day for the relatively dry site and 2.4 m/day for the peaty wetland soil; Δh is water table elevation above the lake shore at a distance of ΔL from the ground water well so that the hydraulic gradient is approximated by Δh / ΔL; HS is thickness of the saturated zone, being the depth difference between the water table and the frost table; and w is the width of each section for which flow is calculated.

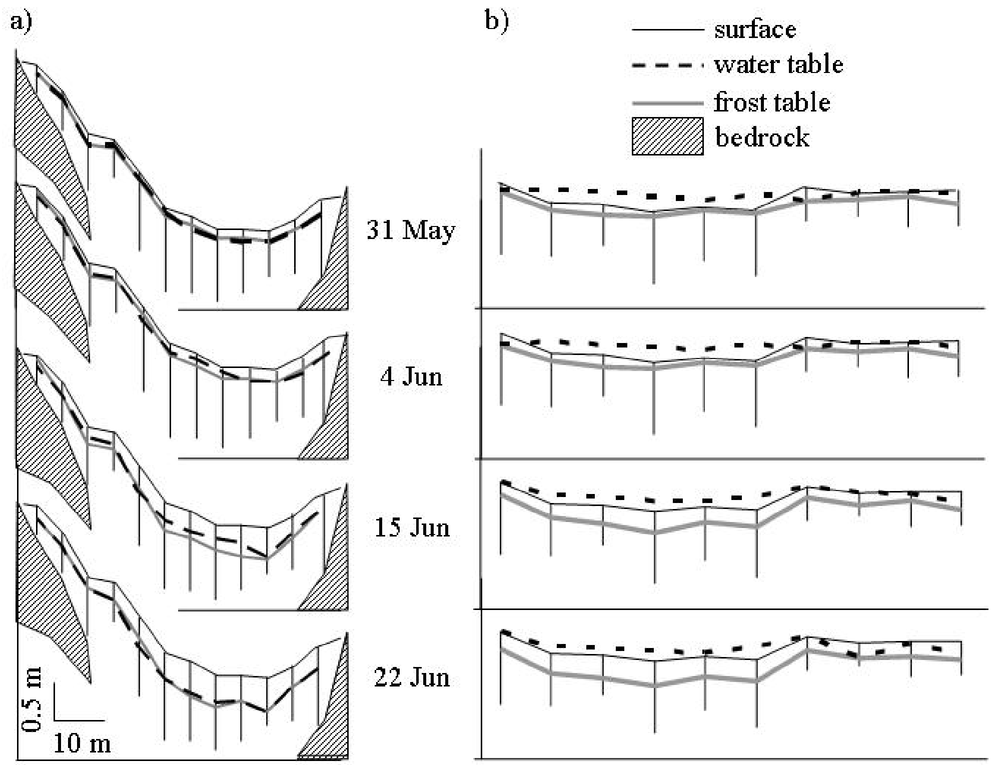

Figure 6: Water table and frost table positions on selected days, along a) a relatively dry bottomland and b) a wetland transect. Vertical lines indicate location of wells.

For the period May 14 to June 24, matrix flow totaled 0.3 m^3 across a width of 60 m in the relatively dry valley. Thus ground water as matrix flow from the valleys was likely to be minimal. On the other hand, a number of soil pipes were observed in the peat and they can convey quick flow down the valley (Carrey and Woo, 2000). Wetlands in the bottomland zone remained inundated throughout the spring (Fig. 6), delivering surface and subsurface flows directly to the lake. For the period May 18 to June 24, matrix and surface flow totaled 22 m^3 across a width of 100 m in a wetland adjacent to Shadow Lake.

Inflow from an upper lake catchment

A small catchment area 0.12 km^2 containing a lake (area of 0.01 km^2) began delivering water to Shadow Lake. Initially the runoff came from its catchment slopes because the lake had not yet started its outflow. On May 25, lake outflow began, rising to a peak on June 4 and then declined until it dropped to a trickle after June 20. The total amount delivered from this catchment reached 7800 m^3. It is significant to note that the beginning of outflow from this upper lake was two days later than when outflow commenced from Shadow Lake. This indicates that within a complex chain of lakes, the response to snowmelt runoff can vary from one lake to the other.

Land evaporation

Evaporation from the land reduces runoff available to the lake. For a snow free site, there was an increasing trend in evaporation as the days advanced towards the summer solstice. For the land area as a whole, these evaporation rates have to be adjusted by the fractional catchment area that becomes snow free. Evaporation loss was negligible when the snow was extensive over the basin. As snow cover diminished, land evaporation was enhanced by an increasing snow free fraction. Following the disappearance of snow, daily evaporation averaged 2 mm/d for the month of June. This value lies between those reported by Spence and Rouse (2002) for a wet (3 mm/d) and a dry

(0.9 mm/d) landscape. For the entire study period of May 1 to June 24, total land evaporation amounted to 17,000 m^3.

LAKE HYDROLOGY

Runoff that reaches the lake is partly lost to evaporation and partly held in storage to raise the lake level. When lake level rises above the outlet lip or threshold, outflow commences. Should lake level drop below the outflow threshold, streamflow will cease. In this way, flow connectivity is controlled by storage through fluctuations in the lake level.

Lake Ice and Evaporation

Arctic and subarctic lakes are invariably covered by ice at the end of winter. The presence of lake ice cover limits evaporation loss until open water condition develops. In 2004, meltwater first reached the lake, slushing the snow along the rim of the lake. Next, the shorefast ice began to melt, creating a moat of open water on May 18 as the ice floated free from the shore. Progressive attrition of lake ice increased the fraction of open water in the lake until the ice was fragmented and finally dissipated. The lake ice decay process is similar to that described by Heron and Woo (1994). Evaporation occurred at the open water portion of the lake. Evaporation loss from the lake increased as the fractional ice cover diminished (figure 7). For the period after the loss of the ice cover (June 6–24), lake evaporation averaged 4.8 mm/d which compares favourably with the average rate of 4.5 mm/d obtained at Skeeter Lake when the latter became ice-free (Spence, 2000). Over the entire period of study, Shadow Lake evaporation totaled 10,000 m^3.

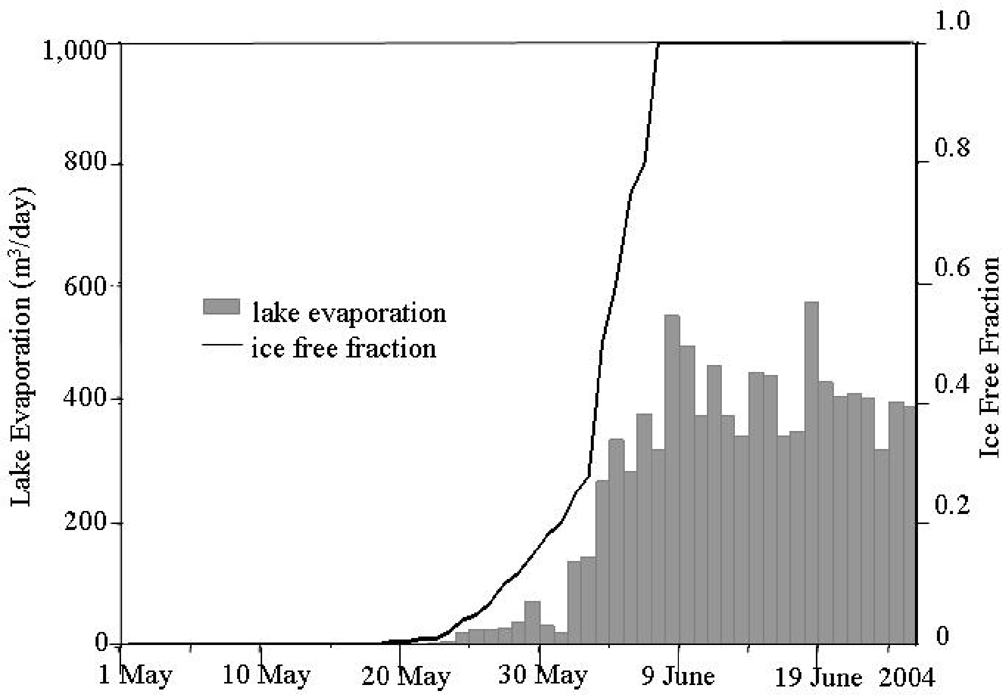

Figure 7: Daily evaporation from open water surface, and ice-free fraction on Shadow Lake.

Lake Storage and Outflow

Lake storage increased steadily as meltwater runoff reached the lake (figure 8). The early storage increase was gradual until May 22, when there were only small patches of residual snow in the catchment. Then, there was an abrupt increase in storage and a sharp hydrograph rise. The timing corresponded with the rapid delivery of flow at the slope runoff plot (Fig. 5), suggesting that it is the fast delivery of meltwater to the lake that occasioned the production of substantial outflow.

Like many small northern lakes, Shadow Lake outlet was blocked by snow at the end of winter (FitzGibbon and Dunne, 1981; Woo et al., 1981). This blockage prevented the occurrence of outflow until it was topped by the rising lake level. Initial outflow was gradual (about 100 m^3/day), yet it was able to deepen the channel and remove the snow and ice blockage, thereby lowering the threshold for lake discharge. Following the sharp hydrograph rise on May 23, outflow attained a peak that exceeded 2000 m^3/day. After this major peak, outflow declined but was interrupted briefly by a secondary rise due to rainfall. When lake storage level dropped below the lip of the outlet, outflow terminated. Afterwards, lake storage depletion continued and outflow was not revived for the rest of the year.

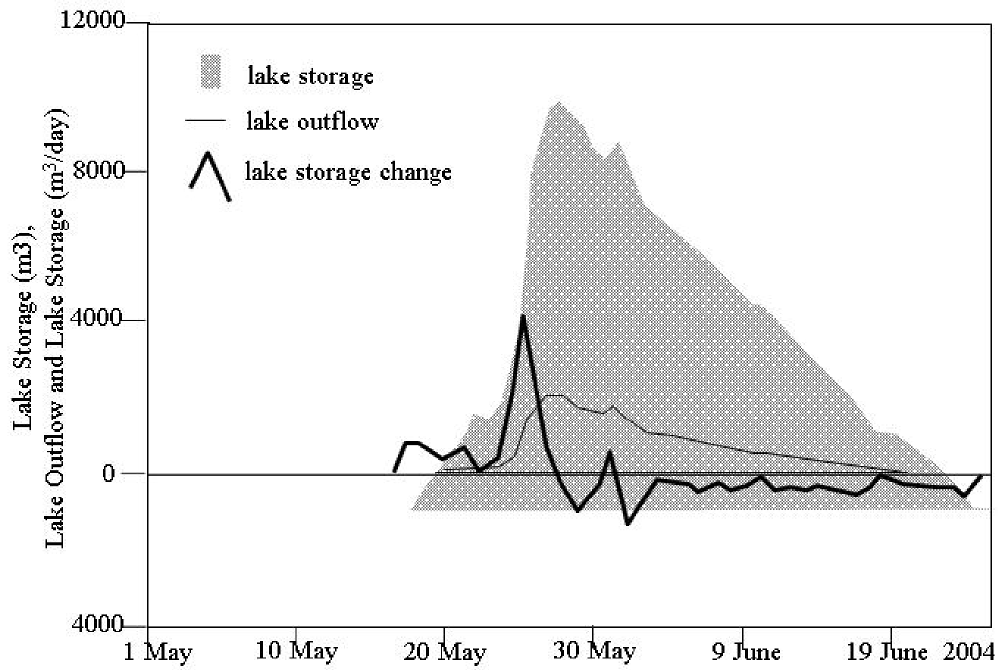

Figure 8: Daily change in lake storage, lake level and outflow from Shadow Lake.

Water Balance

Examination of the water balance of Shadow Lake basin enables an assessment of the relative magnitudes of its inputs, losses and storages:

$$M+R-E+Q_{in}-Q_{out}=\Delta S \tag{4}$$

where M is snowmelt, R is rainfall, E is evaporation, Q_{in} is inflow from upper lake catchment and Q_{out} is Shadow Lake outflow, and ΔS is change in lake storage. Substituting into eq. 4 the measured and calculated magnitudes (in thousands of m^3) for the study period, defined here as between the initiation of snowmelt and the cessation of lake outflow, yields 55 + 6.3 – 27 + 7.8 – 24 = 17.

The storage surplus of 17,000 m^3 in the basin at the end of the study period can be attributed to the recharge of groundwater and soil moisture in the uplands and bottomlands. Some of this water would eventually be released to the lake in the summer (based on field data not yet published).

For the study period, the runoff ratio of Shadow Lake basin, i.e. $Q_{out}/(M+R)$, was 0.4. This value is considerably larger than the ratio of 0.07 for the melt period in Skeeter Lake basin (0.043 km^2 lake area in a 0.575 km^2 catchment) located 100 km north of Yellowknife (Spence, 2000), and for Pocket Lake basin (lake area 0.048 km^2, catchment area 0.052 km^2, located 7 km from Shadow Lake) which generated outflow only once in a seven year period (Gibson et al. 1998). These limited comparisons indicate that outflow generation can vary greatly even within the same Shield environment.

After the study period, evaporation loss from the lake continued drawdown of its storage which did not recover to the level of its outlet threshold in spite of rainfall inputs of 23,000 m^3 in the summer. This highlights the hydrological significance of snowmelt in sustaining outflow from the Shield Lake.

DISCUSSION

This study indicates that in terms of runoff from the land portion of Shadow Lake catchment, there are zones of (1) fast flow delivery where the uplands are adjacent to the lake, (2) slow flow delivery from the bottomlands, mostly through surface flow in the soil, and (3) retarded flow delivery from areas that show no evidence of surface flow connection with the lake. The mechanisms and rates of snowmelt runoff delivery to the lake, as well as the lake ice cover duration and open water evaporation rates, are significant considerations with respect to outflow generation.

We propose a conceptual framework to relate runoff delivery to the frequency and magnitude of outflow occurrences. Within a catchment, zones not contributing directly to immediate runoff will find the meltwater and rainfall either infiltrating (and may re-emerge much later as groundwater flow) or evaporating from the ponded and soil-filled areas. Those catchments with large tracts of slow or retarded flow delivery will infill the lake very gradually, often at a rate slower than evaporation of the lake water. On the other hand, a catchment with a large portion of bedrock upland in direct contact with its lake will yield much runoff through fast delivery to the lake, thereby reducing the opportunity for evaporation loss in transit. Furthermore, runoff can reach the lake quicker than lake evaporation can bring down the water level, thereby permitting a net rise in the lake to reach and then exceed the outflow threshold.

A contrast of Shadow Lake with Pocket Lake and Skeeter Lake provides an illustration. In the case of Pocket Lake, there is a large ratio of lake area relative to the runoff contributing area. Runoff input from the uplands is often insufficient to match the cumulative lake evaporation so that lake storage level would fall much below the outflow threshold. The consequence is infrequent occurrence of outflow from this lake. In the case of Skeeter Lake, its low runoff ratio may be attributed to the large proportion of slow delivery areas to the total basin area (see Figure 1 in Spence, 2000), with very limited bedrock uplands adjacent to the lake. Although runoff is produced on the uplands during the melt season, it enters the soil-filled bottomlands which permit only slow delivery that does not match the rate of lake evaporation. Thus, lake level seldom rises above its outflow threshold.

From the above discussion, three factors may be considered important in terms of outflow generation from Shield lakes during the snowmelt season. (1) Antecedent lake storage condition: cumulative storage drawdown of the previous year is important as much snowmelt input may be needed to raise the storage to the outflow threshold. (2) The ratio of lake to catchment area: a small ratio of lake-to-basin area will likely have the benefit of bringing much runoff from the land surface into the lake to satisfy lake evaporation and storage demands, allowing outflow to be sustained. (3) Rate of runoff delivery: basins with a large area of fast runoff delivery will provide input quickly to the lake, before lake evaporation has the opportunity to lower the lake level.

The effect of snowmelt runoff can extend beyond the spring season as it serves a two-fold function in terms of Shield lake discharge. When the lake level is relatively high in the spring, an addition of melt runoff gives rise to high outflow from the lake, as evidenced in Shadow Lake during May 2004. If there had been considerable lake storage loss in the previous summer due to evaporation drawdown, meltwater input permits a replenishment of the storage deficit so that when the summer rain comes, outflow can be generated from the lake. The hydrographs of Yellowknife River at the outlet of Prosperous Lake (presented by Spence, 2000) offer an illustration. In 1990, this river produced peak flow in the spring in response to snowmelt but in 1993, spring melt did not give rise to peak flow possibly due to large storage deficit. Yet, peak flow occurred in early August, suggesting that lake storage deficit was sufficiently reduced by snowmelt input to permit subsequent high flow generation by rain events.

CONCLUSIONS

The snow on boreal Precambrian Shield is typically uneven, notably on the hilly uplands but less so on the lakes which invariably have a winter ice cover. As the snow melts, the distribution of snow-covered area changes continually, with the snow-free zone subject to evaporation loss. Runoff generation is spatially variable on the uplands. Meltwater can enter the lake directly, flow to the bottomland before released to the lake, or infiltrate through fissures in the bedrock. The bottomlands may have ice-rich seasonal frost near the ground surface to limit infiltration, yielding overland flow where conditions are favourable. As the ground thaws, the frost table descends, the water table drops and runoff to the lake becomes minimal. Runoff to the lakes lags behind the commencement of snowmelt by over a week. Then, there can be fast delivery from the upland directly to the lake, slow delivery via the bottomlands, notably through subsurface flow, and retarded delivery from certain parts of the catchment where there are no obvious surface flow paths to the lake and water there is subject to evaporation loss or infiltration. Both slow and retarded delivery may reach the lake eventually as groundwater discharge. The production and delivery of runoff are therefore spatially heterogeneous and they are subject to different time lags.

Lake storage is raised by catchment runoff, direct rainfall and snowmelt on the lake, and by inflow from upper lake. Evaporation loss from the open water portion of the ice-covered lake diminishes lake storage but a net storage increase will raise the lake level. When the lake level surpasses the height of the outlet lip (or threshold), outflow occurs. Lake storage is gradually reduced by water loss to evaporation and to outflow. When lake level falls below the outlet threshold, outflow ceases. For Shadow Lake, the entire flow duration for 2004 lasted 37 days.

Like other subarctic areas, snowmelt is the principal contributor to streamflow in the boreal Shield but headwater catchment occupied by lakes experience the extreme situation whereby outflow terminates after the melt season. For Shadow Lake catchment, the runoff ratio (outflow expressed as a fraction of snowmelt and rainfall in the spring) was 0.4. This value is higher than two other headwater basins in the boreal Shield. One possible reason is that Shadow Lake yields a larger quantity of water that was delivered quickly to the lake, at a time when lake ice cover was still present to inhibit evaporation, whereas slow delivery over a protracted period will be matched by greater opportunities for water loss to evaporation.

This study gives an account of the major processes responsible for flow generation and cessation in headwater areas of the Canadian Shield. Our findings should be applicable to other drainage systems in other lake-studded landscape under cold and semi-arid conditions.

ACKNOWLEDGEMENTS

This study was funded by the Natural Sciences and Engineering Research Council through the Mackenzie GEWEX (Global Energy and Water Cycle Experiment) Study, and by the Northern Student Training Program. Logistical and instrumentation support was provided by the Prairie and

Northern Region of Environment Canada, through the assistance of Jess Jasper and Chris Spence. We thank the following for their assistance in the field and for offering supplemental data: Bob Reid, Paul Saso, May Guan, Paula Lucidi, Derek Faria and Todd Pagett.

REFERENCES

Allan C, Roulet N. 1994. Runoff generation in zero order Precambrian shield catchments: the stormflow response of a heterogeneous landscape. *Hydrological Processes* **8**: 369–388.

Buttle JM, Sami K. 1992. Testing the groundwater ridging hypothesis of streamflow generation during snowmelt in a forested catchment. *Journal of Hydrology* **135**: 53–72.

Carey SK, Woo MK. 2000. The role of soil pipes as a slop runoff mechanism, Subarctic Yukon, Canada. *Journal of Hydrology* **253**: 110–129.

Environment Canada 1993. *Canadian Climate Normals 1961–1990*. Meteorological Service of Canada, Environment Canada, 58 pp.

FitzGibbon JE, Dunne T. 1981. Land surface and lake storage during snowmelt runoff in a subarctic drainage system. *Arctic and Alpine Research* **13**(3): 277–285.

Gibson J, Reid R, Spence C. 1998. A six year isotopic record of lake evaporation at a mine site in the Canadian subarctic: Results and validation. *Hydrological Processes* **11**: 1779–1792.

Heron R, Woo MK. 1978. Snowmelt computations for a High Arctic site. *Proceedings 35th Eastern Snow Conference,*Hanover, New Hampshire. 162–172.

Heron R, Woo MK. 1994. Decay of a high Arctic lake-ice cover: observations and modelling. *Journal of Glaciology* **40**(135): 283–292.

Landels AL, Gill D. 1972. Differences in volume of surface runoff during snowmelt period: Yellowknife, Northwest Territories. *Proceedings of Banff Symposium, September, 1972.IAHS* **107**: 927–941.

Luthin JN. 1966. *Drainage Engineering*. John Wiley and Sons Inc., New York, 250 pp.

Ohmura A. 1982. Objective criteria for rejecting data for Bowen ratio flux calculations. *Journal of Applied Meteorology* **21**: 595–598.

Peters D, Buttle J, Taylor C, LaZerte B. 1995. Runoff production in a forested shallow soil, Canadian shield basin. *Water Resources Research* **31**: 1291–1304.

Priestley CHS, Taylor RJ. 1972. On the assessment of surface heat flux and evaporation using large scale parameters. *Monthly Weather Review* **100**: 81–92.

Shilts WW, Ayslworth JM, Kaszycki, CA, Klassen RA. 1987. Canadian Shield. In Graf WL (Ed.) *Geomorphic Systems of North America*, Geological Society of America, Centennial Special Volume 2, 119–161.

Spence C. 2000. The effect of storage on runoff from a headwater subarctic Shield basin. *Arctic* **53**(3): 237–247.

Spence C, Rouse WR. 2002. The energy budget of Canadian Shield subarctic terrain and its impact on hillslope hydrological processes. *Journal of Hydrometeorology* **3**: 208–218.

Spence C, Woo MK. 2002. Hydrology of a subarctic Canadian shield: bedrock upland. *Journal of Hydrology* **262**: 151–166.

Spence C, Woo MK. 2003. Hydrology of a subarctic Canadian shield: soil-filled valleys. *Journal of Hydrology* **278**: 111–127.

Thorne G, Laporte J, Clarke D. 1999. Water budget of an upland outcrop recharge area in granitic rock terrane of southeastern Manitoba, Canada. *Proceedings of the twelfth International Northern Research Basins Symposium and Workshop*, Reykjavik, Iceland, August 23–37, pp. 317–330.

Woo, M.K. 2000. Permafrost and hydrology. In: Nuttall M and Callaghan TV (Eds.), *The Arctic:Environment, People, Policy*. Harwood Academic Publishers, Amsterdam, the Netherlands, 57–96.

Woo MK, Heron R, Steer P. 1981. Catchment hydrology of a high Arctic lake. *Cold Regions Science and Technology* **5**: 29–41.