The Effect of Storage on Runoff from a Headwater Subarctic Shield Basin

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ABSTRACT. Canadian Shield runoff production processes have been investigated, but research is needed beyond the hillslope scale to determine the influences on the volume and timing of runoff from headwater basins to higher-order streams. Such research will permit an understanding of the magnitude of climate change impacts at the watershed scale, where changes in the hydrologic regime are felt most by water resource users. To this end, water budget terms were measured over different portions of a headwater lake basin north of Yellowknife, Northwest Territories, during the spring snowmelt of 1998 to determine the relative importance of each component as water moved through the basin. Evaporation made up 16% and runoff 70% of the snowmelt from upland areas. Upland ponding of water during the melt increased the evaporative loss. Headwater lake storage deficits are very important in the timing and volume of runoff that is transmitted downstream: 53% of basin meltwater went to meeting the storage deficit in the lake so that only 7% of the meltwater was routed out of the basin. These results imply that topology of shield headwater basins, notably the location of lakes and antecedent water levels in lakes, are important in determining the runoff response. As storage deficits in surface waters will cause changes over space in runoff response and timing, it may be inappropriate to consider the snowpack or rainfall volume in isolation when forecasting runoff volume. These results are important for water management and hydrological modeling applications of northern shield basins.

Key words: Canadian Shield, Northwest Territories, runoff, snowmelt, water management

RÉSUMÉ. On a étudié les processus qui entraînent la production de ruissellement sur le Bouclier canadien, mais il faut envisager la recherche d'un point de vue plus global que l'échelle des versants, afin de déterminer les facteurs qui influencent le volume et la périodicité du ruissellement provenant des bassins d'amont vers des cours d'eau d'ordre plus élevé. Une telle recherche permettra de mieux saisir l'importance des retombées du changement climatique à l'échelle du bassin hydrographique, où les changements dans le régime hydrologique sont le plus ressentis par les usagers des ressources en eau. À cette fin, on a mesuré le bilan hydrique sur plusieurs sections d'un bassin lacustre d'amont au nord de Yellowknife, dans les Territoires du Nord-Ouest, durant la fonte des neiges au printemps de 1998, pour déterminer l'importance relative de chaque élément, à mesure que l'eau traversait le bassin. L'évaporation représentait jusqu'à 16 p. cent et le ruissellement 70 p. cent de la fusion nivale venant des hautes terres. À ces endroits, l'accumulation d'eau dans des dépressions durant la fonte augmentait la perte due à l'évaporation. Les déficits dans les réserves lacustres d'amont ont une influence majeure sur le rythme et le volume du ruissellement qui passe en aval: 53 p. cent de l'eau de fonte du bassin allait compenser le déficit des réserves du lac, de sorte que seulement 7 p. cent de l'eau de fonte se dirigeait hors du bassin. Ces résultats laissent entendre que la topologie des bassins d'amont du Bouclier, en particulier l'emplacement des lacs et leurs niveaux d'eau antérieurs, est importante pour déterminer la réponse du ruissellement. Vu que les déficits des réserves d'eau de surface amèneront des changements spatiaux dans la réponse et le rythme du ruissellement, il ne convient sans doute pas de considérer simplement le volume du manteau nival ou des précipitations quand on veut prévoir le volume de ruissellement. Ces résultats sont importants pour la gestion de l'eau et les applications de modélisation hydrologique des bassins du Bouclier septentrional.

Mots clés: Bouclier canadien, Territoires du Nord-Ouest, ruissellement, eau de fonte, gestion de l'eau

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INTRODUCTION

Climate models predict that atmospheric warming due to increased greenhouse gases will be strongest in polar regions. Such warming could have a large impact on a country such as Canada, which has a significant portion of its landmass in the subarctic and arctic regions. The taiga, or subarctic, Shield is the largest subarctic ecozone and the third largest ecozone in Canada. Quantitative monitoring of subarctic Shield climate suggests that year-to-year variation in the climate is significant. How the components of the water cycle respond to differences in climatic inputs has not been quantitatively studied in the subarctic Shield. In particular, understanding of how water travels from uplands to major streams is limited. To predict the impact of warming climates on water resources, we must address this information gap.

Numerous studies have investigated the runoff mechanisms of shield terrain. Hortonian runoff occurs from

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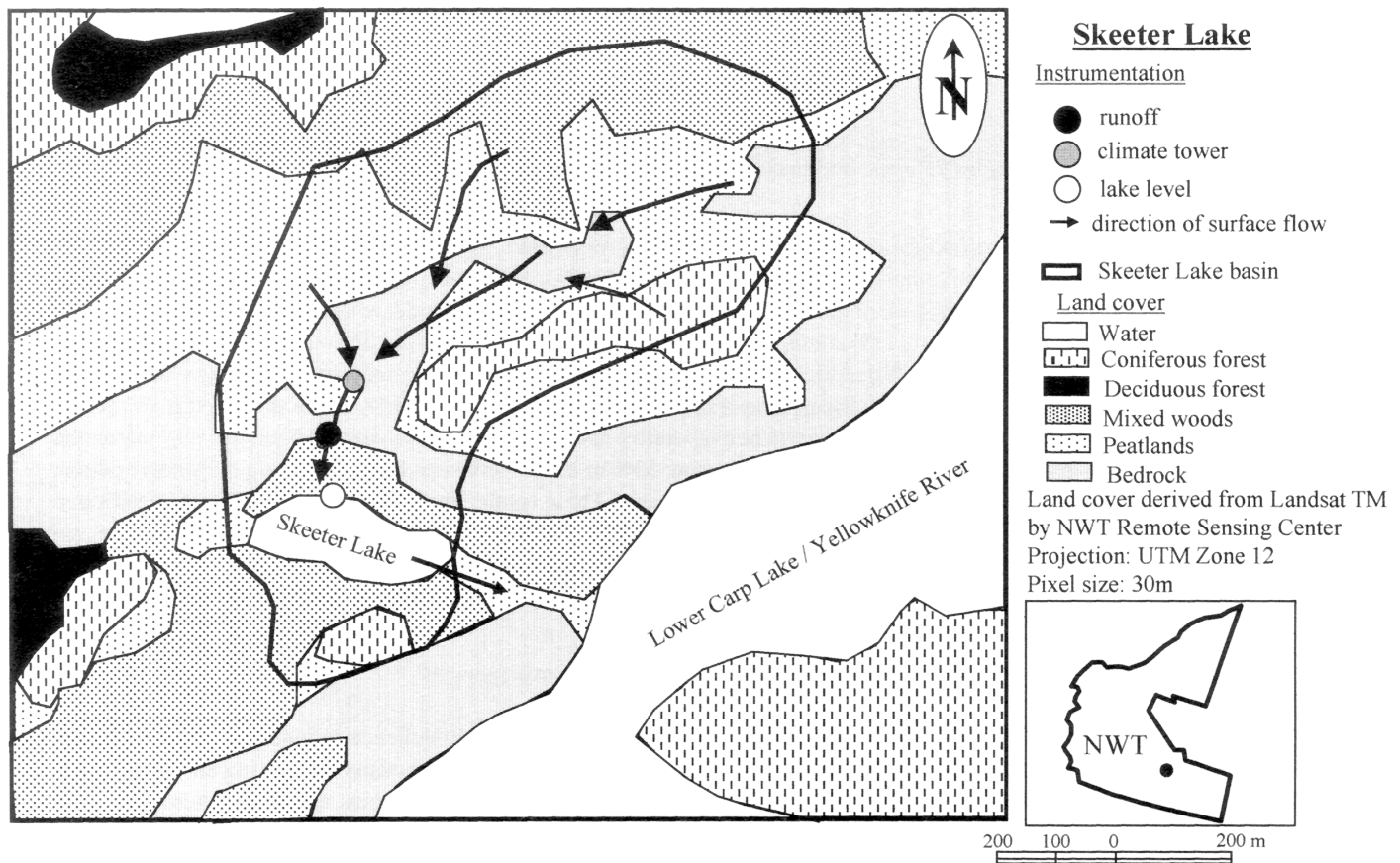
FIG. 1. Location of Skeeter Lake, Northwest Territories, Canada.

exposed bedrock ridges, allowing water to run off into either a stream or into soil-covered areas downslope, where it adds to soil moisture (Renzetti et al., 1992; Peters et al., 1995). Water is transferred from exposed bedrock surfaces to wetlands by moving downslope through soil, in a growing, saturated wedge to saturated wetlands (Bottomley et al., 1984; Wels et al., 1991; Renzetti et al., 1992; Peters et al., 1995). Tensiometer measurements on soil-covered slopes suggest that deep layers become wet first after significant rain. This phenomenon, caused by water moving laterally from the bedrock sources upslope (Renzetti et al., 1992), leaves the middle portions of the soil dry. Work by Peters et al. (1995) supports this idea: they measured almost 100% of slope runoff along the soil/bedrock interface. Flow through soil horizons is only a minute portion of the flow. The disproportionately large increases and delayed responses in piezometric levels downslope (versus upslope) noted by McDonnell and Taylor (1987) also support the wedge hypothesis. Hydrometric results from Branfireun et al. (1996) and isotopic results from Buttle and Peters (1997) imply that much of the event water flux is from upslope areas and has mixed with groundwater from the rising water table.

The above studies were performed at the hillslope scale. The heterogeneity of the Canadian Shield landscape makes the runoff processes that occur during a given storm a function not only of the meteorological and antecedent moisture conditions (Landals and Gill, 1972; Roulet, 1990; Thorne, 1992) but also of the distribution of land cover types in the basin (McDonnell and Taylor, 1987; Moore, 1989; Allan and Roulet, 1994). FitzGibbon and Dunne (1981) found that the effect a lake has on the basin outflow hydrograph is a function of its size relative to the drainage area. Soil-covered slopes downslope of exposed bedrock may not be saturated; this can delay, reduce, or even nullify runoff from some events (Allan and Roulet, 1994). Branfireun and Roulet (1998) found that uplands and low-lying peatlands are hydrologically decoupled during dry conditions. However, when conditions are wet, the two landscape units become coupled, and runoff components are an order of magnitude larger than under dry conditions, given similar rainfall inputs. They also note that changing source areas and runoff pathways result in different response times between the two conditions.

Landals and Gill (1972), who studied runoff from small plots in different land cover types near Yellowknife, Northwest Territories, conducted the only study examining runoff processes in the dry northwest portion of the Canadian Shield, where storage deficits are higher and precipitation lower than in the rest of the Shield. The present study attempts to determine the controls on the volume and timing of runoff from a headwater subarctic shield basin in this region by examining a spring snowmelt event. If the factors determining the amount of runoff transmitted out

TABLE 1. Areas and distribution of land cover types within Skeeter Lake basin.

Cover Type	Area (ha)	% of Basin
Skeeter Lake	4.3	11.6
Coniferous forest	8.5	22.7
Deciduous forest	2.4	6.4
Mixed woods	6.2	16.6
Peatland	6.7	17.9
Bedrock	9.3	24.8
Total	37.5	100

of headwater basins can be defined, this will provide a better understanding of how larger-scale shield basins react to different climatic inputs.

STUDY SITE

A site representative of the subarctic Shield, unofficially known as Skeeter Lake, was chosen as the study site (Fig. 1). Skeeter Lake basin is a 57.5 ha basin about 100 km north of the city of Yellowknife. It drains into Lower Carp Lake, a medium-sized lake on the Yellowknife River system, which drains south to Great Slave Lake. Skeeter Lake has a subarctic continental climate characterized by short, cool summers and long, cold winters. The annual mean temperature of the region is -7.5°C, and the mean annual precipitation is 280 mm, with just under half falling as snow (Wedel, 1990). Table 1 shows the areas and distribution of land cover. Skeeter Lake is a 4.3 ha oval-shaped lake that occupies the bottom of the basin. It is 2 m deep at the shore and has a maximum depth of 6.5 m. The remaining 88% of the basin, referred to here as the upland, is rimmed by exposed bedrock ridges and an open canopy forest of black spruce and birch. A peatland and more exposed bedrock occupy the middle of the upland. Average soil depths observed in the peatland during sensor and tower installation and soil sampling were in the order of 1 metre. Soils outside the peatland are very shallow (< 20 cm) or nonexistent.

METHODOLOGY

Surface runoff beyond exposed bedrock ridges in this part of the Canadian Shield generally occurs only once or twice a year—during the spring snowmelt and perhaps in the autumn, depending on precipitation conditions. The daily water budgets of two portions of the basin, the upland that drains into the lake and Skeeter Lake itself, were determined for the period of 15 April to 15 May during the spring snowmelt of 1998. The water budget is defined as:

$$\Delta S = P - E + I + M - O \qquad (1)$$

where ΔS is change in storage, P is precipitation, E is

evaporation, I is inflow, M is snowmelt, and O is outflow. All terms are in mm/day. The inflow and outflow terms represent surface flows. While subsurface flow occurs in this environment, it has been neglected in the water budget because of the difficulty in measuring it in such a remote region.

Climatic data were collected half hourly and averaged over daily time steps at a remote climate station in the vicinity of Skeeter Lake. Net radiation was measured with a REBS Q7 net radiometer positioned 2.5 m above the ground. A Vaisala HMP35CF was used to measure air temperature and relative humidity at a height of 1.7 m. Wind speed at 2 m height was measured using a Met One cup anemometer. Terrestrial evaporation was calculated daily using the Penman Monteith method, as described in Shuttleworth (1993):

$$E = \frac{1}{\lambda}\left(\frac{\Delta A + \rho \cdot c_p \cdot D / r_a}{\Delta + \gamma(1 + r_s / r_a)}\right) \qquad (2)$$

where Δ is the slope of the saturation vapour pressure curve (kPa/°C), A is available energy ($A = Q^*-Q_h-Q_g-Q_e$), ρ is air density (kg/m³), c_p is the specific heat of air (1.013 kJ/kg·°C), D is the vapour pressure deficit (kPa), γ is the psychrometric constant (kPa/°C), λ is the latent heat of vaporization (J/kg), and r_a is the aerodynamic resistance (s/m) calculated as

$$r_a = \frac{\ln\left[(z_u - d)/z_{om}\right] \cdot \ln\left[(z_e - d)/z_{ov}\right]}{0.1681 \cdot u_z} \qquad (3)$$

where z_u and z_e are the measurement heights (m) of wind speed and humidity and u_z is the wind speed (m/s). It is assumed that $z_{om} = 0.123$ h, $z_{ov} = 0.0123$ h and $d = .67$ h, where h is the mean height of the vegetation measured at 2.8 m. Surface resistance, r_s, (s/m) was calculated as:

$$r_s = \frac{200}{LAI} \qquad (4)$$

where LAI is the leaf area index, given a value of 1.5 on the basis of measurements in similar vegetation by Kite and Spence (1995). This method is expected to underestimate evaporation in subarctic Shield terrain by 20% (Lafleur, 1992).

Open water storage was measured by monitoring Skeeter Lake water levels. The accuracy of the pressure transducer as described by the manufacturer is ±5%. To determine upland storage, soil moisture measurements were made near the climate tower using a Campbell Scientific 615 TDR probe in soil with a depth of 20 cm. Because the soil is so shallow, once the entire soil column had thawed, the measurement of soil moisture by the TDR probe was assumed to be representative of the whole 20 cm column. Change in soil storage was calculated as the change in soil moisture (%) times the entire depth of the column (in mm).

Before the soil column had thawed, the active layer depth was calculated from Woo and Steer (1983):

$$z(t) = \beta\sqrt{t} \qquad (5)$$

where z is active layer depth (cm), t is time since the beginning of ground thaw (days), and β is a constant defined as 0.19 on the basis of subsequent measurements of ground thaw rates in the basin.

Observations of soil saturation during soil sampling and tower maintenance in the fall of 1997 suggest that the entire soil column in the peatland was unsaturated, and the water table was located below the bedrock surface prior to freeze-up. It was not feasible at the study site to drill water table wells in the bedrock.

Lake evaporation was estimated at a constant rate of 4.5 mm/day once ice had come off the shore. This estimate was based on average changes in lake level measurements immediately after the spring snowmelt was complete, on days when there was neither precipitation nor surface inflow or outflow from the lake. This evaporation value is similar to that for the same time of year from Pocket Lake, a similarly proportioned lake (4 ha, 6 m maximum depth) near Yellowknife, Northwest Territories (Reid, 1997). As the lake evaporation estimate is based on the change in lake level, its accuracy should also be ±5%.

Stratified snow surveys were based on land cover type. One ten-point snow survey was taken within each land cover type within the basin. No separate surveys of different aspects or slopes were taken. Since certain land cover types tended to occur in specific topographic zones, it was felt that by sampling land cover, one could also consider the effect of topography. For example, deciduous tree cover occurs only on south-facing, high-angle slopes, and low, sloping ridges often contain only exposed bedrock.

Melt was calculated daily using an energy balance method as in Roulet and Woo (1986):

$$M = \frac{Q_m}{\rho \cdot \lambda_f} = \frac{\left(Q^* + Q_h + Q_e + Q_g\right)}{\rho \cdot \lambda_f} \qquad (6)$$

where M is melt (mm/day), ρ is the density of water (kg/m^3), λ_f is the latent heat of fusion (J/kg), Q_m is the total energy available for snowmelt, Q^* is net radiation measured over the snowpack, Q_h is sensible heat flux, Q_e, latent heat flux, and Q_g, the ground heat flux. All the energy terms are daily averages and in W/m^2. Q^* was measured using the net radiometer located at the climate tower. Q_g was estimated by a calorimetric method (Rouse, 1984), using ground temperature data from a thermopile array between the surface and 10 cm depth and the soil moisture sensor described above at the climate tower. Q_h was estimated using an aerodynamic method as described by Price and Dunne (1976). Q_e was obtained from the daily evaporation estimate calculated using the Penman Monteith method as:

$$Q_e = \rho\lambda E \qquad (7)$$

where λ is latent heat of vaporization (J/kg) and E is the evaporation rate (m/s).

This method provides a single melt rate for every land cover type in the upland. Different melt rates occur in different land cover types because of slight differences in the energy available for melt (Price and Dunne, 1976); however, instrumentation was not available to measure melt rates in different cover types. The daily melt rate was applied to each land cover until the average depth measured for that land cover was melted. The upland melt was determined by weighting each land cover's melt rate by the proportional area of each land cover in the upland. This did not make any difference in the rate of upland melt rate until the last few days of melt, when the shallower snowpacks were gone. The standard deviations of the snowpack measurements were used to estimate their accuracy. The accuracy of the lake snowpack measurement is ± 10%. The accuracy of the terrestrial snowpack measurement is ± 17%.

Upland runoff and inflow into Skeeter Lake was calculated as the residual of the Skeeter Lake water balance, as long as no lake outflow occurred. Outflow from Skeeter Lake towards Lower Carp was not measured as no suitable cross section could be found. When outflow from the lake occurred, the remainder of the inflow record was estimated by a recession curve, and the outflow was determined as the residual of the lake water balance. To determine the recession curve, discharge in the stream connecting the upland and Skeeter Lake was measured twice by the float method, as the stream was too small to measure with a standard current meter. As these terms were estimated as residuals, their accuracy is determined by the additive error of all the other water budget terms. Lake inflow-and-runoff error is ± 20%. Lake outflow error is ± 25%.

RESULTS

Energy Balance

The energy balance during the period of snowmelt is illustrated in Figure 2. Net radiation and sensible heat dominated the energy budget throughout the melt period. It is suspected that as the declining snowpack reveals more vegetation, the albedo drops and the land surface absorbs more energy, which is then released as sensible heat.

Upland Water Balance

Water budget totals for the upland during the study period are listed in Table 2 and shown in Figure 3. There was no precipitation during the period of record. The dates when the streamflow was measured during the recession are identified in Figure 3. The snowmelt can be divided into two distinct periods. The first, 15–25 April, experienced 60 mm of melt, averaging 6 mm/day. The high melt
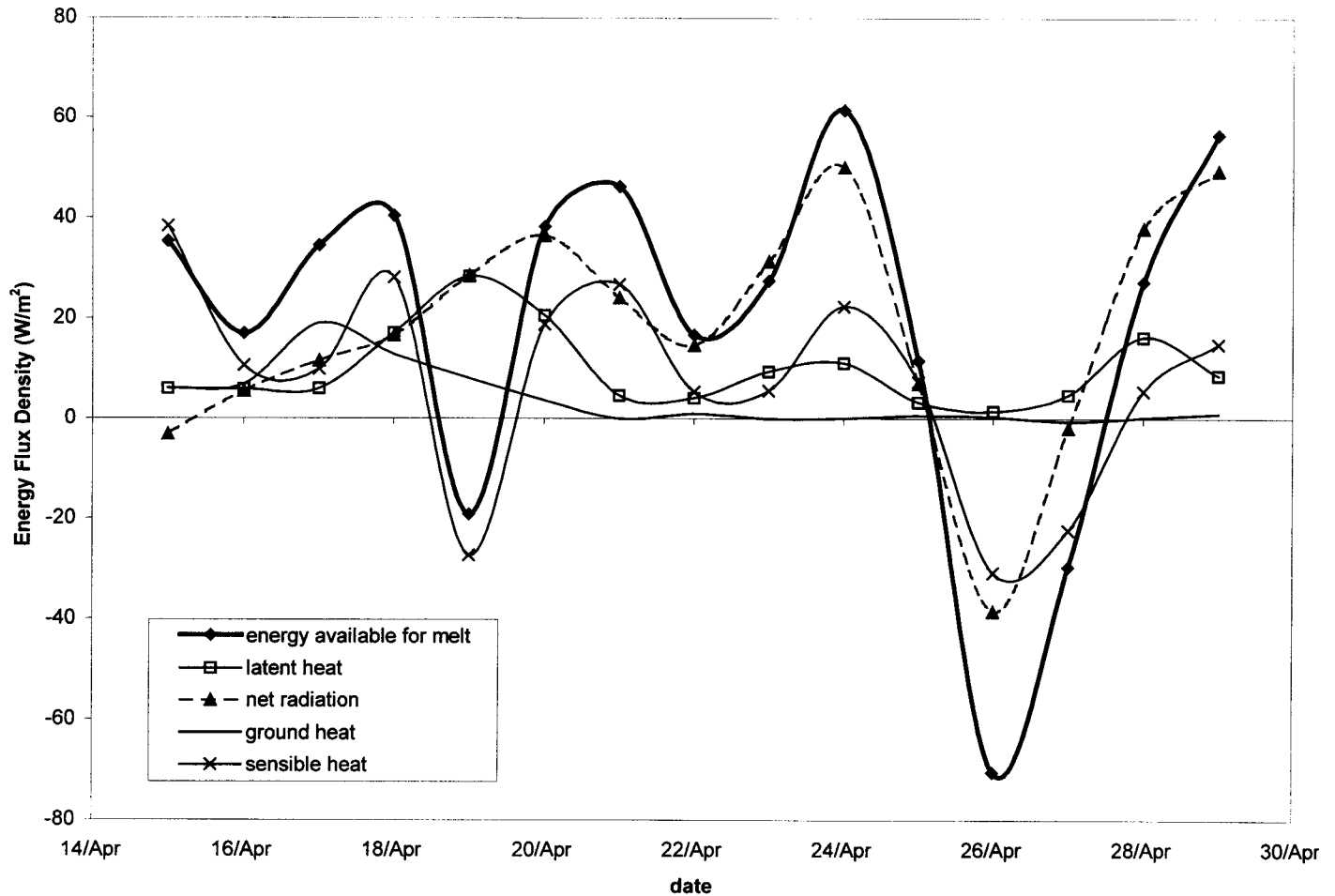
FIG. 2. Daily snowmelt-period energy budget, Skeeter Lake basin, 1998.

TABLE 2. Water budget components. Units are expressed in millimetres (mm) over the entire Skeeter Lake basin to the left of the slash and in millimetres over the local portion to the right of the slash.

	Precipitation	Evaporation	Melt	Upslope Inflow	Runoff and Outflow	Change in Storage
Upland	0	15/17	90/102	N/A	64/71	12/14
Lake	0	7/63	7/67	70/609	7/63	52/452

rate can be attributed to the warm temperatures. Before 15 April, the air temperature had been consistently below freezing (Fig. 4). On 10 April, the temperature began to rise; on 15 April it rose above freezing, and it did not drop below zero until 22 April. Such a long period above zero allowed the snowpack to melt exceedingly quickly. The second period of melt followed a brief cold spell and lasted four days, ending on April 30 when the remainder of the snowpack had melted.

Losses to the atmosphere (through sublimation and evaporation) were low, averaging 0.6 mm/day with a maximum of 1.2 mm/day. There are two distinct periods. During the first period, from 15 April to 29 April, losses to the atmosphere (sublimation loss from the snowpack) fluctuated around 0.45 mm/day. This rate is similar to the 0.46 mm/day sublimation rate found by Carey and Woo (1998) on a subarctic slope near Whitehorse, Yukon.

During the second period, after 29 April, losses rose steadily, averaging 0.7 mm/day.

Runoff began on 21 April, two days after the beginning of melt, and rose slowly to a peak on 26 April, in response to the first period of snowmelt. A decrease in runoff on 27 and 28 April reflected a decline in the snowmelt rate between 25 and 27 April. Runoff began to rise again on 29 April, peaking on 30 April, as the remainder of meltwater began contributing to runoff. After 30 April, runoff receded slowly to zero on 7 May.

The residual of the water balance can be interpreted as the change in surface storage in the upland, calculated using equation (1). Figure 3 shows how ponding in the upland peaked with the first melt on 24 April. The two-day delay between this peak and the first runoff peak reflects the time required for meltwater to move through the snowpack and out of the upland. It is expected that depressions
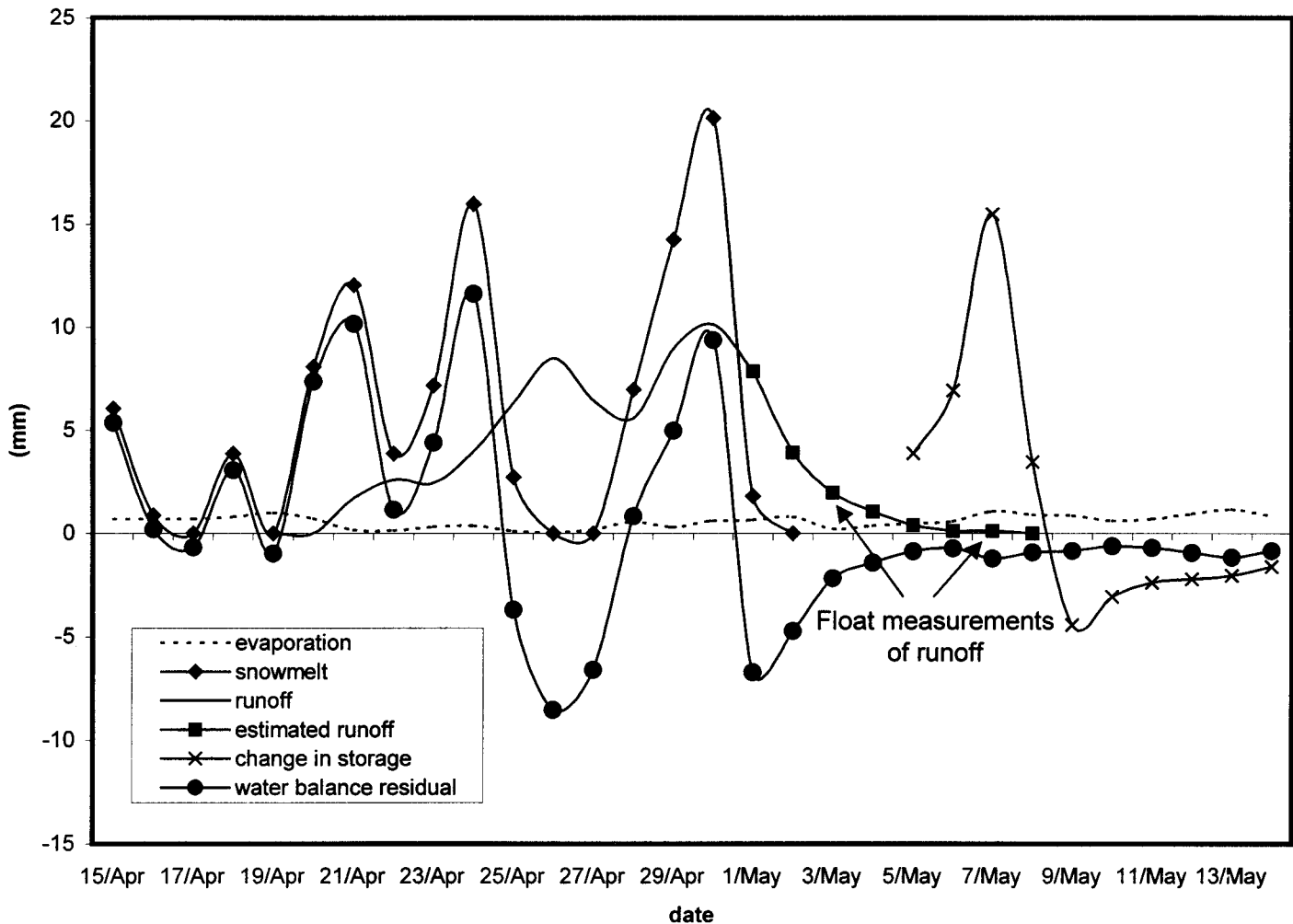
FIG. 3. Daily upland water budget, Skeeter Lake basin, spring 1998.

in the upland delayed the runoff of water. A second increase in storage coincided with the second melt period. There was no delay in runoff at this second snowmelt peak, presumably because ponding already present in the upland left less available surface storage.

The entire soil column where soil moisture was measured did not thaw until 6 May (Fig. 4). Prior to this the soil responded to changes in air temperature and melting of ground frost did occur, as shown by increases in soil moisture (Fig. 4). It was not until 5 May that the soil moisture increased to the freeze-up level, and the change in storage can be attributed to meltwater infiltration. The measured increase in storage was short-lived, as on 9 May the change in storage became negative (Fig. 3). The rate of change in storage after 9 May is much less than the evaporation rate. The fact that storage changes more slowly than evaporation estimates would suggest supports Lafleur's (1992) statement that the Penman Monteith method underestimates evapotranspiration from subarctic forest, although some of the soil moisture loss may be downward. The change in storage pattern reflects the downward soil moisture loss. It is believed that the change in storage pattern shown in Figure 3 is an error, an artifact

of the sensor placement angle. As the sensor was placed horizontally, it detected an infiltration wave that presumably proceeded deeper into the soil following the drop in the frost table. The peak measurement of water that infiltrated should be representative of what entered the soil. This theory can be confirmed by comparing the peak soil storage as detected by the TDR probe with the residual of the water balance immediately prior to the period during which soil moisture increased. The peak of the infiltration wave is 15 mm, equal to the cumulative sum of the water balance residual on 4 May, which represents the amount of meltwater that had not run off or evaporated to 4 May and which was available for infiltration.

Lake Water Balance

The most important term of the lake water budget is the change in storage (Fig. 5 and Table 2). The lake rose close to 0.5 m over the course of the month. The change in storage coincided with the snowmelt rate until the disappearance of the snowpack immediately adjacent to the lake on 22 April. The second and third storage peaks, on 26 April and 31 April, coincide with inflow inputs. Once
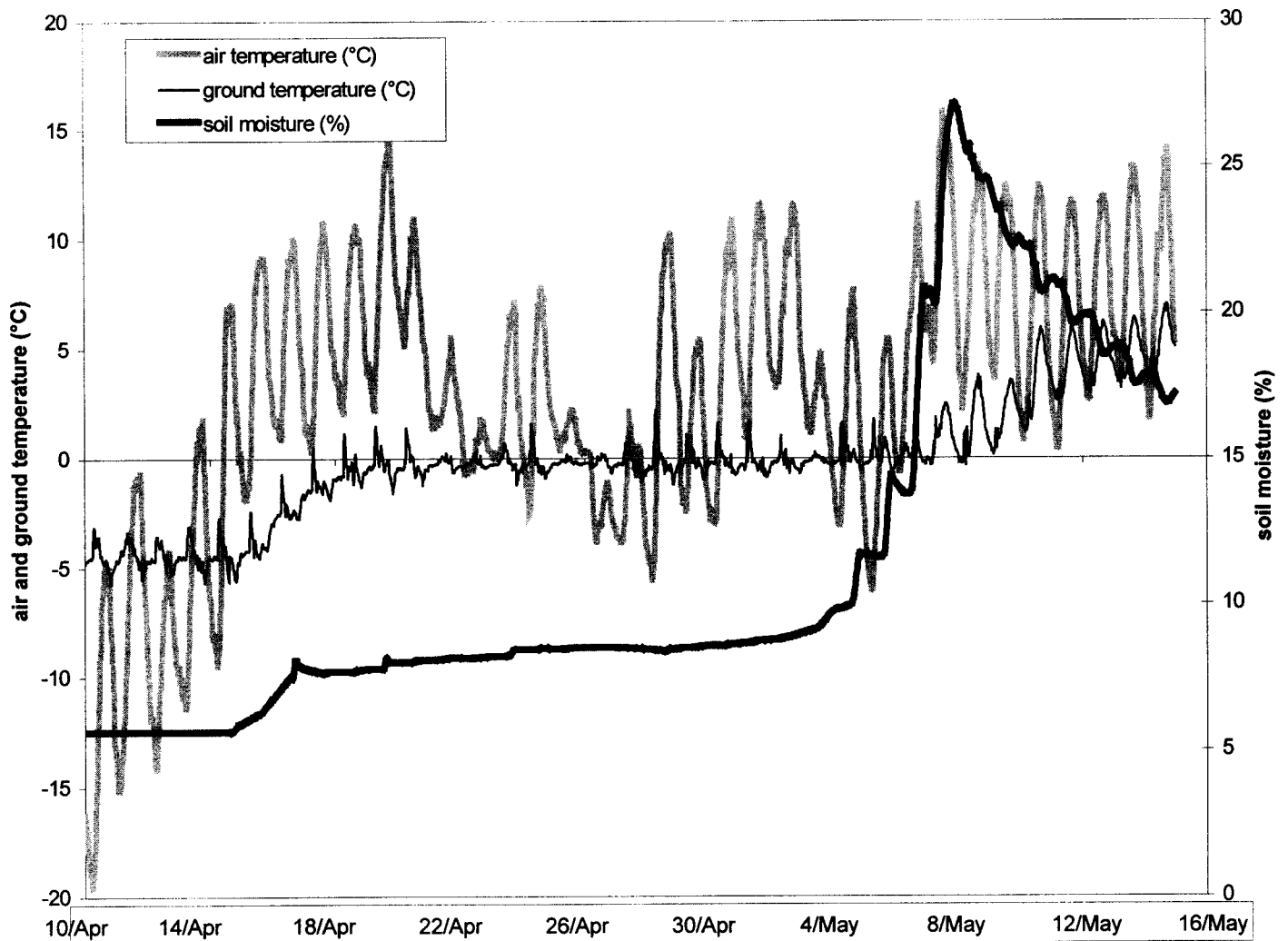
FIG. 4. Air temperature, ground temperature, and soil moisture in the Skeeter Lake basin during the period of the study, spring 1998.

outflow began on 1 May, the change in storage declined quickly to a negative term on 3 May. This quick decline is due to a decreasing rate of inflow, outflow from the lake, and the beginning of evaporation as the lake ice broke off from the shore.

The lake level had to rise almost 0.5 m before the lake surface reached the outlet elevation of 1.9 m (local datum) and outflow could occur. Outflow lasted for only one week while the lake level was above the elevation of the outlet elevation. The attenuation of runoff by the lake storage is significant. There were six days between the centroids of the inflow (28 April) and outflow (3 May). Peak flow was decreased by 75%. Only 63 mm, or 10%, of the 609 mm of inflow was released from the basin.

DISCUSSION

There was 97 ± 17 mm of snowmelt available over Skeeter Lake Basin. Evaporation over both the lake and upland accounted for 22 ± 2.4 mm, or 19–24% of the snowmelt. Just over half (54%) of the loss to the atmos-

phere from the upland occurred after the snowpack was gone and ponded water lay in the upland. The majority of snowmelt, 52 ± 9 mm or 45–64%, went to lake storage. This only left 5–9% of the snowmelt (7 ± 2 mm) available for transport downstream. The residual of the upland water balance is a good check of the error in the water budget. The change in storage is 14 mm as calculated by equation (1) and 14 mm according to the TDR sensor. The water budget values still allow some conclusions to be drawn despite the large magnitude of the error around some of the water budget terms.

The slow runoff recession relative to snowmelt implies that there was temporary storage in the upland. The terrain in the upland has a low slope and numerous depressions for water to pond. FitzGibbon and Dunne (1981) found that runoff from shield upland areas was delayed one to three days by temporary storage in the snowpack and topographic depressions, a result similar to the degree of attenuation noted here. Runoff attenuation by ponding in the upland also allows more water to be evaporated than if the water were to run off more quickly. The evaporation rate after 30 April was 30% higher than before 30 April.
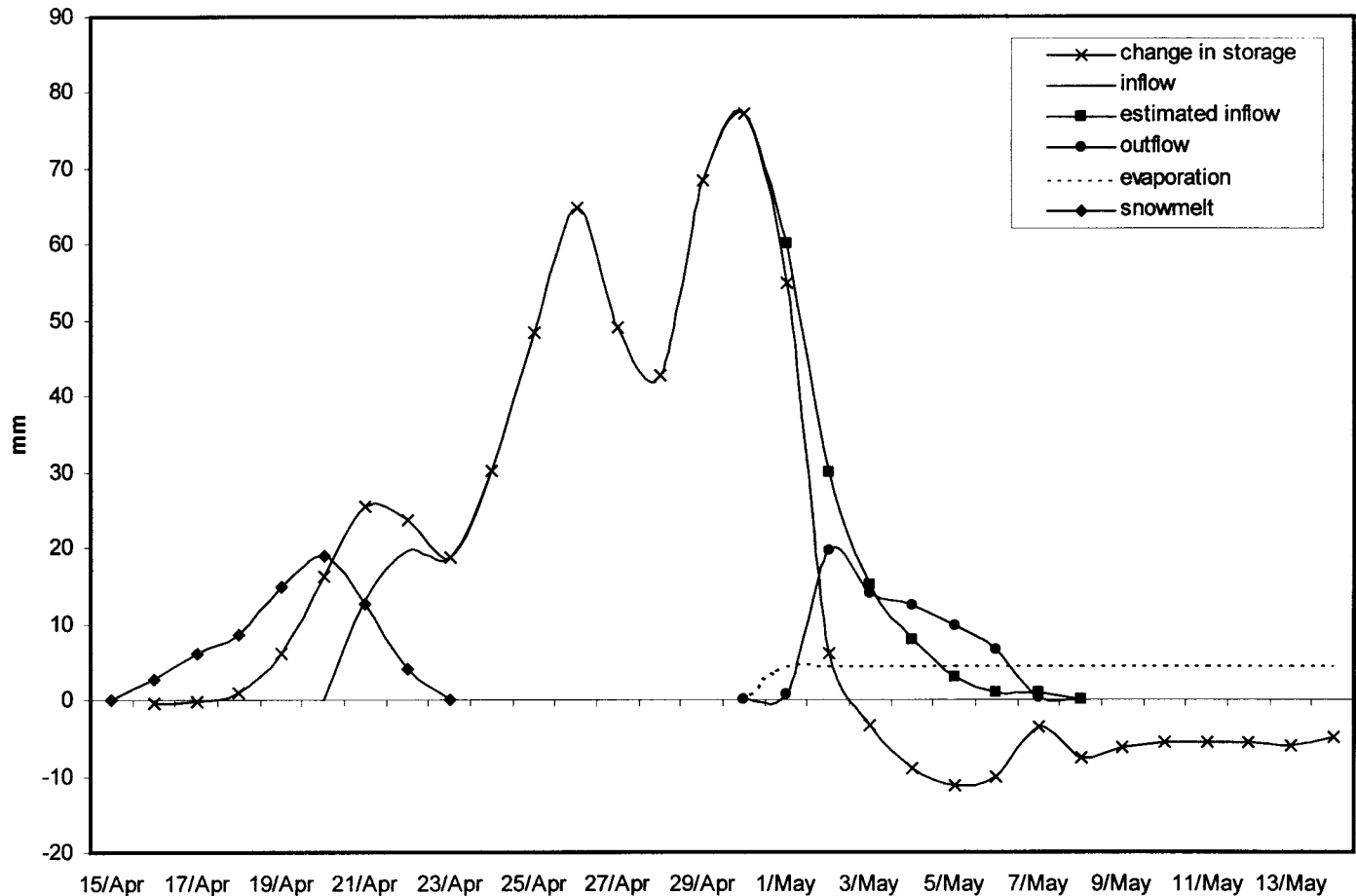
FIG. 5. Daily lake water budget, Skeeter Lake, spring 1998.

The increase in the loss to the atmosphere is attributed to the increasing availability of melt and soil water in these depressions after melting was complete on 30 April.

In the Canadian Shield, all runoff from upland or source areas (except runoff immediately adjacent to large rivers) must travel through headwater lakes before it reaches higher-order streams. Much of the snowmelt runoff from the upland was never routed all the way to the outlet of Skeeter Lake basin because of the high storage capacity of Skeeter Lake prior to snowmelt. The amount of spring snowmelt water that is transmitted out of the Skeeter Lake basin is as much a function of the lake storage deficit as of the volume of water in the snowpack.

Headwater lake storage deficits also affected the timing of runoff. The period when the entire basin was contributing to flow below the lake was only one week, or one-quarter of the period of study. Once the upland stopped contributing, the lake only contributed flow downslope for one day. Had the lake stage been higher before spring melt, the lake surface would have reached the outlet elevation sooner, and outflow supplied by upland water would have been maintained for a longer time.

If timing and volume of basin runoff are a function of surface storage deficits in headwater lakes, then the topology of land cover types has an important role in determining the

proportion of runoff that reaches higher-order streams. McDonnell and Taylor (1987), Moore (1989), and Allan and Roulet (1994) have suggested that wetlands have a similar effect. The location of a headwater lake within its basin may be critically important to determining basin response. Given the outlet elevation and observed water levels, it is expected that under normal climatic conditions, Skeeter Lake experiences outflow at least once a year. Conversely, Gibson et al. (1998) studied Pocket Lake, just north of Yellowknife, Northwest Territories, which has experienced outflow only once in the last five years. Pocket Lake is roughly the same size as Skeeter Lake, but its contributing upland is 6.5 times smaller. Given the similar precipitation inputs that both basins should receive, this contrast shows how contributing areas and the relative location of lakes within their watersheds can influence basin response. There may be a threshold lake size necessary for a lake to have an effect on the basin response. FitzGibbon and Dunne (1981) found that lakes are efficient in smoothing the hydrograph only when they comprise more than 5% of the basin. Skeeter Lake occupies 13% of its basin and appears to have the ability to attenuate the basin hydrograph.

Indian and Northern Affairs Canada conducts annual spring measurements of the snowpack in the Yellowknife River basin. The Water Survey of Canada operates a
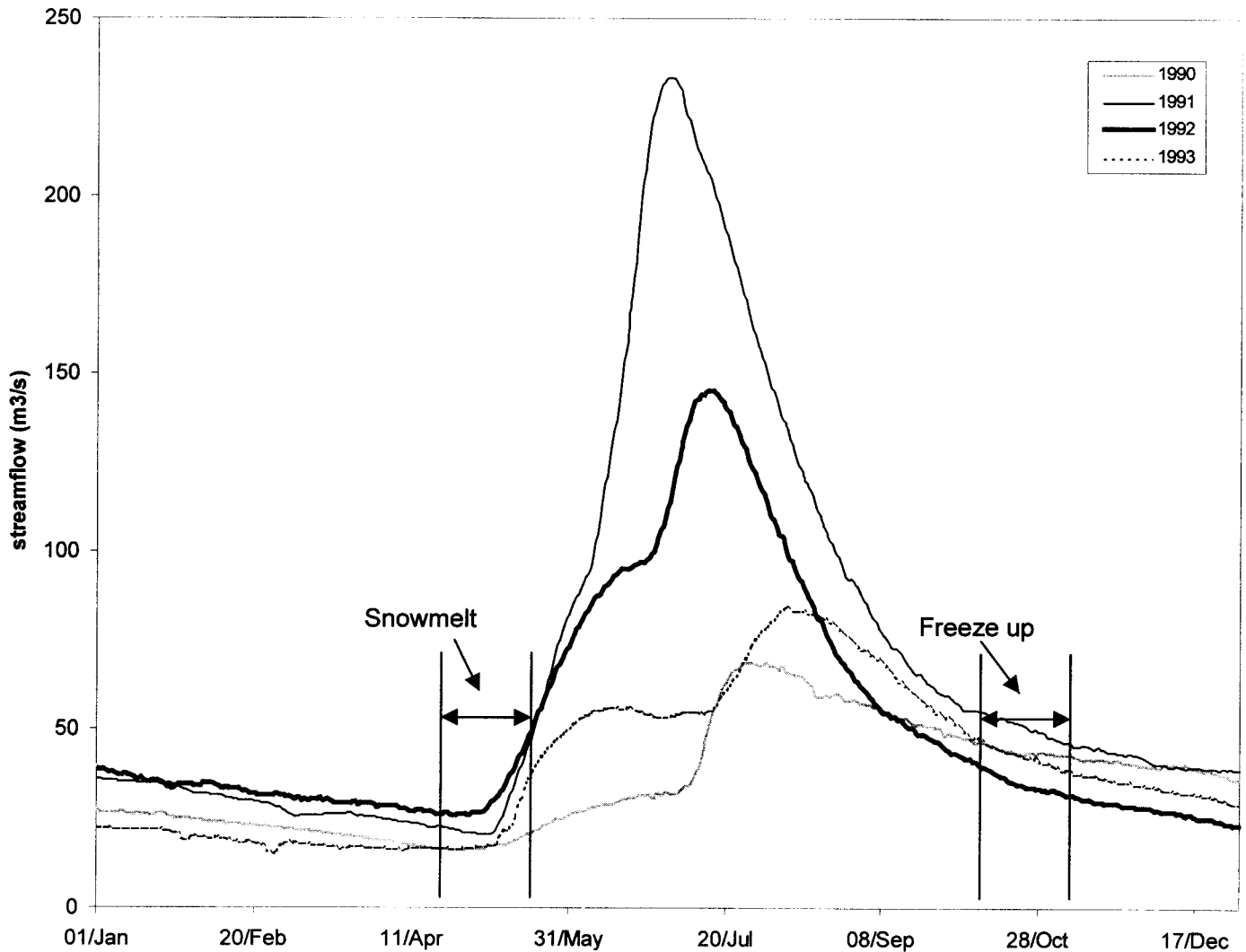
FIG. 6. Selected annual hydrographs of the Yellowknife River at the outlet of Prosperous Lake.

gauging station at the Yellowknife River below Prosperous Lake (drainage area is 16 500 km²). The nature of the Yellowknife River's annual hydrograph is shown in Figure 6. Snowmelt occurs from mid-April to mid-May. The river tends to peak in July and recede slowly until the next year's snowmelt. It is difficult, if not impossible, to determine an accurate volume of snowmelt runoff from this hydrograph, as the event covers several months during which the snowmelt signal is obscured by rainfall, lake evaporation, and freeze-up. Measurements of the first two variables in the basin are nonexistent and poor, respectively. As the annual hydrograph has the same general shape every year, the magnitude of the peak runoff (the difference between the streamflow just prior to snowmelt and the peak summer flow as measured at the Water Survey gauge) can be used as a measure of the total runoff volume. This value was compared with the snow water equivalent as determined by the Indian and Northern Affairs snow surveys.

Figure 7 shows how snow water equivalent and peak snowmelt runoff varied from 1990 to 1997 in the Yellowknife

River Basin. It is clear that the magnitude of the spring freshet in major subarctic shield streams is controlled by additional factors beyond the snow water equivalent. The results here indicate that the annual variation in headwater lake storage deficits must be considered when attempting to explain the runoff volume of northern shield rivers.

Scientists are improving attempts to model northern climate by using global and regional climate models that are linked to hydrological models with land surface schemes (Stewart et al., 1998). The results here indicate that any hydrologic model or land surface scheme applied over the subarctic Shield should account for the way in which temporal changes in storage affect the timing and volume of runoff. However, the author is not aware of any hydrologic model or land surface scheme linked to a general circulation model that does account for such processes. The inability of most hydrologic models to account adequately for changing storage and storage capacity explains in part the numerous problems encountered in applying those models to northern basins (Pietroniro et al., 1998; Soulis, 1998; Spence, 1998).
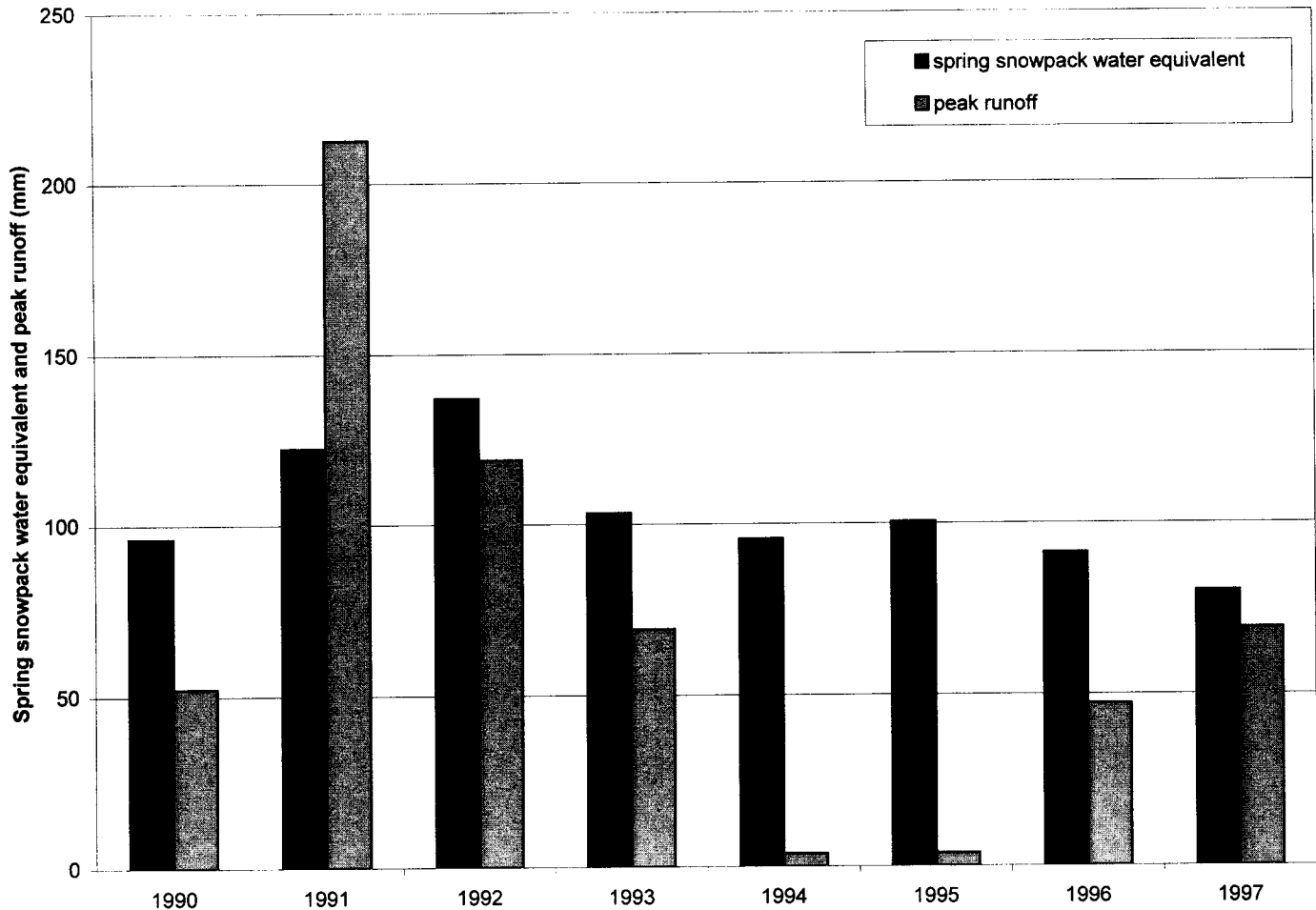
FIG. 7. Spring snowpack water equivalent and peak snowmelt runoff (both in mm) in the Yellowknife River basin, 1990–97.

CONCLUSION

This study shows that a major influence on the volume and timing of runoff from a headwater subarctic shield basin is the lake storage deficit. This deficit will influence the response of higher-order streams, like the Yellowknife River, to precipitation inputs. These results may be of interest to water managers of northern Shield basins, particularly in terms of forecasting runoff volume and response. As well, the results show the importance of correctly accounting for changing storage and storage capacity in hydrologic models that are applied to northern basins.

ACKNOWLEDGEMENTS

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