RUNOFF PROCESSES OF A SHIELD LAKE STREAM SYSTEM

RUNOFF PROCESSES OF A SUBARCTIC CANADIAN SHIELD LAKE STREAM SYSTEM

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ABSTRACT

Many lake-stream networks consisting of numerous lake basin elements linked by surface flow channels occupy Precambrian Shield and lowland areas in the boreal region. To investigate the processes causing flow generation and seasonal severance of flow connection in the lake-stream system, a chain of lakes in northern Canada was studied in 2004. Water balance shows that rapid and substantial runoff from the local basin slopes during the snowmelt period led to a rise of lake levels above their outlet elevations to generate outflow. Continued summer evaporation caused draw down of lake storage below the outflow thresholds, represented by the lake outlet elevations. Outflow ceased and the lakes became disconnected. Summer rainfall in a semi-arid environment was insufficient to overcome storage deficit to re-establish flow connectivity among all lakes. Individual lake outflow generation is dependent on the rate of runoff delivery, the initial antecedent storage level with respect to the critical outflow threshold level and the ratio of catchment to lake area. For the drainage system as a whole, streamflow interruption or continuity depends on linkage of its lake-stream sub-units. The principle of fill and spill governs runoff generation and flow connection between the lake elements. This principle is applied to model the flow along a chain of lakes, taking account of antecedent storage in individual lakes, their storage change calculated through water balance and the thresholds to be exceeded for outflows to occur.

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Mahsi cho to the Yellowknives Dene First Nation for granting me permission to perform my research in the study watershed by supporting my application for a N.W.T. research licence.

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CHAPTER 1: 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. These lakes form hydrologic chains and networks of series of lakes linked by streams, often with the lakes dominating the network. 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 runoff ratios (runoff / precipitation), from under 0.05 to over 0.8 (Landals and Gill, 1972, Spence and Woo, 2002).

Runoff generation and groundwater flow in the Canadian Shield has been studied [Allan and Roulet, 1994; Buttle and Sami, 1992; Branfireun and Roulet, 1998; Devito et al., 1996; Peters et al., 1995; Spence and Woo, 2002; 2003; 2006], though very little attention has been placed on the influence of small lakes on the flow in headwater catchments and tributary streams. 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 $2x10^{-5}$ 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. During dry subarctic summers, lakes have been observed to experience a draw down that sometimes cuts off their outflow, isolating the lakes into disjoint hydrological entities. Spence (2000), for example, observed flow cessation in a small headwater lake that produced flow for only 13 days in an entire year. For the lake-stream network, this can have major influences on flow connectivity. Attention has been paid to flow connectivity in relation to runoff generation in a variety of landscape, including northern wetlands (Bowling et al., 2003; Quinton et al., 2003) and bedrock upland with soil-filled valleys in temperate and boreal latitudes (Branfireun and Roulet, 1993; Buttle et al., 2004; Spence and Woo, 2003). Flow connectivity in lake basins in the subarctic and the Arctic (FitzGibbon and Dunne, 1981; Woo et al., 1981) have been examined however, the processes relating surface flow connections with lake storage remain inadequately understood. The loss of connectivity witnessed by Spence (2000) may be a large scale version of the fill and spill runoff generation mechanism.

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 have not been analysed in detail.

The objective of this study is two-fold. Firstly, 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 when water is plentiful, and later in the dry summer season. Secondly, to investigate the processes related to surface flow connections and the seasonal severance of flow connections. A conceptual framework relating runoff delivery to the frequency and magnitude of lake discharge occurrences will be presented. A framework applicable to the modelling of flow connectivity in a lake-stream system in a semi-arid Shield environment will be proposed.

The thesis is organized as follows. Chapter 2 describes the study basins physical features and climate of the study basins and an experimental study plot utilized to study slope runoff. Chapter 3 outlines the methods of field data. Chapter 4 investigates lake basin runoff processes utilizing a basin water balance approach. Factors controlling lake outflow generation are presented along with a conceptual framework relating runoff delivery to the frequency and magnitude of lake outflow occurrences. Chapter 6 investigates lake – stream network surface flow connectivity. The lake water balances are used to investigate the processes at work. A framework to model stream flow connectivity in lake – stream networks in the semi-arid subarctic Shield is presented. Results of this study will increase knowledge on the behaviour of the subarctic hydrologic system that can lead to improved algorithm development to model Shield hydrology in a semi-arid environment abundant with lakes.

Much of chapter 4 has been published and chapter 5 has been submitted for publication.

Mielko C and Woo MK 2006. Snowmelt runoff processes in a headwater lake and its catchment, subarctic Canadian Shield. *Hydrological Processes* 20, 987-1000.

Woo MK and Mielko C. Hydrologic connectivity of a lake-stream system in a semi-arid environment. *Hydrological Processes.* Submitted.

CHAPTER 2: STUDY AREA

2.1: Study Basin

A tributary catchment (62°33'N; 114°21'W) of Yellowknife River located 15 km north of Yellowknife, Northwest Territories, Canada was chosen as a study basin representative of the subarctic Canadian Shield in a semi-arid environment with numerous lakes. The main study catchment has an area of 1.61 km^2 and contains a chain of four lakes. The lowest lake of the system also receives additional water sources from a lake - stream network with more than ten additional lakes, increasing the total watershed size to 4.76 km^2 (Figure 2.1). To study the runoff and connectivity of a system of lakes, the study basin was subdivided into four sub-basins: Melvin, Shadow, Hazel and Lois (areas of 0.12, 0.56, 0.17 and 0.76 km^2 respectively), each containing a lake from which it is named (varying in size from 0.1 km^2 to 0.6 km^2). These lakes are linked by channels, with Melvin Lake draining into Shadow Lake which connects to Hazel Lake which flows to Lois Lake, forming a chain of lakes that controls streamflow. Lois Lake also receives streamflow input from another chain of lakes drained through Rater Lake (basin area 3.15 km^2). The study area was chosen because it was typical of the lake dominated headwater stream watersheds and was relatively accessible, via an abandoned mining road which ran along part of the western edge of the study basin.

2.2: Terrain Units

The basin can be divided into three terrain units (Figure 2.2), lake, upland and bottomlands. The upland is occupied by crystalline granite and gneissic bedrock that is fractured to varying degrees. The rock surface has been sculpted into undulating mounds and shallow depressions some of which are filled with a thin layer of soil (typical thickness values between 0.01 to 0.5 m). The bedrock upland is covered by lichens, mosses and other vegetation including creeping junipers (*Juniperus horizontalis*) that grow along the cracks, isolated stands of Jack Pine (*Pinus banksiana*) and occasional trembling aspen (*Populus tremuloides*) and white birch (B*etula papyrifera*). The bottomlands consist of wetlands and soil filled valleys. Soils usually have a surface organic layer between 0.15 to 0.45 m in the dry valleys and between 0.25 to greater than 1 m in the wetlands. Below are mineral soils and glacial tills which extend to the bedrock. Permafrost and late lying seasonal frost are prevalent. Drier parts of the bottomlands have an open canopy of white spruce (P*icea glauca*), white birch, black spruce (*Picea mariana*) and tamarack (*Larix laricina*), an understory of willow (S*alix sp.*) and birch (*Betula sp.*) shrubs and a ground cover of mosses, club lichens, common and alpine bear berry (*Arctostaphylos uva-ursi* and *A. rubra*) and small bog and bog cranberry (*Oxycoccus microcarpus* and *Vaccinium vitis-idaea*). Saturated bottomlands are wetlands with moss, grasses, sedges, Labrador tea (*Ledum groenlandicum*), sometimes with stands of tamarack and black spruce.

The main study catchment consists of 35 % uplands, 52 % bottomlands (46 % valley and 6 % wetland) and 13 % lakes. The area and percentage breakdown of terrain units for each sub-basin is provided in Table 2.1.

2.3: Climate

The climate is subarctic and semi-arid. Climate normals of the area observed at Yellowknife Airport based on the period of 1969-1989 [*Environment Canada*, 1993] are summarized in Figure 2.3. The area receives an average of about 280 mm of annual precipitation, 116 mm of which is snowfall. Mean January temperature falls to -27 °C and mean July temperature reaches 17 °C. Mean temperature is above freezing between mid April to mid October, though this can vary greatly from year to year.

2.4: Experimental Slope

An experimental slope of 5300 m^2 was used to study snowmelt runoff delivery from the upland (Figures 2.1 and 2.4). This slope extends from the bedrock upland to a bowl shaped depression at the base which captured runoff from the entire slope to form a temporary pond. The depression contains an aspen stand and has a poorly developed mineral soil underlying a thin (0.02 – 0.05 m) organic soil layer and a thin (up to 0.15 m) leaf detritus layer.

Table 2.1 : Area and percentage breakdown of terrain present in the four sub-basins: Melvin, Shadow, Hazel and Lois.

Sub-basin	Bottomland				Bedrock Upland		Lake	
	Soil-filled Valleys		Wetlands					
	km^2	%	km^2	%	km^2	%	km^2	%
Melvin	0.04	31.0	0.01	8.9	0.06	52.0	0.01	8.1
Shadow	0.18	31.7	0.03	6.0	0.27	47.0	0.09	15.3
Hazel	0.07	43.2	0.01	5.6	0.08	46.3	0.01	4.9
Lois	0.45	58.7	0.04	5.7	0.16	20.5	0.11	15.2
Total	0.73	45.6	0.09	6.0	0.56	34.8	0.22	13.6

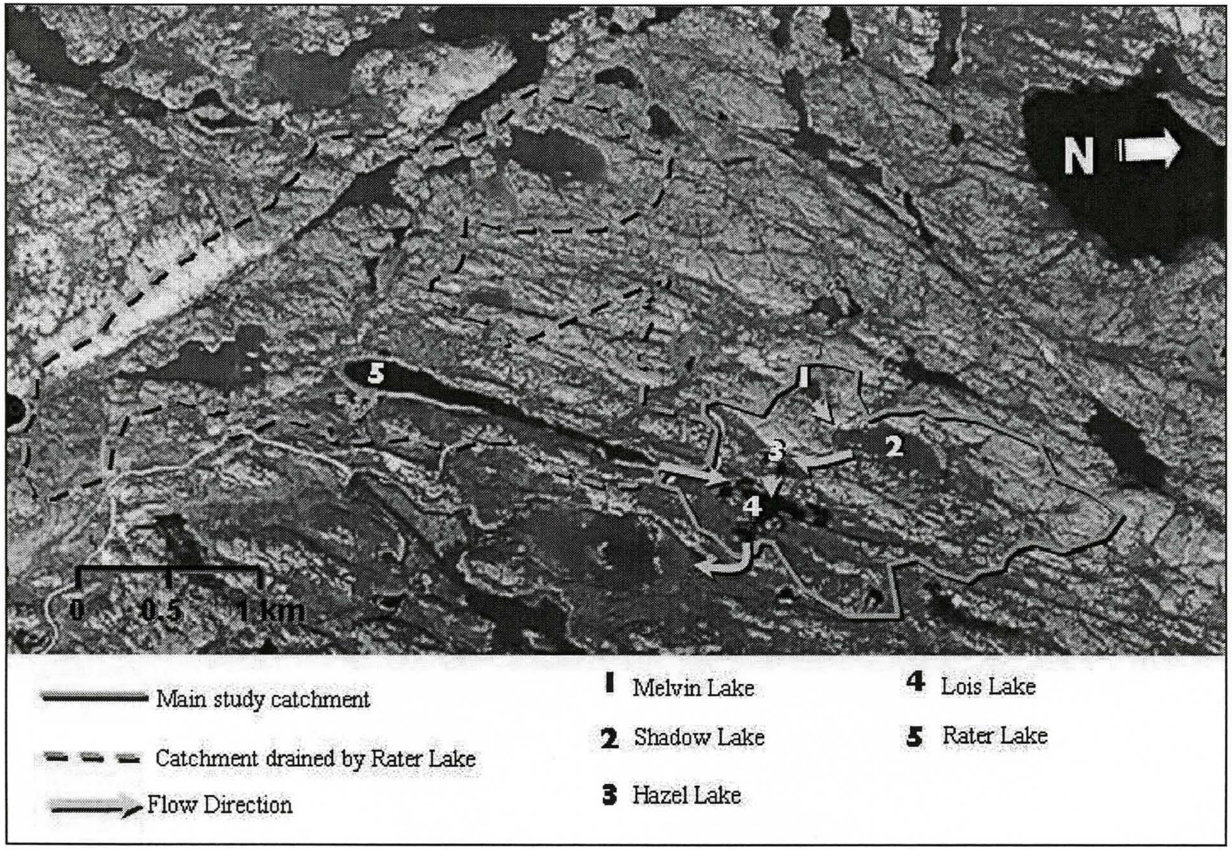

Figure 2.1: Aerial photo of study watershed and surrounding area. The main study watershed is outlined by a solid line and the watershed drained by Rater Lake is outlined by the dashed line. The channeled flow paths of the study lakes are indicated with arrows.

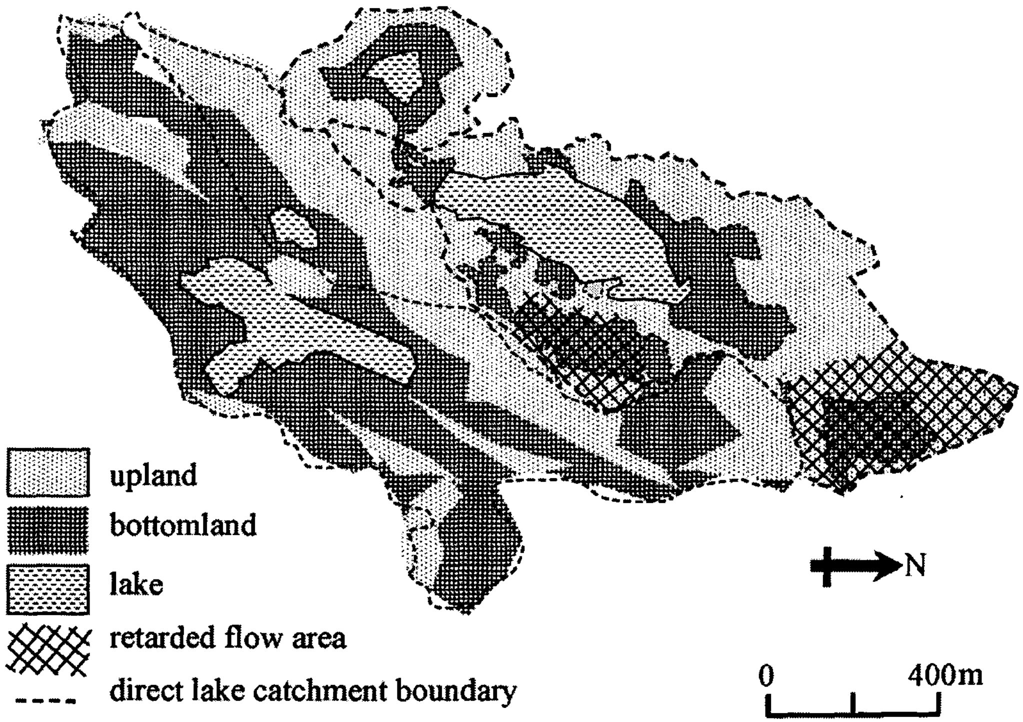

Figure 2.2: Study catchment with sub basins outlined by dashed lines, divided into topographic units of upland, bottomland and lake; and retarded flow areas which are zones without effective flow pathways to convey meltwater runoff to the lake during snowmelt season.

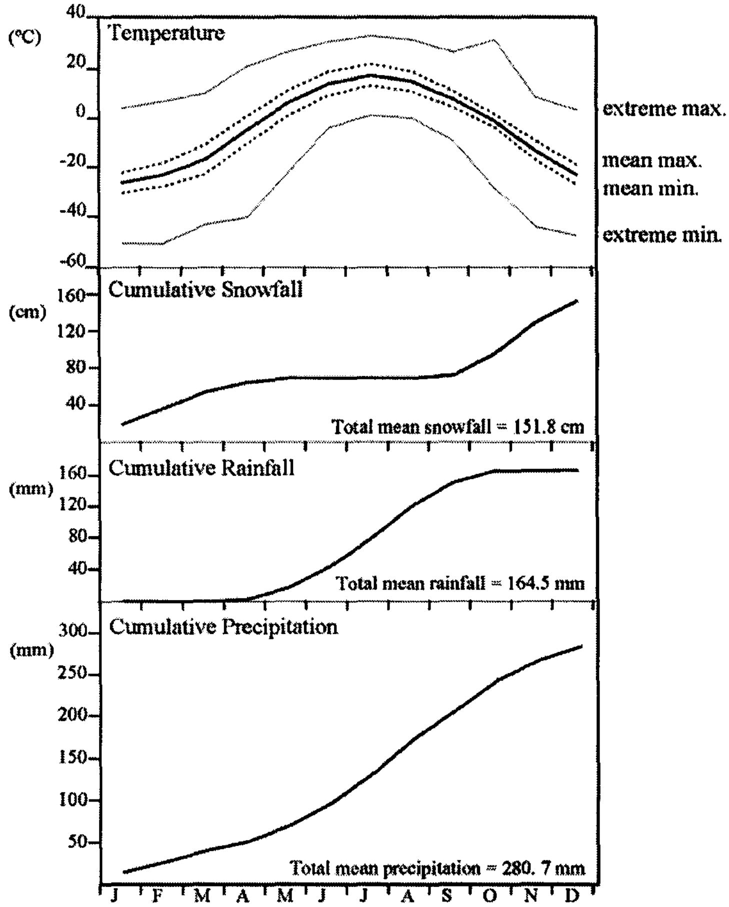

Figure 2.3: Yellowknife airport 1961-2000 climate normals (Environment Canada, 2003).

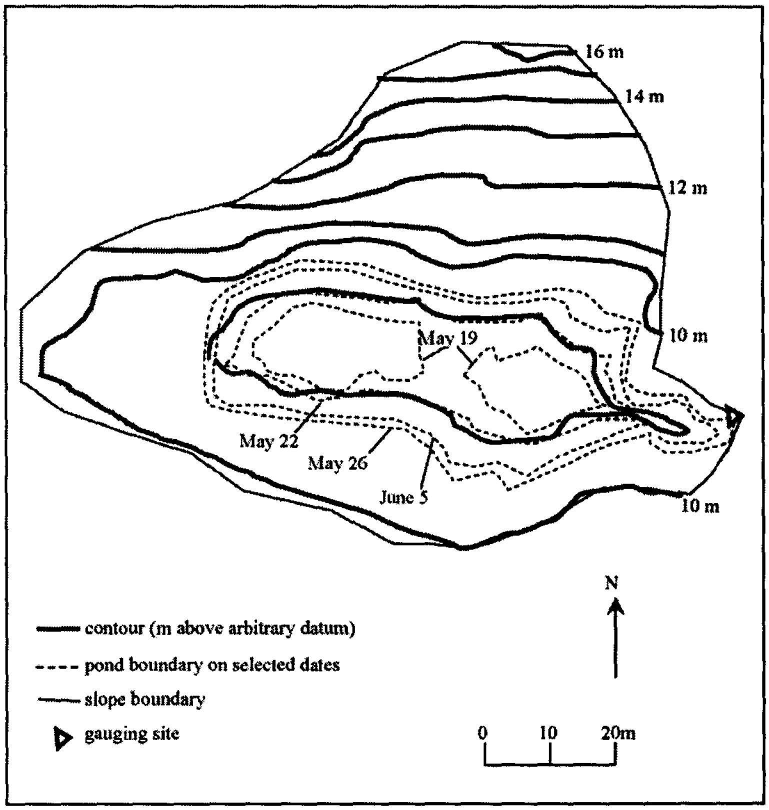

Figure 2.4: Topographic map of the experimental slope, showing direction of drainage into a depression that collects the slope runoff.

CHAPTER 3: METHODS

This study was carried out in the springs and summers of 2003 and 2004. The main study period was between May 4 and August 31 of 2004 with snow data collected in April 2004 and supplemental data gathered in 2003. Measurement sites and field instrumentation deployment is shown in Figure 3.1.

Unless otherwise stated, the magnitude of all hydrologic variables will be reported in cubic meters. Volumetric instead of depth measurements were used here to avoid the complication of having to make conversions to adjust the depth measures every time the discussion is switched between basin, lake and plot scales.

3.1: Precipitation Data

A snow survey was performed prior to the commencement of melt in early April 2004, following methods described in Woo (1998). A series of transects throughout the study catchment classified according to terrain types were used, employing 540 depth measurements and 50 density determinations (using Meteorological Service of Canada snow sampler) to obtain the average snow water equivalent (SWE) for each terrain type. Snow boards, with an area of approximately 0.3x0.3 m^2, were deployed after the snow survey to measure new snowfall. Following a snowfall event, new snow on the boards was collected and weighed to obtain the snow water equivalent. To produce snow distribution maps, 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.

Daily snow ablation was obtained by measuring the lowering rate of the snow surface along several 2-m lines at various terrain types and then by converting the depth change into snow water equivalent unit using surface density measurement (Heron and Woo 1978). The daily spatial pattern of melt was obtained by subtracting the daily ablation rates for various terrain types, weighted by the fraction of each terrain type in the grid, from the SWE values in each grid. This was done for each grid until it became bare, and the total daily amount subtracted represented the daily melt. 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.

Rainfall was measured using a tipping bucket rain-gauge and supplemented by 10 manual gauges spread across the catchment (Figure 3.1). Thiessen polygons were used to obtain the spatially weighted average for rain falling within the sub basins and directly onto the lakes.

3.2: Climate Towers

Two climate towers were installed in the basin, one in a bedrock upland area and the other in a soil filled bottomland area. The bedrock uplands tower (Figure 3.2) was located on the southern portion of Shadow Lake sub basin while the soil filled bottomlands tower (Figure 3.3) was located in a treed valley located in the north eastern portion of Lois Lake sub basin. The bottomlands tower was secured to the top of a small narrow bedrock outcrop which protruded 3.72 m above the bottomland floor in the center of the valley. This location enabled the tower to be securely fastened to the ground and provided additional height to extend the attached instrumentation above the forest canopy.

The towers were equipped to measure and record (Campbell Scientific CR 10 and CR 10X data loggers) at half-hour intervals, air temperature and relative humidity (HMP 35c and 35cf Temperature and Relative Humidity Probes, housed in multiplate radiation shields) at two heights, net radiation (NR Lite Net Radiometers) and windspeed (Met-one 013A Wind Speed sensor: 3-cup anemometer). The bedrock tower also recorded incoming solar radiation (Kipp & Zonen CM3 pyranometer) and rainfall intensity and magnitude (TE525 Texas Tipping Bucket Raingauge).

3.3: Data for Evaporation Calculation

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, air temperatures and relative humidity at two heights, and ground temperatures (TidbiT temperature sensors) at depths of 0.05, 0.1, 0.15 m in the bedrock upland and 0.05, 0.1, 0.15, 0.2, 0.5, 0.8, 1.2 m in the bottomlands. Lake and pond evaporation was calculated at half hourly intervals by the Priestley and Taylor (1972) method, with an alpha value of 1.26 (Eaton et al., 2001). Data employed for the calculation included net radiation and air temperature measured over the open water surface of a nearby lake (unpublished data from Bob Reid, 2004), together with lake temperatures (TidbiT temperature sensors) at surface, 0.1, 0.15, 0.3, 0.45 and 1m for each lake converted into heat storage using the method outlined by Oswald and Rouse (2004). Shadow Lake acquired a seasonal ice cover that reached a maximum thickness of 0.6m at end of the 2003 - 2004 winter. Photographs were taken of the decaying lake ice cover during the spring to estimate the changing ice free fraction.

3.4: Lake Level and Outflow

Lake level for five lakes (Melvin, Shadow, Hazel, Lois and Rater) was recorded at half hourly intervals using an automatic water level sensor (Ecotone (r) CP Series Water Level Monitoring Instruments manufactured by Remote Data Systems Inc., North Carolina). Channel flow into and out of the lakes was gauged periodically using a Scientific Instruments pygmy current meter. Very low flows were determined where possible by noting the time required to collect a measured volume of discharge. Triplicate samples were taken to obtain an average flow. A 45 degree sharp crested weir (Figure 3.4) was installed to enable gauging of flow from Melvin Lake into Shadow Lake as no suitable gauging locations existed. The discharge measurements were related to the lake level to empirically establish rating curves that convert lake level records into hourly discharges (Figure 3.5).

3.5: Experimental Slope Runoff

The outlet of the experimental slope catchment was equipped with a 45° v-notch weir to enable the gauging of slope discharge (Figure 3.6). Water level of the meltwater runoff pond was monitored using an automatic water level sensor (Ecotone (r) CP Series Water Level Monitoring Instruments) and a rating curve was derived (Figure 3.5 and Table 3.2) allowing half hourly discharge to be calculated. The areal extent of the pond was mapped daily.

3.6: Groundwater Monitoring

To obtain an approximation of the magnitude and timing of groundwater storage and flow in the study catchment, two transects, each with ten groundwater wells, were installed, in a wetland and in a relatively dry bottomland. Each well was reinforced by a perforated PVC pipe of 35 or 48 mm diameter, extending down to the bedrock or the permafrost. The level of three wells was recorded half-hourly by Ecotone Water Level Monitoring Instruments, while the remaining well levels were measured manually on a daily to weekly basis depending on changing groundwater conditions. Hydraulic conductivity was obtained by pumping tests described by Luthin (1966). Frost table adjacent to each well was measured by pounding a steel rod into the ground until the ice-rich frozen soil was encountered. The emergence of surface flow along a channel in a wetland was measured using velocity-area method, with velocity obtained by a Scientific Instruments pygmy current meter.

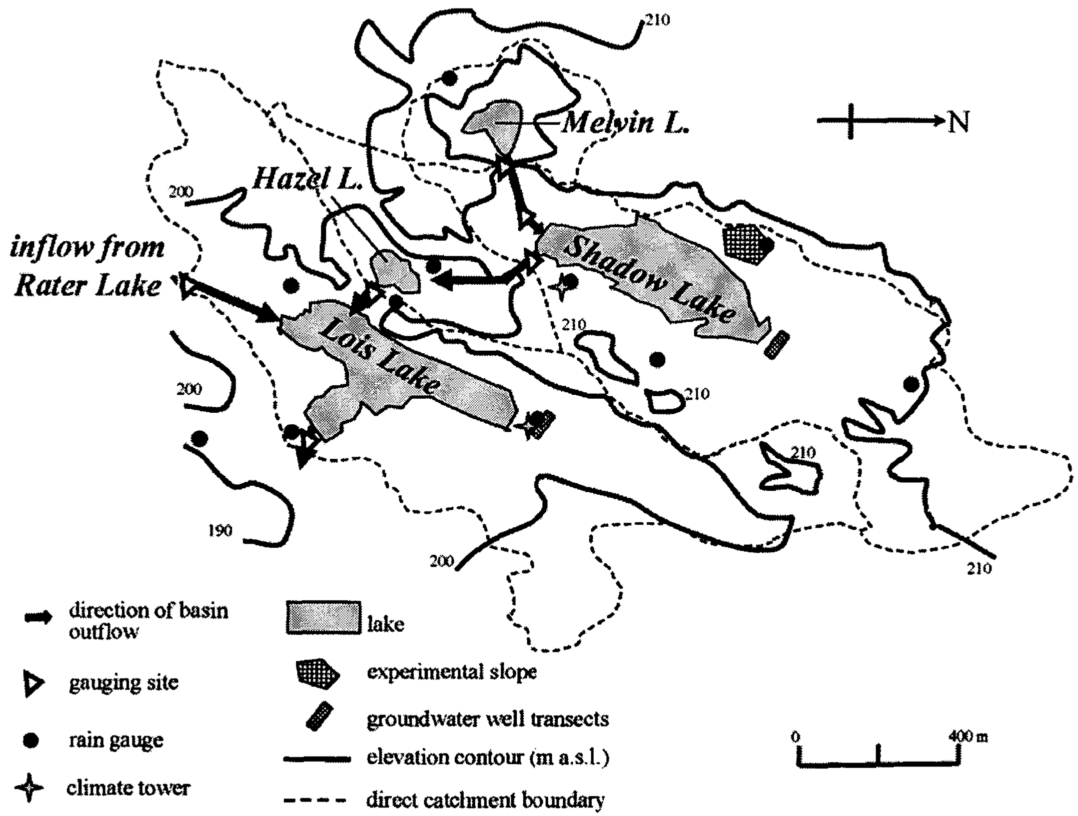

Figure 3.1: Topographic map of study catchment showing measurement sites, instrument deployment.

Figure 3.2: Photograph of bedrock upland climate tower.

Figure 3.3: The upper section of the soil-filled bottomland climate tower showing instrument deployment.

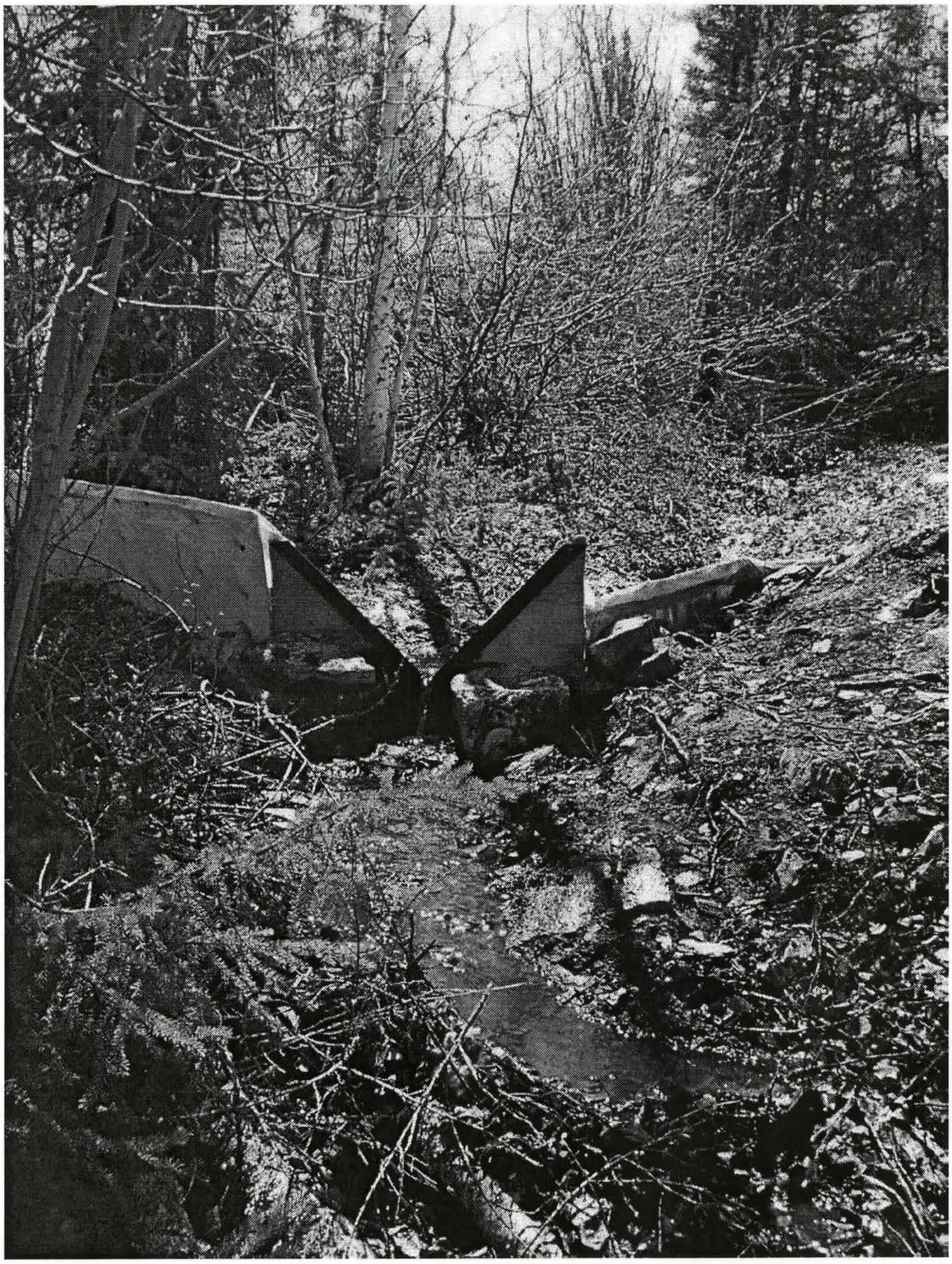

Figure 3.4: Weir along outlet flow path connecting Melvin Lake discharge to Shadow Lake.

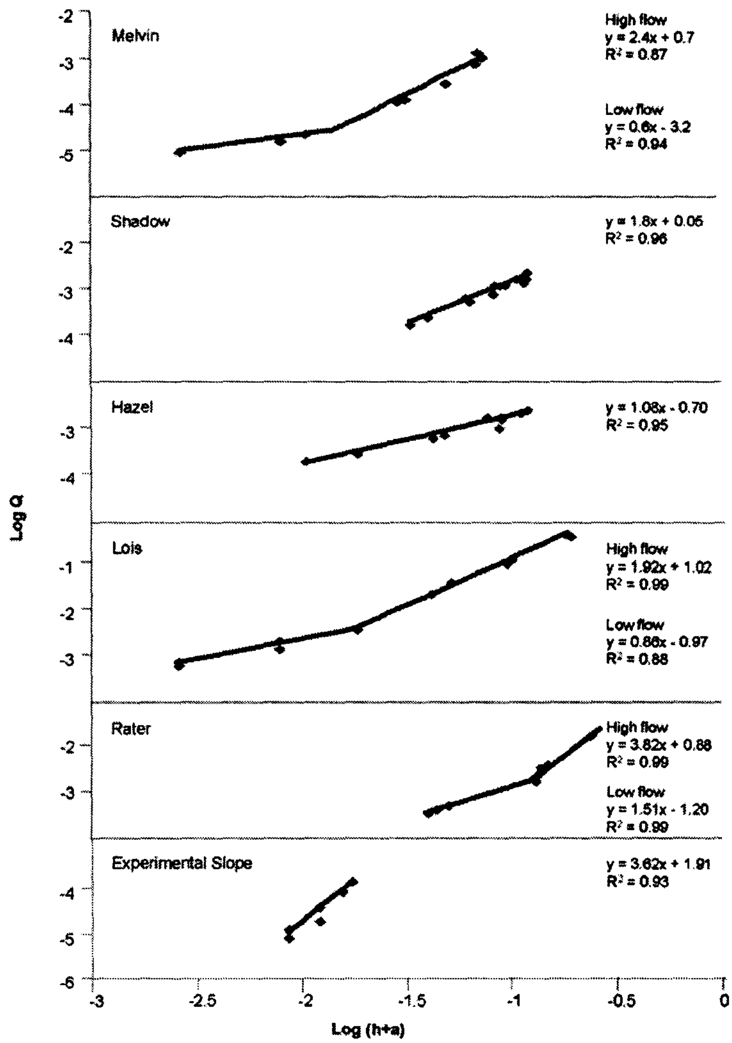

Figure 3.5: Empirical rating curves for Melvin, Shadow, Hazel, Lois, Rater and the experimental plot outflow. Equation in the form of log Q = N log h + log C, where Q is discharge in cubic meters per second, h is lake level above outlet threshold in meters, and both C and N are empirically derived constants.

Figure 3.6: Photograph of Experimental slope outlet weir and Ecotone Water Level Monitoring Instrument at the experimental slope outlet.

CHAPTER 4: SNOWMELT RUNOFF PROCESSES IN LAKE CATCHMENTS AND LAKE BASIN WATER BALANCES

Objective of this section are to investigate the manner in which runoff is delivered to lakes and to understand the role of Shield lakes in streamflow generation during the wet spring snowmelt period (May 1 – June 24, 2004). 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?

Volumetric instead of depth measurements were used here to avoid the complication of having to make conversions to adjust the depth measures every time the discussion is switched between basin, lake and plot scales. Depth measurements are periodically provided where appropriate to enable easier comparison between the results of this study with others. Lake water balance in depth measurements for the period of May 1 to August 31 will be provided in chapter 5.

4.1: Snowmelt and rainfall

At the end of winter, much snow was captured in the low lying areas, whereas the uplands had less but more variable snow due to exposure to drifting and microtopography (Figure 4.1). Average snow water equivalent (SWE) for the bedrock, valley, wetland and lake were 92, 105, 102, and 107 mm, respectively. Weighting these values by the areas occupied by the various terrain types yielded a pre-melt snow storage of 99 mm SWE, the equivalent of 159 000 m^3 SWE for the entire study basin. The four sub-basins, Melvin, Shadow, Hazel and Lois had SWE depths of 98, 98, 97 and 100 mm which correspond to snow storages of 12 000, 55 000, 16 000 and 76 000 m^3 (Table 4.1) respectively.

Snowmelt commenced on May 5 and daily snow ablation for each terrain type is presented in Figure 4.2. These ablation values were used to produce gridded daily maps of the residual snow cover (Figure 4.3). The snow cover was relatively continuous until May 12. In the next two days, most parts of the eastern catchment became bare. By May 15 only the lake and major valleys retained residual snow while over 90% of the basin snow cover disappeared by May 20. The entire study basin became snow free on May 24 which corresponded to field observations of the snow-free condition.

The weighted daily snowmelt and the changing fraction of snow free area of each sub basin are shown in Figure 4.4. Initial snowmelt proceeded at a moderate rate yielding roughly 3 mm/day, corresponding to quantities of 500, 2500, 700 and 3300 m^3/day. The cold spell of May 9 produced less than 300 m^3 of melt. Accelerated melt on May 14 reached almost 19 600 m^3, and the high melt rate continued on the next day, quickly depleting the residual snow cover as an additional 15 600 m^3 was melted. A reduction of the daily melt contribution after May 15 was due to a rapid decrease in the snow coverage.

Rain fell on May 30, June 2 and 6, after all the snow had disappeared. Total rainfall ranged between 11 to 13 mm in the four sub basins, which amounted to 1500, 6300, 1900 and 8200 m^3 (Figure 4.5). For the entire snowmelt study period a combined melt and rainfall of 110 mm (equivalent to 177 100 m^3 tallied from the four sub basins) reflects the relative dryness of the environment.

4.2: 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 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 and depressions. Field observations confirmed that the water in these zones either infiltrated or was lost to evaporation. Such areas do not contribute direct runoff to the lakes and are considered to be areas where flow delivery to the lake is retarded (Figure 2.3). A final source of water comes from upper lakes which can drain into a lake through either a poorly or well defined channel.

Runoff from upland

Result from the experimental slope enables an estimation of runoff contribution from the upland areas. The bowl shaped depression at the foot slope collected slope runoff, creating a temporary 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 loss from the pond increased as the pond area expanded but diminished when the pond shrank. Using these results, daily upland runoff (Q_{upland}) was evaluated by

$$Q_{upland} = \Delta S_{pond} + Q_{pond} + E_{pond.} \qquad (4.1)$$

Figure 4.6 shows the daily values of change in pond storage (ΔS_{pond}), pond outflow (Q_{pond}) and evaporation (E_{pond}). The storage change from day t to day t+1 was obtained by:

$$\Delta S_{pond} = (H_{t+1} - H_t)\,(A_{t+1} + A_t)/2 \qquad (4.2)$$

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 Equation 4.1 was 259 m^3, equivalent to 49 mm per unit slope area.

Runoff from bottomlands

The presence of ice rich frozen soil (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 the descent of the water table. Measurements of the frost and water tables along two lines of wells, one in a wetland and the other in a non-wetland setting showed that the saturated layer was relatively thin (in the order of centimetres) (Figure 4.7). Darcian flow across each transect line was calculated by

$$Q_g = K\sum(\Delta h_i / \Delta L_i) z_i w_i \qquad (4.3)$$

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; z is thickness of the saturated thawed 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.

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 dry bottomlands was likely to be minimal. For the wet site, matrix flow totaled 22 m^3 across a width of 100 m, for the period May 18 to June 24. Although several soil pipes were noted in the peat (c.f. Carey and Woo, 2000), their limited occurrence suggests that they contribute very little to ground water delivery. The wetlands were inundated throughout the spring (Figure 4.7) but most of the water was retained as depression storage. Surface flow was from one wetland, however, yieldeing a measured discharge of 4500 m^3 to Shadow Lake on May 25-26.

Inflow from upper lake catchments

Melvin is the uppermost lake in the study basin and as such does not receive this input, however it acts as a source for the lower Shadow Lake. On May 25, inflow to Shadow Lake from Melvin Lake Catchment began. Inflow rose to a peak on June 4 and then declined until June 20 when flow into Shadow Lake ceased, with less than 20 m^3/day discharged from the upper lake. The total amount delivered to Shadow Lake reached 7800 m^3. Hazel Lake began to receive inflow from Shadow Lake sub basin on May 22. Inflow peaked May 27 and ceased on June 24, after a total of 980 m^3 of water was inputted into Hazel Lake. It is significant to note that the beginning of outflow from Melvin 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. Lois received input both from Hazel Lake sub basin and from the Rater Lake catchment, which drains into Lois. Hazel inputted 930m^3 to Lois during the period of May 20 to June 14. Inflow from Rater Lake catchment did not reach Lois Lake until May 30 due to the presence of solid ice blocking the interior of a culvert which drains the lake. Inflow peaked on June 6 and steadily declined until July 26 when it rose slightly before finally ceased on August 5. Inflow from this catchment was substantial, contributing 170,000 m^3 to Lois Lake by June 24.

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 land evaporation averaged 1.2 mm/d. These values lie 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 spring period total catchment evaporation amounted to 122 200 m^3, with 8700 from Melvin, 26 800 from Shadow, 13 100 from Hazel and 73 600 from Lois.

4.3: 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 its effect on lake level.

Lake ice and evaporation

Arctic and subarctic lakes are invariably covered by ice at the end of winter. The presence of a lake ice cover limits evaporation loss until open water condition develops. In 2004, all four lakes experienced very similar patterns of ice decay. Meltwater first reached the lakes from the hillslopes to slush the snow along the rims of the lakes. Next, the shorefast ice began to melt, creating moats of open water on May 16 (Melvin and Hazel) and May 18 (Shadow and Lois) as the lake ice floated free from the shores. Progressive attrition of lakes ice increased the fraction of open water on the lakes until the ice was fragmented and finally dissipated (Figure 4.8). The two smaller lakes, Melvin and Hazel became ice free on June 3 with the larger Shadow and Lois lakes become ice free on June 6. 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 4.8). For the period after the loss of the ice cover, lake evaporation averaged between 3.9 and 4.8 mm/d which compares favourably with the average rate of 4.5 mm/d obtained at Skeeter Lake (100 km north of Yellowknife) when the latter became ice-free (Spence, 2000). Over the entire spring study period, Melvin, Shadow, Hazel and Lois lake evaporation totaled 1200 m^3, 9700 m^3, 900 m^3, and 10,800 m^3 respectively.

Lake storage and outflow

Lake storage increased steadily as meltwater runoff reached the lakes (Figure 4.9). 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 (Figure 4.6), suggesting that it was the fast delivery of meltwater to the lake that occasioned the production of substantial outflow.

Like many small northern lakes, the 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. Lake storage depletion continued for the remainder of the summer and outflow was not revived for the rest of the year.

Neither Melvin, Hazel nor Lois experienced any apparent ice or snow outlet blockage, with discharges beginning when the lips of their outlets were topped. Melvin and Hazel both experienced sharp hydrograph rises peaking at 0.01 m^3/s and 0.02 m^3/s respectively. These were followed by similar outflow recessions with brief responses to the late rain event, also evidenced by Shadow. Flow stoppage for these Melvin and Hazel lakes occurred on June 26 and June 14 respectively. Some outflow seeped through a small soil covered beaver dam at the outlet of Lake Hazel until July 6. Seepage through the dam was measured as it flowed down the steep bedrock drop to Lois Lake and averaged 8 m^3/day.

Lois Lake began discharging on May 24 and experienced two large peaks in outflow (Figure 4.9d), the first peak is attributed to basin snowmelt runoff and the second to the delayed streamflow input from Rater Basin.

4.4: Catchment Water Balance

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

$$M+R-E+Q_{in}-Q_{out}=\Delta S \qquad (4.1)$$

where M is snowmelt, R is rainfall, E is evaporation, Q_{in} is inflow from upper lake catchment and Q_{out} is lake outflow, and ΔS is basin storage. Substituting into

Equation 4.1 the measured and calculated magnitudes of Shadow (in thousands of m^3) for the spring period, defined here as between the initiation of snowmelt and the cessation of lake outflow (May 5 to June 24, 2004), yields

$$55 + 6.3 - 27 + 7.8 - 24 = 18.1$$

After subtracting the 900 m^3 of storage increase in the lake from the total basin storage of 18,100 m^3, there was still a surplus of 17,200 m^3. This surplus can be attributed to the recharged groundwater and soil moisture that were still retained in the uplands and bottomlands. Some of this water would eventually be released to the lake in the summer (based on summer field data presented in Chapter 5). For comparison purposes, Table 4.1 provides seasonal catchment water balance values for all 4 sub basins.

For the study period, the runoff ratio of Shadow Lake basin, i.e. $Q_{out}/(M+R)$, was 0.4 (Table 4.1). 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) (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 to draw down 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 lakes.

4.5: Discussion

Field mapping shows 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 groundwater flow in the soil, and (3) retarded flow delivery where no effective flow pathways exist to convey meltwater runoff to the lake during the melt season. The mechanisms and rates of snowmelt runoff to the lake, as well as the lake ice cover duration and open water evaporation rates, are significant considerations with respect to outflow generation. Losses from evaporation much greater than rainfall inputs place a significant demand on lake storages, which hinders the re-establishment of outflow in the summer period.

A conceptual framework is proposed to relate runoff delivery to the frequency and magnitude of outflow occurrences (Figure 4.10). Within a catchment, zones not contributing directly to immediate runoff will find that meltwater and rainfall either infiltrates (and may re-emerge much later as groundwater flow) or evaporates from the ponded and soil-filled areas. Those catchments with large tracts of slow and retarded flow delivery can only 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 limiting the opportunity for evaporation loss in transit. Furthermore, runoff can reach the lake quicker than lake evaporation can draw 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.

Given that the snow cover is usually thin on the lake, most water comes from the basin slopes instead of directly from the snow on the lake ice. To generate lake outflow in the spring, as in other seasons, three factors are considered important in terms of raising the lake storage to the critical level at which outflow can commence.

(1) Rate of runoff delivery (Figure 4.11a, b): basins with a large area of fast runoff delivery can provide input quickly to the lake to raise its level to the outflow threshold, before lake evaporation has the opportunity to lower the lake level.

(2) Antecedent lake storage condition (Figure 4.11c): a low antecedent storage below the outflow threshold requires much runoff input to raise the water level to the outflow stage. 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.

(3) The ratio of catchment to lake area: a large ratio of basin-to-lake area will likely have the benefit of bringing much runoff from the land into the lake to satisfy lake evaporation and storage demands (and these demands become proportionally smaller as the basin-to-lake ratio increases), thus facilitating the occurrence of outflow.

Table 4.1: Spring water balances (in 10^3 m^3) of each of the four sub-basins and combined for the total study catchment for the period of May 1 to June 24.

	Melt	Rain	Evaporation	Inflow	Outflow	Storage change
Melvin	12	1.5	8.7	0	8.3	-3.5
Shadow	55	6.3	26.8	7.8	24.5	17.8
Hazel	16	1.9	13.1	24.5	6.5	22.8
Lois	76	8.2	73.6	355.3	253.2	112.7
Total	159	17.9	122.2	387.6	292.5	149.8

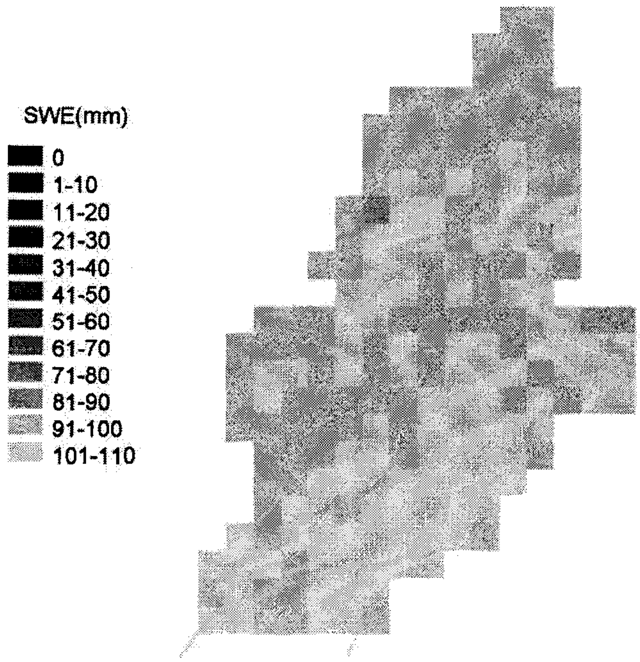

Figure 4.1: Snow water equivalent (SWE) distribution map (May 4) showing initial snow conditions at beginning of the study period. Full line shows sub-basin boundary.

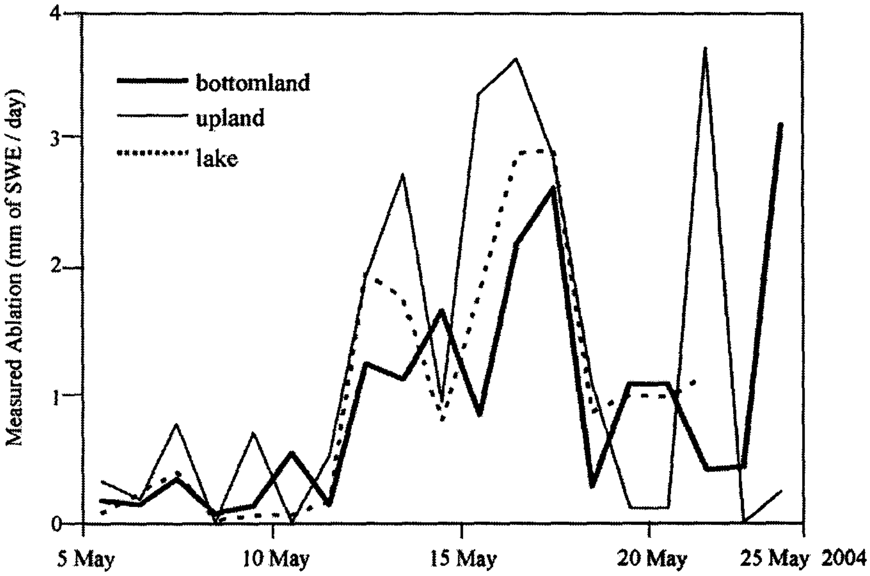

Figure 4.2: Daily snow ablation on three types of terrain, including upland, bottomland and lake.

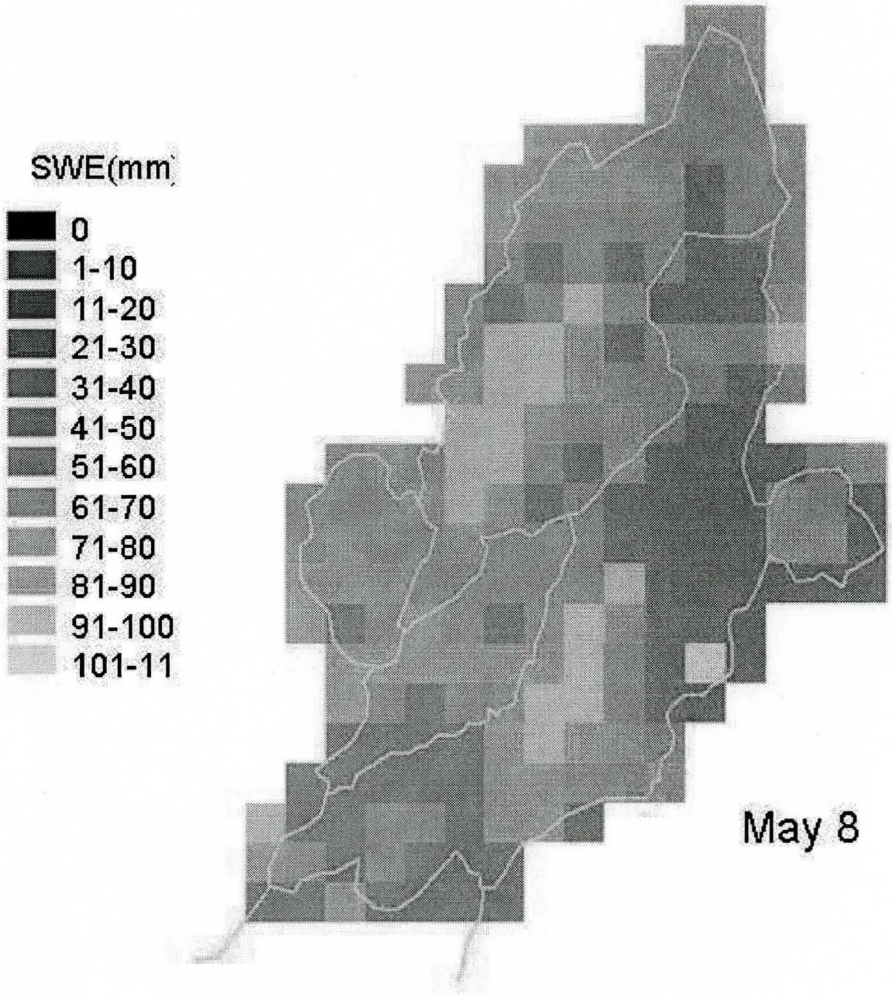

Figure 4.3a: Snow water equivalent (mm) distribution pattern on May 8. Full line shows sub-basin boundary.

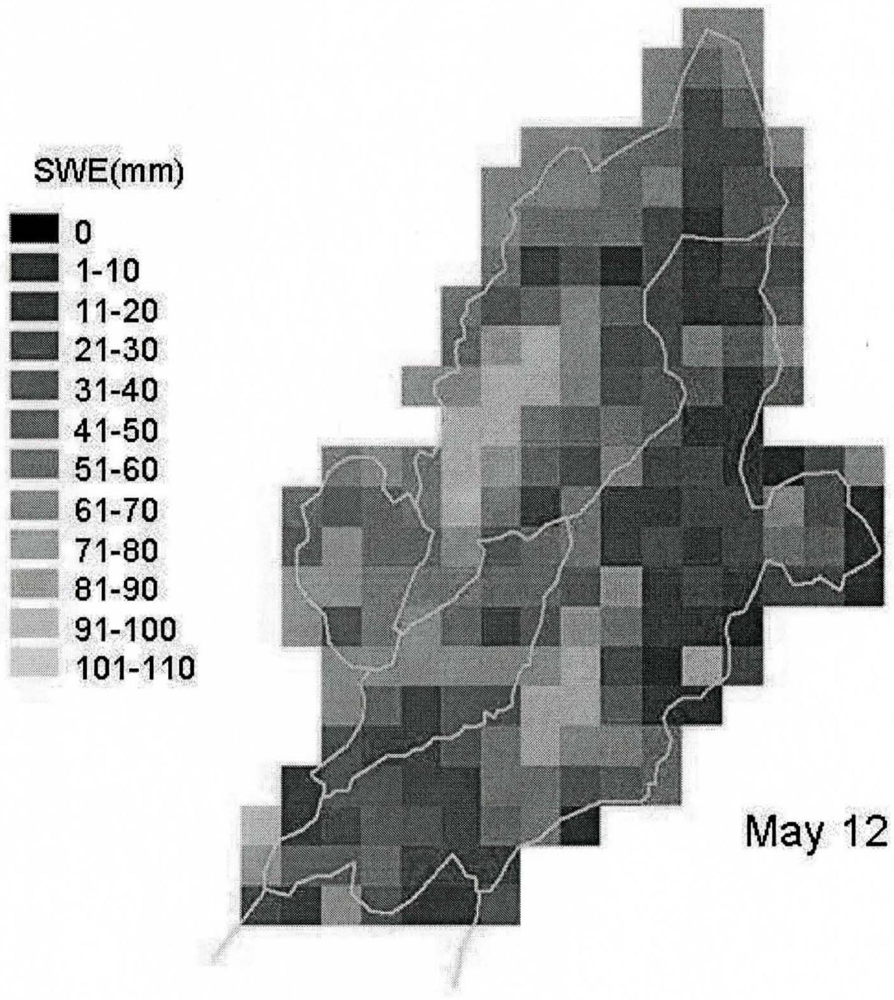

Figure 4.3b: Snow water equivalent (mm) distribution pattern on May 12. Full line shows sub-basin boundary.

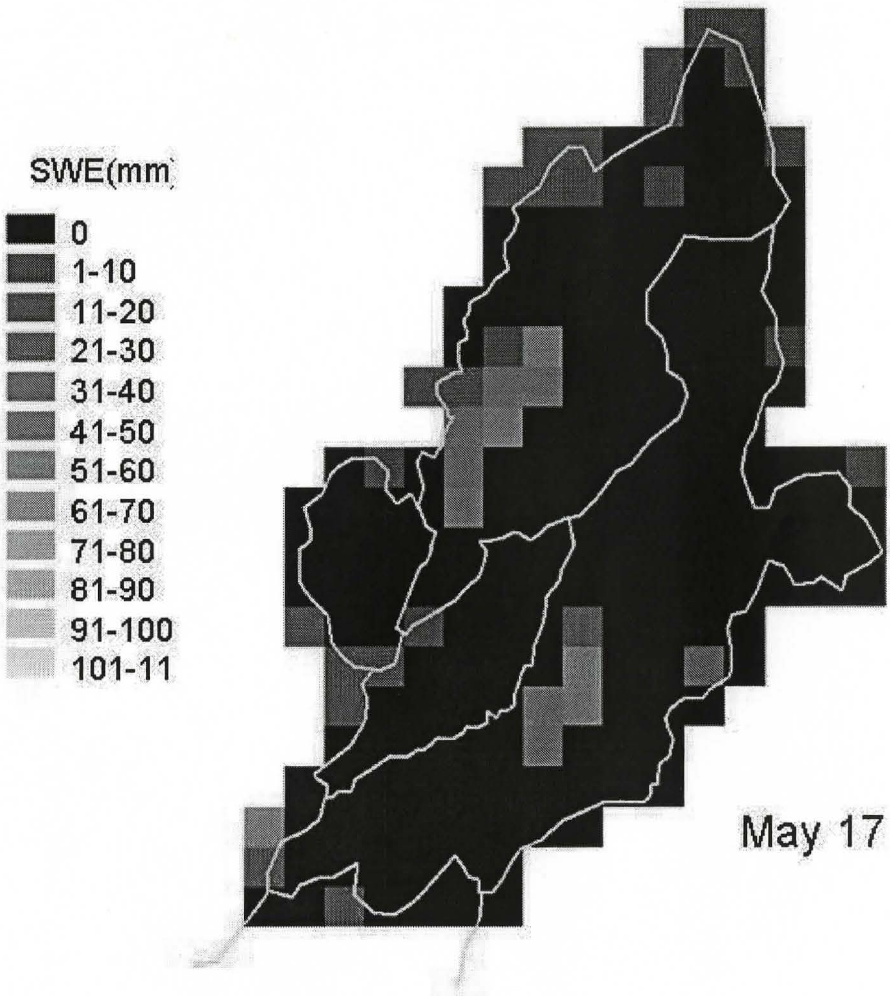

Figure 4.3c: Snow water equivalent (mm) distribution pattern on May 16. Full line shows sub-basin boundary.

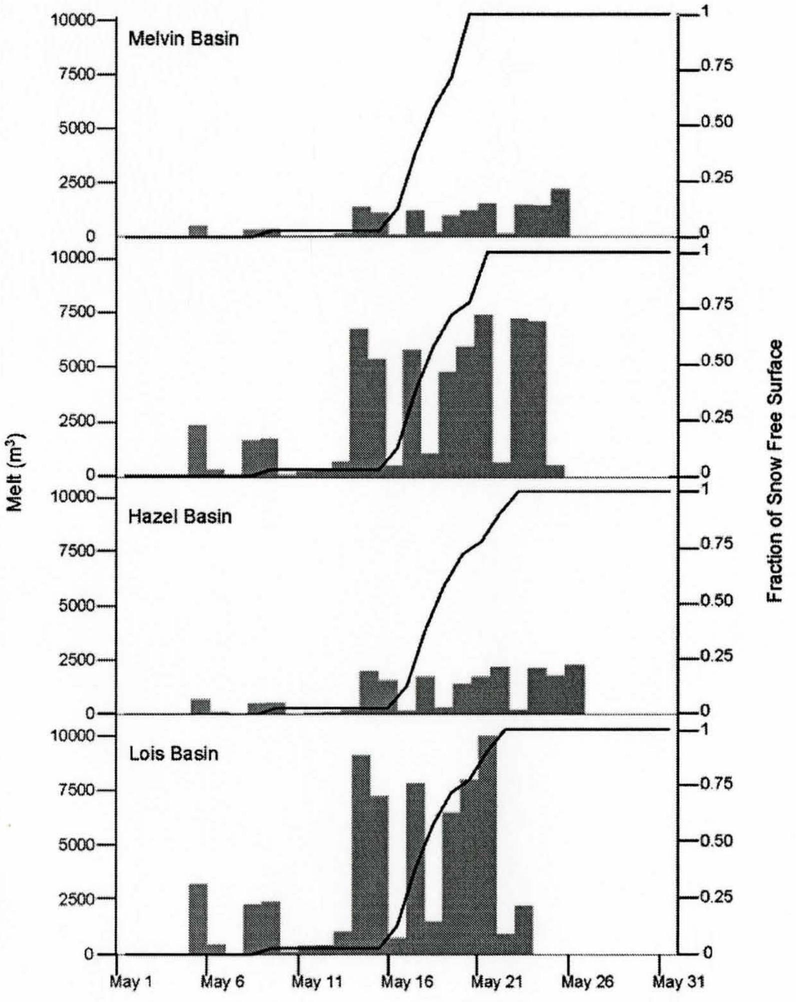

Figure 4.4: Daily snowmelt (cubic meters) and snow-free fraction for each of the four sub-basin catchments.

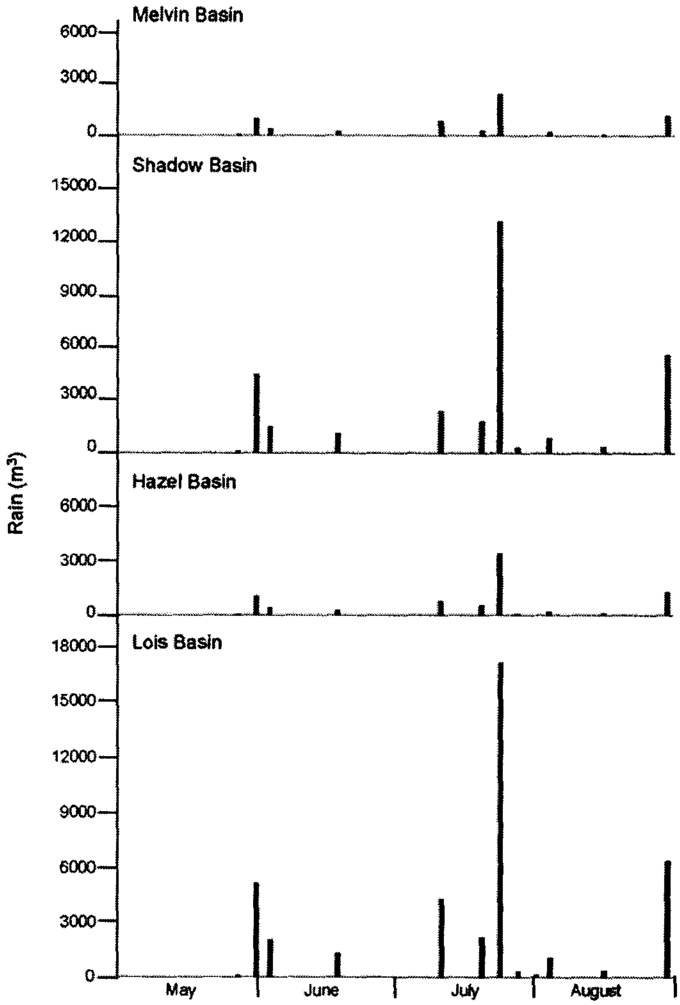

Figure 4.5: Rainfall in cubic meters of water for each of the four sub-basins.

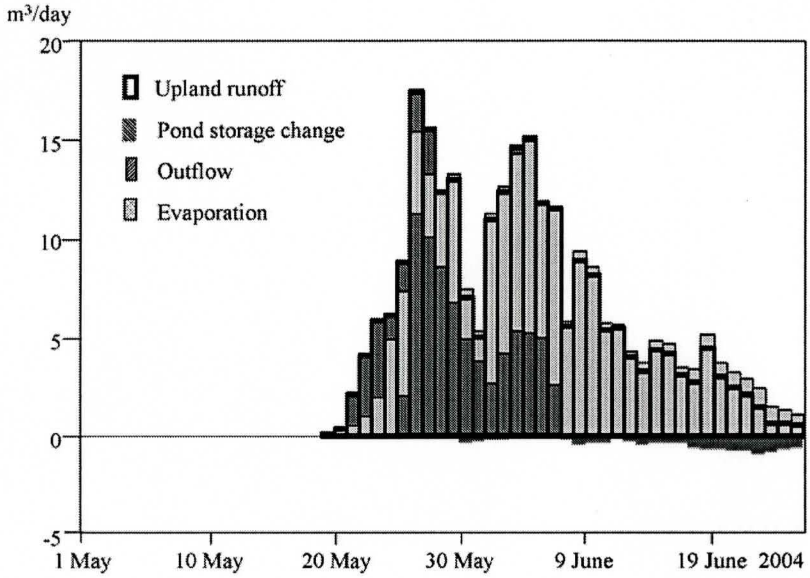

Figure 4.6: Daily upland runoff from the experimental plot, estimated as the sum of pond storage change, pond outflow and evaporation.

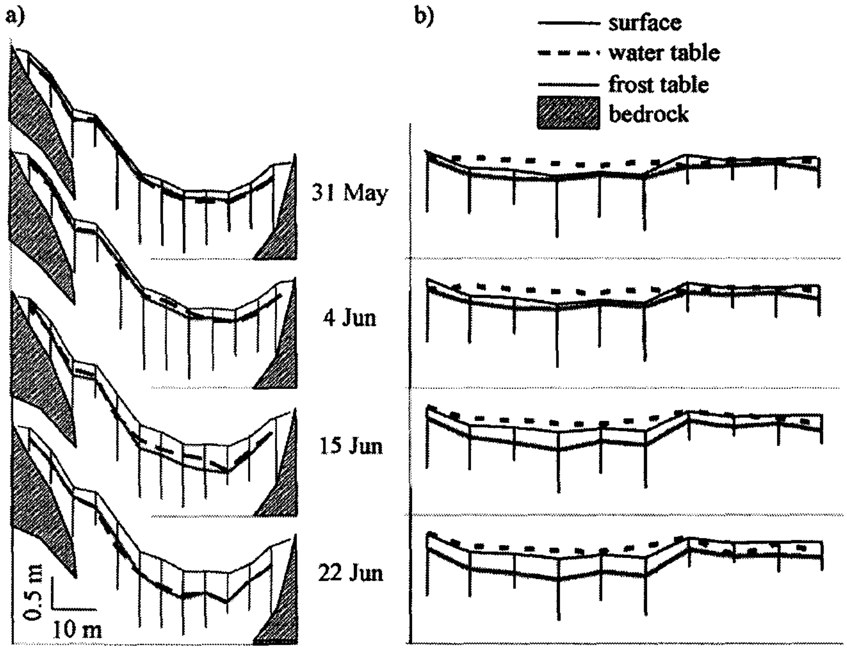

Figure 4.7: Water table and frost table positions on selected days at bottomland sites, along a (a) relatively dry and (b) wetland transect, both oriented perpendicular to the direction of ground water flow to the lake. Vertical lines indicate location and depth of wells.

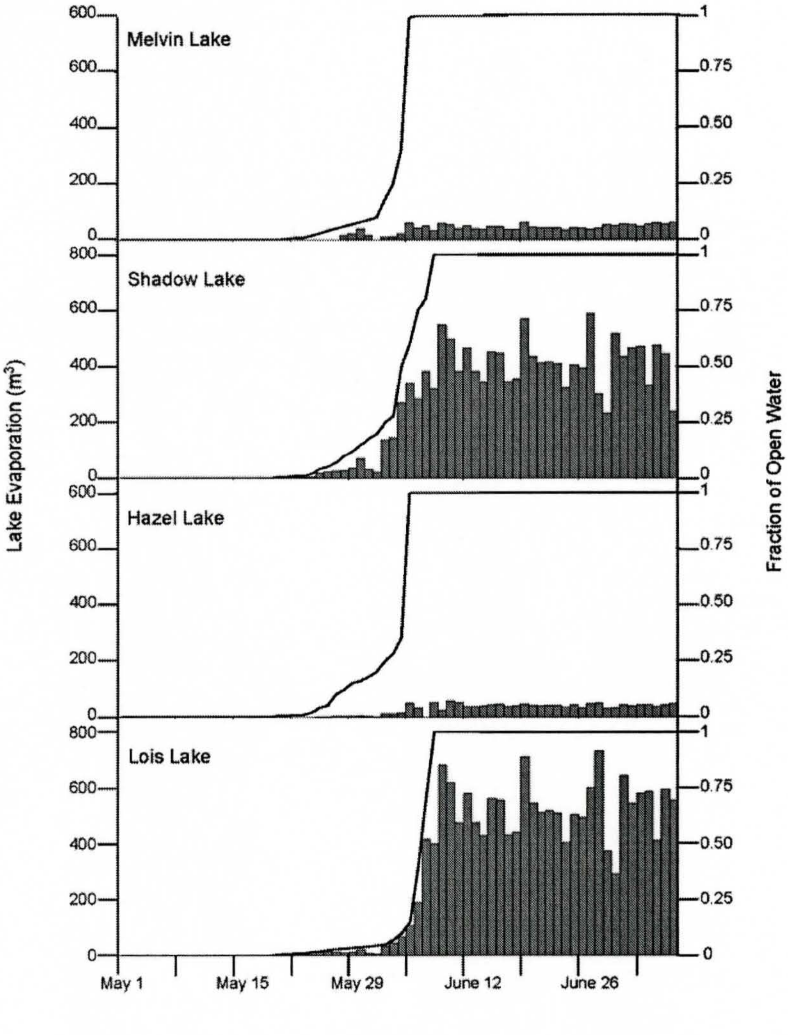

Figure 4.8: Daily evaporation from open water surface, and ice-free fraction of Melvin, Shadow, Hazel and Lois lakes.

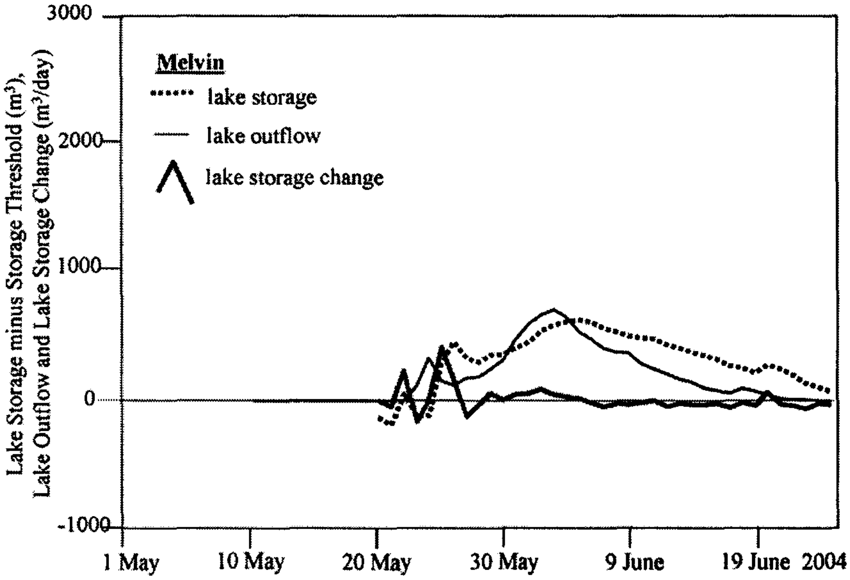

Figure 4.9a: Daily change in lake storage, lake level and outflow from Melvin Lake and storage minus antecedant storage starting at the initiation of lake ice breakup. Zero represents the critical threshold above which outflow occurs.

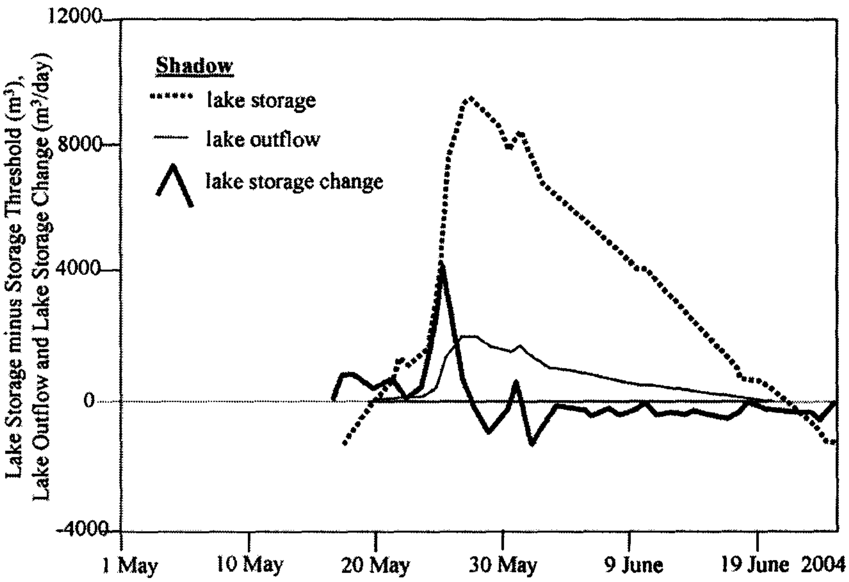

Figure 4.9b: Daily change in lake storage, lake level and outflow from Melvin Lake and storage minus storage threshold starting at the initiation of lake ice breakup. Zero represents the critical threshold above which outflow occurs.

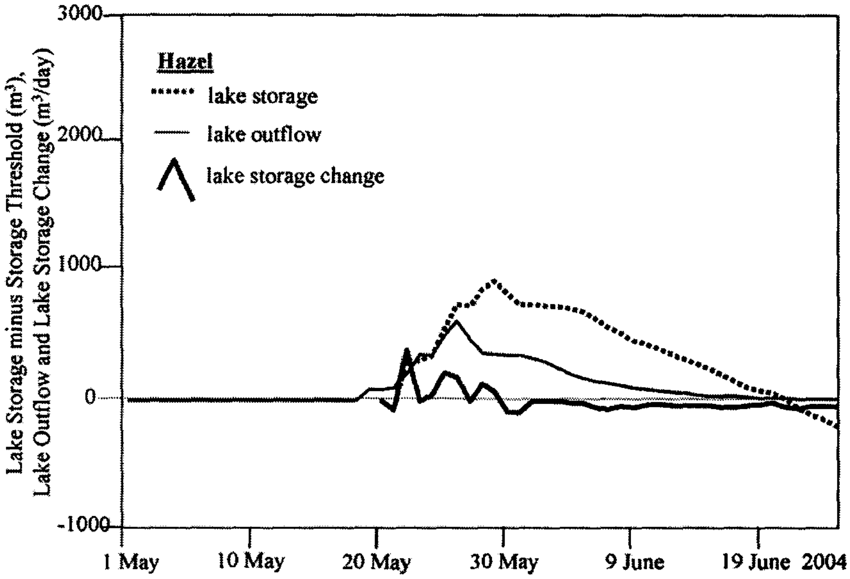

Figure 4.9c: Daily change in lake storage, lake level and outflow from Melvin Lake and storage minus storage threshold starting at the initiation of lake ice breakup. Zero represents the critical threshold above which outflow occurs.

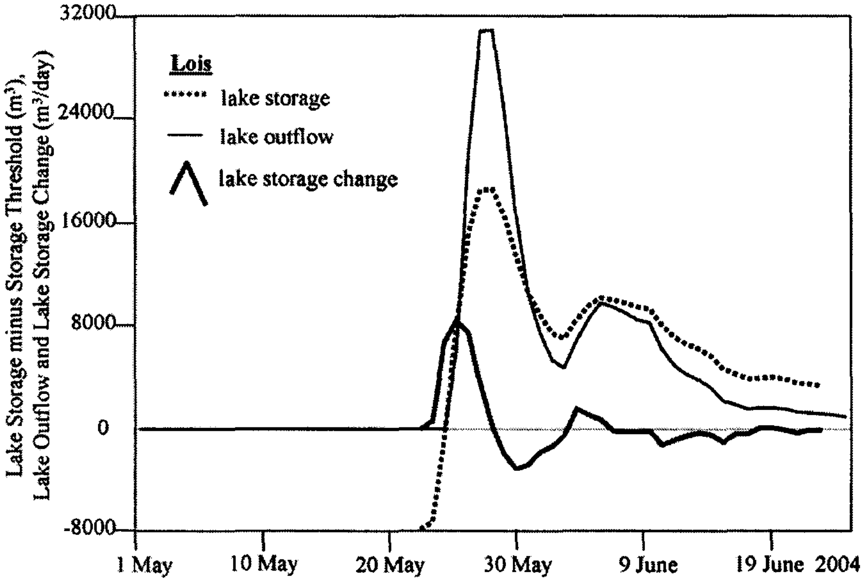

Figure 4.9d: Daily change in lake storage, lake level and outflow from Melvin Lake and storage minus storage threshold starting at the initiation of lake ice breakup. Zero represents the critical threshold above which outflow occurs.

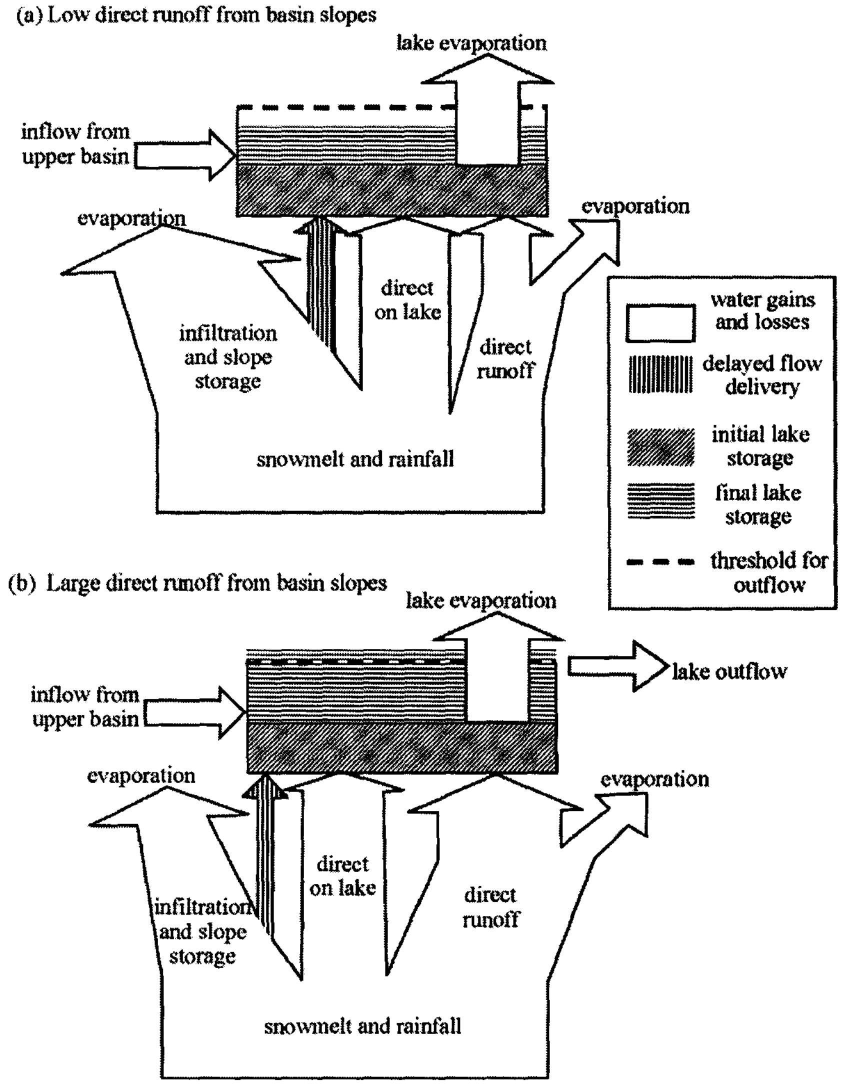

Figure 4.10: Conceptual framework relating runoff delivery to the frequency and magnitude of outflow occurrences; a) with limited direct runoff to the lake, lake storage increase during the melt season is insufficient to reach the outflow threshold, b) large direct runoff from basin slopes enables rapid lake level rise above the critical level to generate outflow.

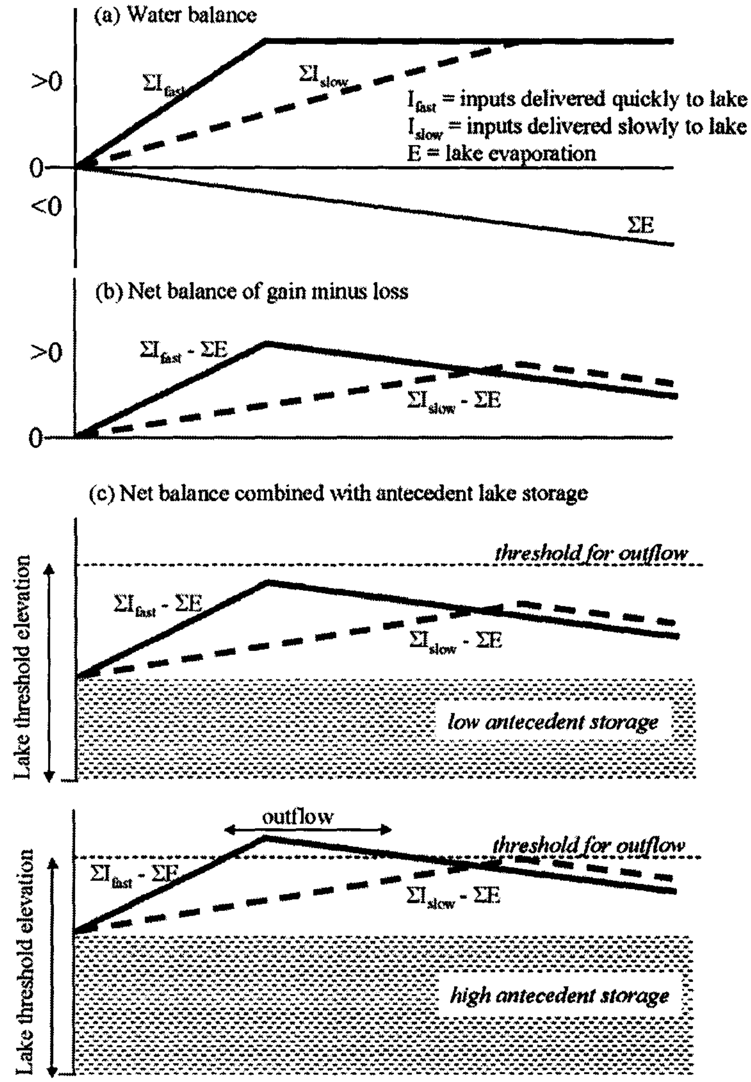

Figure 4.11: Illustration of how a) rate of runoff delivery is important in terms of raising lake storage, as b) a fast rate of delivery can overcome evaporative demands, resulting in a larger storage rise for the same amount of water input. c) Antecedent lake storage is also a critical factor in determining if rising lake storage will exceed the critical level at which outflow can commence.

CHAPTER 5: FLOW CONNECTIVITY OF A LAKE-STREAM SYSTEM

Examination of lake water balance permits an understanding of how each of its component influences lake storage at different times of the thawed season. Chapter four investigated the runoff processes of a subarctic Shield lake and its catchment during the snowmelt season and confirmed the importance of lake storage status in streamflow generation. The chapter will investigate the flow connectivity to understand the processes operating in a lake-stream drainage network and to suggest a method for improved algorithm development to assist in the modelling of Shield hydrology in a semi-arid environment abundant with lakes.

5.1 Lake Hydrology

Lake Water Balance

Examination of lake water balance permits an understanding of how each of its components influence lake storage at different times of the thawed season. The water balance of a lake is (expressed in mm/d):

$$\Delta S = R - E + Q_i - Q_o + Q_b^* + \xi \qquad (5.1)$$

Here, rainfall on the lake (R), storage change based on lake level fluctuation (ΔS), flows along stream channels into (Q_i) and out of the lake (Q_o) were measured directly and lake evaporation (E) was calculated by the Priestley and Taylor method using measured variables. Q_b^* is the net flux of water delivered to or leaked from the lake, via overland or subsurface flows in its direct catchment area; and ξ is the error in the evaluation of the water balance components. Both Q_b^* and ξ cannot be assessed independently. Assuming that ξ is small relative to the magnitude of various components of the water balance, Q_b^* was obtained as a residual term by rearranging Equation 5.1. Water balance calculations were performed after much of the basin snow has melted, at the onset of lake ice break-up. Figure 5.1 is a plot of the daily values of water balance for the four lakes studied and Table 5.1 provides their total magnitudes for the study period.

There was a large water influx Q_b* derived from snowmelt on the basin slopes (Chapter Four). It was mainly the overland and subsurface contributions that raised the lake storage, since there were few rainfall events during the spring of 2004. After the melt period (all the basin snow disappeared by 26 May), Q_b* became negative, suggesting that there has been a continuous groundwater loss from the lakes. Despite the bedrock structure, seepage loss is highly plausible because the rock fractures can be effective conduits to convey water (Thorne et al. 1999). Lake evaporation increased steadily in May when the lake ice cover diminished. Afterwards, lake evaporation averaged 3.3 mm/d and was the main process responsible for lake storage decline. A 20-22 mm rainstorm event in late July raised the water levels in all four lakes. Lake inflow and outflow were observed at all the lakes during the snowmelt period but summer rainfall events raised channeled outflow from only one lake (Lois).

Several generalizations can be drawn from the water balance study. At the beginning of the study period and prior to snowmelt, there was zero flow in the channels that enter or leave the lakes. This was a time when lake levels were below the elevation of their outlets. Slope runoff in the melt season was the dominant process that led to a sharp rise in storage. In this regard, both the intensity and the amount of flow are important, hence the rate and duration of snowmelt on the basin slopes are significant considerations (Chapter 4). Evaporation, inhibited when a lake ice cover was present, continued throughout the summer to deplete lake storage. Although rainfall can produce a rapid rise in lake level, the magnitude can seldom match the rise due to snowmelt runoff to the lake because intense rain is seldom sustained in the semi-arid environment. Water balance calculation indicates that a lake can be recharged by slope runoff and can also lose water through subsurface conduits. Major hydrologic exchanges are between the lakes and their local catchment area, but other than the uppermost Melvin Lake, all other lakes receive periodic stream inflow from the lakes above them.

Lake Outflow Generation

The beginning and terminating dates of outflow from various lakes are shown in Table 5.1. The snowmelt season is the primary period for outflow generation. Of note is that the starting and ending dates differed among the lakes. Hazel Lake, the third along the chain of lakes, was the first to generate outflow, followed by Shadow Lake above it. Thus, there was no systematic downstream sequencing in the starting or ending dates of flow, implying that the discharge from a lake higher up in a lake-stream system is not necessarily responsible for the generation of outflow from a lake lower down the lake-stream network.

For the entire study period, Melvin Lake, the highest in elevation but with the smallest drainage area, had the lowest volume of outflow. On the other hand, Lois Lake which receives inflows from both Hazel and Rater lakes, yielded the largest volume of outflow. Shadow and Hazel lakes had intermediate and comparable volumes but different timing of outflow. For all the lakes, snowmelt was the dominant, if not the only period when lake outflow was produced. Rainfall was able to rejuvenate outflow only for Lois Lake.

Field observation of outflow occurrence shows that a lake has to rise above the lip of its outlet which marks the flow threshold. This threshold is normally the lowest point along the perimeter of the lake and can be a bedrock sill, a channel carved in soil, the bottom of culvert such as for Rater Lake, or a blockage by a beaver dam as in the case of Lois Lake. In the spring, snow drift and ice often raises the threshold elevations so that the lakes would be impounded to a greater height than in the summer (Figure 5.2). Such situations have been reported for other lakes in the subarctic (FitzGibbon and Dunne 1981) and the Arctic (Woo et al. 1981).

Lake Storage and Outflow

Field information can be combined with water balance analysis to relate lake storage with outflow. The fill and spill concept can be extended to the lake-stream flow system. For a lake, outflow (Q_o) occurs when its storage level exceeds the flow threshold (S_T):

$$Q_o = 0, \quad \text{if } (S_{t-1} + \Delta S_t) < S_T \tag{2}$$

where S_{t-1} is lake storage level at end of the past period t-1; ΔS_t is the change in storage for the current period t, obtained by water balance through Equation 6.1. Thus, whether outflow can be produced is predicated upon antecedent storage (S_{t-1}), the change in storage for time t, and the threshold that must be exceeded and this threshold can be higher in the spring when the lake outlet is blocked by snow and ice.

It is previously thought that a small lake area relative to its catchment size (<5%) will not significantly attenuate streamflow because lake storage then plays a relatively little role compared with the magnitude of basin runoff (FitzGibbon and Dunne 1981, p. 282). Table 5.1 shows that Lois Lake with a low lake to basin ratio of 2% was able to arrest the flow from Rater Lake throughout most of the summer. This suggests that a low ratio is not a satisfactory indicator of flow attenuation; rather, it is the capacity of lake storage relative to water input that is of concern.

5.2: Flow Connectivity of Lake-Stream Systems

Outflow generation does not necessarily proceed systematically from the uppermost to the lowest lake, nor does outflow stoppage follow any ordered sequence along the chain of lakes. Hazel Lake in the middle of the chain was the first to generate and terminate outflow, whereas Shadow Lake above it continued to discharge for almost two more weeks. As flow connectivity does not progress systematically along the drainage network, each lake can form its subsystem that may or may not coalesce after certain hydrologic events.

Physical setting of a lake plays an important role in terms of it hydrologic connectivity within the flow system. The Precambrian Shield rock is mainly composed of impervious granite and gneiss with an occasional veneer of soil cover on the uplands to allow efficient shedding of rain and meltwater, though infiltration is encouraged where fissures abound (Spence and Woo 2002). Fast delivery of water from the basin slopes enables rapid increase in lake storage, permitting quick rise of lake level above the outlet threshold to produce outflow. Under cold subarctic conditions, a long duration of the lake ice cover shortens the evaporation season. Lake level drawdown due to evaporation is confined to the ice-free period but daily evaporation in the summer can be large because of long daylight hours. Finally, a semi-arid climate ensures that post-snowmelt evaporation loss exceeds rainfall input so that the summer water balance favors lake level decline that leads to periods of no outflow. Outflow cessation is less commonly reported for Shield lakes in temperate areas due to lake storage replenishment by higher rainfall.

In a humid environment, discharge often increases downstream and this may be attributed to an enlarged area of flow contribution from the basin slopes to exceed the retention capacity of channel storage. The presence of lakes represents a significant storage along the drainage channels so that lakes modify the shape of the hydrographs. In semi-arid and arid areas, the vertical water balance of the lake and its direct catchment area (here referred to as the basin area that feeds directly to the lake and exclusive of the area contributing to streamflow above the lake inlet) can overwhelm the magnitude of channel inflow-outflow to distort the downstream change in flow normally exhibited in rivers of the humid region. Flow interruption arises when flow connection between a lake and its adjacent streams are severed. The hydrograph of Lois Lake outlet offers such an example. Outflow from this basin was maintained for only 49 days in 2004.

The same fill and spill processes apply to other headwater catchments and to larger drainage systems in a lake-studded landscape under cold, semi-arid conditions. Baker Creek (62°30'48"N; 114°24'34"W) offers examples that a larger basin (area 121 km^2 or two orders of magnitude larger than the study basin) also experienced flow discontinuities in the summer (Water Survey of Canada, 2003).

This Shield basin has a number of lakes that form parts of a lake-stream network. Figure 5.3 shows that in 1998, flow ceased on August 19-24, but was revived by rainfall in early November. In 2001 and 2004, Baker Creek discharge fell below 0.01 m^3/s after August 19 and remained so for half a month. Similarly in 2004, its flow dropped below 0.01 m^3/s in mid September. Such zero or negligible flows occurred late in the dry summers after much water was lost to evaporation. It is proposed that the severance of connection among the lakes effectively reduced the source areas that maintained flow for the Creek, as discussed by Spence (2006) who mapped the changing runoff contributing areas in the basin for two summers.

Fill and spill of individual lakes represents the major mechanism for streamflow generation in the Shield lake-stream system. For each lake in the system, not only the channel inflow but its storage capacity and the water balance of its direct catchment area should be represented explicitly. Figure 5.4 schematizes the flow connectivity of three lakes in a cascading system, each with its storage threshold (S_T) that needs to be surpassed, and the antecedent storage (S_{t-1}) that is updated by the lake water balance at time t (ΔS_t). Inflow and outflow in this case may play a minor role in ΔS compared with the vertical and direct catchment exchanges (Q_b* in Equation 1) in the water balance. Then, each lake and its direct catchment may be considered to operate as an independent entity in the chain, subject to variable channel flow linkages with the entities above or below.

5.3: Modelling a Lake-Stream System

The fill and spill concept represents the principal mechanism of streamflow generation in the Shield lake-stream system and it can be applied to the modeling of flow connectivity along a chain of lakes. Hydrologic models usually use routing procedures to treat flow attenuation along channels, or reservoir schemes to represent flow retention and release. Lake storage plays a deciding role in outflow production in natural lake-stream systems, particularly in a dry environment. A hydrologic model conceptualization and framework will be presented to simulate flow connectivity within a lake-stream network by account for lake storage regulation effects on streamflow.

A lake-stream network consists of a number of lake basin (Figure 5.5) elements linked to each other topologically by surface channels. Each element may serve the hydrological functions of providing storage, receiving inflow and losing water to evaporation and outflow. Modelling the network requires both the land phase and channel phase hydrology to be represented fully. Land phase hydrology, through the water balance of areas that drain into each lake, provides input to the lake. Channel phase hydrology centres on the fill and spill concept, with the lakes being the focal points for the receipt of inputs from their catchment areas and from inflow, if any. Outflow occurs only when lake storage exceeds some threshold value (e.g. elevation of the lake outlet) specific to each element.

Dividing a lake-stream flow system up into a collection of lake – basin elements, allows the issue of dynamic flow connectivity throughout the system to be better represented using the fill and spill principle. Figure 5.6 is a flow chart depicting the computational procedures, starting from the uppermost lake basin element and continuing until the outlet of the lake-stream network is reached. The model requires the topologic linkages of various lake elements to be specified, and the storage threshold for each lake (equivalent to its maximum storage capacity) to be assigned. Any land surface scheme or water balance computational method can be used to calculate runoff from the basin slopes to the lake. An accounting of lake storage change is needed for every time step to check that if $S<S_T$, outflow is zero. Otherwise, the generated outflow is sent down the channel using any routing algorithm desired. The next lake basin elements will receive runoff from their own catchment areas plus any channel flow from the lake basin elements upstream. In this way, each lake and its direct catchment may be considered to operate as an independent unit in the chain, subject to variable channel flow linkages with the entities above or below.

The topologic linkages between lakes can be determined through a naming classification system presented in Figure 5.7. Each lake within the system is classified based on its order within the lake chain to which it belongs. Lake number increases sequentially down a particular chain. When two chains meet, the new chain downstream takes on the number of the higher order. Lakes within each order are then numbered and a unique identification is assigned to each lake consisting of its order and number (order.number). The topologic linkages of the lake – stream flow system reported in this study can be represented by the sub system shown in Figure 5.7 contained within the dashed box.

The procedures are adaptable for incorporation into existing modular, water balance based models. This proposed model can be used not only for lake – stream flow networks but for all systems subject to the fill and spill process.

Table 5.1: Lake water balance (in mm) of four lakes, for the period May 5 to August 31, 2004.

	Melvin Lake	Shadow Lake	Hazel Lake	Lois lake
Snowmelt	125	107	118	83
Rainfall	44	44	42	45
Evaporation	362	320	359	337
Stream inflow to lake	0	91	980	3523
Stream outflow from lake	880	283	933	2409
Net exchange with basin	829	190	-641	-664
Net storage change	-11.86	-11.80	-26.39	-2.51

Table 5.2: Starting and ending dates, and volume of lake outflow.

	Melvin Lake	Shadow Lake	Hazel Lake	Lois lake	Rater Lake
Direct catchment area[1] (km^2)	0.116	0.565	0.168	0.760	3.150
Basin area[2] (km^2)	0.116	0.681	0.849	4.759	3.150
Lake area (km^2)	0.009	0.087	0.007	0.108	0.214
Lake/basin ratio (%)	7.8	12.8	0.8	2.3	6.8
Snowmelt period outflow					
Start date	May 24	May 22	May 20	May 24	May 30
End date	June 25	June 25	June 15	July 5	-
Total flow volume (m^3)	8,265	24,490	23,329	260,693	
Summer season outflow					
Start date				July 26	-
End date				July 31	Aug. 24
Total flow volume (m^3)				4	173,328

[1] direct catchment area refers to the basin area that feeds directly to the lake and exclusive of the area contributing to streamflow above the lake inlet.
[2] total basin area that drains into a lake, including the direct catchment and the areas upstream.
[3] outflow for Rater Lake was continuous between May 30 and August 25; number refers to total flow during this period.

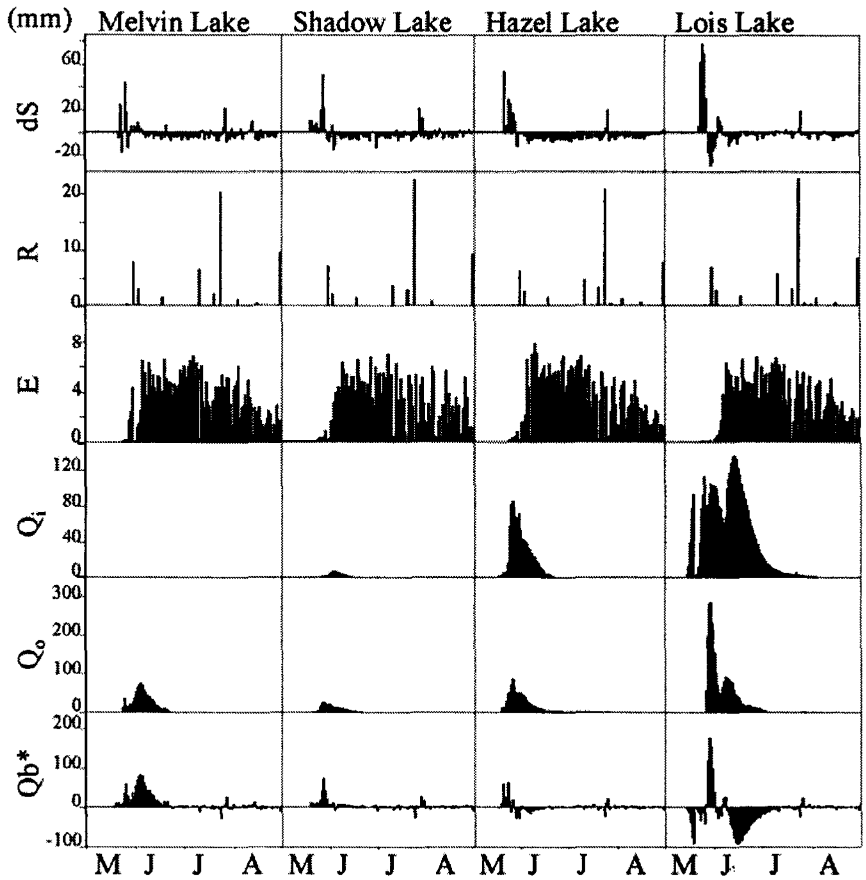

Figure 5.1: Daily variation of water balance components of four study lakes during snowmelt to late summer period, 2004.

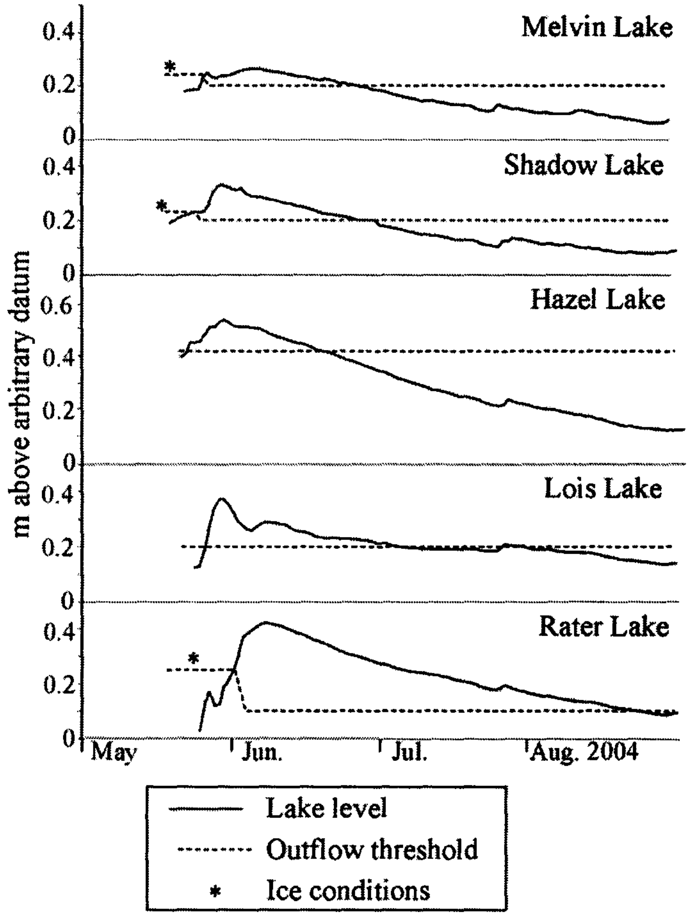

Figure 5.2: Fluctuations of lake storage above outflow thresholds for five lakes.

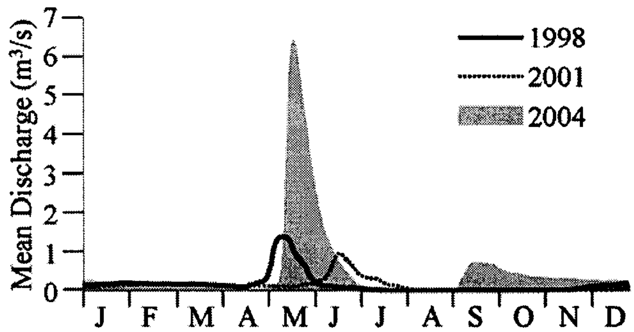

Figure 5.3: Hydrographs of northern Shield rivers showing occurrences of zero flow conditions during open channel conditions.

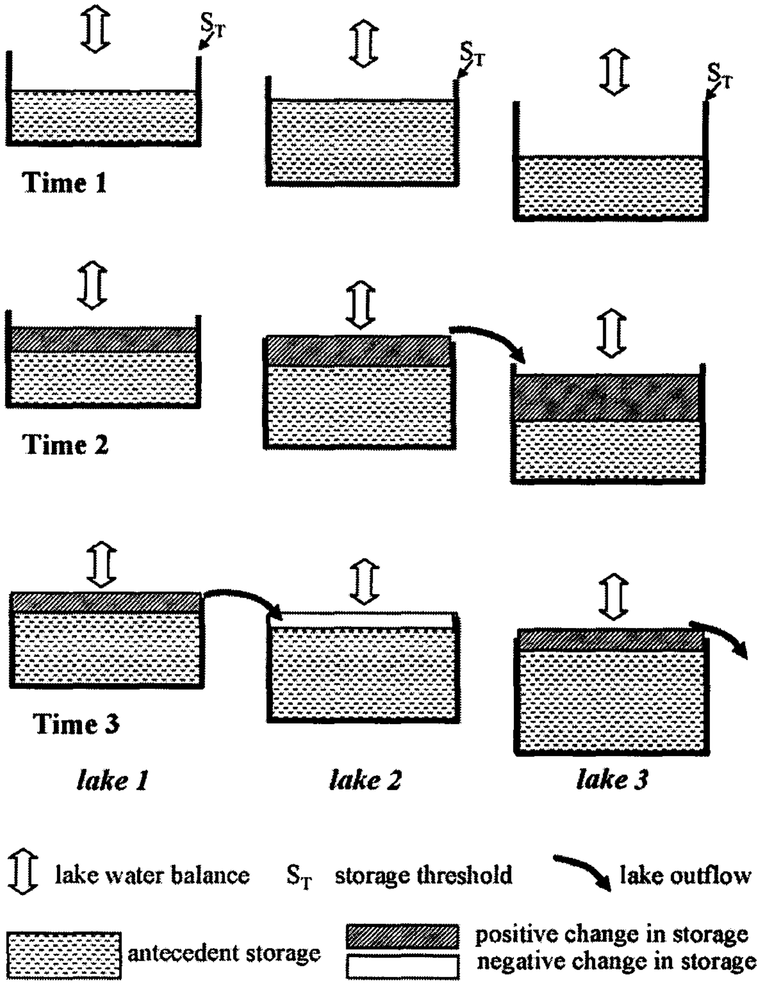

Figure 5.4: Conceptualization of fill and spill framework applicable to modeling of flow connectivity in a lake-stream system in a semi-arid Shield environment. The principle that each lake can operate as an independent entity is illustrated as follows. Initially, antecedent storage is below the outflow threshold (ST) which differs among lakes. Water balance causes change in storage for each lake. In period 2, storage increase allows the level in Lake 2 to exceed it threshold to produce outflow. In period 3, negative storage change for Lake 2 leads to water level drawdown and cessation of outflow, but other lakes experience positive storage change that enable outflow generation. Consequently, flow connectivity can be discontinuous and variable in time.

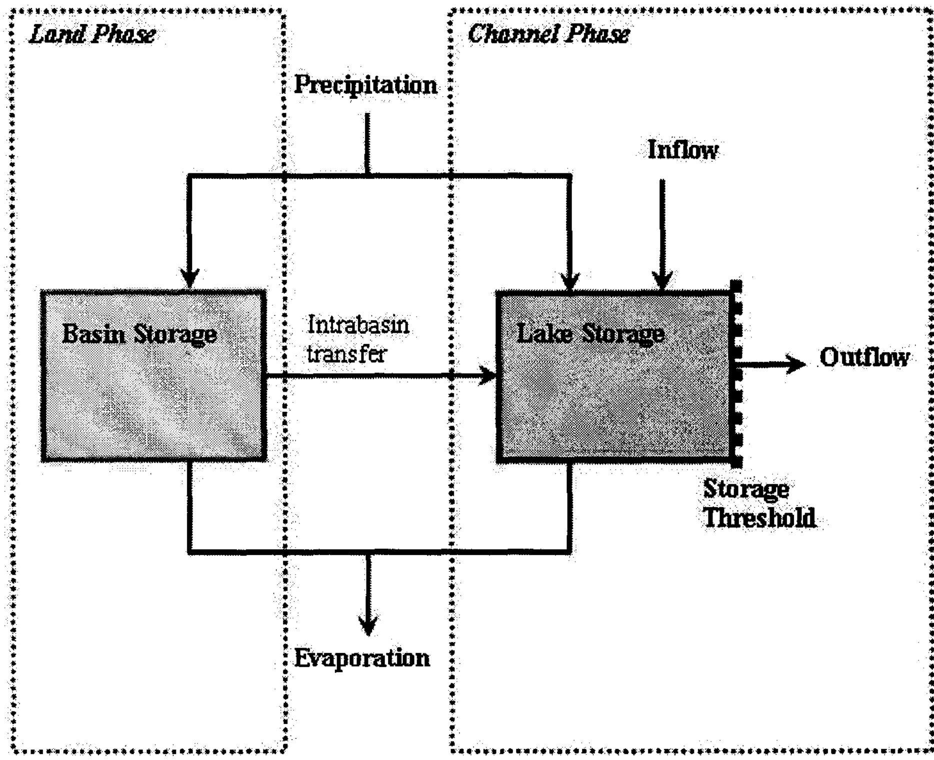

Figure 5.5: Conceptualization of a lake basin as a hydrologic element containing both land and channel phases.

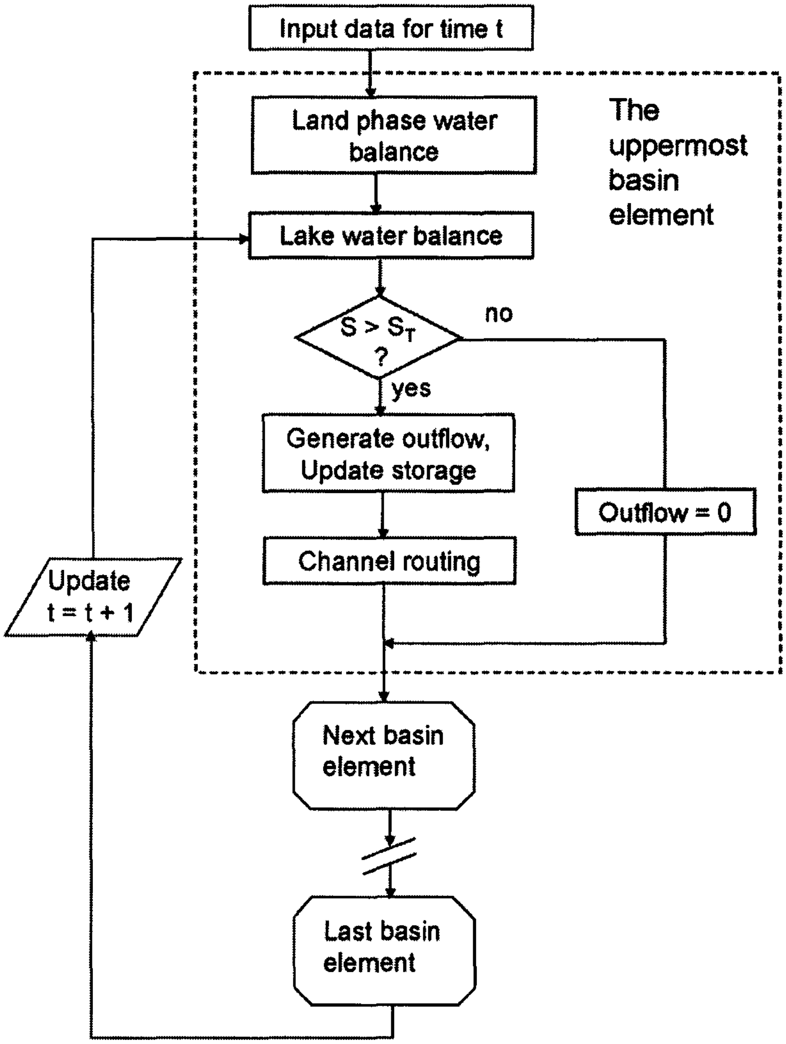

Figure 5.6: Flow-chart showing the simulation of land-phase and channel-phase linkages in a lake-stream network.

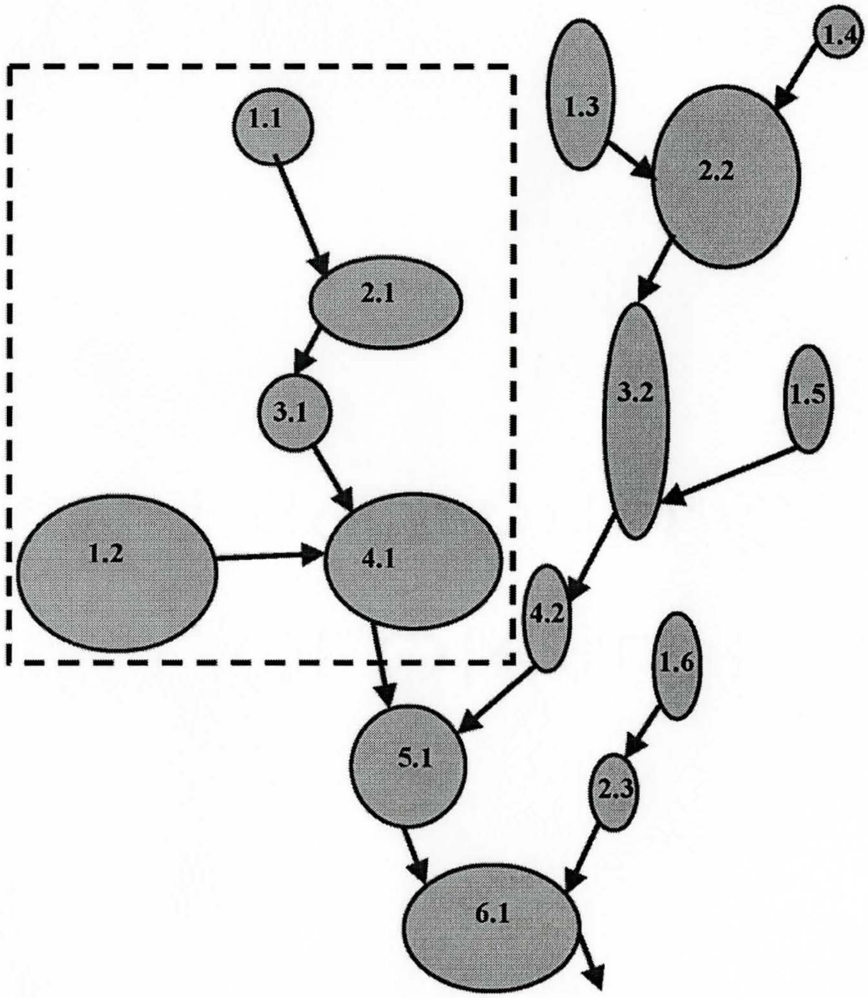

Figure 5.7: Topological lake basin classification for use in applying model calculations. The topology of the study catchment is represented within the dashed box. The basin drained by Rater Lake (which includes several upstream lakes) has been simplified into one element.

CHAPTER 6: CONCLUSION

This is the first comprehensive study of a lake-stream system in the semi-arid Canadian Shield environment. It examined the hydrologic behaviour and provided an understanding of the processes influencing the linkages between lakes and their surrounding catchment areas. The fill and spill principle offers explanation of the flow connectivity of lakes in a lake-stream chain. Major findings of the study are summarized below.

Results from the headwater lake-stream system demonstrate that outflow production is governed by (1) antecedent storage in the lake which is a product of cumulative water balance in the past periods, (2) lake water balance of the current period that includes vertical fluxes due to snowmelt, rainfall and evaporation, lateral exchanges with the slopes that drain directly to the lake, and inflow and outflow, (3) lake storage capacity which represents the threshold for lake outflow. Outflow occurs only when the threshold is exceeded, hence channel flow is variable in time and flow linkage need not be continuous along a chain of lakes. This confirms the significance of the fill and spill concept in outflow generation from small lakes. The principle can be extended to drainage systems beyond the headwater zone, as is supported by the discharge record of a larger river in the subarctic Shield region.

This study gives an account of the major processes responsible for flow generation and cessation in a typical headwater lake basin of the subarctic Canadian Shield. The processes governing flow linkages in a lake stream network are presented in the context of lake fill and spill. A simple process based model incorporating land phase and channel phase hydrology is presented to model flow release and stoppage along lake stream flow networks. The fill and spill concept is utilized with the lakes as the focal points for the receipt of inputs from their catchment areas and from inflow, if any. When lake storage exceeds some threshold value (e.g. elevation of the lake outlet) specific to each element, lake outflow will occur.

The processes described in this study are applicable to arid and semi-arid areas subject to alternating flooding and drying, where lake storage controls the retention or release of flow. The lake fill and spill principle is also relevant to the humid region, but with ample water inputs to the lakes, the severance of channel flow linkage between lakes in the humid Shield areas is uncommon. The concept can be incorporated into existing hydrologic models to improve the modeling of Shield hydrology. A process-based model of flow release and stoppage for lakes will permit improved prediction of flow reliability which can be an important consideration in water resource management in regions with many small lakes.

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