RUNOFF HYDROLOGY OF A LOW ARCTIC DRAINAGE BASIN

by

NIGEL T. ROULET

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AUTHOR: Nigel T. Roulet (B.Sc., Trent University)
(M.Sc., Trent University)

SUPERVISOR: Dr. M.K. Woo

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ABSTRACT

This is a study of the hydrology of a low Arctic drainage basin in the continuous permafrost region. The basin has a varied surface cover of upland tundra, wetlands and lakes. A water balance computed at the basin scale determines the relative importance of various hydrological components including snowmelt, rainfall, inflow, evaporation and outflow. Studies conducted at sites representative of uplands, wetlands and lakes quantify the magnitude and temporal variability of runoff. By combining the process and water balance studies results, an explanatory framework is developed for the spring and summer hydrology.

Snowmelt is the largest input of water to the basin and runoff is one third greater than evaporation as a loss. Peak discharge occurs during snowmelt when meltwater exceeds the storage capacity available in the basin. Snowmelt contribution is greatest from the wetland areas, but, because of a larger areal extent, upland meltwater contribution is also significant. The wetland areas provide a route through which lake and channel overbank flow and meltwater are conveyed to the basin outflow. Surface flow is the primary runoff process during snowmelt in both wetland and upland areas.

In summer, evaporation is the major process through which water is lost from the basin. Upland areas become dry, but wetlands have enough stored water to sustain evaporation

at potential rates. Runoff response to a rainstorm is negligible when the basin is dry, but is significant if previous storms have saturated the upland and the wetland areas. Storm runoff is produced first from hummocky areas at the base of hillslopes, then from uplands and finally from the wetlands. When all parts of the basin contribute storm runoff the basin response will be both large and rapid. Stormflow recession is long for all storms.

The upland-wetland-lake hydrological system examined in this study is typical of the North American low Arctic environment. Through an understanding of both the hydrological processes within each subsystem and the linkages between subsystems an explanation of the runoff regime of a low Arctic drainage basin is obtained.

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NOTATION

A	proportional area	dimensionless
α	evaporation coefficient	dimensionless
b_0	regression intercept	
b_1	regression slope	
β	Bowen ratio	dimensionless
C	volumetric heat capacity	$J\ m^{-3}\ ^oK^{-1}$
c	specific heat capacity	$J\ kg^{-1}\ ^oK^{-1}$
C_e	coefficient of discharge	dimemsionless
γ	psychrometric constant	$mb\ ^oC^{-1}$
D_h	drag coefficient	$m\ s^{-1}$
d_s	saturated zone thickness	m
e	evaporation rate	$mm\ t^{-1}$
g	acceleration due to gravity	$m\ s^{-2}$
h	hydraulic head	m
h_e	effective head	m
h_w	head above V notch weir	m
θ	volumetric soil moisture	dimensionless
θ_v	V notch weir interior angle	degrees
n	porosity	dimensionless
I	volumetric ice content	dimensionless
i	ice contribution	$mm\ t^{-1}$
K	hydraulic conductivity	$m\ d^{-1}$
$K\downarrow$	solar radiation flux density	$J\ m^{-2}\ t^{-1}$
k	the von Karman constant	dimensionless

L_f	latent heat of fusion	$J\ kg^{-1}$
L_v	latent heat of vapourization	$J\ kg^{-1}$
m	meltwater	$mm\ t^{-1}$
m_c	recession constant	t
Q	energy flux density	$J\ m^{-2}\ t^{-1}$
Q^*	net radiation flux density	$J\ m^{-2}\ t^{-1}$
q	discharge	$m^3\ s^{-1}$
q_s	specific humidity	$kg\ kg^{-1}$
Ri	the Richardson number	dimensionless
r	rainfall	$mm\ t^{-1}$
ρ	density	$kg\ m^{-3}$
s	water storage	mm
σ	slope of the saturation vapour pressure curve	$mb\ ^oC^{-1}$
T	temperature	^oC
T^*	mean air temperature	^oC
t	time	d, hrs, min, s
u	wind velocity	$m\ s^{-1}$
w	width	m
x	linear distance	m
x	fractional component	dimensionless
z	depth	m
z_m	mean depth	m
z_0	aerodynamic surface roughness	m

Subscripts

a	air

c	main meteorological site
G	ground
H	sensible heat
i	ice
L	lake
E	latent heat
m	mineral soil
o	organic soil
s	surface
ss	subsurface
w	water

CHAPTER ONE INTRODUCTION

Over 10 percent of the land area of Canada is within the continental Arctic, yet little is known of the region's physical hydrology. Intensive resource exploration in this region has inspired geophysical and engineering research, but this research has been largely confined to the Beaufort Sea coast which is unrepresentative of most of the area north of the tree line. More recently, mineral exploration in the eastern portion of the mainland Northwest Territories, in the District of Keewatin, has raised environmental concerns. Interaction between mining and the regional water resources could produce serious problems, the impact of which cannot be assessed given the present state of knowledge of the region.

Central Keewatin is within the zone of continuous permafrost and it has one of the highest proportions of land covered by surface water in Canada (National Atlas of Canada, 1982). Regional topography is gently rolling with large flat valleys, many of which contain wetlands, defined as areas where wet soils occur and have a water table near or above the ground surface for most of the thawed season (Zoltai et al., 1973), because of poor drainage. It has been estimated that five percent of the land area of this region is wetland (Zoltai, 1979). Most of the region has continuous or patchy upland tundra vegetation developed on shallow layers of poorly developed soil (Zoltai and Johnson, 1978). The rest of the region consists of limited areas of rock outcrop.

This physical setting in a continental location makes this region different from other areas in the Arctic and Subarctic. Hydrological understanding gained in the high Arctic Islands (Ryden, 1977; Woo, 1983) or the Subarctic (Dingman, 1975) may not be transferrable to this low Arctic environment. It is therefore the purpose of this research to examine the hydrology of a drainage basin in central Keewatin to gain an understanding of low Arctic runoff processes.

1.1.0 Permafrost Hydrology

Permafrost is defined as ground that freezes one winter and remains frozen the following summer and into the next winter (Brown and Pewe, 1973). In the continuous permafrost region all water movement of significance to runoff occurs either as suprapermafrost ground water flow in the active layer or as surface flow and for all practical purposes, the permafrost table can be viewed as an impermeable lower boundary (Tolstikhin and Tolstikhin, 1974).

There are two notable features related to the runoff regime in the permafrost region. In this cold environment, a large portion of the annual precipitation falls as snow, ranging from 40 percent near the tree line (eg. Ennadai Lake: 61°N 110°W) to over 90 percent in parts of the high Arctic (eg. Alert: 83°N 67°W) (AES, 1982). Snow is stored for up to 10 months of each year but melts rapidly in spring to generate as much as 80 percent of the total annual runoff in a period of several weeks (Woo, 1982). The second characteristic of permafrost hydrology is the changing water storage capacity produced by the annual freeze-thaw cycle of the ground. This storage capacity is the least in winter and immediately before ground thaw begins, and greatest when the frost table has reached its maximum depth.

The effects of permafrost on storm runoff have been studied in the discontinuous permafrost region where both permafrost dominated and non-permafrost basins occur side by side. Runoff from permafrost dominated basins shows a higher peak and lower base flow, a greater flow volume (Haugen et al., 1982) and a 'flashier' response (Slaughter and Kane, 1979). Within individual basins, areas that are underlain by permafrost generate the bulk of the surface runoff (Dingman, 1975) while the non-permafrost areas provided water to the regional groundwater system (Kane and Stein, 1983).

Superimposed on the influence of snowmelt and permafrost is that of such factors as topography, soils, vegetation and open water. These locally affect the flow regimes, storage capacities and evaporative losses and hence the basin water balance. For example, Arctic basins with poorly developed soils and little vegetation have an annual runoff ratio of 0.7 (Woo, 1983), but basins containing wetlands have a runoff ratio of 0.5 (Brown et al., 1968) because of greater evaporation loss (Ryden, 1977).

In an attempt to describe the seasonal runoff patterns that are evident in permafrost regions, Church (1974) devised a classification system comprising four regimes; Arctic nival, Subarctic nival, Proglacial and Muskeg. There are no glaciers in the continental western Arctic so the proglacial regime is not relevant to this study. Characteristic of the remaining three regimes are (Church, 1974):

1) Arctic Nival regime: dominated by snowmelt floods in spring and low flows in the summer, punctuated by occasional rain induced high flows. There is no winter discharge.

2) Subarctic Nival regime: same characteristics of the Arctic Nival, but rain is more important as a water input and there may be winter discharge sustained by ground water flow.

3) Muskeg regime: characterized by poor drainage. Flood level discharges are attenuated because of the water holding capacity of organic soils and irregular surface topography. The influence of muskeg on runoff response is similar to that of a 'relatively large lake'. However, Ryden (1977) observed large spring floods and Brown et al. (1968)

observed large summer storm responses from Arctic wetland basins.

The difference in runoff regimes indicates that the magnitude of various runoff processes must vary spatially with differences in basin cover types. Ryden (1977) worked in an Arctic basin that had a varied cover but did not identify the interaction between different elements within the basin. Other researchers have discussed runoff generation in the discontinuous permafrost region (Dingman, 1975; Wright, 1981) but did not quantify the fluxes of water by various runoff processes. To improve an understanding of basin runoff generation in the permafrost environment, the dynamic relationship among various hydrological subsystems within a basin should be studied.

1.2.0 Research Objectives

The central Keewatin District possesses a wide assemblage of different land cover types that have different hydrological characteristics. The runoff regime and the water balance of a low Arctic drainage basin reflect both timing and the manner that these subsystems are hydrologically combined. It is the purpose of this thesis to study the hydrological processes of a small low Arctic basin to:

1) determine the major components of its hydrological subsystems,

2) establish the linkages within and between these subsystems,

and 3) develop an explanatory framework for the observed runoff in this area of continuous permafrost:

1.3.0 Research Approach

This research examined the hydrological processes at several temporal and spatial scales. Snow storage, snowmelt and rainfall were examined at the basin scale. Hydrological processes including suprapermafrost ground water movement and surface water flow were studied at several runoff plots. At specific points in the basin, evaporation, soil water storage capacities and hydrological parameters were monitored. Various observations were made at short time intervals, including snowmelt, evaporation, surface flow and basin outflow, while other phenomena, such as changes in soil moisture and ground water flow, were monitored over longer time intervals. For logistical reasons, this research focuses on the period between late winter and late summer.

This thesis is divided into seven chapters. The next chapter describes the physiographic and climatic conditions of the research area and the methods of study. Chapters Three through Five contain the research results, focusing on basin input and output (Chapter Three), upland area hydrology (Chapter Four) and wetland area hydrology (Chapter Five). In Chapter Six a basin water balance is computed and runoff processes are discussed. The concluding chapter summarizes the major findings.

CHAPTER TWO STUDY SITE AND METHODS

Field studies were undertaken from May to early August, during 1982 and 1983 in a small drainage basin in the central Keewatin District of the Northwest Territories (64°27'N, 97°47'W). Little is known of the region's bio-physical environment. Cummingham and Shilts (1977) have mapped the surficial geology and Zoltai and Johnson (1978) and Edlund (1977) catalogued plant species and soil characteristics of the region. More than thirty years of climatological records are available for the town of Baker Lake, N.W.T., 80 km to the east. The closest Water Survey of Canada stream gauging site to the study basin, maintained until November 1984, was Kingyouk Creek, which flows into Baker Lake. In this chapter the physiography, vegetation and climate of the central Keewatin region will be reviewed and the research site described. The methods used to obtain the required hydrological and meteorological data will also be discussed.

2.1.0 Regional Physiography, Vegetation and Climate

Central Keewatin lies within the low Arctic region (Zoltai, 1979). The rolling topography of the area is underlain by bedrock of the Precambrian

Shield. Local relief is no greater than 50 m. The orientation of the surface topography is controlled by the bedrock structure as well as the erosional and depositional pattern of the southeast or northwest glacial flow (Cummingham and Shilts, 1977).

Soils in the region are dominated by a red-clay rich till derived from Dubawnt sandstones (Cummingham and Shilts, 1977). Soil development is influenced by cryoturbation. On drier sites, a weak B horizon has developed, but active periglacial processes have destroyed some of this (Zoltai and Johnson, 1978). Mud boils ranging in size from less than 1.0 m to greater than 3.0 m are prevalent in better-drained till (Shilts, 1978). In wetter areas organic cryosols, mostly peats, are present (Ferris, 1980). Peat is moderately decomposed and fibric to mesic in structure with rubbed fibre contents of 10 to 40 percent depending on location and depth of sample.

The study site is well within the zone of continuous permafrost. Brown (1978) estimated the depth of permafrost to be greater than 500 m. The active layer ranges from 0.70 to 1.50 m in well drained tills to 0.30 m in well drained peatlands and 0.80 m in wet peatlands (Zoltai and Johnson, 1978).

Vegetational composition is closely related to soil and moisture characteristics. Zoltai and Johnson (1977) recognize four floristic groupings. On poorly-drained soils with shallow freestanding water or a water table near the ground surface, *Carex* spp. and *Eriophorum* spp. meadows dominate. With better drainage, tussock meadows of *Eriophorum* spp., *Sphagnum* spp. and *Drepanocladus* spp. develop. Well drained soils support a lichen heath of *Alectoria* spp., *Cetraria* spp., and *Cornicularia* spp. and dwarf shrubs *Empetrum nigrum*, *Ledum palustre* and *Vaccinium vitis-idaea*. Also *Betula glandulosa* heath will develop in dry, sheltered areas. Rock outcrops and bouldery till support communities of crustose lichens of *Rhizocarpon* spp., *Lecidea* spp. and *Lecanora* spp.

The climate is cold continental dominated by Arctic pressure systems and is described as a low energy climate by Hare and Thomas (1979). For 8 months of the year the mean daily temperature remains below 0°C (Figure 2.1a). The mean annual temperature is -12°C, with a January minimum of -33°C and a July maximum of 11°C. The annual precipitation total is 235 mm of which 42 percent falls as snow. Rainfall occurs in the summer and early fall while most of the snowfall comes in October. During the study period the mean monthly summer temperature and the 1982 precipitation at Baker Lake were similar to the 30 year averages, but the distribution of summer precipitation in 1983 was anomalous (Figure 2.1b). Precipitation for June and July 1983 was only 51 and 18 percent respectively of the long term average while for August it was two times greater.

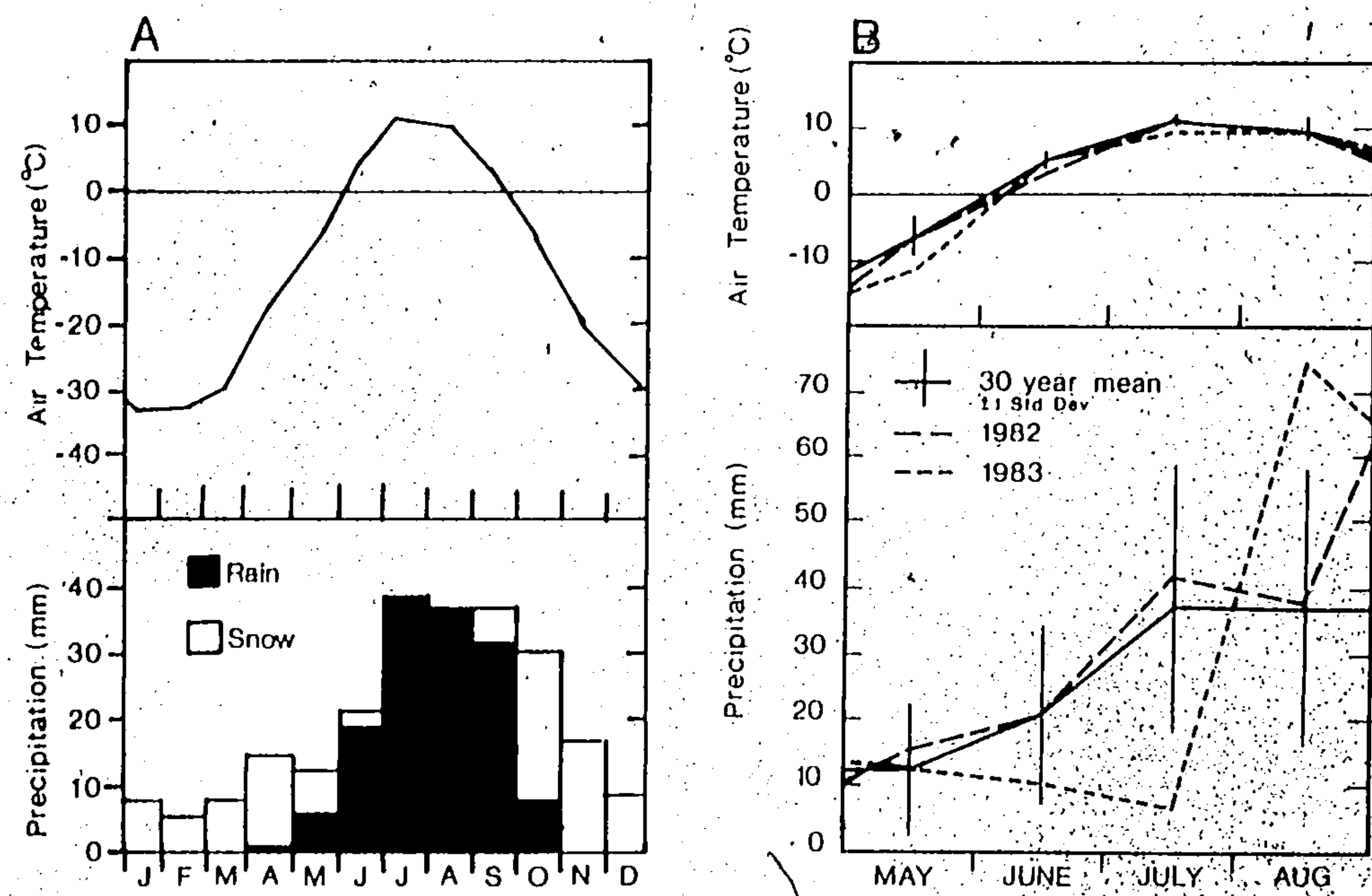

Figure 2.1 A. 1951-1981 mean monthly air temperature and precipitation at Baker Lake, N.W.T.
B. Mean monthly air temperature and precipitation at Baker Lake, N.W.T. during the 1982 and 1983 field seasons.

2.2.0 Study Area and Research Sites

The study basin is situated 80 km west of the town of Baker Lake. The military grid reference for the basin outflow is 465629, Judge Sissons Lake 66A/5 map sheet from Survey and Mapping Branch, Energy Mines and Resources, Ottawa. The study basin area is 1.36 km^2 and the basin outflow elevation is approximately 180 m a.s.l. Maximum difference in elevation within the basin is 25 m.

The basin consists of two plateaus divided by a small scarp running east-west through the basin (Figure 2.2). The basin contains four lakes[1]; Lost Lake, Hardill Lake, Heart Lake and one unnamed lake. Moose Pond is located east of Heart Lake. The outflow of Hardill Lake, draining a 10.6 km^2 area, forms the only channel input and the basin outflow is at the outlet of Heart Lake. A stream passing through the centre of the basin joins these lakes. There are two intermittent streams. One flows into Heart Lake and the other into the main stream below the ridge (Figure 2.2). Open water forms 6 percent of the basin area. Flanking the basin are well drained upland hillslopes. Flatter portions of the basin comprise well drained peatlands while the lowest parts are occupied by flat wetlands. Uplands, well drained peatlands and wetlands comprise 55 ,18 and 21 percent of the basin area respectively.

Three study sites, an upland hillslope, a fen wetland

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[1] The names used in this thesis are not offical.

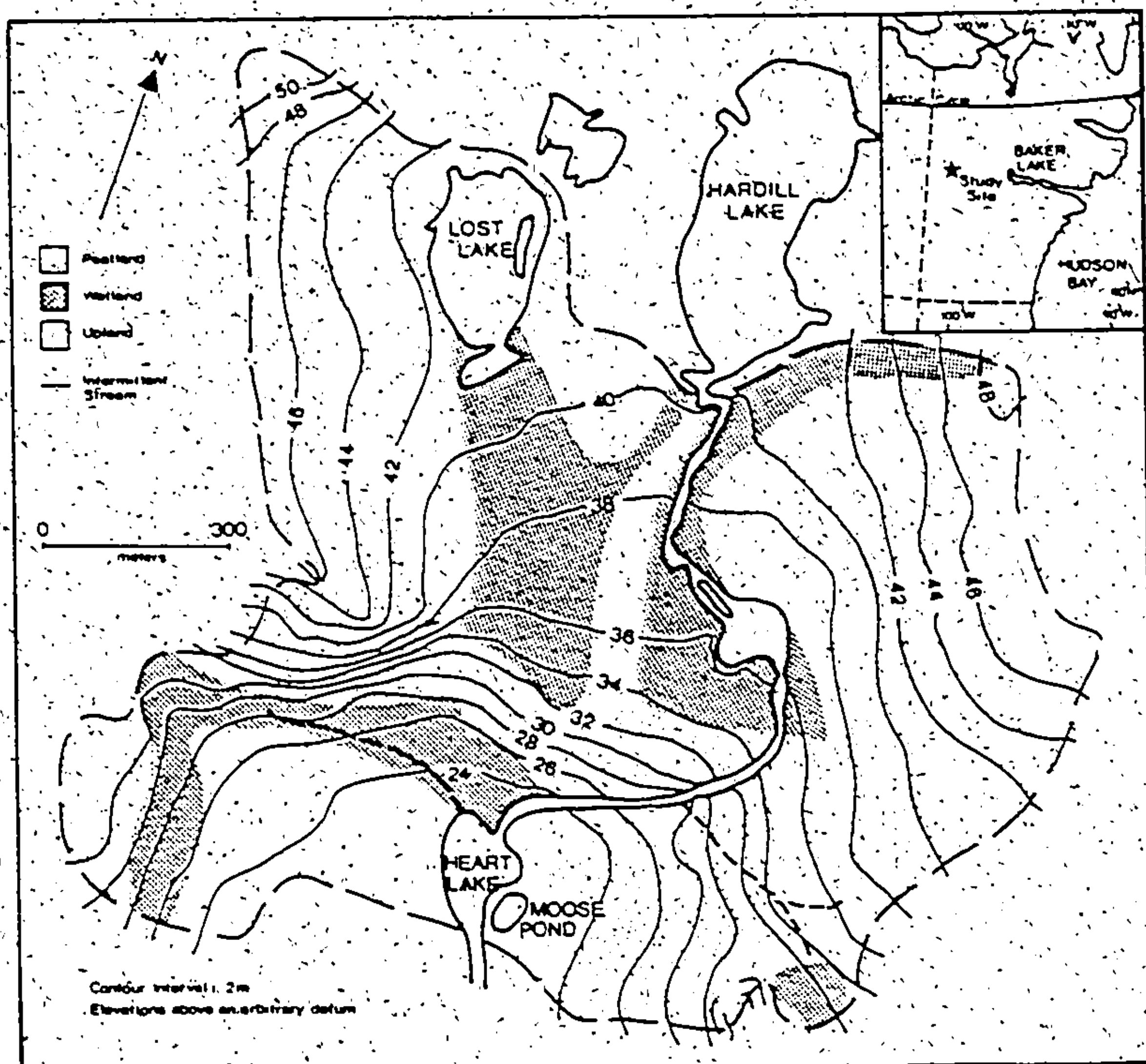


Figure 2.2 Study basin surface topography and cover type. Insert shows the location of the study area in the Keewatin District, N.W.T.

and a small pond were used during the two field seasons (Figure 2.3). The wetland site was situated downslope from Lost Lake. The site area was 0.15 km², with a surface gradient of 2 percent and was covered with 0.25 m of peat. The upland hillslope site was located in the south east corner of the basin, stretching 300 m from a rock outcrop to Moose Pond. Average surface gradient was 5 percent, ranging from 2 percent at the base to 8 percent upslope. The base of the slope was occupied by earth hummocks (Tarnocia and Zoltai, 1978) while the mid and upper portions were approximately 60 percent vegetated, with the remainder comprising mud boils (Shilts, 1978), or bare soil. Soil and vegetation characteristics of the upland and wetland sites are discussed in Chapters Four and Five. Moose Pond provided an open water site and had a surface area of 2,000 m² and a maximum depth of 1.1 m. There was no surface inflow or outflow to Moose Pond except during the spring melt.

2.3 Methods

This study involved field measurements of hydrological and meterological variables and parameters which can be used to characterize the low Arctic hydrological system. Figure 2.3 shows the locations referred to in the following section. In addition, a weather station was maintained during both summers at a base camp 0.9 km southeast of the

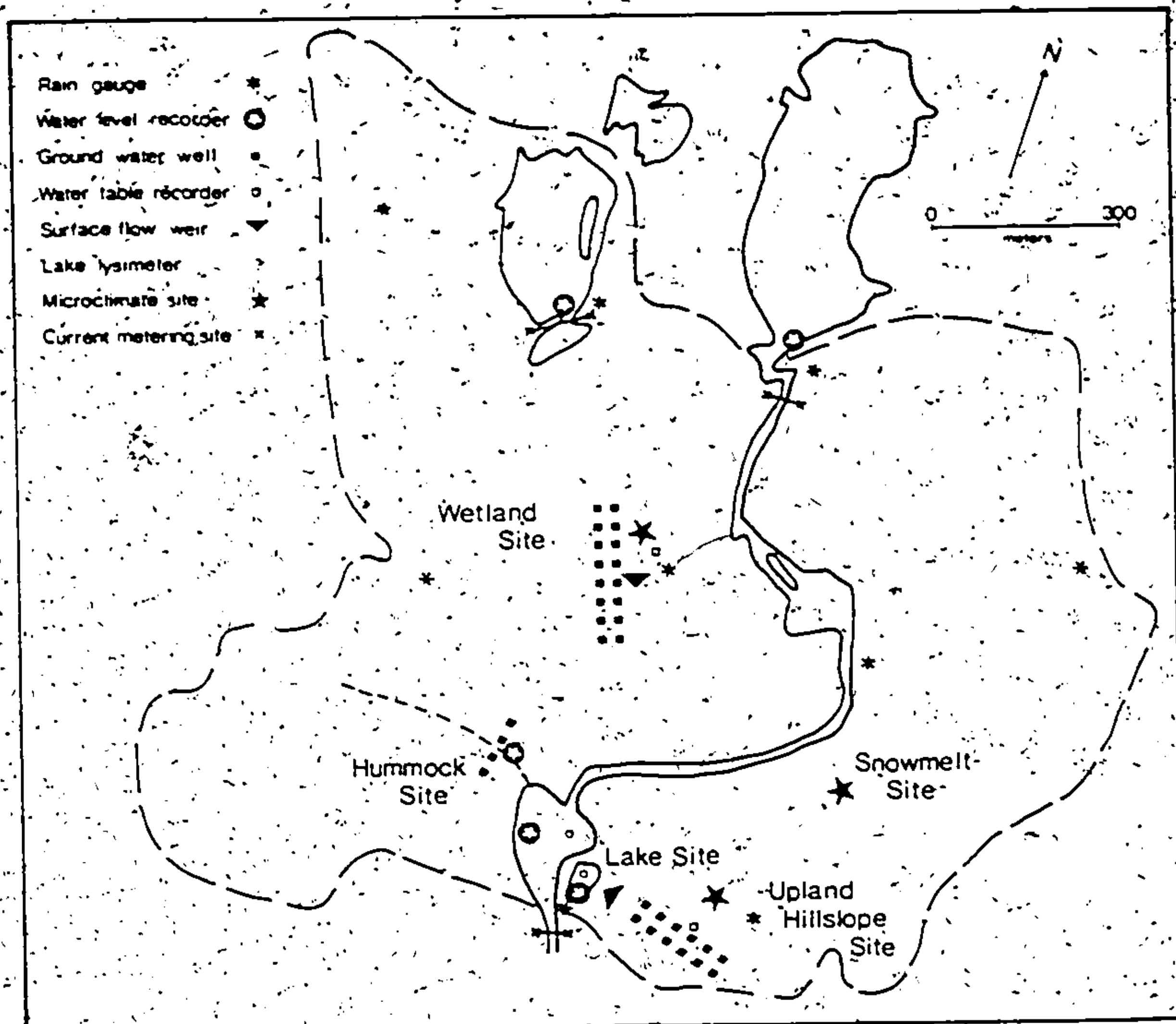


Figure 2.3 Location of study sites and instrumentation in the basin.

basin.

2.3.1 Soil Parameters

Soil bulk density was obtained by taking a known volume of soil and after drying in an oven for 24 hours at approximately 50 °C, weighing it on a triple-beam balance. Specific retention and yield were measured using the gravity drainage method (Johnson, 1962). Hydraulic conductivity was obtained by single well pumping tests (Luthin and Kirkham, 1949). This technique was developed for mineral soils but has been found applicable to peat soil with small hydraulic gradients (Dai and Sparling, 1973; Ingram, 1984).

2.3.2 Meteorological Measurements

Precipitation and Snowmelt

To estimate the snow input to the basin, systematic snow depth surveys using 156 samples in 1982 and 449 samples in 1983 were undertaken late in the winter. Snow density was obtained for six different terrain types (Table 2.1) using the Meteorological Service of Canada snow sampler. Water equivalent was then computed by a method outlined by Woo and Marsh (1978). During the melt period the daily distribution of snow water equivalent was computed for grid cells (625 m^2) superimposed on the study area. The water equivalent of each cell was updated daily assuming an equal rate of snow

TABLE 2.1 Terrain Types for Snow Density Survey

Terrain Types	Percentage of Basin Area
Lakes	5.1
Ridge tops	2.8
Valley bottoms	3.8
Scarp slopes	8.1
Slopes	38.6
Flat wetlands	41.7

loss for the entire basin (Woo et al., 1981). Predicted and actual snowcover distribution during the 1982 and 1983 melt periods are compared in Figure 2.4. There is an approximate 20 percent systematic overestimate of snowfree area by this method.

Snowmelt was calculated using the energy balance approach (Heron and Woo, 1978). Melt, m, in a given time period is,

$$m = (Q^* + Q_H + Q_E + Q_G) / L_f \rho_w \qquad (2.1)$$

where L_f is the latent heat of fusion and ρ_w is the density of water. Net radiation, Q^*, was measured using a Swissteco net radiometer and Q_G is ground heat flux. The sensible, Q_H, and latent heat flux, Q_E, were derived using bulk transfer equations (Price and Dunne, 1976). Temperature and relative humidity were measured using a Campbell 201 temperature and relative humidity probe (accuracy ±0.4°C and ±2% RH) and wind speed using a Cassella-Shepherd anemometer. All measurements were made at 1.0 m above the snow surface. Readings were taken once every minute and averaged hourly by a Campbell CR21 datalogger. Snow roughness was taken as 0.003 m, a value verified by occasional field measurements using a Thornthwaite wind profile system. Melt values determined by equation 2.1 were compared with observed ablation which was measured using a method described by Heron and Woo (1978) (Figure 2.5). At the snowmelt and wetland sites, snowcover temperature was measured daily

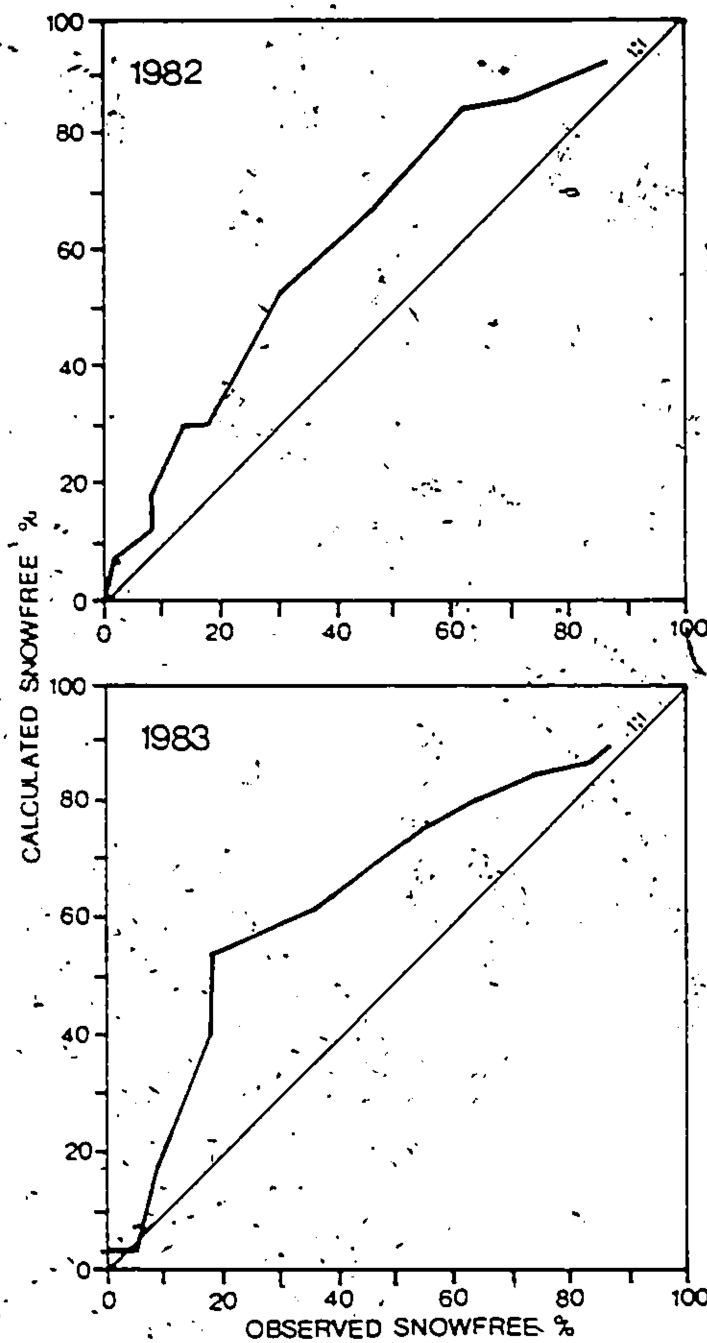


Figure 2.4 Comparison of observed and computed percent snowfree area during the 1982 and 1983 snowmelt periods.

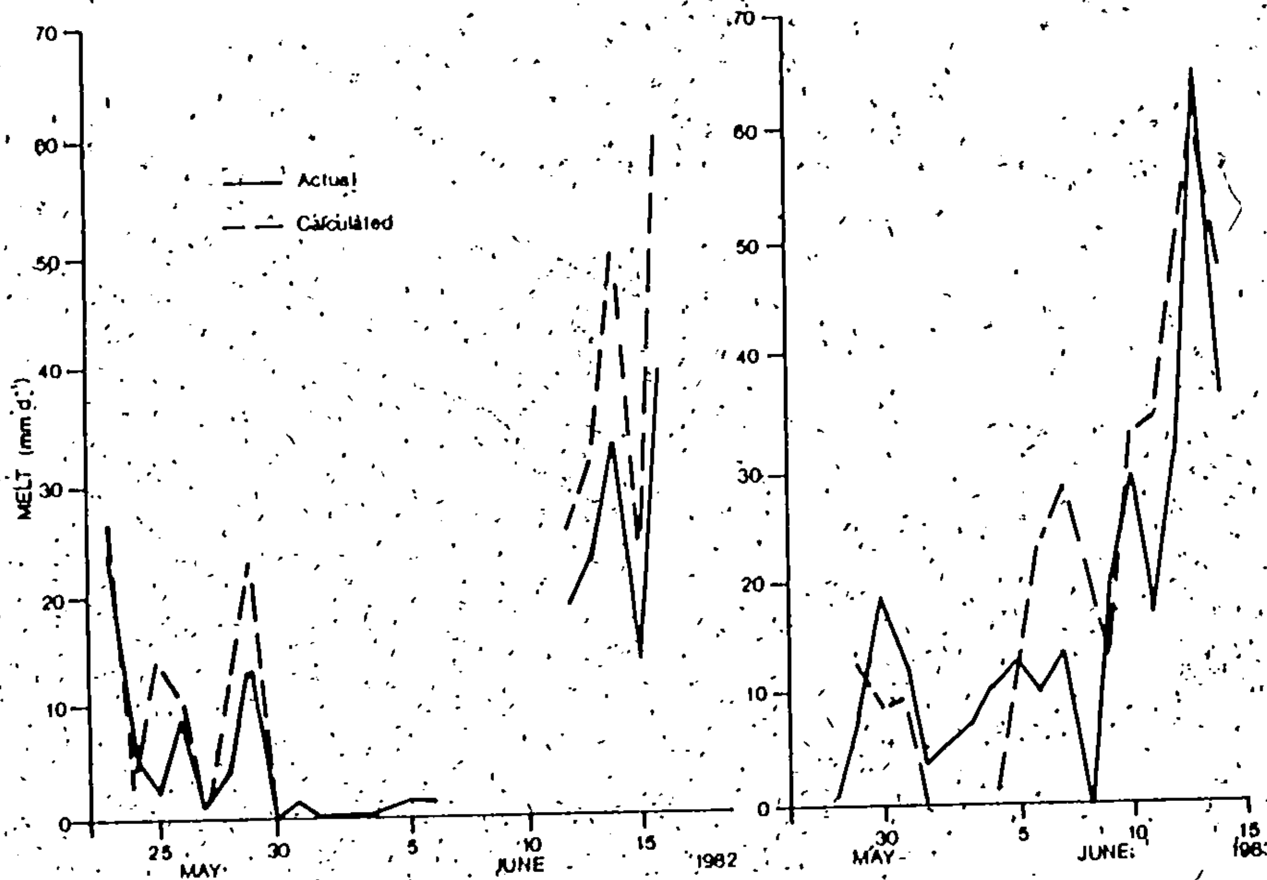


Figure 2.5 Measured and computed daily snowmelt during the 1982 and 1983 melt periods.

from the ground surface to the snow surface at 0.1 m intervals with Fenwall thermistors. Ground and snow thermistors were calibrated in an oil bath accurate to ±0.5°C.

Rainfall was measured at the main meteorological site with a tipping-bucket rain gauge, calibrated to 0.25 mm. This measurement was supplemented by a spatial network of 8 bulk rain gauges (orifice diameter 145 mm, orifice height 250 mm) that were checked manually after all rainfall events exceeding 2.0 mm.

Evaporation

For the computation of evaporation, estimates of ground heat flux are required. Land surface ground heat flux, Q_G, was calculated by (Rouse, 1984),

$$Q_G = \int_0^{z_1} C(z) \frac{\partial T}{\partial t} dz + \frac{\rho_w L_f}{\Delta t} \int_{z_1}^{z_2} I(z)\, dz \qquad (2.2)$$

where C, the volumetric heat capacity of the soil was computed as (deVries 1952, from Rose 1966),

$$C = c_a \rho_a x_a + c_w \rho_w x_w + c_m \rho_m x_m + c_o \rho_o x_o \qquad (2.3)$$

Here c is the specific heat, ρ is density and x is the volumetric fractional component of each soil type. Subscripts a, w, m, and o represent air, water and mineral or organic soil (Table 2.2). Soil temperature, T, was obtained at depths of 0.03, 0.07, 0.12, 0.24, 0.28 and 0.47 m in the wetland and 0.04, 0.13, 0.24, 0.40, 0.60, 0.80 and 1.00 m in the upland. I is volumetric ice content and was assumed equal to soil porosity (i.e. the soil was frozen saturated) and t and z are time and depth, z_1 and z_2 are the depths at which the frost table was located at the beginning and end of each time period Δt. The position of the frost table was measured daily by hammering a rod into the ground at several sites until the frozen zone was reached. Equation 2.2 is only correct if the heat flux across the frost table is zero. In reality this will not be the case, but since the heat flux in frozen soil is small in comparison to the heat consumed during the melting of ground ice, only small errors will be introduced.

The daily flux of sensible heat to the lake was estimated by,

$$Q_L = \rho_w c_w z_m \frac{\Delta T_L}{\Delta t} \qquad (2.4)$$

where $\Delta T_L/\Delta t$ is the average daily temperature change of the lake estimated from measurements of surface water temperature using a Campbell 101 temperature probe. Z_m is the mean depth of the lake. The use of surface temperature to approximate the mean lake temperature may lead to an overestimation of Q_L, but a limited set of Heart Lake temperature profiles from 1982 showed there was little vertical variation in temperature in this shallow lake.

Roulet and Woo (1985a) describe in detail the methods used to obtain actual evaporation from the wetland and lake. Briefly, lake evaporation was measured using three floating lysimeters on Moose Pond. Each lysimeter had a

TABLE 2.2 Soil Specific Heat Capacities

Soil Component	Specific Heat $Jkg^{-1}\,{}^{o}K^{-1}$	Reference
Water	4.18×10^{6}	Oke (1978)
Air	1.20×10^{3}	Oke (1978)
Mineral	1.96×10^{6}	Oke (1978)
Organic	1.89×10^{6}	Smith (1984)

surface area of 1250 cm². Evaporation was obtained by measuring the daily amount of water needed to return the water level in the lysimeter to a fixed mark. From June 22 to August 1, 1983, 22 days of data were obtained. Several days of data were discarded because of rainfall and the remaining days were rejected because lake water had splashed into the lysimeter. Evaporation, e, from the wetland surface in 1983 was calculated using the energy balance Bowen ratio approach,

$$e = \frac{(Q^* - Q_G)}{(1 + \beta)} / \rho_w L_v \tag{2.5}$$

where L_v is the latent heat of vapourization and the Bowen ratio, β, is

$$\beta = \frac{Q_H}{Q_E} = \frac{c_a \Delta T}{L_v \Delta q} \tag{2.6}$$

Here q is specific humidity. Temperature and specific humidity were measured with a four level, self aspirating, wet and dry bulb thermocouple system. Out of a possible 39 days, 21 days of data were obtained. The major loss of data was due to an insufficient supply of moisture to the wet bulb thermocouples. Measurement of lake and wetland evaporation in 1982 failed because of poor lysimeter design and placement.

Upland evaporation was estimated as a residual of the surface energy balance,

$$e = (Q^* - Q_H - Q_G) / \rho_w L_v \tag{2.7}$$

where the sensible heat flux was computed using a bulk transfer equation.

$$Q_H = \rho_a c_a D_h \Delta T (\Phi_m \Phi_H)^{-1} \tag{2.8}$$

where the drag coefficient, D_h, was computed by

$$D_h = u_z \left[\frac{k}{\ln (z/z_0)} \right]^2 \tag{2.9}$$

where u_z is the wind speed at height z measured with a Met-One anemometer, and k is the von Karman constant. The surface roughness, z_0, ranged from 0.001 m to 0.004 m depending on zero plane displacement (J. Drake, personal communications). A value of 0.003 m was used in this study. The adjustment for stability in equation 2.5, $(\Phi_m \Phi_H)^{-1}$, was computed using the Richardson number, Ri, (Oke, 1978);

$$Ri = \frac{g}{T^*} \frac{(\Delta T/\Delta z)}{(\Delta u/\Delta z)^2} \tag{2.10}$$

where g is acceleration due to gravity and T^* is the average air temperature (°K). The following stability factors were used:

Ri	$(\Phi_m \Phi_H)^{-1}$	Condition
$Ri > 0.01$	$(1 + 5Ri)^2$	Stable
$-0.01 \leq Ri \leq 0.01$	1.0	Neutral
$Ri < -0.01$	$(1 - 16Ri)^{3/4}$	Unstable

Temperatures were measured at 0.25, 0.50, 1.00 and 2.00 m above the ground surface using Campbell 101 probes and averaged every 15 minutes. Temperature probes were calibrated against each other and had an interchangeability of 0.2°C. In 1983, 28 daily evaporation estimates were obtained. Days with wind speeds less than 3.0 m s^{-1} were discarded because proper ventilation of the temperature probes could not be guaranteed. In 1982 lysimeters similar to those described by Marsh et al. (1983) were used to measure upland evaporation when the soil was moist.

To estimate evaporation on days that measured evaporation was missing, the Priestley-Taylor (1972) model was used;

$$e = \frac{\alpha}{\rho_w L_v} \cdot \frac{\sigma}{\sigma + \gamma} (Q^* - Q_G) \qquad (2.11)$$

Net radiation was measured with Swissteco net radiometers over the three surface types. The slope of the saturation vapour pressure curve, σ, was calculated as a function of air temperature (Buck, 1981) measured at 1.0 m above each surface with Campbell 101 probes and γ is the psychrometric constant. The α parameter was derived empirically as the ratio of actual evaporation (equation 2.5 for wetland, 2.7 for uplands and lysimeter measurements for the lake) over equilibrium evaporation (equation 2.11 with α=1.0).

On days when measurements of net radiation or air temperature were missing, they were estimated by regression equations (equation 2.12) developed between base camp solar radiation, $K\downarrow$, and air temperature:

$$Q^* = b_0 + b_1 K\downarrow_c$$
$$T = b_0 + b_1 T_c \qquad (2.12)$$

where $K\downarrow$ is downward solar radiation measured with a Eppley pyranometer. Subscript, c, denotes base camp measurements. Regression coefficients, b_0 and b_1, for these relationships are given in Table 2.3.

2.3.3 Hydrological Measurements

Discharge and Surface Water Levels

Stage-discharge relationships were determined for the outflows of Heart Lake, Hardill Lake and Lost Lake (Figure 2.6). Rating curve coefficients are presented in Table 2.4. When the scatter of data about a rating curve is less than 5 percent this method yields reliable estimates of streamflow (Bruce and Clark, 1966). Lake levels were measured near the lake outlets using Leupold-Stevens type F water level recorders mounted on stilling wells. The velocity-area method was used to measure discharge, with velocity measured by a Price Type current meter. Current meters were calibrated to ±0.004 $m\ s^{-1}$ by the National Hydraulics Laboratory, Burlington, Ontario and Gray and Wigham (1970) estimate the total error in the velocity-area method to be between 5 and 10 percent. During the 1982 spring thaw the stilling wells at the Heart and Hardill Lake

TABLE 2.3 Net radiation and temperature equations

Site	Year	b_0	b_1	δ	r^2
Net Radiation					
Uplands	1982	0.88	0.57	0.92	0.96
	1983	1.77	0.34	0.83	0.85
Wetlands	1982/83	0.93	0.57	1.57	0.84
Lake	1982	0.01	0.69	0.86	0.90
	1983	0.53	0.54	1.14	0.89
Air Temperature					
Wetlands	1983	1.15	0.86	1.04	0.93
Lake	1983	0.83	0.90	0.91	0.96

r^2 is the coefficient of determination and δ is the standard error of the predicted in MJ m^{-2} d^{-1} for net radiation or ^{o}C for temperature.

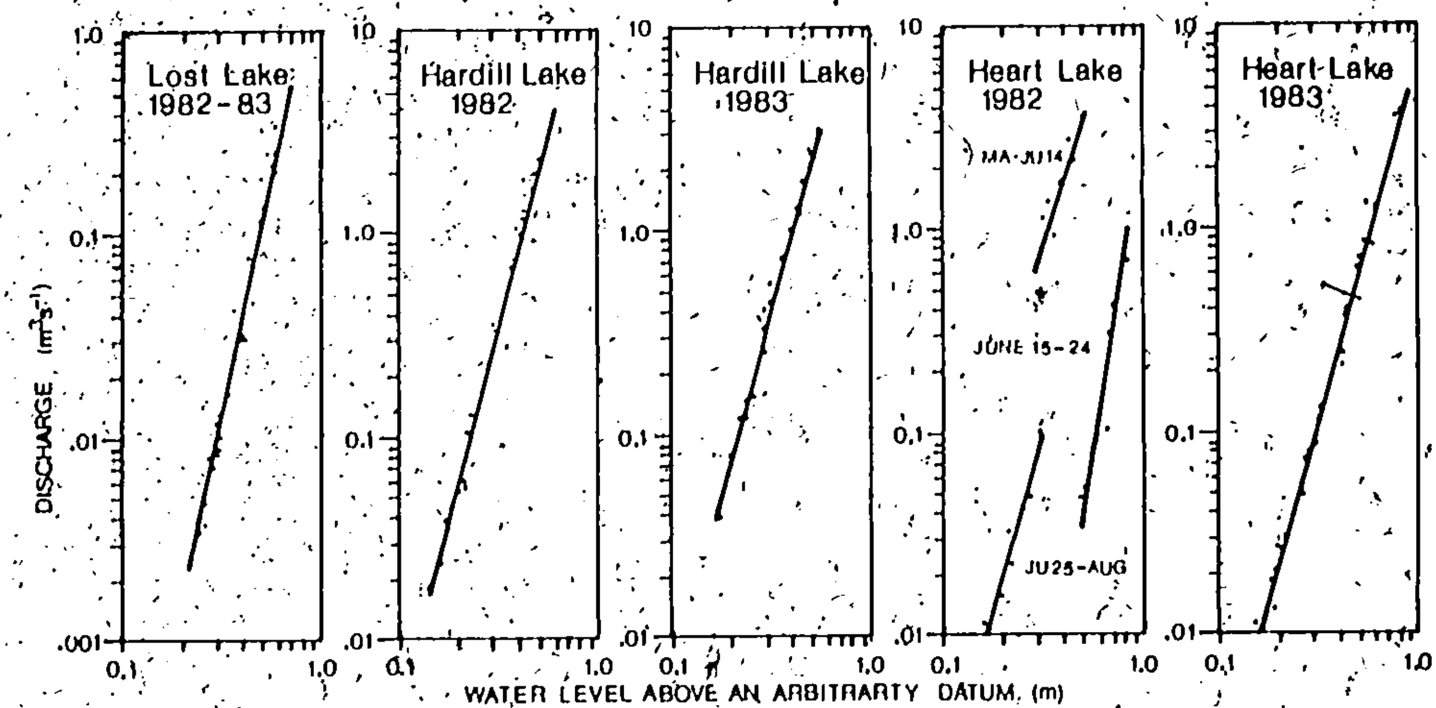


Figure 2.6 Stage-discharge relationships for Lost Lake, Hardill Lake and Heart Lake outflows, 1982 and 1983.

TABLE 2.4 Rating Curve Error Coefficients

Outflow	Date	a	b	r^2	Std. Error*
Heart	1982(1)	4.619	5.536	0.94	0.315
	1982(2)	0.793	5.439	0.86	0.281
	1983	1.742	3.380	0.98	0.215
Hardill	1982	3.904	4.283	0.98	0.211
	1983	2.870	3.225	0.96	0.202
Lost	1982-83	3.957	7.313	0.97	0.219

(1) May 31 - June 10 (2) June 11 - Aug. 4

The form of the rating curve is $\ln(q)=\ln(a)+b\ln(S)$, where q is discharge in m s^{-1} and S is stage in m.

* The standard error relates to the error of the transformed model.

outlets were removed several times by ice and high flows. During spring 1983 Lost Lake discharged for two days while the outlet was still blocked with snow. Discharge was estimated by assuming an exponential increase for the first two days. Flows were small at that time so that little error would be introduced.

In 1983 water level in Moose Pond and the intermittent stream west of Heart Lake were measured using Type F water level recorders. The maximum area of both Heart Lake and Moose Pond in spring were surveyed.

Surface Flow

Surface flow from a 20 m width of wetland and upland site was collected during the 1983 field season. Flows were routed through a 70°V-notch weir and water level behind the weir was recorded. For flow below 2.0 l s^{-1}, discharge was estimated from a stage-discharge relationship. Discharge was measured as the rate at which a known volume passed through the V notch. For higher flows discharge was estimated using an empirical triangular weir calibration formula (Shen, 1981),

$$q = C_e \frac{8}{15} (2g)^{1/2} \tan \frac{\theta_v}{2} h_e^{5/2} \qquad (2.13)$$

where the effective head, h_e, is,

$$h_e = h_w + \frac{0.002}{\sin \theta_v/2} \qquad (2.14)$$

Here θ_v is the angle formed by the V notch and hw is the measured head above the V notch. C_e is the coefficient of discharge and has a value of 0.577 for a 70° V-notch weir (Shen, 1981). Imperial units were converted to SI after computation. Equation 2.13 can be applied to discharges of 2.0 l s^{-1} or greater using a 70°V notch with little error (Ackers et al., 1978).

Subsurface Flow

Subsurface flow was estimated from flow nets derived from a network of ground water wells at the wetland and upland sites. Sixteen wells were installed in the wetland and 12 wells in the upland. Water level and frost table depth were manually sampled at two day intervals and daily subsurface discharge, q_{ss}, was computed using Darcy's equation,

$$q_{ss} = -Kwd_s \frac{dh}{dx} \tag{2.15}$$

where K is hydraulic conductivity, h is water elevation and x is horizontal distance. Thickness of the saturated subsurface zone, d_s, is the elevation difference between the frost and water tables and w is the width of the subsurface flow zone. The gradient, dh/dx, was estimated by the gradient of the flow paths between wells. Subsurface flow was computed applying Dupuit's assumptions: all flow lines in the saturated zone are horizontal and parallel to the water table and flow is of uniform velocity through the vertical cross section of the saturated zone (Dunne and Leopold, 1978). When a shallow active layer occurs the depth of saturation is small and therefore the error introduced by invoking these assumptions will be small.

Frozen and Unfrozen Soil Infiltration

Infiltration was measured both in frozen and unfrozen soil using double ringed infiltrometers. Two infiltrometers were installed in the wetland and upland site in the summer of 1982 and were used in the spring of 1983 to estimate frozen soil infiltration capacity. Unfrozen soil infiltrometers were installed in the wetland and on both vegetated and bare upland soil. Infiltrometers were installed to depths of 0.03, 0.08 and 0.12 m. Absolute infiltration capacity derived using double ringed infiltrometers may be as much as 30 percent in error (Burgy and Luthin, 1957) but for relative comparisons they yield satisfactory results (Wilm, 1941).

Ground Ice Contribution

The amount of water released by daily ground ice melt was computed as,

$$\bar{I} = I(z)\ (\rho_i/\rho_w)\ \Delta z/\Delta t \qquad (2.16)$$

where ρ_i is the density of ice and $\Delta z/\Delta t$ is the daily increase in thaw depth. The volumetric ice content at depth z, I(z), was assumed to equal porosity. The validity of this assumption is examined in Chapters Four and Five.

Soil Water Storage

The depth of water stored in the upland or wetland unfrozen soil layer, S, was derived as,

$$S = \int_0^{z_1} \theta(z)\,dz + \int_{z_1}^{z_2} n(z)\,dz \qquad (2.18)$$

where θ is the soil moisture content and n is the soil porosity. z_1 and z_2 represent the depth of water table and frost table. Water table depths were continuously recorded at one well in the wetland site and at one in the upland site during both field seasons. Soil moisture content was determined weekly in 1983 using the gravimetric approach and converted to volumetric measures using soil density. To avoid introducing errors the density of individual soil samples, instead of an overall density was used in this calculation.

Additional daily water and frost table measurements were made in a hummock region west of Heart Lake. A row of wells, perpendicularly bisecting the intermittent stream, was maintained in July and August, 1983.

2.3.4 Comment on Error

In hydrological data there can be several kinds of errors. Errors discussed so far are mostly measurement errors. Additional errors can be introduced during data processing. Finally errors are produced by the variability, fluctuation and uncertainty in the sample data obtained. It is usually assumed that sampling errors are random and that the sample's statistical properties approach those of the population when the sample size is large (Haan, 1977). Large sample sizes were obtained whenever possible, but the samples of certain variables, for example soil moisture, were limited because of logistics.

It is difficult to asses the overall error in hydrological studies that combine the measurement of many variables and parameters, particularly when this data is extended to the basin scale. There are some estimates of the relative size of the sampling error produced by some of the techniques used in this study. The snow sampler used has an error of -1.8 to 13.0 percent (Goodison et al., 1981) and the survey technique yields a 15 percent error on average (Woo and Marsh, 1978). Rain gauges are generally believed to under catch rainfall. At a typical central Keewatin wind speed of 6 m s^{-1} the standard rain gauge catches approximately 70 percent of 'true' rainfall (Bruce and Clark, 1966). Some recent evidence suggests, however, that at wind speeds of 3 m s^{-1} or less the standard rain gauge tends to over catch (B. Goodison, personal communication). The error in the technique used to estimate evaporation is not known. Random errors could be as great as 20 percent at any one time (J. Davies, personal communication), but more important is the possible systematic error that would be introduced with measurement or calculation of, for example, net radiation or ground heat flux.

In the remaining chapters several comparisons are made between computed and measured water storage through the evaluation of a water balance. At the plot scale this approach illustrates where the errors are most likely to exist. In Chapter Six an error analysis is undertaken on the basin water balance equation.

CHAPTER THREE BASIN INPUTS AND OUTPUTS

An important element of basin hydrological analysis is the definition of the system boundaries and the quantification of the water flux across boundaries. If inputs and outputs can be correctly determined, then emphasis can be placed on the understanding of the internal linkages within the system itself. In the previous chapter, the basin boundaries were illustrated. In this chapter, the inputs of snow meltwater, rainfall and streamflow into the basin and the outputs of evaporation and basin discharge are examined. These components are used in the following two chapters to discuss the storage and flow of water in individual elements of the basin system and in Chapter Six to evaluate the basin water balance.

3.1.0 Snow Distribution and Melt

Woo et al. (1983) estimated that snow provided 70 percent of the total precipitation input to a high Arctic drainage basin. For central Keewatin the climatic norms indicate that 42 percent of the precipitation occurs as snow. Snow that falls on the ground is subject to redistribution by wind. In the Arctic, Woo and Marsh (1978) found greater depths of snow in depressions and very little snow in exposed elevated areas. In the present study, the total snow water equivalent was dominated by an area of large accumulation in the lee of the scarp running through the centre of the basin. The distribution of water equivalent prior to the 1982 and 1983 snowmelts is shown in Figures 3.1 and 3.2 respectively. Table 3.1 lists the mean snow depth and density for each of the six basin terrain types used in the snow survey. The mean basin water equivalent in 1982 was 135 mm and in 1983 it was 201 mm.

Water equivalent at the upland site was 106 mm in 1983 and McMillan (1983) estimated it to be 33 mm in 1982, but the 1982 survey was conducted after melt had begun. At the wetland site water equivalent was 142 mm in 1982 and 146 mm in 1983. Based on these results the difference in basin water equivalent reported above could be due, in part, to the limited sample size used in the 1982 survey.

The snowmelt pattern for the two study years was quite different (Figure 3.3). In 1982, snowmelt began on May 22 and proceeded rapidly until May 29. During this time, the snowcover became isothermal and localized water flow was discernible in certain parts of the basin. A cold spell followed and the meltwater refroze during the next 12 days. By then, approximately 30 percent of the total basin area was snow free (Figure 3.4). Melt resumed on June 10 and the remaining snowcover was soon depleted. This pattern

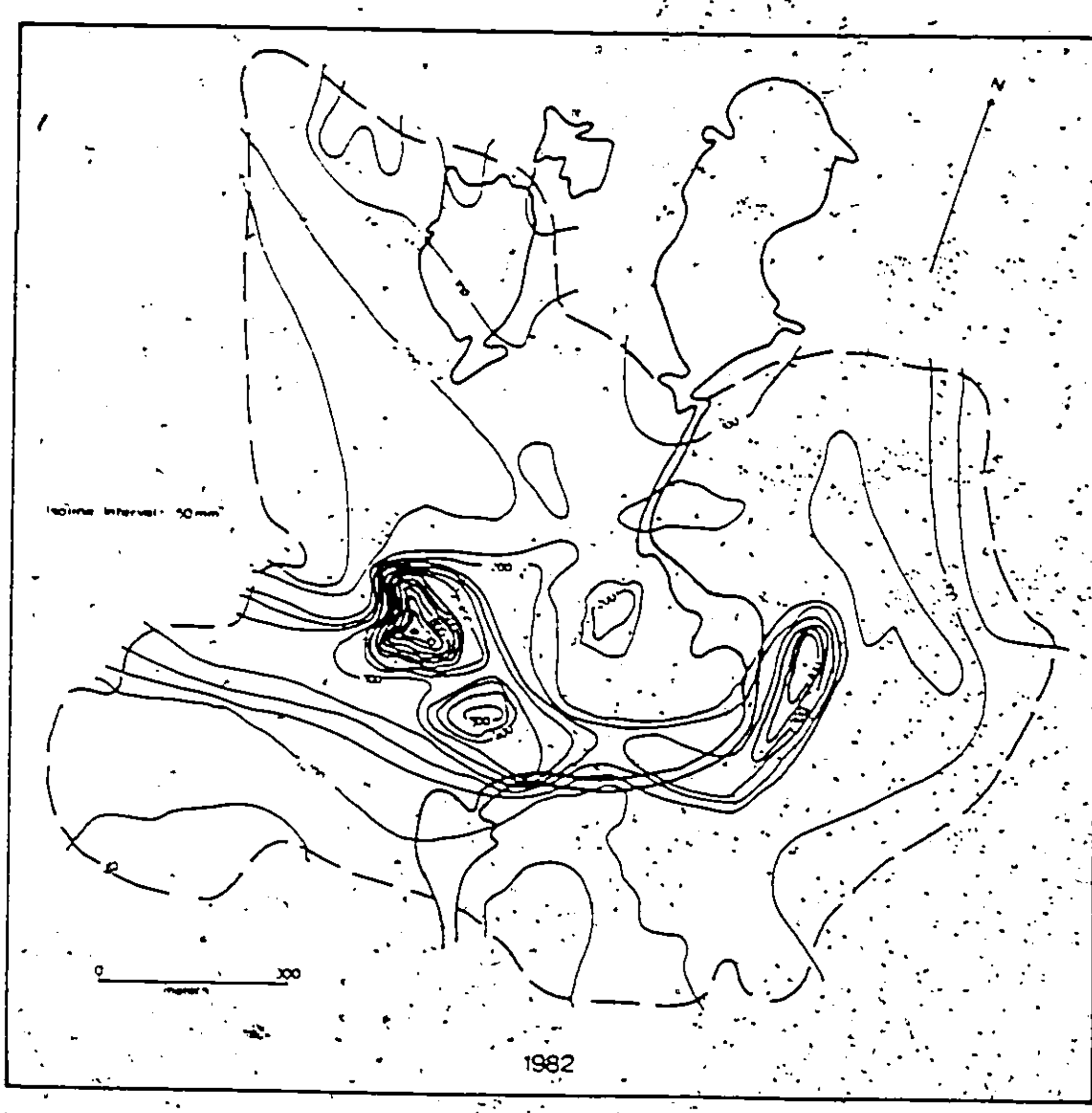


Figure 3.1 The distribution of snow water equivalent, May 23, 1982. Note: the number of sites sampled was much lower in 1982 than in 1983 producing the difference in detail between the above Figure and Figure 3.2.

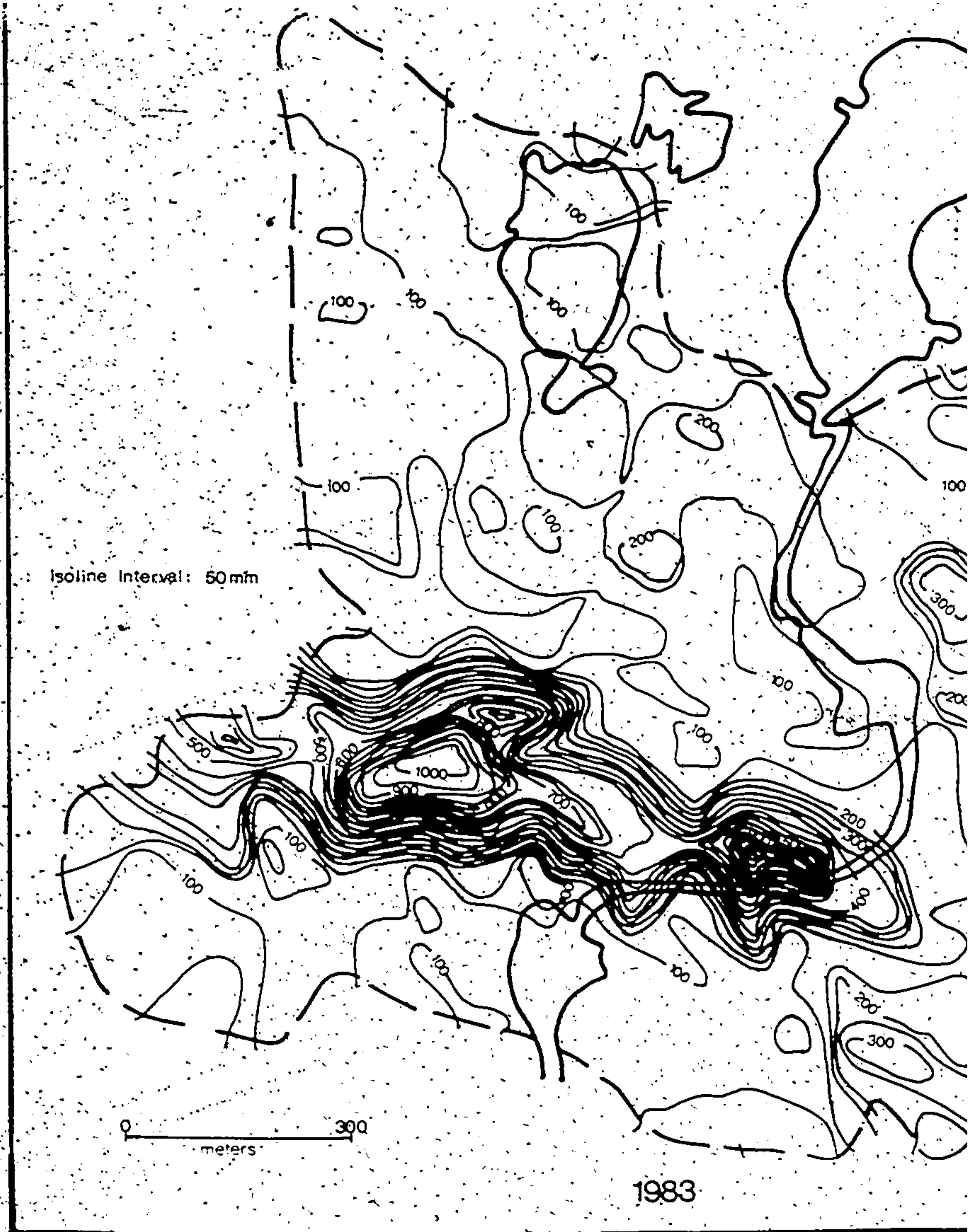
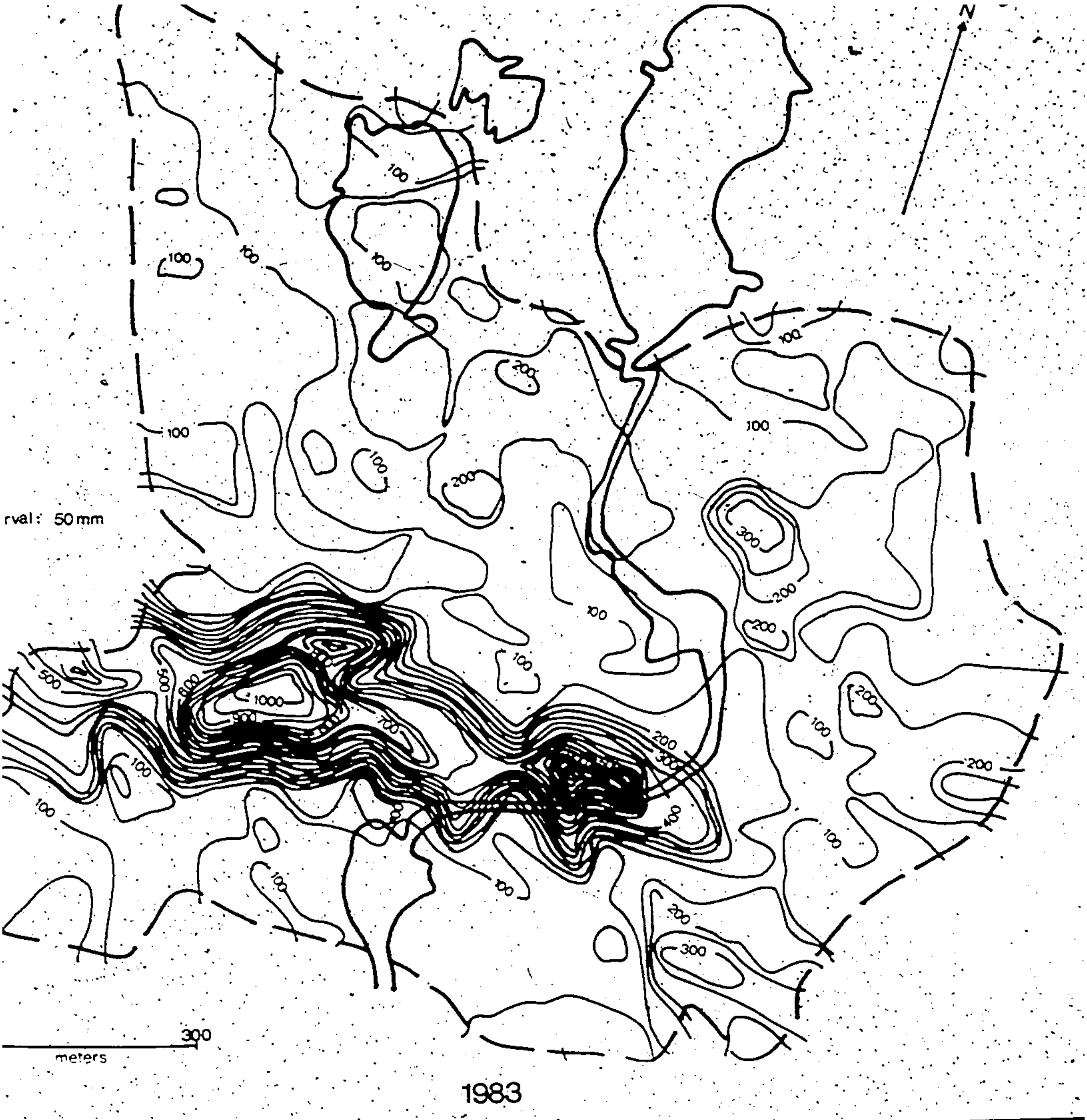


Figure 3.2 The distribtuion of snow water equivalent, May 18, 1983.



The distribtuion of snow water equivalent, May 18, 1983.

TABLE 3.1 Terrain Snow Survey

Terrain Types	N	Depth (mm) Mean	s	N	Density (kgm^{-3}) Mean	s
Lakes	88	0.259	0.296	23	384	87
Flat Wetlands	76	0.457	0.725	20	366	47
Slopes	87	0.278	0.128	23	324	83
Ridges Tops	87	0.200	0.243	21	394	85
Valley Bottoms	90	0.354	0.099	23	323	39
Scarp Slopes	91	2.270	1.194	22	394	85

N is the number of samples and s is the standard deviation.

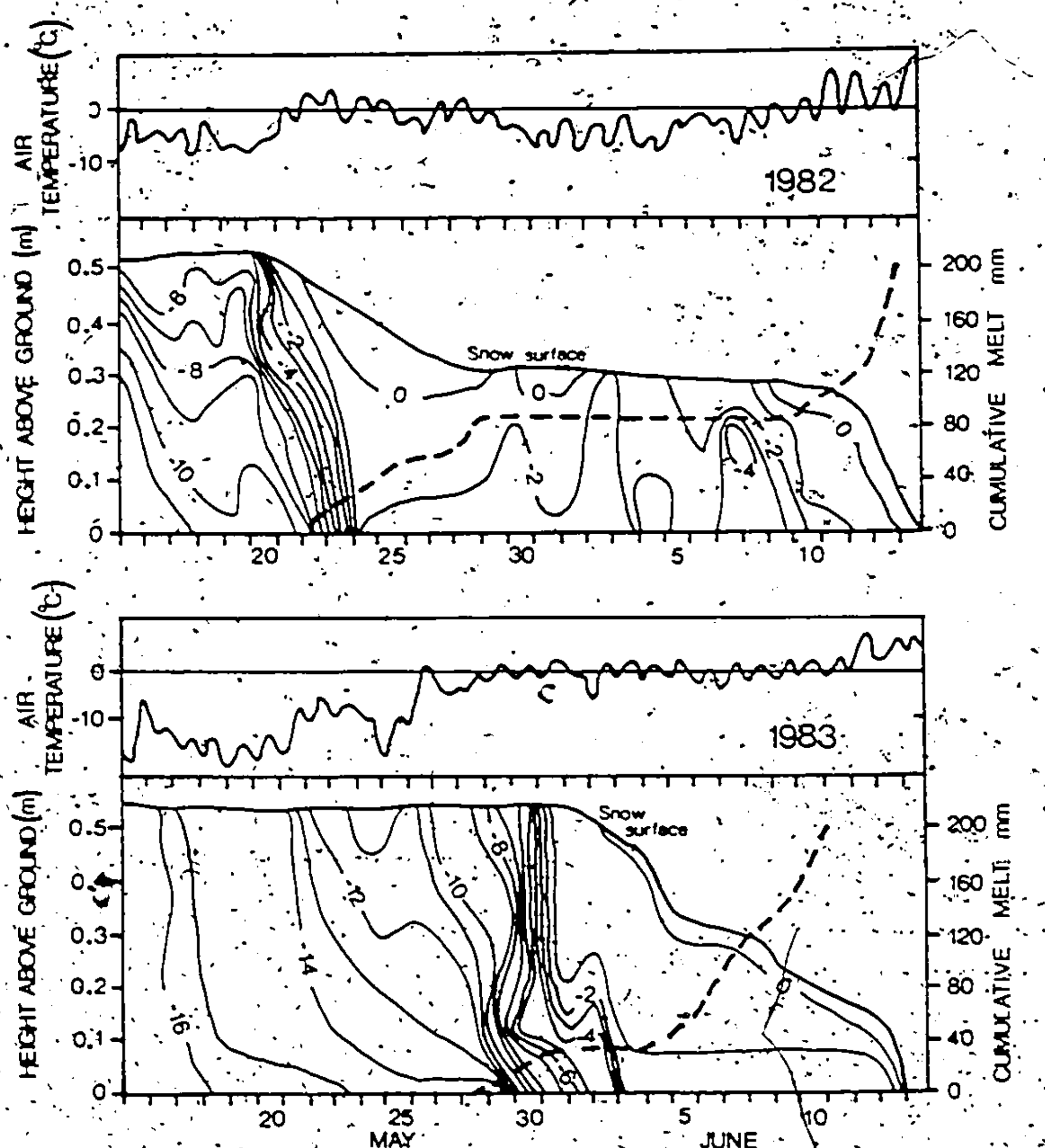


Figure 3.3 Snow depth and temperature, air temperature and cumulative melt in mm (dashed line) at the snowmelt site, 1982 and 1983.

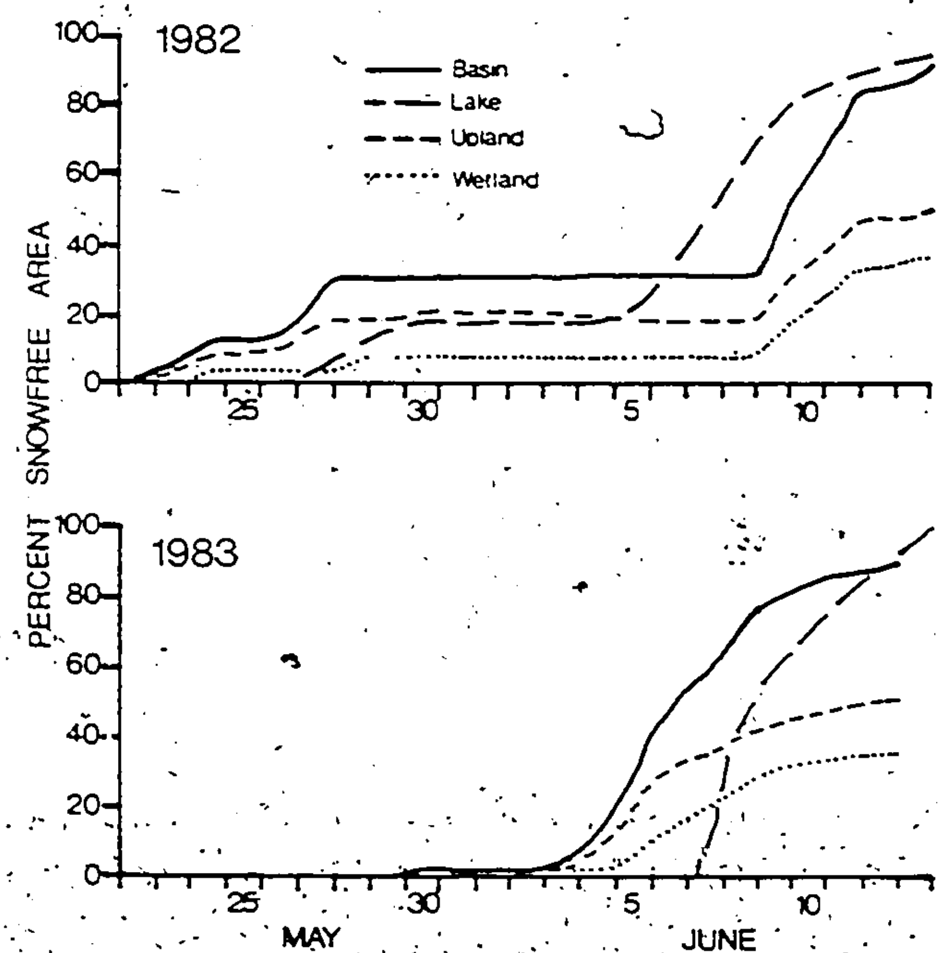


Figure 3.4 Percent snowfree area of the basin and individual cover types during the 1982 and 1983 snowmelt. Percent is expressed relative to the total area of each cover type.

contrasted sharply with 1983 when the melt began on May 28. The snowcover ripened quickly and the basin was completely snow free by June 15, except for the area below the central ridge. For a portion of the 1982 melt there were lower daily air temperatures and lower nocturnal temperatures than in 1983. It is interesting to note that while basal ice of the form discussed by Woo and Heron (1981) was not common in the basin, at the snowmelt site it did form on June 3, 1983 and melted by June 16.

In Figure 3.4 the percentage of ice free lake surface is shown. Ice cover melt began somewhat later than that of the land snowcover but proceeded rapidly.

3.2.0 Rainfall

A comparison of the 1982 and 1983 rainfall measured at Baker Lake with the 30 year mean showed abnormalities (see Figure 2.1). A significant difference in measured precipitation also occurred in the basin between the two years during the field periods. In 1982, 108 mm of rain fell between May 24 and August 5 of which approximately 50 percent fell during two storms in July (Table 3.2). There was no single large rain storm in the 1983 field period and 39 mm of rain fell between June 9 and August 10.

Depth of rain was estimated for all storms greater than 2 mm by the Thiessen polygon method. There was no systematic spatial variability to the rainfall pattern

TABLE 3.2 Basin Rainfall for 1982 and 1983

Date 1982	mm	Date 1983	mm
May 24	1.3	June 9	12.4
		June 14	0.7
June 11	1.3	June 24	0.3
June 20	14.0		
June 23	15.7	July 1	0.7
June 26	1.3	July 5	0.3
		July 7	0.3
July 14	6.5	July 9	0.7
July 15	25.9	July 10	0.3
July 21	27.4	July 12	9.8
July 28	9.0	July 13	0.7
July 29	1.3	July 16	0.3
July 30	1.3	July 22	1.0
July 31	3.2	July 28	1.0
		July 31	1.0
		Aug 5	9.2
Total	108.2		38.7

(Table 3.3). The arithmetic and areally weighted mean rainfall were quite similar for most storms. The main tipping bucket rain gauge (see Section 2.3.2) systematically recorded less rainfall than the basin rain gauge network, suggesting a lower catch efficiency for the tipping-bucket rain gauge. The ratio between the two measures was 1.3. Smaller rainfalls were estimated from the tipping bucket measurement and adjusted according to this ratio.

3.3.0 Evaporation

The study basin was divided into three different surface types for the purpose of hydrological investigation (see Section 2.1.0). These are lakes and channels, wetland and upland surfaces. Because of differences in moisture characteristics, vegetation and microtopography, evaporation was expected to be different from each surface. Studies were conducted to quantify such differences.

3.3.1 Lake and Wetland Evaporation

Roulet and Woo (1985a) discuss in detail the results of a study of evaporation from Moose Pond and the wetland site. They found that over a period of two weeks, total lake and wetland evaporation were the same, but there were large day-to-day variations in evaporation losses from each surface. The mean α value for the lake was 1.29 ± 0.41 and for the wetland it was 1.56 ± 0.32 (Table 3.4). These two α

TABLE 3.3 Comparison of Basin and Main Rain Gauges

	1982						1983	
	June		July				July	Aug
Gauge	20	23	15	21	28	31	12	5
1	13.1	14.9	28.6	28.5	9.7	3.3	na	11.4
2	20.2	15.7	22.4	30.7	8.4	na	11.9	9.9
3	12.8	15.8	23.4	37.0	8.9	2.6	9.4	10.8
4	14.8	16.4	27.5	22.7	9.8	2.6	12.0	8.4
5	13.4	16.2	22.8	26.4	9.3	3.6	11.4	8.1
6	13.7	16.6	28.7	32.9	8.4	5.4	11.4	9.4
7	12.6	15.1	24.6	24.6	7.6	3.2	10.6	8.0
8	12.8	15.5	30.7	33.7	9.3	na	8.8	15.2
Mean	14.2	15.8	26.1	29.6	8.9	3.5	10.8	10.2
s	2.5	0.6	3.2	4.9	0.8	1.0	1.3	2.4
Thiessen	14.0	15.7	25.9	27.4	9.0	3.2	9.8	9.2
Main	8.0	14.0	22.0	29.0	8.0	3.0	5.8	6.3
Ratio	1.8	1.1	1.2	0.9	1.1	1.1	1.7	1.5

1. Thiessen refers to the Thiessen polygon method for estimating rainfall total over a basin. See Figure 2.3 for rain gauge network.
2. Main refers to the main meteorological site tipping bucket raingauge at base camp (see Sect. 2.3.2).

TABLE 3.4 Evaporation Parameter (α)

Period	Upland 1[a]	Upland 2[b]	Wetland 1	Wetland 2[d]	Lake
N	21	28		21	15
Mean	1.29	1.01	1.29[c]	1.56	1.29
s	0.47	0.24		0.32	0.41
Maximum	2.53	1.42		2.14	1.56
Minimum	0.49	0.43		1.06	0.72

a. Period 1 for upland was calculated using the 1982 lysimeter data.
b. Period 2 for upland was calculated using the 1983 energy balance.
c. When the wetland flooded it was assumed α was equal to the lake condition.
d. Period 2 for the wetland was calculated using the energy balance Bowen ratio approach.

values are significantly different at the 95 percent confidence level. The lake α value is similar, but the wetland value is larger than those reported by Rouse et al. (1977) for similar conditions. Factors influencing α are discussed by Roulet and Woo (1985a). The relationships between actual evaporation and equilibrium evaporation for the lake and the wetland are shown in Figure 3.5.

Daily evaporation for the wetland was estimated for 1982 and 1983 using equation 2.11. An α of 1.56 was used for the period when the water table was below the surface and 1.29 for the time when the water table was above the surface (Figure 3.6). Similar estimates were made for lake evaporation assuming an α of 1.29 (Figure 3.7). Radiation and air temperature were estimated for days with missing record using regression equations presented in Table 2.3. There was no measurement of net radiation over the wetland during the 1982 field season so that the estimated wetland evaporation for that field season is prone to error. Wetland evaporation was considerably lower in 1982 (Table 3.5). This results because the water table was above the ground surface for a greater duration in 1982 than in 1983 thus an α value of 1.29 was used for a longer period of time in 1982. This had the effect of reducing the daily evaporation estimate in 1982. Whether this in fact occurred is not known, but the 1983 wetland water balance calculation presented in Chapter Five supports the 1983 estimate of wetland evaporation. The

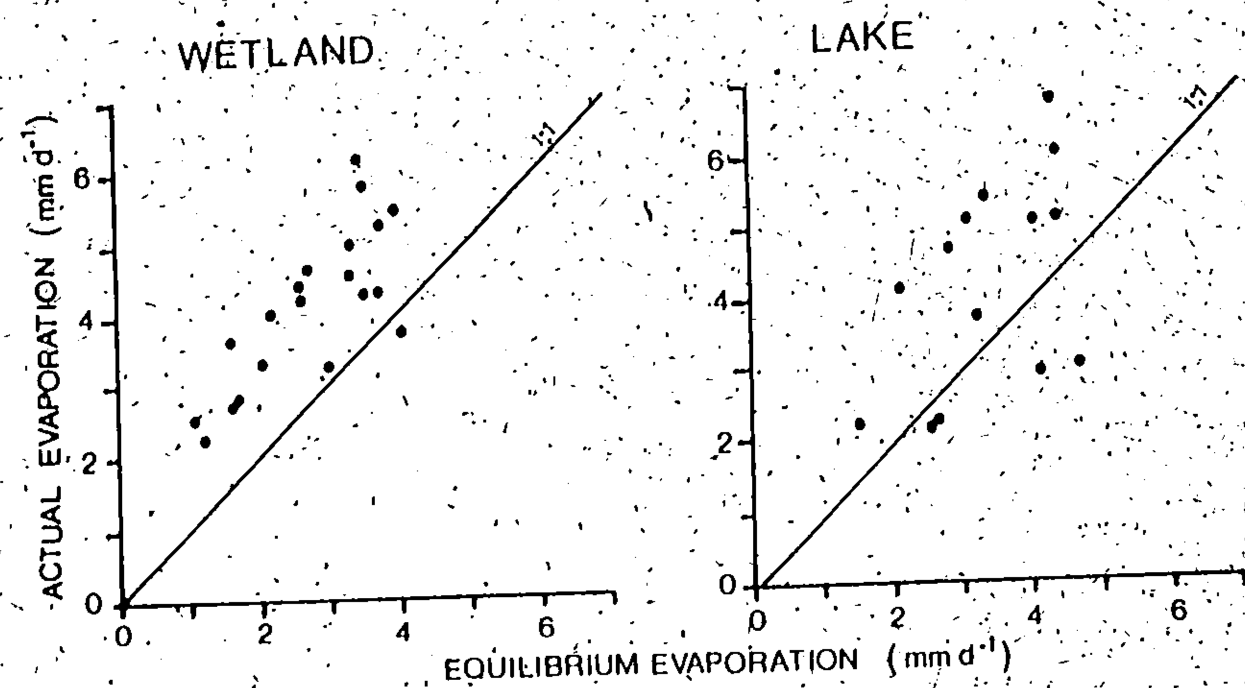


Figure 3.5 Wetland and lake measured and equilibrium evaporation estimates.

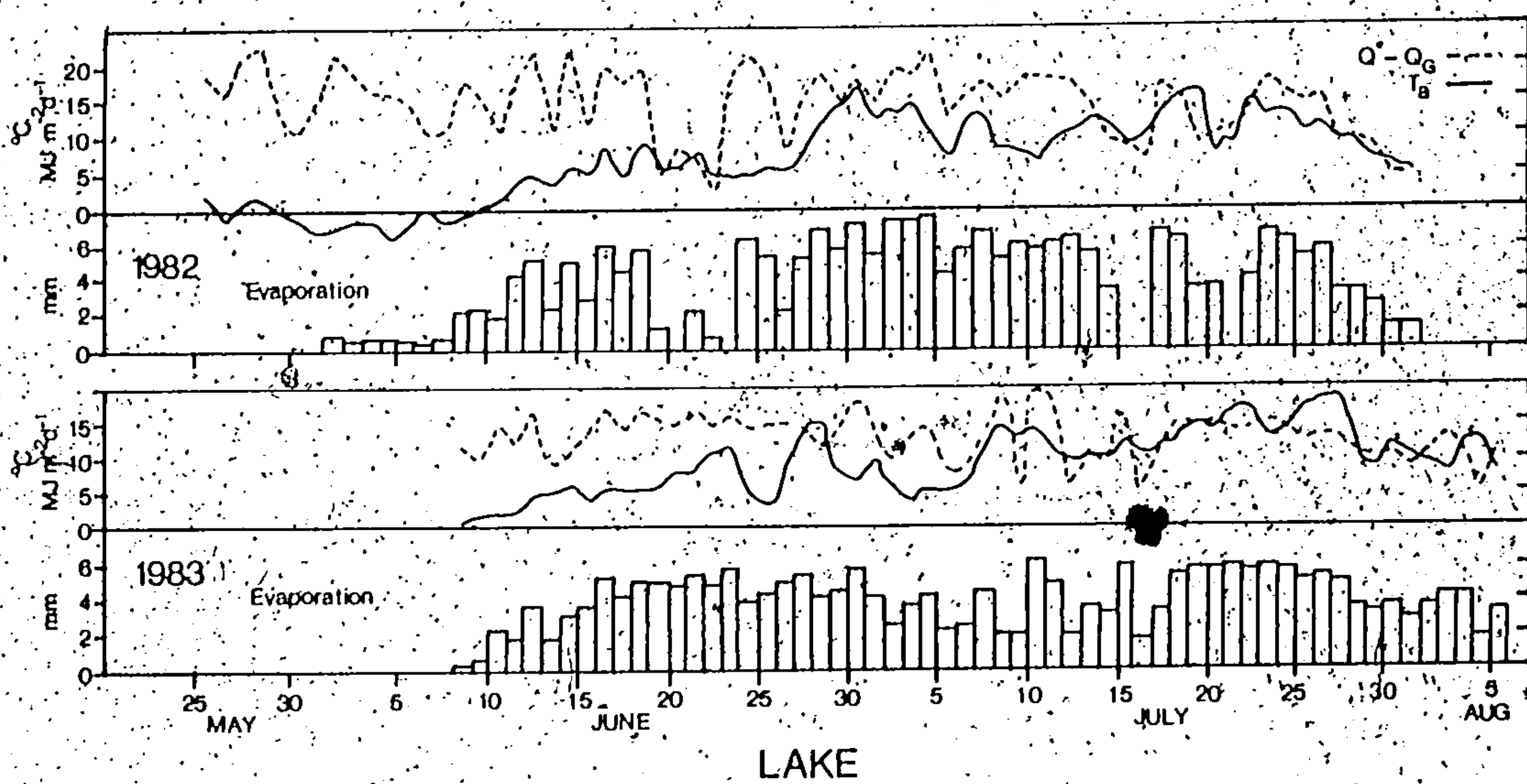


Figure 3.6 Daily lake evaporation computed using the Priestley-Taylor model, air temperature and net radiation for 1982 and 1983.

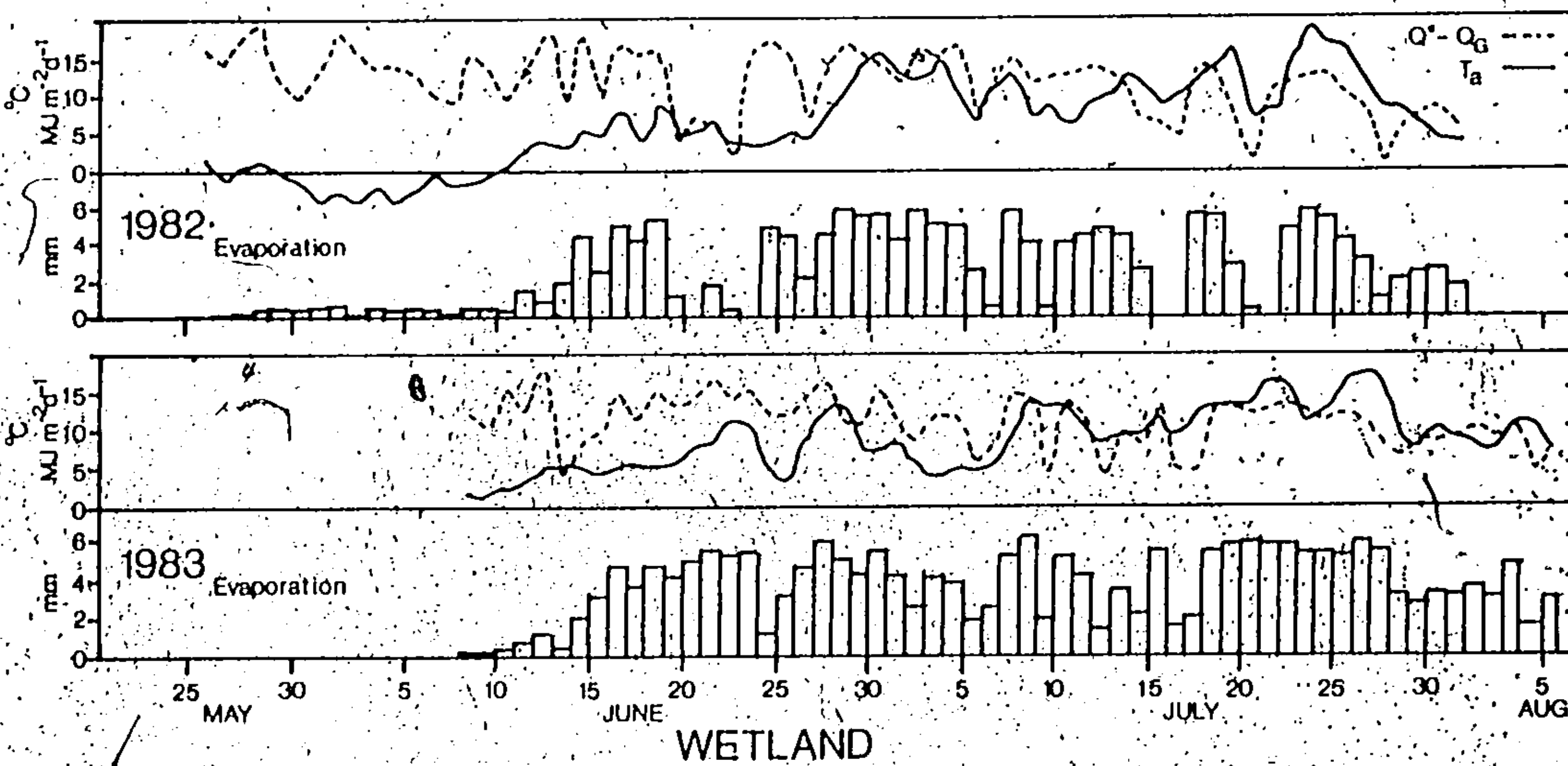


Figure 3.7 Daily wetland evaporation computed using the Priestley-Taylor model, air temperature and net radiation for 1982 and 1983.

TABLE 3.5 Evaporation, e (in mm) 1982-1983

	Upland 1982	Upland 1983	Wetland 1982	Wetland 1983	Lake 1982	Lake 1983
No. of Days	69	59	69	59	61	59
Total e	153	133	175*	223	239	232
Mean daily e	2.35	2.25	2.76*	3.78	4.12	3.93
s	1.31	0.98	5.96*	1.81	2.18	1.41
Maximum daily e	5.20	4.25	5.96*	1.81	2.18	1.41
Minimum daily e	0.0	0.35	0.0*	0.08	0.0	0.29

These values were derived using the α values shown in the Table 3.4. * Q^* for 1982 wetland was estimated by regression.

lake evaporation was similar in both years.

3.3.2 Upland Evaporation

The loss of water by evaporation at the upland site was estimated by two different methods. In 1982 lysimeters were used, but only after snowmelt and large rainstorms did they yield reliable estimates. In 1983 the temperature gradient over an upland surface was continuously measured, and with the bulk tranfer approach the sensible heat flux was calculated. The latent heat flux was then derived as the residual of the surface energy balance. The daily upland energy balance for 27 days in 1983 is shown in Table 3.6. It should be noted that any error in the evaluation of net radiation, ground heat flux or sensible heat flux will be accumulated in the estimate of latent heat. Since errors in the energy balance are additive the maximum error in the upland evaporation estimate could be as great as 40 percent. It is, however, unlikely all errors will have the same direction and therefore the mean error would be much less. Bowen ratios have been calculated for these days and it can be seen they are low, suggesting no moisture restriction (Table 3.6).

From the estimation of latent heat, an α of 1.01 ± 0.24 was obtained for periods when the water table was below 0.10 m (Table 3.4). Comparison of estimated evaporation with equilibrium evaporation is presented in Figure

TABLE 3.6 Daily Upland Energy Balance 1983

Date	Q^*	Q_G	Q_H	Q_E	B
June 28	11.59	1.81	2.44	7.34	—0.33
June 29	12.23	1.88	4.94	5.41	0.92
June 30	15.09	1.95	6.17	6.97	0.86
July 2	9.61	2.09	3.36	4.16	0.81
July 3	13.50	2.16	8.01	3.33	2.41
July 4	13.10	2.23	8.65	2.22	3.90
July 5	5.55	2.29	1.57	1.69	0.93
July 10	13.58	2.57	3.66	7.36	0.50
July 11	12.34	2.61	6.28	3.46	1.82
July 12	4.85	2.65	0.72	1.48	0.49
July 13	10.33	2.68	3.59	4.06	0.88
July 14	8.41	2.71	2.06	3.64	0.57
July 15	12.90	2.73	5.41	4.77	1.13
July 16	4.50	2.74	0.17	1.59	0.11
July 17	8.63	2.75	1.07	4.81	0.22
July 18	13.20	2.75	3.87	6.58	0.59
July 20	13.35	2.72	4.42	6.21	0.71
July 22	12.74	2.69	3.88	6.16	0.63
July 23	12.35	2.66	4.48	5.21	0.86
July 28	8.41	2.24	2.47	3.70	0.67
July 29	5.77	2.13	1.57	2.08	0.76
July 30	7.03	2.01	2.15	2.87	0.75
July 31	8.67	1.87	1.86	4.94	0.38
Aug 2	11.62	1.56	5.40	4.66	1.16
Aug 3	9.95	1.39	2.35	6.22	0.38
Aug 5	9.08	0.99	2.81	5.27	0.53
Aug 6	6.46	0.77	1.60	4.08	0.39

1. All components of the energy balance are expressed in MJ m^{-2} d^{-1}

3.8a. The 1982 lysimeter data were used to derive α for periods when the water table was near the soil surface. The mean α was 1.29 ± 0.47 under these conditions (Table 3.4), but there was considerable scatter in the results (Figure 3.8b).

These two α values were used to estimate evaporation from the upland surface for the 1982 and 1983 study period (Figure 3.9). Computed evaporation is lower than that of the lake and the wetland (Table 3.5). Evaporation was greater in 1982.

The estimates of water loss by evaporation for upland, wetland and lake surfaces will be used in water balance computations in later chapters. The lake α is similar to that reported elsewhere for shallow lakes (DeBruin and Keijman, 1979; Stewart and Rouse, 1976). Any large errors in estimated lake evaporation will likely be due to a measurement error of net radiation because the Priestley-Taylor model is not very sensitive to air temperature. The errors in evaporation from the upland and wetland will become obvious when the water balances are calculated. Using the Priestley-Taylor (1972) model to estimate evaporation from surfaces with temporally and spatially variable conditions is questionable, but no simple alternative exists. In 1983 there was little rain and as will be seen in the next chapter, the upland soil moisture and ground water became depleted. Unfortunately upland evaporation was not

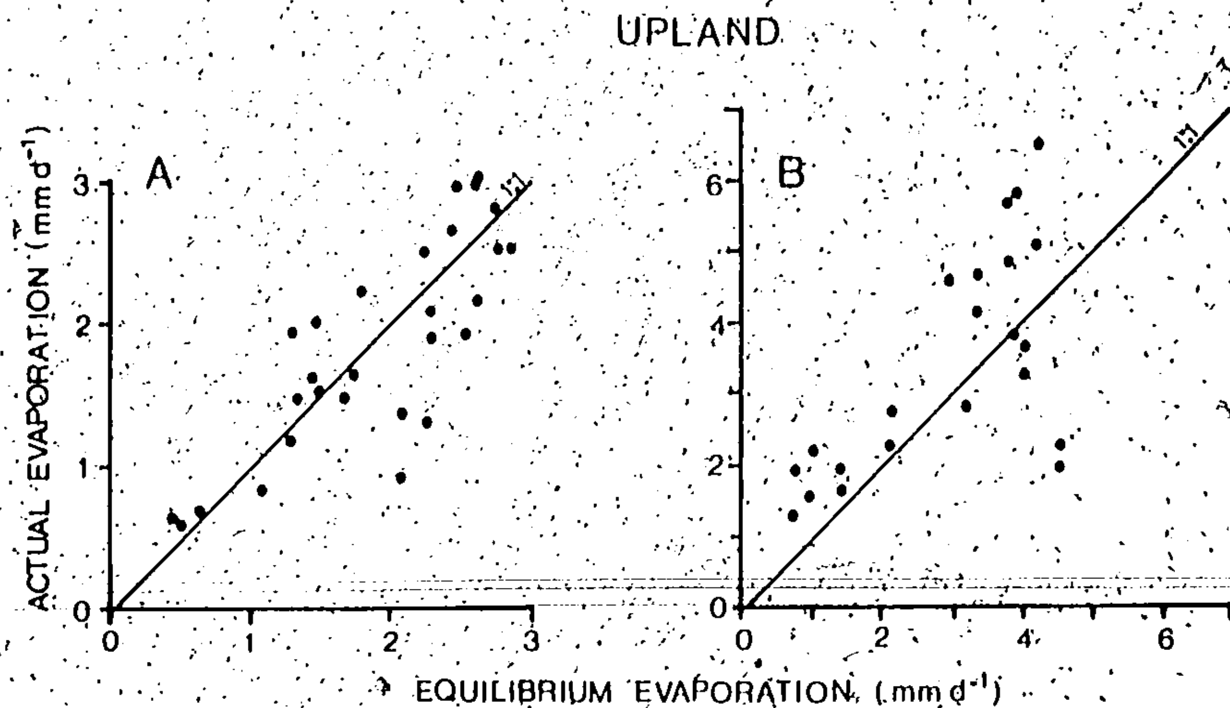


Figure 3.8 Upland measured and equilibrium evaporation.
A. Energy balance evaporation and a water table greater than 0.1 m below the ground surface.
B. Lysimeter evaporation and a water table within 0.1 m of the ground surface.

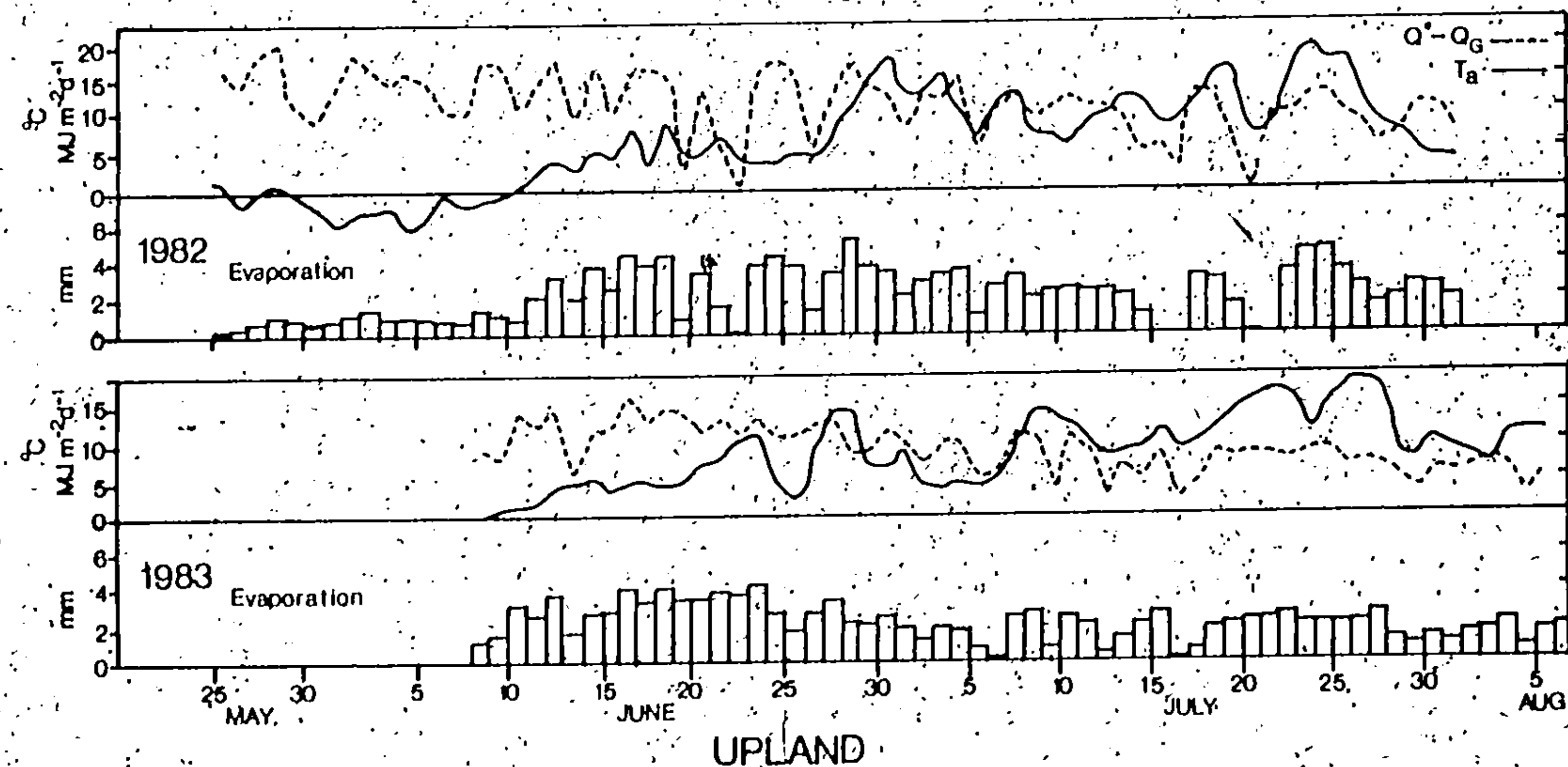


Figure 3.9 Daily upland evaporation computed using the Priestley-Taylor model, air temperature and net radiation for 1982 and 1983.

directly measured as was wetland evaporation. An attempt was made in 1982 to relate α to upland moisture conditions in an similar fashion as Marsh et al (1981) but this failed because surface soil moisture estimates were too unreali- able. Despite the potential for errors, the mean upland α value conforms to the result of Rouse et al (1977).

3.4.0 Discharge and Lake Storage

If not lost to evaporation, water entering the study basin is either stored within the basin or leaves as outflow from Heart Lake. While streamflow is typically a smaller loss than evaporation at low latitudes (Hewlett, 1980; pg. 66), at high latitudes over half of all input is usually discharged (eg. Brown et al, 1968; Ryden, 1977; Woo et al, 1983). The lake and channel system is important in conveying water through the basin, and if flooding, it provides a dynamic linkage with adjacent portions of the land hydro- logical system. Discharge and the storage of water in lakes and channels is discussed in the following section.

3.4.1 Basin Inflow and Outflow

Basin discharge was computed from continuous records of Heart Lake outflow for the 1982 and 1983 field seasons. In 1982 the recording site was disturbed several times during ice breakup and high flow, disrupting the record, but the 1983 discharge record was complete. From the hydrographs it is obvious that the snowmelt season was the dominant discharge period (Figure 3.10). In 1983, over 60 percent of the total discharge had occurred by June 18, only 10 days after streamflow initiation. A similar pattern was evident in 1982, after the period of snowmelt stoppage had passed.

Flow began prior to the initial measurement of water level on May 30, 1982. A peak discharge of 4.3 m^3s^{-1} occurred on June 16, 1982. Discharge decreased to 0.25 m^3s^{-1} by June 21 and was 0.06 m^3s^{-1} on July 5 (Figure 3.10). Flow began on June 8, 1983 and a peak of 2.96 m^3s^{-1} was recorded on June 14. Peak flow was lower than that of 1982 and also lacked variation in the daily flow pattern. This resulted from a difference in the snowmelt conditions as has been discussed previously.

The discharge response to rainstorms was quite variable. During the storm of July 16, 1982, discharge almost doubled from 0.047 m^3s^{-1} to 0.084 m^3s^{-1} (Figure 3.11). This, however, was a small increase compared with a five fold increase in discharge that resulted from a later rainstorm of similar size on July 20 and 21, 1982. Discharge increased from 0.074 m^3s^{-1} to 0.353 m^3s^{-1}. Such a difference in response will be discussed in greater detail in Chapter Six.

The Hardill Lake outflow provided the largest input of water to the study area. For most of the field season water leaving Hardill Lake was confined within the main

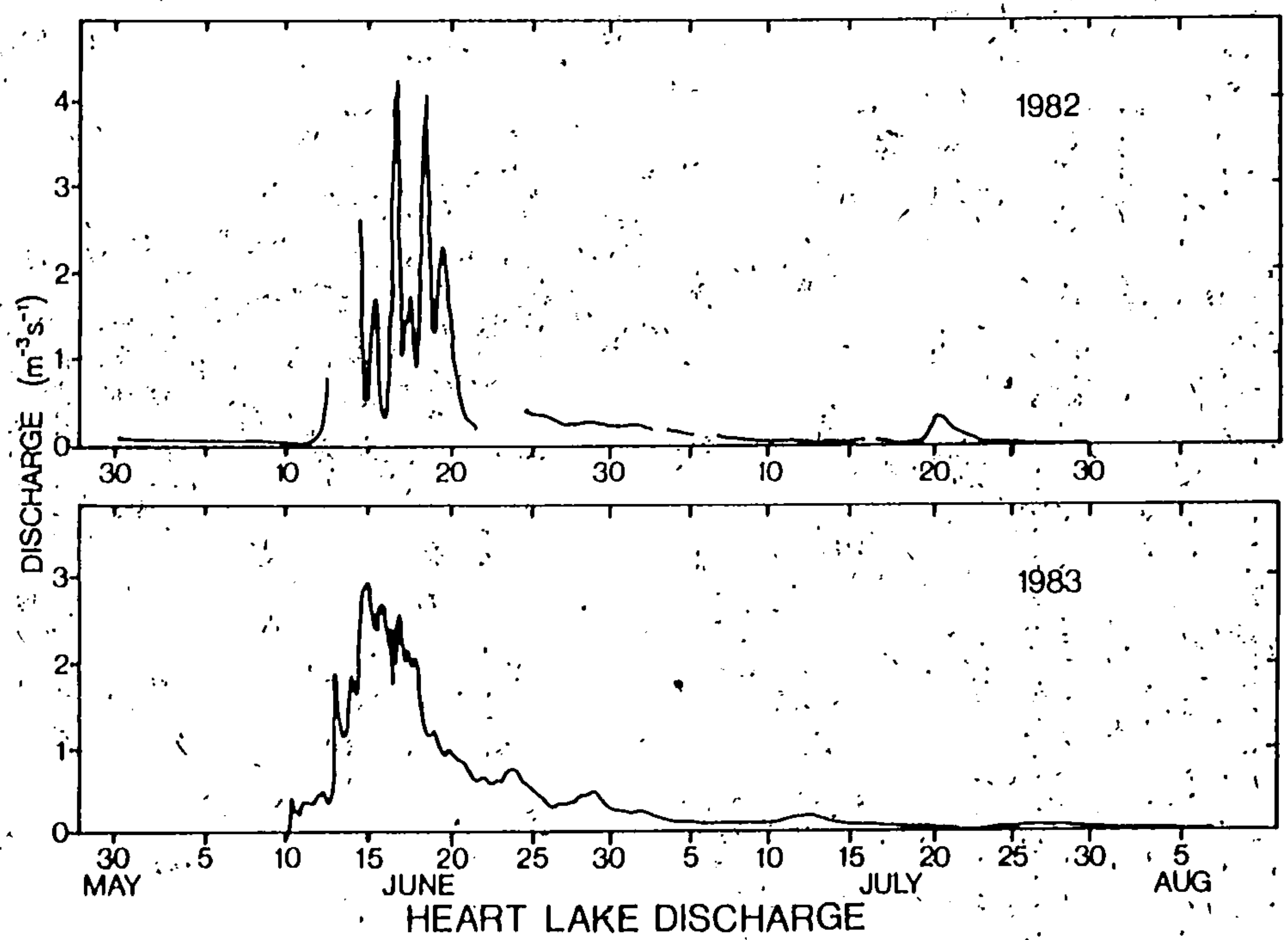


Figure 3.10 Heart Lake discharge, 1982 and 1983.

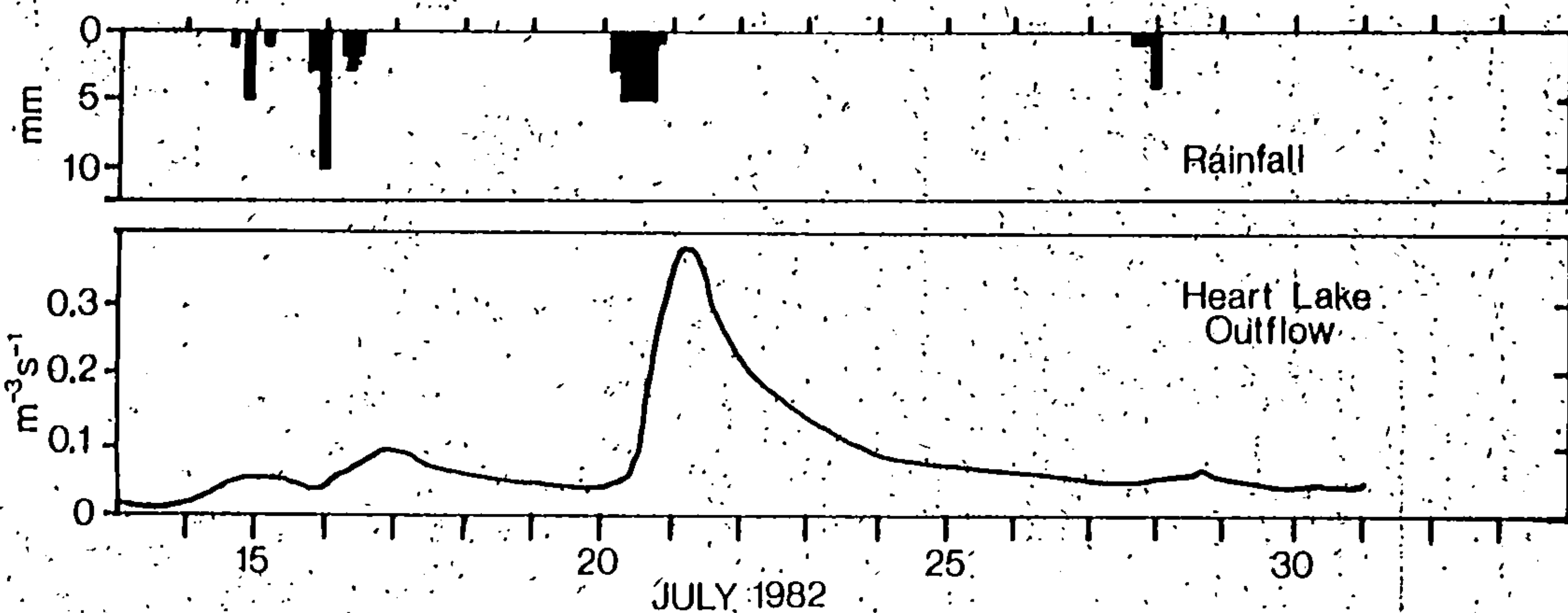


Figure 3.11 Heart Lake discharge during two 1982 storms.

channel, but during snowmelt when overbank flow was common, discharge from Hardill Lake flooded a portion of the land component of the basin. The Hardill Lake discharge record for 1982 was poor but a complete record exists for 1983 (Figure 3.12). In 1983, low flows began on June 10, but increased sharply on June 13, reaching a peak of 2.34 m^3m^{-1} on June 16. Similar to the Heart Lake discharge, diurnal variation was not apparent during the snowmelt period in 1983. Discharge recession was long from snowmelt.

Discharge records from Heart and Hardill Lakes are used to compute the basin water balance (Chapter Six). The seasonal streamflow hydrographs of 1982 and 1983 are similar to those reported for basins of a similar size in the High Arctic (Ryden, 1977; Woo, 1983). While these basins, including the study basin, are quite different physiographically and topographically, the overwhelming influence of melting snow yields similar streamflow regimes. Chyurlia (1977) indicated that discharge from a 21 km^2 basin near Baker Lake did not reflect such a dominance by snowmelt. While his basin was twice the size of the study basin, Chyurlia recorded discharges an order of magnitude lower than those observed in this study. This lower discharge may, in part, be attributed to a larger open water area. There was also a large difference in the water available for runoff. Chyurlia estimated total precipitation, snowmelt and rainfall combined, to be 93 mm

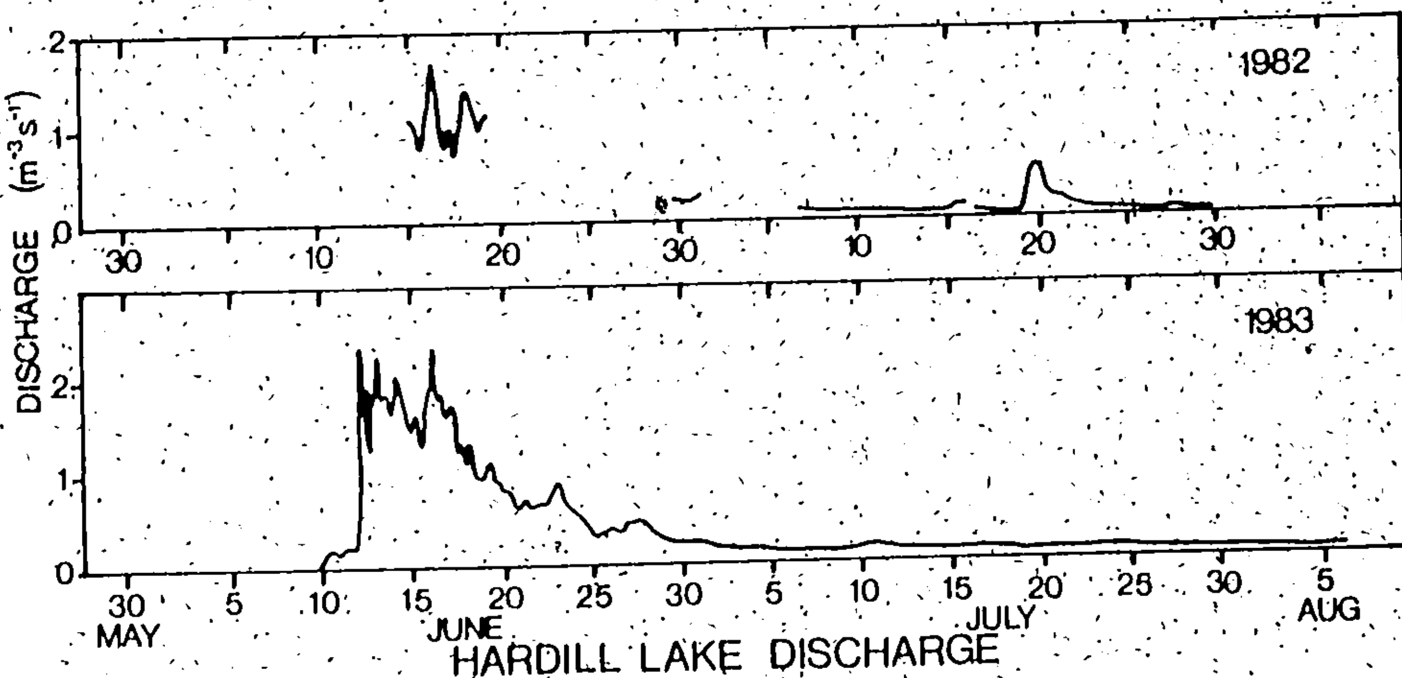


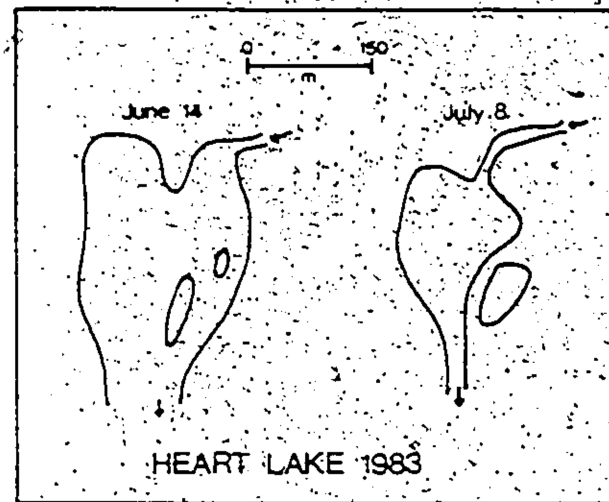
Figure 3.12 Hardill Lake discharge, 1982 and 1983.

(June 7 to July 2 inclusive) which is less than half the input observed in this study. However, the estimates of precipitation were taken from the Baker Lake weather station and Woo et al. (1983) have shown that this can lead to an underestimate of basin snow of 300 percent. It is also possible that Ghyurila's observations began after peak flow had occurred.

3.4.2 Lake and Channel Storage

Lake and channel storage were not quantified on a daily basis for either field season. It is obvious from the long discharge recession that water may be stored for a protracted period of time each summer. A portion of this storage was accommodated by the areal expansion of the lakes and channels. The outflow of Heart Lake, normally 1.25 m wide, increased to a width of 34 m during the snowmelt period. Figure 3.13 shows a profile of the Heart Lake outlet on three different dates in 1983. Overbank discharge occurred from June 12 to July 2, 1983, providing direct interaction between the lake and channel discharge and the land portion of the basin. Heart Lake itself also flooded during the snowmelt period. Maximum areal extent of the lake was reached at the time of peak discharge. At this time the surface area of the lake had doubled (Figure 3.13).

During the summer, the Heart Lake outflow never reached a bankfull stage. Peak discharge observed during the summer



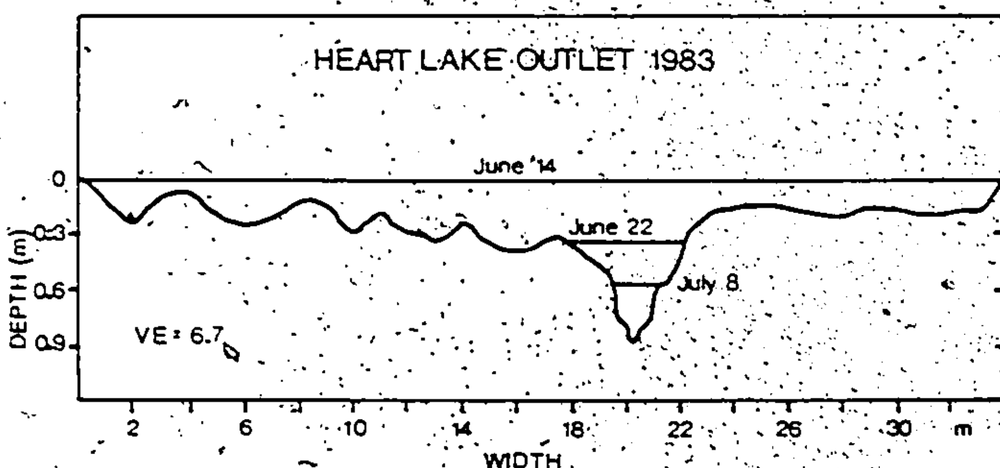


Figure 3.13 Expansion of Heart Lake (top diagram) and width of the Heart Lake outlet (lower diagram) during the 1983 snowmelt runoff period. Discharge at the time of width measurements were: June 14, 2.958 m^3s^{-1}; June 22, 0.778 m^3s^{-1}; and July 8, 0.109 m^3s^{-1}.

storm of July 20 and 21, 1982 was about half that required to initiate overbank flow. It should be noted, however, that the outflow channel at Heart Lake was particularly well defined compared with many other lake outlets in the region.

In 1983 lake levels were continuously recorded for Heart and Lost Lake. On July 9 the Lost Lake and July 21 the Heart Lake water level dropped below the premelt level (Figure 3.14). There was no substantial difference in the surface area of these lakes between June 9 prior to the snowmelt flood and August 10, so the total loss of stored water can be estimated directly as the water level difference between these two times. The average net loss of water from lakes was therefore 115 mm during the 1983 field season.

3.5.0 Summary

Basin input is dominated by snowmelt water. Rainfall can be important over short periods but the magnitude is variable from year to year. Discharge from Hardill Lake was the largest source of input. Evaporation and streamflow were water losses. Evaporation was large from lakes and wetlands and less from the upland areas. Discharge was greatest during snowmelt. Meltwater entered into lake storage during the spring period and was gradually released over the summer thus sustaining a slowly declining streamflow. Discharge response to rainfall can also be immediate and large as was

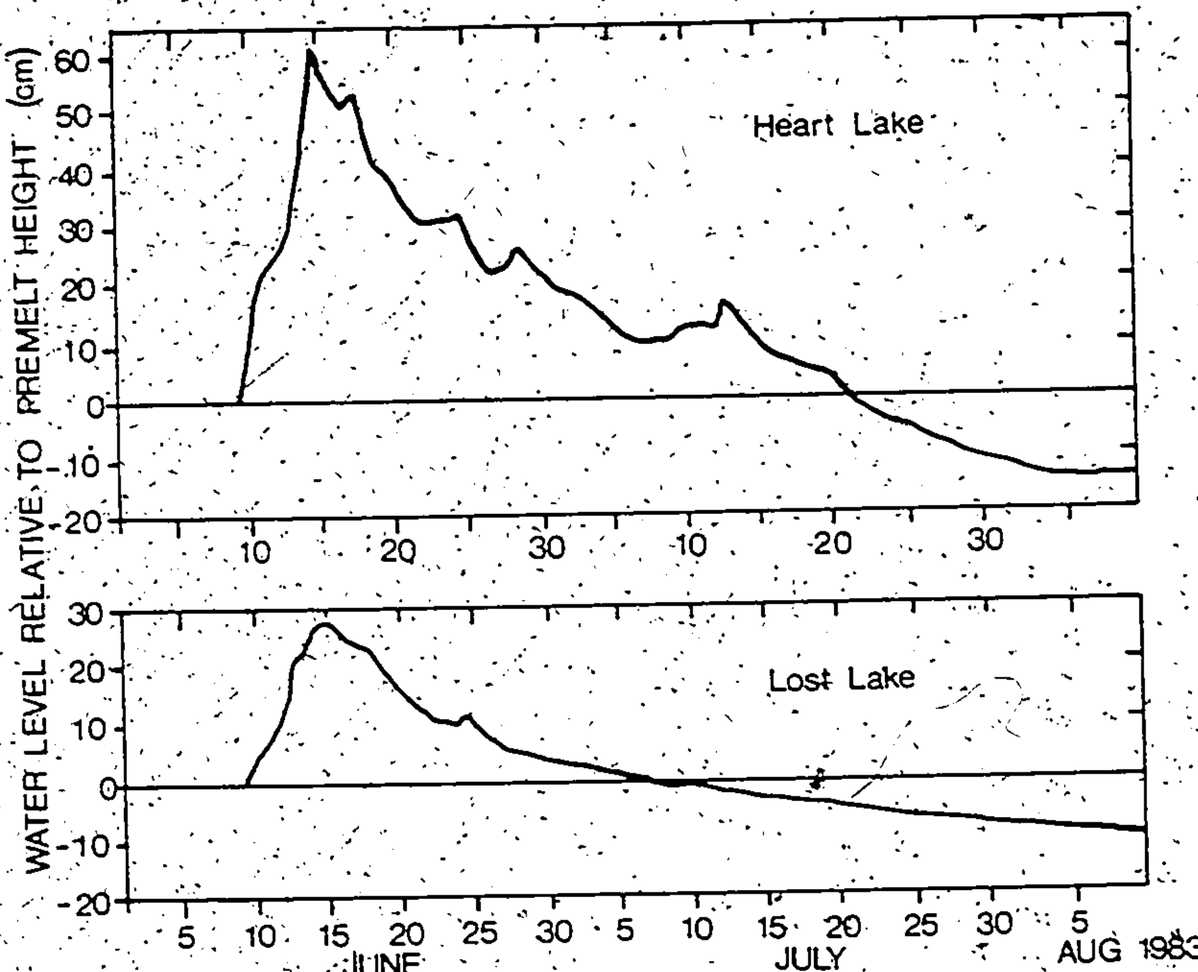


Figure 3.14 Heart and Lost Lake water level changes during the 1983 field season relative to their premelt water level.

observed during one summer rainstorm in the 1982 field season.

CHAPTER FOUR UPLAND HYDROLOGICAL SYSTEM

Over one half of the study basin is classified as upland (Section 2.1.0), comprising rock outcrops, ridges and slopes. The dominant type of upland consists of partially vegetated slopes that flank the basin. Storage and runoff from this large area influences basin hydrology. In this chapter, the spring and summer hydrology of an upland hillslope will be examined. The literature available on northern hillslope hydrology will be briefly reviewed, followed by a discussion on the hillslope soil structure and active layer thaw, storage and flow of water on a slope and a slope's water balance. The chapter concludes with a discussion on the upland hydrological system.

4.1.0 Previous Work

In the continuous permafrost region, snowmelt releases a large amount of water when the frost table is close to the ground surface, hence storage is limited. As a result, considerable surface flow is observed during the melt season (Lewkowicz and French, 1982a; Woo and Steer, 1982). Surface flow occurs both as overland or sheet flow and as rill flow (Woo and Steer, 1982). Woo and Heron (1981) found that when a basal ice layer formed at the bottom of a melting snowpack, the soil pores are sealed with ice, thus eliminating infiltration.

During summer when the soil storage capacity increases through ground thaw, surface flow occurs only in response to large rain storms. Summer surface flows are generated as return flow, occurring at locations where the water table intersects the ground surface (Lewkowicz and French, 1982a; Woo and Steer, 1982). Dingman (1975) suggested that return flow at the base of slopes strongly affected the runoff response of a central Alaskan drainage basin.

Subsurface flow becomes dominant as the depth of the thawed layer increases, but flow decreases through summer as sources of water are depleted. Subsurface flow response to rain storms depends on the depth of the frost table and the size of the input (Woo and Steer, 1982). For an entire runoff season, the subsurface discharge from a high Arctic slope was determined by Steer (1982) to be 28 percent of total surface discharge, but Lewkowicz and French (1982b) showed that it could vary between 23 and 83 percent from one year to the next. They concluded that differences in runoff pattern depended on initial snow distribution and climatological conditions during melt (Lewkowicz and French, 1982a).

Woo and Steer (1982) showed that subsurface flow can emerge downslope as surface flow during rainstorms but also occasionally when no input occurs. They hypothesized that suprapermafrost ground water may be ponded behind subsurface frozen barriers produced by differential ground thaw and is released in surges when these barriers are melted out. Another important result of their work was that the subsurface hydrological divide was different than the surface topographical divide so that some water was contributed by the areas adjacent to the hillslope concerned.

In the discontinuous permafrost region, subsurface discharge can be more important than surface discharge. Wright (1981) observed no surface flow on a subarctic hillslope in northern Quebec. In the same region Lewis (1977) found that subsurface flow was concentrated in depressions where there were discontinuities in soil hydraulic conductivity. He believed that increased subsurface flow produced greater thaw depths by increasing the lateral flux of heat. In Alaska, Santeford (1979) did not observe any runoff response from a moss filled lysimeter to rainstorms until the soil storage capacity was exceeded. Chacho and Bredthauer (1983), however, observed an immediate response from a similar, but sloping soil filled lysimeter. The last two studies demonstrate that subsurface storm response is a function of both the soil conductivity and the slope of the soil complex.

4.2.0 Soil Characteristics and Active Layer Thaw

The upland hillslope used in this study is typical of much of the sloping ground in central Keewatin described by Zoltai and Johnson (1978). The slope vegetation cover is approximately 0.03 m thick, overlying a humus organic cryosol. On average the cryosol is 0.10 m thick but can range from 0.02 m to 0.28 m. Beneath this organic layer and in areas where there is no vegetation cover, there is a mineral soil layer composed of till, sand, silts and clays. On the upper slope, bare patches occupy approximately 40 percent of the surface area. On the lower slope there are organic soil hummocks. The organic layer in the hummock zone ranges from 0.20 to 0.30 m in thickness. The physical properties and hydrological parameters for the slope organic and mineral soils are shown in Table 4.1.

The distribution of vegetation is closely related to moisture conditions. On the upper slope prominent lichens are *Alectoria ochroleuca*, *Cetraria cucullata* and *nivalis*, and several species of both *Cladina* and *Cladonia*. Few shrubs are evident except in areas sheltered by rock outcrops where *Vaccinium* spp., *Andromeda polifolia* and *Ledum palustre spp. decumbens* are found. Near the base of the slope in the hummocky area, *Carex* spp. and *Eriophorum* ssp. prevail. *Sphagnum rubellum* abounds in the extremely wet interhummock areas.

The hillslope soil temperatures and frost table

Table 4.1 Upland Hillslope Soil Characteristics

Soil Type	Depth (m)	Density (kgm^{-3})	Porosity	Specific Yield (%)	Specific Retention (%)	Hydraulic Conductivity (md^{-1})
Organic	0-0.10	536	0.61	20.0±6.1	50.6±17.8	6.391±1.558
Mineral	0.10+	1682	0.18	2.0±2.2	16.4±7.2	1.112±1.159

changes for the 1982 and 1983 field seasons are shown in Figure 4.1. No detailed soil temperature measurements were made in 1982. In 1983, soil temperatures in the organic layer (0-0.10m) increased rapidly during the snowmelt period. After snowmelt, the soil warmed slowly for the next ten days and then there was a steady increase in temperatures until July 20 to 25. After this the soil began to cool.

In 1983 the frost table dropped gradually with time. Maximum frost table depth was 1.12 m. In 1982, ground thaw began earlier. From May 22 to 28, the ground thaw penetrated the top few centimeters of soil, but the soil refroze in early June. Uninterrupted thaw began on June 5. Maximum observed frost table depth was 1.09 m.

Both the soil structure and the depth of thaw affect the storage capacity of the hillslope, as is shown in the section below.

4.3.0 Water Storage and Flow on a Upland Hillslope

Water movement on the hillslope began immediately after the onset of snowmelt in 1983. The shallow snowcover ripened quickly. Portions of the ground surface were exposed within several days and soil thawing was initiated. Meltwater could easily infiltrate the organic soil cover, but the frozen mineral soil had a measured infiltration capacity of zero. The infiltration capacity of the frozen organic

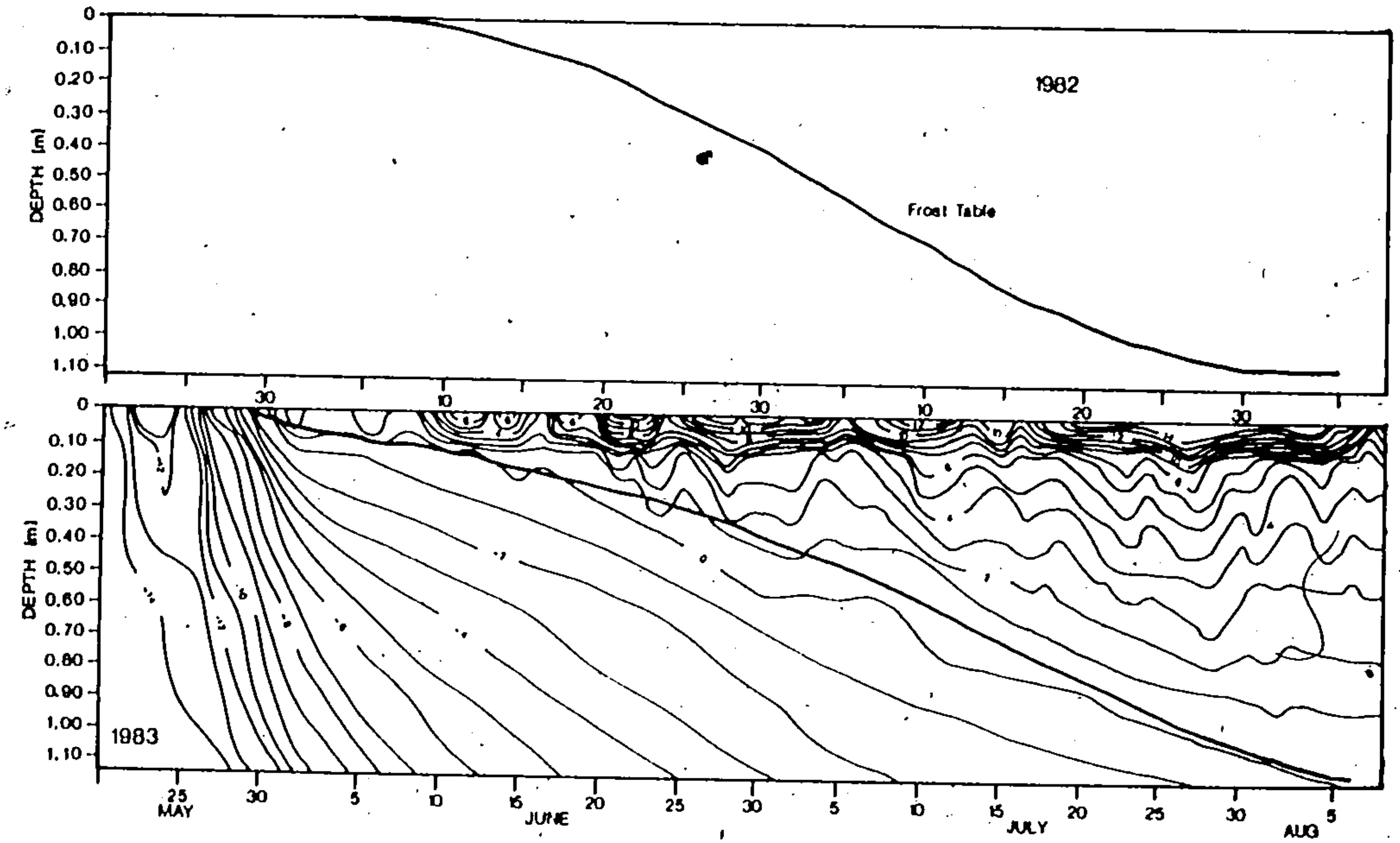


Figure 4.1 Upland soil temperature and frost table depth during the 1982 and 1983. Isotherm interval is 1.0 °C.

layer was so high that a constant head could not be maintained during infiltration experiments.

Surface discharge at the base of the slope began on May 31. Water filled the area behind the weir for two days, but no flow passed through the V-notch. Based on water storage changes behind the weir, surface flow was estimated to reach a maximum of approximately 0.36 l s^{-1} during these two days (Figure 4.2). The catch width of the surface flow weir was 7 m. Flow ceased until June 5. A peak surface flow of 0.856 ls^{-1} occurred on June 12 at 10:00 hours (CST). Flows were also large on June 8,(0.464 l s^{-1}), and on June 16, (0.557 l s^{-1}). Discharge throughout the entire melt period showed a high diurnal variability. Initially runoff was supplied by melting snow next to the weir, but several days later the surface flow was supplied from upslope. This produced a delay in flow from June 1 to June 5. By the end of the surface flow period, only a late lying snowpack at the lee of a rock outcrop furnished water to the slope.

The nature of the surface flow was similar to that described by Woo and Steer (1982). Water was temporarily stored in surface depression created by the vegetation growth around patches of exposed mineral soil. Woo (personal communication) measured this depression storage to be 6.0 mm. Water moved downslope from depression to depression by small rills or percolated through the vegetation and the upper organic soil layer. No surface flow was observed

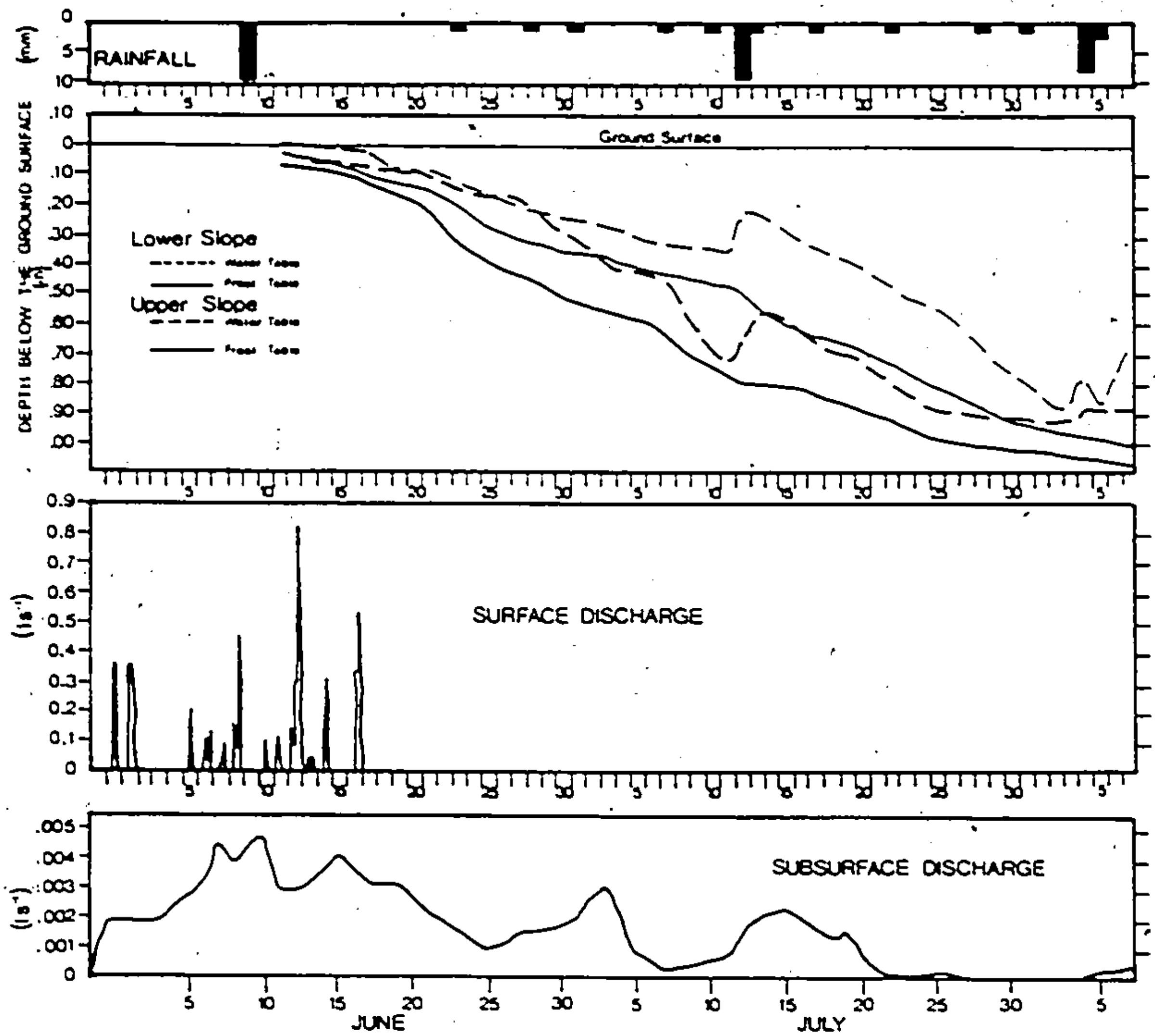


Figure 4.2 Upland ground water levels, frost table depth and surface and subsurface discharge during the 1983.

on the vegetated soil except at the base of the slope.

Ground thaw and subsurface flow began on May 29, 1983. Subsurface discharge was calculated using the same cross sectional width as the surface flow collector. Subsurface flow increased until June 10, reaching 4.84 x 10^{-3} l s^{-1} which is two orders of magnitude less than peak surface flow (Figure 4.2). At this time the zone of saturation was almost totally within the organic layer. Despite a thin saturated thickness compared with later in the summer period when the water table fell below the organic/mineral soil interface, the greater hydraulic conductivity of the upper layer produced larger flows. Subsurface flow decreased throughout the summer with occasional small increases. The second and third of these increases resulted from rainfalls, but the first increase did not correspond with any observable external input of water. This could have been caused by a release of ponded ground water due to differential ground thaw as suggested by Woo and Steer (1983).

As summer progressed the soil storage capacity of the hillslope increased. The water table dropped below the soil surface on the upper slope before measurements began on June 11. This happened on the lower slope on June 15 (Figure 4.2). At the base of the slope the water table remained above the soil surface within the interhummock depressions for an additional week. In addition to snowmelt and rainfall, the melting of ground ice supplied a large portion of the stored ground water. Assuming that the mineral soil was initially saturated with ice, the conversion of ice to groundwater would be approximately 2.4 mm d^{-1}. The water table dropped on the hillslope because of increased thaw depth and loss of water by evaporation and subsurface flow. Soil moisture decreased over time and was consistently greater on the lower slope and in the organic soil layer (Figure 4.3).

The only water table rise in 1983 resulted from the storms of July 12 and August 4. Subsurface flow increased but no surface flow was generated. Moisture content in the organic soil layer of the lower slope increased. In 1982, McMillan (1983) monitored the ground water level at many locations on the same hillslope. He recorded increased water levels during the storms of July 14-16, 20 and 28 (Figure 4.4). The upper slope water level rose to the ground surface while the water level on the lower slope rose well above the ground surface. Widespread surface flow occurred on the lower slope, particularly in the hummocky area. On the upper slope, ponding occurred in the mineral soil depressions where the water table intersected the ground surface. Even though the mineral soil has a low infiltration capacity compared with the organic soil (Figure 4.5) rainfall intensity was never great enough to produce rainfall excess. However, there was surface flow in places, but no measurement of its magnitude was attempted (McMillan,

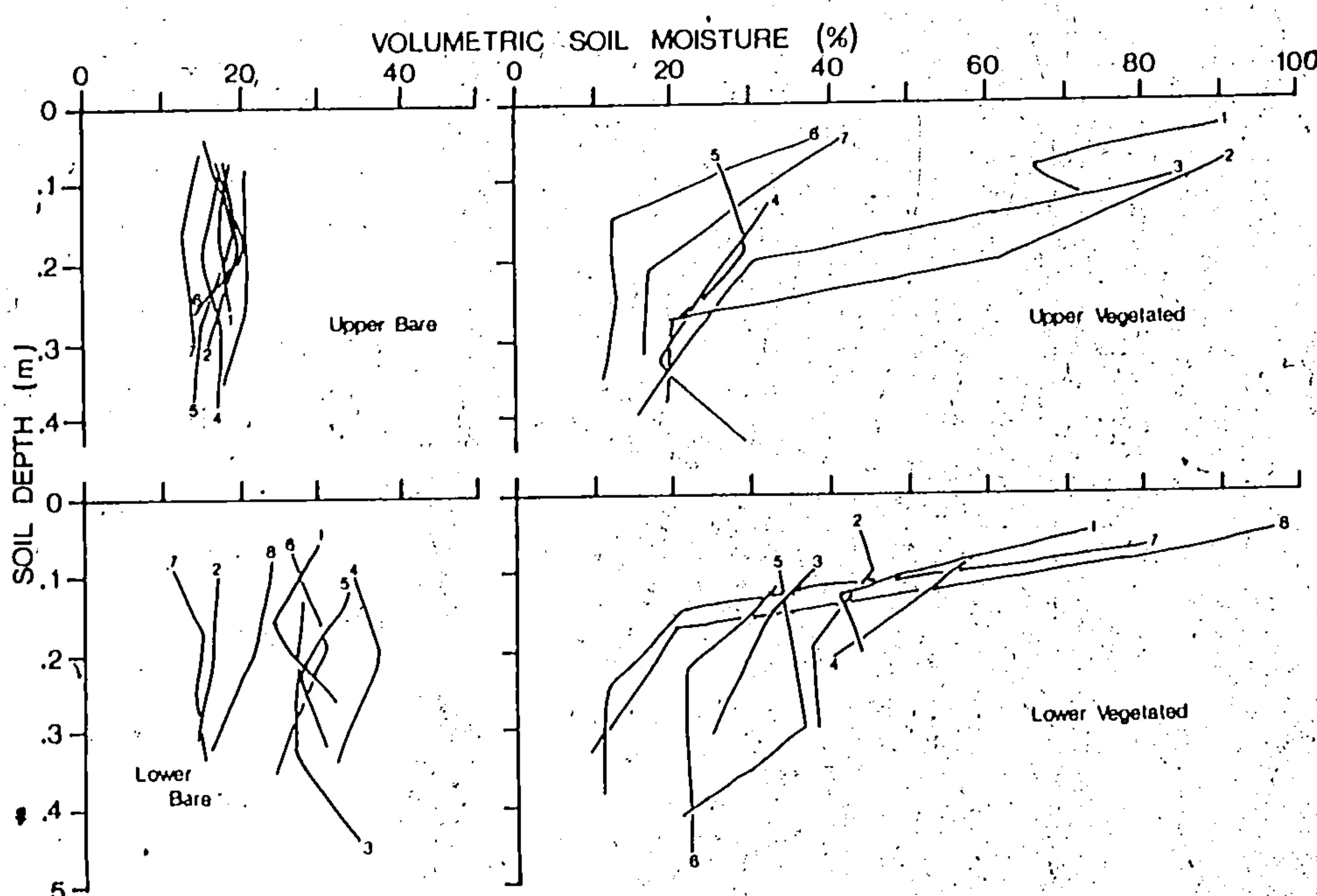


Figure 4.3 Upland soil moisture during the 1983. Numbers indicate dates: 1. June 21, 2. June 29, 3. July 6, 4. July 13, 5. July 20, 6. July 27, 7. August 4 and 8. August 6.

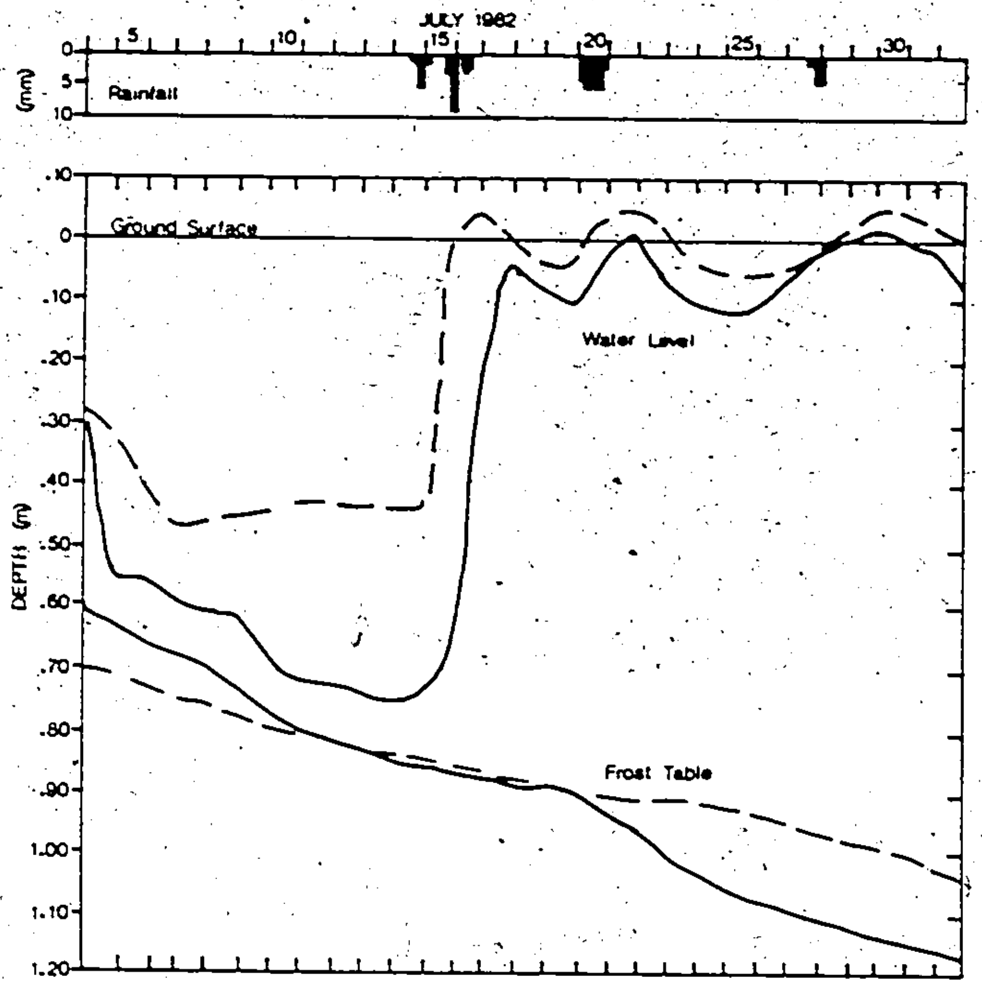


Figure 4.4 Upland ground water level and frost table depth for the 1982 field season. Dashed line represents lower slope site; solid line represents the upper slope site, Data obtained from McMillan (1983).

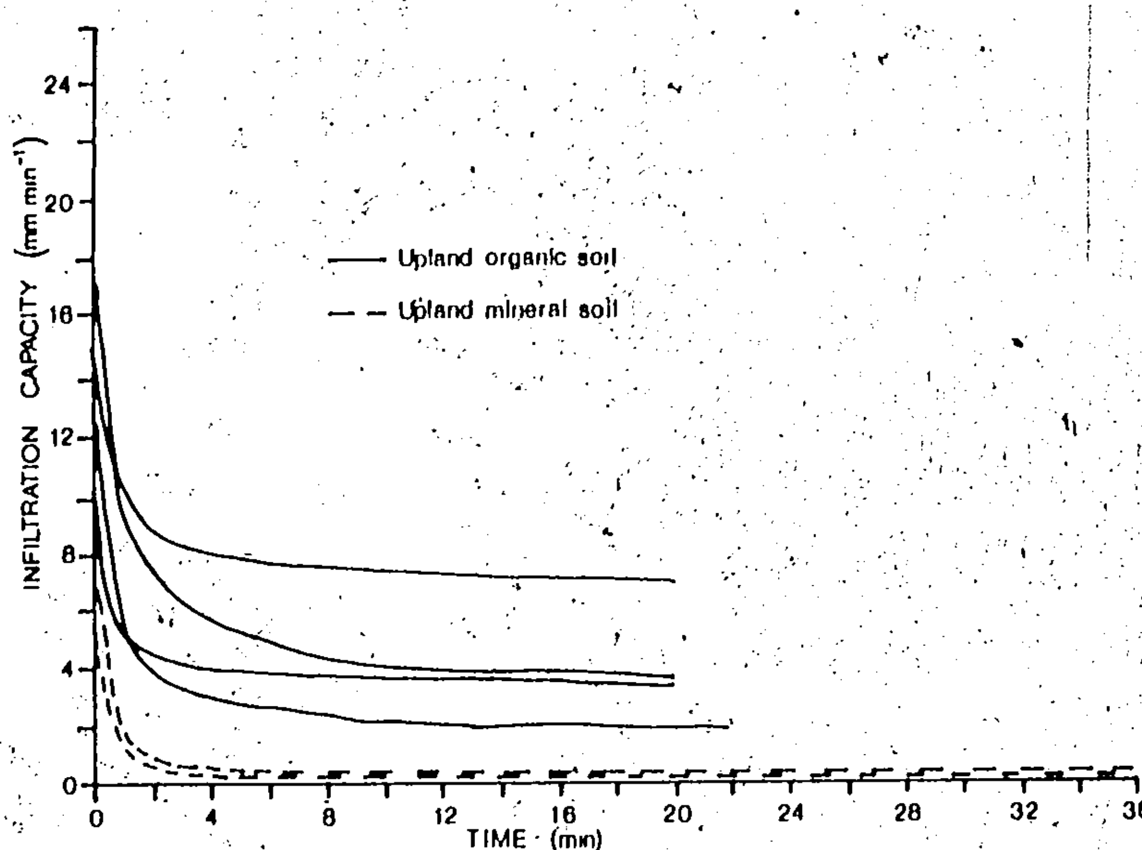


Figure 4.5 Infiltration capacity for the upland surface soil types.

personal communication).

Since the lower slope provided runoff during rain events before other areas, the water level in a small stream and adjacent hummock region below the slope west of Heart Lake were monitored from July 20 to August 5, 1983, to observe storm response. Ground water level between hummocks adjacent to the stream was about 0.10 - 0.15 m below the surface on July 20 and decreased to between 0.20 and 0.39 m by August 3 (Figure 4.6). Streamflow continued until July 28. The rainstorm of August 4 and 5 produced a water table rise of 0.20 m and streamflow began again. In contrast to the ground water levels further up the hillslope (see Figure 4.2 for example), the hummocky zone ground water system was very responsive. The importance this region plays in basin storm response will be discussed in Chapter Six.

4.4.0 Upland Hillslope Water Balance

To evaluate the relative importance of the various components of the upland hydrological system, a water balance was computed from May 1 to August 7, 1983, using:

$$m + r + i - e - q_s - q_{ss} = \Delta S/\Delta t \qquad (4.1)$$

where m and r are inputs by snowmelt and rainfall, i is the contribution of water from ground ice and e, q_s, and q_{ss} are losses of water by evaporation, surface discharge and

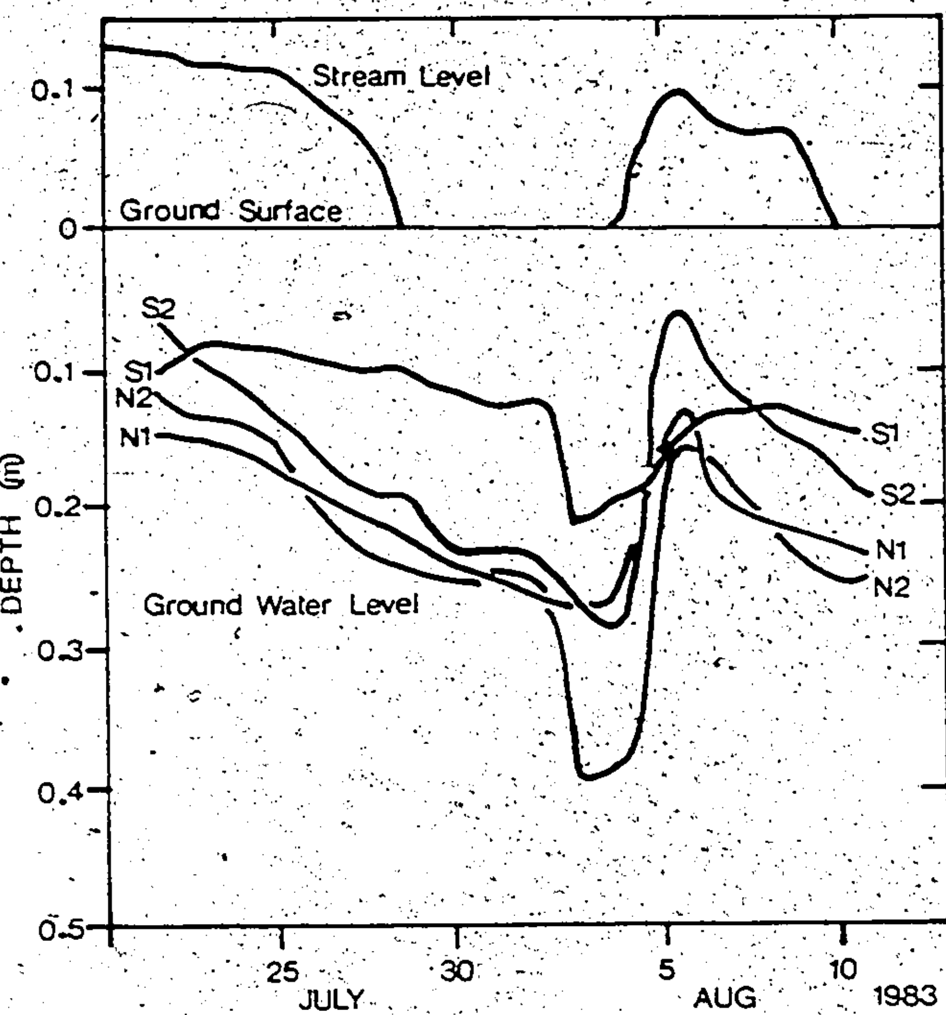


Figure 4.6 Hummock area ground water and stream level from mid July to early August, 1983. N and S indicate north and south; 1 indicates wells closest to the stream.

subsurface discharge respectively. $\Delta S/\Delta t$ is the change in storage over a given time period.

Over the study period the largest source of water to the hillslope was melting snow. Rainfall was not a large input in 1983, but its magnitude in 1982 indicate that it could be much greater. Evaporation was the most significant loss of water, followed by discharge (Table 4.2, Figure 4.7).

On thawing, ground ice provided a large portion of the stored water. It was calculated assuming that the organic soil layer contained little ground ice (10 percent of porosity) and the mineral soil was ice saturated. McMillan (1983) found hoarfrost at the base of the hillslope snowcover and in the vegetation cover, indicating an upward vapour flux of soil water had occurred during the winter. This assumption of low organic soil ice content is further borne out by the low runoff yielded during the snowmelt period and the high infiltration capacity. From May 1 until surface discharge ceased, inputs totalled 119 mm and losses were -66 mm (the sum of surface and subsurface discharge and evaporation), so that 53 mm of water, the difference between inputs and ouputs, must have been stored. In the 0.10 m thick layer of organic cover 36 mm of water could be retained. The remaining 17 mm could easily be accommodated by space made available by the thawing of ground ice in the mineral soil. This calculation

Table 4.2 Upland Hillslope Water Balance (May 1-August 7, 1983)

Snowmelt Water	Rainfall	Ground Ice	Surface Flow	Subsurface Flow	Evaporation	Storage Change
107	39	183	40	4	137	148*

Note: All values are in mm. * Measured storage was 181 mm, Aug. 7, 1983.

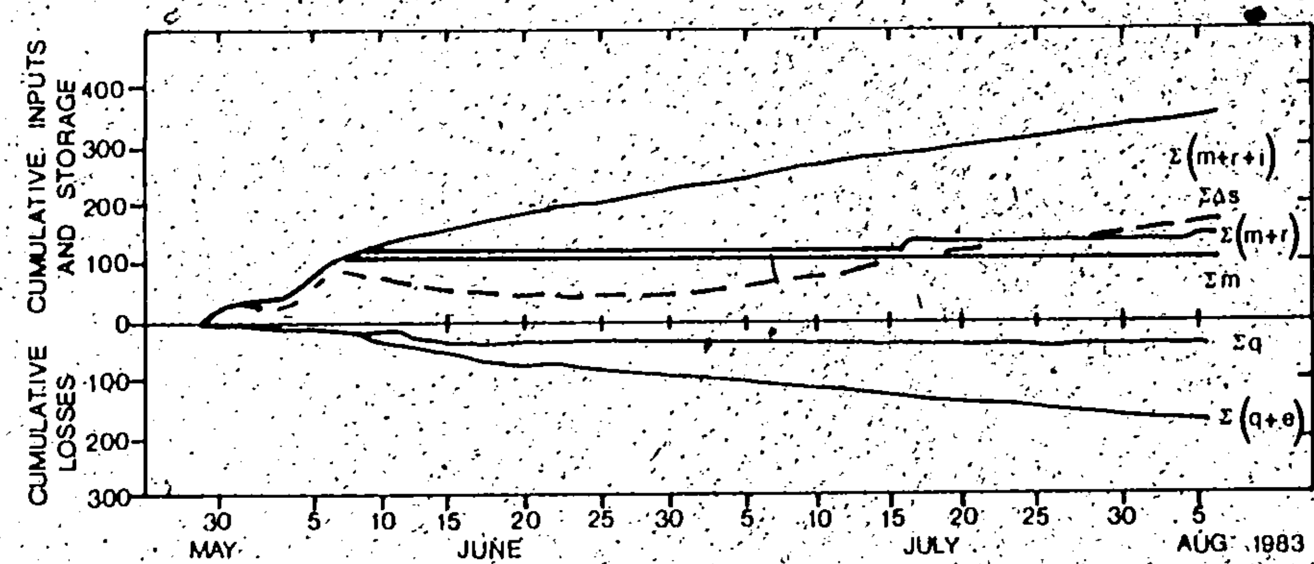


Figure 4.7 Upland cumulative gain and loss of water, 1983.

was weighted for a 60 percent slope cover of organic soil (see Section 2.2).

The measured stored water on August 7, 1983 was 181 mm. This yields a -33 mm disagreement between computed and measured storage (Table 4.2). Small negative storage changes are expected, but as the water table and soil moisture decrease so should the evaporation loss. However, in 1983, there were two long periods of persistently negative storage changes with no correspondly large drop in the loss of water through evaporation (Figure 4.8). Evaporation was calculated as a residual of the energy balance for the upland and extrapolated using the Priestley-Taylor model (Section 3.3.2). This extrapolation did not take into account the availability of soil moisture, so that evaporation could have been overestimated for the extended dry periods of 1983. To reconcile the computed storage with the measured storage, evaporation needs to be reduced to 104 mm. This also implies that the reported α for upland surfaces is too large.

4.5.0 Discussion of the Upland Hydrological System

Areally, upland hillslopes represent the dominant surface type in the study basin. The results presented above provide insight into the upland hydrological system. Some slopes in the basin are steeper with less vegetation while others are gentler than the study hillslope (for

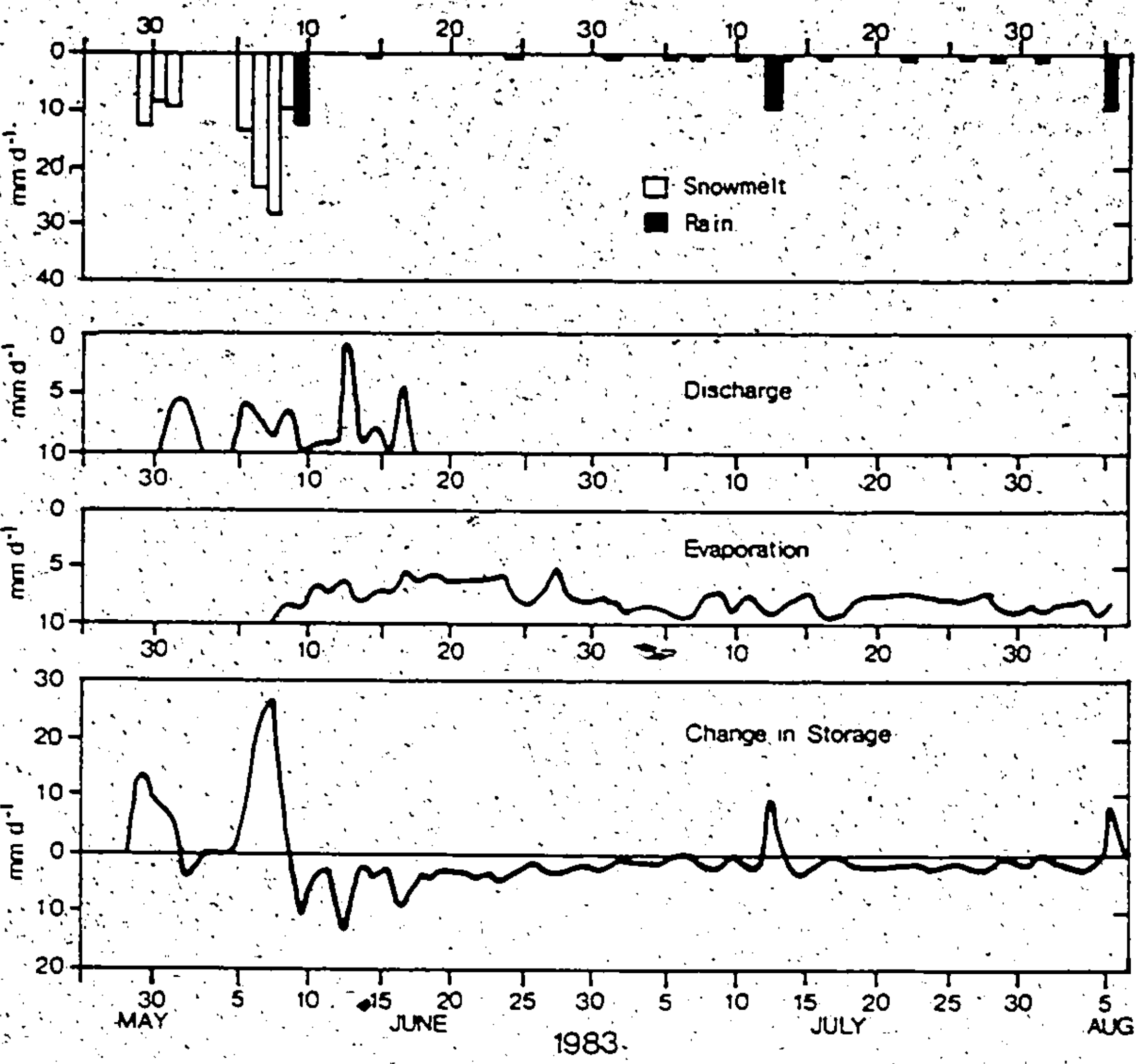


Figure 4.8 Upland inputs, outputs and changes in storage excluding ground ice input, 1983.

example the dry peatlands would fall into the latter category). By selecting a moderate hillslope an average condition is obtained.

The hillslope system became hydrologically active as soon as snowmelt began. Ground thaw started in areas that became snowfree on the first day of melt, allowing both surface and subsurface flow to play a role in snowmelt runoff. Surface flow was dominant and confined to inter-hummock regions on the lower slope and mineral soil depressions on the upper slope. Diurnal fluctuations in surface discharge was greater than that observed on high Arctic slopes (Lewkowicz and French, 1982a; Woo and Steer, 1982) because the nighttime negative heat flux in the low Arctic produced a prolonged daily stoppage in snowmelt. This was similar to Subarctic snowmelts (Price, 1975). A slightly greater percentage of total runoff occurred as surface flow than was reported by Steer (1982). Also, large quantities of water infiltrated and on freezing raised the temperature of the organic layer.

Surface and subsurface flow were intimately linked. The unfrozen soil on the lower slope remained saturated and surface flow continued after the slope snowcover had dissipated, sustained by return flow from upper slope subsurface flow. The peak in subsurface flow coincided with the rain on snow event of June 9 and emerged a day and a half later as peak surface flow down slope.

During the 1983 summer, slope discharge was sustained by subsurface flow. As the water table fell below the organic/mineral soil interface, subsurface flow decreased greatly. In early July, 1983, subsurface flow increased with no apparent external source of water. It was possibly due to a release of impounded subsurface water. Differences in saturated zone thickness between the upper and lower slope, before and after this date, tend to support this. Subsurface flow also increased slightly during the two summer rainstorms of the 1983 field season.

In 1983, subsurface flow represented 9 percent of the total hillslope discharge. It is not possible to extract a seasonal discharge for 1982 from McMillan (1983), but instantaneous subsurface discharge prior to July 16, 1982 were similar to those calculated for 1983. After the 1982 rainstorms the water table rose to the ground surface, but McMillan (1983) computed only small increases in subsurface flow. However, he did not include an upper organic layer in his analysis. Incorporating the organic soil layer, the July 17, 1982 subsurface discharge would be approximately 0.056 $m^3d^{-1}m^{-1}$ (0.392 m^3s^{-1} over the flow net width used in this study), which is similar in magnitude to snowmelt subsurface discharge. Assuming a slope length of between 200 and 250 m, this would amount to 0.25 mm d^{-1} runoff. Since in 1982 the lower slope remained saturated for a longer period, subsurface flow would have been substantially larger than that calculated for 1983. Similarly, surface flow would have been greater, but it is not possible to compute if the ratio of surface to subsurface flow varied between years.

The response of the hillslope system in terms of the timing and magnitude of stormflow was dictated by the moisture conditions of the hummocky area at the base of the slope. Observations showed that the hummocky area groundwater level rose dramatically in response to even a small rainstorm. The hydraulic conductivity of the interhummock organic soil must be sufficiently high to allow substantial flow or the stream would not have been filled with water as quickly. Several dynamic processes, related to the location and structure of the hummock area can explain both the generally wetter conditions and greater response. A combination of flow processes are envisaged (Figure 4.9). In early snowmelt, when no ground thaw has taken place, surface flow follows the lowest route between hummocks. When thaw penetrates the soil, the interhummock zone thaw depth is slightly greater (Figure 4.10) because of a higher thermal conductivity and subsurface flow begins while surface flow decreases. In summer, after water supply on the hillslope has diminished, the flow regime becomes entirely subsurface. The third flow regime can be enhanced or transformed into the second flow regime during rainstorms. The response is exaggerated in this zone because the hummock topography and underlying frost table contours are

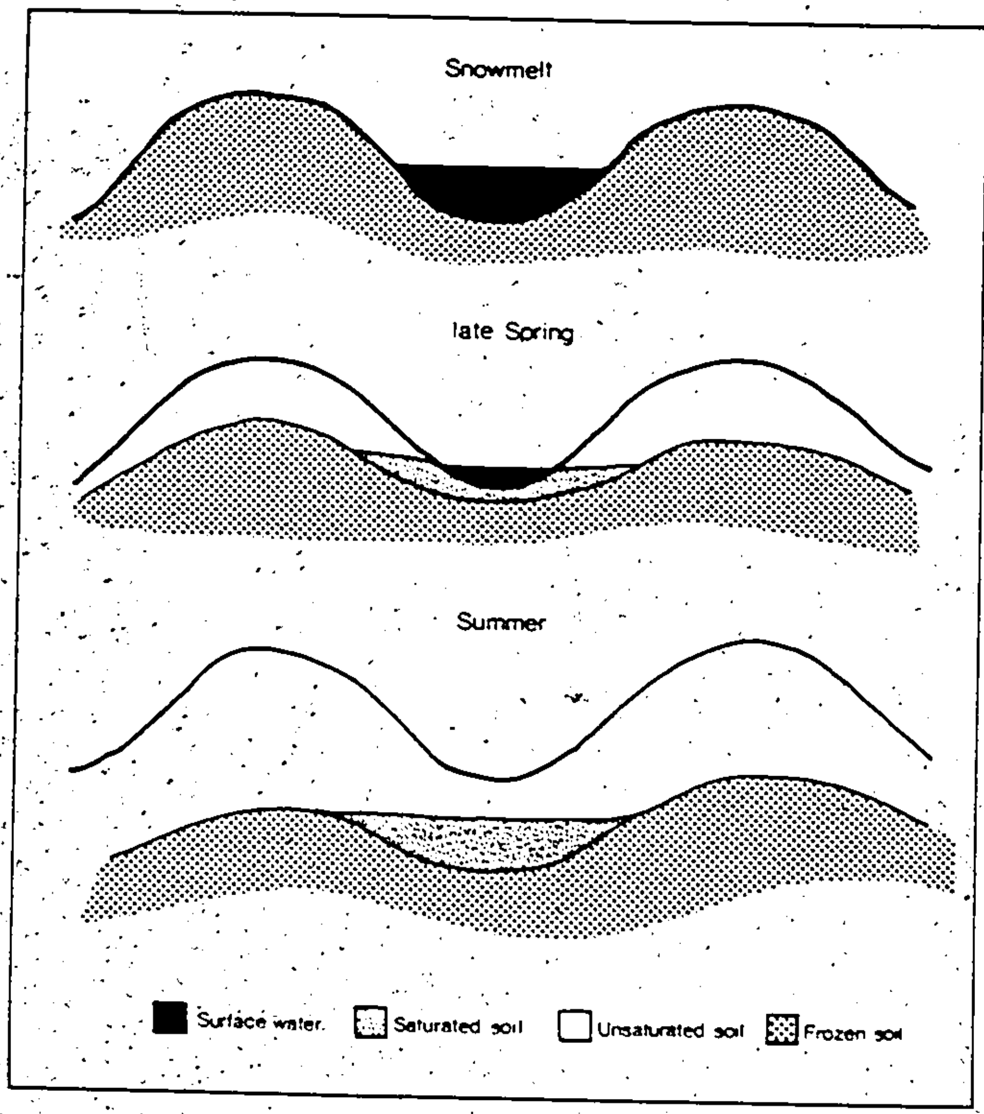


Figure 4.9 Diagramatic representation of water levels through the spring and summer between hummocks.

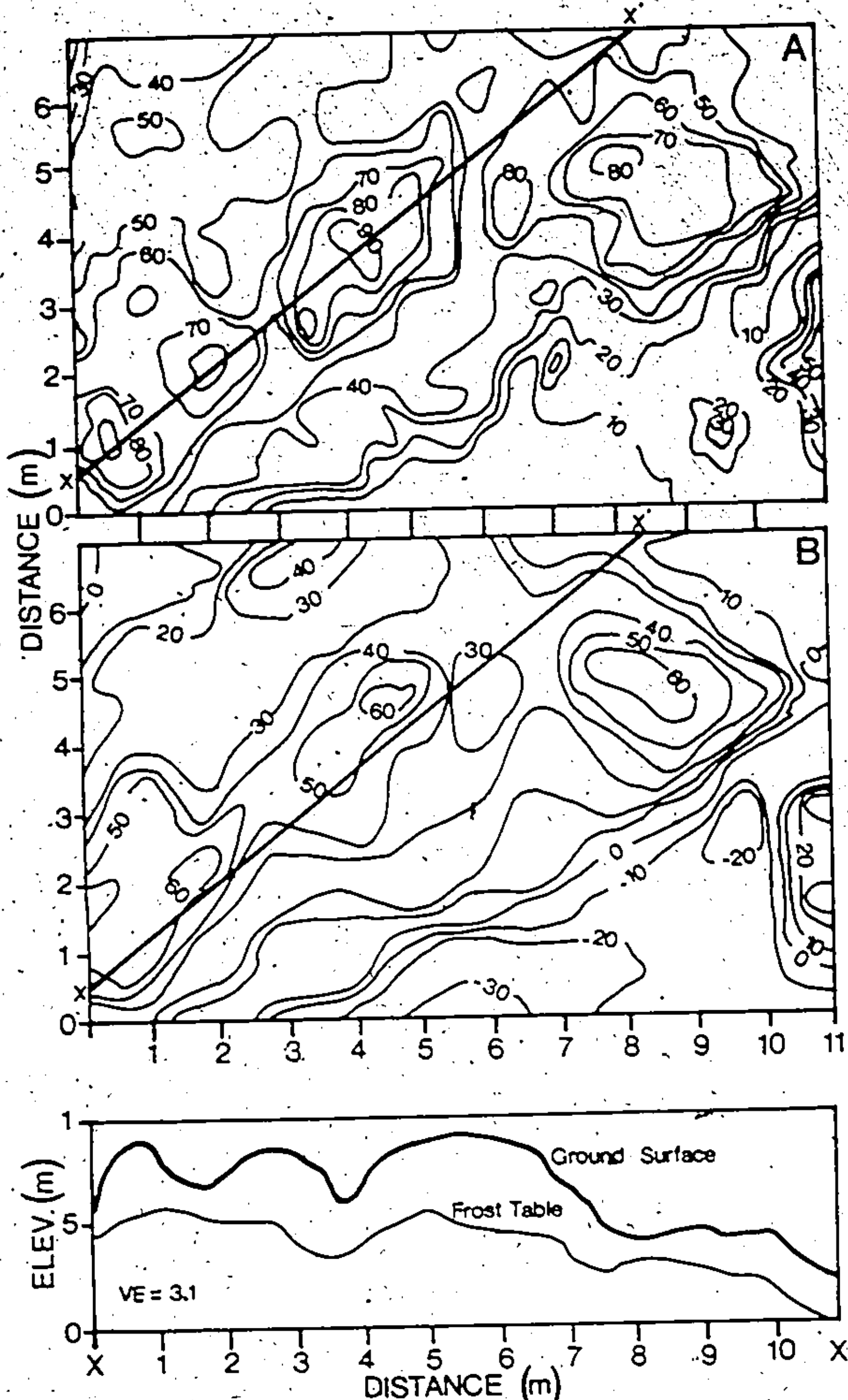


Figure 4.10 Areal view of the hummock area north of the intermittent stream west of Heart Lake, July 18, 1982. The upper diagram (A) is surface elevation. The second diagram (B) is frost table elevation. Contour interval is 10 cm. Elevations are above an arbitrary datum. The bottom diagram is a horizontal profile through X and X'. Data supplied by M.K. Woo.

structured such that they redirect flow into the confined interhummock depressions (Figure 4.10). The magnitude of the rain event and antecedent moisture conditions on the hillslope and hummock area will dictate if the water level will rise high enough to generate surface flow such as was seen during the 1982 field season.

When no snowmelt water or rain are added to the hillslope, the amount of water stored as soil moisture and groundwater is reduced by subsurface flow and evaporation. Close examination of the soil moisture diagrams shows that the unsaturated soil moisture values sometime exceed measured porosity (see Figure 4.3 and Table 4.1). Spatial variability of soil characteristics may produce this inconsistency. Because the gravimetric soil moisture technique is destructive the same soil sample can not be sampled repeatedly. In the case of organic soil it was difficult to distinguish the vegetation/soil interface when sampling near the ground surface. However, despite these inconsistencies, the data shows that the soil moisture decreases over time providing moisture to evaporation.

The dominant loss of water from the hillslope was by evaporation. Error in evaporation estimates in 1983 is largest when soil moisture is depleted and the water table is at a low level. Wright (1981) found the α parameter to be 0.13 lower than the value used in this study for 'dry' conditions on a similar slope. In other Arctic locations α

has been shown to be small in the range of soil moisture levels experienced in this study (Marsh et al, 1981).

In conclusion, the hillslope generates the bulk of its discharge during snowmelt as surface flow and to a lesser extent, subsurface flow. Storage is available in the organic soil layer at the time of melt. As the frost table increases in depth, surface flow ceases and subsurface flow and evaporation deplete the water stored. During large rainfall events the hillslope can yield water as subsurface and surface flow resulting from saturated conditions and return flow in the hummocky zone at the base of the slope. During low rainfalls, the lower slope hummocky area can yield some water, but not the rest of the hillslope. The upland storm response is governed by the size of the rain event and the antecedent moisture conditions on the hill-slope and in the hummocky zone.

CHAPTER FIVE WETLAND HYDROLOGICAL SYSTEM

The previous two chapters discussed lakes, channels and uplands, which represent approximately 75 percent of the basin area. The remaining portion of the basin area is occupied by flat peat covered fens which occur adjacent to the lakes, streams and below ridges that contain late lying snowcovers (Roulet and Woo, 1985b). The hydrological regime of these fens is different from that of the upland system because of the nature of water input from other systems and conditions within the wetland itself. In this way the wetlands, while being a small portion of the basin area, play a significant role in the hydrology of the basin. This point is pursued further in Chapter Six. In this chapter, after a brief review of northern wetland hydrology, the soil characteristics and active layer thaw, water flow and storage and the water balance of the largest fen in the study basin will be examined. It concludes with a discussion of the results in the context of northern and wetland hydrology.

5.1.0 Previous Work

There have been very few wetland hydrology studies undertaken in the northern permafrost region of North

America. Goode et al. (1977) reviewed some of the general aspects of muskeg hydrology in North America, but the major work on wetland hydrology has been done in Europe and in the Asian portion of the Soviet Union. This research is summarized by Dooge (1972), Ingram (1983), Ivanov (1975) and Romanov (1968a, 1968b). Of this work only a few studies from the Soviet Union apply to permafrost regions. Unfortunately this work is micro or regional in scale. Little work has been done at the mesoscale of a small basin.

In the North American high Arctic, Rydén (1977) found that snowmelt discharge from a small sedge wetland increased rapidly and large lateral flows were produced within the wetland itself. After the annual peak flow occurred, discharge receded gradually and ceased entirely during the latter portion of the summer even though the wetland soil remained near saturation. Ryden attributed this to the high soil moisture retention capabilities of peat. Soil moisture retention of 0.8 for various peat types is not uncommon (Boelter, 1965).

The magnitude and response of wetland runoff during snowmelt and summer rainstorms depends on a wetland's ability to store water. Landils and Gill (1972), working in the region of discontinuous permafrost near Yellowknife, N.W.T., found that the soil moisture condition of the previous autumn was the critical factor in determining the magnitude of spring snowmelt flows. When the peat was saturated prior to freeze-back, discharge the next spring was large. In summer, antecedent moisture conditions and rainfall quantity governed storm runoff from a small wetland (1.6 km^2) on the Arctic Alaskan coastal plain (Brown et al., 1968). Runoff ratios for individual storm events varied from 0.01 for small rainfalls and low antecedent moisture conditions to 0.70 for large rainfalls and saturated soil conditions. Storm runoff response lagged 7 to 11 hours behind initiation of rainfall. Brown et al (1968) attributed these delays to surface storage and the flatness of the wetland. Lag time and antecedent moisture conditions were not related.

Seasonal and annual discharge from wetlands in permafrost regions represents approximately 50 percent of the total precipitation (Brown et al., 1968; Ryden, 1977). This is about 20 percent lower than what discharges from non-wetland basins at the same latitude (Woo, 1983). Thus, in these latitudes wetlands must lose more water by evaporation than non-wetland basins. Rouse (1977) found that a subarctic sedge meadow lost water by evaporation at near potential rates. Actual evaporation from northern European fens is greater than regional estimates of potential evaporation (Ingram, 1983). Roulet and Woo (1985a) also found high evaporation losses from the same fen that is investigated in this chapter.

In the following sections the dynamic nature of the wetland flow and water storage system is examined. Through this study the importance of the individual components of the wetland hydrological system will be established.

5.2.0 Soil Characteristics and Active Layer Thaw

The soil at the wetland study site is comprised of three layers. The top layer is a 0.02 to 0.03 m thick mat of vegetation. Dominant plants are mosses and sedges (*Drepanocladus* spp. and *Carex* spp.). Beneath this mat is 0.22 m of wet hemic peat (see Kivinen, 1977 for classification), while the lower layer is composed of a sand, silt and clay material (Table 5.1).

The specific yield, specific retention and hydraulic conductivity of peat are highly variable, but there was no clear vertical variation with depth as reported by Boelter (1965). Specific yields range from 0.12 near the surface to 0.04 at the peat/mineral soil interface (Figure 5.1). The small vertical variation is due to the high degree of compaction of the peat. Peat hydraulic conductivity was similar to that reported by Irwin (1966) but is several orders of magnitude less than that measured by Boelter (1965) for a similar depth of peat. The mineral soil under the peat had a very low hydraulic conductivity and specific yield and large retention.

Wetland ground temperature for 1983 and frost table progression for the 1982 and 1983 field seasons are shown in

TABLE 5.1 Wetland Soil Charactersitcis

Soil Type	Depth (m)	Density (kgm^{-3})	Porosity	Specific Yield (%)	Specific Retention (%)	Hydraulic Conductivity (md^{-1})
Peat	0-0.25	189	0.87	5.7±3.1	81.2±16.0	3.21±1.40
Clay	0.25+	1456	0.37	2.9±2.6	33.7±16.0	0.31±0.30

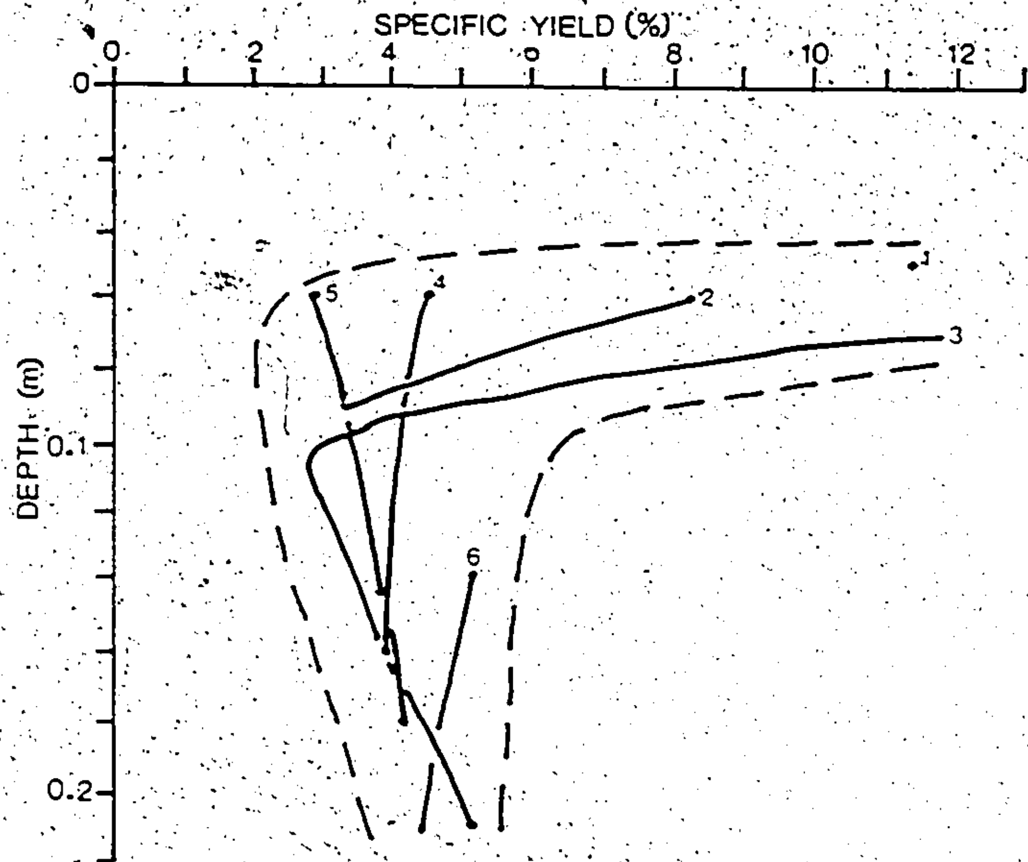


Figure 5.1 Vertical variation of specific yield in wetland peat. Dashed line indicates the general envelope of specific yield. Dates: 1. June 25, 2. July 3, 3. July 8, 4. July 14, 5. July 22. and 6. July 29, 1983.

Figure 5.2. In 1983, the temperature in the upper 0.10 m of soil rose rapidly between May 30 and June 1, from -12°C to -6°C. The same layer warmed more slowly over the next 9 days to 0°C. Temperature changes at depths greater than 0.20 m were comparatively conservative, even in the middle of the summer. By July 27, the maximum soil temperature had been reached and the soil began to cool in the first week of August.

The drop in the 1983 frost table followed the 0°C isotherm except for the first few days of ground thaw. This difference could easily be due to the temperature and frost table measurements being taken in slightly different locations. The average rate of decline of the frost table was slightly less than 0.01 md^{-1} in 1983. A similar pattern was observed in 1982. Maximum measured frost table depths were 0.49 m on August 5, 1982 and 0.53 m on August 8, 1983.

5.3.0 Water Storage and Flow in a Wetland

Surface discharge from the wetland began on June 9, 1983, 12 days after snowmelt commenced. This delay was due to the melt water being held in the snowcover (Figure 3.3) and the slow ripening of the snowcover (Figure 5.2). The snowcover broke up on June 10 and discharge peaked at 0.13 m^3d^{-1} on June 13 (Figure 5.3). The diurnal melt cycles were reflected in the daily discharge. The frost table was at the ground surface until June 11 and this restricted

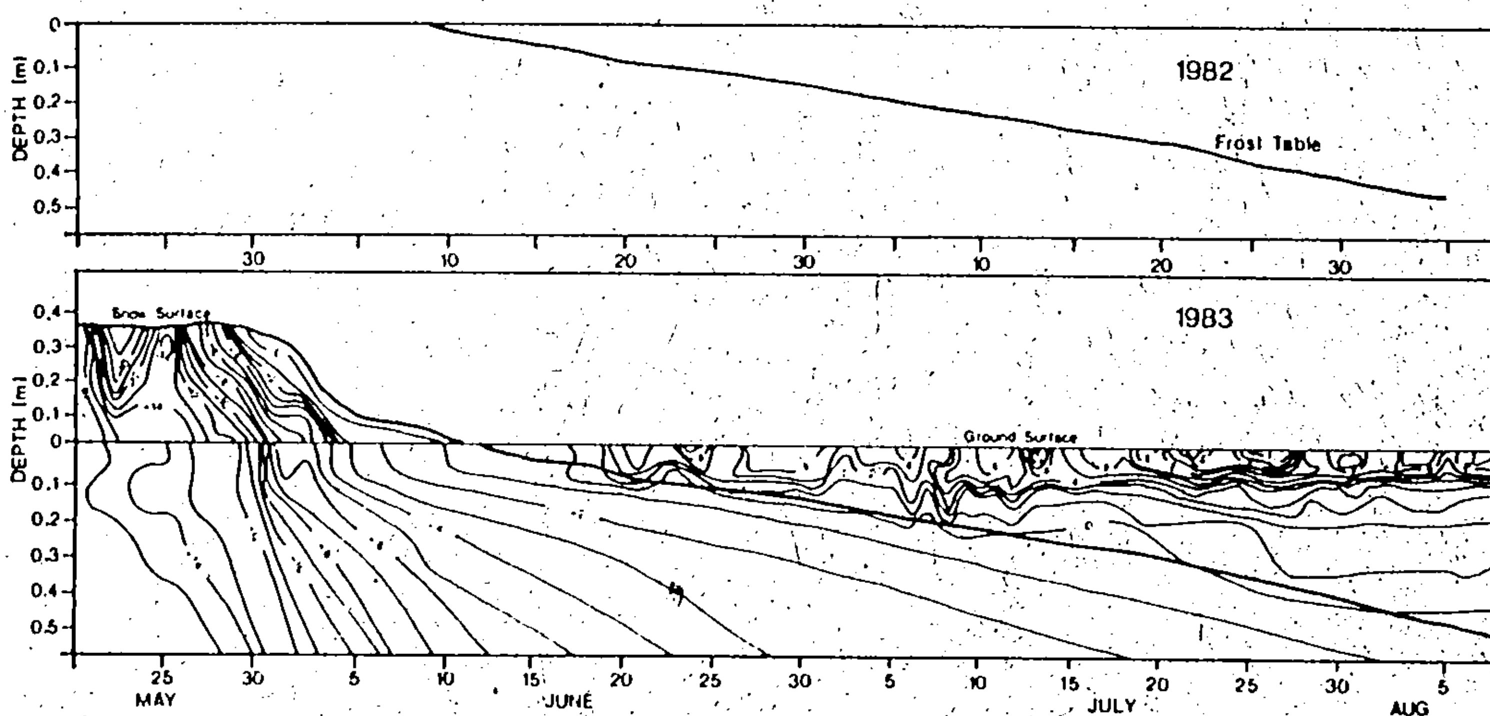


Figure 5.2 Wetland snow and soil temperatures and frost table depth, 1982 and 1983. Isotherm interval is 1 °C.

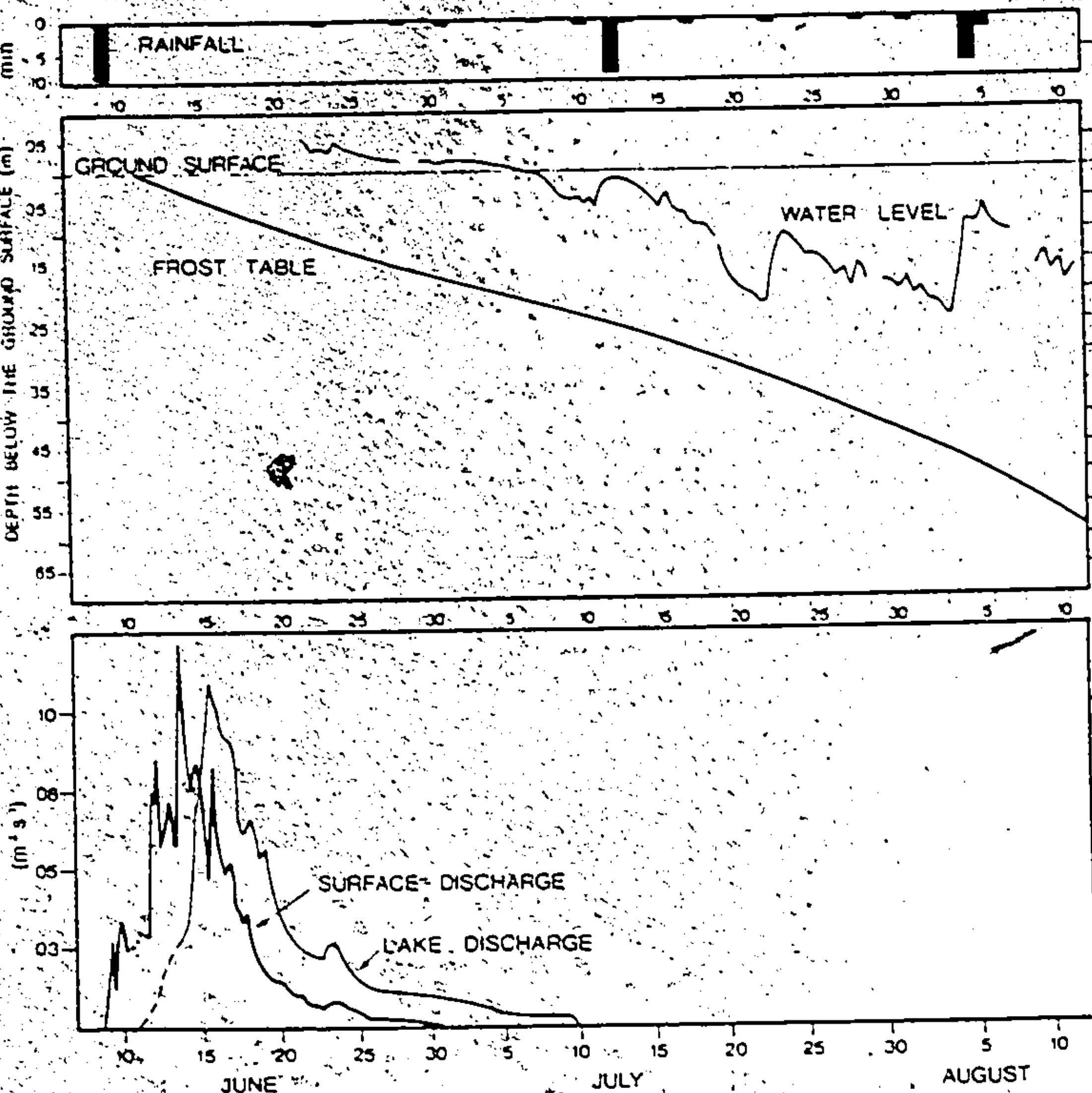


Figure 5.3 Wetland ground water level and frost table depth, surface discharge and Lost Lake discharge into the wetland, 1983. Dashed portion of the Lost Lake hydrograph is extrapolated.

infiltration.

Lost Lake discharge into the wetland which reached the wetland as overbank flow from the small pond, was added to the water already in the wetland. When lake discharge became significant on June 15, 1983, the wetland had begun to thaw, thus increasing its storage capacity. This enabled the surface outflow from the wetland to decline even though the inflow from the lake continued for 22 days after surface discharge peaked. With inflow sustained by the lake, water remained above the wetland surface long after the snow had disappeared. As the water level dropped toward the ground surface, surface flow became channelized in small rills and depressions created by the wetland microtopography (Figure 5.4). Surface discharge ceased entirely on July 1, 1983.

Figure 5.5 shows the relationship between daily mean water depth above the wetland surface calculated from measurements made at the sixteen groundwater wells and daily surface discharge. At water depths in excess 0.04 m, the relationship is linear with a slope of 34 $(m^3 d^{-1}) mm^{-1}$. Below 0.04 m, the slope decreases. Discharge ceased when the mean water depth was less than 0.024 m. This represents an approximation for surface depression storage. Woo (personal communication) estimated depression storage to be 0.032 m for an arbitrarily chosen plot in the study wetland. The difference between these values reflects the variation in wetland microtopography.

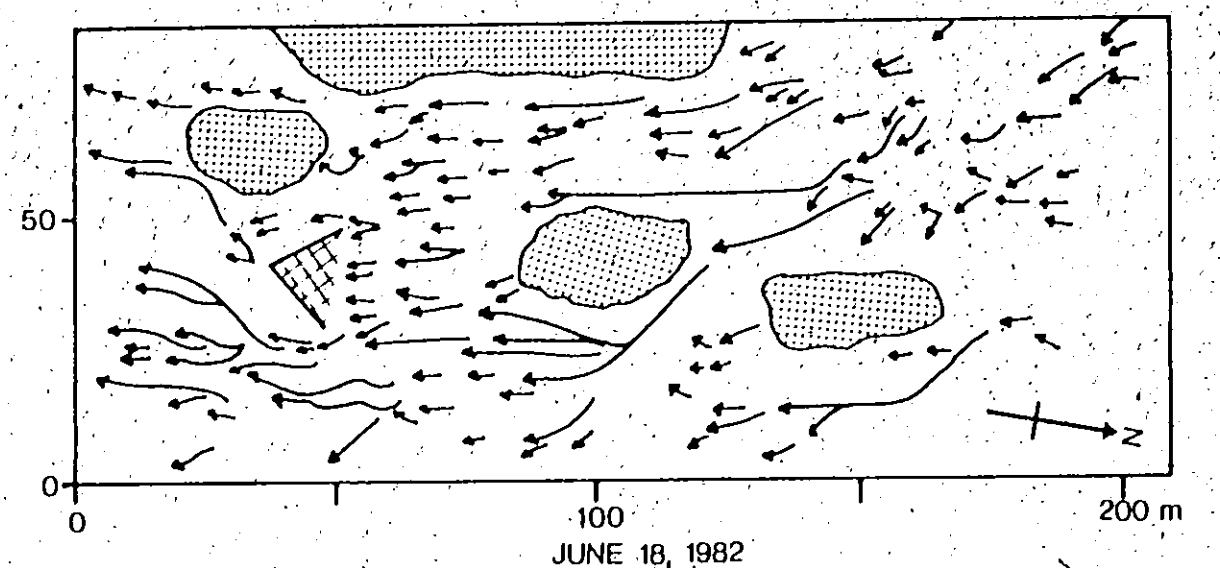


Figure 5.4 Wetland surface flow vectors, June 18, 1983. Scale is approximate. Arrows indicate flow paths. Shaded areas are raised grass tussocks and the cross hatch are areas flooded by the weir. Mapped by M.K. Woo and J. Drake, June, 1983.

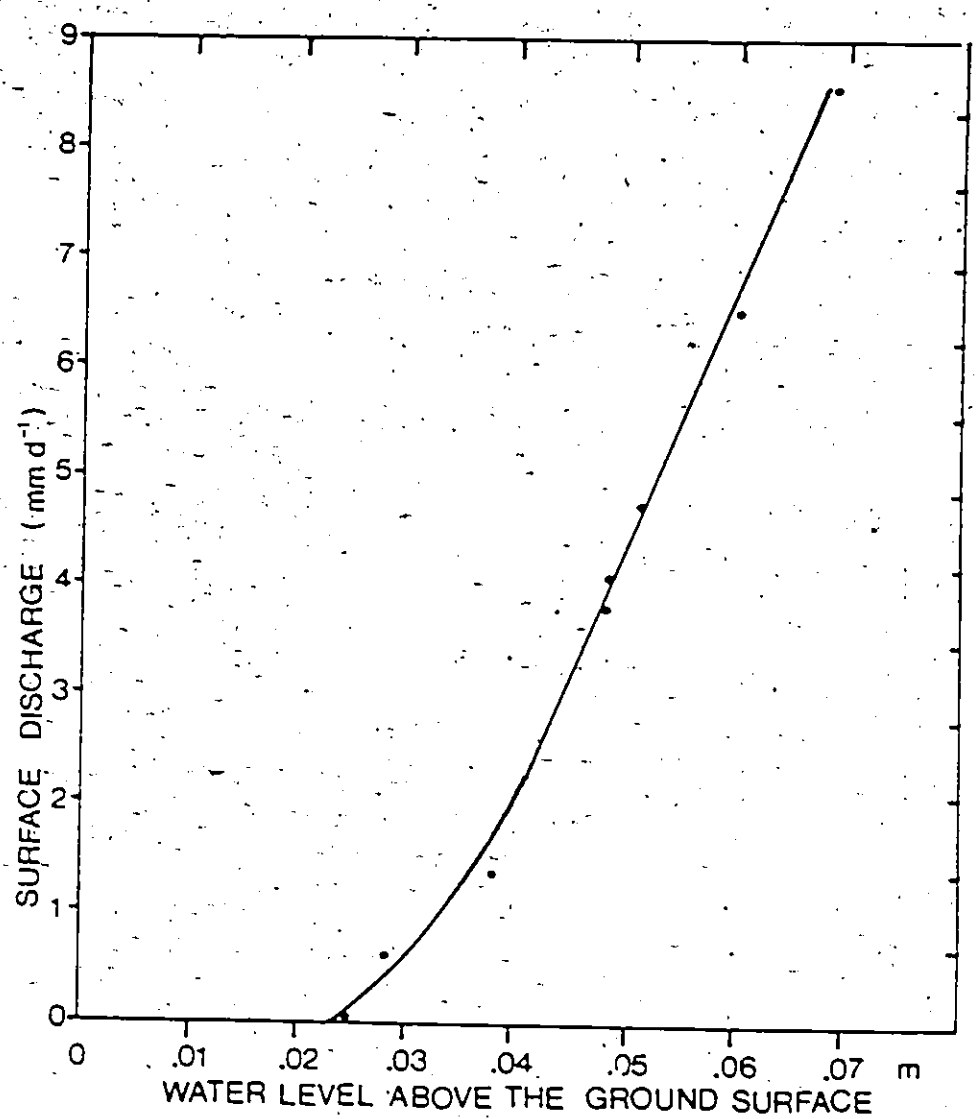


Figure 5.5 Wetland mean depth of water above the ground surface versus mean depth of daily wetland surface runoff, 1983. Line fitted by eye.

The wetland soil had a large specific retention (Table 5.1). If the soil was near saturation at the time of freeze-back a substantial amount of water would be released back to the wetland when the ground thaws. The ground ice content was not measured directly but experiments prior to snowmelt yielded zero infiltration to the frozen soil. Ground thaw began on June 11, 1983. Assuming that the wetland peat and mineral soils were saturated when frozen, ground ice thaw yielded between 3 to 7 mm d^{-1} of water equivalent, depending on the position of the frost table and the rate of thaw. Owing to a density difference between ice and water, 10 percent of the void volume in the newly thawed zone had to be filled by water. With a saturated ice content of 0.87 in peat, 0.7 to 0.9 mm d^{-1} of water is required to maintain the soil at saturation.

Calculated subsurface discharge ranged from zero in frozen ground to a maximum of 4 m^3d^{-1} based on a strip of wetland 300 m wide. Subsurface discharge was several orders of magnitude less than that of surface discharge. In fact subsurface flow for the 1983 study period yielded less water than two hours of surface discharge at peak stage. Two factors caused this low yield. The hydraulic gradient never exceeded 0.24 and the ground thaw was shallow, thus restricting the saturated zone thickness (Figure 5.6). The largest subsurface discharge occurred on July 12, 1983, when the frost table was within the mineral soil and the water

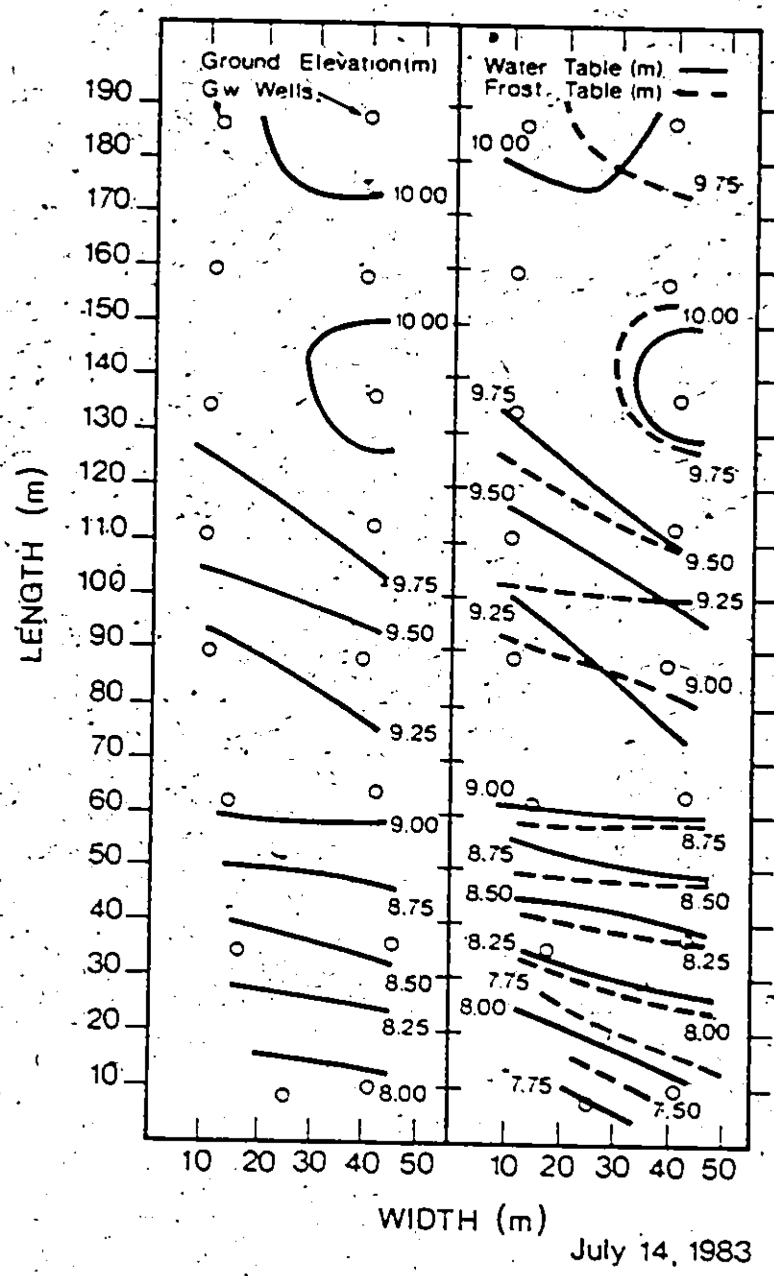


Figure 5.6 Wetland ground water flow net and frost table depth on July 14, 1983. Length and width refer to the area through which subsurface flow was calculated (see Figure 2.3)

table was just beneath the ground surface. Later in the summer the same depth of water produced a much lower discharge because the bulk of the saturated zone was in the mineral soil layer which has a much lower hydraulic conductivity (Figure 5.7).

The low subsurface discharge indicates that once the water table has fallen below the ground surface, discharge does not greatly reduce the soil moisture. However, field data show that for most of July and early August, a falling water table was accompanied by a reduction of soil moisture in the unsaturated zone (Figure 5.8). The decrease in moisture content from saturated conditions on July 8 to 70 percent volumetric contents on August 4 amounted to a moisture deficit of 50 mm. The difference between precipitation and wetland evaporation for the same period was 65 mm. This means the bulk of evaporated water was supplied from the unsaturated zone.

The high specific retention of peat kept the wetland near saturation during periods with no rain. Rainstorms caused an immediate rise in the water table. In the 1983 field season small rainstorms produced marked increases in water level (Figure 5.3). In the summer field season of 1982, the July 14-15 rainstorm (32 mm) raised the water table from 0.137 m below to 0.02 m above the ground surface (Figure 5.9). Water filled many rills and depressions and some localized lateral flow was discernible.

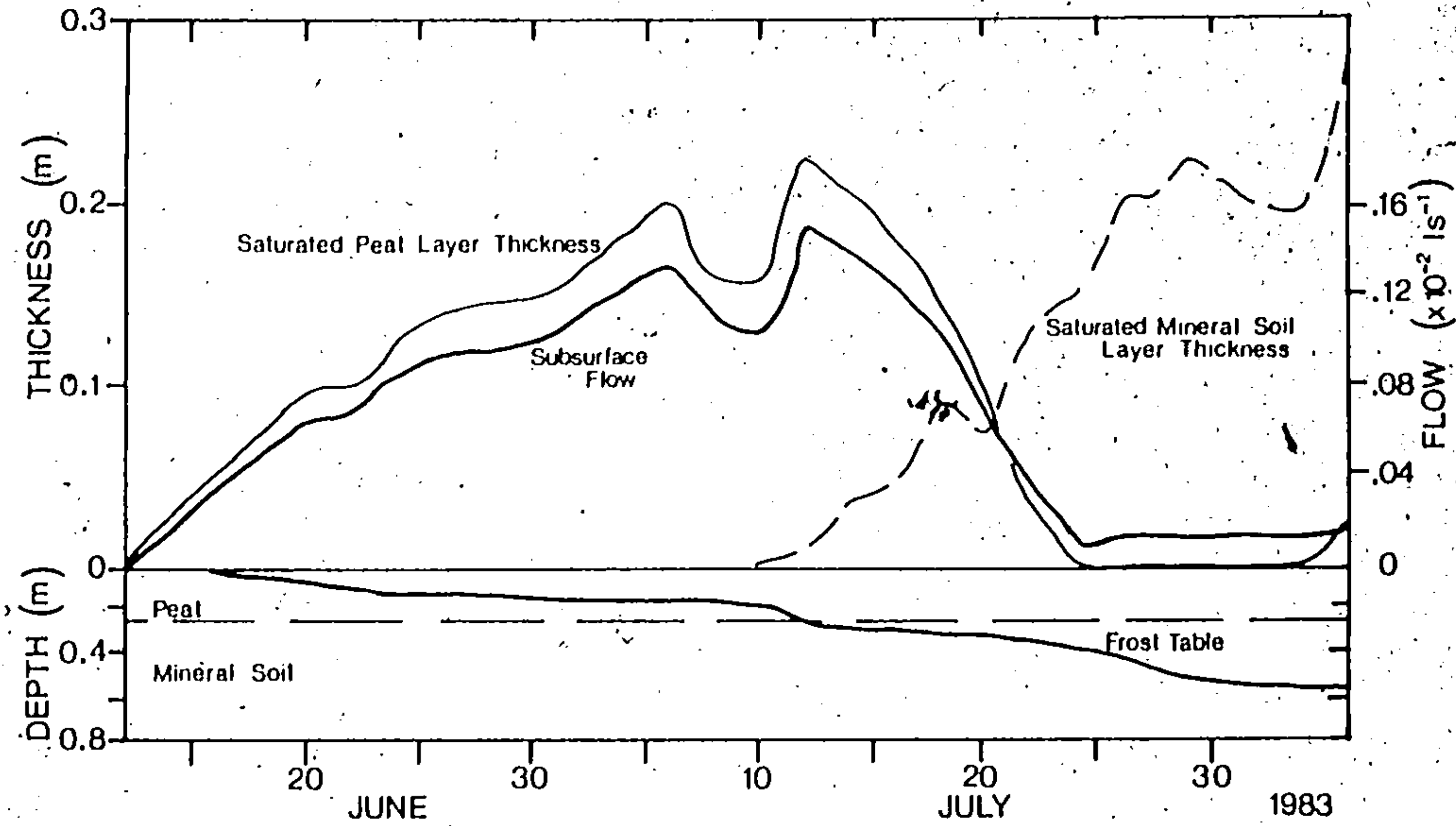


Figure 5.7 Wetland saturated zone thickness and daily subsurface discharge, 1983.

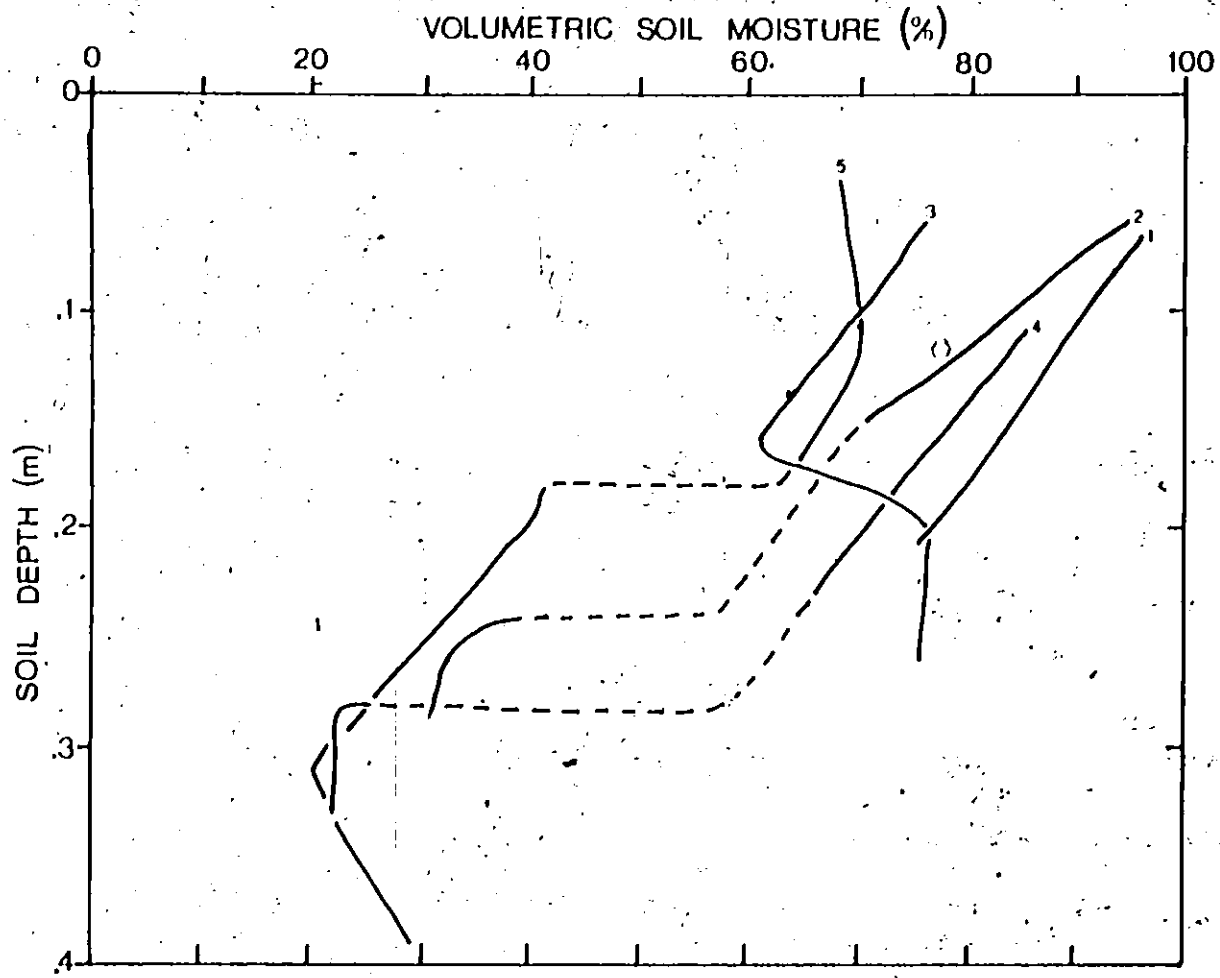


Figure 5.8 Wetland soil moisture, 1983. Numbers indciate dates: 1. July 8, 2. July 15, 3. July 22, 4. July 29 and 5. Aug. 4.

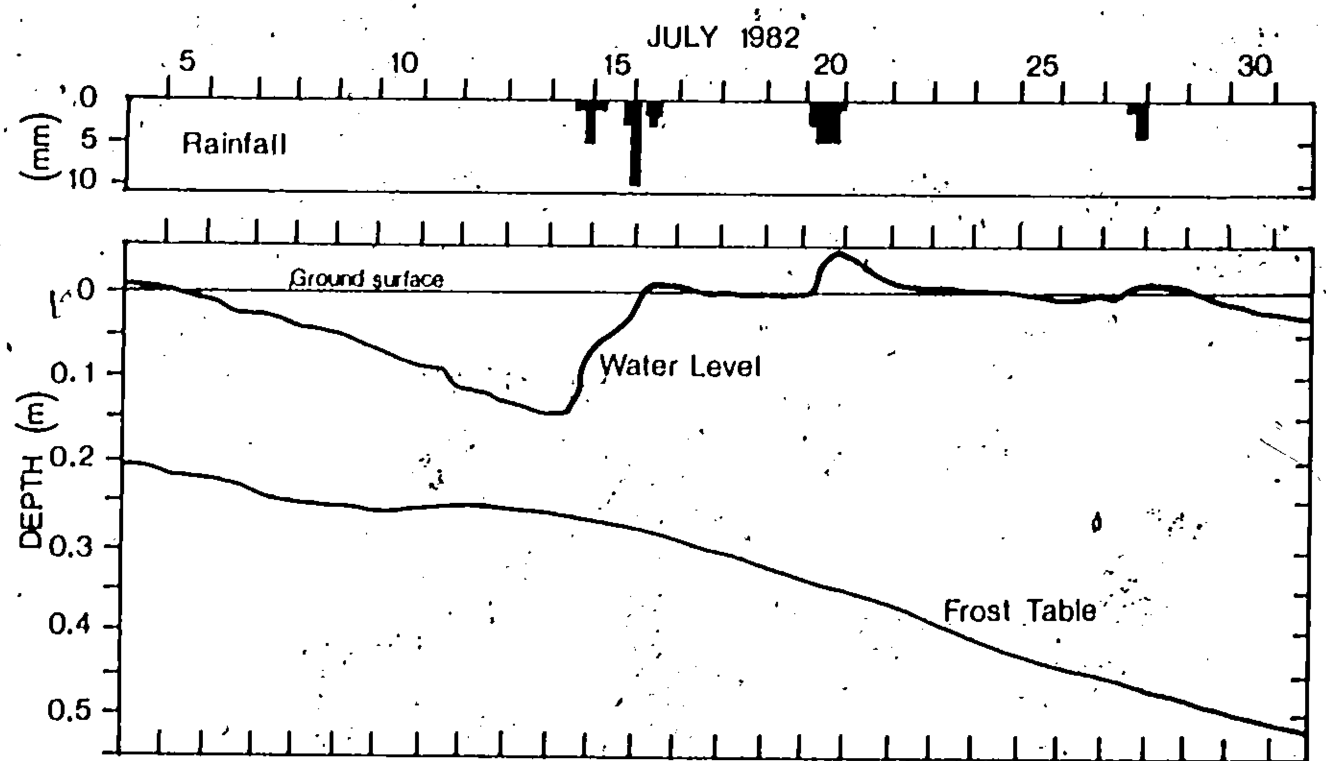


Figure 5.9 Wetland ground water level and frost table depth, 1982.

An infiltration capacity of 300 mm hr^{-1} was measured for the surface layer of the peat and 30 mm hr^{-1} for depths of 0.08-0.14 m, therefore rain water quickly penetrated the ground surface and recharged subsurface soil moisture deficits. The second rainstorm on July 20 (28 mm) fell on an already saturated wetland with little remaining capacity to store additional water. Prior to this storm event the water level was 0.008 m above the ground surface, therefore approximately one third of the depression storage was already occupied. The water level rose to 0.045 m. Although surface flow was not measured, its magnitude can be estimated using the 1983 relationship shown in Figure 5.5. Surface discharge generated by this storm could have reached 3.0 mm d^{-1}. Also the rainstorm was sufficient to produce discharge from the lake upslope of the wetland. However, all of this water was held in storage in the pond immediately below the lake.

Subsurface flow was calculated for the period between July 8 and August 5, 1982. Since the rainstorms occurred when the thaw zone was well developed, peak subsurface discharge reached 6 m^3d^{-1}. Greater flow depths persisted for the entire 1982 field season, though the subsurface discharge for the period amounted to slightly more than 1.0 mm.

5.4.0 Wetland Water Balance

As has been done for the upland hydrological system (see Section 4.4.0) the relative importance of the components of the wetland hydrological system were computed using the water balance equation (eq. 4.1) from May 1 to August 7, 1983. Lake discharge constituted by far the largest source of water, followed by melt water (Table 5.2). The greatest loss of water from the wetland was by surface flow, but evaporation was also prominent. Subsurface discharge was negligible, representing less than 0.5 percent of the total loss.

The wetland retained water mostly as ground ice in winter. The melting of ground ice provided 251 mm or 80 percent of the total water held in subsurface storage, thus satisfying the bulk of the considerable water retention capacity of peat. This value was based on the assumption that both the peat and the mineral soil were completely saturated at the time of freeze-back.

As a check on the reliability of the computed water balance, water stored in the wetland was measured on August 5, 1983. Assuming total saturation below the water table and using the August 4 measurement of soil moisture above the water table, there was 294 mm of stored water. Compared with the 317 mm obtained by the water balance calculation, the difference is small and well within the error limits of the techniques used.

The seasonal change of the water balance components is shown in Figure 5.10. There was a delay after melt had begun because water was stored in the snowcover.

TABLE 5.2 Wetland Water Balance (May 1 - August 7, 1983)

Snowmelt/Rain	Ground Ice	Lake Discharge	Surface Flow	Subsurface Flow	Evaporation	Storage Change
146/39	251*	430	331	1	217	317*
	210+					276+

Note: All Values in mm   * Assuming 100% saturated with ice
+ Assuming 70% saturated with ice. Measured storage was 294 mm.

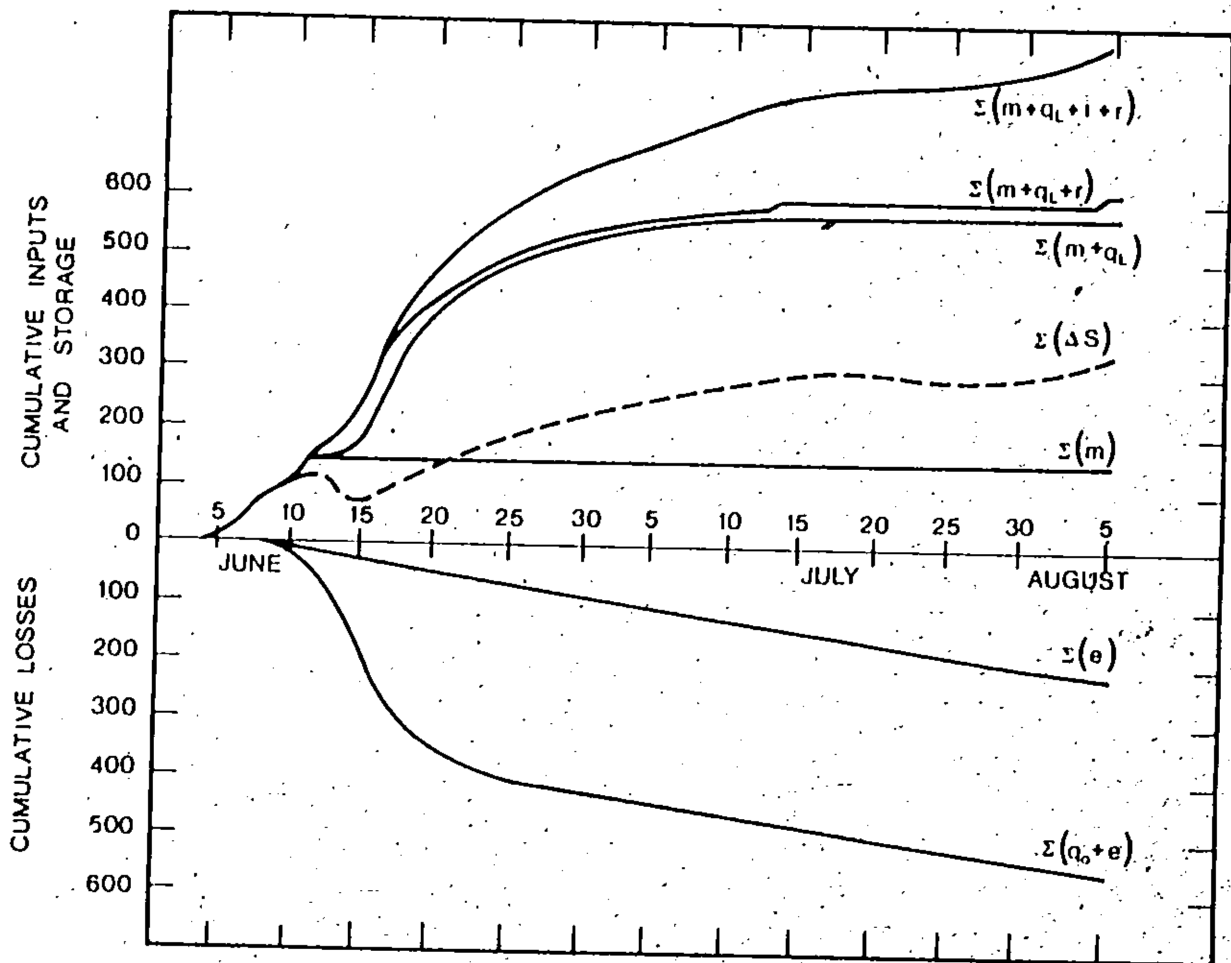


Figure 5.10 Wetland cumulative gain and loss of water, 1983.

Later, while lake discharge to the wetland became prominent, surface discharge from the wetland declined. This appears inconsistent unless some of the lake discharge is actually stored in the wetland.

The infiltration experiments suggested that little water can penetrate the frozen peat, but the ground temperature indicated the possibility of meltwater infiltration. Immediately after snowmelt began temperature in the upper layers of peat rose rapidly (Figure 5.2). The total amount of energy required to produce this rise is 2.99 MJm^{-2}, estimated by

$$Q = C \cdot \Delta T \; \Delta z \qquad (5.1)$$

where C, the volumetric heat capacity, was computed from the fractional components of ice and peat (deVries, 1952; eq. 2.3). Conduction could not provide this heat as no significant temperature gradient existed. If the temperature rise was entirely due to the latent heat released by the freezing of infiltrated water (Woo and Heron, 1981) approximately 4 mm d^{-1} could have infiltrated during this two day period. This water would refreeze blocking further entry of water into the soil, thus accounting for the experimental observation that water could not penetrate the frozen peat.

Examination of the lake inflow into the wetland and surface discharge hydrographs (Figure 5.3) reveals that discharge from the wetland peaked before the maximum inflow of water to the wetland. By integrating the area between the two hydrographs from their points of intersection (June 15 and July 9) and subtracting the difference between precipitation and evaporation, the wetland gained 61 mm of water. This water could not be stored at the wetland surface as depression storage was already exceeded. During the same period the frost table dropped 0.20 m but, if the peat was originally saturated with ice, only 17 mm of storage would have been available. To accommodate the remaining 44 mm the peat must have had void space in its frozen column. A 70 percent saturated ice content in the upper 0.2 m can satisfies this storage requirement for the remaining water. Based on this result the ground ice contribution was recalculated as 210 mm which is still significant (Table 5.2). The seasonal difference between computed and measured water storage is now 23 mm, 3 percent of the total water input.

A major portion of lake discharge into the wetland was not measured in 1982 so an adequate assessment of the water balance components cannot be made for that year.

5.5.0 Discussion of the Wetland Hydrological System

This wetland study reveals several characteristics that are unique to the northern wetland system. This discussion will concentrate on the wetland hydrological system and in the next chapter the role of the wetland at the basin scale will be examined.

The temporal variability of the hydrological components and processes are directly linked to the wetland's dynamic storage capacity. Storage capacity is limited both by the rate and depth of ground thaw and the hydrological properties of the peat. Even though peat is highly porous, it has a high specific retention. A large amount of water is held against gravity within the unsaturated zone, so that much of the void space cannot store additional input.

There were three external sources of water to the wetland. Snowmelt water produced in the wetland itself was of secondary importance, being less than half the magnitude of discharge from Lost Lake. Not only is the magnitude of the lake and snowmelt water important but the temporal distribution of these inputs has equal significance. When snowmelt began the soil was frozen and possibly became impermeable after some initial infiltration so that surface depressions provided the only available storage. Once depression storage was exceeded surface discharge increased rapidly. When the lake input became significant, ground thaw had increased access to subsurface storage. From the recalculated water balance it was seen that a portion of the lake discharge had to be stored as ground water in the wetland. This storage of water prolonged the effect of the lake input and kept the wetland saturated until July 9, thus satisfying the peak evaporation demand around the summer solstice. If this lake input was significantly reduced, the wetland would be unable to sustain the high moisture levels essential for wetland development.

This process of wetland recharge contrasts with runoff in the high Arctic wetland system reported by Ryden (1977). The wetland areas of Truelove Lowland, where Ryden worked, are situated in depressions between raised beach ridges so that the flow is confined and excess water from snowmelt is partially retained. Similarly, the wetland investigated by Brown et al. (1968) was in a depression formed by an old lake bed.

Rainfall was the third input of water to the wetland. Ground ice, while not an external input in the same sense as rainfall, snowmelt and lake discharge, was a notable storage component in that it satisfied a large portion of the storage capacity of the wetland.

Northern wetlands generate a seasonal pattern termed 'muskeg regime' by Church (1974) who suggested that wetlands attenuate high flows because of a 'large water retaining capacity of the muskeg vegetation' and 'high resistance' to flow caused by vegetation and irregular surface conditions. Results from this study demonstrate that a wetland's ability to moderate flow and reduce response is small because of a limited storage capacity. Runoff from the wetland investigated did not follow Church's 'muskeg regime' in spring when the soil is frozen and inputs were large. In summer the storage capacity increased as the peat thawed and the runoff response became variable. During the summers of 1982 and 1983, minor rain events raised the water table substantially. This was accompanied by increases in wetland discharge commensurate with water table rises above the surface depression storage limit. During such storms when surface and subsurface storage were exceeded there was little attenuation of flow. As a result of the July 20, 1983 storm, wetland discharge could have reached 3.0 mm d^{-1} which would have been 100 times larger than the normal summer wetland flow. This pattern of response is similar to the dynamic relationship between storage and summer runoff reported for temperate wetlands (Bay, 1969; Woo and Valverde, 1981).

It is widely held that temperate wetlands sustain flow during periods of low flow by a 'slow release of water' (Bertulli, 1981). To sustain low flow, a significant amount of water should be discharged from the wetland as subsurface flow. The small subsurface flow computed for this wetland site is typical of many wetlands (Boelter, 1972; Rothwell, 1982). In all wetlands there is a small hydraulic gradient, but in permafrost regions the cross sectional flow area is further reduced by the shallow thaw zone. Thus, this study shows that even saturated wetlands may not produce significant discharge once surface flow has ceased. Ryden (1977) reached the same conclusion for a high Arctic wetland.

While subsurface flow from the wetland was insignificant, evaporation was a major process through which water was lost in the summer. Evaporation was responsible for the decrease in soil moisture content between July 8 and August 5, 1983. Water is easily supplied to the evaporating surface by the vertically connected pores and an adequate capillary rise (Romanov, 1968a). Romanov (1968b), summarizing many years of Soviet research on bog hydrology and evaporation, demonstrated that if the water table is within 0.25 m of a wetland's surface evaporation is a function of atmospheric conditions, but when the water table falls below 0.25 m evaporation becomes limited by the moisture supply. This is because of a break in the capillary rise when the water table depth exceeds 0.25 m, causing the water in the larger soil pores to drain under gravity (Romanov, 1968a). The water table during the two summer field seasons was within this upper zone of capillary influence.

In conclusion the wetland can be seen as a dynamic storage system generating a variable discharge through time. The lake and wetland system are intimately linked. Surface flow is the chief mechanism through which water is lost from the wetland and the bulk of it occurs in spring when the ground is frozen. Surface flow is also significant in summer, when rainstorms exceed the limited wetland storage capacity, while subsurface flow is insignificant in storm response and as a component of the seasonal water balance. The limited storage capacity is increased by evaporation and ground thaw. With a limited storage capacity, the wetland plays a minor role in flow regulation in spring and during summer when large storms occur.

CHAPTER SIX DISCUSSION ON LOW ARCTIC BASIN HYDROLOGY

The hydrology of a low Arctic basin reflects the combined effect of the temporal and spatial variability of linkages among several hydrological subsystems. In the previous chapters the hydrological processes in the three main subsystems were discussed, providing the basis of an explanatory framework for the low Arctic hydrological system. In this chapter, the discussion will concentrate on the spring snowmelt and summer runoff processes and the basin water balance for the 1983 study period.

6.1.0 Basin Water Balance

A basin water balance was computed for the period May 15 to August 5, 1983. The depth of runoff originating from the study basin was initially calculated by subtracting the Heart Lake discharge from the Hardill Lake discharge and dividing by the study basin area. However, since the study basin constitutes less than 10 percent of the Hardill Lake basin, errors in the measurement of discharge would be grossly amplified. Runoff per unit area computed using the entire Heart Lake basin area of 11.9 km² was therefore assumed to be the same as that for the study basin. Woo and Drake (personel communication) reported that this basin contained 19, 73 and 8 percent of wetland, upland and lakes respectively. These proportions are similar to those of the smaller study basin.

The basin water balance was computed using

$$m + r - e - q = \Delta S \qquad (6.1)$$

where m is the basin snow water equivalent, r is rainfall, e is evaporation calculated on an areal basis from the snowfree area of each surface type, q is Heart Lake discharge and ΔS is the change in basin water storage over the summer. For the purposes of this water balance calculation the dry peatlands were considered as uplands.

The magnitude of the water balance components is shown in Table 6.1. The result shows a large negative storage. Re-evaluating the basin water balance with the adjusted upland evaporation (24 percent decrease: see section 4.4.0) the seasonal storage change is reduced to -76 mm. It is impossible to measure basin storage, but based on estimates of ground ice content, I, and measured soil water content, θ, at the end of the field season, an estimate can be made for the land portion using:

$$\Delta S = \frac{\rho_I}{\rho_W}\left(\left[\int_0^{z_1} I(z)\,dz\right]_{t_1} - \left[\int_0^{z_1} \theta(z)\,dz\right]_{t_2}\right) \qquad (6.2)$$

where t_1 and t_2 represents premelt and August 5 respectively, and z_1 is the depth of the frost table on August 5. Values for the first and second terms in equation 6.2 are

TABLE 6.1 Basin Water Balance May 15 - Aug. 5, 1983

Inputs		Outputs		Change in
Snow	Rain	Evaporation	Discharge	Storage
201	39	159	182	-101
		134*		-76*

* Calculated using the adjusted upland evaporation
• All values are expressed in mm.

presented in Tables 4.2 and 5.2 for the upland and wetland. The results of this calculation are shown in Table 6.2. Both Lost Lake and Heart Lake water levels dropped well below their premelt level (Figure 3.14) indicating a negative lake storage change, averaging -115 mm (see Section 3.4.2). The total change in basin storage was computed as +9 mm using,

$$\Delta S_b = xA_u \Delta S_u + xA_w \Delta S_w + xA_L \Delta S_L \tag{6.3}$$

where xA is the fractional basin area occupied by each surface type and subscripts u, w, L and b signify upland, wetland, lake and basin. These estimates of basin storage change indicate that inputs almost balanced outputs over the summer of 1983.

The disagreement between storage change calculated using equation 6.1 and 6.3 is large but not unrealistic considering the possible accumulated errors. Water balance change in storage expressed as percentages of total precipitation, discharge and evaporation are 32, 42 and 57, but it is likely that errors are spread among all the components. Using the error margins quoted in Section 2.3.4 the magnitude of possible errors in snow water equivalent, rainfall, discharge and evaporation are 30, 12, 27 and 27 mm respectively. Assuming these errors are representative and random the root mean square error of the basin water balance is 50 mm, but the error in some components such snow water equivalent and rainfall are probably systematic. If the

TABLE 6.2 Basin Storage Change in mm, 1983

Surface Type	Initial Storage	Final Storage	ΔS	Area	Basin ΔS
Wetland	210	294	84	0.21	18
Upland	183	181	-2	0.73	-2
Lake	0	-115	-115	0.06	-7

direction of all errors is the same the maximum error would be 96 mm. The absolute value of the computed storage change is 76 mm (Table 6.1), which is only three quarters the size of the maximum potential error. There is no way of establishing the actual water balance error and these estimates are provided to illustrate the difficulties of establishing confidence limits around such calculations.

It is important to establish how representative the estimate of the basin water balance components are in a regional context. However, this is difficult since no long term records exist for any of the measured variables. Snow on the ground at Baker Lake measured the spring of 1983 represented 86 mm of water, while a total of 135 mm of snow water equivalent fell from October, 1982 to April, 1983 at the same location. These values are respectively 115 mm and 66 mm less than the measured basin snow water equivalent. The snow survey data is not unreasonable because Arctic weather stations are known to under-represent basin snowfall by up to 300 percent (Woo et al, 1983). The normality of the basin rainfall was discussed in Section 2.1.0. Annual open water evaporation for the region was estimated by Ferguson et al. (1970) to be approximately 150 mm. This is similar to the basin evaporation for the 1983 field season, but it should be noted that there remained the month of August and the first part of September for additional, though decreasing, evaporation losses to occur.

Several water balances have been computed for other basins in the continuous permafrost region (Table 6.3). In the high Arctic snow represents over 80 percent of water input (Ryden, 1977; Woo, 1983) but rain becomes more significant further south. Based on the 30 year climate norm of Baker Lake, rainfall and snow are of equal importance. Based on the 1983 measured basin snow water equivalent and the 30 year rainfall norm, snow represented 60 percent of basin water input. The seasonal runoff ratio is similar to that of high Arctic non-vegetated basins (Woo, 1983). However, if most of the error lies in the estimation of inputs, the runoff ratio could be as low as 0.58. Mean annual runoff ratio for several rivers in central Keewatin ranged from 0.56 to 0.64 (Woo et al, 1982) but all these river catchments are several orders of magnitude larger than the study basin.

The only other Keewatin water balance was computed for a small basin on the north side of Chesterfield Inlet (Newbury et al., 1979). Snow water equivalent was grossly underestimated, but even after adjustment the results suggested that far more precipitation ran off in this coastal environment than elsewhere in central Keewatin. The large water body of Hudson Bay adjacent to this basin may have suppressed evaporation. Another permafrost basin located at Barrow, Alaska, contains wetlands and had a mean annual runoff ratio of 0.5 (Brown et al, 1968). The present

TABLE 6.3 Arctic Water Balance Studies

Location and Basin Size	P (mm)	Q (mm)	E (mm)	Q/P	Reference
Cornwallis Is. (74°45′N,94°50′W) 33 km^2, 5 yrs.	233	199	38	0.71	Woo(1983)
Truelove Lowland (75°33′N,84°40′) 3 yrs.	185	84	101	0.45	Ryden(1977)
Chesterfield Inlet (63°20′N,90°43′W) 1.5 km^2, 1 yr.	397	290	107	0.73	Newbury et al (1979)
This study	240 (316)*	182	134	0.76 (0.58)*	

* Calculated assuming no change in storage and P = Q + E

study and others (Brown et al, 1968; Ryden, 1977) demonstrate that permafrost basins containing wetlands yield less runoff because of higher evaporation losses.

The seasonal change of basin water balance observed by this study (Figure 6.1) is similar to other Arctic basins that contain wetlands (Ryden, 1977). Snowmelt runoff dominated in spring while in summer evaporation was dominant. Unlike the high Arctic, summer rainstorms in central Keewatin can be quite large. Positive changes in storage occurred in the summer in response to these rainstorms. In 1983, immediately after this research project was terminated, the Baker Lake area received the largest 24 hour storm on record. This rainfall would have satisfied the storage deficit and generated runoff. If the runoff response of the streams at Baker Lake was an indication, the storm discharge from the study basin would have been considerable.

6.2.0 Basin Response Processes

Based on the basin water balance two seasons are apparent. A spring period is dominated by large inputs and high runoff and a summer period is characterized by rainfall induced runoff events and large water losses through evaporation. The runoff mechanisms of the spring and summer periods are discussed in the following two sections.

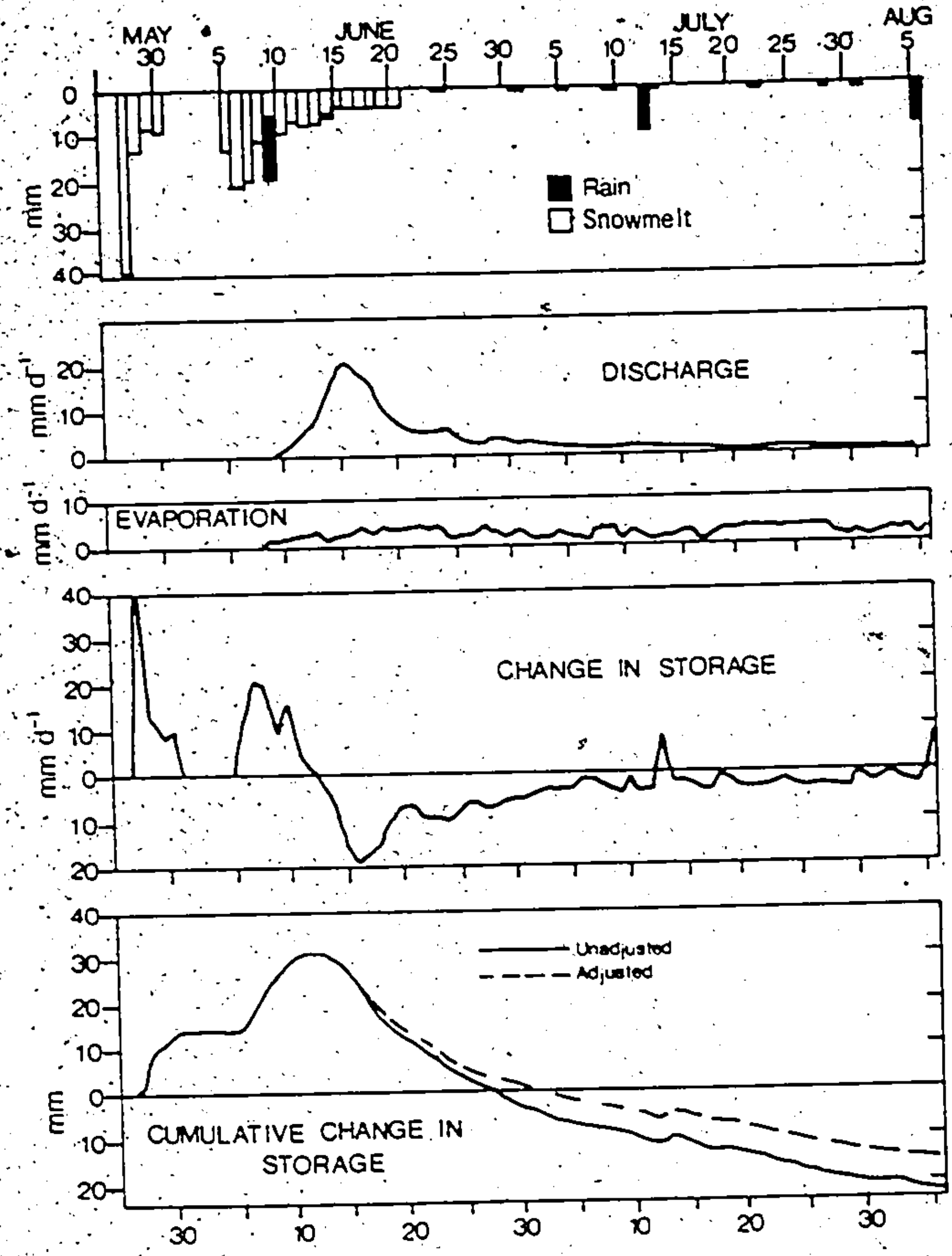


Figure 6.1 Basin inputs, outputs, changes in storage and cumulative storage change, May 25 - August 5, 1983. The adjusted cumulative storage incorporates the change in upland evaporation discussed in section 4.4.

6.2.1 Runoff Processes in the Spring Period

Water movement began in the uplands (May 31) immediately after the initiation of snowmelt (May 28) in 1983. There, the snowcover was the shallowest, being 106 mm water eqivalent compared with a basin average of 201 mm. Some of the snow may have been lost through sublimation (Ryden, 1977). Kane et al. (1981) measured this loss to be one third of the total snowcover water equivalent. This can reduce considerably the water available for runoff, but in the present study sublimation was not measured. A portion of the upland meltwater infiltrated and was stored in the organic soil layer (36 mm) and later in the mineral soil (17 mm). Other studies in permafrost regions have also indicated that an organic soil layer can provide a significant amount of meltwater storage (Santeford, 1979; Kane et al., 1981). Initial subsurface storage was limited to the organic layer only and 40 percent of the snowmelt water ran off. In response to this water supply, lake and stream levels began to rise noticeably on June 5 (Figure 6.2). Over the next few days upland discharge rapidly increased from 20 to 60 percent of its cummulative runoff total and lake levels continued to rise, reaching 50 percent of their maximum rise on June 10. Up to this time the bulk of the wetland meltwater was held in the deeper snowcover. As this snowcover disintegrated, wetland depression storage was quickly exceeded and surface runoff increased rapidly. The uplands

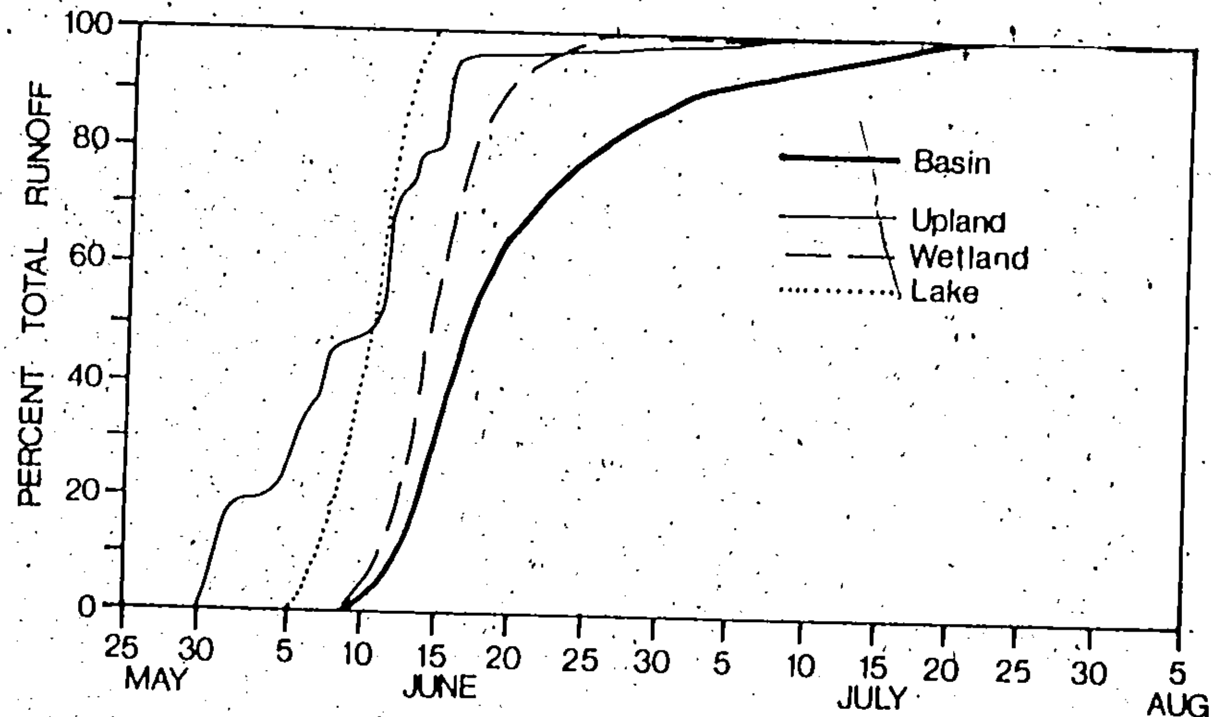


Figure 6.2 Cumulative percent of total runoff for the basin, wetland and upland hillsope, 1983. Cumulative lake level rise relative to maximum rise is also shown.

had released most of their runoff by June 12, causing the water level of downstream lakes and channels to overflow their banks, discharging into the adjacent wetlands. Although basin discharge began before this date, sizable outflow occurred only when the runoff system became integrated. The wetlands provided the route through which snowmelt runoff was conveyed to the basin outlet. In six days the wetland discharged 75 percent of its total runoff. During the entire spring period, the wetland provided little storage relative to the volume of the water input nor was the flow greatly attenuated. Once the wetland outflow receded due to diminished input (June 18 to June 30), there was a similar but slightly delayed recession in basin discharge (to July 10).

The overwhelming influence of snowmelt on basin discharge can be seen in Figure 6.2. Seventy-five percent of basin discharge had occurred by June 25, 1983, only 16 days after flow began. The uplands and wetlands offer little storage capacity to absorb the large quantity of snowmelt water because of frozen soil. There was ample infiltration into frozen soil in the upland area until flow occurred as a result of storage exceedance. In the wetland surface flow was generated because of little infiltration, possibly as a result of blockage of soil pores when the meltwater refroze (Woo and Heron, 1981). High surface flows continued even after the ground began to thaw, as the supply of lake water exceeded the wetland storage capacity. Landals and Gill (1972) observed that soil moisture conditions of the previous fall were important for the quantity and timing of snowmelt runoff in the following spring. Larger discharges were observed during the 1982 spring and in the year previous to that there were large rainstorms late in the summer (Woo et al., 1982). The 1982 snow survey indicated that there was much less water available but this was the result of a poor snow survey which underestimated the availability of snow.

Discharge during the snowmelt period can vary from year to year. Corresponding to daily temperature fluctuations and night-time freezing were large diurnal runoff cycles in 1982. In contrast, the 1983 snowmelt discharge showed much less fluctuation. Snowmelt began earlier in 1982 and then stopped for a week. The second melt period produced a sharp rise, yielding 70 percent of total water discharge in only six days (Figure 6.3).

Based on the results of this study, the spring streamflow regime of this central Keewatin drainage basin could be classified as 'Arctic nival' (Marsh and Woo, 1981) and is not overly different from snowmelt runoff in other Arctic regions. The classification of muskeg regime (Church, 1974) cannot be applied to central Keewatin for the spring period because even though wetlands occupy much of the basin there is little available storage to dampen the

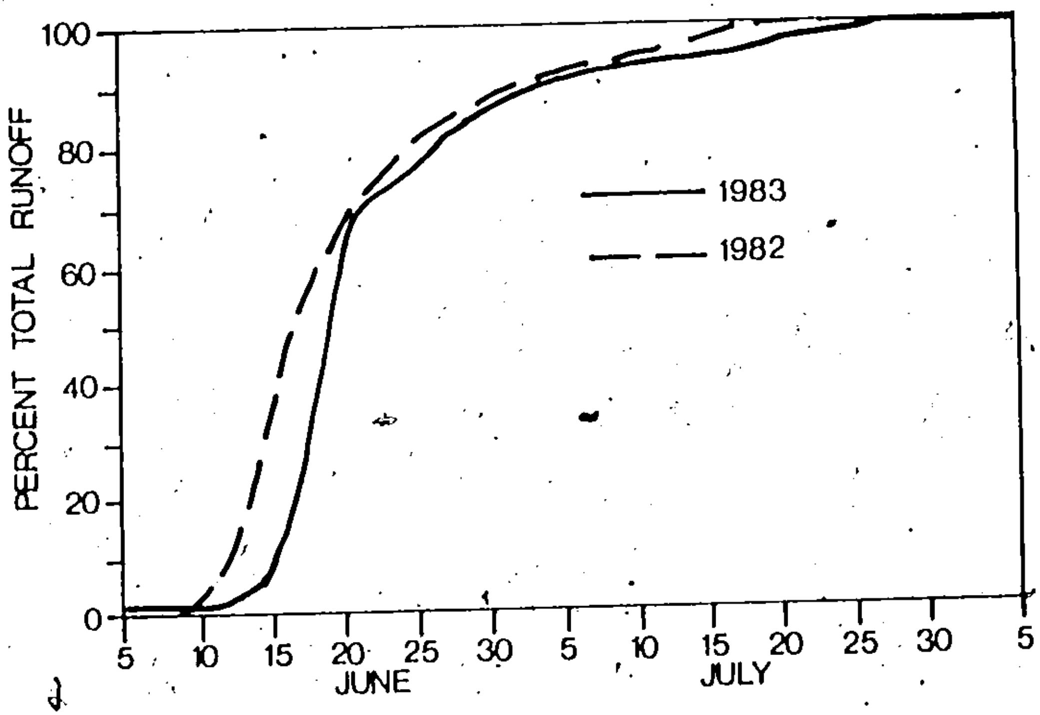


Figure 6.3 Cumulative percent of total runoff for the basin, 1982 and 1983.

large flows. The large snowmelt input and occurrence of frozen ground during spring overwhelms any intra-basin differences that become more apparent in the summer.

6.2.2 Runoff Processes in the Summer Period

Summer is a low flow period with occasional increases in discharge caused by summer storms.

As the frost table deepens in the wetland and the upland, the water table dropped. Discharge from the wetland was negligible once the water level dropped below the ground surface since wetland subsurface flow was insignificant (maximum of 3.3×10^{-2} mm d^{-1}). Upland discharge was sustained longer by subsurface flow, but became much reduced when the water table dropped below the organic- mineral soil interface. These patterns of declining summer flow have been observed on high Arctic hillslopes (Woo and Steer, 1983) and in other Arctic wetland where summer flow ceased entirely (Ryden, 1977).

From early July, 1983 onwards, streamflow was sustained principally by lake drainage. On July 1, 1983, basin discharge was 2.0 mm d^{-1} and was less than 0.10 mm d^{-1} on July 20. The cumulative water loss from Heart Lake between July 1, the date when wetland discharge ceased, and August 5, assuming no change in lake surface area, was -395 mm (Figure 3.14). One hundred and forty five millimetres of this loss was through evaporation (Figure 3.6). Distributing the remaining -245 mm over the entire study basin area, Heart Lake provided 13 mm of the 20 mm total basin discharge during this period.

As was discussed in Section 5.5.0 it is commonly thought that wetlands sustain low flows during dry periods (Bertulli, 1981), but the results of this study conform more to Bay's (1969) findings that "some types of peatlands do not .... make any substantial contribution to streamflow during dry periods"(p. 101). Boelter (1972) showed that some wetlands cannot produce adequate lateral subsurface flow to maintain a moderate water level in adjacent streams. In diplotelmic wetlands, the upper 0.2 to 0.3 m of peat called the acrotelm, is the only hydrologically active zone (Ivanov, 1975; Romanov, 1968a). Below this, a layer called the catotelm, is hydrologically inactive because of low hydraulic conductivity and large specific retention (Boelter, 1965). The wetlands in the study basin lack an acrotelm so that flow becomes negligible once the water table moves below the soil surface. Ingram (1984) calls this type of wetland a 'haplotelmic' wetland.

Evaporation in 1983 was a principal factor (138 mm) in the creation of the large storage deficit (-76 mm). The evaporation from lakes, uplands and wetlands has been discussed previously and placed in the context of subsystem water balance calculations. In terms of summer runoff, the amount of evaporation loss between rain storms will strongly affect the timing and magnitude of runoff response.

Four summer storms occurring on July 15-16 and July 20-21, 1982 and July 12 and August 5, 1983 were analyzed to determine such streamflow characteristics as lag time, flow recession and runoff volume. Lag time was computed as the time interval between the centroid of rainfall mass and the peak in basin discharge and the recession constant was calculated using

$$m = \frac{-\Delta t}{\ln (q_2/q_1)} \tag{6.4}$$

where Δt is the time interval between discharge measurements q_1 and q_2. The volume of storm runoff was calculated by base flow hydrograph separation (Gray and Wigham, 1970). The storm hydrographs are shown in Figure 6.4 and 6.5 and the streamflow characteristics are presented in Table 6.4. For comparison, a range of streamflow characteristics for other permafrost basins containing wetland areas are given in Table 6.5.

Differences in streamflow characteristics between the July 15-16 and the July 20-21, 1982 storms may be attributed to the influence of antecedent moisture conditions. Both Brown et al. (1968) and Dingman (1971) found that runoff ratios were related to antecedent moisture conditions. Brown et al. (1968) observed a 40 percent increase in runoff ratios from a storm sequence similar to

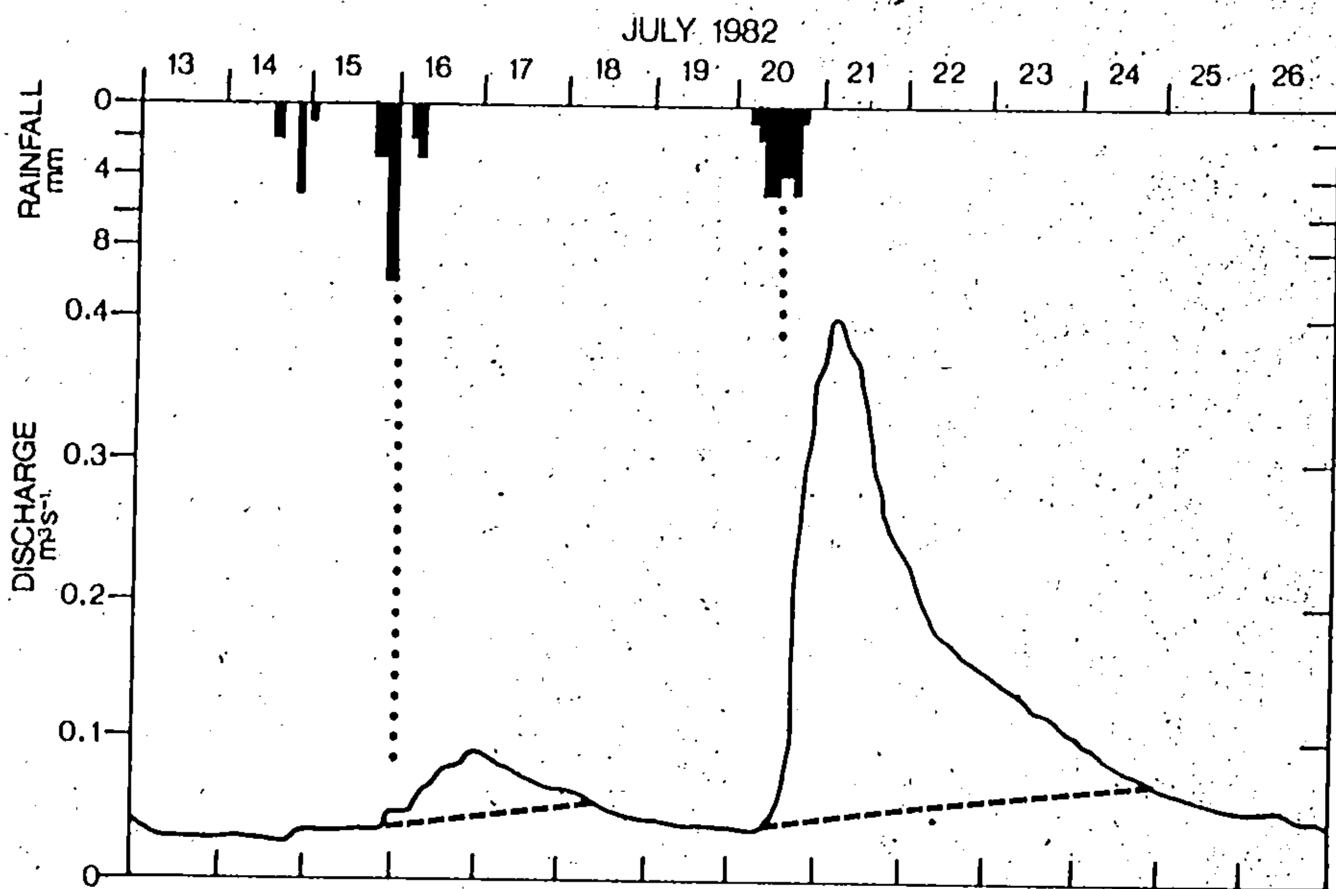


Figure 6.4 Storm hydrographs and rainfall, 1982. Dotted line indicates the centriod of rainfall and the portion of the hydrograph above the dashed line is considered stormflow runoff.

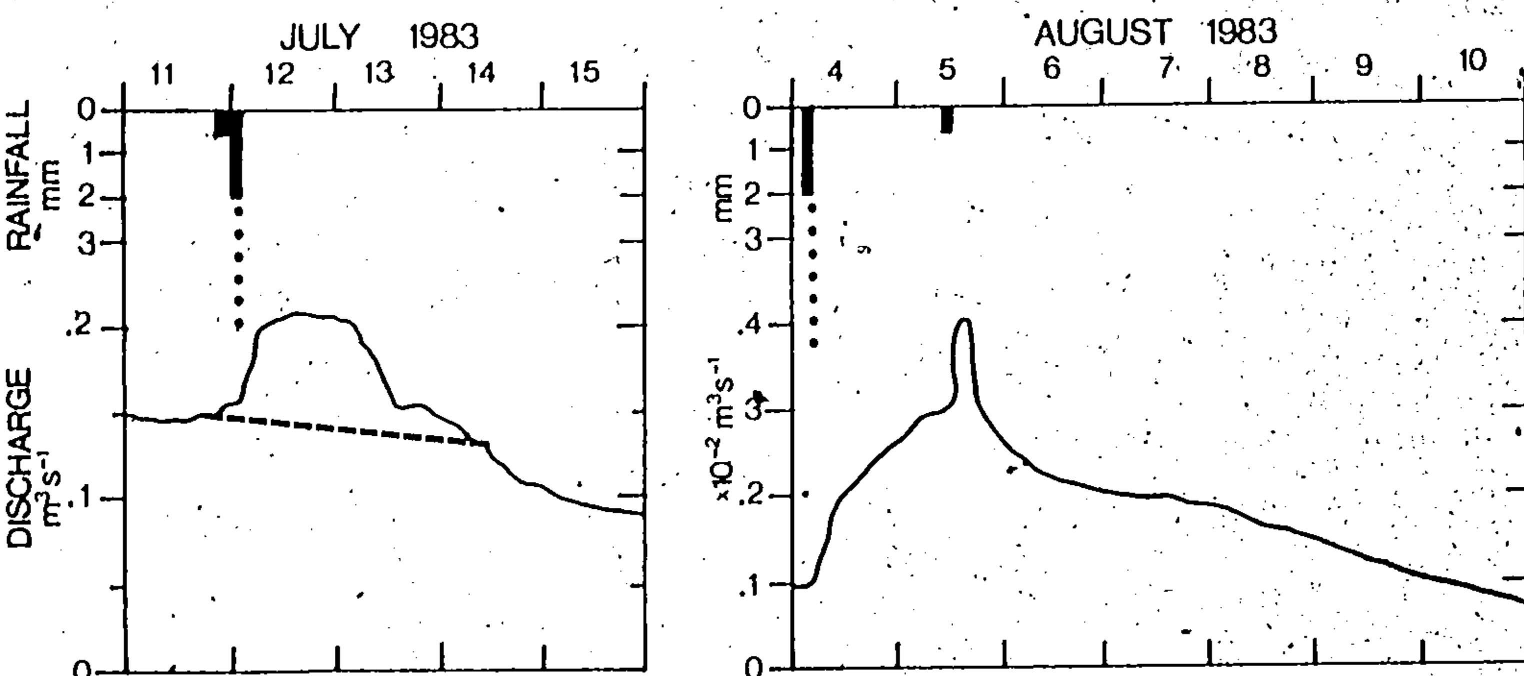


Figure 6.5 Storm hydrographs and rainfall, 1983. (see Figure 6.4 for explanation).

TABLE 6.4 Streamflow Characteristics 1982-1983

Date	P (mm)	Q (mm)	Q/P	Lag Time (hr)	Recession Constants (hr) a	b	c
1982							
July 15-16	32.4	0.5	0.02	27	59	109	
July 20-21	27.4	4.5	0.16	16	30	56	112
1983							
July 12	9.8	0.6	0.06	13	44	87	223*
Aug. 5	9.2	na		37	141		

* Recession constant used for base flow separation.
a. 24 hrs, b. 24-96 hrs, c. >96 hrs.

TABLE 6.5 Streamflow Characteristics of other Arctic Basins

Location and Basin Size	n	Lag Time	Recession Constants	Q/P	Reference
Barrow, Alaska (71°18'N,156°47'W) 1.6 km²	16	2.5-21.7	15-162	1-69	Brown et al (1968)
Central Alaska (64°51'N,147°43'W) 1.8 km²	12	1.5-44.0+	21.7-76.9	14-42	Dingman (1971)

n is the number of storms analyzed.
+ Lag time computed from beginning of rainfall.
Time lag and recession constants are in hours.

the one which occurred in the present study during 1982. In the present study the first storm produced a runoff ratio of 0.06, while the second four days later yielded a runoff ratio of 0.16. In the same storm sequence, time lag was considerably shorter for the second storm, but if the storm of July 12, 1983 is also considered there seems to be little relation between antecedent moisture conditions and response time. Brown et al. (1968) had similar results, but Dingman (1971) was able to establish a relationship between basin wetness and storm response time. There are not enough storms from the present study to assess the cause of these differences in results. The long recessions found in this study (Figure 6.6) were also observed in the previously mentioned studies. Brown et al. (1968) attributed the long recessions to the detention of flow by many surface depressions, but in the present study they are probably due to water detention in lakes.

The different storm responses in this study were probably due to a variable contributing area which expanded and contracted according to the degree of linkage between basin subsystems. Small storm responses were produced by direct precipitation on open water and by groundwater flow from the hummock areas. The hummock area responded readily to rain, providing discharge even after a prolonged dry period (Figure 4.6). When larger storms occurred after prolonged dry periods some water which entered the upland

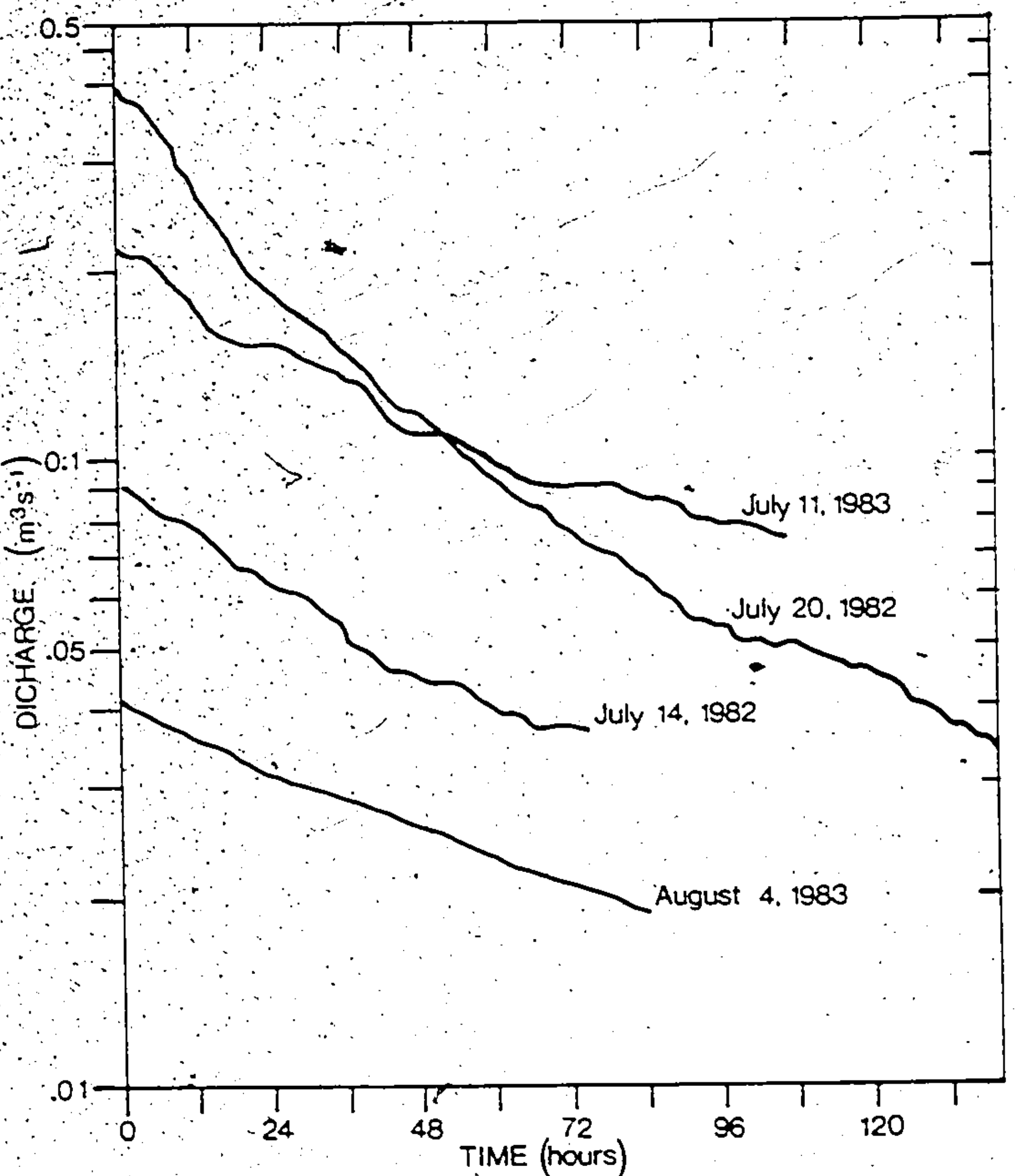


Figure 6.6 Storm recession limbs, 1982 and 1983. 0 hours is taken as the time of peak storm discharge.

system emerged as return flow (Figure 4.4). However, the Heart Lake hydrograph indicates that this process had little effect on basin discharge (Figure 6.4, Table 6.4). During this storm and the two storms in 1983, the wetland contributed little runoff. The wetland discharged only during the July 20, 1982 storm when the wetland, saturated by the previous storm, could retain little additional moisture. Storage overflow was generated in a fashion analogous to that described by Santeford (1979). For this storm, both the wetland and upland areas produced surface discharge. As the flow linkages among various hydrological subsystems became well established, basin discharge increased markedly.

An estimation of the storage capacity within each subsystem prior to the July 20, 1982 rain storm suggests that the uplands contributed more stormflow than the wetland. The amount of available upland storage was calculated as

$$\text{Available storage} = \int_0^{z_l} S_y(z)\, dz \qquad (6.5)$$

where S_y is specific yield and z_l is the depth of the water table. This calculation was adjusted for the difference in the upper and lower slope water levels (Figure 4.4) and a 60 percent areal extent of organic soil cover was assumed. This calculation yielded 11.5 mm of subsurface storage and with 6.0 mm of depression storage (see Section 4.3.0), the total upland storage capacity on July 19, 1982 was 17.5 mm. The wetland water table was 0.008 m (8 mm) above the ground surface, leaving a remaining depression storage capacity of 20 mm. This value of depression storage was the average based on the measurements made for this study and for other sites (see Section 5.3.0). The July 20, 1982 rainfall of 27.4 mm, produced 9.9 mm m^{-2} from the uplands and 7.4 mm m^{-2} from the wetlands. Taking the areal extent of the two systems into consideration, the uplands yielded 5 times more runoff than the wetlands during this storm. The water level rise above the ground surface in the wetland (37 mm) was greater than the depth of rainfall (27 mm). This rise suggests that additional water was input to the wetland from adjacent hydrological systems, but it was established in Section 5.3.0 that there was no lake input. A hillslope, approximately twice the size of the study wetland, lies immediately to the west and may have contributed additional water. There is no data available to evaluate this hypothesis, but it is highly likely that the low-lying position of the wetland enables it to act as a routing linkage between the upland and the lake and channel system.

Both the Arctic wetland basins studied by Brown et al. (1968) and this project generated much larger summer storm runoff than non-wetland Arctic basins of a similar size (cf. Anderson, 1974; Woo, 1980). This study shows that in this area of Keewatin the existence of wetlands can reduce summer stormflow under certain conditions, as Church (1974) stated, but under other conditions they produce a significant storm response.

In conclusion, the type of summer runoff pattern observed in this study basin did not match any one of Church's (1974) permafrost streamflow regimes. Attributes of several regimes were found. The runoff pattern was similar to that observed in other Arctic basins that contain wetlands. The presence of wetlands and lakes modify the runoff pattern yielding an unique runoff regime in this area.

CHAPTER SEVEN CONCLUSION

This study has examined the hydrological processes in a small low Arctic drainage basin. By analyzing the basin's subsystems water balance and runoff processes an explanatory framework was developed for the spring and summer hydrological periods.

The research basin is typical of other basins in the low Arctic, containing sloping uplands, wetland fens and many lakes. Fens occupy the valley bottoms and have an extensive peat cover, on a clay rich glacial till. The upland areas have a till based soil which is partially covered by an organic cryosol. Streams and lakes are generally small, but can double in size during spring.

During the two years of study, snowmelt contribution and ground ice were important inputs to the basin, while rainfall was less significant and variable in nature. Runoff and evaporation were of equal importance as water losses.

In spring, surface flow was found to be the chief mechanism through which snowmelt water was conveyed to the lake and channel network from the wetland and upland areas. It was primarily caused by a limited soil water storage capacity because of the frozen soils. Flow first began on upland slopes in late May and both lake and stream levels began to rise in response. Initially flow was on the ground surface, but after several days of ground thaw some subsurface flow did occur. Upland flow was supplemented by larger flows from the wetland areas two weeks after melt began. Such a delay in runoff was due to meltwater storage in the deeper wetland snowcover. As melt proceeded, the lakes and streams overflowed and water entered the adjacent wetlands. Overbank discharge exceeded the wetland storage capacity and large surface flows were sustained for an additional 10 days. All flow of significance in the wetlands was above ground.

Peak snowmelt discharge occurred on June 17, 1982 and June 15, 1983, 25 and 20 days after snowmelt began. In spring, surface flow was found to be the chief mechanism through which snowmelt water was conveyed to the lake and channel system from the wetland and upland areas. During the spring period 40 percent of the upland snowcover and all of the wetland snowcover eventually yielded run-off. Because of the much larger areal extent of the upland areas they provided one third of the basin runoff in spring while the wetland provided the remaining two thirds. As wetland discharge ceased and the upland water table dropped below the organic-mineral soil interface, lake and stream water levels also dropped. This in turn led to a decrease in streamflow, bringing to an end the spring runoff period.

In the summer period, the principal water loss was evaporation. Lake and wetland surfaces lost water at potential rates, but upland evaporation was limited by low soil moisture and a deeper water table. Maximum frost table depths were approximately 1.10 and 0.55 m in the upland and wetland areas respectively. Both subsystems had a restricted subsurface storage capacity because of high specific water retention of the peaty wetland soil and because of low porosity and high water retention in the upland soil.

Summer streamflow was sustained primarily by lake drainage. Observations clearly showed that the wetland areas do not maintain low flow during dry periods. This is due to the lack of an acrotelm or hydrologically active layer in the peat. The low Arctic fens can therefore be classified as haplotelmic wetlands.

Storm runoff response depended on the size of the rainstorm and antecedent moisture conditions. With minor storms and dry basin conditions, runoff was produced by direct precipitation and flow from wet hummock areas at the base of slopes. For larger rainstorms and dry basin conditions, water from upland areas reached the lower slopes as return flow and produced a small basin outflow. A greater response was generated if the water table in the upland area rose to the organic soil layer to produce a significant increase in subsurface flow. When several large storms followed each other in quick succession, surface flow was generated in the wetland and upland areas and the basin storm response was large. The wetlands play a significant role conveying the storm water to the basin outflow. When this linkages is established runoff ratios for individual storms are larger. The timing of response did not depend on basin wetness but streamflow recession generally took several days.

The variable runoff pattern of the low Arctic hydrological system most closely follows the 'Arctic nival' regime described by Church (1974) but is modified by the presence of wetlands and lakes. In spring the runoff pattern is overwhelming influenced by snowmelt and and the lack of storage in the frozen active layer. In summer the spatial diversity of cover types produces a varying storm response which depends on the degree of linkage between basin subsystems. Whether the summer runoff pattern follows the 'muskeg' type regime depends on the role played by the wetlands during different storms.

The main factor in determining if sizable discharges occur during the spring and summer is the degree of linkage within the basin. Whether various portions of the basin provide runoff through a linkage is dependent on storage capacity and the runoff generating mechanisms in that area. By examining the hydrological processes of various subsystems in the basin and the dynamic linkages between these subsystems, an improved understanding of low Arctic basin hydrology can be obtained.

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