HILLSLOPE HYDROLOGY AND RUNOFF PROCESSES IN A SUBARCTIC, SUBALPINE ENVIRONMENT

BY

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ABSTRACT

Within the subalpine zone of a subarctic basin, hydrological processes were studied on four hillslopes within 5 km^2 in an attempt to determine the factors that cause the variability in the magnitude and timing of water balance components. The hillslope was chosen as the scale of study as it links process operating at the point with streamflow, and exhibits strong contrasts in microclimate, vegetation, frost and soils, providing an ideal natural laboratory.

Hillslopes showed strong asymmetry in the timing and magnitude of processes during the melt and summer period. On slopes with well drained soils and seasonal frost, vertical hydrological exchanges predominate over the entire year and slopes rarely contribute runoff for streamflow. In contrast, hillslopes underlain with permafrost and/or poorly drained soils with a capping organic layer produce strong lateral fluxes. Water balance information highlighted the principal factors that lead to differences in process magnitudes and timing. This information is important in understanding basin hydrology, as streamflow is a summation of lateral fluxes from slopes.

The presence of ice-rich layers blocking soil interstices encourages runoff by restricting drainage. Runoff exhibits a two-layer flow system consisting of quickflow (pathways in the porous organic layer, pipes, rills, and interconnected depressions) and slowflow (pathways in underlying mineral soils and highly decomposed and compacted peat). Quickflow controls the shape and timing of the runoff hydrograph, which is influenced by properties of hillslope wetness and organic layer thickness. Recession analysis revealed variable source areas for runoff generation and highlighted the role of wetness-controlled hydrologic connectivity of a slope segment.

Results from this thesis have implications for water resource inventories and predicting hydrologic behaviour of subarctic, subalpine hillslopes.

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The friendship of Richard Petrone, Tim Griffis and Mark "Rocco" Stradiotto has provided me the distractions to maintain my sanity and put a smile on my face. If only I could wind the clock back four years and have that much fun again....... There have been so many people whom have been more than friends over the last four years. It is hard to name them all, yet Sarah and Heather come to mind as I write this. Around the lab, Augustine, Mark and Erica have provided for a comfortable, relaxed and engaging atmosphere. All my love to Krysha Dukacz, my greatest friend and someone whom I'm sure is glad to see this thesis complete. My family is the rock upon which I stand. I would like to thank my mother and father, Mary and John Carey, from the bottom of my heart for their continued support and love.

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PREFACE

The thesis is made up of a collection of papers which have been published or submitted for publication. The study problems, objectives and relationship among the papers are described in the introduction. The research papers include the following:

Chapter Two: Carey SK, Woo MK. 1998. Snowmelt hydrology of two subarctic slopes, southern Yukon, Canada. *Nordic Hydrology* **29**: 331-346.

Chapter Three: Carey SK. Woo MK. 1999. Hydrology of two slopes in subarctic Yukon, Canada. *Hydrological Processes* **13**: 2549-2562.

Chapter Four: Carey SK, Woo MK. Variability of hillslope water balance, Wolf Creek basin, subarctic Yukon. *Hydrological Processes*. Submitted.

Chapter Five: Carey SK. Woo MK. Slope runoff processes and flow generation in a subarctic, subalpine catchment. *Journal of Hydrology*. Submitted.

Appendix A: Carey SK, Woo MK. 2000. The role of soil pipes as a slope runoff mechanism, subarctic Yukon, Canada. *Journal of Hydrology*, **223**: 206-222.

While all papers were co-authored with the research supervisor Dr. Ming-ko Woo, the first author and candidate conducted the actual research involving problem formulation, literature review, data collection, field work and writing. Dr. Woo provided guidance on the direction of the research, critiqued all the papers and provided editorial advice. Several figures were also hand-drawn by Dr. Woo. References at the end of chapters conform to the journal specifications.

CHAPTER ONE

INTRODUCTION

1.1 BACKGROUND

The subarctic region covers approximately 30 % of Canada, lying between the closed canopy boreal forest in the south and the treeless arctic tundra in the north. It is characterized by open-canopied coniferous forests or by patches of open canopied forests and tundra. Permafrost is widespread in subarctic soils (Zoltai *et al.*, 1988). The term subarctic has the connotation of a northern cold environment, while for mountain areas, the terms alpine and subalpine are applied. In practical terms, subarctic environments can range from flat to alpine terrain.

Our hydrologic understanding of the subarctic with strong relief is poor as there has been little research within this environment. Table 1.1 summarizes important hydrological studies in the subarctic subalpine zone. It is noteworthy that most of this research has been conducted in the hilly region of central Alaska, but the applicability of principles found there to the more rugged terrain in the western Canadian subarctic has not been evaluated. Also included in Table 1.1 are studies from tundra areas that have particular value to subarctic, subalpine hydrology.

Hydrological research in the subarctic has generally taken two primary approaches: 1. plot studies of hydrological processes, and 2. analyses of streamflow which tends to integrate most processes operating within the basin. At the plot scale, winter studies have highlighted the role of forests on snow interception, sublimation and accumulation patterns (Pomeroy *et al.*, 1999). Strong winter temperature gradients between the ground and the atmosphere induce upward vapour fluxes from frozen organic soils, rendering them desiccated and porous (Santeford, 1979*a*). To study snowmelt, energy balance approaches are frequently used (Price and Dunne, 1976; Eaton and Wendler, 1982), and more recently, the role of forests on altering melt has been highlighted (Pomeroy *et al.*, 1999; Giesbrecht and Woo, *in press*, Woo and Giesbrecht, *in press*). Infiltration and percolation of meltwater into frozen soils, both permanent and seasonal, has been measured, stressing the role of ice-filled pores on impeding subsurface movement of water (Kane *et al.*, 1981; Kane and Stein, 1983). Plot studies of runoff show that slopes with surface organic mats contribute large portions of their snowpack to runoff while plots on slopes without permafrost produce little or no runoff (Slaughter and Kane, 1979; Kane *et al.*, 1981; Chacho and Bredthauer, 1983). The existence of ice-rich layers at the interface of the organic and mineral soils is cited as a principal cause of lateral runoff during melt (Kane *et al.*, 1981). Water storage and transmittance properties of surface organic layers on poorly drained slopes has received particular attention in subarctic and tundra environments as saturated hydraulic conductivity is often orders of magnitude greater than the underlying mineral soils (Santeford, 1979*b*; Hinzman *et al.*, 1993; McNamara *et al.*, 1998; Quinton and Marsh, 1999). During the summer, studies of evapotranspiration have been particularly sparse. Except for the work of Granger (1999), evapotranspiration studies in the subarctic have focussed upon the flat areas surrounding the Hudson's Bay Lowlands (e.g. Lafleur *et al.*, 1992) or the northern limit of the boreal forest zone, which has vegetation characteristics similar to the southern subarctic (e.g. Blanket *et al.*, 1997).

At the basin scale, Slaughter *et al.*, (1983) observed that catchments underlain predominantly by permafrost produce flashier flows and greater runoff responses than those underlain with seasonal frost only. Dingman (1966; 1971) inspected streamflow hydrographs and noted fast responses and long recessions compared with temperate zones, which was explained by the slow release of water from organic soils. Chacho and Bredthauer (1983) tested this explanation by comparing streamflow characteristics with runoff from lysimeters on permafrost slopes. Dingman (1973) hypothesized that runoff rates and pathways are dependent on the elevation of the water table on poorly drained permafrost slopes, invoking the partial source area concept of runoff generation (Betson 1964). By evaluating streamflow chemistry and isotopes during freshet, Gibson *et al.*, (1993) concluded that only permafrost-underlain slopes contribute water to runoff. Water balance studies within the subarctic, subalpine areas are scarce. Wright (1979) studied lichen tundra underlain by permafrost near Schefferville, Quebec, and noted that confinement of infiltrated water sustained wet conditions in the active layer, encouraging groundwater flow and providing wet conditions that maintained evapotranspiration at or near the equilibrium rate. In relatively flat terrain, Metcalfe and Buttle (1999) analysed differences in sub-basin water balance in a boreal forest catchment with discontinuous permafrost and noted large differences in fluxes controlled by variations in snowpack properties, rainfall characteristics, thaw depths and storage characteristics.

1.2 Research Issues

There are several areas where understanding of hydrological processes in subarctic, subalpine environments are inadequate.

1) *Spatial Variability*. These environments consist of a mosaic of landscape types as aspect and elevation strongly influence microclimate. This in turn affects vegetation, the presence or absence of permafrost, and the magnitude and timing of hydrological processes. To date, there has been no

attempt to quantify the variability in water balances at a hillslope or sub-basin scale. Understanding spatial variability in water balance components is important in evaluating basin water balance while providing insight into the factors (e.g. vegetation, permafrost) that cause this variability.

2) *Process Interactions*. Most studies have either examined individual processes or streamflow. Thus, although information has been gained on specific processes (e.g. infiltration into frozen soils), the interaction among processes (e.g. the role of soil moisture on evapotranspiration) has been mostly neglected. The examination of processes within a water balance framework must be completed to analyse interactions and place them in a seasonal context to obtain greater understanding of the annual water cycle.

3) *Hillslope-stream linkages*. Typically, hillslope hydrology has been concerned with the processes of runoff generation, yet the hillslope is an ideal scale to bridge the gap between the plot and basin scale as streamflow is the summation of lateral fluxes from slopes. As yet, there has been no attempt to link process operating on hillslopes to streamflow in subarctic, subalpine environments. The processes of runoff generation have also not been investigated. Understanding the mechanisms whereby water moves from the hillslope to the stream network is critical for determining runoff timing, magnitude and streamflow chemistry.

1.3 OBJECTIVES

In light of these issues, the general objective of this thesis is to improve understanding of hillslope hydrology in a subarctic, subalpine setting. This goal is attained through several specific objectives:

1) *Measure the seasonal water balance on subarctic, subalpine hillslopes.* Identifying the range in annual water balance on hillslopes is critical for water resource inventories, indicate what slopes contribute lateral fluxes for streamflow, and improve understanding the basin water balance.

2) *On these hillslopes, resolve the critical factors that produce variability in the timing and magnitude of hydrological processes.* Process studies will reveal the factors that cause the variability in hillslope hydrology and allow the extension of principles (both conceptual and numerical) to other locations. By studying all major processes operating on subarctic hillslopes, interactions and feedbacks among processes will also be observed.

3) *Determine the processes of runoff generation and water delivery to streams on subarctic slopes.* Water discharged from the base of hillslopes results in streamflow, and the resultant pathways affect runoff timing, magnitude and streamflow chemistry. Knowledge of runoff generation and contributing areas for runoff allows coupling of the hillslope to the stream.

1.4 STRUCTURE OF THE THESIS

To address these objectives, the thesis has been written as a series of papers which form the various chapters. A literature review and description of the methods pertaining to each aspect of the thesis are contained within each paper.

Chapter two, entitled *Snowmelt hydrology of two subarctic slopes, Southern Yukon, Canada,* examines hydrological processes during the snowmelt period on two subalpine slopes that have large contrasts in their physical setting. Process studies highlight the causes for variability in timing and magnitude of water balance components during this important period.

Chapter three, entitled *Hydrology of two slopes in subarctic Yukon, Canada,* expands on chapter two by examining the hydrological processes over the entire year on the same two slopes. The annual rhythm of the hydrological cycle is characterized, emphasizing interactions among processes and causes of the dramatic differences between hillslope hydrology.

Chapter four, entitled *Variability of hillslope water balance, Wolf Creek basin, subarctic Yukon,* provides snowmelt and summer water balances for four distinct hillslopes within the study basin. The variability in the hillslope water balance provides the basis to ascertain the factors that control the variation in the timing and magnitude of hydrological processes on subarctic, subalpine slopes.

Chapter five is entitled *Slope runoff processes and flow generation in a subarctic, subalpine catchment.* Processes of runoff generation are investigated while hydrograph recession analysis is used at the hillslope scale to evaluate runoff contributing areas.

Chapter six summarizes and discusses the key findings and presents suggestions for future research.

Appendix A, entitled *The role of soil pipes as a slope runoff mechanism, subarctic Yukon, Canada,* is included as it provides details on the hydrology of soil pipes and preferential flow on one of the study slopes, and places their role within a larger framework. It is included as an appendix due to the smaller scope of this study compared with the thesis chapters. Preferential flow is a widely cited phenomenon on permafrost hillslopes, and Appendix A fills an important gap in understanding the pipeflow processes.

1.5 References

Betson RP. 1964. What is watershed runoff?. *Journal of Geophysical Research* **69**: 1541-1552.

Blanken PD, Black TA, Yang PC, Neumann HH, Nesic Z, Staebler R, den Hartog G, Novak MD, Lee X. 1997. Energy balance and canopy conductance of a boreal aspen forest: Partitioning overstory and understory components. *Journal of Geophysical Research*, **102 (D24)**: 28915-28927.

Chacho Jr. EF. 1990. Water and suspended solids discharge during snowmelt in a discontinuous permafrost basin. In *Proceedings of the 5th Canadian Permafrost Conference*, National Research Council, Ottawa, pp. 167-173.

Chacho Jr. EF, Bredthauer S. 1983. Runoff from a small subarctic watershed, Alaska. In *Proceedings, Fourth International Conference on Permafrost*, National Academy Press, Washington, D.C., pp. 115-120.

Dingman SL. 1966. Characteristics of summer runoff from a small watershed in central Alaska. *Water Resources Research* **2**: 751-754.

Dingman SL. 1971. *Hydrology of Glenn Creek watershed, Tanana basin, Central Alaska*. U.S. Army Cold Region Research Engineering Laboratory Research Report 297, Hanover, New Hampshire, p. 112.

Dingman SL. 1973. Effects of permafrost on stream flow characteristics in the discontinuous permafrost zone of central Alaska. In *North American Contribution to Second International Conference of Permafrost*, National Academy of Sciences, Washington, D.C., pp. 447- 453.

Eaton F, Wendler G. 1982. The heat balance during the snow melt season for a permafrost watershed in Interior Alaska. *Archives for Meteorology, Geophysics, and Bioclimatology, Series A* **31**: 19-33.

Gibson JJ, Edwards TWD, Prowse TD. 1993. Runoff generation in a high boreal wetland in northern Canada. *Nordic Hydrology* **24**: 213-224.

Giesbrecht, MA, Woo MK. 2000. Simulation of snowmelt in a subarctic spruce woodland: 1. Open woodland model. *Water Resources Research* (in press).

Granger RJ. 1999. Partitioning of energy during the snow-free season at the Wolf Creek research basin. In *Wolf Creek Research Basin: Hydrology, Ecology, Environment*, Pomeroy JW, Granger RJ. (eds.), Environment Canada, Saskatoon, pp. 33-43.

Hinzman LD, Kane DL, Everett KR. 1993. Hillslope hydrology in an Arctic setting. In *Proceedings, Sixth International Conference on Permafrost*, South China Press, Beijing, pp. 257-271.

Kane DL, Stein J. 1983. Water movement into seasonally frozen soils. *Water Resources Research* **19**: 1547-1557

Kane DL, Bredthauer SR, Stein J. 1981. Subarctic snowmelt runoff generation. In *Proceedings of the Specialty Conference on The Northern Community*, Vinson TS (ed.), ASCE, Seattle, Washington, pp. 591-601.

Lafleur PM, Rouse WR, Carlson DC. 1992. Energy balance differences and hydrologic impacts across the northern treeline. *International Journal of Climatology* **12**: 193-203.

McNamara JP, Kane DL, Hinzman LD. 1998. An analysis of streamflow hydrology in the Kuparuk River Basin, Arctic Alaska: a nested watershed approach. *Journal of Hydrology* **206**: 39-57.

Metcalfe RA, Buttle JM. 1999. Semi-distributed water balance dynamics in a small boreal forest basin. *Journal of Hydrology* **226**: 66-87.

Price AJ, Dunne TD. 1976. Energy balance computations on snowmelt in a subarctic area. *Water Resources Research* **12**: 686-694.

Pomeroy JW, Hedstrom N, Parviainen J. 1999. The snow mass balance of Wolf Creek, Yukon: Effects of snow sublimation and redistribution. In *Wolf Creek Research Basin: Hydrology, Ecology, Environment*, Pomeroy JW, Granger RJ. (eds.), Environment Canada, Saskatoon, pp. 15-30.

Quinton WL, Marsh P. 1999. A conceptual framework for runoff generation in a permafrost environment. *Hydrological Processes* **13**: 2563-2581.

Santeford HS. 1979*a*. Snow soil interactions in interior Alaska. In *Proceedings, Modeling of Snow Cover Runoff*, Colbeck SC, Ray M. (eds.), Cold Regions Research and Engineering Laboratory, Hanover, NH, pp. 311-318.

Santeford HS. 1979*b*. Toward hydrologic modeling of the black spruce/permafrost ecosystem of interior Alaska. In *Proceedings 30th Alaska Science Conference*, Fairbanks, Alaska, 9 pp.

Slaughter CW, Kane DL. 1979. Hydrologic role of shallow organic soils in cold climates. In *Proceedings, Canadian Hydrology Symposium 79 - Cold Climate Hydrology*, National Research Council of Canada, Ottawa, pp. 380-389.

Slaughter CW, Hilgert JW, Culp EH. 1983. Summer streamflow and sediment yield from discontinuous-permafrost headwater catchments. In *Proceedings, Fourth International Conference on Permafrost*, National Academy Press, Washington, D.C., pp. 1172-1177.

Woo MK, Giesbrecht MA. 2000. Simulation of snowmelt in a subarctic spruce woodland: 1. tree model. *Water Resources Research* (in press).

Wright RK. 1979. Preliminary results of a study on active layer hydrology in the discontinuous zone at Schefferville, Nouveau-Quebec. *Geographic Physique et Quaternaire* **33**: 359-368

Zolatai SC, Tarnocai C, Mills GF, Veldluis H. 1988. Wetlands of subarctic Canada. In *Wetlands of Canada*. Ecological Land Classification Series, No. 24. Sustainable Development Branch, Environment Canada, Ottawa and Polyscience Publications Inc. Montreal. pp 57-96.

Table 1.1 Important hydrological studies with implications for subarctic, subalpine catchments.

Reference	**Key Findings / Importance**	**Location**
Chacho and Bredthauer, (1983)	Lysimeters on hillslopes exhibited similar response time and recession characteristics to streamflow, yet runoff ratios are much lower.	Glenn Creek, AK 65°N, 147°W
Dingman (1966)	Examined streamflow characteristics of a subarctic basin and observed that recessions are distinct from temperate regions and can be quantified using a single recession constant.	Glenn Creek, AK 65°N, 147°W
Dingman (1971)	Comprehensive review of the hydrology of the Glenn Creek Watershed.	Glenn Creek, AK 65°N, 147°W
Dingman (1973)	Proposed mechanism whereby runoff is controlled by the position of the water table on permafrost-underlain organic-covered hillslopes.	Glenn Creek, AK 65°N, 147°W
Eaton and Wendler (1982)	Calculated surface energy balance of a melting snowpack and quantified sublimation during melt.	Glenn Creek, AK 65°N, 147°W
Gibson *et al.*, (1993)	Stable isotopes used to investigate runoff generation. Snowmelt contributions were secondary to active-layer storage contributions throughout freshet. Only permafrost-underlain hillslopes contributed to runoff.	Manners Creek, NWT 61°N, 121°W
Granger (1999)	Variability in energy balance components and evapotranspiration in a subarctic basin quantified. Forests have a different water use strategy than those in the boreal zone.	Wolf Creek, YK 60°N, 135°W
Hinzman *et al.*, (1993)	Organic layer controls runoff rates. Preferential flow features termed water tracks rapidly convey water from hillslopes to the stream.	Imnavait Creek, AK 68°N, 149°W
Kane *et al.*, (1981)	Differences between infiltration into organic-covered permafrost soils and permafrost-free mineral soils studied. Ice-rich layers block percolation on organic-covered hillslopes.	Fairbanks, AK 65°N, 147°W
Kane and Stein (1983)	Infiltration rates related to soil moisture conditions near the ground surface in non-permafrost soils. Horton's equation used to model infiltration.	Fairbanks, AK 65°N, 147°W

Table 1 *continued*. Important hydrological studies with implications for subarctic, subalpine catchments.

Reference	**Key Findings / Importance**	**Location**
McNamara *et al.*, (1998)	In a tundra environment, examined runoff characteristics using a nested watershed approach. Stormflow characteristics were presented and related to hillslope and basin characteristics. Importance of organic soils and permafrost highlighted.	North Slope, AK > 68°N
Pomeroy *et al.*, (1999)	Factors of snow sublimation and redistribution reviewed using four-year record in a subalpine basin. Forests canopies affect sublimation rate and end-of-winter snow water equivalent.	Wolf Creek, YK 60°N, 135°W
Price and Dunne (1976)	Related melt at the snow surface to runoff measurements beneath the snowpack.	Schefferville,Qc 55°N, 67°W
Quinton and Marsh (1999)	Examined processes of runoff generation on tundra slopes. Linked runoff source areas to the elevation of the water table.	Trail Valley Creek, NWT 68°N, 133°W
Santeford (1979*a*)	Organic layers become desiccated over the winter due the temperature-induced upward vapour flux from the porous soils.	Fairbanks, AK 65°N, 147°W
Santeford (1979*b*)	Organic covered permafrost hillslopes control streamflow runoff in subarctic environments. Modelled organic layer as a bucket to explain response.	Fairbanks, AK 65°N, 147°W
Slaughter *et al.*, (1983)	Permafrost-underlain terrain more responsive to precipitation than permafrost-free terrain. Compared streamflow characteristics of permafrost-dominated and a permafrost-free first order basin.	Carribou-Poker Creeks, AK 65°N, 147°W
Woo and Giesbrecht, *in press*	Effect of forest shadows and tree wells on snowpack energy balance quantified.	Wolf Creek, YK 60°N, 135°W
Wright (1979)	Development of active layer strongly influenced by conditions at freeze-up. Permafrost confined water, increasing soil moisture to enhance evaporation.	Schefferville, Qc 55°N, 67°W

CHAPTER TWO

SNOWMELT HYDROLOGY OF TWO SUBARCTIC SLOPES, SOUTHERN YUKON, CANADA

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and

Ming-ko Woo

A modified version of this chapter appears as: Carey SK, Woo MK. 1998. Snowmelt hydrology of two subarctic slopes, southern Yukon, Canada. *Nordic Hydrology* **29**: 331-346.

ABSTRACT

Large quantities of water are discharged from subarctic basins during snowmelt season. Runoff contributing areas as well as timing and magnitude of meltwater generation from different slopes are highly variable. Two slopes in the lower Wolf Creek basin, southern Yukon, were studied in 1997. The south-facing slope has a dense aspen forest that is leafless in the melt period (April - May) and was underlain by seasonal frost. The north-facing slope has open stands of spruce and an organic layer that rests on mineral soils with permafrost. In 1997, snowmelt is advanced by over 10 days on the south slope, which receives more solar radiation than the north aspect. All meltwater on the south slope infiltrates the frozen silt without generating runoff. By the time significant melt events occur on the north slope, the frost and snow are gone from the south. Meltwater is able to infiltrate the frozen organic soil but deep percolation is prevented by the ice-rich substrate. Lateral flow begins after the organic layer is saturated, with much runoff along intermittent rills fed by diffuse and pipe flows. Rills and pipes are interconnected but the drainage network and runoff contributing area change depending on the disposition of the snow as well as water and frost table positions relative to local topography. Contrasts between the north and south slopes have important implications on direct runoff generation during the melt period. Situations similar to the study site can be found elsewhere in subarctic North America and the observed processes have a bearing upon hydrological modelling for the subarctic environment.

2.1.0 INTRODUCTION

In the subarctic environment, snowmelt yields large amounts of water to streamflow. Heterogeneity within subarctic basins produces considerable variations in the flow characteristics of headwater catchments. For instance, Slaughter *et al.*, (1983) noted that flashier flows are associated with catchments with large percentages of permafrost. Research in central Alaska (Dingman, 1973; Kane *et al.*, 1981) reported the occurrence of snowmelt runoff from organic-covered, permafrost slopes. On the other hand, Kane *et al.*, (1981) and Gibson *et al.*, (1993) observed that non-permafrost slopes contributed little to streamflow. An examination of subarctic hillslopes with different orientations indicates physical contrasts which should have a bearing upon their hydrological characteristics. In particular, slopes of north and south aspects are highly different in: 1) radiation regime, 2) vegetation (spruce vs. birch-aspen forests), 3) soil surface cover (organic mat vs. leaf litter) and 4) frozen ground conditions (permafrost vs. seasonal frost). These features affect the snowmelt and meltwater delivery processes on the slopes.

Investigations of hydrological processes have been carried out in subarctic sites underlain by seasonal frost or permafrost, ranging in scale from experimental plots to entire drainage basins (Chacho and Bredthauer, 1983; Dingman, 1966; 1973; Kane *et al.*, 1981; Kane and Stein, 1983; Santeford, 1979*a;* Wright, 1979). The energy balance approach is frequently used to study snowmelt processes (Eaton and Wendler, 1982; Price and Dunne, 1976). Infiltration into frozen soils, be they seasonal frost or permafrost, has been measured in Central Alaska (Kane *et al.*, 1981; Kane and Stein, 1983). The hydrological behaviour of the organic layer on permafrost slopes has often been emphasized (Slaughter and Kane, 1979) particularly because most snowmelt runoff is transmitted down the slopes in these soils (*cf.* Hinzman *et al.*, 1993).

Results from these studies have improved our understanding of subarctic hydrology. On the other hand, there are no direct inter-slope comparisons regarding their hydrological responses to the snowmelt and runoff processes. Since drainage basins comprise slopes of different orientations, contrasts in quantity, timing and mechanisms of water delivery from the snowpack greatly influence the slope runoff contribution to the basins. This study focuses upon the comparative issue and addresses the spatial differences in the snowmelt hydrology of subarctic slopes.

2.2.0 STUDY AREA AND METHODS

Two slopes in the subalpine zone of the Wolf Creek basin (61°31'40" N, 135°31'14"W), located 15 km south of Whitehorse, Yukon Territory, Canada, were selected for this study (Figure 2.1). These slopes are separated by the Wolf Creek valley, which at this point is 120 m wide and at an elevation of 1175 m above mean sea level. The north-facing (N) slope has a gradient of 0.18 and is underlain by boulder-clay (till), capped by an organic layer consisting of peat, lichens, mosses, sedges and grasses. Permafrost occurs at depths of 0.6-1.2 m except at the slope bottom where seasonal frost depth exceeds 2 m. Most parts of this slope are covered by Labrador tea and willow shrubs, growing to a height of 1.6 m. Black spruce trees, approximately 20 m apart, reach heights of 9 m. The south-facing (S) slope has a gradient of 0.65. A leaf litter of 0.04 m overlies silty materials. No permafrost is found on this slope but winter frost penetrates to a depth of 1.5 m. A dense aspen forest, approximately 11 m high, covers the slope but the absence of leaves during the snowmelt period allows much radiation to reach below the canopy.

Wolf Creek basin has a subarctic continental climate. The long-term climatic record at Whitehorse yields a mean January temperature of -21°C and a July mean of 15°C. Temperatures during the snowmelt months of April and May average 1°C and 7°C respectively. Mean annual precipitation is 260 mm with about 55 per cent falling as rain.

The study was carried out from April 8 to May 28, 1997. Net radiation on each slope was measured by REBS net radiometers set at 1 m above the snow, air temperature and relative humidity were measured with a Vasiala HMP35CF sensor housed in a Gill shield. Ground temperatures were measured by two arrays of thermocouples on each slope, down to a maximum depth of 1.5 m. Soil moisture was determined by TDR probes (MoisturePoint) buried at 0.05, 0.1, 0.2, 0.4 and 0.6 m in the N-slope, and at an additional depth of 1 m on the S-slope. Wind speed was measured by a Met One 014A anemometer set at 1.5 m above snow surface. These data were recorded by Campbell Scientific CR10 dataloggers. In addition, an eddy correlation device (Gill propeller anemometer Model 27106 and a Campbell Scientific FW3 fine wire thermocouple) was mounted at a 6 m height to provide information for the calculation of sensible heat flux over the N-slope. Twelve snow pits were excavated in early April on each slope to determine snowpack thickness and density. Daily ablation was obtained by measuring the rate of snow surface lowering and converting the depth change into water equivalent using density measurements. During the snowmelt period, five flumes installed at the outlet and along three rills on the N-slope provided measurements of runoff.

2.3.0 Hydrological Components

2.3.1 Snow cover and conditions during melt

On April 8 when this study began, the snow was isothermal on the S-slope and some melting occurred despite daily mean air temperature being <0°C. Mean snow depth on this date was 0.44

m which yields a water equivalent of 137±14 mm. The snow cover consisted of three visible layers (Figure 2.2): a thin surface crust, a dense wet middle layer and a less dense layer that included some depth hoar. In contrast, temperature of the snow on the N-slope was <0°C. Its mean snow depth was 0.71 m, representing a water equivalent of 187±24 mm (Figure 2.2). Differences in end-of-winter snow storage on these slopes may be partly due to early melt on the S-slope, though a snow survey conducted in April 1996, before any ripening started, indicated that snow depth on the N-slope was 0.7 m compared to 0.5 m on the S-slope. A lower winter accumulation may be due to higher interception and sublimation losses from the S-slope.

Air temperature during the study period was similar for both slopes (Figure 2.3). Daily means fell from near freezing to -5°C on April 18 and then rose sharply over the next five days, corresponding with rapid depletion of the snow on the S-slope. Afterwards, temperature remained around +5°C and then rose to >10°C when most of the snow on the N-slope also disappeared. Figure 2.3 shows the positive values of daily net radiation (i.e. for the daylight hours only) for both slopes which increased rapidly when the air temperature rose above 0°C. Values for the S-slope exceeded that of the N-slope throughout the melt season, with larger differences occurring after the snow was gone from the S-slope. Wind speed was low along the sheltered valley during most of the study period. Only once did the daily average wind speed at 1.5 m above the snow exceeded 2 m s^{-1} (Figure 2.3).

2.3.2 Snowmelt processes

The energy balance of a melting snowpack is

$$Q_M = Q^* + Q_H + Q_E + Q_P + Q_S \tag{2.1}$$

where Q_M is the melt energy, Q^* net radiation, Q_H and Q_E sensible and latent fluxes, Q_P melt energy convected by rainfall, Q_S change in heat storage within the snow and all the terms are in MJ m^{-2}d^{-1}. In this study, Q_M was determined using ablation measurements:

$$Q_M = L_f \rho M \tag{2.2}$$

where M (in m d^{-1}) is measured melt, L_f and ρ are the latent heat of fusion (MJ kg^{-1}) and the density of snow (kg m^{-3}). Only trace precipitation events occurred during the melt period and Q_P is ignored. The snow on the S-slope was already ripened and Q_S was unimportant, but on the N-slope, the cold content in the snow was calculated by

$$Q_S = c\sum_{i=1}^{3} d_i\left(T_i - T_0\right) \tag{2.3}$$

where c is the volumetric heat capacity of ice (MJ m^{-3}C^{-1}), T_o=0°C, d_i is snow thickness and T_i is snow temperature for layer i, for a total of three layers (Figure 2.2). The low wind speed near the snow surface complicated calculations of Q_H and Q_E using aerodynamic equations. These terms were lumped as turbulent flux (Q_T), obtained as a residual of the energy balance

$$Q_T = Q_H + Q_E = Q_M - Q^* - Q_S \tag{2.4}$$

The magnitude of these energy balance terms was assessed (Table 2.1).

Figure 2.3 shows the rates of snow cover depletion as represented by snow depth changes on the two slopes. The main melt period for the S-slope ended on April 26 except for several patches at upslope locations which lingered until May 8. The snow on the N-slope did not become isothermal until April 21 but after that, the main melt period continued until between May 10 to May 18 when the snow cover disintegrated into patches. Where there were icings in the hollows or along the rills, melting and erosion of the icings followed.

Net radiation was 45 MJ and Q_M was 50 MJ for the main melt period on the S-slope, leaving a net turbulent flux contribution of about 5 MJ. Such a dominance of radiation melt in a subarctic setting has also been reported elsewhere (Price and Dunne, 1976). On the N-slope, Q^* was 11 MJ for the early melt period while Q_M was only 3 MJ. About 1 MJ was spent in eliminating the cold content, leaving a net Q_T of -6 MJ, with the negative sign indicating energy consumption by sublimation. In the main melt period, Q^*=58 MJ, Q_M=74 MJ and $Q_S = 0$ because the snow was isothermal. Turbulent flux was -17 MJ but it combined Q_H and Q_E which were likely to be opposite in flux direction. A rough assessment of Q_H was made for a large area of the slope using an eddy-correlation device placed at 6 m above the snow. Although the footprint of the sensors included some trees and protruding shrubs, the computed $Q_H = 33$ MJ was considered a reasonable approximation of sensible heat flux to the snow. Then, from eq. 4, $Q_E = -50$ MJ which is equivalent to 18 mm of sublimation. This magnitude is similar to the atmospheric water loss of 21 mm reported by Kane *et al.*, (1981), and 23 mm reported by Eaton and Wendler (1982) for the melt season on a north-facing slope near Fairbanks, Alaska.

2.3.3 Infiltration and ground temperature

In early April, ground temperature was <0°C at all sites except the lower S-slope which had seasonal frost extending down to 1.2 m only. The presence of a snow cover insulated the ground from conductive heating, but between April 5 and 8, ground temperature at the top 1-m layer of the upper S-slope rose abruptly by >2°C (Figure 2.4). Similar rapid temperature rises were observed in the top layer of the N-slope sites but at later dates (April 22-24). This phenomenon may be attributed to the re-freezing of infiltrated meltwater, releasing heat to warm the soil (Woo and Heron, 1981).

Temperature profiles of the S-slope indicate that only seasonal frost was present (Figure 2.4). The 0°C isotherm could not descend from the surface while there was a snow cover, but ground thaw proceeded upward from the base of the seasonal frost. When the snow melted, two-sided thawing of the frost occurred. Ground frost was eliminated by April 25 at the lower slope, and the loss of frost at the upper slope was delayed until May 22 as the snow remained until May 8. Throughout the study period, no surface runoff was observed on the S-slope, suggesting unimpeded infiltration into the frozen soil (Kane and Stein 1983). The presence of seasonal frost on this slope had little noticeable effect on infiltration.

Ground thaw on the N-slope began on May 12-13 with thawing proceeding downward only (Figure 2.4), a feature typical of soils underlain by permafrost. By May 30, thaw depth reached only 0.35-0.4 m. The slow thaw may be attributed to the insulating properties of the organic cover and the abundance of ground ice which accentuates the 'zero curtain' effect, thereby retarding the descent of the thawing front (Woo and Xia, 1996). Unlike the frost of the S-slope, the ice-rich zone at the organic-mineral interface limited meltwater percolation. This is evidenced by the continual saturation of the organic layer without a corresponding rise in the moisture content of the mineral soil below (Figure 2.5).

2.3.4 Soil moisture changes

Liquid water content in the soil changes in response to (1) phase change as ground ice melts, (2) infiltration into the soil, and (3) lateral water movement. Figure 2.5 shows selected profiles of liquid water content in the soils as measured by time domain reflectometry.

Between April 13 and 20, the lower site of S-slope showed an increase in liquid water content at its top 0.4 m layer. Since the soil was still frozen to a depth of 1 m, such an increase was due to infiltration of meltwater into frozen soil. After snow disappeared, the ground thawed quickly and further liquid water increase may be related to ground ice melt. This contribution to liquid water content was uniform throughout the profile, preserving the curvature established earlier in the season. After May 10, surface soil moisture decreased in response to evaporation and percolation to lower levels indicated by comparing the profiles of May 20 and 25. It is unclear whether the moisture increase at depth >0.8 m was caused by drainage from upslope.

On the upper S-slope, an increase in liquid water content near the ground surface between April 20 and 30 was likely due to infiltration into the frozen soil. Subsequent infiltration and percolation sustained a gradual moisture increase down to 0.4 m until May 10. After that, the snow was depleted; infiltration ceased but percolation continued to raise the moisture level down to a depth of 0.5 m. The moisture profiles at this site are similar to those presented by Kane and Stein (1983) for a non-permafrost slope in Fairbanks, Alaska.

Liquid water content of the organic layer on the lower N-slope did not increase notably in the melt period, suggesting a low infiltration rate which was a consequence of the considerable amount of ice that filled the pores in the organic layer. Thawing of the surface zone released much of the ice as liquid water (May 15). On the upper N-slope, the liquid water content of the top layer rose between April 25 and early May when the ground was still frozen. The ease of infiltration into frozen organics at this site is due to the large amount of air space partly created by the vapourization of the ice in the pores during the winter, a phenomenon documented elsewhere in the subarctic (Santeford, 1979*a;* Smith and Burn, 1987). Despite the high moisture content in the organic layer, percolation to the mineral substrate is hindered by an ice-rich zone at the interface of the organic and mineral soils, a feature also noted by Slaughter and Kane (1979). Consequently, once the storage deficit in the organic soil is satisfied, additional water input generates runoff (Santeford, 1979*b*) which moves downslope as surface and subsurface flows.

2.3.5 Slope runoff

At no time was the soil on the S-slope saturated and surface runoff did not occur. In contrast, many parts of the N-slope were saturated by the infiltrated meltwater and slope runoff was prevalent. When runoff started on May 12, much snow remained and water was issued at the base of the pack to topographic lows and the rills. As melt continued, water moved along preferential paths consisting of intermittent rills and pipes in the organic layer (*cf.* Gibson *et al.*, 1993; Hinzman *et al.*, 1993). Together, they formed a network which conveyed the flow rapidly downslope (Carey and Woo, 2000). In addition, subsurface diffused flow began when the saturated ground thawed, facilitated by the high hydraulic conductivity of the organic layer ($\approx 1 \times 10^{-4}$ ms^{-1}). Both the pathways and the source areas of runoff changed as new surface and subsurface hydrological connections were made. Snowpack location, micro-topography, water table position and thaw depth all played a role in altering the source area and the pattern of flow.

Flumes set up at the base of N-slope along three rills recorded snowmelt runoff between May 12 and 24 (Figure 2.6). The areas draining into flumes are not constant as their source areas are altered by changes in the drainage network. In relative terms, flumes N-1 has the largest catchment area and N-2, the least. All flow data showed prominent diurnal cycles, with daily peaks occurring around 2130 h and daily lows at 0800 h (Pacific Standard Time). Runoff rose to high values on May 12 and then declined during the next three cool and overcast days. Increased temperature and radiation led to the second high flow period on May 18 but the flow at N-2 already dropped because of snow depletion in its catchment area. May 18 produced the highest flow at N-3 but N-1 peaked on May 21. The discordance in flow among the rills marked the differences in their water sources as the meltwater contribution area changed. As the snow cover on the slope disintegrated, the diurnal flow cycles disappeared and runoff at N-3 ceased on May 24. The recurrence of flow at N-2 on May 23 was not related to snowmelt but was due to replenishment by subsurface flow through a deepening thawed zone.

The timing of flow lagged behind the initiation of major snowmelt by at least two weeks (Figure 2.3). Runoff began on May 12 when about half of the storage on N-slope and all the snow on S-slope was depleted. Such time lag is likely the result of meltwater movement and storage in the snowpack, the replenishment of water storage in the organic layer, seepage to the pipes and rills and transmission of water to the lower slope by surface and subsurface drainage. The highly variable source area and the changing pattern of the drainage network complicate the timing of meltwater delivery.

To assess the contribution of snowmelt to runoff, a water balance for the melt period is estimated. Since the total area draining into the three flumes cannot be determined accurately, two additional flumes were placed higher up the slope to catch the flow from upslope. The difference in the flows between the upper and the lower flumes (bottom of Figure 2.3) is considered to represent the contribution from the area that drained between these flumes (inset of Figure 2.6). Another difficulty is the missing data for the period from 1300 h, May 18 to 1500 h, May 20. A crude interpolation was made based on flow relationship with air temperature. Using a runoff contributing area of 840 m^2 (Figure 2.6), the magnitudes of the water balance components for the melt season (April 21 to May 25) are calculated. The snow cover (187 mm water equivalent) lost 18 mm to sublimation, but icing on the slope provided 19 mm of water. Runoff was 155 mm, leaving 33 mm as water added to the active layer storage. The amount of meltwater that recharged the active layer is comparable to the 30 mm for a Fairbanks site (Santeford, 1979*b*) and it is a plausible value in terms of change of moisture content measured in the organic zone (Figure 2.5). The runoff ratio, being 155/(187-18+19), was 0.8. This value is higher than the ratio (0.5) obtained by Kane *et al.*, (1981) for an experimental plot near Fairbanks where the snowfall is substantially less; but it falls within the range of snowmelt runoff ratios (0.6-0.8) obtained from runoff plots set up near Yellowknife, Northwest Territories, Canada (Landals and Gill, 1972).

2.4.0 DISCUSSIONS AND CONCLUSIONS

The selection of two subarctic slopes close to each other eliminates meso-scale variabilities so that differences in their hydrological behaviour can be attributed to local considerations alone. Slope aspect and its attendent physical attributes such as incoming radiation, vegetation, frost and soil development, lead to large contrasts in snowmelt and meltwater delivery. Results from one field season clearly demonstrate large differences between a north and a south facing slope in relation to the magnitude, timing and the mechanisms of melt and runoff processes.

There are several major contrasting hydrological features between the two slopes.

(1) At the end-of-winter, there is less snow on the south than the north slope and this seems to be a function of interception and sublimation loss over the winter or the pre-melt period. By the time snowmelt begins in earnest on the north slope, the snow has gone from the opposite slope even though the latter has a much higher density (though leafless) of trees.

(2) The snow on the south slope ripens more than 10 days earlier than the north slope. Radiation melt is important and this may be one factor that accentuates the contrasts in the timing and the magnitude of snowmelt between slopes with different amounts of radiation receipt.

(3) Seasonal frost is found in the south slope but this is no deterrent to meltwater infiltration. The frost, >1 m thick, thaws out quickly at the end of snowmelt.

(4) Snowmelt does not generate direct runoff on the south slope. However, runoff may occur in certain years when pre-melt soil moisture conditions are high, as was observed in a similar environment near Fairbanks, Alaska (Slaughter, pers. comm).

(5) The north slope is underlain by permafrost with a thin active layer, insulated by an organic soil with living moss and lichen. Ground thaw below the organic zone is retarded by an ice-rich layer at the organic-mineral soil interface.

(6) Meltwater infiltration into frozen organic soil depends on the ice content in the soil pores. Percolation to the mineral soil is limited by frost and runoff is generated when the organic layer reaches saturation.

(7) Delivery of water downslope is facilitated by piping in the organic layer and by rills that channel the flow. The source area and the drainage network for slope runoff are highly dynamic, changing according to the snow disposition and the convergence or divergence of surface and subsurface flows.

(8) There is a considerable time lag (>10 days) between the initiation of snowmelt and the occurrence of runoff on the north slope because of the various storages (snow, organic soil) and delays in water transmission (water movement in snow, surface and subsurface flows).

From the perspective of subarctic catchment hydrology, it is important to understand the mechanisms of meltwater production and delivery, and to quantify the timing, magnitude and spatial distribution of slope runoff. Characteristics such as those manifested by the south slope will dominate catchments with only seasonal frost while hydrologic features present on the north slope will prevail in basins with considerable permafrost. Thus, the former type of basins yields little runoff in the melt period but the latter group will have flashy flows. This is verified by Slaughter *et al.*,'s (1983) comparison of streamflow for three headwater basins in the Caribou-Poker Creeks watershed, Central Alaska, where the catchments with larger percentages of permafrost have higher and flashier discharges. In terms of snowmelt modelling for small catchments, particularly for distributed models (Kirnbauer *et al.*, 1994), proper account should be taken of the spatial contrasts in slope runoff production in order to generate streamflow that reflects the variable source areas. Macro-hydrological models which lump large areas into single units should not ignore the intra-unit variability in the timing, the area and the magnitude of meltwater production as demonstrated in this study. This, however, involves scaling considerations (Blöschl and Sivapalan, 1995) which cannot be easily resolved.

2.5.0 REFERENCES

Blöschl, G. and Sivapalan, M. (1995) Scale issues in hydrological modelling: a review, *Hydrological Processes*, *Vol. 9*, pp. 251-290.

Carey, S.K., and Woo, M.K. (2000) The role of soil pipes as a slope runoff mechanism, subarctic Yukon, Canada, *Journal of Hydrology*, *Vol. 203*, pp. 206-222.

Chacho, Jr. E.F., and Bredthauer, S. (1983) Runoff from a small subarctic watershed, Alaska. In: Proceedings, Fourth International Conference on Permafrost, National Academy Press, Washington, D.C., pp. 115-120.

Dingman, S.L. (1966) Characteristics of summer runoff from a small watershed in central Alaska, *Water Resources Research*, *Vol. 2*, pp. 751-754.

Dingman, S.L. (1973) Effects of permafrost on stream flow characteristics in the discontinuous permafrost zone of central Alaska. In: North American Contribution to Second International Conference of Permafrost, National Academy of Sciences, Washington, D.C., pp. 447- 453.

Eaton, F., and Wendler, G. (1982) The heat balance during the snow melt season for a permafrost watershed in Interior Alaska, *Archive for Meteorology, Geophysics, and Bioclimatology, Series A*, *Vol. 31*, pp. 19-33.

Gibson, J.J., Edwards, T.W.D., and Prowse, T.D. (1993) Runoff generation in a high boreal wetland in northern Canada, *Nordic Hydrology, Vol. 24*, pp. 213-224.

Hinzman, L.D., Kane, D.L., and Everett, K.R. (1993) Hillslope hydrology in an Arctic setting. In: Proceedings, Sixth International Conference on Permafrost, South China Press, Beijing, pp. 257-271.

Kane, D.L., Bredthauer, S.R., and Stein, J. (1981) Subarctic snowmelt runoff generation. In: T.S. Vinson (Ed.), Proceedings of the Specialty Conference on The Northern Community, ASCE, Seattle, Washington, pp. 591-601.

Kane, D.L., and Stein, J. (1983) Water movement into seasonally frozen soils, *Water Resources Research*, *Vol. 19*, pp. 1547-1557.

Kirnbauer, R. Blöschl, G. And Gutknecht, D. (1994) Entering the era of distributed snow models, *Nordic Hydrology* , *Vol. 25*, pp. 1-24.

Landals, A.L., and Gill, D. (1972) Differences in volume of surface runoff during the snowmelt period: Yellowknife, Northwest Territories. In: The Role of Snow and Ice in Hydrology Symposium, Banff, Alberta, IAHS Pub. No. 107, pp. 927-941.

Price, A.J., and Dunne, T.D. (1976) Energy balance computations on snowmelt in a subarctic area. *Water Resources Research*, *Vol. 12*, pp. 686-694.

Santeford, H.S. (1979*a*) Snow soil interactions in interior Alaska. In: Colbeck, S.C. and Ray, M. (Ed.) Proceedings, Modeling of Snow Cover Runoff, Cold Regions Research and Engineering Laboratory, Hanover, NH, pp. 311-318.

Santeford, H.S. (1979*b*) Toward hydrologic modeling of the black spruce/permafrost ecosystem of interior Alaska. In: Proceedings 30th Alaska Science Conference, Fairbanks, Alaska, 9 pp.

Slaughter, C.W., and Kane, D.L. (1979) Hydrologic role of shallow organic soils in cold climates. In: Proceedings, Canadian Hydrology Symposium 79 - Cold Climate Hydrology, National Research Council of Canada, Ottawa, pp. 380-389.

Slaughter, C.W., Hilgert, J.W., and Culp, E.H. (1983) Summer streamflow and sediment yield from discontinuous-permafrost headwater catchments. In: Proceedings, Fourth International Conference on Permafrost, National Academy Press, Washington, D.C., pp. 1172-1177.

Smith, M.W. and Burn, C.R. (1987) Outward flux of vapour from frozen soils at Mayo, Yukon, Canada: results and interpretation, *Cold Regions Science and Technology*, *Vol. 13*, pp. 143-152.

Woo, M.K. and Heron, R. (1981) Occurrence of ice layers at the base of High Arctic snowpacks, *Arctic and Alpine Research*, *Vol. 13*, pp. 225-230.

Woo, M.K., and Xia, Z.J. (1996) Effects of hydrology on the thermal conditions of the active layer, *Nordic Hydrology, Vol. 27*, pp. 129-142.

Wright, R.K. (1979) Preliminary results of a study on active layer hydrology in the discontinuous zone at Schefferville, Nouveau-Quebec, *Géographie Physique et Quaternaire, Vol. 33*, pp. 359-368.

Table 2.1: Snowmelt energy balance for North and South slopes, 1997

Energy Fluxes (MJ)	North slope April 9-20	North slope Apr. 21-May 18	South slope April 8-24
Melt energy (Q_M)	+3.0	+74.3	+50.3
Net radiation (Q^*)	+11.3	+57.6	+45.1
Turbulent fluxes (Q_T)	-5.9	-16.7	+5.2
Heat storage change (Q_S)	+1.2	0	0

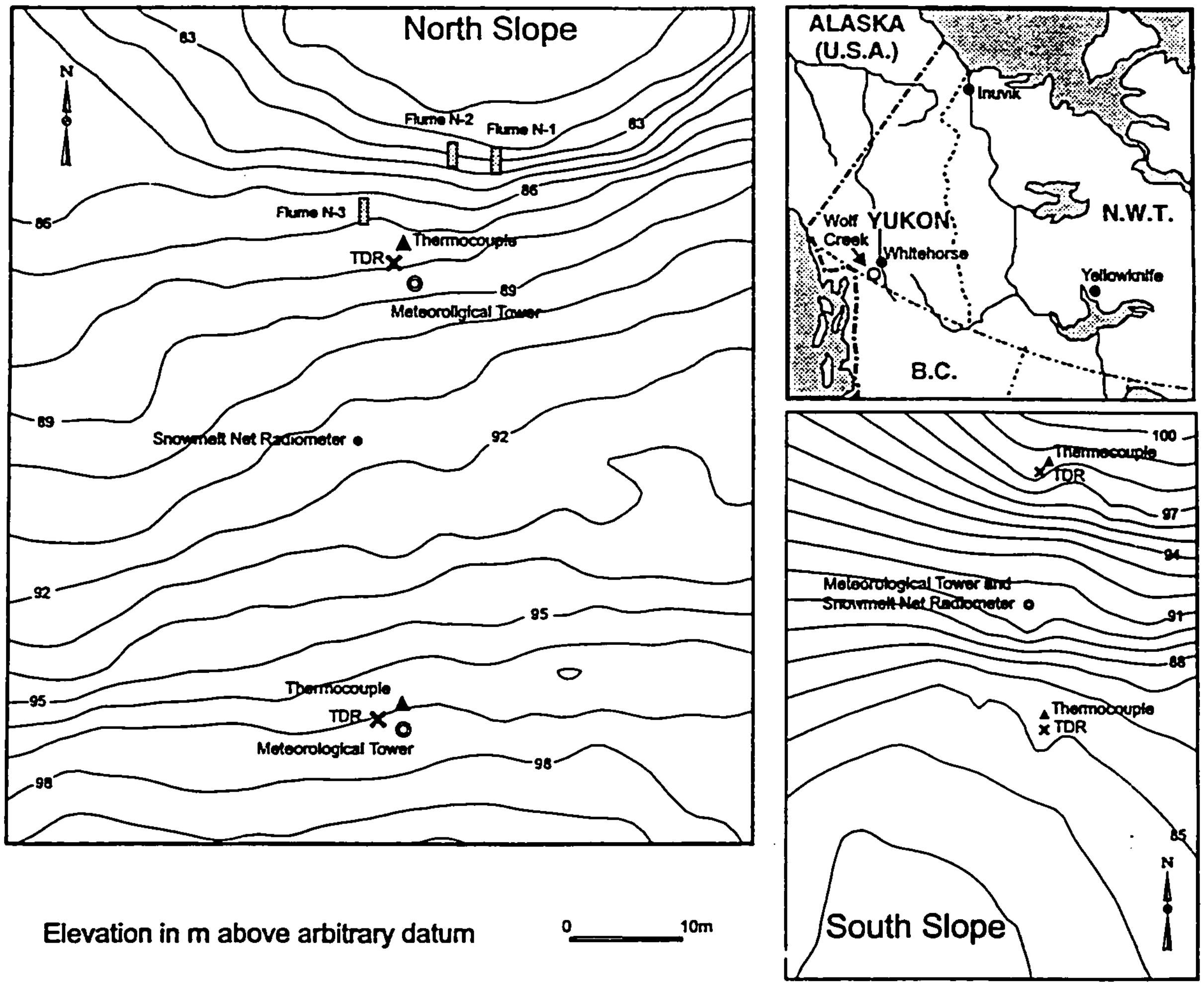

Figure 2.1 Topography and instrumentation of tow experimental slopes. Inset shows location of Wolf Creek Basin.

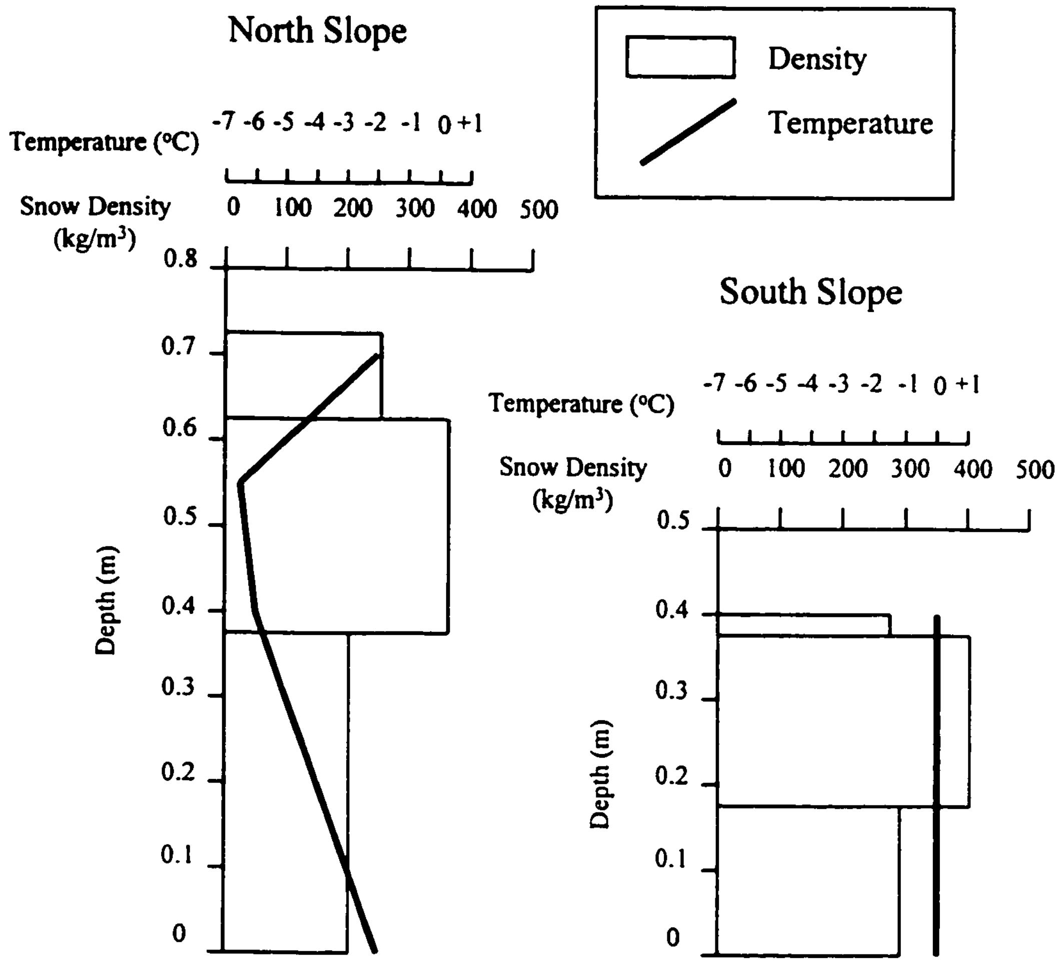

Figure 2.2. Vertical variation of temperature and density in typical snow profiles on North and South slopes as measured on April 8, 1997.

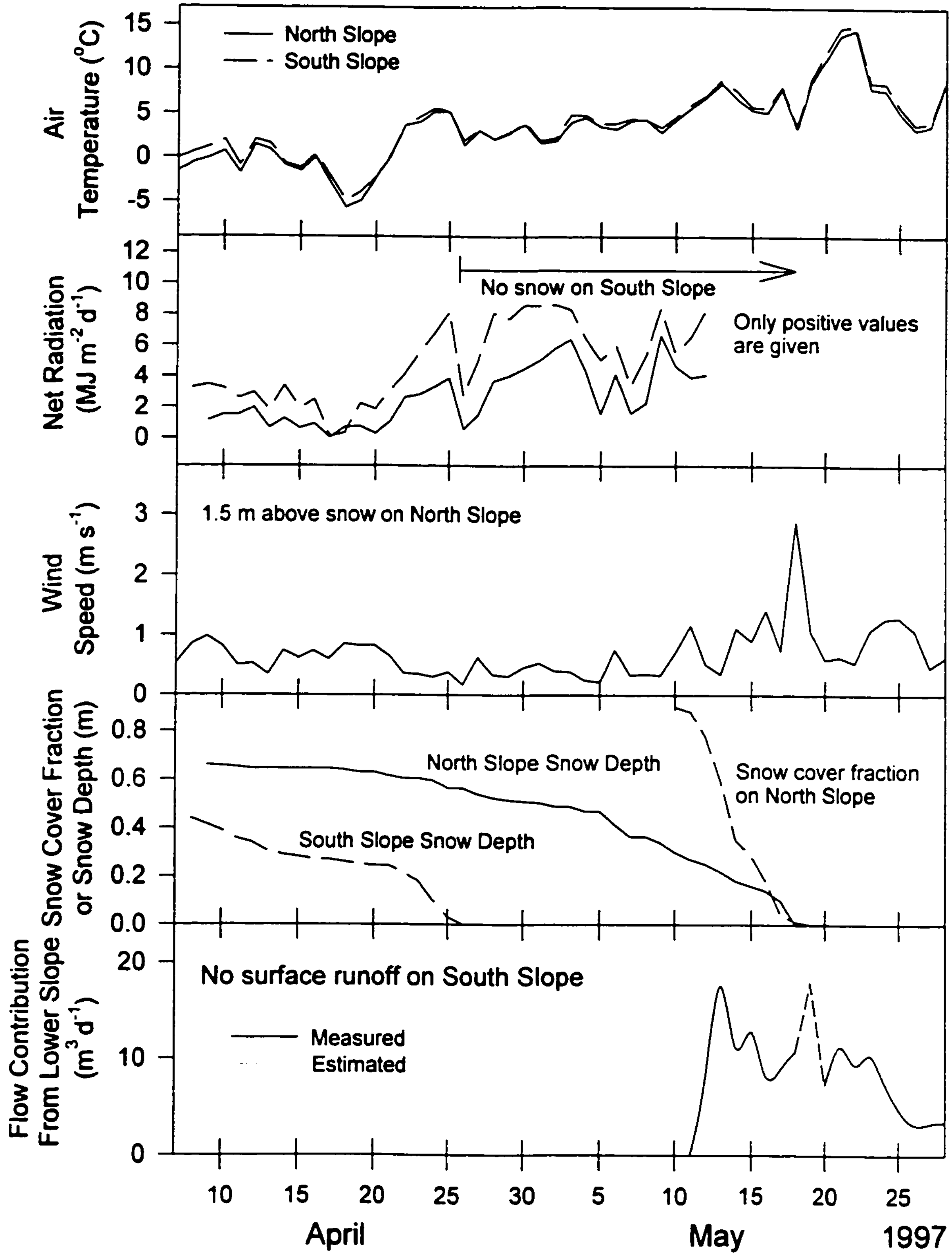

Figure 2.3. Mean daily air temperature, daily net radiation (positive values only), mean wind speed, snow depth and snow cover fraction, and daily runoff contribution from lower North slope.

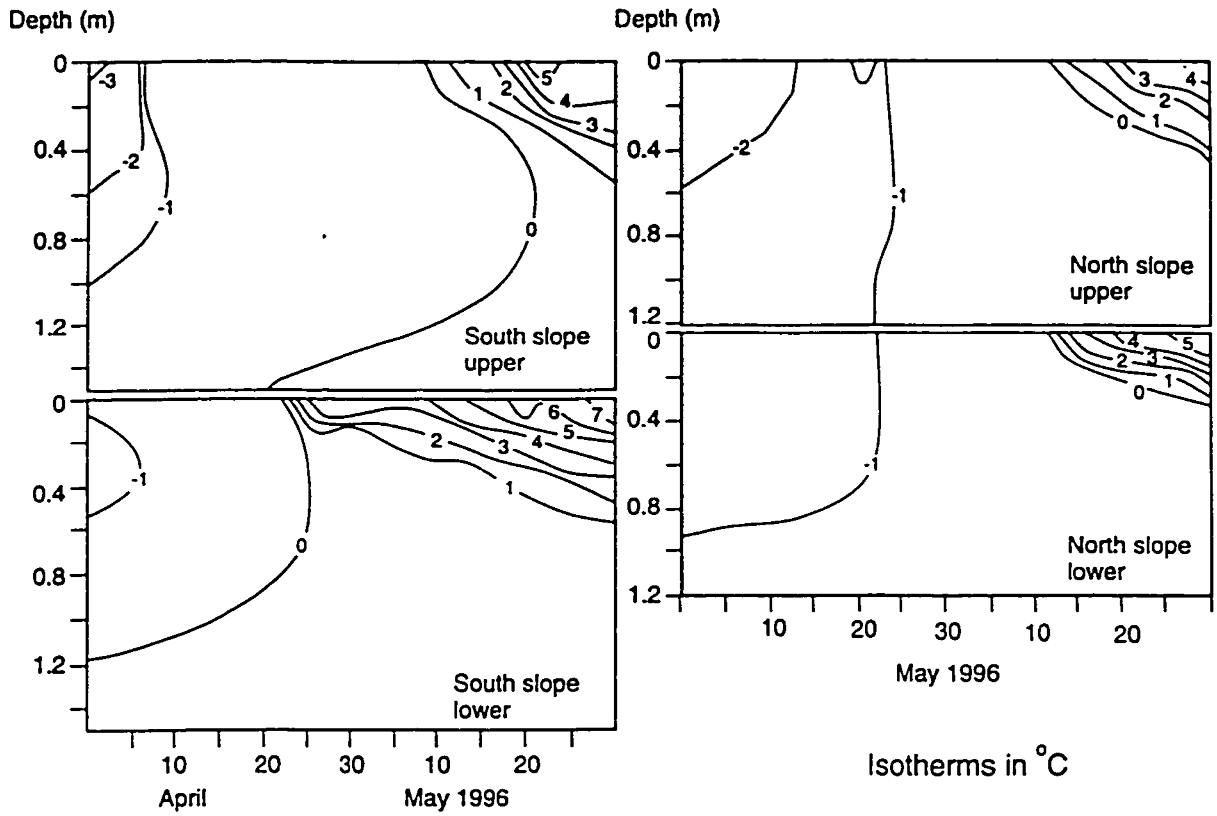

Figure 2.4 Temperature changes in the near surface zones at four plots. For each plot, the disappearance of its snow cover corresponded with the rise of surface temperature above 0°C.

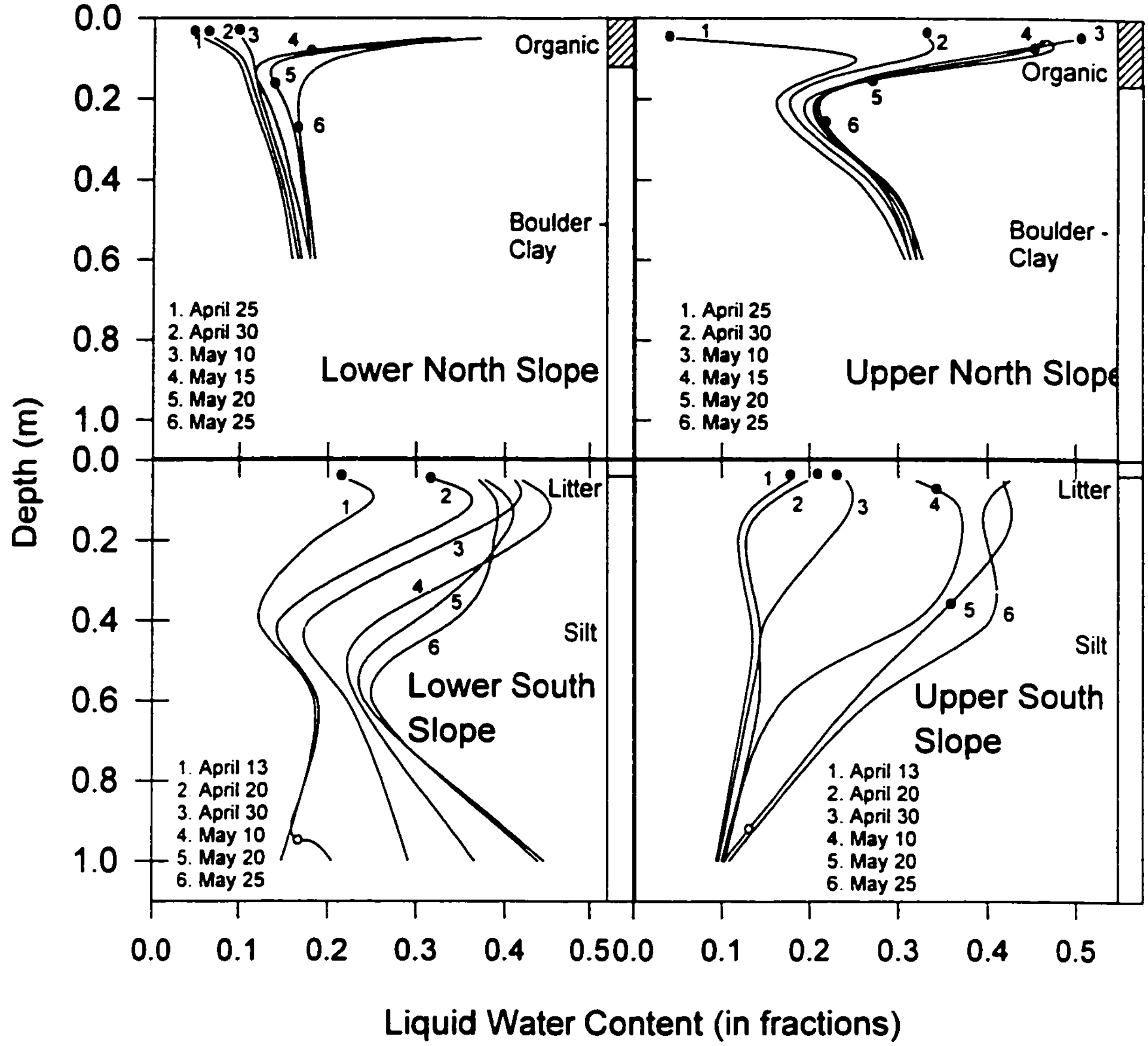

Figure 2.5. Vertical profiles of liquid water content at four plots on six dates in the study period, 1997. Vertical columns on the right of each panel show the soil stratigraphy. Black circles indicate position of the downward thawing front; shaded circles for the South slope profiles show the thaw boundary at the base of the seasonal frost. Profiles without any circle attached are frost-free.

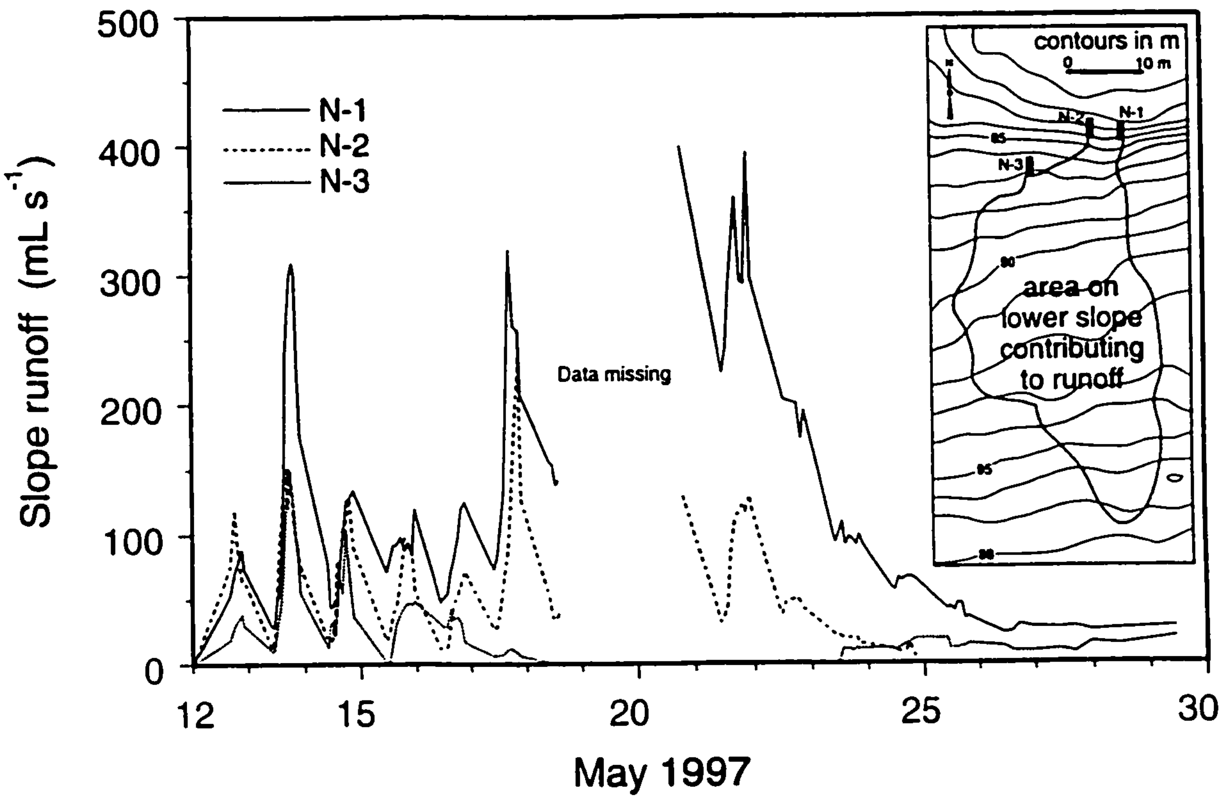

Figure 2.6 Runoff at flumes N-1, N-2 and N-3 located in rills at the base of the North slope. Inset shows the slope area that contributed to runoff differences between the upper and the lower flumes.

CHAPTER THREE

HYDROLOGY OF TWO SLOPES IN SUBARCTIC YUKON, CANADA

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Abstract

Two subarctic forested slopes in central Wolf Creek basin, Yukon, were studied in 1996-97 to determine the seasonal pattern of the hydrologic processes. A south-facing slope has a dense aspen forest on silty soils with seasonal frost only and a north-facing slope has open stands of black spruce and an organic layer on top of clay sediments with permafrost. Snowmelt is advanced by approximately one month on the South-facing slope due to greater radiation receipt. Meltwater infiltrates its seasonally frozen soil with low ice content, recharging the soil moisture reservoir but yielding no lateral surface or subsurface flow. Summer evaporation depletes this recharged moisture and any additional rainfall input, at the expense of surface or subsurface flow. The North-facing slope with an ice rich substrate hinders deep percolation. Snow meltwater is impounded within the organic layer to produce surface runoff in rills and gullies, and subsurface flow along pipes and within the matrix of the organic soil. During the summer, most subsurface flows are confined to the organic layer which has hydraulic conductivities orders of magnitudes larger than the underlying boulder-clay. Evaporation on the north-facing slope declines as both the frost table and the water table descend in the summer. A water balance of the two slopes demonstrates that vertical processes of infiltration and evaporation dominate moisture exchanges on the south-facing slope, whereas the retardation of deep drainage by frost and by clayey soil on the permafrost slope promotes a strong lateral flow component, principally within the organic layer. These results have the important implication that permafrost slopes and organic horizons are the principal controls on streamflow generation in subarctic catchments.

3.1.0 INTRODUCTION

Large areas of the Subarctic are characterized by permafrost and seasonal frost. In addition to topography, microclimate, soil and vegetation, the presence of ground frost exerts a strong control on subarctic hydrology.

Previous subarctic research on surface and subsurface drainage has focused on hydrologic responses at the basin scale or has studied specific processes at the plot scale. From a basin perspective, Slaughter *et al.*, (1983) observed flashier flows in catchments largely underlain by permafrost whereas basins dominated by seasonal frost only produce moderate responses to rainfall. In Central Alaska, Chacho and Bredthauer (1983) concluded that non-permafrost areas yield virtually no near-surface runoff during the snowmelt or summer season, and that large portions of the watershed, including some permafrost slopes, do not contribute to streamflow. Dingman (1966, 1971, 1973) inspected hydrograph characteristics and proposed a mechanism whereby surface runoff occurs when high water tables, restricted from vertical drainage by permafrost, rise to intersect the ground surface during rainstorms. Slaughter and Kane (1979) linked hydrologic role of the shallow organic soils to basin responses in order to explain the shape of streamflow hydrographs. Gibson *et al.*, (1993) used isotopic and geochemical indicators to assess water sources and runoff contributing areas during snowmelt.

Plot studies have emphasized specific processes operating on permafrost or permafrost-free terrain at various times of the year. During snowmelt, lysimeters and runoff plots show that slopes with surface organic mats and permafrost substrates contribute large percentages of their snowpack to runoff while plots on slopes without permafrost produce little or no runoff (Landals and Gill, 1971; Kane *et al.*, 1981; Chacho and Bredthauer, 1983). The existence of ice-rich layers at the interface of the organic and mineral soils is cited as the principal cause of lateral runoff

(Kane *et al.*, 1981). The hydrologic role of surface organic layers on poorly-drained permafrost slopes has been investigated, emphasizing water storage and transmittance properties (Santeford, 1979*b*; Slaughter and Kane, 1979; Hinzman *et al.*, 1993; McNamara *et al.*, 1998). The infiltration of water into seasonally frozen soils (Kane *et al.*, 1981; Kane and Stein, 1983) shows considerable complexity of water movement conditions. Snowmelt energy balances have been performed in the Subarctic (Price and Dunne, 1976; Eaton and Wendler, 1982), yet there are few evaporation studies that apply to slopes where a mosaic of land-surface conditions complicates the evaporative regime.

Despite progress in basin and plot-scale hydrology, there remains a need to examine the subarctic hydrological processes at an intermediate scale and to investigate the relative importance of these processes during different times of the year. The hillslope offers an ideal scale to bridge plot and basin studies. The purpose of this study is to examine hydrologic processes of two subarctic slopes, one with permafrost and the other with seasonal frost only, for the period between snowmelt and autumnal freeze-back. Results from this study will advance understanding of water fluxes and pathways, permitting improvements in the prediction of hydrologic behaviour of subarctic drainage basins.

3.2.0 STUDY AREA

Two slopes approximately 120 m apart across a river valley were instrumented in the fall of 1995 (Figure 3.1). These slopes are located in Wolf Creek Basin (61°31'40" N, 135°31' 14"W, elevation 1175 m), 15 km south of Whitehorse, Yukon Territory, Canada. A north-facing (N) slope with an average gradient of 0.18 is underlain by boulder-clay of glacial origin, capped by an organic layer 0.05-0.3 m thick consisting of peat, lichens, mosses, sedges and grasses. All but the extreme footslope portion of this slope is underlain by permafrost at depths from 0.6 to over 2 m. Vegetation consists of Labrador tea, willow shrubs, black and white spruce with a stand density of 200 trees/ha and maximum heights exceeding 9 m. The south-facing (S) slope has a dense aspen forest approximately 11 m high. The soil consists of a thin leaf litter overlying homogeneous silt to a depth of > 2 m. The average slope gradient is 0.6 and no permafrost is found under this slope. Porosity, bulk density and hydraulic conductivity of the soil layers for the two slopes are shown in Table 3.1.

The study area experiences a subarctic continental climate with mean January and July temperatures of −21°C and 15°C at the Whitehorse airport. Mean annual precipitation is approximately 260 mm, half of which falls as rain.

3.3.0 METHODS

Hydrologic parameters were obtained at various points on the study slope. Horizontal hydraulic conductivity was determined for each soil layer on the N-slope using pumping tests conducted in wells and piezometers (Freeze and Cherry, 1979). For the S-slope, samples were returned to the laboratory and constant head permeameter tests were used to assess the vertical hydraulic conductivity. Porosity and bulk density were determined using gravimetric methods. Ice content was determined by chipping samples from different horizons and then thawing the samples, subtracting the unfrozen moisture content by time domain reflectometry (TDR). In addition, ice content was visually estimated from frozen samples chipped from the soil.

Two 6 m towers were erected on the N-slope, 30 m apart. The lower tower is located approximately 20 m from the slope base and is instrumented with an incident short wave pyranometer (Eppley) and a net radiometer (REBS), a Vaisala HMP35CF air temperature and relative humidity probe housed in a Gill shield, and wind speed sensors (Met-One) at 2 and 6 m. The upper tower is instrumented with a net radiometer (Middleton) and five wet and dry thermocouples at various heights to provide profile data for evaporation calculations. A 17 m tower is positioned at the approximate centre of the S-slope aspen forest. Above the canopy, the tower has a short wave pyranometer (Eppley) and a net radiometer (REBS), four wet and dry thermocouples and a wind sensor (Met-One). A temperature and relative humidity probe (Vaisala HMP35CF) and a net radiometer (Middleton) are also located at 3 m beneath the canopy. Ground temperatures within each slope were measured using seven arrays of thermocouples in the N-slope and three in the S-slope down to a maximum depth of 1.5 m. A Texas Instruments tipping bucket rain gauge was installed in the valley at equidistance between the slopes. Signals from all sensors were sampled every 60 seconds and recorded on Campbell Scientific CR-10 data-loggers at 30 minute intervals.

Snow conditions on each slope prior to melt were surveyed using a Mount Rose snow sampler, supplemented by 12 snow-pits. During melt, snow ablation was monitored daily at several sites on each slope. Detailed description of the field methods is presented in Carey and Woo (1998).

Soil moisture was measured using TDR probes (MoisturePoint) buried at 0.05, 0.1, 0.2, 0.4 and 0.6 m at three sites on the N-slope and at two sites on the S-slope with an additional sensor at 1 m. Water tables were monitored continuously with electronic water level recorders (10 turn potentiometers with floats) at three sites on the N-slope and manually measured daily at an additional 12 wells. The S-slope had four wells dug to a depth of 2 m without reaching the saturated zone. Surface runoff on the N-slope was monitored using nine small portable flumes to define intra-slope drainage and to assess the changing source areas during the snowmelt period. On occasion, some flumes were moved in order to determine the variable nature of snowmelt runoff. Surface runoff was measured approximately once every 3 hours during melt. Through the summer, water level was recorded electronically every half-hour behind the outlet of a flume at the slope base, and discharge was determined from a calibration relating spot measurements of discharge to water level. Subsurface flow was computed on a daily basis using data obtained from 15 wells. By monitoring the frost and water table positions, the flow was calculated by dividing the soil into n layers and using Darcy's equation:

$$Q_{SS} = \sum_{i=1}^{n} - K_i w(d_i - z_i)(\Delta h/\Delta x) \quad (3.1)$$

where K_i is the hydraulic conductivity of a soil layer, $(d_i$-$z_i)$ is the thickness of the thawed zone for the soil layer, $\Delta h/\Delta x$ is the hydraulic gradient estimated between a pair of wells and w is the width of the slope, taken to be 30 m at the study site.

Evaporation (E) for each slope was determined using the Bowen ratio approach augmented by lysimeters.

$$E = (Q^* - Q_g)(1 + \beta)\rho_w L_v \quad (3.2)$$

where E is evaporation, Q^* and Q_g are net radiation and ground heat flux, ρ_w is the density of water and L_v the latent heat of vaporization and the Bowen ratio, β, is

$$\beta = C_a \Delta T / (L_v \Delta q) \quad (3.3)$$

with C_a being the specific heat of air, and ΔT and Δq the difference in air temperature and in vapour pressure. Objective criteria given by Ohmura (1982) were used to remove spurious β values. Wherever these criteria were not met, a modified Priestley and Taylor (1972) approach was used

$$E = \alpha \frac{\sigma}{\sigma + \lambda} \frac{Q^* - Q_g}{\rho_w L_v} \tag{3.4}$$

where σ is the slope of the saturated vapour pressure curve , λ is the psychometric constant and α is the Priestley and Taylor parameter which relates actual to potential evaporation. The value of α is not static and is modified on a daily basis based on comparing the Bowen ratio and Priestley and Taylor evaporation. Daily lysimeter measurements and soil moisture budgets were also used as a check on the values thus calculated.

3.4.0 HYDROLOGICAL PROCESSES

Hydrological processes were studied between April 8 and September 22, 1997. Additional data from April 6 to August 12, 1996 were used to reinforce the findings of the 1997 field season.

3.4.1 Snow Storage and Snowmelt

Snow constitutes approximately half of the annual water input to the study area. Prior to melt, the mean snow water equivalence was 187 ± 24 mm on the N-slope and 160 ± 14 mm on the S-slope. Greater snow accumulation on northerly exposures was also observed in the spring of 1996 (0.7 m vs. 0.5 m), and can probably be explained by higher winter sublimation rates on the S-slope. Carey and Woo (1998) presented a detailed study of snowmelt, energy balances and runoff hydrology during this period for both study slopes. On April 8, the S-slope snowpack was already isothermal at 0°C while the snow on N-slope retained considerable cold content. Although the air temperature was similar above both slopes, radiation receipt was higher on the S-slope and melt proceeded rapidly until the slope became snow-free on April 26 (Figure 3.2 *a-c*). The rapid increase in melt rate after April 21 was due to heat advection from the bare areas. Snowmelt on the N-slope lagged behind that of the S-slope by approximately one month, not reaching isothermal state until April 21 when average daily air temperatures rose above freezing. The main melt period continued until mid-May when the snow cover disintegrated into patches. Using the energy balance approach, Carey and Woo (1998) determined that 18 mm and 10 mm of the snowpack sublimated on the North and South slope respectively, a result consistent with the findings of Eaton and Wendler (1982).

3.4.2 Icing

The role of icing on hillslope hydrology has received much less attention than large icings found in northern river channels (Woo, 1986). Small icings in tributary valleys have been observed (Gibson *et al.*, 1993) and studied by Reedyk *et al.*, (1995) in Manners Creek, N.W.T, a small basin with characteristics similar to the present study area. There, icing accounted for 20 % of surface water storage in the catchment, and supplied 6 % of total streamflow for the spring of 1990. On the scale of the N-slope, icing in gullies and rills was observed in both 1996 and 1997. In late September 1997, icing formed when warm groundwater exfiltrated and froze in gullies due to low air temperatures. Hydrologically, icing represents an over-winter surface storage of groundwater which is released gradually after snowmelt. For the winter of 1996-1997, icing accounted for approximately 10 % of surface water storage (19 mm of water) as determined from surface mapping, and prolonged the runoff in gullies after snowmelt, until the slope became ice-free on May 28. Icing in topographic lows also blocks drainage routes in the spring, retarding the delivery of runoff downslope.

3.4.3 Rainfall

At the study site, rain fell on 70 of the 170 days during the study period, depositing 150 mm to the slopes (Figure 3.3*a*). May accounted for only 2 % of the seasonal total. June and July were the wettest months with 59 mm and 58 mm of rain respectively, compared with only 22 mm in August and 10 mm in September. The largest event on June 5 added 21 mm of rain in 17 hours. Between June 28 and July 9, rain fell on all but one day. Most precipitation events were of convectional origin, usually lasting less than 1 hour in the late afternoon or early evening. According to the Whitehorse precipitation records, 1997 was a normal rainfall year.

3.4.4 Snowmelt Infiltration and Surface Runoff

At no time on the S-slope in both 1996 and 1997 was surface runoff observed. The high infiltration capacity of these silty soils combined with unsaturated conditions, low ice content and an absence of segregated ice layers allowed water to move into and percolate unrestricted down the frozen soil column. Figure 3.4 shows the rapid increase in S-slope soil moisture content at the upper and lower slopes over approximately one month. Although not observed, surface runoff may occur in years with a high pre-melt soil moisture content and/or rapid melting of a thick snowpack.

On the N-slope, meltwater infiltration into the frozen organic layer was uninhibited until this layer reached saturation on May 11. Lateral runoff then began because percolation into the boulder-clay substrate was restricted by the presence of an ice-rich layer (reaching 90% by volume) at the organic-mineral interface (Figure 3.2*d*). Santeford (1979*b*) obtained similar relations for permafrost slopes near Fairbanks, Alaska, and developed a conceptual bucket-type model for organic soils where runoff occurs only after the winter-desiccated organics are saturated. Runoff increased rapidly on the N-slope to reach a maximum on May 13 of 18 mm per day. Diurnal cycles were prominent, with maximum flow occurring at 2130 and minimum at 0800 Pacific Standard Time. Water issued from the base of the snowpack along gullies, surface rills and depressions as well as in soil pipes developed at the organic-mineral interface (Carey and Woo, 2000). Standing water occurred in hollows that gradually drained as thawing proceeded. Surface flow ceased on May 25 when downslope water movement was confined to the organic layer. For 1997, 155 mm of water, or approximately 80 % of the snowpack, was delivered as runoff during this period. This value lies within the range of runoff coefficients found for other permafrost slopes and basins (Landals and Gill, 1972; Hinzman *et al.*, 1993; McNamara *et al.*, 1997, 1998; Woo and Young, 1997).

Streamflow contributing areas during melt are dictated by runoff from slopes underlain by permafrost. In both 1996 and 1997, streamflow did not begin in Wolf Creek until approximately one month after snowmelt terminated on the southerly exposures, but flow commencement corresponded with the main melt period on the N-slope. The dominance of permafrost slopes as runoff contribution areas was also suggested by Gibson *et al.*, (1993) using stable isotopes and geochemical methods for Manners Creek, N.W.T., a basin with discontinuous permafrost.

3.4.5 Subsurface Runoff

At no time during the 1996-1997 study period was the water table within 2 m of the surface on the S-slope. Preliminary geophysical investigations carried out in 1997 indicated that the water table was more than 6 m beneath the slope surface. A similar condition was documented in Manners Creek, (Gibson *et al.*, 1993) where the water table was >3.4 m below the non-permafrost slopes.

Subsurface inflow and outflow for the lower section of the N-slope (30 m wide by 45 m long) were studied. Typically, as summer progresses, the thawed zone in the active layer thickens, facilitating the transmission of water downslope (Woo and Steer, 1982). The organic layer generally has hydraulic conductivities (K) several orders of magnitude greater than mineral soils, increasing runoff rapidly when the phreatic surface rises into this layer. For the N-slope, average K is 2.5×10^{-4} m/s for the lower organic layer and 5×10^{-9} m/s for the boulder-clay. Subsurface flow through this frozen mineral soil is considered negligible.

Subsurface flow in the N-slope is dominated by lateral drainage in the organic layer (termed quick-flow), leaving about 1 % of the water transmitted as slow-flow in the boulder-clay zone. Flow within the organic layer follows local zones of enhanced wetness, occupying topographic lows or moving along sinuous pathways (including pipes) that unite topographic concavities. Quick-flow within the organic layer on the N-slope, as experienced in subarctic catchments elsewhere (Slaughter and Kane, 1979) represents the main mode of subsurface flow on subarctic permafrost slopes. In the continuous permafrost zone near Inuvik, N.W.T., Quinton and Marsh (1998*a,b*) found that runoff is strongly controlled by the level of the water table within the organic layer, where hydraulic properties vary greatly with depth, and by hummocks which impede drainage.

Figure 3.3*b* shows the cumulative inflow, outflow and net groundwater flux from the experimental area. Inflow to the slope segment was highest in late May and June due to snowmelt water input, with inflow exceeding outflow to yield a net gain of 165 mm, equivalent to an average of 4 mm a day until mid-July. After that, the water table at most parts of the upslope dropped into the boulder-clay substrate, and outflow exceeded inflow by about 15 mm between mid July and mid August. After that, groundwater inflow and outflow were approximately equal. Taken over the entire field season, there was a net groundwater inflow of 245 mm and a net groundwater loss of 97 mm from the experimental slope segment.

3.4.6 Soil Moisture

Seasonal frost in the S-slopes did not limit vertical movement of water during snowmelt. Infiltration and percolation during and after snowmelt did not saturate the upper 2 m of the silt profile. Following the large recharge of soil moisture over the short melt period, moisture declined steadily throughout summer (Figure 3.4). Rainwater infiltration produced minor and brief effects on soil moisture enhancement in the top 0.1 m. Moisture differences between the upper and lower slopes may be attributed to larger evaporation losses at the lower segment which is subject to greater radiation receipt and more intense drying at the surface. Between May 18 and September 16, the lower slope experienced a moisture loss of 138 mm compared to 126 mm at the upper site.

For the N-slope, infiltration of meltwater into the frozen desiccated organic layer proceeded until its moisture deficit was satisfied and then percolation into the mineral soil was impeded by the presence of ice-rich zone at the organic/mineral interface. As the frost table descended past the icy zone, percolation ensued but the rate was limited by its very low hydraulic conductivity. Large contrast in K between the organic and mineral layers maintained a local perched water table in various parts of the hillslope, notably at the upslope locations (Figure 3.4). Above the frost table was a more continuous saturated zone which extended throughout the entire slope in early summer. By late summer, however, the saturated zone was found only at the footslope areas, being sustained by drainage from upslope.

3.4.7 Evapotranspiration

Evapotranspiration on the S-slope began earlier than the N-slope due to the earlier disappearance of its snow cover. Between late April and early May, evapotranspiration (Figure 3.3*c*) rates were low due to low radiation receipt and because bud-burst of the aspen forest did not occur until mid-May. On the N-slope, evaporation did not start in earnest until May 18, but evaporation from the bare ground cannot be separated from sublimation from the remaining patchy snowcover. From May 18 to early June, evapotranspiration from the N-slope was greater than the S-slope due to the ponding of water in surface depressions. Evapotranspiration on both slopes became similar by early June, proceeding at rates greater than equilibrium (Priestley and Taylor's $\alpha > 1.1$). As surface drying began on the N-slope in mid-July, its evapotranspiration rate dropped. During the rain free period (August 6 to 12) evapotranspiration for the N-slope was approximately 85 % of the S-slope ($\alpha < 1.0$ for N-slope). Taken over the entire snow-free period, evapotranspiration from the N-slope was 315 mm compared with 372 mm from the S-slope, as calculated by the Bowen Ratio method.

3.5.0 SEASONALITY OF PROCESSES

The annual rhythm of the hydrologic processes operating on the experimental slopes are conceptually represented in Figure 3.5*a,b*. Moisture exchanges in winter are confined to precipitation, sublimation and moisture migration within the soil and through the snowpack. On the N-slope (Figure 3.5*a*), vapour transport from the wet organics into the snowpack desiccates the surface layer (Santeford, 1979*a*). Migration of water within the soil to the mineral/organic interface also occurs, creating the ice-rich zone described above. Upward vapour transfer continues within the snow cover, as is evidenced by the formation of hoar at the snowpack base. On the S-slope (Figure 3.5*b*), dry conditions prior to freeze-back limit upward moisture migration and the formation of ice lenses.

Snowmelt was the single largest hydrological event of the year, releasing approximately half of the annual precipitation over a period of several weeks. The snow on the S-slope melts approximately one month earlier than on the north, releasing meltwater that infiltrates and percolates the frozen unsaturated silt and recharges the soil moisture reservoir. Surface runoff may occur only in years with exceptionally high antecedent moisture conditions to produce large ice contents, or in years when a deep snow cover is melted rapidly. Deep percolation to the local groundwater table is likely to be small, and so is evapotranspiration during snowmelt since the leaves have not year appeared in the aspen forest. While vertical processes prevail during snowmelt on the S-slope, lateral runoff dominates the N-slope with permafrost. During snowmelt, meltwater easily infiltrates the organic soil but the shallow icy zone impedes deep percolation. Water is impounded in depressions, rills and soil pipes or within the organic layer (Carey and Woo, 1998; 2000). The runoff contributing area is highly variable at this time as differential thaw, snowcover disposition and water table position affect the flow connectivity and runoff pathways. Wet surface conditions during this period also favours higher evaporation.

Vertical processes dominate the summer hydrology of the S-slope, with rainfall and evapotranspiration being the principal moisture exchanges. Evapotranspiration from the aspen forest consumes the rainfall input and depletes the soil water reservoir, causing a gradual drying of the soil profile. In contrast, both lateral and vertical exchanges of water are important on the N-slope. In early summer, upslope areas yield large volumes of meltwater to create high water tables in a shallow thawed zone. Runoff at this time is dominated by flow within the organic layer, and the maintenance of a perched water table above the mineral soils of low hydraulic conductivity provides a rapid means of water delivery to the stream. In the course of the summer, inflow from upslope decreases, the frost table descends deeper into the mineral soil and the water table drops accordingly. Lateral flow is not sustained except during heavy rainstorms. Evapotranspiration decreases as the organic surface layer is mostly covered by non-transpiring lichens and mosses, though the spruce and shrubs can withdraw water from deeper soil horizons.

During the fall, the freezing front descends on the S-slope. Senescence and leaf-fall along with a decline in available energy reduce evapotranspiration and surface-atmospheric moisture exchanges decline. The onset of a winter snow cover by mid to late October effectively shuts down hydrologic exchanges unless occasional mid-winter thaw occurs. On the N-slope, exfiltration of warm groundwater and expulsion of soil moisture due to two-sided freezing (MacKay, 1983) facilitate the formation of icing in gullies, rills and slope concavities. Full freeze-back of the active layer occurred by mid and late October in 1996 and 1997 respectively, providing only a short period for icing formation. Evaporation declines steadily until a winter snow cover is established. Then, sublimation of the snow becomes the principal processes of water loss from the subarctic forest (Pomeroy and Gray, 1995).

3.6.0 WATER BALANCE

The magnitude of various hydrologic processes operating on the two slopes are compared by performing water balance computations for the 1997 study period. The major components include

$$R + M + I \pm Q_S \pm Q_{SS} - V - ET = \Delta S \tag{3.5}$$

where ΔS is seasonal change in storage, R, M and I are rainfall, snowmelt and icing, Q_s is surface and Q_{ss} subsurface flow, V is sublimation from the snowpack and ET is evapotranspiration. The slopes showed significant differences in their water balance (Table 3.2). Subsurface inflow to the N-slope study segment was the largest single source of water, accounting for 40 % of the total water input. In contrast, there was no inflow or outflow to the S-slope, leaving snowmelt and rainfall as the only two water sources, each supplying approximately equal amounts of water. Atmospheric losses dominated the export of water on the S-slope as no runoff occurred. Surface and subsurface outflows were important loss terms for the N-slope, removing over 40 % of the water, with the remaining losses generated by evaporation and sublimation. Greater sublimation on the N-slope was due to its extended snow covered period, but evapotranspiration was larger on the S-slope.

Despite possible measurement and computation errors in the determination of the magnitudes of the various water balance components, the water gains and losses at the N-slope were roughly in balance, as shown by a low storage change term. On the other hand, the S-slope yielded a seasonal deficit of about 55 mm. This deficit may in part be caused by the uncertainty of the total frozen water content within the soil. In mid August 1997, there was on average 28 mm more soil water than was measured in mid-August 1996 in the S-slope. However, between mid-August and the freeze-back in October, 1996, about 70 mm of rain was recorded. This, combined with low evapotranspiration during the period, provided additional moisture storage in the soil from which water could be withdrawn in the 1997 field season.

3.7.0 DISCUSSIONS AND CONCLUSIONS

A study was carried out on two subarctic slopes for two years and the results can be generalized to the forested, discontinuous permafrost zone, where frost plays a special role in hydrology. Ice content in the frozen soil is a major consideration. For the S-slope where the seasonal frost has low ice content, snow meltwater and rainfall can infiltrate and percolate easily. Conversely, the ice-rich permafrost soil in the N-slope restricts drainage, maintaining a high water table in the thawed zone and promoting lateral runoff during snowmelt and high rainfall periods.

Poor drainage imposed by shallow frost favours the development of thick organic soil cover on the permafrost slope which strongly affects the hydrological processes. In summer, the non-transpiring lichens and mosses curtail evapotranspiration. More importantly, subsurface flow in the organic layer offers an efficient mechanism for water delivery downslope. Two flow systems then develop on the N-slope: (1) quick-flow along soil pipes and rapid matrix flow in the organic layer and (2) slow-flow in the mineral soil below, controlled by its very low hydraulic conductivities and a changing thawed depth in the active layer.

Water balances demonstrate that the S-slope without permafrost is dominated by vertical moisture exchanges, whereas permafrost slopes have both a vertical and a lateral component. Infiltration and percolation are not inhibited by soil properties nor by frost. Moisture from both snowmelt and rainfall is used to recharge the soil moisture reservoir annually, which in turn is used to satisfy evapotranspiration. The lack of runoff and groundwater recharge is confirmed, as evapotranspiration is of a magnitude similar to precipitation. On the other hand, the N-slope with permafrost experiences both vertical and horizontal fluxes. Lateral runoff, particularly in the melt season, is a major component of the water balance.

By contrasting the hydrology of two slopes, it is surmised that the non-permafrost slope does not produce stormflow to the stream. It is even doubtful whether the S-slope contributes baseflow to Wolf Creek because a preliminary geophysical survey indicates that the stream elevation is higher than the deep water table below this slope. Thus, the runoff source area to the stream is confined to the permafrost slopes. In terms of subarctic basins which are aggregations of heterogeneous hillslopes, findings confirm that the streamflow contributing areas are variable. Future research in this environment may benefit from linking slope processes to basin responses.

3.8.0 References

Carey, S.K. and Woo, M.K. 1998. 'Snowmelt hydrology of two subarctic slopes, Southern Yukon, Canada', *Nord. Hydrol.*, **29**, 331-346.

Carey SK, Woo MK. 'The role of soil pipes as a slope runoff mechanism, subarctic Yukon, Canada', *J. Hydrol.*, **203**, 206-222.

Chacho, Jr. E.F. and Bredthauer, S. 1983. 'Runoff from a small subarctic watershed, Alaska', In *Proceedings, Fourth International Conference on Permafrost*, National Academy Press, Washington, D.C., pp. 115-120.

Dingman, S.L. 1966. 'Characteristics of summer runoff from a small watershed in central Alaska', *Water Resour. Res.*, **2**, 751-754.

Dingman, S.L. 1971. Hydrology of Glenn Creek watershed, Tanana basin, Central Alaska', *U.S. Army Cold Region Research Engineering Laboratory Research Report 297*, p. 112.

Dingman, S.L. 1973. 'Effects of permafrost on stream flow characteristics in the discontinuous permafrost zone of central Alaska', In *North American Contribution to Second International Conference of Permafrost*, National Academy of Sciences, Washington, D.C., pp. 447- 453.

Eaton, F. and Wendler, G. 1982. 'The heat balance during the snow melt season for a permafrost watershed in Interior Alaska', *Arch. Meteorol. Geophys. Bioclimatol Ser. A*, **31**, 19-33.

Freeze, R. A. and Cheery, J.A. 1979. *Groundwater*, Prentice-Hall Inc., New Jersey, p. 604.

Gibson, J.J., Edwards, T.W.D. and Prowse, T.D. 1993. 'Runoff generation in a high boreal wetland in northern Canada', *Nord. Hydrol.*, **24**, 213-224.

Hinzman, L.D., Kane, D.L. and Everett, K.R. 1993. 'Hillslope hydrology in an Arctic setting', In *Proceedings, Sixth International Conference on Permafrost*, South China Press, Beijing, pp. 257-271.

Kane, D.L., Bredthauer, S.R. and Stein, J. 1981. 'Subarctic snowmelt runoff generation', In T.S. Vinson (Ed.), *Proceedings of the Specialty Conference on The Northern Community*, ASCE, Seattle, Washington, pp. 591-601.

Kane, D.L. and Stein, J. 1983. 'Water movement into seasonally frozen soils', *Water Resour. Res.*, **19**, 1547-1557.

Landals, A.L. and Gill, D. 1972. 'Differences in volume of surface runoff during the snowmelt period: Yellowknife, Northwest Territories', In *The Role of Snow and Ice in Hydrology Symposium*, Banff, Alberta, IAHS Pub. No. 107, pp. 927-941.

Mackay, J.R. 1983. 'Downward water movement into frozen ground, western arctic coast, Canada', *Can. J. Earth Sci.*, **20**, 120-134.

McNamara, J.P., Kane, D.L. and Hinzman, L.D. 1997. 'Hydrograph separations in an Arctic watershed using mixing model and graphical techniques', *Water Resour. Res.*, **33**, 1707-1719.

McNamara, J.P., Kane, D.L. and Hinzman, L.D. 1998. 'An analysis of streamflow hydrology in the Kuparuk River Basin, Arctic Alaska: a nested watershed approach', *J. Hydrol.*, **206**, 39-57.

Ohmura, A. 1982. 'Objective criteria for rejecting data for Bowen ratio flux calculations', *J. Appl. Met.*, **21**, 595-598.

Pomeroy, J.W. and Gray, D.M. 1995. *Snow Cover: Accumulation, Relocation and Management.* NHRI report No. 7, Supply and Services Canada, Saskatoon, 144 pp.

Price, A.J. and Dunne, T.D. 1976 'Energy balance computations on snowmelt in a subarctic area', *Water Resour. Res.*, **12**, 686-694.

Priestley, C. H. B. and Taylor, R.J. 1972. 'On the assessment of surface heat flux and evaporation using large scale parameters', *Mon. Weather Rev.*, **100**, 81-92.

Quinton, W.L. and Marsh, P. 1998a. 'Melt water fluxes, hillslope runoff and stream flow in an Arctic Permafrost Basin', in *Proceedings, 7th International Conference on Permafrost, Yellowknife, Canada,* 921-926.

Quinton, W.L. and Marsh, P. 1998b. 'The influence of mineral earth hummocks on subsurface drainage in the continuous permafrost zone', *Permafrost and Periglacial Processes*, **9**, 213-228.

Reedyk, S., Woo, M.K.. and Prowse, T.D. 1995. 'Contribution of icing ablation to streamflow in a discontinuous permafrost area', *Can. J. Earth Sci.*, **32**, 13-20.

Santeford, H.S. 1979*a*. 'Snow soil interactions in interior Alaska', In Colbeck, S.C. and Ray, M. (Ed.) *Modeling of Snow Cover Runoff*, Cold Regions Research and Engineering Laboratory, Hanover, NH, pp. 311-318.

Santeford, H.S. 1979*b*. 'Toward hydrologic modeling of the black spruce/permafrost ecosystem of interior Alaska', In *Proceedings 30th Alaska Science Conference*, Fairbanks, Alaska, 9 pp.

Slaughter, C.W. and Kane, D.L. 1979. 'Hydrologic role of shallow organic soils in cold climates', In *Proceedings, Canadian Hydrology Symposium 79 - Cold Climate Hydrology*, National Research Council of Canada, Ottawa, pp. 380-389.

Slaughter, C.W., Hilgert, J.W. and Culp, E.H. 1983. 'Summer streamflow and sediment yield from discontinuous-permafrost headwater catchments', In *Proceedings, Fourth International Conference on Permafrost*, National Academy Press, Washington, D.C., pp. 1172-1177.

Woo, M.K. 1986. 'Permafrost hydrology in North America', *Atmos. - Ocean*, **24**, 201-234.

Woo, M.K. and Steer, P. 1982. 'Occurrence of surface flow on arctic slopes, southwestern Cornwallis Island', *Can. J. Earth Sci.*, **19**, 2368-2377.

Woo, M.K. and Young, K.L. 1997. 'Hydrology of a small drainage basin with polar oasis environment', *Permafrost and Periglacial Processes*, **8**, 257-277.

Table 3.1. Soil properties of the experimental slopes. *K* is hydraulic conductivity.

	North Slope *Upper organic* (0-0.1 m)	*Lower organic* (0.1-0.3 m)	*mineral* (> 0.3 m)	South Slope *silt*
Bulk Density (kg/m^3)	55 ± 20	90 ±20	1340 ± 180	1420 ± 70
Porosity (%)	92 ±4	84 ±10	52 ±7	55 ±4
K (m/s)	$7 \pm 4 \times 10^{-3}$	$2.5 \pm 2 \times 10^{-4}$	$5 \pm 3 \times 10^{-9}$	$6 \pm 1 \times 10^{-6}$

± represents standard deviation

Table 3.2. Water balance of south and north slopes for the period of April 6 to September 22, 1997. All values in mm of water.

	South Slope	**North Slope**
Gains		
Snow	160	187
Icing	0	19
Rainfall	150	150
Subsurface Flow	0	245
Losses		
Sublimation	10	18
Evapotranspiration	372	315
Surface Flow	0	155
Subsurface Flow	0	97
Change in Storage	-72	+16

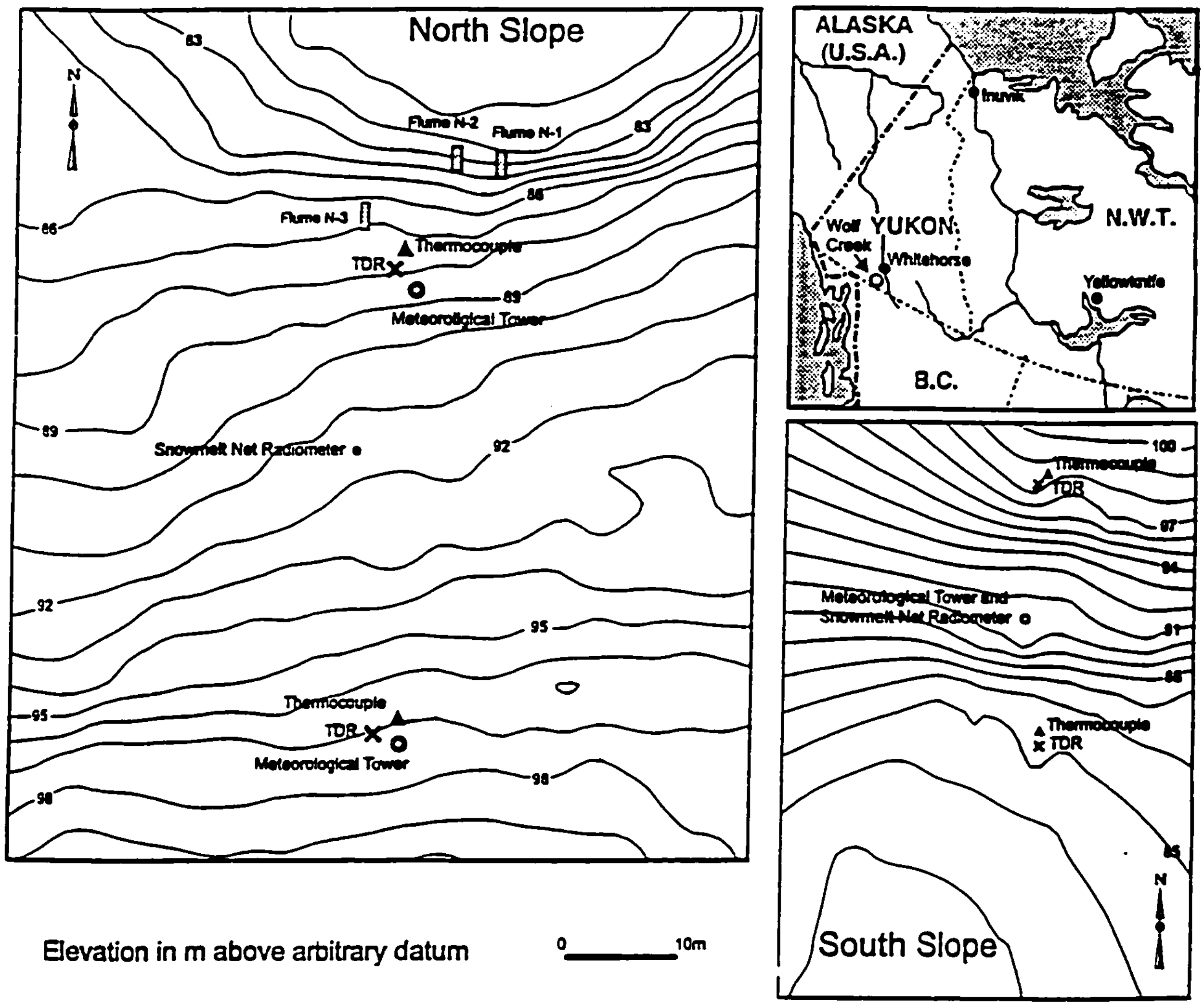

Figure 3.1 Topography and instrumentation of two experimental slopes. Inset shows location of Wolf Creek Basin.

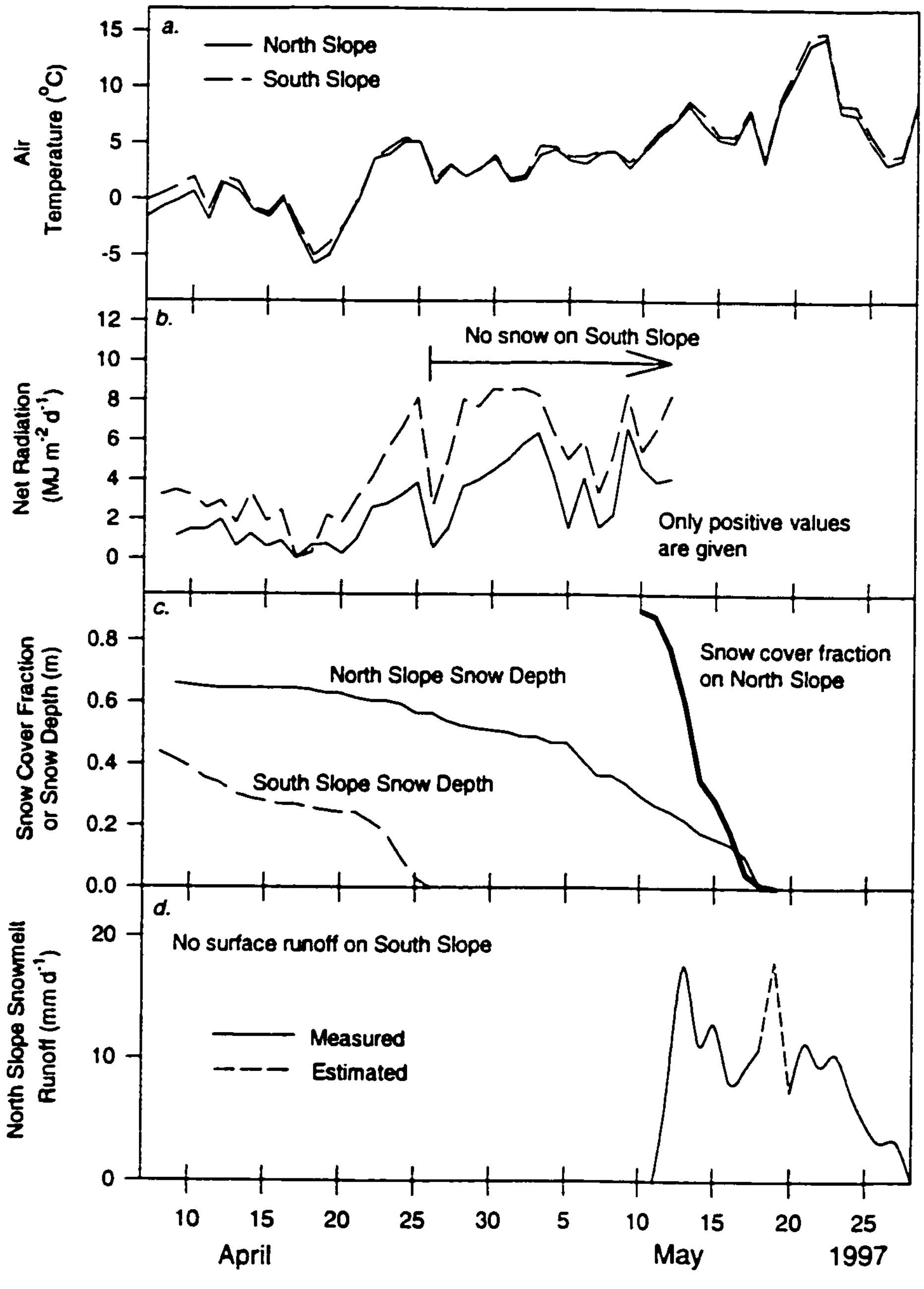

Figure 3.2 Mean daily air temperature, daily net radiation (positive values only), snow depth and snow cover fraction (N-slope only), and surface runoff from the North slope.

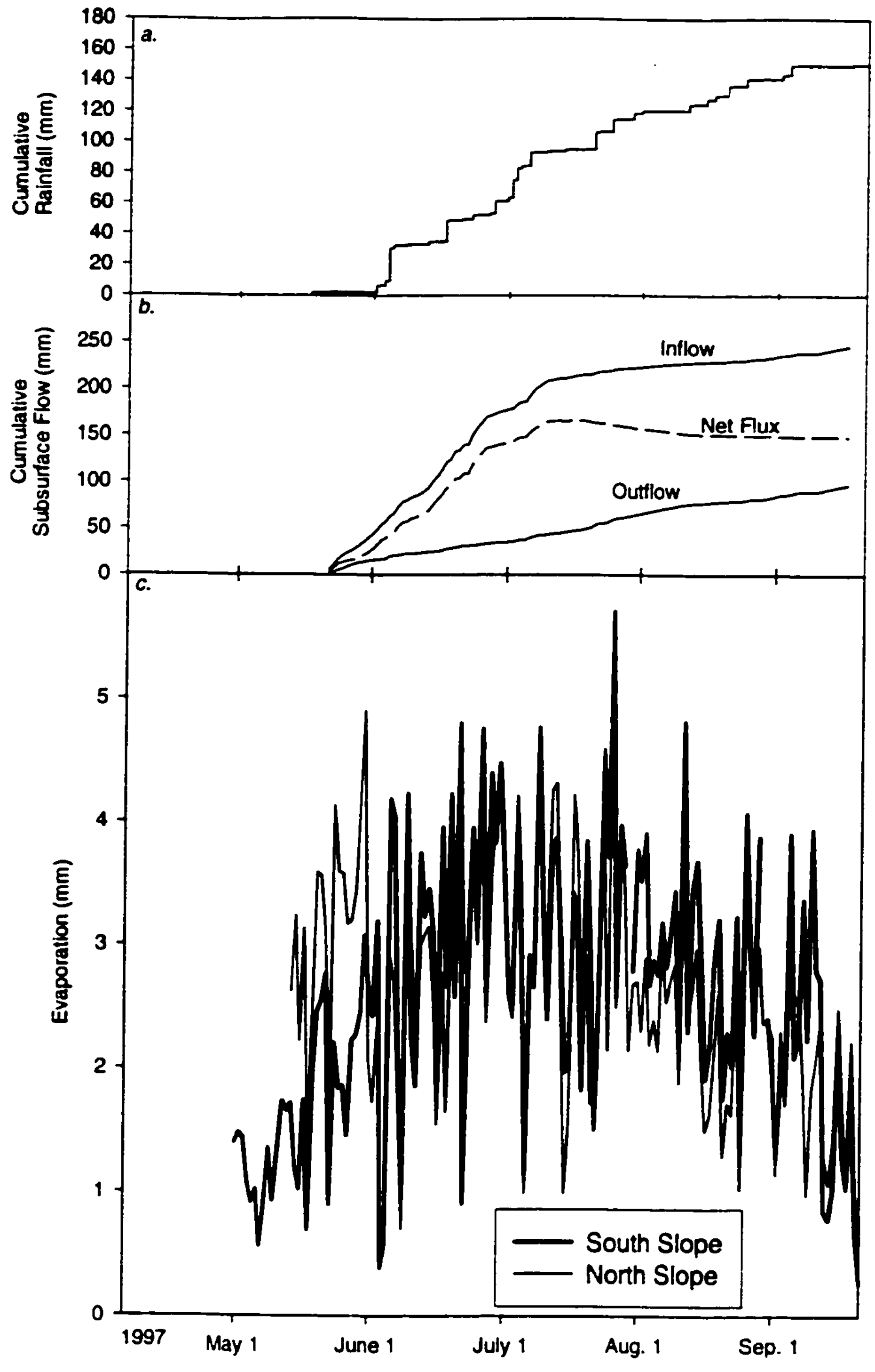

Figure 3.3 Cumulative rainfall, cumulative subsurface inflow, outflow and net flux for the N-slope and daily evaporation for the summer period.

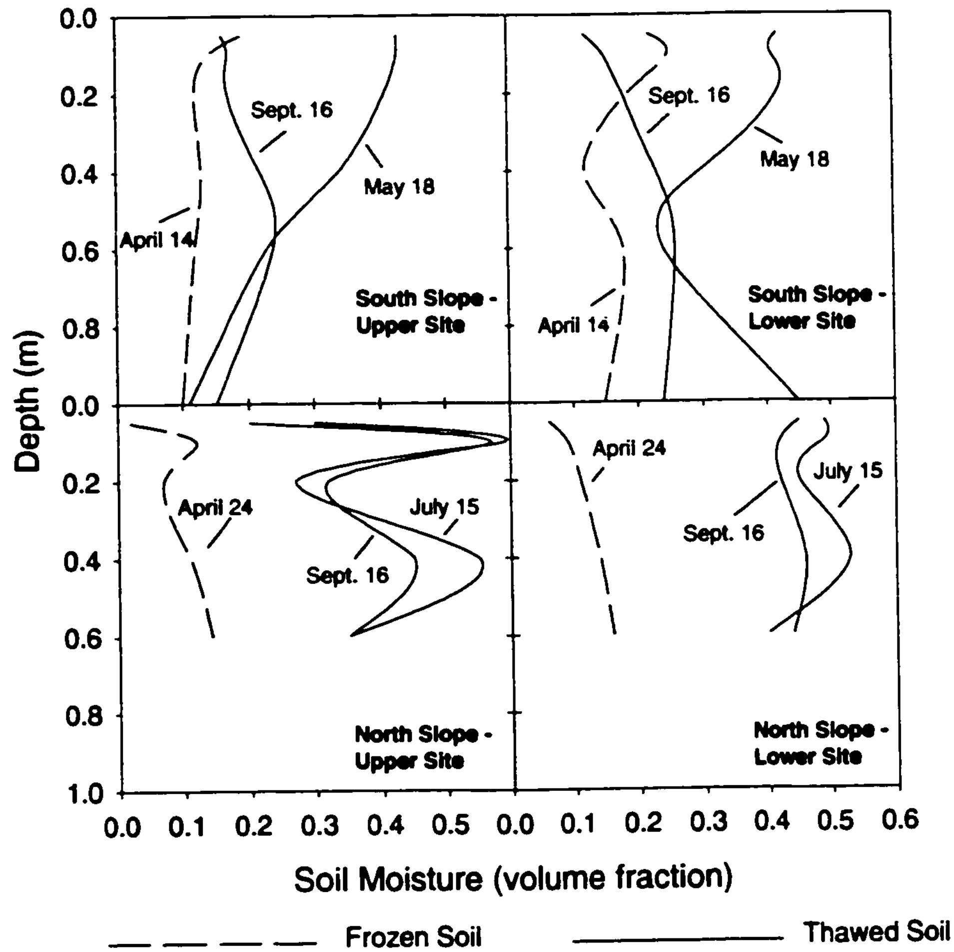

Figure 3.4 Changes in soil water content at four sites between snowmelt and the freeze-back period.

a

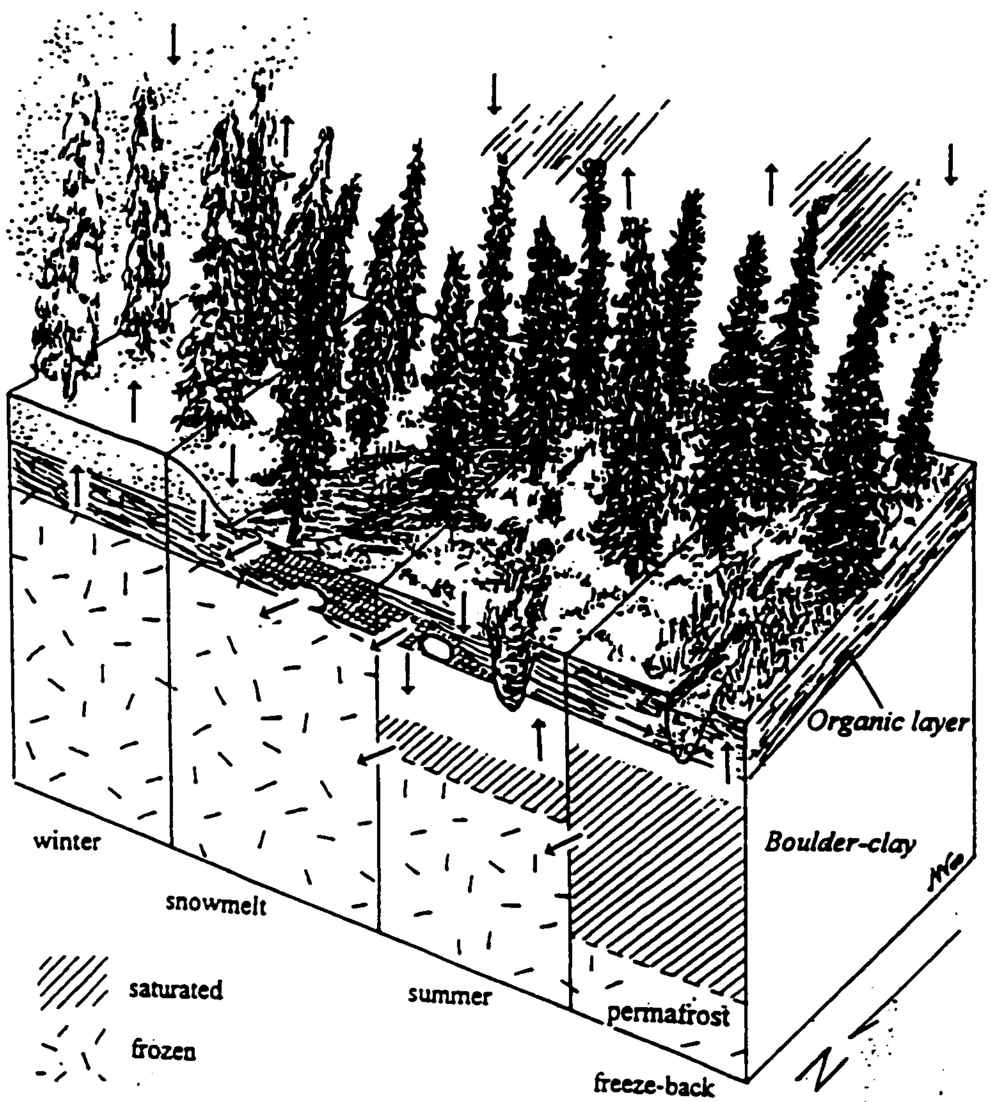

Figure 3.5*a* Seasonality of hydrological processes at north slope. Arrows indicate directions of moisture flux. Water in gullies is frozen as icing during freeze-back. Vertical scale for subsurface zone is greatly exaggerated.

b.

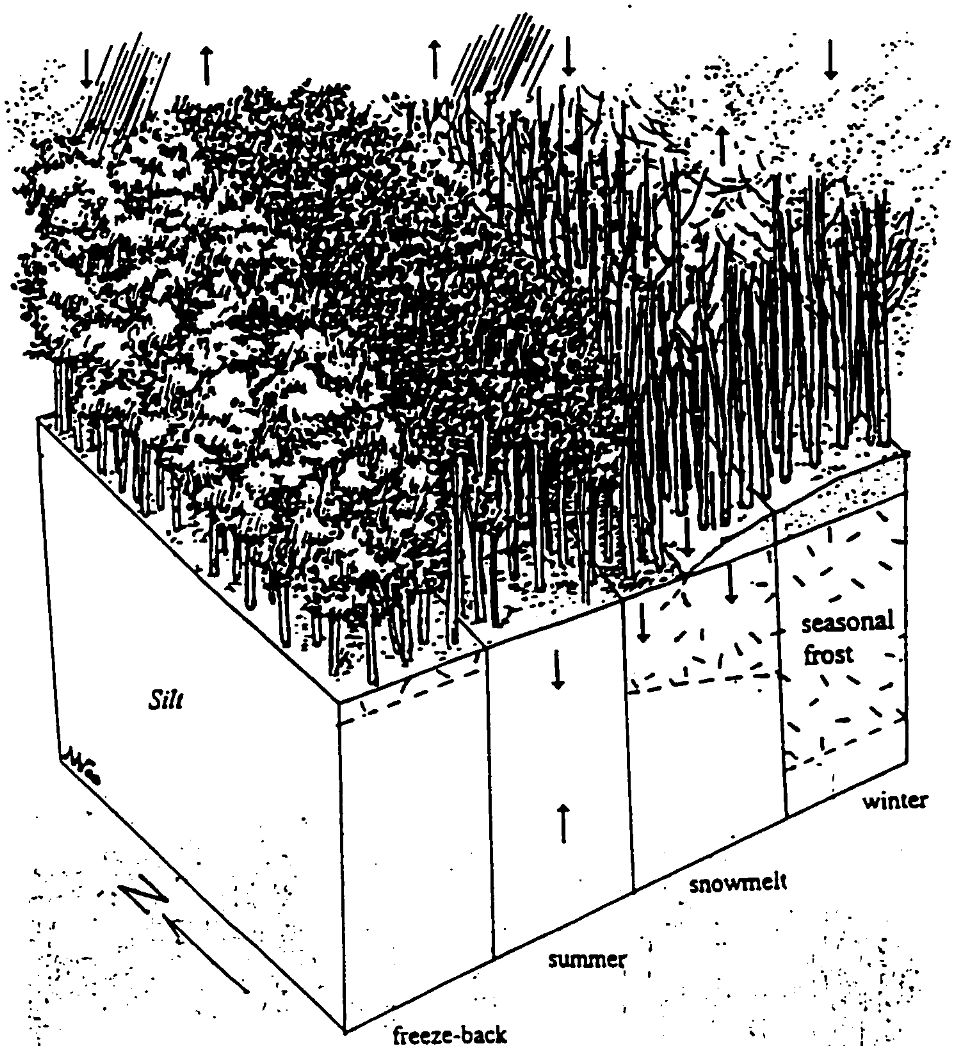

Figure 3.5*b* Seasonality of hydrological processes at south slope. Arrows indicate directions of moisture flux. Vertical scale for subsurface zone is greatly exaggerated.

CHAPTER FOUR

VARIABILITY OF HILLSLOPE WATER BALANCE, WOLF CREEK BASIN, SUBARCTIC YUKON

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and

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Submitted to *Hydrological Processes*

ABSTRACT

A hydrological study was conducted between 1997 and 1999 in the subalpine open woodland of the Wolf Creek basin, Yukon, to assess the inter-slope water balance variability. The water balance during the snowmelt and summer periods on four hillslopes revealed strong contrasts in process magnitudes and highlighted important factors including frost, vegetation, soils and microclimate that controlled vertical and lateral fluxes of water. Snow accounted for approximately half the annual water input, while differences in accumulation among hillslopes were related to interception properties of vegetation. Available energy at the snow surface controlled the melt sequence and the snow on some slopes disappeared up to two months earlier than others. Snowmelt runoff was confined to slopes with ice-rich substrates that inhibited deep percolation, with the runoff magnitude governed by the snow storage and the antecedent moisture of the desiccated organic soils prior to melt. During summer, evapotranspiration exceeded rainfall, largely sustained by water from the soil moisture reservoir recharged during the melt period. Differences in net radiation on slopes controlled the potential evapotranspiration, with the actual rates limited by the phenology of the deciduous forests and shrubs. Evapotranspiration was further suppressed on slopes where the organic soils became dry in late summer. Summer runoff was confined to slopes with porous organic layers overlying mineral soils to form a two-layer flow system: 1) quickflow in the surface organic layer and 2) slowflow in the mineral soil. Differences in the rates of flow were related to the position of the water table which may rise into the organic layer to activate quickflow. The presence of ice-rich frost and permafrost impeded vertical drainage and indirectly regulated the position of the water table. The location of the hillslope within a basin influenced recharge and discharge dynamics. Slope segments with large inflows sustained discharge throughout the summer to enhance basin runoff. In this way, the present study provides insight into basin hydrology.

4.1.0 INTRODUCTION

Hillslopes are dominant physiographic features of subarctic, subalpine basins. On these slopes, both vertical and lateral water fluxes exhibit large variability as topography, microclimate, soil, frost and vegetation vary widely over short distances. Whereas streamflow integrates the basin response to hydrologic inputs, understanding the hillslope water balance is central to the quantification of basin hydrology.

Most studies of subarctic hydrology have focused upon processes, including canopy interception and sublimation (Pomeroy *et al.*, 1999), snowmelt (Price and Dunne, 1976; Eaton and Wendler, 1982; Carey and Woo, 1998; Giesbrecht and Woo, *in press*; Woo and Giesbrecht, *in press*), evaporation (Lafleur *et al.*, 1992; Granger, 1999), infiltration, soil storage and runoff (Dingman, 1971; Landals and Gill, 1971; Santeford, 1979*b*; Slaughter and Kane, 1979; Kane *et al.*, 1981; Chacho and Bredthauer, 1983; Kane and Stein, 1983; Slaughter *et al.*, 1983; Carey and Woo, 1999; 2000). Few studies have examined the spatial and seasonal variability of the water cycle. Metcalfe and Buttle (1999) analyzed sub-basin water balance in a boreal forest catchment with discontinuous permafrost and noted large differences in fluxes controlled by variations in snow conditions, rainfall characteristics, thaw depth and storage properties.

Carey and Woo (1998; 1999) examined two hillslopes with strong contrasts in their physical setting and hydrological responses. While this study captured the extreme range in the water balance dynamics, the spatial variability in the magnitude of water balance components needs further refinement in order to account for the hydrological differences among basin slopes. It is the purpose of this paper to examine the factors that cause the variation in timing and magnitude of hydrological processes in a subarctic, subalpine environment. This study expands on previous research to investigate the water balance components of two additional hillslopes in the same catchment.

Comparison of results amongst the four hillslopes will enhance the insight into the hydrological variability with the subalpine, subarctic zone.

4.2.0 STUDY AREA

Four slopes of different orientation were studied within the Wolf Creek Basin (61°31' N, 135°31' W), 15 km south of Whitehorse, Yukon Territory, Canada. Climatic records from the Whitehorse airport (elevation 703 m) indicate that the study area has a subarctic continental climate characterized by large temperature range and low precipitation. The mean January and July temperatures are -21°C and +15°C. Mean annual precipitation is approximately 260 mm, half of which falls as rain (Wahl *et al.*, 1987). Hydrological investigations carried out since 1996 suggest that Whitehorse airport precipitation may underestimate basin precipitation by approximately 25 to 35 % (Pomeroy and Granger, 1999).

The experimental slopes are located within an area of 5 km^2 in order to reduce climatological differences due to distance (Figure 4.1). The physical setting of the North-facing (N) and South-facing (S) slope is provided in Carey and Woo (1998; 1999). The N-slope has an average gradient of 0.18, is underlain by clayey soils with stone inclusions, and capped by an organic layer (0.05-0.3 m thick) consisting of peat, lichens, mosses, sedges and grasses. All but the extreme footslope portion of the slope is underlain by permafrost at depths from 0.6 to 2 m. Vegetation consists of Labrador tea, willow shrubs and stands of black and white spruce. The S-slope has a dense aspen forest canopy approximately 11 m high. The soil consists of thin leaf litter overlying homogeneous silt to a depth of > 2 m. The average gradient is 0.6 and no permafrost is found under this slope. An East-facing (E) slope with an average gradient of 0.1 is underlain by sandy soils with large stone inclusions, capped by an organic layer (0.05-0.25 m thick) consisting of peat, lichens, mosses, sedges and grasses. This slope retains a saturated zone at or near the surface for most of the thawed season, and an acrotelm-catotelm transition has developed within the organic layer. Only seasonal frost is found down to a maximum depth of approximately 1 m. Tree growth is sparse with stands of black and white spruce widely separated by open areas vegetated by Labrador tea and willow shrubs. Physiographically, the E-slope is located at the base of a large alpine upland contributing area that supplies water to the study site throughout most of the summer. A West-facing (W) slope, with isolated stands of spruce has a shrub cover consisting of willow and Labrador tea. Its mineral soil is similar to that of the E-slope but the organic layer is much thicker (0.25-0.45 m) and consists of peat, lichens and mosses with few sedges and grasses. This slope is underlain by permafrost at depths between 0.45 and 0.6 m. Physiographically, it is a straight slope with a gradient of 0.17. Tables 4.1 and 4.2 provide comparisons of the physical setting and soil properties of the four hillslopes.

4.3.0 FIELD METHODS

Hydrological parameters were obtained at no less than four locations on each hillslope. Horizontal hydraulic conductivity was determined for each soil layer using pumping tests conducted in wells and piezometers (Freeze and Cherry, 1979). Porosity and bulk density were determined using gravimetric methods. Ice content was determined by chipping samples from different horizons and then thawing samples, subtracting the unfrozen moisture content measured by time domain reflectometry (TDR).

Microclimate information was obtained from towers located on each slope. Information on the tower instrumentation is given in Table 4.3. Ground temperature measurements on the N and S-slope are given in Carey and Woo (1998). Within the E and W-slopes, temperatures were measured using a thermocouple array at depths of 0.02, 0.05, 0.10, 0.20, 0.40, 0.60 m on both slopes and additional sensors at 1.0 and 1.4 m on the E-slope. A Texas Instruments tipping bucket rain gauge was installed on the E-slope tower. Due to the their close proximity, rainfall was assumed to be the same for the E and W-slopes. For the N and S-slopes, a tipping bucket rain gauge was placed equidistant between the two slopes. Signals from the sensors were sampled every 60 seconds and recorded on Campbell Scientific CR-10 data-loggers at 30 minute intervals.

Snow density and depth were surveyed using a Mount Rose snow sampler and supplemented by snow-pit observations prior to the main melt periods. During melt, snow ablation was monitored twice-daily at several sites on each slope by measuring the rate of snow-surface lowering and converting the depth change into water equivalent using density measurements. Mean snow water equivalent and snow-cover fraction were also mapped daily using 50 point snow surveys.

Soil moisture and runoff measurements for the N and S-slopes are described by Carey and Woo (1999). Soil moisture was measured using calibrated Campbell CS-615 water content reflectometer probes buried at 0.10, 0.20, 0.40 m on the W-slope and at 0.10, 0.20 and 0.6 m on the E-slope. In 1999, water table was monitored using an electronic water level recorder at one site on each slope. In 1997, water table was monitored at four locations when on-site investigations were undertaken. Runoff was measured with sharp-crested 90° V-notch flumes and water-level recorders installed in small micro-catchments on each hillslope (Figure 4.1). These flumes were 0.6 m long and 0.4 m wide, with plastic sheeting that directed water into the flumes. Water level in the flumes was recorded at half-hour intervals and discharge was determined using a rating curved based on spot measurements of discharge. The accuracy of the stage-discharge curve is ±20 %. Catchment area determined for the melt period from topographic analysis and flow path tracing are 840 m^2 (N-slope), 510 m^2 (E-slope) and 425 m^2 (W-slope). Subsurface flows were computed using Darcy's equation (Carey and Woo, 1999).

Soil moisture changes were calculated for both the saturated and unsaturated zones. For the saturated layer, change in storage (ΔS_s) is

$$\Delta S_s = \theta_{sy} \Delta d \tag{4.1}$$

where Δd is the change in water table depth, θ_{sy} is the specific yield obtained by measuring the volumetric fraction of water drained from a block of soil over a 24 hr period. Storage change in the unsaturated layer (ΔS_u) was determined from the water content reflectometry probes above the water table. Unsaturated soil moisture profiles were assumed linear between the soil moisture probe depths.

Evaporation (E) was calculated using two approaches, including the Bowen ratio method

$$E = \left(Q^* - Q_G\right)\left(1 + \beta\right)\rho_w L_v \tag{4.2}$$

where Q^* and Q_G are net radiation and ground heat flux, ρ_w is the density of water and L_v is the latent heat of vaporization and the Bowen ratio, β, is

$$\beta = C_a \Delta T \,/\, \left(L_w \Delta q\right) \tag{4.3}$$

with C_a being the specific heat of air, and ΔT and Δq the difference in air temperature and vapour pressure. Objective criteria given by Ohmura (1982) were used to remove spurious β. Whenever these criteria were not met, a modified Priestley and Taylor (1972) approach was used

$$E = \alpha \frac{\sigma}{\sigma + \lambda} \frac{Q^* - Q_G}{\rho_w L_v} \tag{4.4}$$

where σ is the slope of the saturated vapour pressure curve, λ is the psychometric constant and α is the Priestley and Taylor parameter that relates actual to potential evaporation. The value of α was modified weekly by comparing the Bowen ratio and Priestley and Taylor evaporation rates. Soil moisture budgets were also used to check on evaporation calculations.

4.4.0 Hydrologic Components

Hydrological processes were studied between the spring of 1996 and the fall of 1999. Intense investigation on the N and S-slope was carried out during the melt and summer season of 1997, and on the E and W-slope during melt in 1998 and summer in 1999.

4.4.1 Snowcover and Snowmelt

Snow constitutes approximately half of the annual water input to the study area. Detailed study of the snowmelt hydrology of the N and S-slopes showed extreme contrasts in the timing and pathway of meltwater (Carey and Woo, 1998). Mean snow water equivalent prior to melt was greater on the N-slope (187±24 mm) than the S-slope (160±14 mm) which may be attributed to the higher winter sublimation on the S-slope. Although air temperature was similar above these slopes, radiation receipt was higher on the S-slope and melt proceeded rapidly as the slope became snow free on April 26, whereas complete melt of the N-slope was delayed until May 19 (Figure 4.2*a*). In 1998, snow surveys on the E and W-slope were completed when the snowpack was approaching isothermal conditions. Prior to this, there was minor melt in tree wells and around large shrubs. Mean snow depths on the E and W-slopes were 0.26 m and 0.28 m, yielding water equivalents of 92±34 mm and 90±22mm respectively. The similarity in end-of-winter snow storage was confirmed in 1999 when 78 mm and 88 mm of snow water equivalent was measured on the E and W-slopes before the snowpack ripened. During the 1998 melt season, air temperature and net radiation were slightly greater above the W-slope than the E-slope (Figure 4.2*b*), resulting in earlier melt initiation on the W-slope. Intense warming and increased radiation after April 29 accelerated melt on the E-slope and both slopes became snow-free at the same time except for a few small patches. Sharp increases in ground temperature measured at the probes were used in 1998 and 1999 to approximate the dates when the slopes became snow free (Table 4.4). In both years, the sequence of melt begins with the S-slope followed by the E and W-slopes at approximately the same time and finally the N-slope. In 1999, there was almost a two-month difference in the timing of melt between the N and S-slopes.

4.4.2 Snowmelt Infiltration

Unsaturated conditions, low ice content, the absence of segregated ice and the high infiltration capacity of soils in the S-slope allowed water to infiltrate and percolate the silty soils unimpeded in both 1996 and 1997 (Figure 4.3). In the N, E and W-slopes, the organic layer was desiccated by upward vapour flux during the winter, resulting in a low ice content that permitted meltwater infiltration (Santeford, 1979*a*). Prior to melt, total organic layer water content (ice + liquid water) for the N, E and W-slope was 18 % (40 mm), 34 % (55 mm) and 26 % (58 mm). Thus, most frozen organic-covers offered little resistance to infiltration which continues until saturation occurred. Meltwater infiltration and melting of ground ice, mostly as hoar crystals, gradually re-wet the organic layers. Liquid water content increased notably in the organic soils during the main melt period within all three slopes (Figure 4.3). The presence of ice-rich horizons (> 70 %) at the interface of organic and mineral soils may prevent deep percolation, hence the development of a perched saturation zone began at the base of the organic layer after 102 mm (N-slope), 20 mm (E-slope) and 60 mm (W-slope) of water was released from the snowpack. While this zone was maintained by meltwater entry into the organic layer, the ice-rich organic-mineral interface prevented deep percolation. Lateral drainage then was generated in the saturated organic layer.

4.4.3 Snowmelt Runoff

At no time was surface runoff observed on the S-slope between 1996 and 1999, although in years with a high ice content and/or rapid melting of a thick snowpack, runoff may be produced. For the N-slope, a description of snowmelt runoff timing and pathways is presented in Carey and Woo (1998)

Unlike the N-slope where preferential flowpaths such as rills and pipes were important flow pathways (Carey and Woo, 2000), diffuse subsurface flow within the frozen and unfrozen organic soils was the main pathway on the E-slope. In the E-slope, the saturated hydraulic conductivity (K_s) of the frozen organic layer (lower layer) determined by pumping tests was 4 x 10^{-5} m/s, approximately half the K_s of the layer when thawed. Large K_s value for the frozen organic soil is due to its high porosity and low ice content. The flume recorded an abrupt start in runoff on April 26 (80 % snowcover) when daily maximum temperatures reached +5°C and continued until April 28 when the daily minimum fell to <-6°C (Figure 4.4*a*). As daily average temperatures fell below freezing, runoff was confined to the afternoon and early evening hours. Rising temperatures and radiation on April 30 accelerated melt and sustained continuous runoff that showed prominent diurnal cycles that peaked around 1900 h and dropped to their lows near 0800 h (PST). Similar to patterns of seasonal melt on the N-slope (Figure 4.4*b*), runoff from the organic layer declined sharply as the snow disappeared.

On the W-slope, flow was confined to the frozen organic layer (K_s=1.1 x 10^{-3} m/s) and subsurface flow began 5 days later than the E-slope (Figure 4.4*a*). Runoff gradually increased from May 1 to May 3 with weak diurnal cycles. The weak oscillations compared to the E and N-slope were likely due to the thicker organic layer and greater percolation time from the surface to the water table. Compared with the N-slope, both the E and W-slopes have much less snowmelt runoff. The greater snowpack on the N-slope in 1997 far exceeded the capacity of the organic soil storage, resulting in increased runoff and greater runoff ratios. Over the 1998 study period, approximately 50 mm and 19 mm of water drained from the E-slope compared with 155 mm for the N-slope in 1997. Runoff ratios for the slopes were 0.8 (N-slope), 0.6 (E-slope) and 0.25 (W-slope).

4.4.4 Rainfall

At the study sites, rainfall was 150 mm in 1997 and 141 mm in 1999. Due to instrumentation difficulties, rainfall was not measured continuously in 1998. Rainfall statistics for the two years were similar, with rain falling on 48 and 47 days of the 117 day study period in 1997 and 1999 respectively (Figure 4.5*a*). In both years, June and July were the wettest months with September being the driest. The majority of summer rainfall events were small magnitude convectional storms of short duration. The largest events in 1997 and 1999 were 19 and 21 mm respectively. Over 80 % of the events in both years were less than 5 mm (Figure 4.5*b*). The thirty-year mean rainfall (May to September) at the Whitehorse airport is 149 mm, suggesting that both 1997 and 1999 are average years.

4.4.5 Summer Runoff

No runoff was observed on the S-slope during the summer of 1996 and 1997. On the N-slope, Carey and Woo (1999) measured subsurface inflow and outflow for the lower slope section (30 m wide by 45 m long), which was dominated by lateral drainage in the organic layer (termed quickflow). Inflow to the lower slope was highest in late May and June due to snowmelt water input from upslope locations that were transferred downslope through pathways within the organic soils (Figure 4.6). After July 1, the water table at the top of the study segment fell into the mineral layer and slope inflow sharply declined. Drainage at the hillslope base was steady throughout the summer as the water table remained within the organic layer. For the 1997 field season, there was a net groundwater inflow of 245 mm and a net groundwater loss of 97 mm from the experimental slope.

Continuous measurements in 1999 of water table position in the E and W-slope showed that drainage on the E-slope was continuous, whereas the permafrost-underlain W-slope experienced periods of negligible drainage when the saturated layer disappeared (Figure 4.7). Monitoring the water table at only one well precluded the analysis of inflow and outflow on these study slopes, although observations of water table position on the slopes in 1997 indicated that the W-slope drained as a thin saturated wedge that became thinner and retreated downslope (Figure 4.8), whereas the entire E-slope has a water-table at or near the surface throughout the summer. To assess the volume of runoff generated directly on the E and W-slopes, runoff was calculated as a residual by monitoring changes in soil water content following rainfall events. The difference between the rate of soil moisture loss at the recorded well/TDR site and the evaporation rate was considered runoff. Direct runoff generated from the E-slope was 30 mm compared with 11 mm from the W-slope.

4.4.6 Evapotranspiration

Evapotranspiration began as the snow disappeared and represented the single greatest loss of moisture from all hillslopes. On the S-slope in 1997, evapotranspiration began in late April and was suppressed as bud-burst of the aspen forest was delayed until mid-May (Figure 4.9*a*). On the N-slope in 1997, rates were large following melt as surface ponding in depressions supplied ample water for evaporation. On the E and W-slope in 1999, no ponding occurred and evapotranspiration was low due to the delay in leafing of shrubs even through there was ample moisture within the near-surface soils following snowmelt (Figure 4.9*b*). On the N and S-slopes, evapotranspiration was similar in June and July, proceeding at rates greater than equilibrium. Surface drying on the N-slope resulted in a reduction in evapotranspiration whereas S-slope evapotranspiration proceeded at potential rates until senescence and reduced radiation lowered values in September. Evapotranspiration on the E and W-slopes was similar from mid-July until mid-August, fluctuating primarily with radiation receipt. Evapotranspiration on the E-slope was greater than the W-slope due to wetter soil conditions. The decline in net radiation and the beginning of senescence in late August signalled evapotranspiration decline. Over the measurement period (May 15-September 12, 1997), evapotranspiration from the E-slope was 260 mm compared with 220 mm from the W-slope, and in 1997 between the disappearance of snowpack and September 22, evapotranspiration was 315 mm and 372 mm on the N and S-slopes respectively (Table 4.5).

4.5 Hillslope Water Balances

Carey and Woo (1999) present the water balance of the N and S-slopes for 1997. These results, together with those from the E and W-slopes will be compared to examine the variability of water balance components on subarctic slopes. Due to differences in the measurement period and strong seasonal patterns, the water balance is divided into snowmelt and summer periods.

4.5.1 Snowmelt Water Balances

During the snowmelt period, the water balance equation is

$$M + I + R - V \pm Q = \Delta S \tag{4.5}$$

where M is snowmelt, I is icing melt, R is rainfall, V is sublimation from the snowpack, Q is net runoff and ΔS is change in storage. The magnitudes of the various water balance components are provided in Table 4.5 and their cumulative quantities in Figure 4.10. R was zero as there was no rainfall during the spring melt study periods. Changes in storage is plotted as a residual in Figure 4.10, whereas ΔS in Table 5.5 was calculated from direct measurements of total soil water content change (frozen + unfrozen) during the study period. In Table 5.5, b represents the net cumulative error in measurement and calculation of all water balance components considering that some errors may be compensating.

Snow water equivalent, snowpack cold content and microclimatic conditions control the onset and duration of melt. On the S-slope, snowmelt began approximately two weeks earlier than the N-slope and became snow-free by April 25, one month before the N-slope. Compared with the E and W-slopes where snowpacks were thinner and melt proceeded rapidly, the N-slope melt period extended over five weeks. The slow release of meltwater over this long period resulted in greater soil moisture storage before the onset of runoff. In contrast, melt on the E and W-slopes was rapid, and the quick release of meltwater reduced the duration of soil moisture recharge and time between the onset of melt and the beginning of runoff. Runoff on hillslopes began once the storage capacity of surface organic layers was exceeded and continued until melt inputs ceased and the organic layers drainage reduced the storage below field capacity.

Differences in snow water equivalent result from different over-winter accumulation between the two years and the variations in interception and snowpack sublimation among the slopes. On all slopes, sublimation from the snowpack was a small component of the melt water balance and occurred steadily throughout the melt in response to turbulent exchange. On the W-slope, large sublimation were due to high wind-speeds whereas on the N-slope, the long melt period allowed for increased sublimation. Icing in rills was observed only on the N-slope, and melted rapidly after the disappearance of the snowpack. No runoff was observed on the S-slope, but 155 mm, 50 mm and 19 mm of water was discharged from the N, E and W-slopes, giving runoff ratios of 0.8, 0.6 and 0.25 respectively. Increased snow storage on the N-slope yielded greater runoff as once the storage capacity of the organic layer was exceeded, the remaining water ran off. Measured changes in soil water indicate recharges of 45 mm (E-slope), 79 mm (W-slope), 58 mm (N-slope) and 137 mm (S-slope), which compare well with the residuals from the water balance of 32 mm (E-slope), 63 mm (W-slope), 31 mm (N-slope), 149 mm (S-slope). Values of b indicate a 10 to 20 % error in the water balance terms during snowmelt.

4.5.2 Summer Water Balances

The water balance for the summer period is

$$R - ET \pm Q = \Delta S \tag{4.6}$$

where R is rainfall, ET is evapotranspiration, Q is net runoff and ΔS is change in storage. The summer water balances are presented in Table 5.5 and the cumulative values in Figure 4.11. In Figure 4.11, ΔS is plotted as a residual for the N and S-slopes whereas the value in Table 5.5 is determined from direct measurement. Evapotranspiration was the greatest loss of water from the hillslopes, exceeding precipitation by 79 mm (W-slope) to 203 mm (S-slope). There were large differences in ET among the slopes, with S (372 mm) and N-slopes (315 mm) far surpassing the E (260 mm) and W-slopes (270 mm). Larger values on the S-slope are attributed to greater energy receipt above the aspen forest, earlier exposure of bare ground, and no apparent moisture limitation as soils become dry (moisture falling to < 12 %). Larger ET from the N-slope than the E-slope can be explained by dense surface vegetation and forest cover, increased net radiation receipt (1480 MJ N-slope, 1200 MJ E-slope, 1277 MJ W-slope) and daily temperatures that averaged 3°C higher. Values of ET compare closely with those presented by Granger (1999) for sites within the basin between 1995 and 1997. In 1997, Granger (1999) reported 373 mm of ET from a coniferous forest site at elevations much lower (750 m) than the N and S-slopes and 253 mm from a shrub zone whose vegetation and elevation (1250 m) is similar to the E and W-slope. Wetter conditions at the E-slope allowed higher ET than the W-slope, despite lower total net radiation.

Determination of runoff was complicated due to the inflow-outflow dynamics of the study segments. Within the N-slope, subsurface inflow exceeded outflow until approximately July 1, increasing soil moisture storage. After July 1, inflow to the study slope declined and outflow and evapotranspiration combined to deplete the soil moisture. Inflow and outflow dynamics were not evaluated on the E and W-slope in 1999. On the W-slope, runoff was confined to areas near the slope base, and inflow into the study segment was probably minimal. Direct runoff from the study segment was greatest early in the year when the slope was wetter and less rainfall was required to initiate a runoff response. Total runoff from the study area is calculated as 11 mm, which gives a response factor, *RF* (the ratio of runoff to rainfall) of 0.08. On the E-slope, large inflow to the study segment sustained high water table near the surface. Runoff was calculated as 30 mm for the study period (*RF* = 0.21).

During summer, evapotranspiration and drainage depleted the soil moisture on the W and S-slopes, while on the E and N-slopes, soil moisture was replenished by inflow. In Figure 4.11, ΔS began at 31 mm (N-slope) and 137 mm (S-slope), which was the volume of recharge during melt. On the W and E-slopes in 1999, melt recharge is estimated as 10 mm and 63 mm from TDR measurements. The error term, *b*, in Table 5.5 reflects both the measurement and calculation error along with the magnitude of the missing components of the water balance. The large negative value of *b* on the E-slope encompasses the error and the volume of water supplied laterally from upslope of the study segment. In contrast, the small *b* term on the W-slope provides support for the magnitude of water balance components. On the S-slope, the soil moisture reservoir supplied water for *ET* throughout the summer. At the end of the study period in 1997, there was 92 mm less water within the top 1 m than in fall 1996. On the N-slope, measurements showed 37 mm of soil moisture loss and *b*=20, which can be considered an error term as all components of the hydrological cycle were measured.

4.6.0 DISCUSSION

The interactions among factors and water balance components of subarctic, subalpine hillslopes are complex, causing considerable variations in the timing and magnitude of hydrological processes. Figure 4.12 outlines the major interactions that relate the water balance components to their environmental factors.

4.6.1 Surface-Atmosphere Interactions

Heterogeneity in vegetation among hillslopes is a notable characteristic of subarctic hillslopes. Aspen forests predominate the south-facing well-drained slopes without permafrost, whereas other aspects with poorly drained or permafrost-underlain soils are dominated by stands of black and white spruce with an understory of willow and alder shrubs. This aspect-dependent relationship has been cited across the subarctic from the Northwest Territories (Gibson *et al.*, 1983) to central Alaska (Slaughter and Kane, 1979). Vegetation influences snow accumulation, sublimation and melt. Differences in SWE among the slopes is related to the collection efficiency of the canopy as intercepted snow is prone to sublimation (Harding and Pomeroy, 1996). Estimates of snow sublimation from forest canopies vary widely from 4 % to over 30 % depending upon interception efficiency, exposure time and atmospheric conditions (Pomeroy and Gray, 1995). On the N-slope, Woo and Giesbrecht (2000) modelled heat exchange between the spruce forest and surrounding snowpack, highlighting the role of tree shadows and melt within tree wells on slope melt patterns. In summer, properties of the canopy along with storm characteristics determine the amount of rainfall interception, which is quickly returned to the atmosphere as evaporation. On the S-slope, complete canopy cover of the slope soils suggests greater interception loss than the sparse spruce vegetation on the other slopes. The annual leafing cycle of vegetation modifies *ET* as rates were low prior to bud-burst in the spring and declined rapidly during senescence.

Differences in snowmelt and evapotranspiration among the study slopes can be related to the energy receipt on the slopes. In complex terrain, the shortwave radiation is spatially variable being controlled primarily by aspect and exposure (e.g. Hinzman *et al.*, 1992), yet the role of longwave radiation is more complex as radiation from surrounding slopes (especially those that have become snow-free) may influence snowmelt on slopes (e.g. Plüss and Ohmura, 1997). The asynchronous nature of radiation timing and magnitude strongly influences the melt season on various slopes and hence has important implications for the occurrence of spring freshet (Janowicz *et al.*, 1997). In summer, the magnitude of available energy controls the potential for evapotranspiration. Large differences in net radiation on the N and S-slopes compared to the E and W-slopes resulted in much greater ET rates on the prior slopes. On the W-slope, available energy was greater than the E-slope, yet dry soil conditions limit latent heat flux.

4.6.2 Surface-Ground Interactions

Topographic influences include those caused by 1. the local slope, and 2. the position of the slope in relation to the surrounding terrain. Locally, runoff rates increase with greater slope angle, and in the absence of inflow from upslope areas, soil moisture increases downslope due to gravity drainage. This redistribution of moisture within the slope convects heat, resulting in greater active layer thaw near the slope base (Carey and Woo, *in press*). The setting of a slope segment within the surrounding terrain influences slope recharge and discharge. In hillslopes where inflow exceeds outflow, frequent saturated conditions maintain zones of continuous discharge. For instance, inflow to the E-slope sustains a high water table throughout the summer. For the N-slope, inflow exceeds outflow early in the summer, yet declines as drying progresses and hydrologic connections disappear. Although hillslope plan-form influences wetness distribution in a slope, the interplay between inflow and outflow controls the hydrologic connectivity, soil moisture patterns and the ability of a slope segment to transmit water to the stream from upslope areas. The exchange of water between terrain units (e.g. Alpine-Subalpine) and the redistribution of water at the basin scale are topics that require more study.

Soil profiles consisting of porous organic layers overlying mineral substrates are widespread in subarctic and tundra environments (*e.g.* Slaughter and Kane, 1979; Hinzman *et al.*, 1993; Quinton and Marsh, 1999). The difference in hydraulic properties between these layers gives rise to a two-layer flow system described by Carey and Woo (2000). Within the organic layer, variations in the level of decomposition and/or the presence of an acrotelm-catotelm transition within the organic soils may also cause large changes in saturated hydraulic conductivity with depth (Quinton and Marsh, 1999). Where these profiles occur, runoff increases as the water table approaches the ground surface. Furthermore, a perched water table, such as that observed in the N-slope, may yield lateral flow at the organic-mineral interface (Carey and Woo, 1999). Runoff may be reduced by increased storage capacity of the organic soil, which is a function of its thickness and physical properties. Despite shallow thaw on the W-slope, a thick organic layer reduces runoff and retains large amounts of water prior to saturation. In contrast, slopes with mineral soils and no capping layer do not exhibit the two-layer flow system, and the absence of a discontinuity within the profile encourages vertical water fluxes (as observed on the S-slope).

Ice-rich frost influences the water balance by impeding water movement due to ice-filled pores. During melt, ground frost can obstruct the percolation of meltwater, encouraging the development of a perched saturated layer and lateral flow. Typically, this occurs on organic-covered hillslopes where an ice-rich layer forms at the organic-mineral interface (*cf.* Slaughter and Kane, 1979). Where pores are not sealed by ice, infiltration and percolation are unimpeded (*cf.* Kane and Stein, 1983). In summer, ice-rich permafrost seals soil interstices, restricting deep drainage and indirectly regulating the water table level and soil moisture content. This then controls the magnitude of runoff by activating the two-layer flow system such that rapid flow occurs when the water table rises within the organic layer.

4.7.0 CONCLUSIONS

Water balances were determined for four subarctic, subalpine hillslopes to improve understanding of intra-basin hydrologic variability and the factors that control the timing and the magnitude of hydrological fluxes. Results from this study enlarge upon those presented by Carey and Woo (1998; 1999). The major findings are:

1) The magnitude and timing of water balance components vary among the slopes during the snowmelt and summer periods.

2) Snow disappearance varies by up to two months on hillslopes within a similar climate setting. Snowmelt timing and rate are controlled by radiation receipt, which is principally controlled by aspect, exposure and vegetation.

3) Snowmelt runoff is confined to hillslopes with organic soils with an ice-rich base that prevents the percolation of meltwater. Frozen but porous organic materials do not impede water movement and once the storage capacity of the organic layer is exceeded, lateral runoff begins.

4) Evapotranspiration is the largest water loss from all hillslopes, exceeding summer rainfall inputs and depleting the soil moisture that is recharged by snowmelt. Differences in evapotranspiration among hillslopes are related to the length of the snow-free period, radiation receipt, vegetation phenology, and the moisture conditions in the near-surface soils.

5) Runoff during the summer is confined to the wet organic-covered hillslopes, with larger flow from slopes where the water table is close to the surface.

6) For some slope segments, lateral inflow and outflow significantly affect the magnitude of other water balance terms. Where inflow equals or exceeds outflow, the hillslope remains wet and runoff and evapotranspiration are enhanced. Drainage of the hillslope segment without inflow lowers water table elevation, reducing runoff in the organic layer and drying the near-surface soils.

4.8.0 References

Carey SK, Woo MK. 1998. Snowmelt hydrology of two subarctic slopes, Southern Yukon, Canada. *Nordic Hydrology* **29**: 331-346.

Carey SK, Woo MK. 1999. Hydrology of two slopes in subarctic Yukon, Canada. *Hydrological Processes* **13**: 2549-2562.

Carey SK, Woo MK. 2000. The role of soil pipes as a slope runoff mechanism, subarctic Yukon, Canada. *Journal of Hydrology*, **223**: 206-222.

Carey SK, Woo MK. 2000. Within-slope variability of ground heat flux, subarctic Yukon. *Physical Geography, (In press).*

Chacho Jr. EF, Bredthauer S. 1983. Runoff from a small subarctic watershed, Alaska. In *Proceedings, Fourth International Conference on Permafrost*, National Academy Press, Washington, D.C., pp. 115-120.

Dingman SL. 1971. *Hydrology of Glenn Creek watershed, Tanana basin, Central Alaska*. U.S. Army Cold Region Research Engineering Laboratory Research Report 297, Hanover, New Hampshire, p. 112.

Eaton F, Wendler G. 1982. The heat balance during the snow melt season for a permafrost watershed in Interior Alaska. *Archives for Meteorology, Geophysics, and Bioclimatology*, *Series A* **31**: 19-33.

Freeze RA, Cheery JA. 1979. *Groundwater*. Prentice-Hall Inc., New Jersey, p. 604.

Gibson JJ, Edwards TWD, Prowse TD. 1993. Runoff generation in a high boreal wetland in northern Canada. *Nordic Hydrology* **24**: 213-224.

Giesbrecht, MA, Woo MK. 2000. Simulation of snowmelt in a subarctic spruce woodland: 1. Open woodland model. *Water Resources Research* (*in press*).

Granger RJ. 1999. Partitioning of energy during the snow-free season at the Wolf Creek research basin. In *Wolf Creek Research Basin: Hydrology, Ecology, Environment*, Pomeroy JW, Granger RJ. (eds), Environment Canada, Saskatoon, pp. 33-43.

Harding RJ, Pomeroy JW. 1996. The energy balance of the winter boreal landscape. *Journal of Climate* **9**: 2778-2787.

Hinzman LD, Wendler G, Gieck RE, Kane DL. 1992. Snowmelt at a small Alaskan arctic watershed 1. Energy related processes. In *9th International Northern Research Basins Symposium/Workshop*, Canada, 1992. NHRI Symposium No. 10. Prowse TD, Ommanney CSL, Ulmer KE (eds), Saskatoon, pp. 171-197.

Hinzman LD, Kane DL, Everett KR. 1993. Hillslope hydrology in an Arctic setting. In *Proceedings, Sixth International Conference on Permafrost*, South China Press, Beijing, pp. 257-271.

Janowicz JR, Gray DM, Pomeroy JW. 1997. Snowmelt and runoff in a subarctic mountain basin. In Proceedings of the Hydro-ecology Workshop on the Arctic Environmental Strategy, Milburn D (ed). Symposium NO. 16, National Hydrology Research Institute, Saskatoon, pp. 303-320.

Kane DL, Stein J. 1983. Water movement into seasonally frozen soils. *Water Resources Research* **19**: 1547-1557

Kane DL, Bredthauer SR, Stein J. 1981. Subarctic snowmelt runoff generation. *Proceedings of the Specialty Conference on The Northern Community*, Vinson TS (ed.), ASCE, Seattle, Washington, pp. 591-601.

Landals AL, Gill D. 1972. Differences in volume of surface runoff during the snowmelt period: Yellowknife, Northwest Territories. In *The Role of Snow and Ice in Hydrology Symposium*, Banff, Alberta, IAHS Pub. No. 107, pp. 927-941.

Metcalfe RA, Buttle JM. 1999. Semi-distributed water balance dynamics in a small boreal forest basin. *Journal of Hydrology* **226**: 66-87.

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

Plüss C, Ohmura A. 1997. Longwave radiation on snow-covered mountainous surfaces. *Journal of Applied Meteorology* **36**: 818-824.

Pomeroy JW, Gray DM. 1995. Snowcover accumulation, relocation and management. *National Hydrology Research Institute Science Report No. 7.* NHRI, Environment Canada, Saskatoon, 144 pp.

Pomeroy JW, Granger RJ.(eds.) 1999. *Wolf Creek Research Basin: Hydrology, Ecology, Environment.* Environment Canada, Saskatoon, p.160.

Pomeroy JW, Hedstrom N, Parviainen J. 1999. The snow mass balance of Wolf Creek, Yukon: Effects of snow sublimation and redistribution. In *Wolf Creek Research Basin: Hydrology, Ecology, Environment*, Pomeroy JW, Granger RJ. (eds), Environment Canada, Saskatoon, pp. 15-30.

Priestley CHB, Taylor RJ. 1972. On the assessment of surface heat flux and evaporation using large scale parameters. *Monthly Weather Re*view **100**: 81-92.

Price AJ, Dunne TD. 1976. Energy balance computations on snowmelt in a subarctic area. *Water Resources Research* **12**: 686-694.

Quinton, W.L., 1997. *Runoff from Hummock-covered Arctic Tundra hillslope in the Continuous Permafrost Zone.* Unpublished Ph.D. Thesis. University of Saskatchewan, Saskatoon, Canada, p. 277.

Quinton WL, Marsh P. 1999. A conceptual framework for runoff generation in a permafrost environment. *Hydrological Processes* **13**: 2563-2581.

Santeford HS. 1979*a*. Snow soil interactions in interior Alaska. In *Modeling of Snow Cover Runoff.* Colbeck SC, Ray M. (ed). Cold Regions Research and Engineering Laboratory, Hanover, NH, pp. 311-318.

Santeford HS. 1979*b*. Toward hydrologic modeling of the black spruce/permafrost ecosystem of interior Alaska. In *Proceedings 30th Alaska Science Conference*, Fairbanks, Alaska, 9 pp.

Slaughter CW, Kane DL. 1979. Hydrologic role of shallow organic soils in cold climates. In: *Proceedings, Canadian Hydrology Symposium 79 - Cold Climate Hydrology*, National Research Council of Canada, Ottawa, pp. 380-389.

Wahl HE, Fraser DB, Harvey RC, Maxwell JB. 1987. *Climate of Yukon.* Environment Canada, Atmosphere Environment Service, Climatological Studies No. 40, Ottawa.

Woo MK, Giesbrecht MA. 2000. Simulation of snowmelt in a subarctic spruce woodland: 1. tree model. *Water Resources Research* (*in press*).

Table 4.1.Characteristics of experimental slopes, Wolf Creek basin, Yukon.

	N-slope	S-slope	E-slope	W-slope
Elevation (m)	1175	1175	1250	1250
Slope Angle	0.18	0.6	0.10	0.17
Vegetation	Open Black/White spruce Willow shrub Labrador Tea	Closed aspen forest	Sparse Black/White spruce Willow shrub Labrador Tea	Willow shrub Labrador Tea
Frost	permafrost (0.6-2 m)	seasonal frost (1.6 m)	seasonal frost (1.2 m)	permafrost (0.45-0.65 m)

Table 4.2. Experimental slope soil properties. K_s is saturated hydraulic conductivity and θ_{sr} is specific retention.

	North Slope			South Slope
	Upper Organic (0-0.1 m)	Lower Organic (0.1-0.3 m)	Mineral Soil (>0.3 m)	Mineral Soil (silt)
Bulk Density (kg/m^3)	55 ± 20	90 ± 20	1340 ± 180	1420 ± 70
Porosity (%)	92 ± 4	84 ± 10	52 ± 7	55 ± 4
K_s (m/s)	$7 \pm 4 \times 10^{-3}$	$2.5 \pm 2 \times 10^{-4}$	$5 \pm 7 \times 10^{-9}$	$6 \pm 1 \times 10^{-6}$
θ_{sr}	0.44 ± .09	0.49 ± .08	0.42 ± .05	0.31 ± .05

	East Slope			West Slope		
	Upper Organic (0-0.1 m)	Lower Organic (0.1-0.25 m)	Mineral Soil (>0.25 m)	Upper Organic (0-0.2 m)	Lower Organic (0.2-0.4 m)	Mineral Soil (>0.6 m)
Bulk Density (kg/m^3)	74 ± 30	130 ± 50	1644 ± 120	48 ± 20	69 ± 20	1565 ± 100
Porosity (%)	86 ± 5	77 ± 12	4 5± 5	89 ± 4	82 ± 6	47 ± 4
K_s (m/s)	$6 \pm 7 \times 10^{-4}$	$1 \pm 1 \times 10^{-4}$	$1 \pm 3 \times 10^{-7}$	na	$1 \pm 2 \times 10^{-3}$	$2 \pm 4 \times 10^{-7}$
θ_{sr}	0.48 ± .06	0.48 ± .07	0.1 ± .05	0.37 ± 09	0.41 ± .10	0.10 ± .04

Table 4.3. Climatological instrumentation employed on the study slopes. Number in italics represents height of sensors.

Climatological Variable	**N-slope**	**S-slope**	**E-slope**	**W-slope**
Solar Radiation (Pyranometer)	Eppley *6 m*	Eppley *17 m*	Li-Cor *3 m*	Li-Cor *3 m*
Net Radiation (Pyrradiometer)	REBS *6 m*	REBS *17 m* Middleton *3 m*	REBS *3 m*	REBS *3m*
Wind Speed	Met-one *2, 6 m*	Met-one *17 m*	Met-one *3 m*	Met-one *3m*
Air Temperature	Vaisala HMP35CF *2 m* Aspirated thermocouples *2,3,4,5,6 m*	Vaisala HM35PCF *3 m* Aspirated thermocouples *13,14,15,16, 17 m*	Vaisala HM35PCF *1.5, 3 m*	Vaisala HM35PCF *1.5, 3 m*
Relative Humidity	Vaisala HMP35CF *2 m* Aspirated wet thermocouples *2,3,4,5,6 m*	Vaisala HM35PCF *3 m* Aspirated wet thermocouples *13,14,15,16, 17 m*	Vaisala HM35PCF *1.5, 3 m*	Vaisala HM35PCF *1.5, 3 m*
Precipitation	Texas Instruments *3 m* between N and S-slopes.	Texas Instruments *3 m* between N and S-slopes.	Texas Instruments *3 m*	Taken from E-slope

Table 4.4. Estimated dates that experimental slopes became snow-free in 1998 and 1999.

	1998	*1999*
North slope	May 16	June 4
South slope	prior to April 6	April 11
East slope	May 5	May 10
West slope	May 5	May 14

Table 4.5. Water balance components for the study slopes. All values are in mm of water.

	E-slope	**W-slope**	**N-slope**	**S-slope**
Snowmelt Period	24 April - 6 May 1998	24 April - 6 May 1998	21 April - 25 May, 1997	6 April - 26 April, 1997
Snow Water Equivalent	92	90	187	160
Icing	0	0	19	0
Sublimation	8	13	18	10
Runoff	50	19	155	0
Change in storage	+45	+79	+58	+137
b	-11	-21	-25	+13

Summer Period	15 May - 12 Sept., 1999	15 May - 12 Sept., 1999	18 May - 22 Sept., 1997	26 April - 22 Sept. 1997
Rainfall	141	141	150	150
Evapotranspiration	260	220	315	372
Runoff (outflow)	30	11	97	0
Runoff (inflow)	na	na	245	0
Change in storage	-12	-96	-37	-138
b	-137	+17	+20	-84

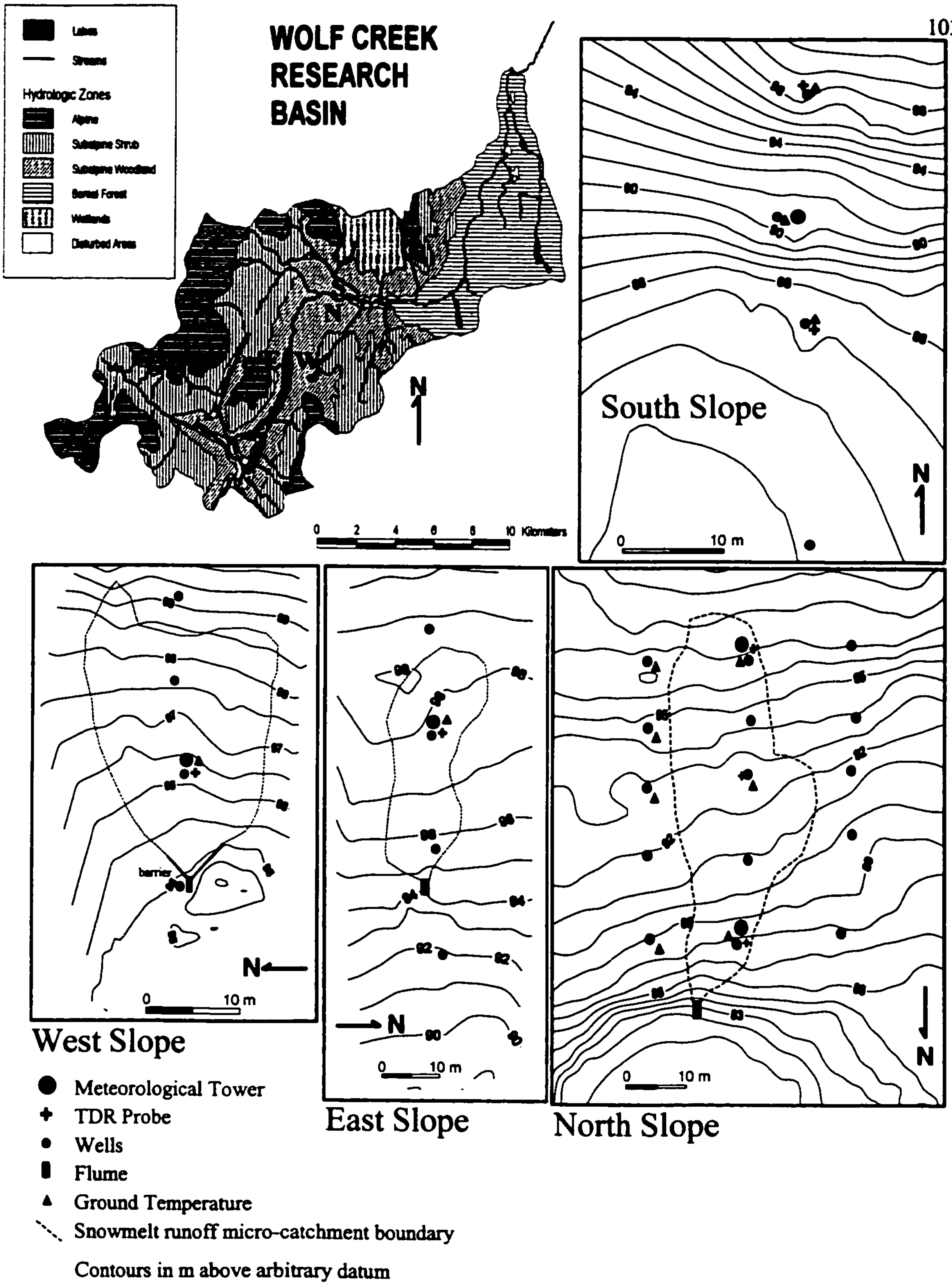

Figure 4.1 Hydrologic zones and location of experimental slope within the Wolf Creek Basin, topography of experimental slopes and deployment of instrumentation.

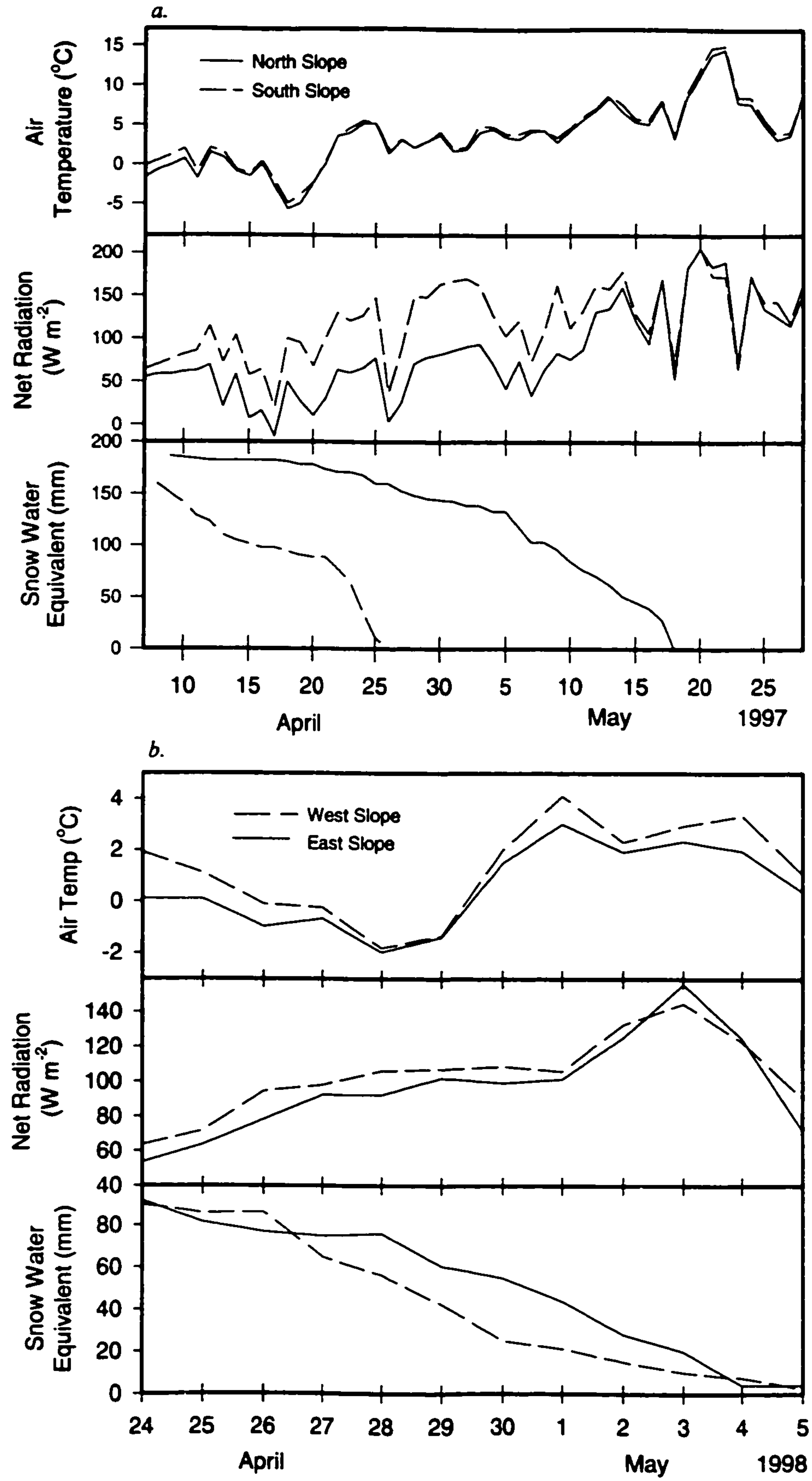

Figure 4.2 Mean daily air temperature, net radiation and snow water equivalent for the (*a*) North and South slopes in 1997, and (*b*) East and West facing slopes in 1998.

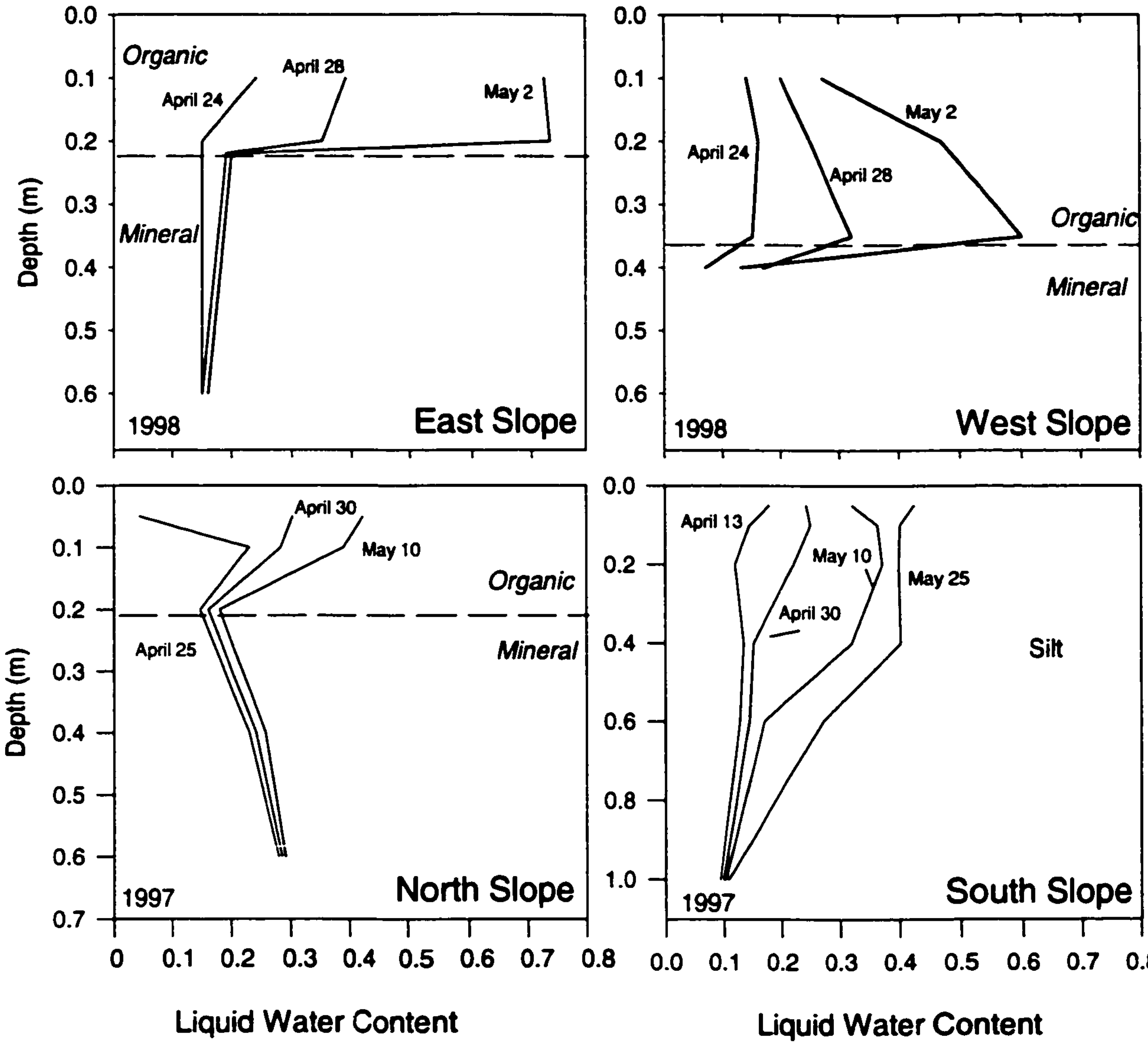

Figure 4.3 Changes in liquid water content for the four slopes during the melt period.

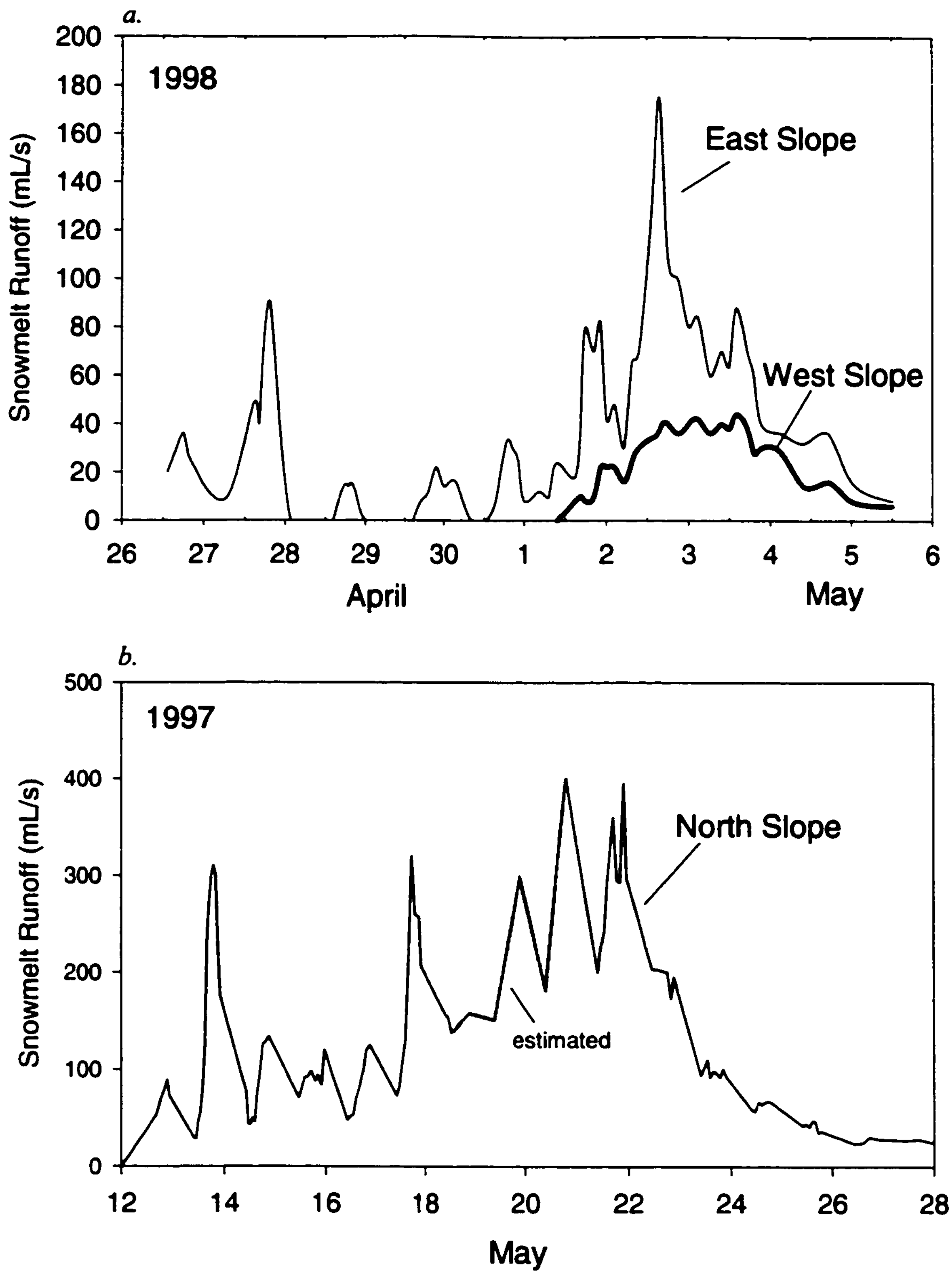

Figure 4.4. Runoff during the (*a*) 1998 snowmelt period for the East and West facing slopes, and (*b*) the snowmelt period in 1997 for the North slope.

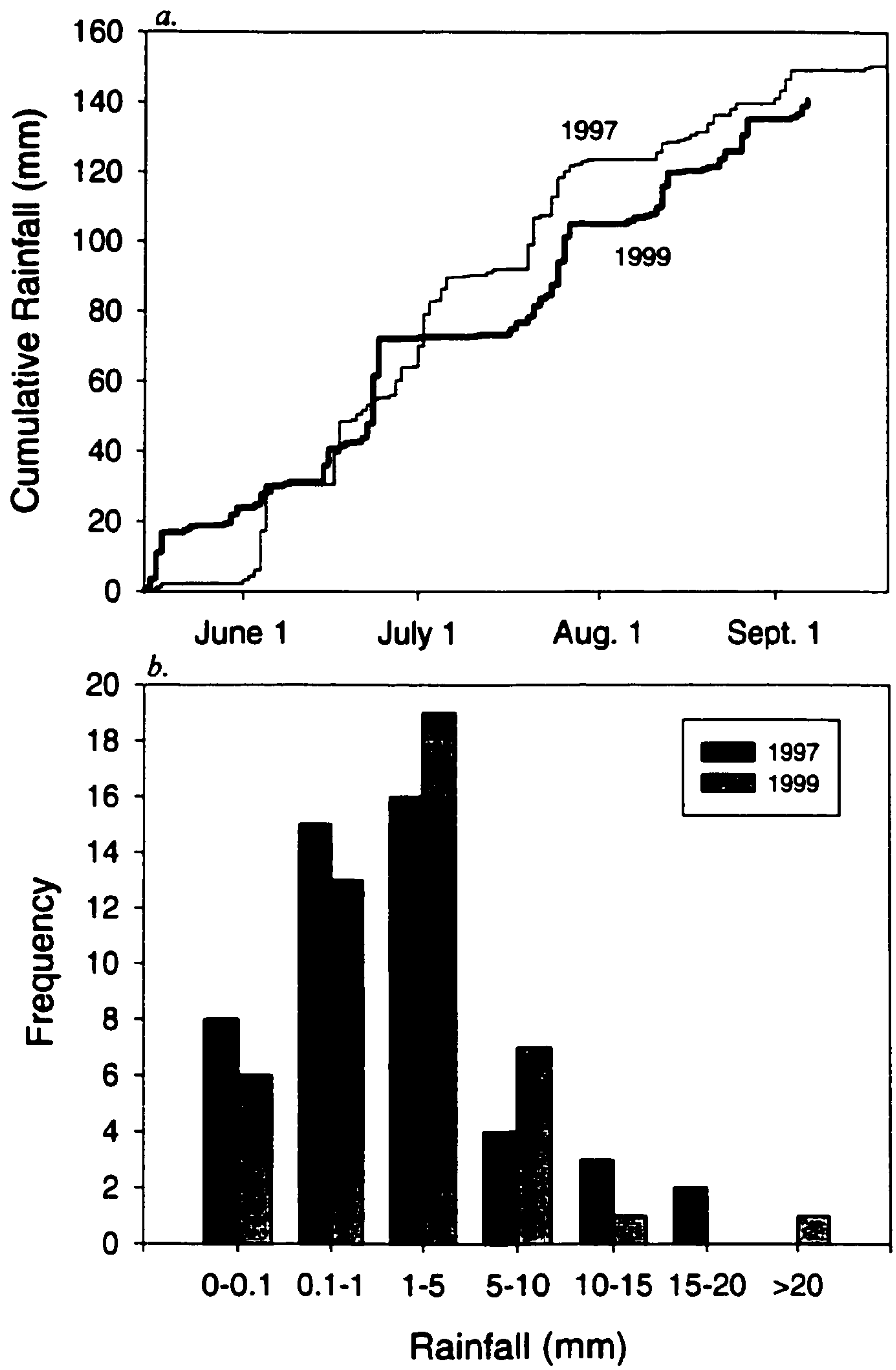

Figure 4.5. (*a*) Cumulative rainfall, and (*b*) storm characteristics for 1997 and 1999.

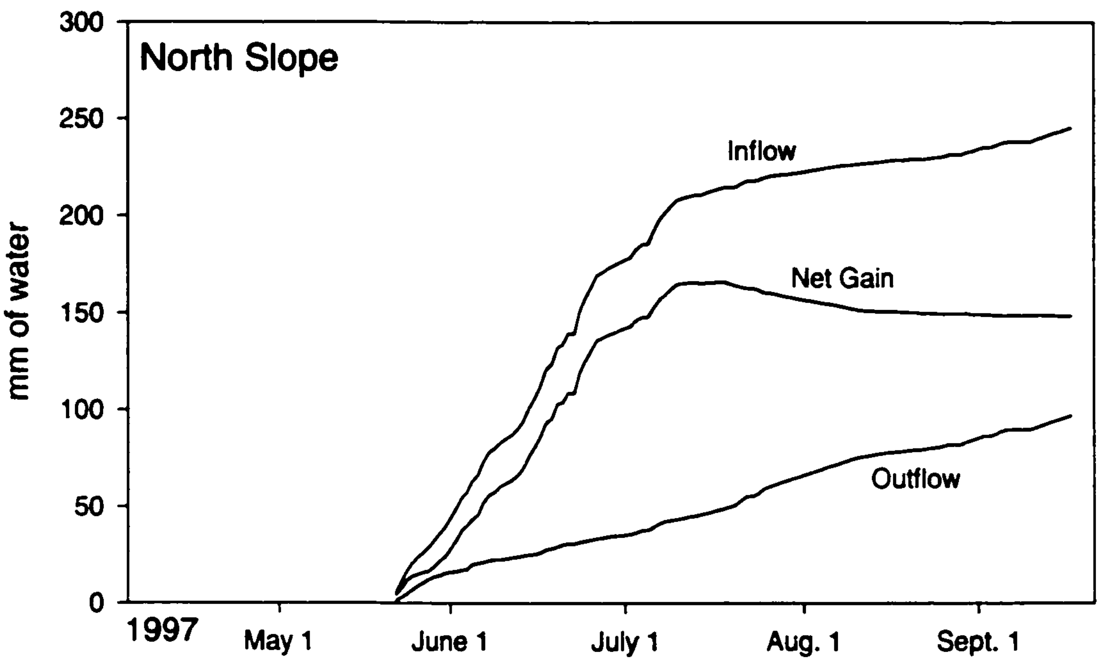

Figure 4.6 Cumulative subsurface inflow, outflow and net gain for the North slope, 1997.

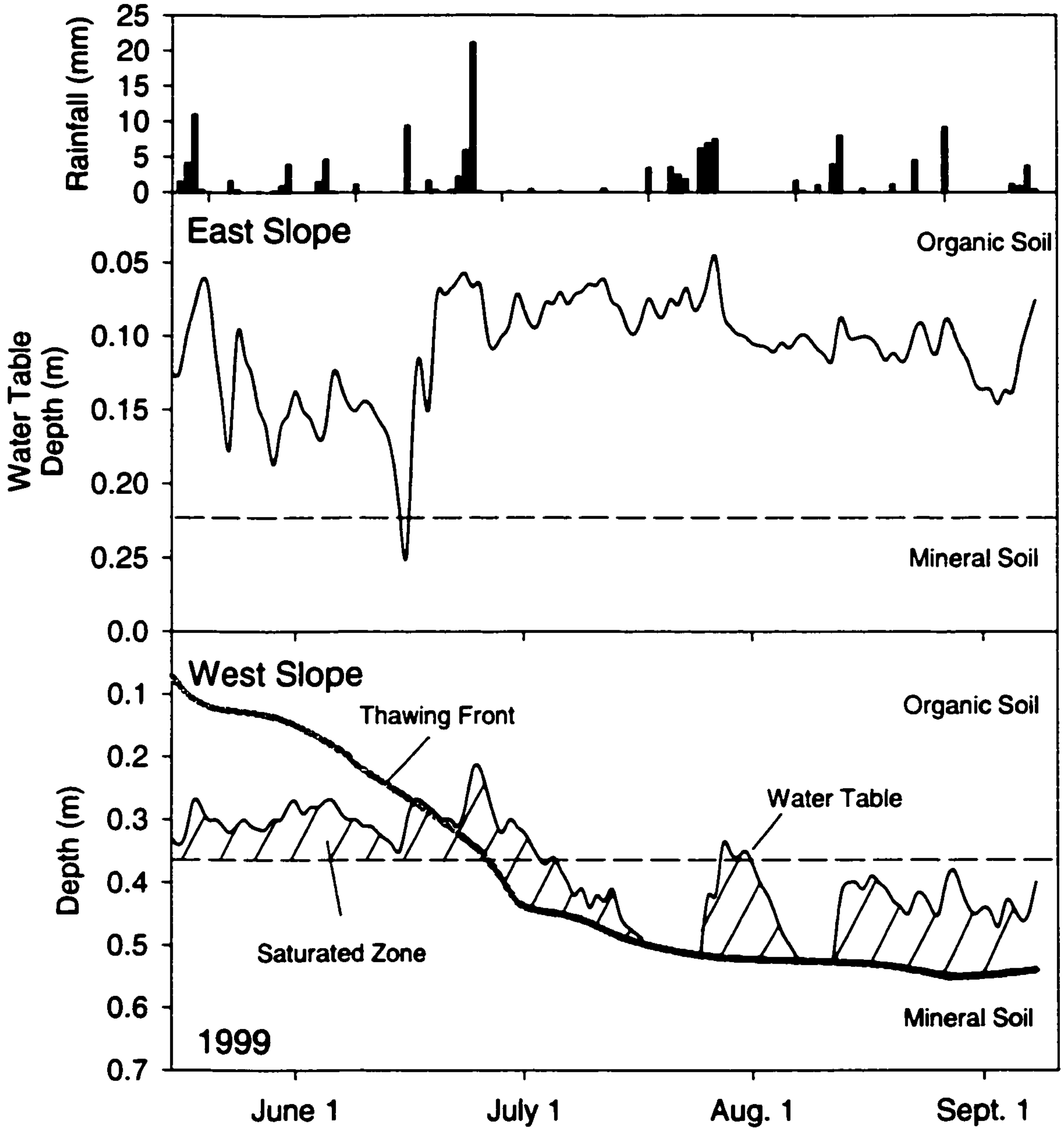

Figure 4.7 Daily rainfall, East slope water table position, West slope water and frost table position, 1999.

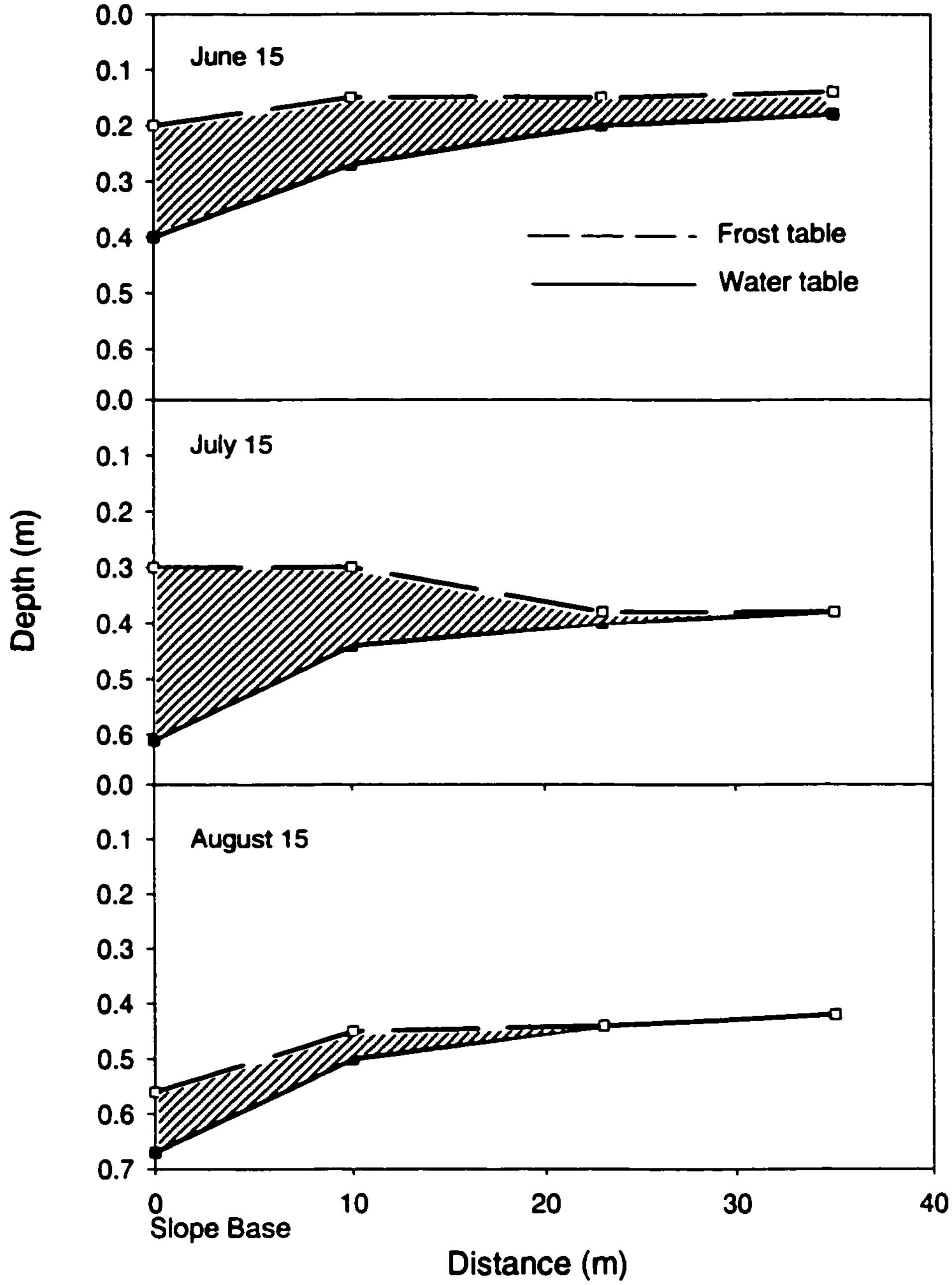

Figure 4.8 Position of the frost table (solid line) and water table (dashed line) at four sites on the West slope, 1997. The abscissa is the distance upslope from the bottom well.

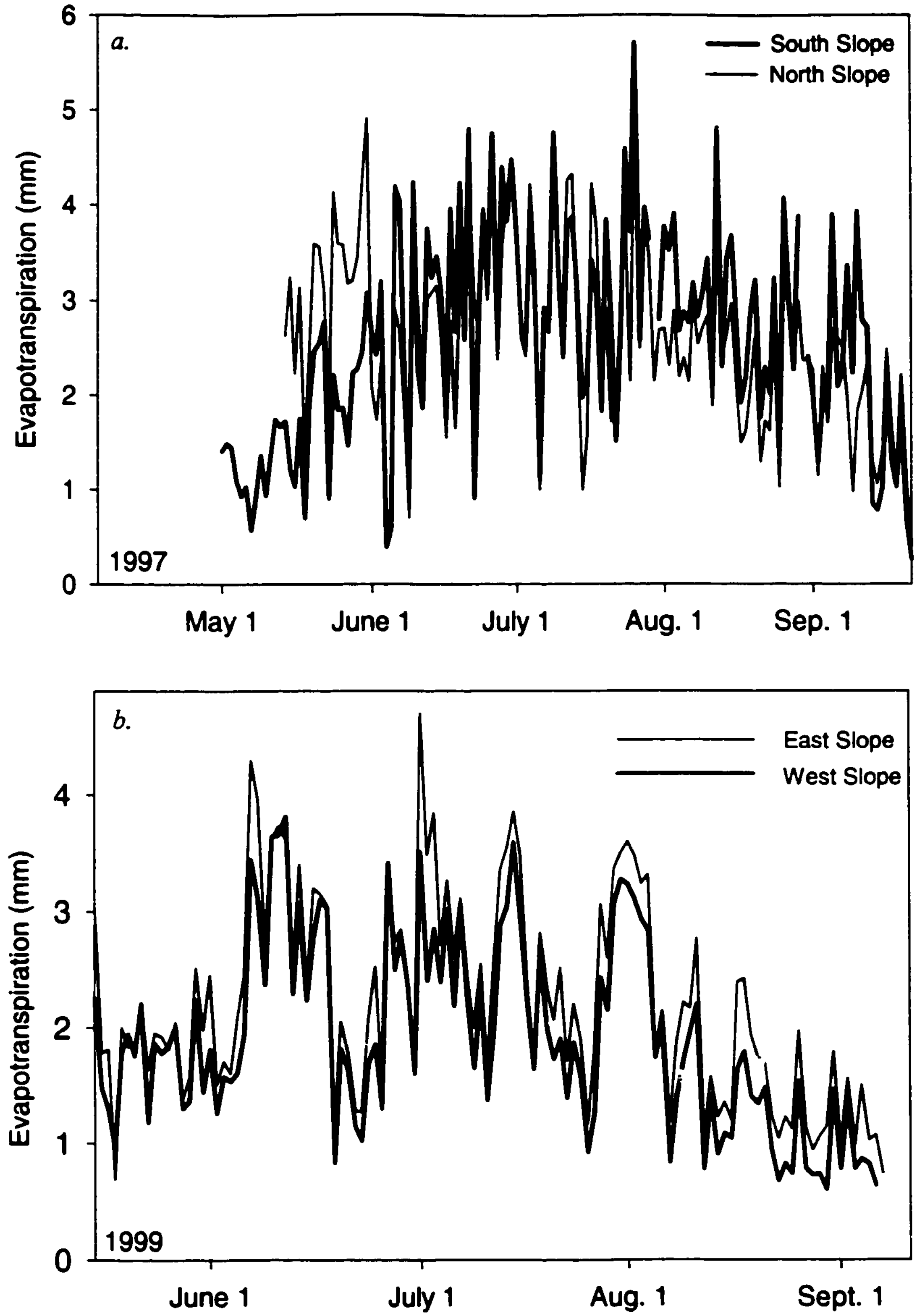

Figure 4.9 Daily evapotranspiration for (*a*) the North and South slope in 1997, and (*b*) the East and West slope in 1999.

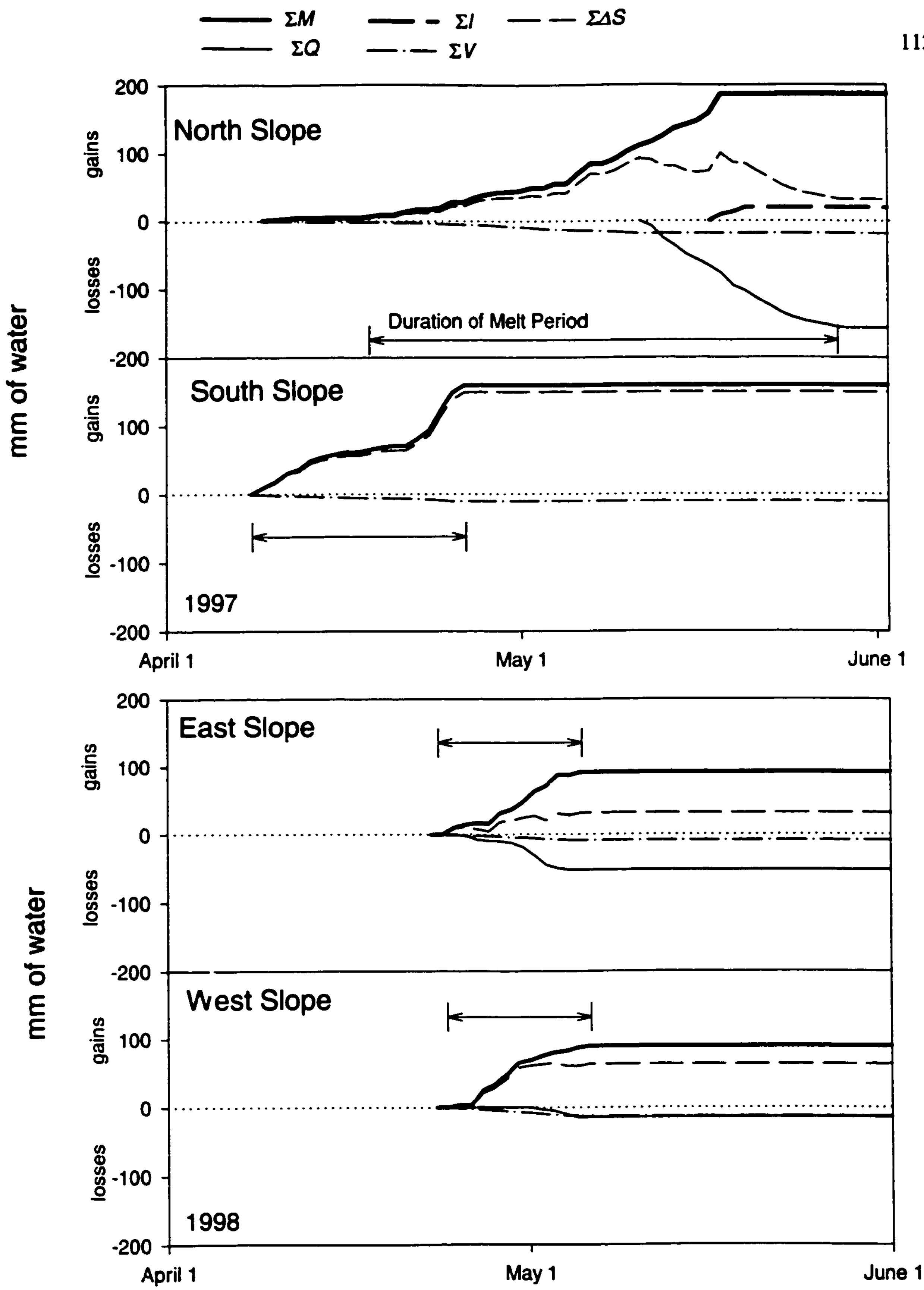

Figure 4.10 Cumulative values of daily water balance for the melt period. Symbols represent cumulative values of melt (ΣM), runoff (ΣQ), icing (ΣI), sublimation (ΣV) and storage change ($\Sigma \Delta S$).

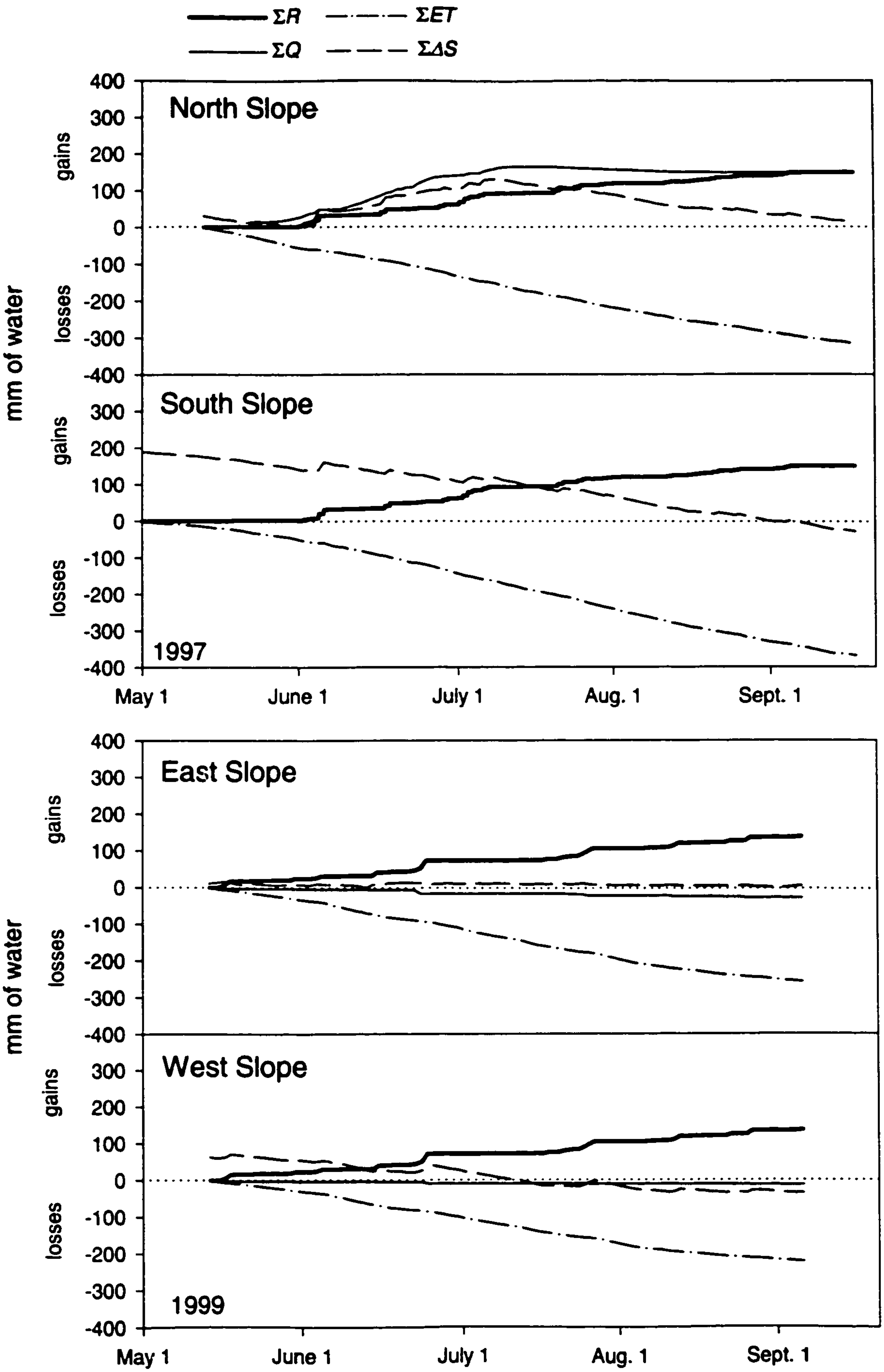

Figure 4.11 Cumulative values of daily water balance for the summer period. Symbols represent cumulative values of rain (ΣR), runoff (ΣQ), evapotranspiration (ΣET), and storage change ($\Sigma \Delta S$). Storage change is begins at the level representing recharge from melt.

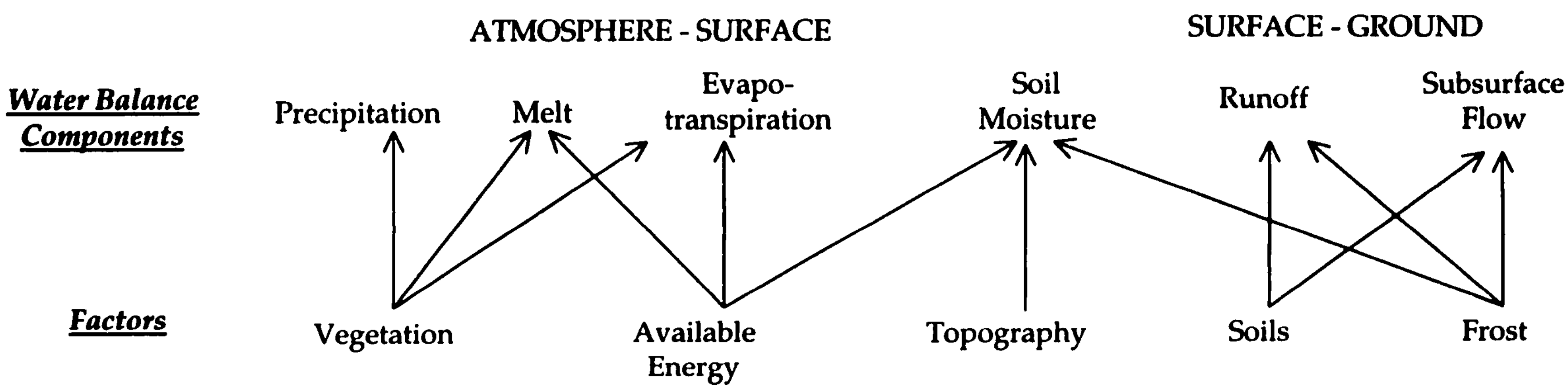

Figure 12. Principal interactions between environmental factors and water balance components.

CHAPTER FIVE

SLOPE RUNOFF PROCESSES AND FLOW GENERATION IN A SUBARCTIC, SUBALPINE CATCHMENT

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and

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Submitted to *Journal of Hydrology*

ABSTRACT

Hillslope runoff was studied in a subarctic, subalpine environment to improve understanding of runoff generation processes and the mechanisms whereby water moves from hillslopes to the stream. Runoff characteristics of four hillslopes were examined between 1997 and 1999, each with distinct soils, frost, topography and vegetation. Lateral fluxes were confined to hillslopes with porous organic soils overlying less permeable mineral substrates, setting up a two-layer flow system whereby most drainage occurs as quickflow in the porous organic layer as matrix flow and/or as preferential flow in pipes, rills and interconnected surface depressions. During snowmelt, meltwater infiltrated and percolated the porous frozen organic layers with little resistance. Percolation ceased at the organic/mineral interface due to the impermeable nature of frozen mineral soils, forming a perched saturated zone and initiating runoff. Snowmelt runoff increased on slopes with greater snow storage and reduced organic layer thickness. In summer, runoff was greatest on slopes where wet conditions were sustained by inflow. Stormflow hydrographs responded rapidly to rainfall while exhibiting extended recessions compared with temperate regions. Where sustained inflow occurred, the recession limb showed two segments to reflect different source areas of stormflow production. Recession analysis was used to quantify contributing areas which was highly variable and controlled largely by hillslope wetness and organic layer properties. Results indicate that the concept of variable source area for runoff generation applies to subarctic, subalpine catchments.

5.1.0 INTRODUCTION

Studies of streamflow and catchment hydrology in the subarctic have shown flashier flows in basins dominated by permafrost compared with more moderate responses from basins with only seasonal frost (Slaughter *et al.*, 1983). Dingman (1966; 1971) noted that the summer hydrographs have longer recession limbs that those in temperate regions and hypothesized that the properties of organic soils on permafrost-underlaid hillslopes control stormflow properties. Chacho and Bredthauer (1983) observed that non-permafrost areas in central Alaska yield virtually no near-surface runoff, and that a large portion of catchments, including some permafrost zones, do not contribute to runoff. Metcalfe and Buttle (1999) examined differences in sub-basin water balance in a boreal forest catchment with discontinuous permafrost and noted large differences in fluxes controlled by variations in snowpack properties, rainfall characteristics, ground thaw depths and storage properties.

Studies of runoff processes at the hillslope and plot scale report a wide range of mechanisms and pathways of lateral water movement. In the Canadian high Arctic, the presence of late-lying snowbanks and frozen soils are important in generating surface and subsurface flows throughout the short thaw season (eg. Woo and Steer, 1982; Lewkowicz and Young, 1990) resulting in runoff regimes similar to those found in glacierized basins (Marsh and Woo, 1981). In more southerly areas of continuous and discontinuous permafrost, the widespread presence of highly permeable organic soils limits the occurrence of overland flow (Dingman, 1973; Slaughter and Kane, 1979; Hinzman *et al.*, 1993; Quinton and Marsh, 1999). Subsurface flow in organic soils is the primary mode of water delivery from hillslopes to streams, although soil irregularities and preferential pathways such as soil pipes (Carey and Woo, 2000), water tracks (Kane *et al.*, 1989) and mineral earth hummocks (Quinton and Marsh, 1998) complicates runoff.

The common approach in cold-region runoff hydrology is to transfer concepts developed for temperate environments and modify them to explain the runoff regimes and processes observed at high latitudes. Lewkowicz and French (1982) suggested that wet conditions maintained downslope of snowbanks in the high Arctic indicated the applicability of the partial area concept of runoff generation (Betson, 1964). Dingman (1973) modified the variable source area concept (Hewlett and Hibbert, 1967) to describe runoff in the subarctic boreal forest zone of Alaska. There, catchment wetness, represented through the position of the water table, controls the extent of the near-stream areas that contribute to streamflow during summer rainfall events. In hummocky permafrost terrain with a peat cover that exhibits large vertical variability in hydraulic conductivity, Quinton and Marsh (1999) present a framework for runoff generation based on the variable source area concept. There, the extent of the contributing area is related to the water table position within the peat layer. When the water table recedes in late summer, the source area for streamflow becomes quite small as the enlargement of the storage capacity of the soils limits runoff production. When the saturated zone rises to the surface, water can flow rapidly through the porous organic soils and along the inter-hummock cracks to contribute to streamflow.

Previous runoff research in the discontinuous permafrost environment has focussed upon streamflow characteristics (eg. Dingman, 1971; Chacho, 1990) which are a reflection of the processes operating a the hillslope scale under the influence of different topographic, microclimatic, vegetation, frost and soil conditions. In environments that exhibit large variability in their physical settings over short distances, the hillslope is therefore the ideal scale to study runoff. At the hillslope scale, it is possible to relate runoff to the physical properties of the slope while eliminating the mixed signal that results from streamflow analysis. Using hillslopes that represent a wide range of hydrological conditions, the objective of this paper is to study runoff processes and the characteristics of source areas that contribute to streamflow in a subarctic, subalpine environment. By examining the characteristics of hillslope runoff and developing a framework for assessing source areas, the factors that account for the variability in runoff magnitude and timing on subarctic hillslopes can be determined.

5.2.0 Theoretical Considerations

Several components of the hydrograph that are considered for analysis are: 1. the response time defined as the time between the beginning of water input and the beginning of a measurable response (or hydrograph rise), 2. the rising limb, which is the component of the hydrograph from the onset of the response to the hydrograph peak, and 3. the recession limb extending from the hydrograph peak to the point where the flow returns to pre-event levels. Response time often reflects the wetness of a watershed or hillslope, whereas the shape and timing of rising limbs show close correspondence with the duration and intensity of storms (Dingman, 1994). Analysis of the recession flow characteristics have been applied widely to infer upon the behaviour of the storages that feed streamflow (for review, see Tallaksen (1995)). The general equation for recession flows derived from basin storage considerations can be applied to hillslopes. For a period with no storage recharge and neglecting evapotranspiration, the slope discharge (q) is

$$q = -\frac{dS}{dt} \tag{5.1}$$

where S is the volume of hillslope storage. Storage and discharge are commonly related as

$$q = cS^n \tag{5.2}$$

where n and c are empirical coefficients. Substitution of Eq. 5.2 into Eq.5.1 and integrating leads to the recession equations

where k is the saturated hydraulic conductivity and h is depth above the base and α is the slope. The volume of water in storage in the saturated layer is

$$S = \theta h l \tag{5.6}$$

where θ is the porosity of the layer and l is the length of saturation up the hillslope (indicating the contributing area).

Three cases can be considered:

(1) Constant contributing areas during recessions

If the length of saturated zone remains unchanged through the recession period (or l is constant), substitution of Eq. 5.6 into Eq. 5.1 gives

$$q = -\theta \frac{dh}{dt} \tag{5.7}$$

Differentiation of Eq. 5.5 yields

$$dh = \frac{1}{k\alpha} dq \tag{5.8}$$

substituting Eq. 5.8 into Eq. 5.7

$$q = -\frac{\theta}{k\alpha}\frac{dq}{dt} \tag{5.9}$$

Integrating and noting that $q=q_0$ when $t=0$

$$q(t) = q_0 e^{-\frac{k\alpha}{\theta}} \tag{5.10}$$

Thus, the recession curve for one reservoir with a constant contributing follows an exponential decay as depicted in Figure 5.1*a*. The upslope contributing area for a hillslope unit, l, can be calculated by substituting in Eq. 5.4

$$l = \frac{k\alpha}{\theta}t' \tag{5.11}$$

Here, $t'=t^*$ is the recession coefficient obtained from the hydrograph.

(2) Changing contributing area during recession

If the contributing area decreases continuously during recession, Eq. 5.11 indicates that t' will increase linearly. This leads to the gradual deviation of the recession limb from an exponential decay (Figure 5.1*b*). If changes in l are abrupt due to rapid disconnecting of flow contributing zones in the hillslope, the recession will comprise several segments, each of which characterized by a lower exponential value as the source area decreases.

(3) Multiple hillslope reservoirs

Runoff from hillslope may be generated from several storage reservoirs (eg. several soil aquifers), each with its recession coefficient, the discharge can be represented by the superposition of recession from these reservoirs:

$$q = \sum_{i=1}^{m} q_0 e^{\left(-t/t^*_i\right)} \tag{5.12}$$

The contributing areas for each reservoir, l_i, will be

$$l_i = \frac{k\alpha}{\theta} t^*_i \tag{5.13}$$

The observed recession coefficient t' will manifest the combined effects of t^*_i on the recession curve (Figure 5.3*c*).

The occurrence of surface and preferential turbulent flow (e.g. pipes) complicates the determination of contributing areas. Preferential flow will lower t' due to increased drainage efficiency. A decrease in t' indicates a reduction in l in Eq. 5.13, which is physically unreasonable considering that preferential flow can increase hydrologic connectivity on hillslopes by conveying water over extended distances. Therefore, the use Eq. 5.13 is only appropriate in circumstances when runoff pathways and the stormflow hydrograph are dominated by Darcian flow in the organic layer, which is the case for most summer storms on subarctic slopes (Dingman, 1971; Carey and Woo, 2000). Error also occurs as t^*_2 may influence the slope of t^*_1 (Figure 5.3*c*) if they are considered to be caused by separate hillslope reservoirs. Assuming that both reservoirs drain immediately following the hydrograph peak, t^*_2 will reduce the calculated value of t^*_1 (thus taken as t'_1), resulting in a smaller contributing source area than is estimated by Eq.5. 13.

Recession characteristics reflect the integrated influence of hillslope topography (α), contributing area (l) and soils (k and θ). Changes in t' provides insight into the aquifer storage and indirectly the effect of frost on stormflow. This analysis provides the theoretical basis to confirm the finding of Dingman (1971) and McNamara *et al.*, (1998) that t' is important in characterizing the flow from organic-covered hillslopes. In this study, recession analysis will be used as a tool for comparison of flow releases from hillslopes with different physical attributes.

5.3.0 EXPERIMENTAL DESIGN

5.3.1 Study Area

To investigate the variation of runoff in a subarctic, subalpine environment, four slopes within the Wolf Creek basin, (61° 31' N, 135°, 31' W)(Figure 5.2) were selected to represent a broad range of frost, soil and topographic conditions. Each hillslope was located within the subalpine woodland eco-zone of the basin where permafrost is discontinuous and vegetation consists of a mix of black and white spruce and willow shrubs on poorly drained slopes and aspen forests on well drained sites. On poorly drained hillslopes, an organic layer overlies less permeable mineral layers but well-drained slopes have only thin accumulations of leaf litter. Basin geology, hydrology, climate and ecology are summarized in Pomeroy and Granger (1999).

The study hillslopes are labelled according to their principal aspect. The physical setting of each slope is summarized in Table 5.1 and the properties of the soils are presented in Table 5.2. Carey and Woo (1999; Chapter 4), present detailed site descriptions and water balance information for these hillslopes. The South-facing (S) slope is distinct from the other three slopes in that it has a dense aspen forest but without a surface organic layer. The North-facing (N) and West-facing (W) slopes are underlain by permafrost and have similar topography. The N-slope has thinner organic soils, deeper thaw and often wetter conditions than the W-slope. The East-facing (E) slope has seasonal frost only and an organic layer which has a distinct acrotelm-catotelm gradation at ≈ 0.1 m. This slope receives lateral drainage from an upslope alpine area that supplies water to the study site throughout the year to maintain a high water table and enhance the runoff response.

5.3.2 Hydrological Measurements

Hydrological parameters were obtained at no less than four sites on each study slope. Horizontal hydraulic conductivity was determined for each soil layer using pumping tests conducted in wells and piezometers (Freeze and Cherry, 1979). Porosity and bulk density were determined using gravimetric methods. Ice content was estimated by chipping samples from different soil horizons and then thawing the samples, subtracting the unfrozen soil moisture content by time domain reflectometry (TDR). In addition, ice content was visually estimated from the chipped frozen samples.

Soil moisture was measured using calibrated TDR probes (MoisturePoint on N, S-slope; Campbell CS-615 on E,W-slope) buried at 0.05, 0.1, 0.2, 0.4, and 0.6 m at three sites on the N-slope and two sites on the S-slope with an additional sensor at 1 m. On the E and W-slopes, soil moisture was measured at depths of 0.1, 0.2, 0.4 m (W-slope) and 0.1, 0.2, 0.6 m (E-slope). In 1997, water tables were monitored continuously with electronic water level recorders at three sites on the N-slope and manually measured daily at 12 additional sites. On the E and W-slopes, water level was monitored continuously in 1999 at one well and manually in 1997 at an approximately weekly basis (at four wells). On the S-slope, four observation wells were dug to depths of 2 m without reaching the saturated zone. At no time between 1996 and 1999 was a saturated zone observed in any of these wells.

Slope runoff was measured on the N, E and W-slopes using flumes dug into the hillslope base. These flumes were boxes approximately 0.6 m long, 0.4 m wide with a 90°V notch. Water was directed by plastic sheeting into the flumes. Sampling schemes for these flumes varied. On the N-slope during melt period in 1997, a graduated cylinder was used to measure discharge at three hour intervals. During the summer of 1997, hillslope runoff was recorded by a stage-discharge rating curve obtained by converting the half-hourly water level with an accuracy to ±5 ml/s. This method was also used to measure snowmelt runoff on the E and W-slopes in 1998. In 1997, runoff was directly measured weekly at the slopes. Additionally, intense measurement periods in 1997 on the E and W-slope provided detailed information on the stormflow characteristics of six rainfall events, including the recording of the water level in the flumes using Stevens type-F water level recorders. Weekly measurements were used to calibrate stage vs. discharge relationships. In 1999, runoff from rainfall events was calculated as a residual of the water balance (Chapter 4). The difference between the rate of soil moisture loss following storms at the auto-well/TDR site and the evaporation rate was considered as runoff.

5.3.3 Hydrograph Analysis

At no time did the S-slope yield any lateral flow during the study period. Stormflow hydrographs from the N, W and E-slopes in 1997 were examined to assess rainfall-runoff characteristics. The events selected were summer storms that produced a single runoff peak, and had limited rainfall during the flow recession period. Discharge was separated into stormflow and baseflow by projecting a straight line across the hydrograph from the point of initial rise to the point where discharge returns to its antecedent level. Properties of the rainfall hyetograph and stormflow hydrograph were computed. Values of t' were determined from the falling limbs of storm hydrographs by regressing correlations of $\ln(q)$ against t for points on the falling limb. The falling limb of each hydrograph was visually inspected and broken into components when there was a notable divergence on plots of $\ln(q)$ vs. t. Values of t'_i were determined as above for each portion of the recession limb.

5.4.0 RESULTS

5.4.1 Snowmelt Runoff

Snowmelt runoff was measured on the N-slope in 1997 and the E and W-slopes in 1998. The absence of observed flow from the S-slope suggests that these well drained hillslopes rarely generate runoff, or perhaps only sporadically when the soil was very wet in the freeze-back period. Snowmelt runoff exhibited two major trends: 1. a general increase in runoff during the main melt period corresponding to the wetting-up of the organic layer followed by a rapid decline after the snow disappears, 2. diurnal fluctuations superimposed upon the seasonal trend, responding to the daily rhythm of melt (Figure 5.3).

In 1997, snow water equivalent (SWE) on the N-slope (187±24 mm) was approximately twice the E (92±34 mm) and W (90±22 mm) slopes in 1998. The main snowmelt runoff period for the N-slope was also longer, extending over 14 days compared with 10 and 4 days for the E and W-slopes. On the N, W and E-slopes, runoff began following the formation of a saturated zone at the base of the organic layer. Ice-rich layers and a discontinuity in the soil properties restricted percolation into the underlaying mineral soils and the perched saturated zone rose towards the ground surface (Figure 5.4). Saturation began one to two weeks following the initiation of snowmelt infiltration (Carey and Woo, 1998; Chapter 4). The delay between infiltration and runoff was caused by the re-wetting of organic layers that were desiccated over the winter from upward vapour flux out of the soil (Santeford, 1979*a*; Smith and Burn, 1987). Within the organic layer, pre-melt total water (frozen + unfrozen) contents for the N, E and W-slopes were 18 % (40 mm), 34 %, (55 mm) and 26 % (58 mm), which indicated significant drying after freeze-back. Saturation of these organic soils and runoff generation began after 102 mm (N-slope), 20 mm (E-slope) and 60 mm (W-slope) of meltwater infiltration, bringing the total water content to 142 mm (N-slope), 75 mm (E-slope) and 118 mm (W-slope).

Runoff followed several pathways that reflected increase hydrological connectivity as the ground thawed. Rills, soil pipes and diffuse flow within the organic layer were the principal modes of flow that conveyed water to the stream at different rates. Despite similar patterns of melt, peak daily runoff for the N and E-slopes occurred at approximately 2130 and 1900, PST, with daily minima for both slopes at 0800. The diurnal cycles on the W-slope were weak and the time of the daily high and low flows cannot be estimated, although visual inspection suggests that they lag slightly behind the E-slope. The weak diurnal signals are attributed to its thick organic layers which increased the time of water percolation to reach the water table and to the lack of preferential flow paths. For the N-slope, Carey and Woo (2000) showed that runoff from soil pipes lagged peak snowmelt by 2-4 hrs, whereas integrated hillslope runoff lagged by 6 hrs. Total snowmelt runoff for the N, E and W-slopes was 155 mm, 50 mm and 19 mm respectively. The ratio of runoff to SWE (termed runoff ratio) for these slopes was 0.8, 0.6 and 0.25.

Meltwater can infiltrate the highly porous frozen organic layer, but runoff will not begin until the moisture retention capacity θ_{sr} is satisfied (Table 5.2). Maximum unsaturated water storage capacity are 102 mm (N-slope), 97 mm (E-slope) and 133 mm (W-slope). These values compare well with the total water content measured after snowmelt, 77 mm (N-slope), 100 mm (E-slope) and 137 mm (W-slope). Together with the SWE and pre-melt soil water content, the water holding capacity is useful in predicting the runoff response from snowmelt.

5.4.2 Summer Runoff

Runoff at the base of the N-slope was monitored from May 12 to September 17, 1997, with a break between August 12 to 30. On the E and W-slopes, runoff was measured in 1997 at no less than a weekly basis between June 10 and August 12 along with six intensive measurement periods. In 1999, runoff ratios were determined for the E and W-slopes as a residual of short-term water balances. At no time during 1996 or 1997 was the water table within 2 m of the surface within the S-slope, and geophysical investigations indicated that the water table was >6 m below the slope surface. A similar condition was documented in Manners Creek, NWT (Gibson *et al.*, 1993) where the water table was >3.4 m below non-permafrost slopes.

5.4.2.1 North Slope

Following the disappearance of snow, the water table dropped to the organic-mineral interface where thawing of the ice-rich soils retarded the descent of the frost table. At this time, upslope hydrological connections were reduced due to uneven ground thaw and small localized ponding where drainage was impeded (Carey and Woo, 2000). Average daily runoff at the base of the N-slope decreased abruptly after melt but gradually increased between June 1 and July 1 (Figure 5.5*a*). Then, runoff steadily declined until mid-September. Low rainfall in the fall of 1996 and 1997 yielded the lowest flows of the season just prior to freeze-back, which began in mid-October in both years. Superimposed on the seasonal trend are spikes that represent enhanced flows from rainfall. Using Darcy's equation to calculate groundwater inflow and outflow at the base of the N-slope, Carey and Woo (1999) noted upslope hydrological connectivity allowed water to enter the lower slope study segment and produced a net water gain, resulting in inflow in May and June. In July, water tables declined into the mineral soils with relatively low hydraulic conductivities and upslope connections became greatly reduced, and runoff gradually decreased through the remainder of the summer.

5.4.2.2 West Slope

Runoff from the W-slope measured at approximately weekly intervals in 1997 showed similar trends to the N-slope although at much smaller flow magnitudes (Figure 5.5*b*). When measurements began on June 10, runoff was small (6 ml/s), increased to a maximum of 22 ml/s on June 17 and then declined until mid-July when continuous drainage ceased. The spatial variability of the saturated zone in 1997 indicated that this slope drained as a thin saturated wedge which shrank throughout the summer (Chapter 4). Following the cessation of continuos runoff, only rainfall events of sufficient magnitude (> 4 mm) were able to induce direct runoff (Figure 5.6*a*). Rainfall-runoff analysis for the 1999 events show the ratio of runoff to precipitation gradually decreased in the summer (Figure 5.6*b*). When the water table resided within the organic layer, runoff rations were between 0.1 and 0.2 while the absolute value of runoff was related to antecedent wetness and storm intensity. Between late June and mid-July, water tables fell within the mineral soils and more water was required to initiate runoff response, lowering the runoff ratio to < 0.1.

5.4.2.3 East Slope

Flow measured at the base of the E-slope was an order of magnitude greater than the N and W-slopes and exhibited less seasonal variability (Figure 5.5*b*). As with the N and W-slopes, runoff increased immediately following melt, yet continued steadily into August with only a slight decline in September. Sustained summer runoff was due to its extensive contributing areas and the lengthy travel times of runoff from the upslope zone to reach the study segment. Rainfall-runoff analysis in 1999 showed that rainfall events > 2.5 mm were sufficient to generate a runoff response (Figure 5.5*a*). The enhanced wetness kept the water table within the organic zone throughout the observation periods between 1997 and 1999. Large scatter in the runoff ratio indicate the sensitivity of flow response to rainfall magnitude as slope wetness was similar throughout the study period (Figure 5.5*b*).

5.4.3 Hydrograph responses to rainstorms

Hydrographs from the N, W and E-slopes were examined in 1997 to assess rainfall-runoff characteristics (Table 5.3). Inability to estimate a micro-catchment area for the E-slope following the melt period precluded the determination of runoff coefficients from its hydrograph.

The amount of precipitation required to initiate stormflow, termed initial abstraction (P_{abst}), was higher for the W than the N and E-slopes. P_{abst} for the E-slope was the lowest, ranging between 0.5 and 8.1 mm with an average of 2.3 mm. The largest value (8.1 mm) was associated with a high intensity rainfall burst at the beginning of the event. P_{abst} on the N-slope ranged between 0.9 and 4.1 mm, which contrasts the 4-9.8 mm range for the W-slope that has a thick organic layer with much storage capacity. P_{abst} was not correlated with properties of hillslope wetness such as antecedent discharge (Q_{ant}) or five-day antecedent precipitation (P_{5d}). The lack of correlation between P_{abst} and watershed wetness has been observed elsewhere for subarctic and arctic watersheds (Dingman, 1971; McNamara *et al.*, 1998) where subsurface storage capacity is limited by topographic position of the presence of permafrost.

Church (1974) noted the rapid response of basins dominated by permafrost. For the experimental slopes, other factors should be considered. The hydrograph response times (T_{resp}) were shortest for the permafrost-free E-slope (0.5-3.5 hr) and longest for the W-slope (2-12 hr) where the permafrost averaged 0.6 m below the surface. Rapid hydrograph rise at the E-slope was due to wet near-surface conditions, requiring little additional moisture to initiate runoff. Average T_{resp} for the N, E and W-slopes were 3.9, 1.5 and 6.3 hr. These can be compared with the response time of 2.15 hr for a small (2.2 km^2) catchment in northern Alaska (McNamara *et al.*, 1998) where a network of water tracks remains saturated for much of the summer (Kane *et al.*, 1989).

As in much larger subarctic basins, the time of hydrograph rise (T_{rise}) on the E-slope closely corresponded to the duration of precipitation (P_{dur}), whereas on the W-slope, hydrograph rise was on average approximately twice P_{dur}. Extended hydrograph rise on the W-slope reflected the increased transmission time through the thick unsaturated organic soil. On the N-slope, there was large variability among storms with T_{rise} being both greater and less than P_{dur}. For the long events of June 4 and July 3, the hydrographs peaked before the end of the rainstorms, although rainfall following the peaks was very small (< 0.5 mm on both occasions).

The response factor (*RF*) expressed as the ratio of direct runoff to rainfall, exhibited considerable variability both among the slopes and within the summer seasons. Direct runoff on the W and N-slope was obtained by subtracting the baseflow from the hydrograph and dividing the volume of water by the catchment area delineated from topographic surveys. In 1999, direct runoff was determined as a residual of the water balance on the E and W-slope study segments. For the N and W-slopes in 1997, *RF* decreased after the widespread descent of the water table into the mineral layer, suggesting that as thaw depth increased, the storage capacity increased at the expense of runoff. Before June 17, *RF* was > 0.4 for the N-slope, and when the extended recessions are included, *RF* exceeded 0.6. As the water table dropped into the mineral layer, *RF* declined to an average of 0.18. On the W-slope, *RF* were approximately half that of the N-slope due to the large storage capacity of its organic soil, yet the same seasonal trend was present. In both 1997 and 1999, the *RF* fell from > 0.2 to < 0.15 as water and frost table declined in mid-summer. In 1999, *RF* from the E-slope ranged between 0.02 and 0.45 with an average of 0.19 (Figure 5.6*b*). No seasonal trends were apparent as the slope maintained a high water table throughout the summer. However, runoff from the study segment showed a positive relationship with rainfall (Figure 5.6*a*).

5.4.4 Recession Characteristics

5.4.4.1 Snowmelt recessions

After the removal of the snowpack, rapid drainage of the surface organic layer occurred before thawing of mineral soils began. Recessions were calculated from the falling limb of the melt hydrograph (Figure 5.3*a,b*), as this recession should reflect a gradual drainage of the organic layer storage. On the N-slope, the recession between may 22 and 26 closely followed an exponential decay, yielding t'= 40 h. Irregularities along the recession limb are probably due to the changes in the source area and hydrological connections along the slope (Carey and Woo, 2000). Recessions for the E and W-slopes are less well represented by exponential decays. Fluctuating recessions are possibly related to evaporation and variable meltwater contribution from the shrinking residual snow patches. Taken between 1600 May 3 and 1200 May 5, t' for the E and W-slopes are 20 and 30 hr respectively.

5.4.4.2 Recessions from summer rainstorms

The recession curves for summer hydrographs often showed two distinct segments; an initial steep portion occurring between 17 and 66 hr after the event peak, followed by a gentler decline lasting between 35 and 159 hr (Figure 5.7).

Recession limbs were separated into segments wherever a plot of ln(q) vs. t showed a distinct break. Recession coefficients (t'_i) for each storm were determined for each segment. Except for two storms in June on the N-slope, the W and N-slopes produced only once recession segment, although t'_1 values varied among storms. The June 4 and 17 storms on the N-slope yielded a sharp flattening of the recession curve, but the extended recessions (t'_2) were poorly represented by exponential decays. All recession curves from the E-slope showed two distinct slopes that closely followed the exponential decay. The exception was t'_2 for the June 17 storm because 3 mm of rainfall occurred during the recession period.

Table 5.3 summarizes the t'_1 coefficients for the initial separation of the recession. It ranged between 21 to 75 hr for the N-slope, 27 to 65 hr for the W-slope and 55 to 86 hr for the E-slope. Values of t'_1 showed weak seasonal trends, being relatively small in early summer and increased to a maximum around July 1 when the hillslopes had water tables nearest the surface in the summer period. Values of t'_1 compared well with other permafrost and organic covered, steep (>10°) basins (eg. t'=39 hr for Glenn Creek (Dingman, 1971) and ranges from 20 hr to > 100 hr in northern Alaska (McNamara *et al.*, 1998), increasing with drainage area).

Recession coefficients for the extended segment (t'_2) averaged 145 hr for the June 4 and 17 storms on the N-slope and 406 hr for the six events on the E-slope (Table 5.3). t'_2 for the E-slope showed a large range in values, with a three-fold difference between the maximum and the minimum values. The seasonal trend followed that of t'_1, with highest t'_2 for the June 28-29 event and the lowest for the September event.

5.5.0 HILLSLOPE-STREAMFLOW CONNECTIONS

The analysis of recession characteristics of event hydrographs for hillslope runoff permits an indirect assessment of the factors that influence stormflow contributing areas. Section 2.1 provides the rationale which can be applied to the interpretation of the field results given in section 4. The contributing areas for slope runoff can be distinguished into those that contribute to the initial recession period (l_1) and those that contribute to the extended recession phase (l_2). In view of the hydraulic conductivity of the mineral soils (Table 5.2), runoff contribution to stormflow from the mineral layer is considered negligible and it is reasonable to consider the runoff responses to be generated only by flows through the organic soils.

Eq. 5.13 and the values of k and θ were used to estimate l_1 and l_2 for various rainstorms.

Values of k and θ were obtained from the layer nearest the surface in which the water table resided at the base of the study area at the time of peak runoff for a storm event.

For the N-slope, l_1 ranged between 514 m on July 4 and 144 m on June 4, the latter low l_1 value may be partly due to the shallow ground thaw and intense rainfall which activated a small number of soil pipes on the hillslopes (Carey and Woo, 2000), invalidating the assumptions in Eq. 5.13. After this event, there was an increase in l_1, followed by a general decline throughout the remainder of the summer with a high degree of variability. The seasonal trend of l_1 suggests that the lower slope underwent increased drying, accompanied by an enlargement of storage capacities of the organic layer. When stormflow is initiated, smaller areas contribute to runoff as more moisture is required to overcome the storage deficit. The lack of correlation between antecedent wetness and T_{resp} does not at first appear to correspond with the above suggestion of hillslope drying. However, hydrograph response times are controlled by the wetness of the lower slope, which remained near saturation for the entire period. Thus, although the slope base continues to respond quickly to rainfall events, successively smaller upslope areas contribute to stormflow as the summer progresses. On the W-slope, l_1 values were small yet their reduction as summer progressed corresponds to observations that the water table moved into the mineral layer progressively downslope throughout the summer (Chapter 4). Values of l_1 for the E-slope were similar to those for the N-slope, but without a noticeable seasonal trend because of its prevalent wet conditions.

Extended contributing areas, l_2, determined from t'_2, occurred on the N-slope for two June storms when ground thaw was shallow. At this time, the occurrence of the water table within the organic layer across most of the experimental area allowed extensive hydrological connections, capturing water from a large upslope area to feed the lower slope study segment over an extended period. For the E-slope, extended recession flows prevailed throughout the summer, and values of l_2 were an order of magnitude greater than l_1, but showed a reduction in value from a maximum of 2792 m on June 28 to a minimum of 891 m on September 5. The large and sustained values of l_2 reflect a large contributing area upslope of the study segment producing runoff that followed pathways in the organic soil over a period of days.

Analysis of contributing areas suggests that subalpine, subarctic hillslopes conform closely to the variable source area concept of runoff generation. When hillslopes are wet and the water table rises within the porous organic layer over most of the hillslope, stormflow may be generated from two separate source areas, each with its own storage reservoir that feeds the reccssion flow: 1. a lower slope zone that provides for the fast response, and 2. zones further upslope that convey water through the saturated organic layer to the hillslope base over extended periods. When the drainage at the hillslope base occurs faster than water supplied as inflow from upslope areas, a gradual drying and water table decline occurs at areas upslope while positions near the hillslope base retain a saturated zone in the organic layer so that only low-slope zones contribute to runoff. The concept of runoff generation dependency on water table elevation and flow is similar to those proposed by Dingman (1973) and Quinton and Marsh (1999); and is analogous to wetness-dependent basin hydrological connectivity described by Roulet and Woo (1988) for a continuous permafrost environment.

5.6.0 THE TWO-LAYER FLOW SYSTEM

Where present, surface organic soils play a dominant role in hillslope runoff in subarctic and arctic environments. This is because of their ability to transmit water rapidly from the hillslope to the drainage network as that they strongly affect: 1. the timing and magnitude of slope runoff (Santeford, 1979*b*; Hinzman *et al.*, 1993), 2. the shape of the streamflow hydrograph (Dingman, 1971; Chacho, 1990), and 3. the extent of the source area of runoff generation (Dingman, 1973; Quinton and Marsh, 1999).

Slopes with an organic cover above the mineral soil exhibit a two-layer flow system consisting of quickflow and slowflow. Quickflow is defined as rapid runoff delivered downslope by matrix flow through highly porous organic material and by preferential flow in interconnected surface depressions (e.g. inter-hummock cracks and hollows), soil pipes and rills. Quickflow regime ranges from laminar flow in porous organic soils to turbulent flow along preferential pathways. Slowflow is defined as laminar flow in the saturated matrices of soils with low hydraulic conductivities (e.g. clayey to silty soils or highly humified and compacted organic materials) with flow velocities that are orders of magnitude lower than quickflow.

5.6.1 Snowmelt Period

In a two-layer flow system, the top organic layer is usually rendered porous by upward vapour flux in winter so that in the snowmelt period, some meltwater can infiltrate and percolate the organic layer under frozen conditions. The impervious frozen mineral substrate with its pores largely sealed by ground ice prevents further percolation and a saturated zone is perched in the organic layer. At this time, the organic layer can be considered an unconfined aquifer without any connections to unfrozen water beneath the seasonally frozen ground or deeper sub-permafrost aquifers. Within the organic layer, quickflow can follow two pathways: 1. Darcian flow in the porous matrix, and 2. preferential flow along pipes, rills and interconnected depressions. Of these pathways, matrix flow is the most widespread as soil pipes and rills are more intermittent features and tend to be associated with steeper slope sections and thinner organic layers. The mechanisms of pipeflow have been discussed by Carey and Woo (2000) and will not be presented here. Uneven ground-thaw during snowmelt hinders extensive hydrologic connections on the slopes, and ponding may occur where drainage is locally restricted. The volume of runoff during the melt period is principally controlled by the snow water equivalent and the liquid water storage of the organic soil. Differences in runoff timing among hillslopes may be caused by: 1. organic layer thickness which affects the time lag between infiltration and lateral flow generation, 2. the presence of preferential flow pathways that can convey water rapidly downslope, and 3. differential melting of snow on the slopes. Runoff ratio increases with snow storage on the slope and with thin organic soil which limits the water holding capacity. Following the disappearance of the snowpack, the organic layer drains rapidly and its water table declines to the organic-mineral interface which tends to be ice-rich and therefore retards thawing due to the large latent heat requirements to melt the ground ice.

5.6.2 Summer Period

As the thawing front enters the mineral soils, the perched saturated zone also subsides. Quickflow pathways near the surface (eg. rills), become inactive, and gradually quickflow in the lower parts of the organic profile also diminish. Slowflow begins in the mineral soils as well as the compacted lower zone of the organic layer. In the study area, only the E-slope presents notable differences in physical properties between upper and lower organic layers. Further descent of the frost table, accompanied by moisture draw-down by evapotranspiration and drainage, cause the water table to pass from the organic into the mineral layer, beginning from upslope locations and progressing downslope throughout the summer. Exceptions are for slopes with significant lateral inflow from upslope source areas (such as the E-slope). Slowflow in the mineral soil contributes little to stormflow due to the limited transmissivity of this layer. Summer rainstorms of sufficient magnitude raise the water table into more porous organic zone where quick-flow can recur. In the late summer, water tables may fall well within the mineral soils or disappear altogether, reducing the ability of rainfall to generate stormflow.

The properties of the organic layer strongly influence the shape of the hydrograph and the volume of slope runoff. Depending upon the antecedent storage in the organic soil, runoff initiation requirements vary from storm to storm. Once runoff is initiated, water is conveyed quickly downslope in the highly transmissive organic layer to produce steep hydrograph rises. Water stored in the organic layer is released over extended periods, resulting in relatively long hydrograph recessions. The slow releases of water from organic soils has been noted also in subarctic Alaska (Dingman, 1971; Santeford, 1979*b*; Kane *et al.*, 1981).

5.6.3 Hillslopes without organic soils

The S-slope offers a strong contrast with the other experimental slopes in the study basin. This slope, without organic soil and devoid of permafrost, has good drainage but no observed saturated zone down to at least 2 m. Such slopes do not generate runoff, except perhaps in very wet years when ice-rich seasonal frosts may form near the ground to inhibit snowmelt infiltration. In terms of basin hydrology, these well drained hillslopes without permafrost or external sources of water to sustain wetness rarely contribute runoff to streamflow.

5.7.0 DISCUSSION AND CONCLUSIONS

This study represents the first attempt to apply recession analysis to slope runoff in a subarctic environment to assess runoff contributing areas at the hillslope scale. Information provided is important for studies of catchment hydrology in subarctic, subalpine settings as it provides an approach for quantifying the contributing source area, and relating it to the processes of hillslope runoff generation.

Hillslopes with an organic layer overlying mineral soil exhibit a two-layer flow system due to a sharp change in saturated hydraulic conductivity at the soil interface. This enables most of the hillslope drainage to occur as quickflow in the porous organic layer via matrix flow and/or as preferential flow in pipes, rills and interconnected surface depressions. It is the quickflow that controls the shape of the stormflow hydrograph which is characterized by fast response and an extended recession. For any storm event, the length of the slope that contributes to stormflow varies according to antecedent wetness and the drainage characteristics of the slope. Where drainage from a low-slope segment exceeds inflow, upslope source areas gradually become detached from the drainage network as drying progresses and the water table passes into the mineral layer. Then, only slowflow follows. On slopes where inflow to the footlsope and near-stream zones is large, a high water tables is maintained; and this sustains hydrological connections in the organic layer through quickflow to deliver water form upslope over protracted periods. We propose that this two-layer flow system enables the variable source area concept of runoff generation to operate on many subarctic, subalpine hillslopes.

5.8.0 REFERENCES

Betson RP. 1964. What is watershed runoff? J. Geophys. Res. 69, 1541-1552.

Carey, S.K., Woo, M.K., 1998. Snowmelt hydrology of two subarctic slopes, Southern Yukon, Canada. Nordic Hydrol. 29, 331-346.

Carey, S.K., Woo, M.K. 1999. Hydrology of two slopes in subarctic Yukon, Canada. Hydrol. Processes 13, 2549-2562.

Carey SK, Woo MK. 2000. The role of soil pipes as a slope runoff mechanism, subarctic Yukon, Canada. J. Hydrol. 223, 206-222.

Chacho, Jr. E.F. and Bredthauer, S. 1983. Runoff from a small subarctic watershed, Alaska. In Proceedings, Fourth International Conference on Permafrost, National Academy Press, Washington, D.C., pp. 115-120.

Chacho Jr., E.F. 1990. Water and suspended solids discharge during snowmelt in a discontinuous permafrost basin. In: Proceedings of the 5th Canadian Permafrost Conference, National Research Council, Ottawa, pp. 167-173.

Church, M., 1974. Hydrology and permafrost with reference to northern North America, In: Proc. Workshop Seminar on Permafrost Hydrology, Canadian National Committee, IHD, Ottawa, pp. 7-20.

Dingman, S.L., 1966. Characteristics of summer runoff from a small watershed in central Alaska. Water Resour. Res. 2, 751-754.

Dingman, S.L., 1971. Hydrology of Glenn Creek watershed, Tanana basin, Central Alaska. U.S. Army Cold Region Research Engineering Laboratory Research Report 297, Hanover, New Hampshire, p. 112.

Dingman, S.L., 1973 Effects of permafrost on stream flow characteristics in the discontinuous permafrost zone of central Alaska, In: North American Contribution to Second International Conference of Permafrost, National Academy of Sciences, Washington, D.C., pp. 447- 453.

Dingman, S.L., 1994. Physical Hydrology. Prentice Hall, New Jersey, p. 575.

Freeze, R. A. and Cheery, J.A., 1979. Groundwater, Prentice-Hall Inc., New Jersey, p. 604.

Gibson, J.J., Edwards, T.W.D., Prowse, T.D., 1993. Runoff generation in a high boreal wetland in northern Canada. Nordic Hydrol. 24, 213-224.

Hewlett, J.D., Hibbert, H. 1967. Factors affecting the response of small watersheds to precipitation in humid regions. In Sopper, W.E., Lull, H.W. (eds.), Forest Hydrology. Permagon Press, Oxford, pp. 275-290.

Hinzman, L.D., Kane, D.L., Everett, K.R., 1993. Hillslope hydrology in an Arctic setting, In: Proceedings, Sixth International Conference on Permafrost, South China Press, Beijing, pp. 257-271.

Holtan, H.N., Overton, D.E., 1963. Analyses and application of simple hydrographs. J. Hydrol. 2, 309-323.

Kane, D.L., Bredthauer, S.R., Stein, J. 1981. Subarctic snowmelt runoff generation. In T.S. Vinson (ed.), Proceedings of the Specialty Conference on The Northern Community, ASCE, Seattle, Washington, pp. 591-601.

Kane, D.L., Hinzman, L.D., Everett, K.R., Benson, C.S., 1989. Hydrology of Imnavait Creek, and arctic watershed. Holarctic Ecol. 12, 262-269.

Lewkowicz, A.G., French, H.M. 1982. The hydrology of small runoff plots in an area of continuous permafrost, Banks Island, N.W.T. In: Proc. 4th Canadian Permafrost Conference, Calgary, NRC, pp. 151-162.

Lewkowicz, A.G., Young, K.L. 1990. Hydrological processes in a small catchment containing a perennial snowbank, Melville Island, N.W.T. In: Prowse, T., Ommanney, C.S.L (eds.), Northern Hydrology Selected Perspectives. NHRI Symposium No. 6. pp. 237-251

Marsh, P., Woo, M.K. 1981. Snowmelt, glacier melt and high arctic stream flow regimes. Canadian J. of Earth Sci. 18, 1380-1384.

Metcalfe R.A., Buttle J.M. 1999. Semi-distributed water balance dynamics in a small boreal forest basin. J. Hydrol. 226, 66-87.

McNamara, J.P., Kane, D.L., Hinzman, L.D., 1998. An analysis of streamflow hydrology in the Kuparuk River Basin, Arctic Alaska: a nested watershed approach. J. of Hydrol. 206, 39-57.

Pomeroy, J.W., Granger, R.J.(eds.) 1999. Wolf Creek Research Basin: Hydrology, Ecology, Environment. Environment Canada, Saskatoon, p.160.

Quinton, W.L., Marsh, P. 1998. The influence of mineral earth hummocks on subsurface drainage in the continuous permafrost zone. Permafrost and Periglacial Processes 9, 213-228.

Quinton, W.L., Marsh, P. 1999. A conceptual framework for runoff generation in a permafrost environment. Hydrol. Proc. 13, 2563-2581.

Roulet, N.T., Woo, M.K. 1986. Hydrology of a wetland in the continuous permafrost region. J. Hydrol. 89, 73-91.

Santeford, H.S. 1979*a*. Snow soil interactions in interior Alaska. In Modeling of Snow Cover Runoff, Colbeck, S.C. and Ray, M. (Ed.). Cold Regions Research and Engineering Laboratory, Hanover, NH, pp. 311-318.

Santeford, H.S. 1979*b*. Toward hydrologic modeling of the black spruce/permafrost ecosystem of interior Alaska. In Proceedings 30th Alaska Science Conference, Fairbanks, Alaska, 9 pp.

Slaughter, C.W. and Kane, D.L. 1979. Hydrologic role of shallow organic soils in cold climates. In Proceedings, Canadian Hydrology Symposium 79 - Cold Climate Hydrology, National Research Council of Canada, Ottawa, pp. 380-389.

Slaughter, C.W., Hilgert, J.W., Culp, E.H., 1983. Summer streamflow and sediment yield from discontinuous-permafrost headwater catchments. In: Proceedings, Fourth International Conference on Permafrost, National Academy Press, Washington, D.C., pp. 1172-1177.

Smith, M.W., Burn, C.R., 1987. Outward flux of vapor from frozen soils at Mayo, Yukon, Canada: results and interpretation. Cold Reg. Sci. Tech. 13, 143-152.

Tallaksen, L.M. 1995. A review of baseflow recession analysis. J. Hydrol. 165, 349-370.

Woo, M.K., Steer, P., 1982. Occurrence of surface flow on arctic slopes, southwestern Cornwallis Island. Can. J. Earth Sci. 19, 2368-2377.

Table 5.1. Characteristics of experimental slopes, Wolf Creek basin, Yukon.

	N-slope	S-slope	E-slope	W-slope
Elevation (m)	1175	1175	1250	1250
Slope Angle	0.18	0.6	0.10	0.17
Vegetation	Open Black/White spruce Willow shrub Labrador Tea	Closed aspen forest	Sparse Black/White spruce Willow shrub Labrador Tea	Willow shrub Labrador Tea
Frost	permafrost (0.6-2 m)	seasonal frost (1.6 m)	seasonal frost (1.2 m)	permafrost (0.45-0.65 m)

Table 5.2. Soil properties. K_s is saturated hydraulic conductivity and θ_{sr} is specific retention.

	North Slope			South Slope
	Upper Organic (0-0.1 m)	Lower Organic (0.1-0.3 m)	Mineral Soil (>0.3 m)	Mineral Soil (silt)
Bulk Density (kg/m^3)	55 ± 20	90 ± 20	1340 ± 180	1420 ± 70
Porosity (%)	92 ± 4	84 ± 10	52 ± 7	55 ± 4
K_s (m/s)	$7 \pm 4 \times 10^{-3}$	$2.5 \pm 2 \times 10^{-4}$	$5 \pm 7 \times 10^{-9}$	$6 \pm 1 \times 10^{-6}$
θ_{sr}	0.44 ± .09	0.49 ± .08	0.42 ± .05	0.31 ± .05

	East Slope			West Slope		
	Upper Organic (0-0.1 m)	Lower Organic (0.1-0.25 m)	Mineral Soil (>0.25 m)	Upper Organic (0-0.2 m)	Lower Organic (0.2-0.4 m)	Mineral Soil (>0.6 m)
Bulk Density (kg/m^3)	74 ± 30	130 ± 50	1644 ± 120	48 ± 20	69 ± 20	1565 ± 100
Porosity (%)	86 ± 5	77 ± 12	4 5± 5	89 ± 4	82 ± 6	47 ± 4
K_s (m/s)	$6 \pm 7 \times 10^{-4}$	$1 \pm 1 \times 10^{-4}$	$1 \pm 3 \times 10^{-7}$	na	$1 \pm 2 \times 10^{-3}$	$2 \pm 4 \times 10^{-7}$
θ_{sr}	0.48 ± .06	0.48 ± .07	0.1 ± .05	0.37 ± 09	0.41 ± .10	0.10 ± .04

Table 5.3. Stormflow hydrograph characteristics for selected rainfall events, 1997.

North Slope

Storm Date, 1997	P_t *(mm)*	P_d *(hr)*	P_{int} *(mm/hr)*	P_{5d} *(mm)*	P_{abst} *(mm)*	T_{resp} *(hr)*
June 4	25.6	42	0.61	4.8	0.9	6
June 17	13.9	15.5	0.90	1.8	3.2	3
June 28	8.8	19	0.46	3.9	1.5	5
July 3	22.5	73	0.31	11.2	2.2	5
July 7	9	30	0.30	22.8	2.6	4
July 22	11	11	1.00	0.2	0.8	2.5
July 26	6.9	4	1.73	11.6	0.9	1.5
July 31	3.8	10	0.38	8.5	1.1	6
September 5	5	5.5	0.9	2.6	4.1	3.9

Storm Date, 1997	T_{rise} *(hr)*	T_{fall} *(hr)*	T_{ef} *(hr)*	Q_{pk} *(mL/s)*	Q_{ant} *(mL/s)*	*RO (mm)*	RO_e *(mm)*	*RF*	t'_1 *(hr)*	t'_2 *(hr)*
June 4 to 6	25	31	159	260	20	8.2	5.7	0.41/0.6	21	127
June 17/18	24	29.5	120	154	30	5.5	3.6	0.43/0.7	40	162
June 28/29	27.5	32.5		68	41	1.7		0.22	35	
July 3	24	66		137	90	2.7		0.14	75	
July 7	13	33.5		147	115	1.2		0.16	32	
July 22	13	36		150	41	4.8		0.33	25	
July 26	12	42		88	42	2.3		0.16	40	
July 31	12	20		80	67	0.3		0.10	44	
September 5	17.5	35		124	52	0.9		0.18	32	

West Slope

Storm Date, 1997	P_t *(mm)*	P_d *(hr)*	P_{int} *(mm/hr)*	P_{5d} *(mm)*	P_{abst} *(mm)*	T_{resp} *(hr)*
June 17	17.3	15.5	1.1	0	9.8	5
June 28	6.4	5.5	1.6	2.2	4.0	8
July 7	6.5	21.5	0.3	19.1	4.6	12
July 22	14	10	1.4	0	7.1	6.5
July 26	8.6	4	2.1	15.3	5.5	2
September 5	5.5	5	1.1	2.9	5.1	4

West Slope cont.

Storm Date, 1997	T_{rise} (hr)	T_{fall} (hr)	T_{ef} (hr)	Q_{pk} (mL/s)	Q_{ant} (mL/s)	RO (mm)	RO_e (mm)	RF	t'_1 (hr)	t'_2 (hr)
June 17/18	27	52		22	4	3.8		0.22	30	
June 28/29	12	40		12	5	0.9		0.14	45	
July 7	32	50		11	3	0.8		0.12	65	
July 22	25	32		8	2	2.2		0.16	23	
July 26	7.5	55		18	4	0.8		0.09	25	
September 5	6.5	29		11	0	0.4		0.07	17	

East Slope

Storm Date, 1997	P_t (mm)	P_d (hr)	P_{int} (mm/hr)	P_{5d} (mm)	P_{abst} (mm)	T_{resp} (hr)
June 17	17.3	15.5	1.1	0	8.1	1
June 28	6.4	5.5	1.6	2.2	1.2	1.5
July 7	6.5	21.5	0.3	19.1	1.5	3.5
July 22	14	10	1.4	0	0.5	1.5
July 26	8.6	4	2.1	15.3	0.7	1
September 5	5.5	5	1.1	2.9	1.6	0.5

Storm Date, 1997	T_{rise} (hr)	T_{fall} (hr)	T_{ef} (hr)	Q_{pk} (mL/s)	Q_{ant} (mL/s)	RO (mm)	RO_e (mm)	RF	t'_1 (hr)	t'_2 (hr)
June 17/18	17	44	114	480	170				56	442
June 28/29	7	28	85	400	245				78	636
July 7	24	34	77	605	320				77	389
July 22	11.5	16	45	380	265				67	362
July 26	5	26	58	405	260				86	405
September 5	6	14	34	335	220				55	203

P_t – total rainfall
P_d – rainfall duration
P_{int} – rainfall intensity
P_{5d} – 5-day antecedent rainfall
P_{abst} – initial precipitation abstraction
T_{resp} – time from start of precipitation to rise in hydrograph
T_{rise} – duration of hydrograph rise
T_{fall} – duration of hydrograph fall
T_{ef} – extended recession period
Q_{pk} – peak runoff
Q_{ant} – antecedent runoff
RO – stormflow runoff
RO_e – extended slope runoff
RF – response factor
$t'_{(1,2)}$ - recession coefficients for two storage reservoirs

Table 5.4. Streamflow contributing areas (l_1) and extended contributing areas (l_2) for selected rainfall events, 1997.

Storm Date	North Slope		West Slope		East Slope	
	l_1 *(m)*	l_2 *(m)*	l_1 *(m)*	l_2 *(m)*	l_1 *(m)*	l_2 *(m)*
June 4	144	871				
June 17	278	1110	112		442	1940
June 28	240		168		636	2792
July 3	514					
July 7	219		243		389	1707
July 22	171		85		362	1589
July 26	274		93		405	1778
July 31	301					
September 5	219		63		203	891

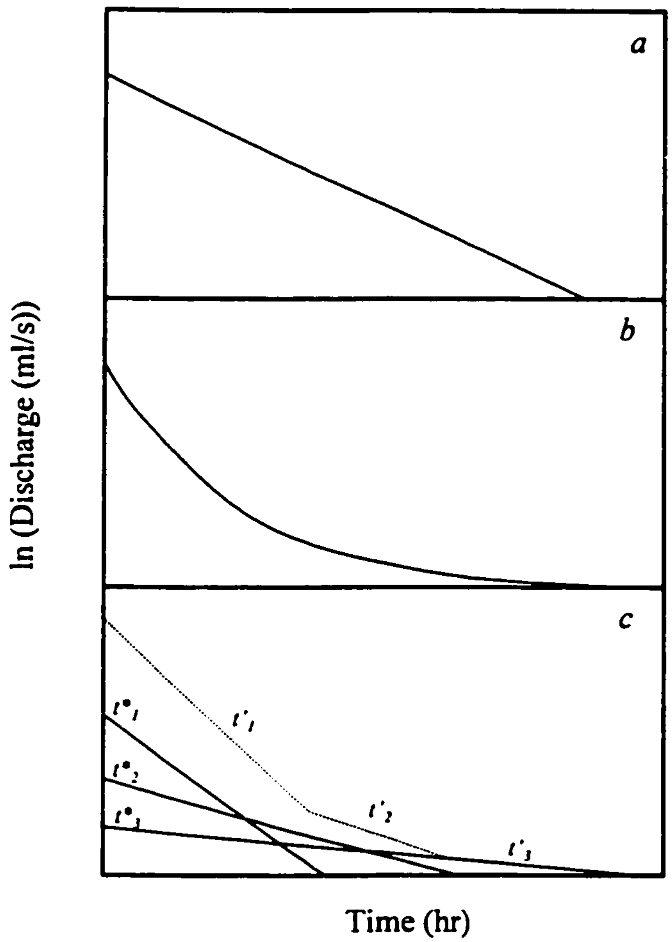

Figure 5.1 Theoretical recessions from a sloping slab considering:
(*a*) discharge from one linear storage-discharge reservoir,
(*b*) discharge from a gradually decreasing contributing area
(*c*) multiple linear storage-discharge reservoirs. Dotted line is observed recession produced by the superposition of the recession flows from these reservoirs.

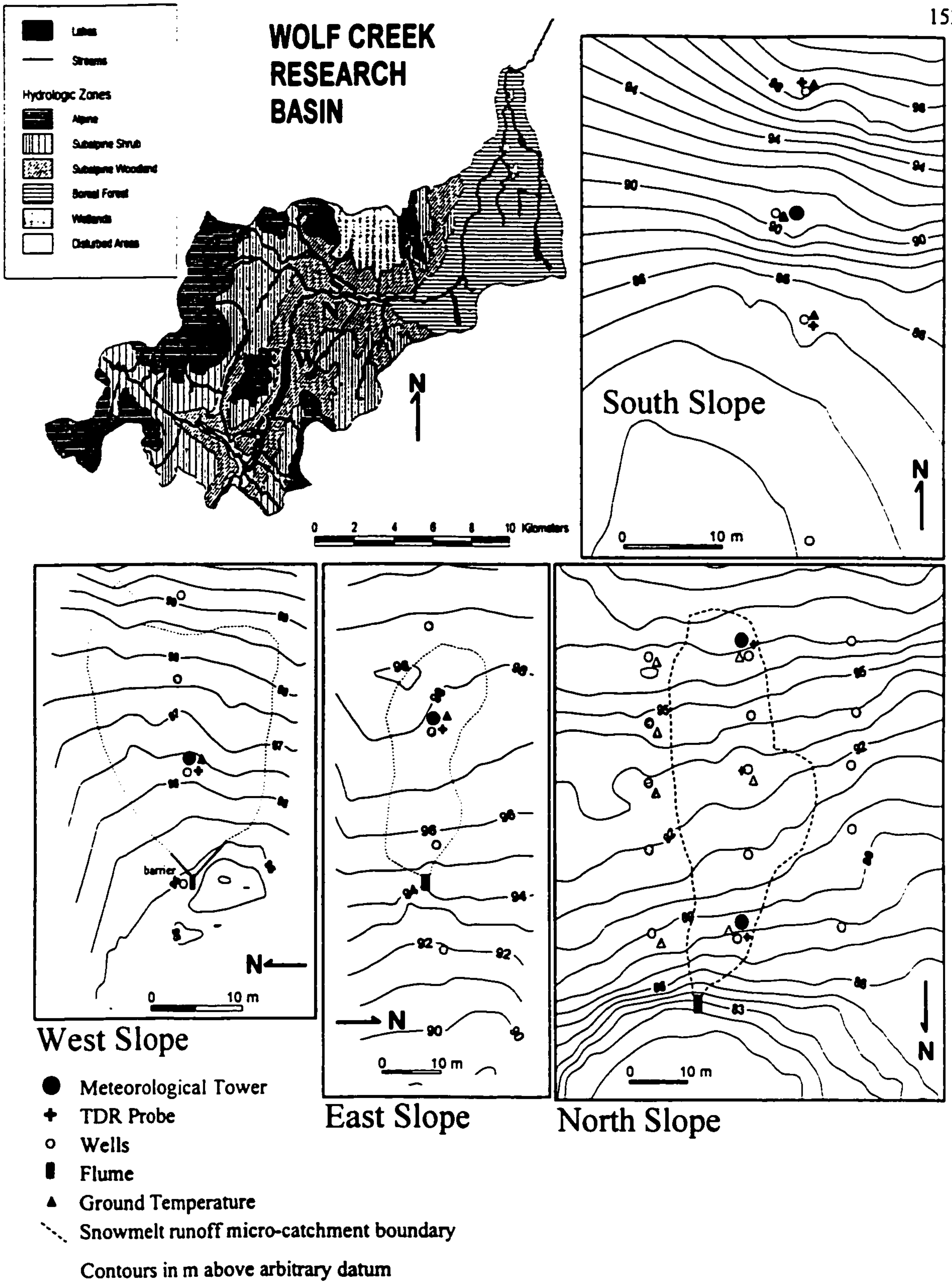

Figure 5.2 Hydrologic zones and location of experimental slopes within the Wolf Creek Basin, topography of experimental slopes and deployment of instrumentation.

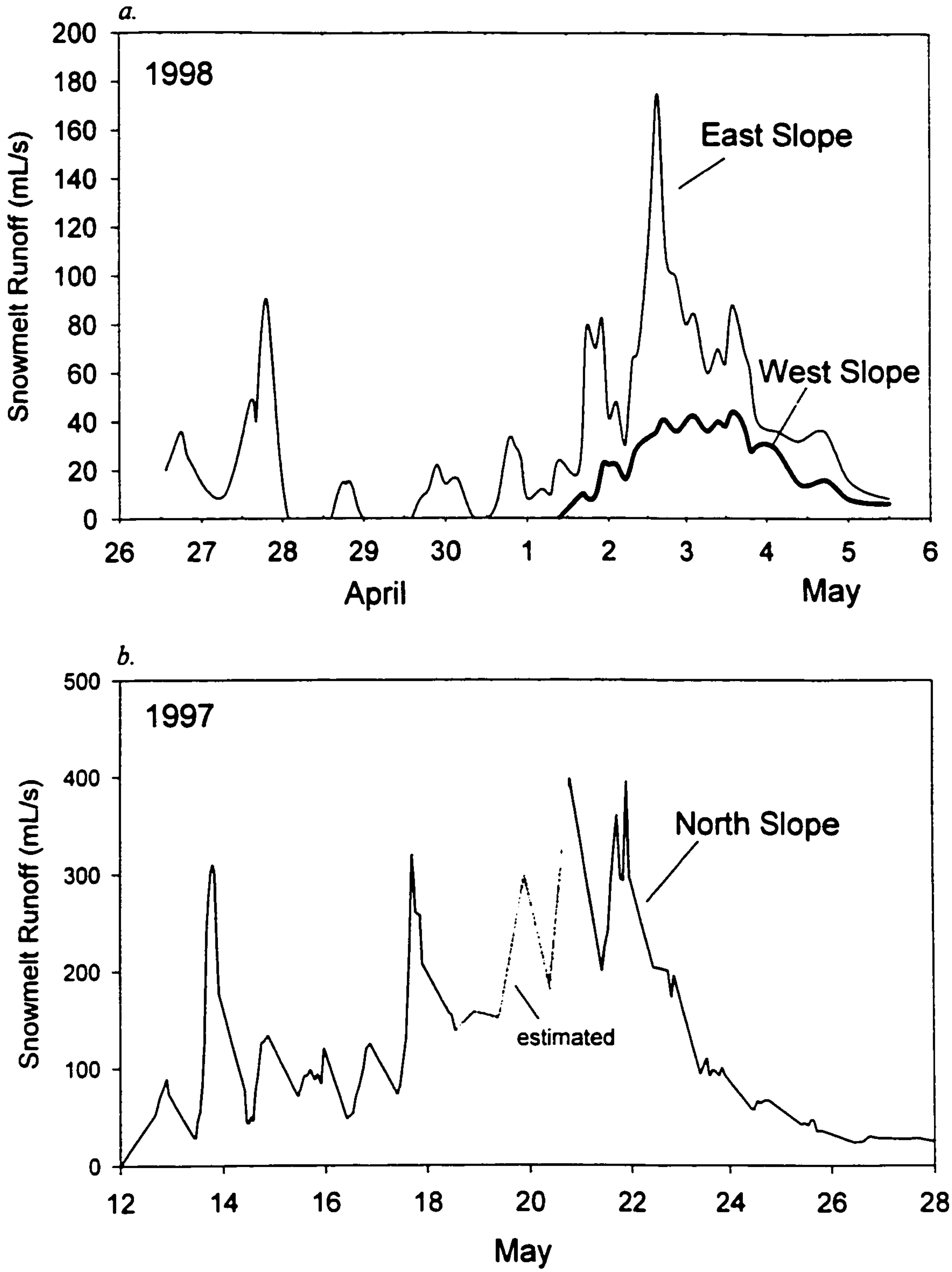

Figure 5.3 Runoff during the (*a*) 1998 snowmelt period for the East and West facing slopes, and (*b*) the snowmelt period in 1997 for the North slope.

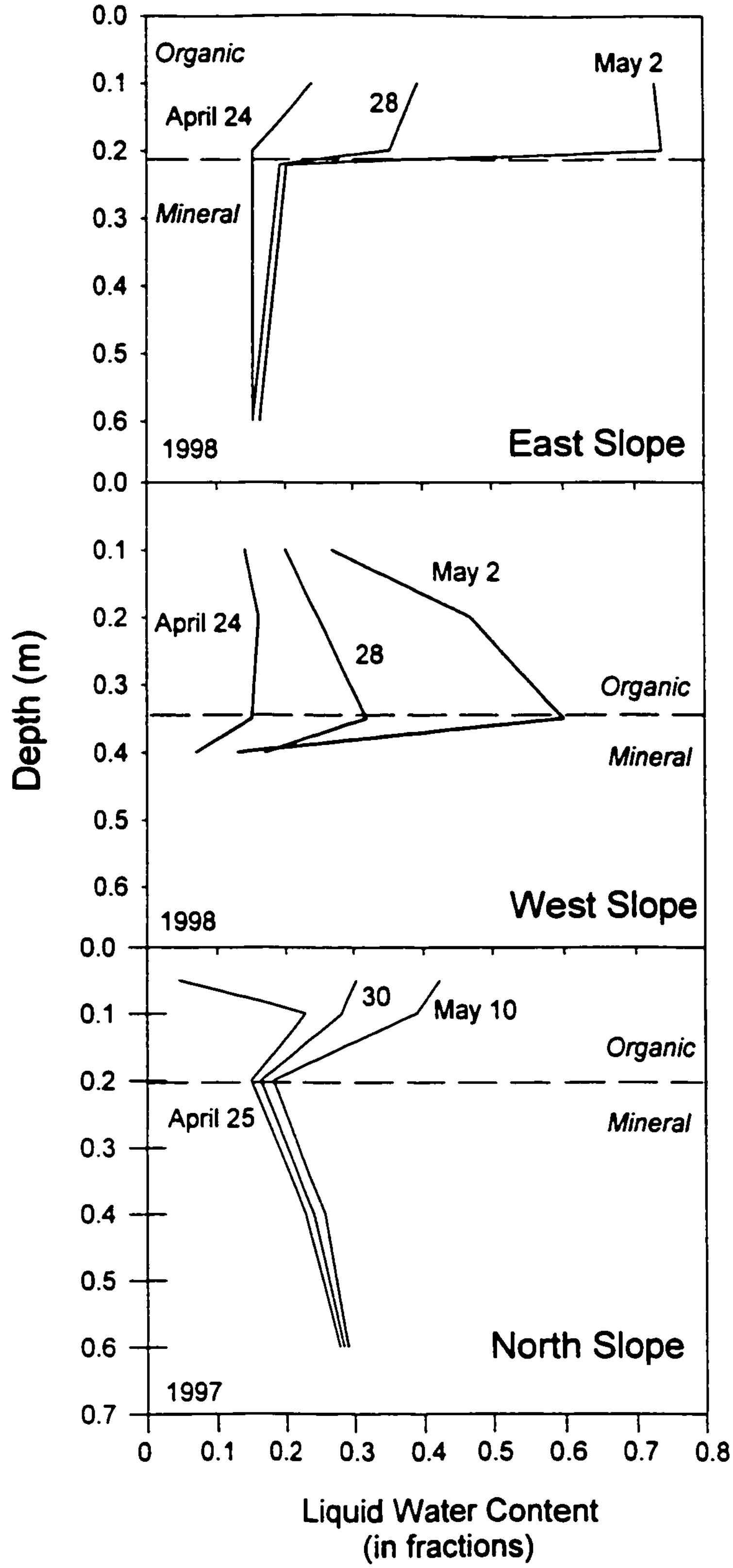

Figure 5.4. Changes in liquid water content for the E, W and N-slope during the melt period.

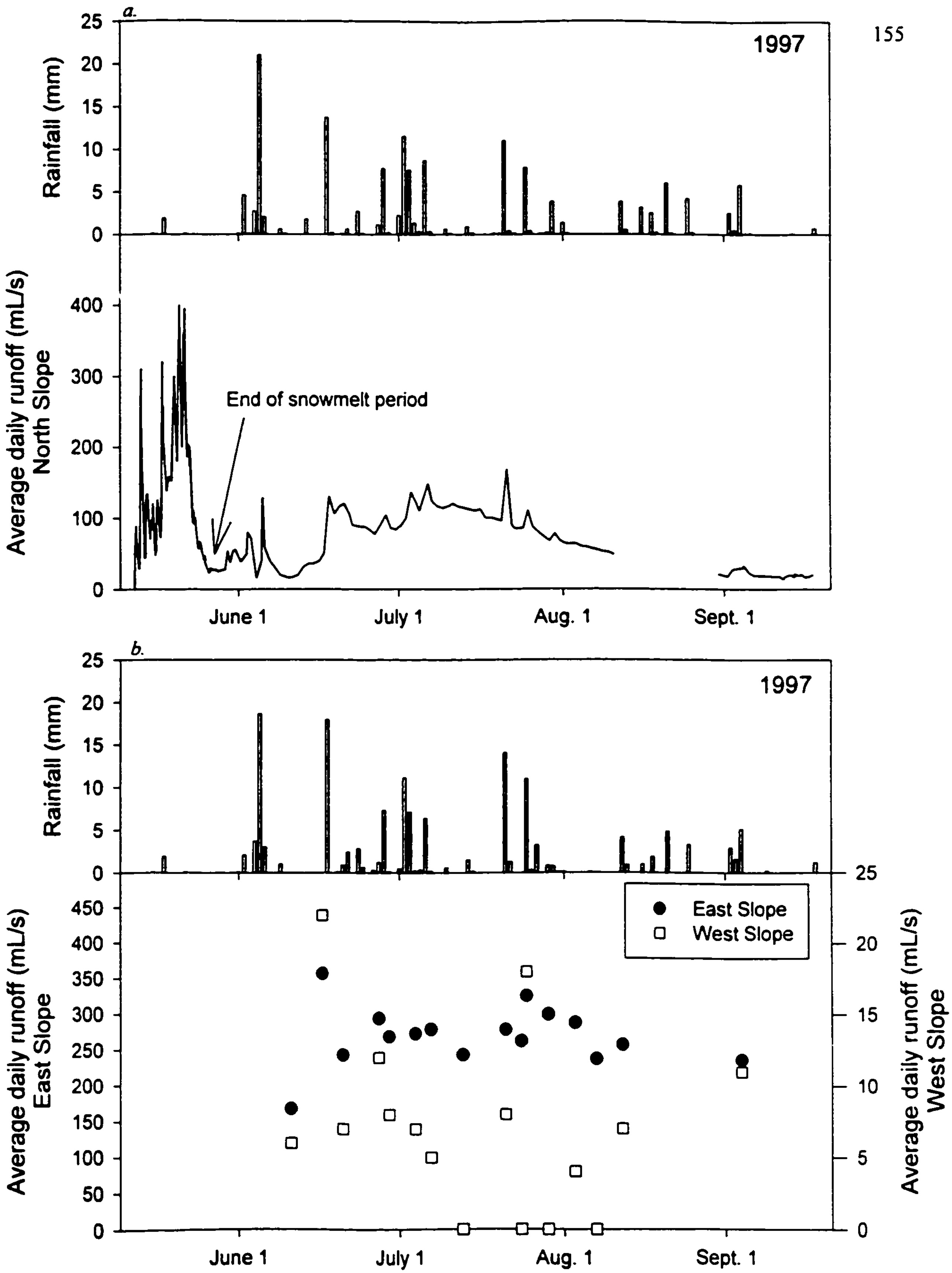

Figure 5.5. Average daily rainfall and runoff for the (*a*) North and (*b*) East and West slope.

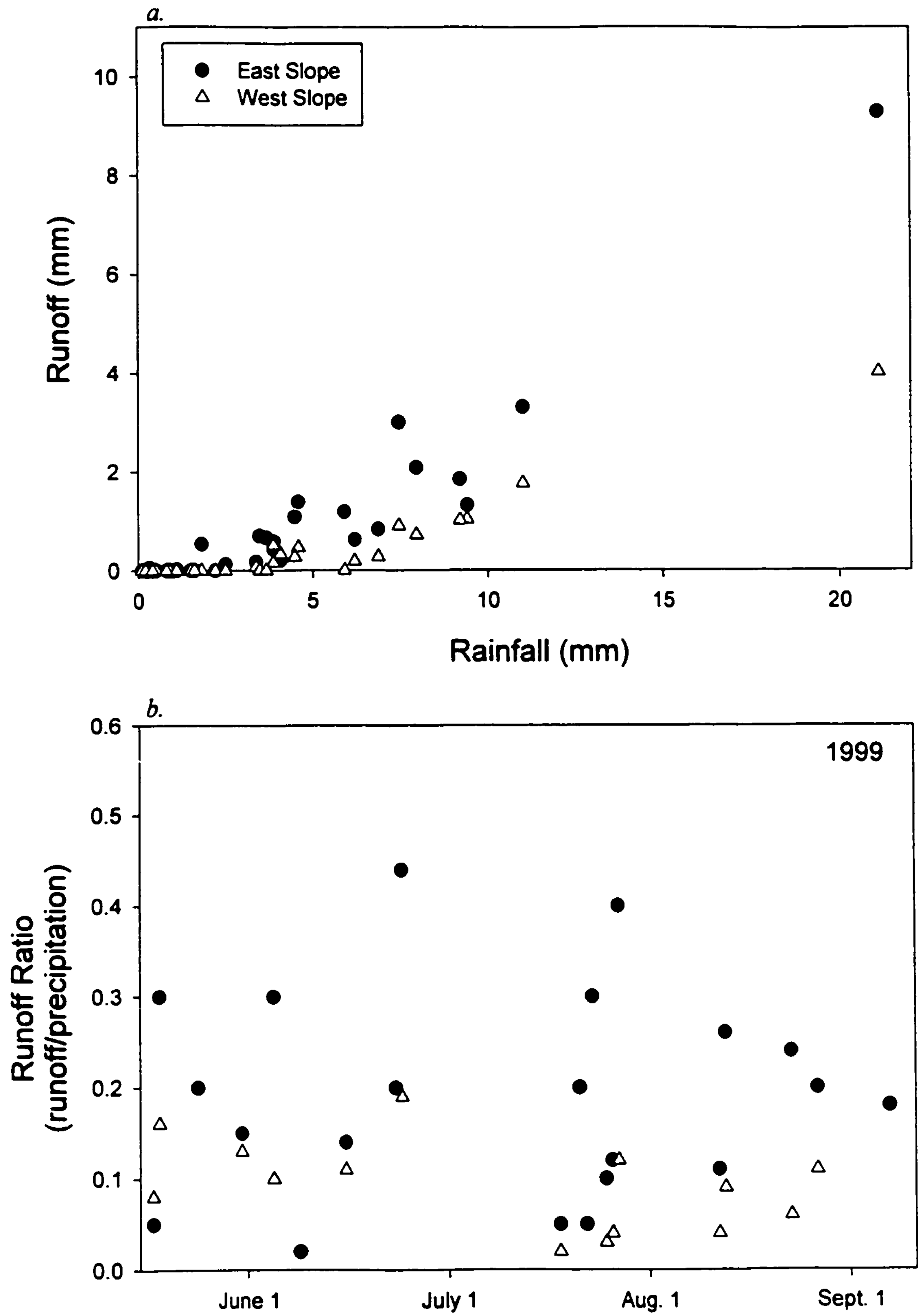

Figure 5.6. (*a*) Volume of runoff for 1999 summer rain events.
(*b*) Ratio of runoff to precipitation.

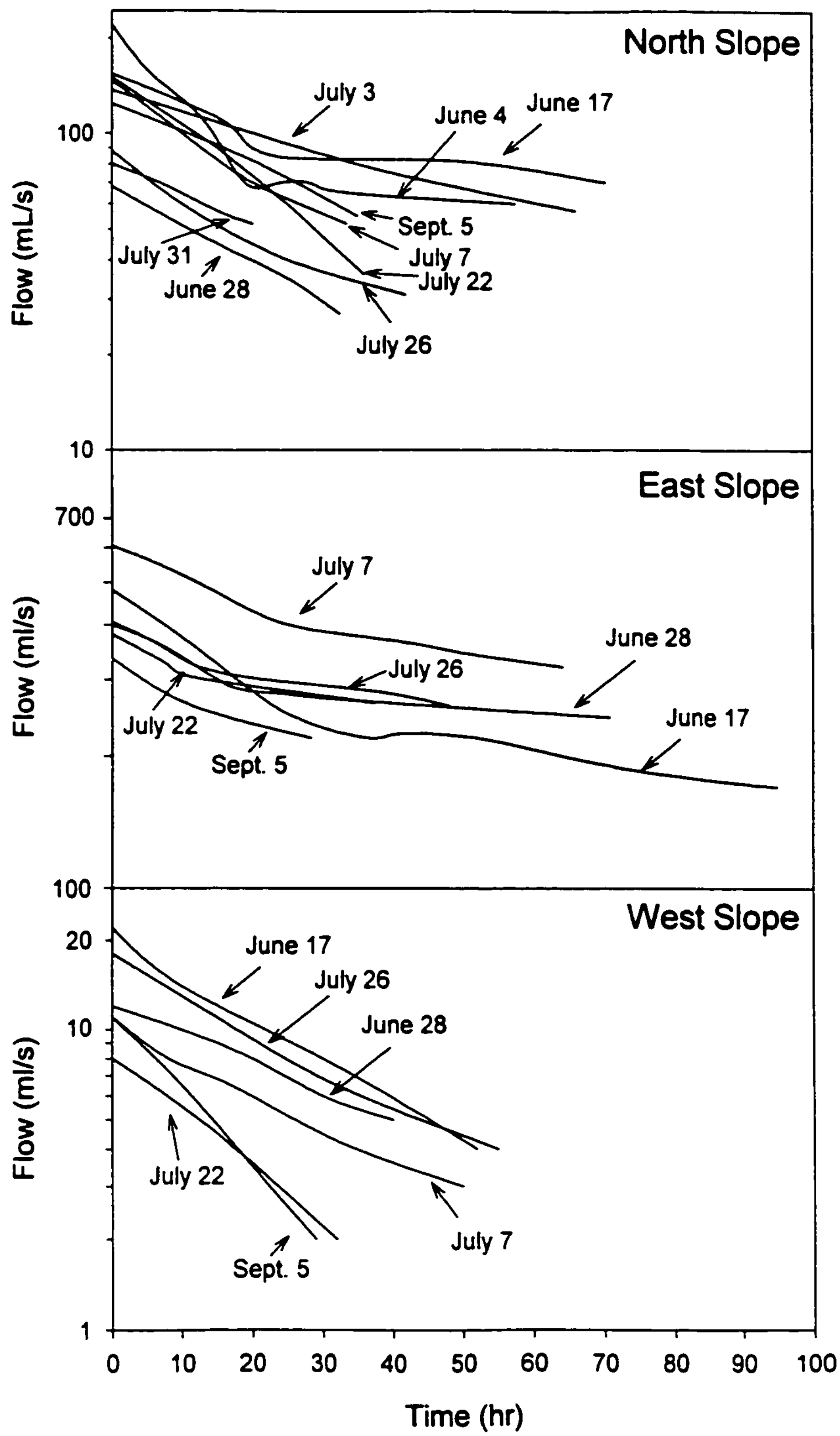

Figure 5.7. Measured discharge vs. time for summer recessions, 1997.

CHAPTER SIX

SUMMARY AND CONCLUSIONS

6.0 RESEARCH PROBLEMS

The paucity of research in subarctic catchments has resulted in an incomplete understanding of hydrological processes and interactions in northern environments. There remains a need for the scientific study of hydrology so that meaningful relations can be obtained to help understand and predict hydrological responses to external forcings, such as changes in the environment. Subarctic terrain ranges from flat to rugged and from sea level to high elevations. In subalpine catchments, a mosaic of landscape types with various aspects and elevation strongly influences the microclimate of hillslopes that in turn influences vegetation, the presence or absence of permafrost, and the magnitude and timing of hydrological processes. Hillslopes in basins with steep topography offer the ideal place to study process-interactions since basin hydrology is an integration of processes operating on slopes. Previous subarctic hydrological investigations have either focussed upon processes studies at the plot scale (eg. Price and Dunne, 1976; Santeford, 1979*a,b*; Slaughter and Kane, 1979; Eaton and Wendler, 1981; Kane *et al.*, 1981; Kane and Stein, 1983) or analyses of streamflow hydrographs which integrate most processes operating within the basins (eg. Dingman, 1966; 1971; Slaughter *et al.*, 1983; Chacho, 1990). There has been little effort to bridge the gap between these two approaches. In the latter approach, streamflow characteristics are used to infer runoff processes within the basin and areas that

6.2 SIGNIFICANT FINDINGS

1) Hillslope water balances were quantified, revealing a large range in the timing and magnitude of hydrological processes during both the melt and summer seasons.

2) Ice rich soil layers prevent vertical water movement and enhance runoff. During melt, the presence of an ice-rich zone at the organic-mineral interface inhibits meltwater percolation, encouraging the development of a perched saturated zone. In summer, ice-rich permafrost and seasonal frost block soil interstices and impedes deep drainage, indirectly influencing the position of the water table. In contrast, water can infiltrate and percolate frozen organic and mineral soils that are porous or have low ice content.

3) Due to ice rich layers, runoff generation during the melt period is confined to hillslopes with a capping organic layer. Once the storage capacity of the desiccated organic layer is exceeded, all remaining water within the snowpack is directed laterally as runoff.

4) The presence of surface organic layers and antecedent wetness dictate the magnitude of summer runoff. Given the large differences in saturated hydraulic conductivity between organic and mineral soils, the closer the water table is to the ground surface, the greater is the magnitude of runoff. On well drained slopes without covering organic layers, vertical hydrological exchanges (ie. precipitation and evapotranspiration) prevail.

5) Soil wetness is influenced by local topography and the position of the hillslope in relation to its surrounding landscape. Steeper slopes encourage drainage while runoff is enhanced on slopes that receive inflow from upslope source areas to sustain a saturated zone in the porous organic soil.

6) The strong contrast in saturated hydraulic conductivity between organic soils and underlying

mineral soils gives rise to a two-layer flow system consisting of quickflow and slowflow. Quickflow is defined as rapid runoff, delivered downslope by matrix flow through highly porous organic material and by preferential flow in interconnected surface depressions, soil pipes and rills. Quickflow regime ranges from laminar flow in porous organic soils to turbulent flow along preferential pathways. Slowflow is defined as laminar flow in the saturated matrices of soils with low hydraulic conductivities, giving rise to flow velocities orders of magnitudes lower than quickflow.

7) Runoff and stormflow are largely controlled by flow within the organic layer. Runoff hydrographs respond quickly to rainstorms and exhibit recessions that last much longer than those observed in the temperate region.

8) The variable source area concept for runoff generation applies to subarctic, subalpine hillslopes.

9) Contributing areas, as deduced by recession analysis, show spatial and temporal variations due to: 1. changing upslope source areas for a given reservoir, and 2. multiple reservoirs each with separate flow contribution zones.

10) Contributing areas and hydrological connections increase with hillslope wetness. When a particular site is not connected to an upslope area, the zone contributing to streamflow declines as the hillslope becomes dryer and the water table passes into the mineral layer. Where the site is connected to an upslope source area, the additional water will contribute to storm runoff. However, such hydrological linkages may become severed as hillslope wetness declines.

11) Evapotranspiration exceeds summer rainfall for all hillslopes, drawing water from the soil moisture reservoir which is replenished during the snowmelt season.

6.3 FUTURE RESEARCH

The major contribution of this research is an improvement in the understanding of hydrological behaviour of subarctic, subalpine hillslopes. There remains, however, a need to directly relate hillslope runoff to streamflow. Geochemical and isotopic signatures of runoff from the hillslopes can shed light on the pathways and residence times of water, thus helping to link slope runoff with streamflow. Chemical analysis in concert with hydrograph analysis may validate the variable source area concept by pinpointing separate reservoirs for streamflow. The processes associated with hydrological linkages between terrain units is poorly understood. In subalpine areas, the physical setting of a hillslope unit is strongly influenced by the surrounding terrain. The mechanisms of water transfer from upland to lowland areas has to be explored.

Water balances of subarctic slopes must be compared with those of the basin as a whole to determine the relative role of various slopes on the catchment water balance. Water balance studies must remain a central focus of research due to their importance in water resources and in highlighting interactions among hydrologic components. Due to boundary layer conditions (insufficient fetch and sloping terrain), the microclimatology of hillslopes within this environment requires more investigation. The atmospheric fluxes of water from this environment should be studied in more detailed to refine and corroborate findings from this study.

Finally, a major thrust in contemporary hydrology is concerned with issues of scaling and representativeness as hydrologists attempt to impart a regional significance to studies that are primarily local in nature. It is important that information and principles from this thesis be tested and applied to other subarctic catchments to confirm the generality of findings.

6.4 References

Chacho Jr. EF. 1990. 'Water and suspended solids discharge during snowmelt in a discontinuous permafrost basin', in: *Proceedings of the 5th Canadian Permafrost Conference*, National Research Council, Ottawa, pp. 167-173.

Dingman SL. 1966. 'Characteristics of summer runoff from a small watershed in central Alaska', *Water Resources Research* **2**: 751-754.

Dingman SL. 1971. *'Hydrology of Glenn Creek watershed, Tanana basin, Central Alaska'*. U.S. Army Cold Region Research Engineering Laboratory Research Report 297, Hanover, New Hampshire, p. 112.

Eaton F, Wendler G. 1982. 'The heat balance during the snow melt season for a permafrost watershed in Interior Alaska', *Archives for Meteorology, Geophysics, and Bioclimatology, Series A* **31**: 19-33.

Gibson JJ, Edwards TWD, Prowse TD. 1993. 'Runoff generation in a high boreal wetland in northern Canada', *Nordic Hydrology* **24**: 213-224.

Kane DL, Stein J. 1983. 'Water movement into seasonally frozen soils', *Water Resources Research* **19**: 1547-1557

Kane DL, Bredthauer SR, Stein J. 1981. 'Subarctic snowmelt runoff generation', in *Proceedings of the Specialty Conference on The Northern Community*, Vinson TS (ed.), ASCE, Seattle, Washington, pp. 591-601.

Price AJ, Dunne TD. 1976. 'Energy balance computations on snowmelt in a subarctic area', *Water Resources Research* **12**: 686-694.

Santeford HS. 1979*a*. 'Snow soil interactions in interior Alaska', in: *Proceedings, Modeling of Snow Cover Runoff*, Colbeck SC, Ray M. (eds.), Cold Regions Research and Engineering Laboratory, Hanover, NH, pp. 311-318.

Santeford HS. 1979*b*. 'Toward hydrologic modeling of the black spruce/permafrost ecosystem of interior Alaska', in *Proceedings 30th Alaska Science Conference*, Fairbanks, Alaska, 9 pp.

Slaughter CW, Kane DL. 1979. 'Hydrologic role of shallow organic soils in cold climates', in *Proceedings, Canadian Hydrology Symposium 79 - Cold Climate Hydrology*, National Research Council of Canada, Ottawa, pp. 380-389.

Slaughter CW, Hilgert JW, Culp EH. 1983. 'Summer streamflow and sediment yield from discontinuous-permafrost headwater catchments', in *Proceedings, Fourth International Conference on Permafrost*, National Academy Press, Washington, D.C., pp. 1172-1177.

APPENDIX A

THE ROLE OF SOIL PIPES AS A SLOPE RUNOFF MECHANISM, SUBARCTIC YUKON, CANADA

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ABSTRACT

Pipeflow in subarctic slopes provides a preferential runoff mechanism that bypasses the soil matrix, rapidly conveying water to the stream. Extensive soil piping occurs in the central Wolf Creek basin, Yukon, at the interface of the organic and mineral horizons. Flow in these pipes is ephemeral, transmitting water only when the water table is within or above the zone where pipes occur. During snowmelt, pipeflow began several days after the onset of surface runoff. Pipeflow contribution increased through the melt season until ground thaw lowered the water tables, leaving matrix flow within the organic layer as the dominant mode of runoff. Pipeflow accounted for 21 % of runoff during the 15 day melt period of 1997. During melt, pipeflow closely followed the daily snowmelt cycles and responded earlier than the integrated slope runoff. Following melt, pipeflow recurred only during two intense summer rainstorms, each yielding different pipeflow response characteristics. In the summer, pipeflow hydrographs rose and fell much quicker than direct storm runoff from the entire slope which was dominated by fast matrix flow within the organic layer. Pipeflow contributing areas were relatively small, but their extent changed with hillslope wetness. Analysis revealed that the Manning flow equation can be combined with contributing areas to simulate pipeflow discharges. Unlike temperate environments where frozen ground is not a major factor, the frost table significantly controls the position of the phreatic surface, and must be considered in any models of pipeflow in permafrost slopes.

A1.0 INTRODUCTION

Runoff from hillslopes underlain by permafrost reaches the stream via different pathways that affect both the timing and magnitude of responses to snowmelt and rainfall. The presence of an organic layer overlying a less permeable mineral substrate with permafrost limits the occurrence of overland flow, which is more common where frozen mineral soils are exposed on the slopes (Woo and Steer, 1982). On many permafrost slopes, saturation overland flow occurs when the entire organic zone becomes saturated (Hinzman *et al.*, 1993). Within the soil matrix, both Darcian and macropore flows may be generated. Large vertical reduction in hydraulic conductivity due to increasing humification of the peat (Quinton, 1997; Quinton and Marsh, 1998) and the discontinuity between the peat and mineral soil (Carey and Woo, 1998) causes the Darcian flow to vary by several orders of magnitude between the mineral and organic horizons. Macropores within the peat and the mineral soil allow water to bypass the soil matrix and accelerate runoff rates.

In addition to vertical variability, flows are also concentrated in linear features on or in permafrost slopes. Inter-hummock zones (Quinton and Marsh, 1998), water tracks (McNamara *et al.*, 1998), rills (Carey and Woo, 1998) and soil pipes (Gibson *et al.*, 1993) all represent modes of enhanced water delivery. These linear features capture local zones of runoff, hence modifying the slope area that contributes to stormflow. An understanding of the role played by these preferential flow paths is critical to hillslope and runoff modelling in permafrost terrain.

Much research on the hydrology of pipes has been conducted in the humid environments of the British Isles (Weyman, 1970; Gilman and Newson, 1980; Jones and Crane, 1984; Wilson and Smart, 1984, Jones, 1971; 1987; 1988; 1997). Jones (1987; 1997) for example, discussed the ability of pipes to modify the contributing areas during storms and their role within the overall framework of runoff responses. Although pipes have been observed in permafrost environments, particularly in the subarctic (Gibson *et al.*, 1993, Quinton, 1997, Carey and Woo, 1998), no study has specifically investigated their hydrologic importance to hillslope runoff. This paper aims to study the occurrence, mechanisms, timing and magnitude of pipeflow on a subarctic hillslope. By comparing pipeflow characteristics with other modes of runoff, the role of pipes will be placed within the broader context of permafrost slope runoff.

A2.0 STUDY SITE AND METHODS

A north facing slope segment in the subarctic open forest (black and white spruce) of the Wolf Creek basin (61°31'40" N, 135°31'14" W, elevation 1175 m) 15 km south of Whitehorse, Yukon, was chosen for this study as part of the Canadian GEWEX programme (Figure A1). All but the footslope is underlain by permafrost at depths between 0.6 and 2 m. The climate is subarctic continental with mean January and July temperatures of –21°C and +15°C at the Whitehorse airport. Mean annual precipitation at the Whitehorse airport is approximately 260 mm, half of which falls as rain. Carey and Woo (1998; 1999) presented the snowmelt hydrology and the annual water balance for this slope.

Soils consist of an organic layer, 0.2±0.09 m thick, of peat and living plants (lichens, mosses, sedges and grass) overlying a stony clay of glacial origin. There are marked differences in the physical properties of the organic and the mineral zones, and significant variation within the organic layer (Figure A2). Bulk density, determined using gravimetric methods, increases downward in the organic layer due to humification to reach a maximum of 400 kg/m^3 at the base. Saturated hydraulic conductivity at each soil layer was obtained using pumping tests conducted in wells and piezometers (Freeze and Cherry, 1979). Well recovery plots indicated that flow within the saturated organic layers was Darcian, although there is some possibility of non-Darcian flow in the porous organic mat within the top few centimeters of the surface. Significant differences between the lower and upper organic zones are due to varying degrees of humification, and from the lower organic zone to the mineral substrate, the hydraulic conductivity drops by seven orders of magnitude (Figure A2).

Hydrological processes were monitored between April 6 and September 22, 1997. Pre-melt snow water equivalent was obtained using a Mount Rose snow sampler, supplemented by 12 snow pits. Snow ablation was monitored daily at several sites using the field method presented in Carey and Woo (1998).

Soil moisture was measured using TDR probes (MoisturePoint) buried at 0.05, 0.1, 0.2, 0.4 and 0.6 m depths at three sites where water table depths were also recorded continuously using electronic water-level recorders. Water tables were measured daily at 12 additional wells. The position of the frost table was determined from ground temperatures monitored at seven points on the slope at various depths down to a maximum of 1.5 m.

Slope runoff from localized drainage areas was measured using nine flumes. These flumes were boxes approximately 0.6 m long, 0.4 m in width with a 90° V notch. Water was directed into the flumes using plastic sheeting where during melt, a graduated cylinder was used to measure discharge once every three hours. During the summer, hillslope runoff was determined by measuring discharge from a flume dug into the slope base (N-1). Water level recorder was recorded electronically every half-hour behind the outlet and discharge was determined from a calibration relating spot measurements of discharge to water level. The calibration curve allowed discharge to be determined to ± 5 mL/s. Pipeflow was monitored by sealing the outlets of three soil pipes (Figure A1) with plastic sheeting to direct the flow into a gutter where the volume was measured once every 3 hours during snowmelt. The volume of water collected in the gutter was converted to discharge assuming a constant discharge rate during the measurement period. During summer rainstorms, pipeflow was measured by directing a volume of water from the pipe outflow into a graduated cylinder which was then converted to a rate. The frequency of pipeflow measurements varied during storm events, but was no less than once every hour when pipeflow was occurring.

Two 6-m micrometeorological towers approximately 30 m apart were erected on the slope. Net radiation (REBS), incident and reflected short wave (Eppley), wind speed (Met-One), air temperature and humidity (Vaisala HMP35CF) were sampled at 1-minute intervals and recorded on Campbell Scientific CR-10 data loggers at 30-minute intervals. Evaporation was calculated using the Bowen ratio method from five wet and dry thermocouples spaced at approximately 1 m intervals above the ground surface (Carey and Woo, 1999).

A3.0 PIPEFLOW OBSERVATIONS

A3.1 Pipe occurrence and drainage areas

Approximately 24 pipes were observed on the study slope segment that measures 3800 m^2. All pipes are found at the base of the organic layer or in the cryoturbated zone at the interface of organic and mineral horizons. Dye tracing, excavation and topographic survey were conducted to map the extent of pipes and their flow contribution areas. There may have been some underestimation in the contributing areas due to dye absorption by organic materials, although in most cases, the dye was readily observed moving through the near-surface pathways on the hillslope. Pipes have a mean length of 2.4±1.7 m and diameter of 77±20 mm, with circular to elliptical cross sections that are commonly exposed only at their effluent ends.

Unlike complex pipes observed in more temperate, humid environments (e.g. McCaig, 1983; Wilson and Smart, 1984; Jones 1987), pipes exist as linear features, often occurring in sequence, oriented predominantly downslope (Figure A1). Pipes and rills occur together as near-surface preferential drainage routes during snowmelt, with water being exchanged between and among pipes and rills through the porous organic mat. The exact direction of water movement depends upon the position and shape of the phreatic surface which is modified by both microtopography and ground frost.

Two methods were used to assess the source areas of pipeflow. During the snowmelt period (May 12-May 27), flow pathways were traced by dye and visual observations (Figure A3), whereas during summer, contributing areas for pipes were determined using the maximum potential dynamic contributing area (DCA) method proposed by Dickinson and Whiteley (1970) as

$$DCA = RO_p \,/\, P \tag{A1}$$

where RO_p is the total storm pipeflow runoff and P is the event precipitation (Table A1). Due to their limited length, proximity to the surface, and a highly variable microtopography which restricts the drainage catchment, contributing areas for the pipes are small compared to those in the humid temperate environments (Jones, 1997). Pipes P-1, P-2 and P-3 have the largest source areas among all the pipes on the study slope. Due to the variability in pipe contributing areas during melt, only the maximum extent as observed on May 20 are given in Table A1. Source areas for snowmelt period (determined from flow tracing) were greater than those for two summer storms that produced pipeflow (estimated by Equation A1). Pipe source areas shrank after their maximum extension during the melt period due to declining slope wetness throughout the summer. After the June 17 event, pipeflow contributing areas were effectively zero as the water table subsided into the mineral zone to deprive the pipes of a water source during subsequent storms.

A3.2 Water movement within pipes

Flow within soil pipes occurs only under wet conditions. When the pipes were active, observed water levels never rose higher than half the diameter of the pipe, and hydrostatic pressures were negligible. Assuming that gravitational force controls pipe discharge, the volume of runoff is directly related to the water level within the pipe. Soil pipes are approximately circular, and the velocity of pipeflow, v, can be calculated using Manning's equation:

$$v = R^{2/3} S^{1/2} n^{-1} \tag{A2}$$

where R is the hydraulic radius, S is the slope and n is the roughness parameter. Observations of pipe dimensions (approximated as circular) and water-level within the pipes were used to determine the hydraulic radius and the cross sectional area of flow. Discharge was calculated as the product of velocity and the cross sectional area of flow. Values of n, determined from observations of mean pipeflow velocity are 0.10-0.17, which compare closely to 0.08-0.15 reported by McCaig (1983). Figure A4*a* relates discharge to depth of water in the soil pipes for n=0.135 with the mean pipe diameter, 77±22 mm. Theoretical and observed discharges compare well (Section 3.3), suggesting the effectiveness of Manning's method.

The efficiency of pipeflow as a runoff mechanism can be determined by combining calculated Manning discharges with pipe contributing areas. Dividing discharge by the pipe contributing area (Table A1), the maximum theoretical pipe runoff can be determined. Using the May 20 contributing areas for pipes P-1 and P-2, the observed pipe dimensions (Table A1) and a mean n=0.135, Figure A4*b* shows the maximum pipe runoff computed for various water levels within the conduit. To sustain pipe discharge, water must feed from the surrounding organic matrix into the pipe at sufficient rates. As an example, using Darcy's Law and a saturated hydraulic conductivity value of K=2.5x10^{-4} m s^{-1} for the lower organic zone in which pipes occur, discharge into P-1 assuming that the pipe is surrounded completely by saturated organic material is approximately 1.3 m^3 hr^{-1} or 360 mL/s. Dividing this discharge by the known contributing area yields a value of 4.8 mm hr^{-1}. These values compare closely with maximum Manning pipe discharge of 380 mL/s for a water depth of 35 mm in the pipe, illustrating that rapid Darcian flow in the organic layer can provide sufficient volumes of water to sustain pipeflow. As water levels within the pipe contributing area fall, matrix flow declines, and the discharge is reduced. In this way, pipe discharge is directly related to the wetness of its contributing area.

A3.3 Pipeflow timing and discharge

Pipeflow was observed only during the snowmelt period and two rainstorms. For clarity, observations will be divided into snowmelt and summer periods.

A3.3.1 Snowmelt period

Runoff began on May 12 when 78 % of the slope remained snow-covered (Figure A5). Initially, water flowed as a thin saturated layer at the snowpack base and was issued to topographic lows and rills in the snow-free zones. Soil pipes uncovered at this time were inactive (Figure A5*a*), an observation similar to those of Roberge and Plamondon (1987) who found no pipeflow at the beginning of the snowmelt period. After the onset of runoff, flow increased rapidly to peak at 12 mm/d on the second day of runoff. Surface flow in rills dominated runoff as shallow thaw hindered subsurface drainage.

As warming continued and lateral advection from bare ground reduced the snow cover to 35 % on May 14, subsurface hydrological connections were established. By then, a large portion of the lower slope was snow-free, with thaw depths reaching 0.1±0.05 m. Ground thaw allowed pipes at the slope base to begin transmitting water (Figure A5*b*), and along with Darcian flow, they became important contributors to runoff volume, comprising 15 % and 5 % respectively.

During the first six days of runoff, snow cover on the slope decreased from 78 % to 5 % on May 17. Daily observations revealed that progressive ground thaw altered drainage modes, pathways and source areas. Between May 16 and 17, rapid melt of the remaining snowpack combined with shallow heterogeneous thaw produced ponding in several shallow depressions. By May 17, approximately 70 % of the soil pipes were transmitting water (Figure A5*c*), accounting for 27 % of slope runoff based on contributing areas. Dye tracing indicated that soil pipes exhibited extensive hydrological connectivity. Water emerging from pipe effluents re-entered the organic layer or a rill segment and was routed to the nearest downslope pipe. In many cases, tracing indicated that flow did not follow the steepest topographic gradient, but passed through the organic layer to adjacent pipes.

By May 20, mean daily temperatures in excess of +12°C reduced the snowpack to small patches. Runoff over this period was intense, with subsurface drainage in the forms of pipe and Darcian flows accounting for 45 % and 25 % of slope drainage respectively. At this time, all pipes on the slope were transmitting water (Figure A5*d*).

Progressive thawing (which increased the soil storage capacity) and continued drainage of meltwater reduced slope runoff after May 20. On May 23, matrix flow accounted for 62 % of runoff but soil pipes declined in importance with only 30 % of the pipes being active (Figure

A5*e*), contributing 22 % of runoff. Pipes that ceased flowing first were those with the smallest contributing areas. Surface flow was confined to the main rill at the centre of the experimental area, delivering 16 % of runoff. By May 27, all runoff from the experimental slope occurred as matrix flow, confined primarily within the organic horizon.

For the entire snowmelt period, approximately 155 mm of runoff occurred, representing 80 % of the winter snowpack. During the 16 day snowmelt period, pipeflow accounted for 21 % of runoff, compared with 53 % and 26 % for rill and matrix flows respectively.

A3.3.2 Snowmelt hydrograph

Sunny days between May 13 and 17 provided clear diurnal runoff signals from the flumes and soil pipes (Figure A6). Strong diurnal trends in pipeflow at P-1 and P-2 and integrated slope runoff at N-1 were generated by the daily melt cycles, although there were significant differences in the timing and shape of the pipe and flume hydrographs. Both daily rise and fall in pipeflow were steeper than the total slope runoff, and there was no extended recession through the night as compared with flow at N-1. Pipeflow rose between 2 and 4 hours prior to flow at N-1, increasing from a base level of approximately 25 mL/s to a daily maximum between 1730 and 1930 PST, and then declined rapidly to the pre-rise values. Serial correlations (Figure A6 inset) between calculated hourly melt and runoff hydrographs show that pipeflow lagged behind the peak daily melt by 2-4 hours compared with 6 hours for integrated slope runoff. Shorter response times and larger correlation values for pipeflow suggest that meltwater was efficiently directed into soil pipes, with little mixing and interaction with the hillslope soils. In contrast, hydrograph signals at N-1 are an amalgamation of several runoff processes which delivered water at different rates, thus delaying the runoff response to snowmelt.

A3.3.3 Summer pipeflow

During summer, matrix flow was the principal runoff mode, particularly within the organic layer. Surface flow was absent except at the slope base where saturation excess and groundwater flow channeled into flume N-1.

Summer pipeflow was first generated by the June 4-6 storm which deposited 25.7 mm rainfall, representing the largest storm of the summer (Figure A7*a*) and generating the highest annual flow in Wolf Creek. During this storm, there were three separate intervals when rain fell. The rise in stormflow at N-1, which is an integrated slope response dominated by fast matrix flow within the organic layer, occurred after an initial abstraction of 0.9 mm, and began to decline after the cessation of the first rainfall event. At the onset of the second rainfall event on June 5, slope runoff increased rapidly, yet pipeflow and intermittent flow in rills were not observed until over 15 mm of rain had fallen. Pipeflow at P-1 and P-2 began and rose rapidly to exceed 150 mL/s over a 5 hour period. Pipe P-2 produced a second peak after a small decline during a lull in the storm. By the time P-3 began flowing 2 hours after P-1 and P-2, about half the pipes were transmitting water. Pipeflow peaked 3-8 hours after the stormflow peak, and declined almost immediately after the intense rain on June 5, well before the decline in stormflow. Between June 4 and 6, soil moisture within the organic layer increased from 33 % to 40 %, representing water storage enrichment of 14 mm. Hydrograph analysis indicated that 16.4 mm of precipitation was directed into stormflow runoff, of which approximately 2.5 mm was pipeflow. Of the 13.9 mm occurring as fast matrix flow, 8.2 mm ran off either during or within 10 hours after the cessation of rain, and the remaining 5.7 mm discharged in the following 7 days.

The second summer pipeflow event occurred on June 17 during a storm that deposited 13.9 mm of rainfall, 9 mm of which fell within a 1.5 hour period (Figure A7*b*). In contrast to the previous storm, pipeflow began and ceased during the rising limb of the stormflow hydrograph. Pipeflow began at P-1 and P-2 within 1 hour of the rain burst and rose steeply to a flow of 250 mL/s in 2 hours. No pipeflow was observed at P-3, and only five pipes on the slope yielded observed flow. Increases in runoff at N-1 began after 3.2 mm of rainfall, yet its hydrograph did not peak until 13 hours after the pipeflow peak. In total, 9.6 mm of stormflow runoff was measured, of which pipeflow accounted for only 0.5 mm. Stormflow was again divided into quick and extended response based on the inflexion point on the hydrograph prior to the return of flows to antecedent levels. Direct stormflow accounted for 5.5 mm of runoff and the extended recession was 3.6 mm over 5 days.

Similar to the melt responses, pipeflow hydrographs for both rainfall events displayed steeper rises and recessions than integrated slope runoff. The time for the onset of pipeflow to pipe hydrograph peak for the two storms were 5 and 2.5 hours compared with 26 and 24 hours for throughflow at N-1. Furthermore, recession times for pipes were rapid, ranging between 2.5 and 8 hours for the June 5 event and 1.5 hours for the June 17 event compared with stormflow recession times of 18 and 31 hours.

The early rise in pipe hydrographs for the June 17 event compares well with the observations in the East Twins catchment near Bristol, England, by Weyman (1970) who stated that where overland flow was absent, pipeflow produced a runoff peak before throughflow. The rapid pipeflow recessions contrasts with the observations of Jones and Crane (1984) and Wilson and Smart (1984) that pipeflow exhibits long recessions, often up to several days. The study pipes were similar to the ephemeral pipes in the Maesnant catchment, Wales, (Jones, 1987) where the timing for the commencement of flow varies from 2 hours before stormflow to almost 6 hours after, and drying up anywhere between one day before the end of stormflow and one-half day afterward. The ephemeral nature of pipes on the study slope suggests considerable variations in the timing of pipeflow response, depending upon rainfall characteristics and antecedent moisture conditions on the slope.

A4.0 Hillslope Stormflow Characteristics

Eight hydrographs measured from flume N-1 in 1997 were analyzed to assess the rainfall-runoff characteristics for the study slope (Table A2), and to compare slope responses with those of pipeflow. Events selected were for later summer storms with a single runoff peak and limited precipitation during the falling limb of the hydrograph.

The volume of precipitation required to initiate stormflow, termed initial abstraction (P_{abst}), was low, ranging between 0.9 to 2.6 mm with an average of 1.6 mm. Unlike relations reported in temperate regions (Dingman, 1994), P_{abst} is not correlated with properties of hillslope wetness such as antecedent discharge (Q_{ant}) and five-day antecedent precipitation (P_{5d}). McNamara *et al.*, (1998) at a continuous permafrost location in Alaska reported an average initial abstraction of 1.52 mm, and also observed no relations between P_{abst} and watershed wetness. This lack of correlation is due to the low subsurface storage capacity limited by the presence of permafrost, requiring little rainfall to saturate the ground sufficiently to raise slope runoff.

Rapid runoff response is typical of permafrost basins (Church, 1974). The response times (T_{resp}) between the start of precipitation and the hydrograph rise were 1.5-6 hours, with an average of 3.8 hours. McNamara *et al.*, (1998) reported an average T_{resp} of 2.15 hours, and credited rapid responses to a network of water tracks (Kane *et al.*, 1989), which remained saturated for much of the summer, except during extreme dry times. Dingman (1971, 1973) attributed the rapid response of a discontinuous permafrost basin near Fairbanks, Alaska, to the wet valley bottoms. On the study slope, a combination of the above two principles applies. The water table is highly variable, existing within the organic layer throughout the summer at the slope base and in intermittently perched zones. During storms, the water table rises into the organic zone at progressive upslope locations, increasing slope flow as hydraulic conductivities change between unsaturated and saturated states. Furthermore, heavy storms change topographic depressions to zones of concentrated wetness, creating preferential flow zones which function like water tracks.

Hillslope recession characteristics varied depending upon rainfall characteristics and slope moisture status. During early summer, stormflow recessions exhibited two distinct phases: 1. an initial phase with steep recession occurring for 20 to 66 hours after the throughflow peak and 2. an extended recession with a gentle decline, lasting in excess of 5 days (Figure A7*a,b*). This extended recession reflected an expanded catchment area and increased upslope hydrological connections when the mean water table position resided within the organic layer. As the source area for slope runoff expanded, it took more time for the water from upslope locations to reach the slope base.

Recession limbs of stormflow typically approximate exponential decays according to

$$q = q_0 e^{\left(-t/t^*\right)} \tag{A3}$$

where q is the flow rate t hours after the beginning of the recession, q_O is the rate at the beginning of the recession, and t^* (in hours) is the recession constant. Values of t^* for the steep stormflow recession of slope runoff range between 25 and 75 hours (Figure A8; Table A2). These values are greater than those for the temperate regions (Holtan and Overton, 1963), but compare well with values from other permafrost basins. Dingman (1973) calculated an average t^* of 39 hour for a 1.8 km^2 basin near Fairbanks, Alaska and McNamara *et al.*, (1998) found t^* to range from 20 to over 100 hour, increasing positively with the drainage area. Although extended recessions observed on the study slope do not follow prominent exponential decays, the approximated t^*

values exceed 150 hours. The large recession constants for permafrost slopes are explained by McNamara *et al.*, (1998) that no rainfall goes into long-term storage, but the water input is released as direct stormflow over an extended time period.

The response factor (*RF*), expressed as the ratio of direct runoff to precipitation, showed marked differences between storms. Direct runoff was calculated by subtracting the baseflow from the hydrograph using a baseflow separation technique. For this study, separation was performed by projecting a straight line across the hydrograph from the point of initial rise to the point where discharge returns to its antecedent value. The response factor showed some correspondence with date, a surrogate for frost table position. A major reduction in *RF* occurred following the descent of the average slope water table (measured at 15 wells) into the mineral zone (Figure A9). As active layer depth increased, the storage capacity increased at the expense of runoff generation. Prior to the descent of the frost table into the mineral layer, *RF* exceeded 0.4, and when the extended recession periods are included, *RF* exceeded 0.6. As the water table dropped into the mineral layer, *RF* declined to an average of 0.19, although there was some variation among storms. The July 22 event was a notable outlier, with a *RF* of 0.33, over 15 % greater than the other storms for the same period. The large *RF* is due to both the magnitude and intensity of the rainstorm.

Response factors in permafrost regions exceeded those in more temperate environments except for basins with a thin soil such as the Canadian shield due to the limited basin capacity to store new water. Response factors exceeding 0.3 are common in permafrost basins (McNamara *et al.*, 1998) compared with less than 0.1 often observed in humid, temperate basins (Dingman, 1994). Slaughter *et al.*, (1983) compared three basins in interior Alaska with differing percentages of their areas underlain by permafrost, and showed that the basins with more permafrost generated higher flows per unit area, regardless of their antecedent moisture conditions. Dingman

(1971) reported an average value of 0.18 for Glenn Creek basin near Fairbanks, Alaska, 60 % of which is underlain by permafrost. The permafrost sections of the Glenn Creek watershed have similar active layer thickness as those for the study slope, thus explaining similar values of *RF*. The most notable reductions in *RF* during the year were related to the descent of the frost table, which affects the storage capacity of slopes.

A5.0 ROLE OF PIPEFLOW IN SLOPE RUNOFF

The observation that soil pipes are important contributors to runoff only during the snowmelt period compares well with other cold-region hillslope studies (Roberge and Plamondon, 1987; Gibson, 1991; Gibson *et al.*, 1993). In temperate latitudes, rainfall and the position of the water table alone control pipeflow responses. However, in permafrost terrain, water table position in the soil is moderated by the position of the frost table along with the frequency and magnitude of precipitation events. When hillslopes are saturated, pipeflow becomes an important matrix bypass mechanism which enhances the rate of runoff. On an annual basis, pipeflow contributions are the greatest during snowmelt when conditions are wettest and a high degree of hydrological connectivity allows soil pipes to drain the saturated thawed zone. For the snowmelt period of May 12- 27, pipeflow accounted for 21 % of runoff, compared with 53 % for rill flow and 26 % for matrix flow in the organic layer. In contrast, pipes play a limited role in summer runoff and stormflow: on two occasions it delivered a total of 3.5 mm, accounting for less than 3 % of summer runoff. However, during large rainfall events in early summer when both water and frost tables are near the surface, pipeflow can contribute over 15% for large storms.

Pipeflow is a runoff mechanism intermediate between matrix flow and surface flow, transmitting water at near-surface velocities through the subsurface zone (Table A3). The faster transmission of water within pipes in part explains the brevity of pipeflow hydrographs during summer storms. Jones (1997), placing pipes within an overall runoff framework developed by Dunne (1978), suggested that peak lag times for ephemeral pipes with limited contributing areas are 30 to 40 % shorter than throughflow. Daily snowmelt cycles, analogous to rainfall pulses, generated daily pipeflow peaks that occurred between 2 and 4 hours ahead of the integrated slope runoff peak. During the later stages of melt, the lag between pipeflow and runoff at flume N-1 increased, as matrix flow contributed a greater portion of total slope runoff.

After snowmelt, pipe responses were more variable due to wetting required to initiate pipeflow. Although pipeflow rises quickly to its peak, followed by a rapid recession, there may be a significant delay in the start of pipeflow, accounted for by the need for the phreatic surface to reach the pipe zone. Variability between events was evident in the summer of 1997. For the June 4-6 event, pipeflow reached its maxima between 3 and 8 hours after runoff peaked, but for the June 16-17 storm, peak pipeflow occurred 13 hours before stormflow.

Contributing areas for individual pipes on the study slope are small compared with elsewhere. For example, the contributing areas (Table A1) for one of the study pipes are an order of magnitude less than for ephemeral pipes in the Maesnant catchment, Wales (Jones, 1997). This limits the ability for the study pipes to convey water over long distances. Contributing areas on the study slope are also dynamic, changing within and between seasons. During snowmelt, source areas change daily as the disposition of the snowpack and thawed ground affect water availability and the flow pathways. When bare ground appears, the contributing areas enlarge gradually as ground thaw improves subsurface flow connections. Contributing areas reach a maximum at the end of the melt period when all pipes are active and when saturated zones are widespread on the slope. Continued ground thaw and water table decline cause a reduction in the source areas, yet during early summer when the phreatic surface lies within the organic zone, contributing areas are large enough to sustain both pipe and matrix flows at the slope base. As the mean hillslope water table drops into the mineral layer, source areas and upslope hydrological connections are reduced. Pipes become ineffective in transmitting water downslope. During summer storms, expansion of the source area is primarily controlled by vertical rises of the phreatic surface so that widespread upward movement of the water table into the organic layer will again increase the hydrological connections and re-activate pipeflow.

A6.0 CONCEPTUAL MODEL OF PIPEFLOW

Despite the common occurrence of pipeflow in permafrost slopes, no model is available to conceptualize the hydrologic processes. Figure A10 is a synthesis of the processes associated with the activation of pipeflow. For organic-covered permafrost slopes, infiltration-excess runoff rarely occurs. Meltwater can enter and percolate the organic layer even under frozen conditions as the top zone is typically made porous by upward flux of moisture to the overlying snowpack during the winter (Santeford, 1979; Smith and Burn, 1987). The very low hydraulic conductivity of the frozen mineral soils limits further percolation and the organic layer becomes saturated from the base upward. Pipe formation is favoured at the organic/mineral interface. During the first stages of snowmelt runoff, uneven ground thaw and the heterogeneous occurrence of ground ice in mineral soils confine water flow to a saturated zone at the base of the snowpack (Dunne *et al.*, 1976) and along snow-free rills. Soon after the widespread appearance of bare ground, thawing reaches the base of the organic layer where soil pipes occur, allowing pipeflow to begin. As thawing deepens and the water table drops, surface flow declines in favour of pipe and fast matrix flow in the organic layer. In early summer, the water table falls below the organic zone due to the termination of snow melt and a descending frost table. At this point, slow matrix flow in the mineral soil prevails but this yields low quantities of runoff relative to other modes of flow. Intense rainstorms in the summer may raise the water table to intersect the zone in which pipes occur, causing re-activation of pipeflow. This mechanism was observed when the average water table on the slope was within 0.17 m of the surface, the approximate depth at which pipes occur (Figure A9). Two features are notable for permafrost slopes. (1) Large spatial variation in ground thaw rate, together with contrasts in organic layer thickness, microtopography and pipe presence, often cause various runoff modes to operate at different locations on the slope, especially during melt. (2) The position of the frost table significantly influences the water table so that in late summer when the frost is deeper, more rainwater percolation is needed to raise the saturated zone to the organic-mineral soil interface.

The above conceptualization of pipeflow captures the idea that water table has to rise above a certain threshold level for activation. Roberge and Plamondon (1987) indicated that the water table has to rise from the mineral substrate into surface organics to activate pipeflow and the intensity of flow is related to the extent of water table rise. Gilman and Newson (1980), McCaig (1983) and Wilson and Smart (1984) working on humid British hillslopes, noted that pipeflow occurs when certain moisture thresholds are exceeded so that lateral drainage can begin in pipes. For permafrost slopes, variable groundwater storage capacity due to changing frost table depth must be considered. Thus, pipeflow re-activation has to be considered in conjunction with both the precipitation input and the frost table depth in the slope.

A7.0 CONCLUSIONS

Soil pipes have been reported at many subarctic permafrost sites. One such site was studied to evaluate the role that pipeflow plays in slope hydrology. Pipeflow generation and flow mechanisms as well as the timing and magnitude of runoff were monitored from snowmelt to the end of summer period, yielding several conclusions.

(1) Quantitatively, pipeflow is important mainly during snowmelt when pipes provide over 20 % of slope runoff. In summer, pipeflow diminishes and matrix flow in the organic layer dominates rainfall-runoff responses.

(2) Manning's equation is applicable to represent flow within natural pipes.

(3) Pipeflow response is intermediate between surface flow and fast matrix flow in the organic soil, transmitting water through the subsurface at velocities similar to overland flow.

(4) During melt, pipeflow hydrographs closely mirror daily melt cycles, with peaks lagging 2 to 4 hours behind maximum daily melt and ahead of integrated slope runoff.

(5) The timing of pipeflow response in summer months is variable compared with matrix flow as the water table must rise into the pipe zone before flow occurs.

(6) Conceptualization of pipeflow requires that the phreatic surface which is strongly influenced by the depth of the frost table intersects the zone in which pipes exist; this then allows the subsurface flow to drain preferentially from the surrounding organic matrix into the pipes.

A8.0 REFERENCES

Carey, S.K., Woo, M.K., 1998. Snowmelt hydrology of two subarctic slopes, Southern Yukon, Canada. Nordic Hydrol. 29, 331-346.

Carey, S.K., Woo, M.K. 1999 Hydrology of two slopes in subarctic Yukon, Canada. Hydrol. Processes 13, 2549-2562.

Church, M., 1974. Hydrology and permafrost with reference to northern North America, In: Proc. Workshop Seminar on Permafrost Hydrology, Canadian National Committee, IHD, Ottawa, pp. 7-20.

Dickinson, W.T., Whiteley, H., 1970. Watershed areas contributing to runoff. IASH Publ. 96, 12-26.

Dingman, S.L., 1971. Hydrology of Glenn Creek watershed, Tanana basin, Central Alaska. U.S. Army Cold Region Research Engineering Laboratory Research Report 297, Hanover, New Hampshire, p. 112.

Dingman, S.L., 1973 Effects of permafrost on stream flow characteristics in the discontinuous permafrost zone of central Alaska, In: North American Contribution to Second International Conference of Permafrost, National Academy of Sciences, Washington, D.C., pp. 447- 453.

Dingman, S.L., 1994. Physical Hydrology. Prentice Hall, New Jersey, p. 575.

Dunne, T., 1978 Field studies of hillslope processes, In: Kirkby, M.J. (Ed.), Hillslope Hydrology. Wiley, Chichester, pp. 227-294.

Dunne, T., Price, A.G., Colbeck, S.C., 1976. The generation of runoff from subarctic snowpacks. Water Resour. Res. 12, 677-685.

Freeze, R. A. and Cheery, J.A., 1979. Groundwater, Prentice-Hall Inc., New Jersey, p. 604.

Gibson, J.J., 1991. Isotope hydrology and water balance investigations in the Manners Creek watershed, District of Mackenzie, Northwest Territories. Unpublished M.Sc. Thesis. University of Waterloo, Canada, p. 215.

Gibson, J.J., Edwards, T.W.D., Prowse, T.D., 1993. Runoff generation in a high boreal wetland in northern Canada. Nordic Hydrol. 24, 213-224.

Gilman, K., Newson, M.D., 1980. Soil pipes and pipeflow – a hydrological study in upland Wales. British Geomorphological Research Group Monograph No. 1, Geobooks, Norwich. p. 114.

Hinzman, L.D., Kane, D.L., Everett, K.R., 1993. Hillslope hydrology in an Arctic setting, In: Proceedings, Sixth International Conference on Permafrost, South China Press, Beijing, pp. 257-271.

Holtan, H.N., Overton, D.E., 1963. Analyses and application of simple hydrographs. J. Hydrol. 2, 309-323.

Jones, J.A.A., 1971. Soil piping and stream channel initiation. Water Resour. Res. 7, 602-610.

Jones, J.A.A., 1987. The effects of soil piping on contributing areas and erosional patterns. Earth Surf. Proc. Landforms 12, 229-248.

Jones, J.A.A., 1988. Modelling pipeflow contributions to stream runoff. Hydrol. Processes 2, 1-17.

Jones, J.A.A., 1997. Pipeflow contributing areas and runoff response. Hydrol. Processes 11, 35-41.

Jones, J.A.A., Crane, F.G., 1984. Pipeflow and pipe erosion in the Maesnant experimental catchment, In: Burt, T.P. and Walling, D.E. (Eds.) Catchment Experiments in Fluvial Geomorphology, Geobooks, Norwich, pp. 55-72.

Kane, D.L., Hinzman, L.D., Everett, K.R., Benson, C.S., 1989. Hydrology of Imnavait Creek, and arctic watershed. Holarctic Ecol. 12, 262-269.

McCaig, M., 1983. Contributions to storm quickflow in a small headwater catchment – the role of natural pipes and soil macropores. Earth Surf. Proc. Landforms 8, 239-252.

McNamara, J.P., Kane, D.L., Hinzman, L.D., 1998. An analysis of streamflow hydrology in the Kuparuk River Basin, Arctic Alaska: a nested watershed approach. J. of Hydrol. 206, 39-57.

Quinton, W.L., 1997. Runoff from Hummock-covered Arctic Tundra hillslope in the Continuous Permafrost Zone. Unpublished Ph.D. Thesis. University of Saskatchewan, Saskatoon, Canada, p. 277.

Quinton, W.L., Marsh, P., 1998. The influence of mineral earth hummocks on subsurface drainage in the continuous permafrost zone. Permafrost and Periglacial Processes 9, 213-228.

Roberge, J., Plamondon, A.P., 1987. Snowmelt runoff pathways in a boreal forest hillslope: the role of pipe throughflow. J. Hydrol, 95, 39-54.

Santeford, H.S., 1979. Snow soil interactions in interior Alaska, In: Colbeck, S.C. and Ray, M. (Eds.) Proceedings, Modeling of Snow Cover Runoff. Cold Regions Research and Engineering Laboratory, Hanover, NH, pp. 311-318.

Slaughter, C.W., Hilgert, J.W., Culp, E.H., 1983. Summer streamflow and sediment yield from discontinuous-permafrost headwater catchments, In: Proceedings, Fourth International Conference on Permafrost, National Academy Press, Washington, D.C., pp. 1172-1177.

Smith, M.W., Burn, C.R., 1987. Outward flux of vapour from frozen soils at Mayo, Yukon, Canada: results and interpretation. Cold Reg. Sci. Tech. 13, 143-152.

Weyman, D.R., 1970. Throughflow on hillslopes and its relation to the stream hydrograph. Bull. Int.Assoc. Sci. Hydrol. 15, 25-33.

Wilson, C.M., Smart, P.L., 1984. Pipes and pipeflow processes in an upland catchment, Wales. Catena 11, 145-158.

Woo, M.K., Steer, P., 1982. Occurrence of surface flow on arctic slopes, southwestern Cornwallis Island. Can. J. Earth Sci. 19, 2368-2377.

Table A1. Maximum contributing areas (in m^2) for study pipes at the end of snowmelt period and during two summer storms.

Date, 1997	P-1	P-2	P-3
May 20	340	220	180
June 5	248	151	40
June 17	106	144	-
Pipe diameter (mm)	89	80	84

Table A3. Hillslope runoff velocities.

Runoff Mode	Velocity (m/s)
Rill flow	0.15-0.45
Pipeflow	0.10-0.30
Matrix flow (organic layer)	10^{-2}-10^{-4}
Matrix flow (mineral layer)	10^{-8}-10^{-9}

Table A2. Stormflow hydrograph characteristics for selected rainfall events, 1997.

Storm Date, 1997	P_t *(mm)*	P_d *(hr)*	P_{int} *(mm/hr)*	P_{5d} *(mm)*	P_{abst} *(mm)*	T_{resp} *(hr)*
June 4 to 6	25.6	42	0.61	4.8	0.9	6
June 17/18	13.9	15.5	0.90	1.8	3.2	3
June 28/29	8.8	19	0.46	3.9	1.5	5
July 3	22.5	73	0.31	11.2	2.2	5
July 7	9	30	0.30	22.8	2.6	4
July 22	11	11	1.00	0.2	0.8	2.5
July 26	6.9	4	1.73	11.6	0.9	1.5
July 31	3.8	10	0.38	8.5	1.1	6

Storm Date, 1997	T_{rise} *(hr)*	T_{fall} *(hr)*	T_{ef} *(hr)*	Q_{pk} *(mL/s)*	Q_{ant} *(mL/s)*	*RO (mm)*	RO_e *(mm)*	*RF*	t^* *(hr)*
June 4 to 6	25	31	159	260	20	8.2	5.7	0.41/0.6	21
June 17/18	24	29.5	120	154	30	5.5	3.6	0.43/0.7	40
June 28/29	27.5	32.5		68	41	1.7		0.22	35
July 3	24	66		137	90	2.7		0.14	75
July 7	13	33.5		147	115	1.2		0.16	32
July 22	13	36		150	41	4.8		0.33	25
July 26	12	42		88	42	2.3		0.16	40
July 31	12	20		80	67	0.3		0.10	44

P_t – total rainfall
P_d – rainfall duration
P_{int} – rainfall intensity
P_{5d} – 5-day antecedent rainfall
P_{abst} – initial precipitation abstraction
T_{resp} – time from start of precipitation to rise in hydrograph
T_{rise} – duration of hydrograph rise
T_{fall} – duration of hydrograph fall
T_{ef} – extended recession period
Q_{pk} – peak runoff
Q_{ant} – antecedent runoff
RO – stormflow runoff
RO_e – extended slope runoff
RF – response factor

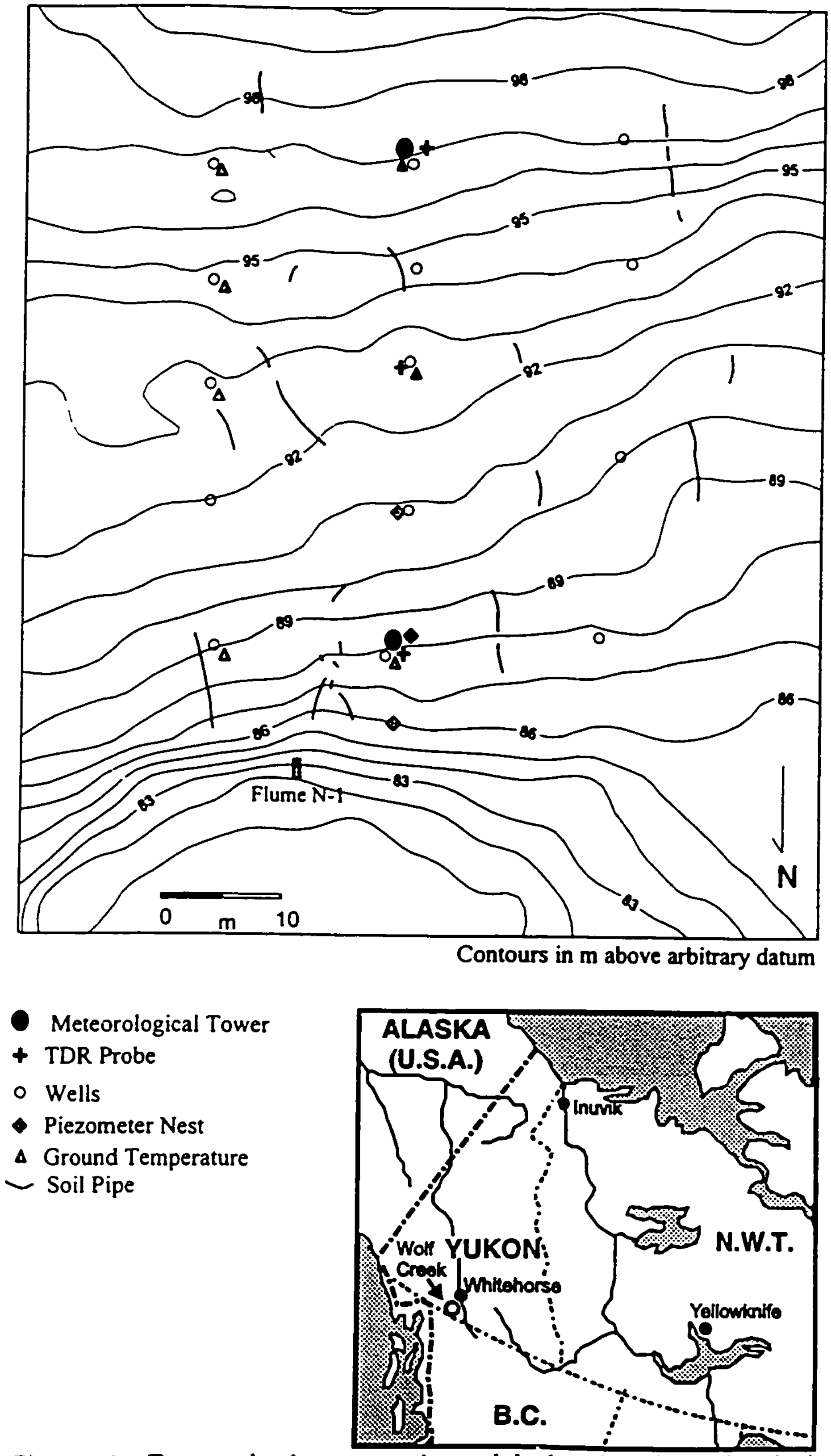

Figure A1. Topography, instrumentation and the location of soil pipes in the experimental slope. Inset shows location of Wolf Creek basin.

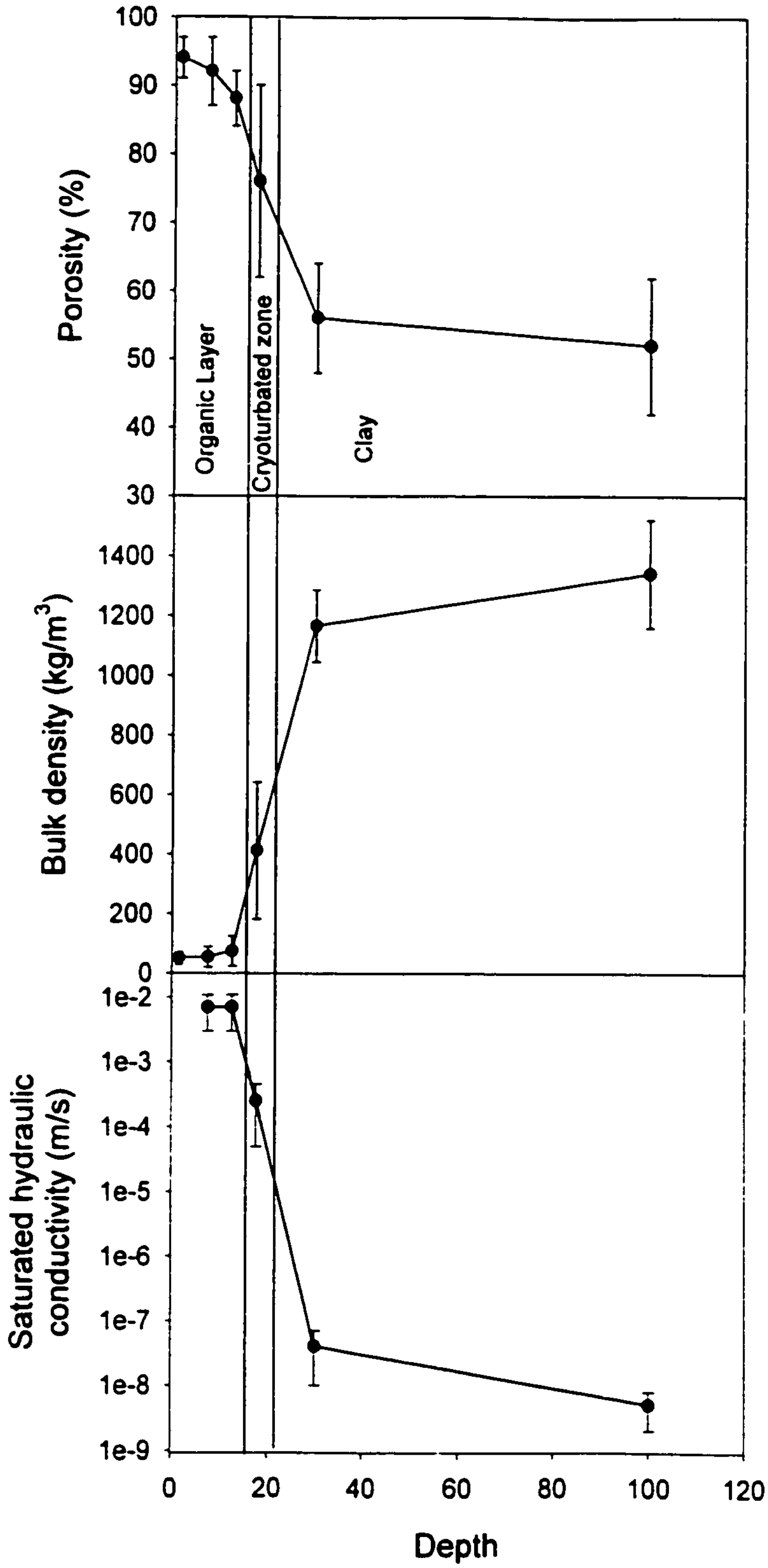

Figure A2. Vertical changes in porosity, bulk density and saturated hydraulic conductivity. Dots represent the location of measurement; horizontal lines are error bars.

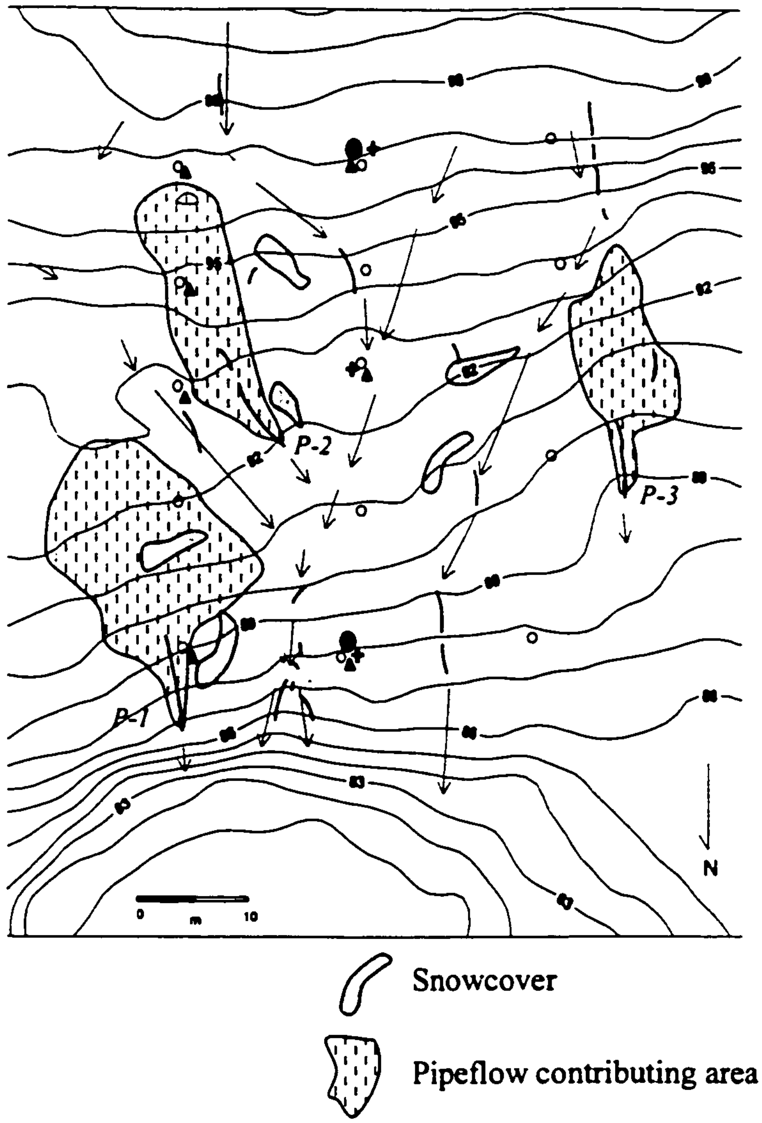

Figure A3. Pipeflow contributing areas (hatched) and drainage directions (arrows) based on dye tracing and field observations, May 20, 1997. Residual snow patches are shaded.

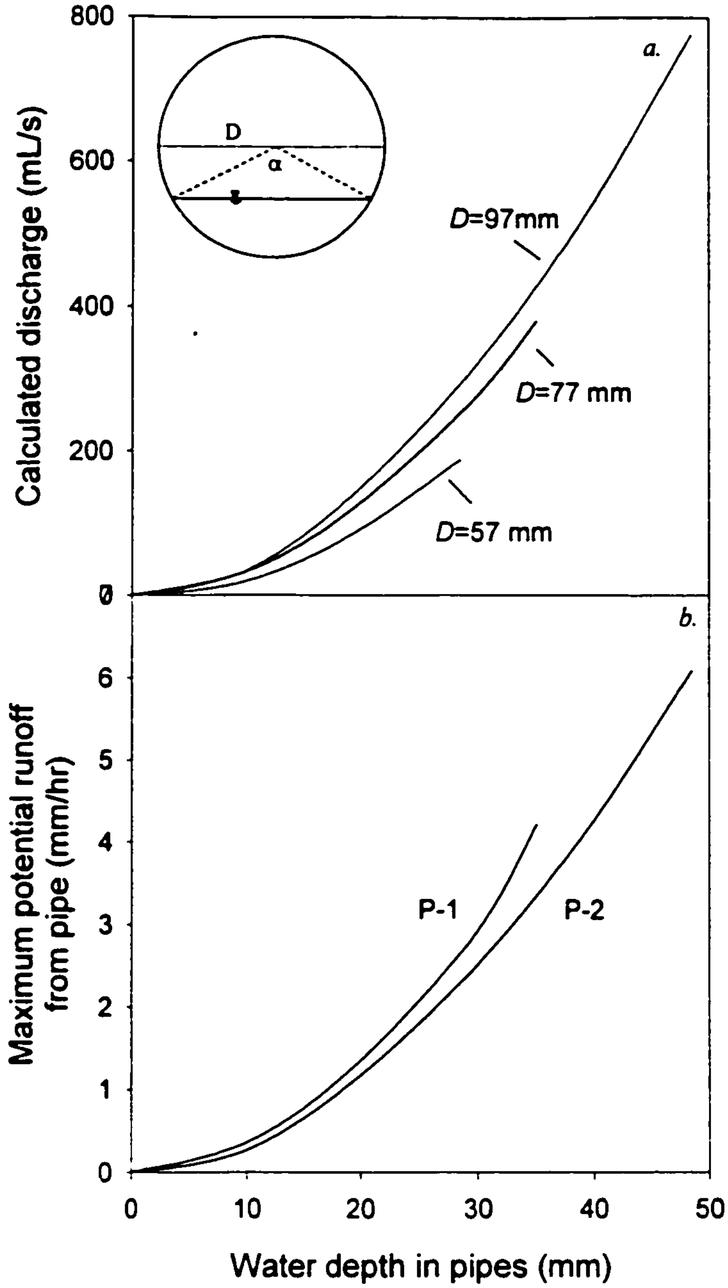

Figure A4: (*a*) Calculated discharge using Manning's equation vs. water level within a theoretical soil pipe. D represents the pipe diameter. Inset is the theoretical pipe dimensions. (*b*) Maximum potential runoff from P-1 and P-2 vs. water level in the pipes.

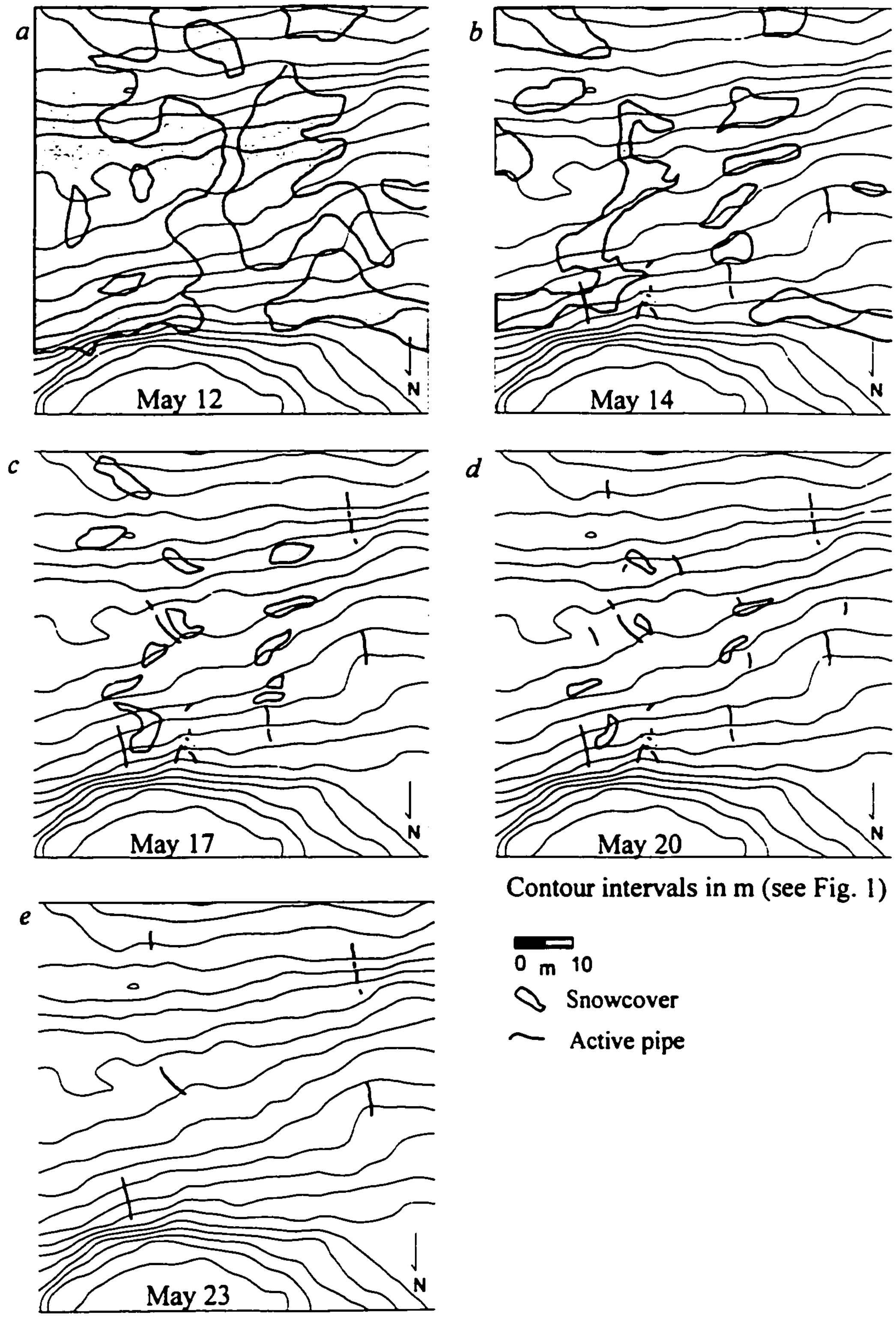

Figure A5. Active soil pipes for selected days during the snowmelt period.

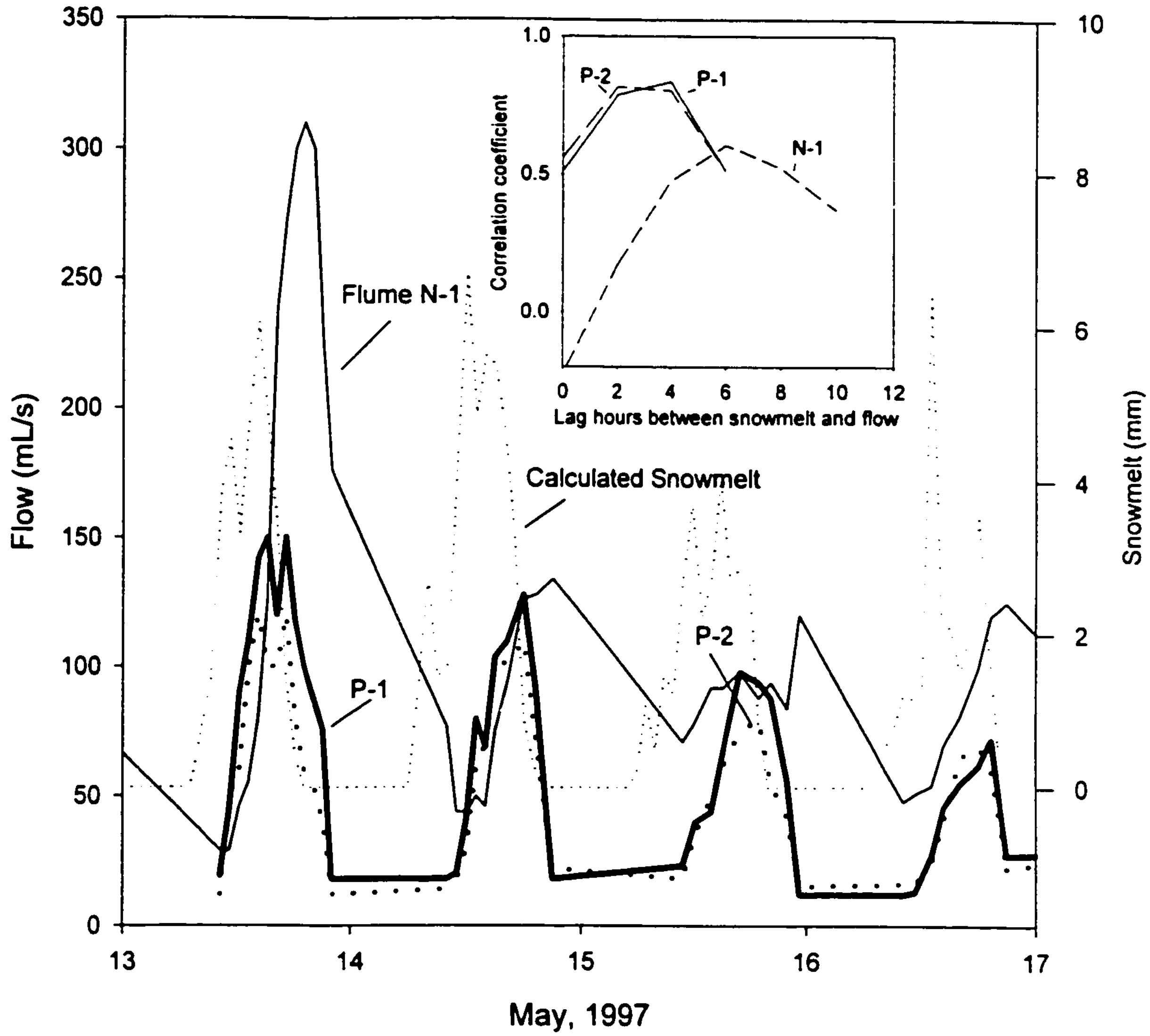

Figure A6: Pipeflow at P-1, P-2 and integrated slope runoff at N-1 compared with calculated snowmelt (dashed line). Inset shows serial correlations between snowmelt and runoff from soil pipes and the slope.

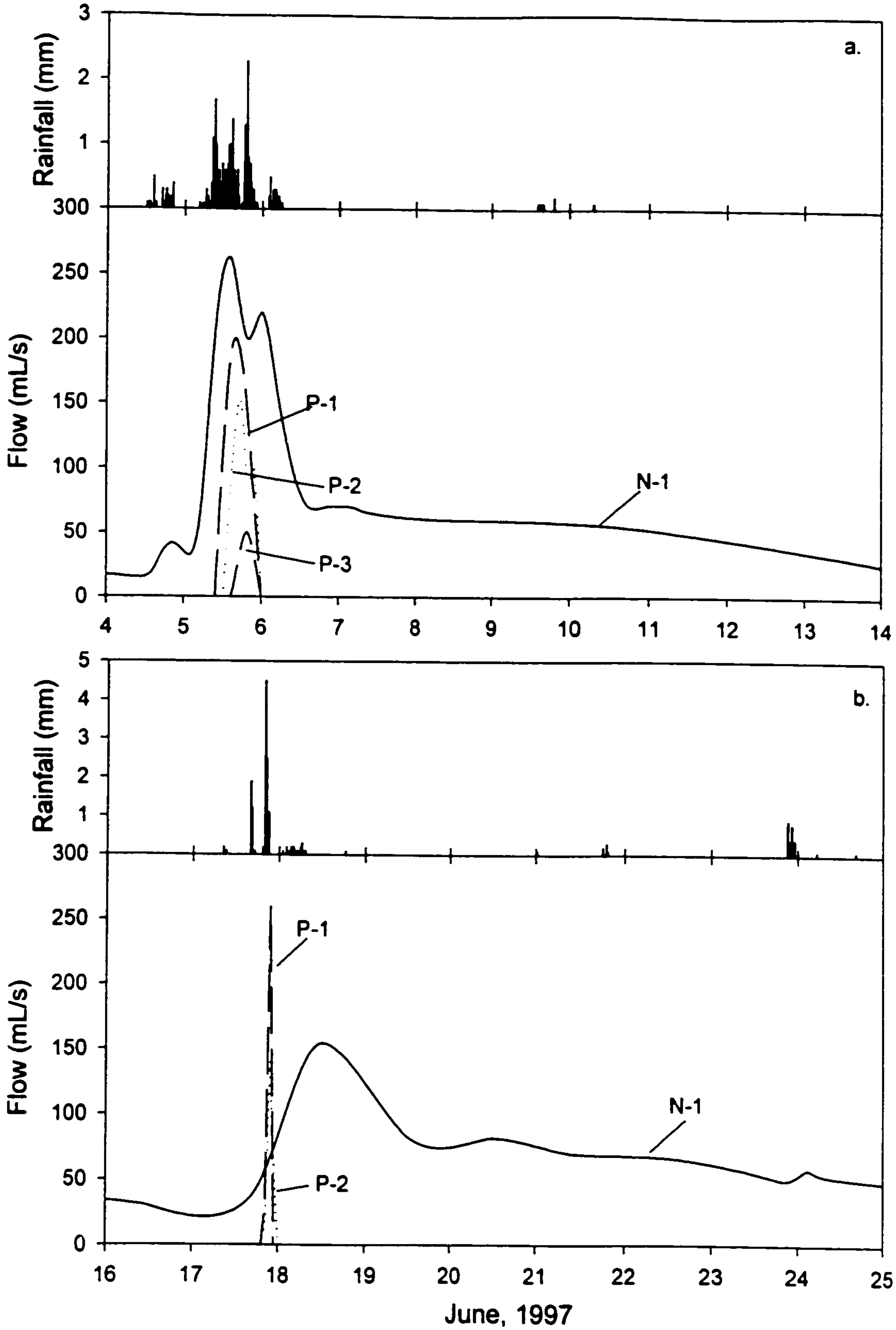

Figure A7. Hyetographs and hydrogaphs for the two summer storms that produced pipeflow responses.

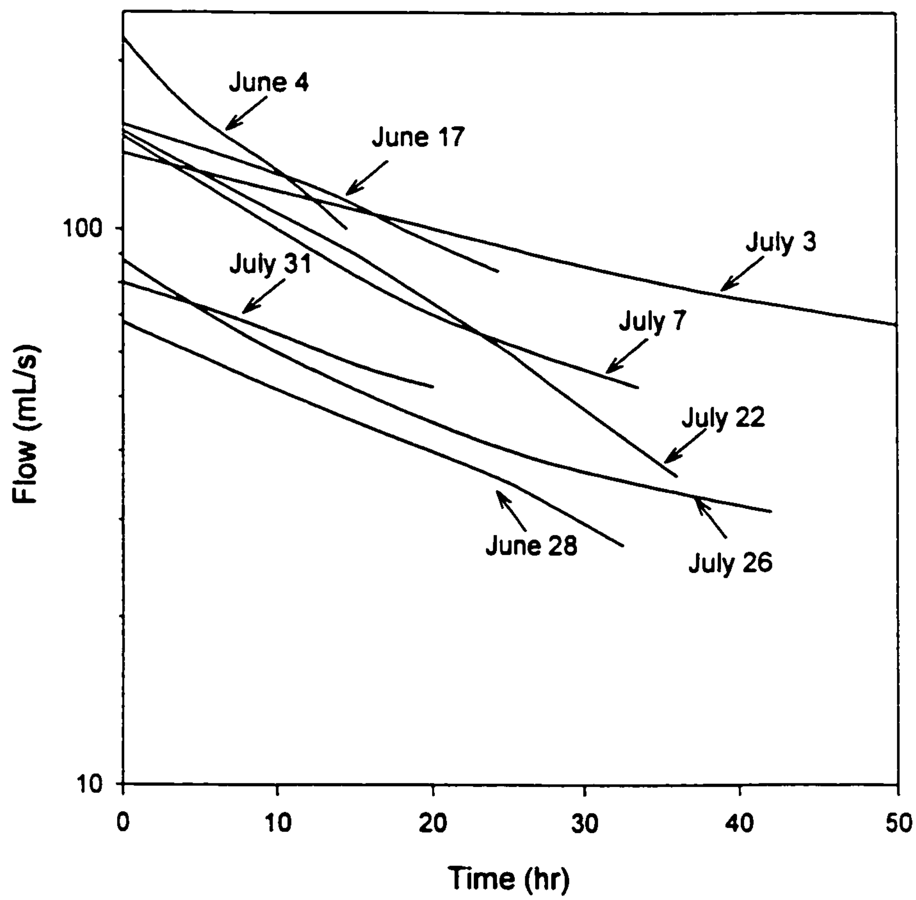

Figure A8: Measured discharge versus time for eight simple flow recessions.

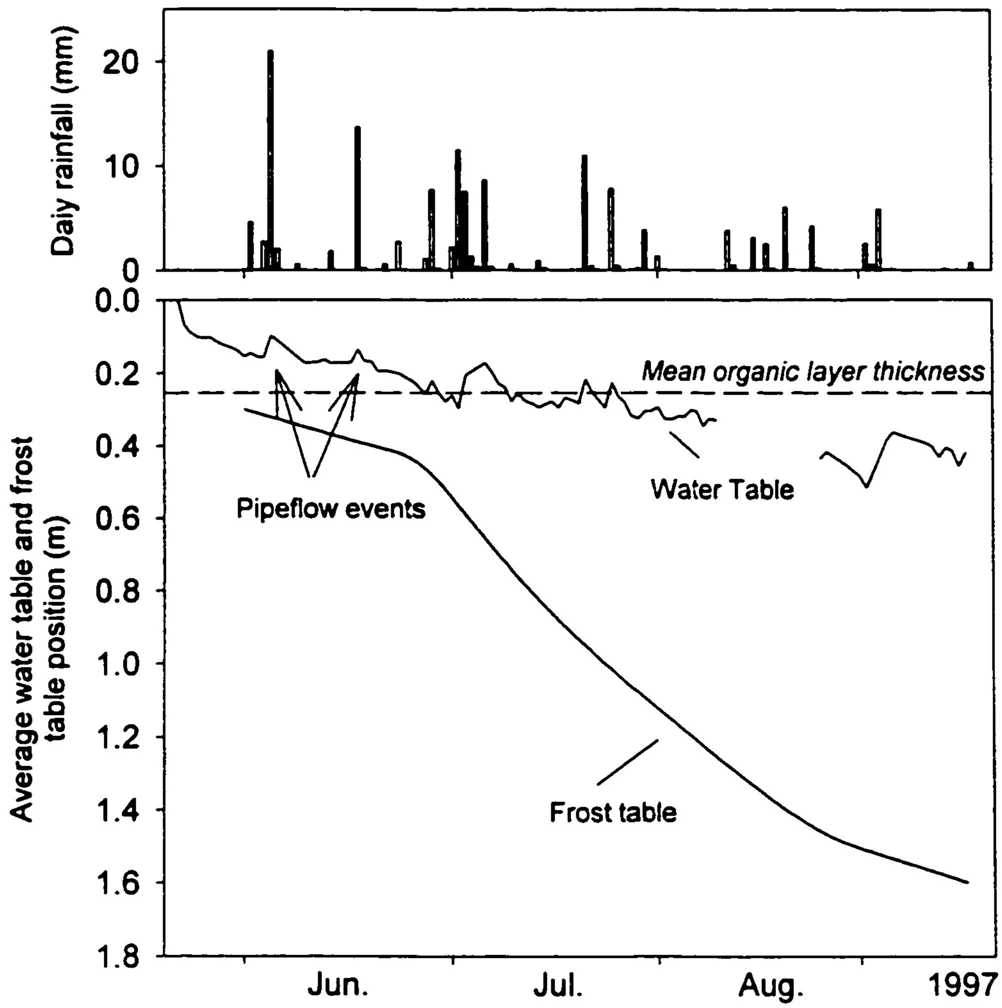

Figure A9. Daily precipitation, average frost and water table positions, summer 1997.

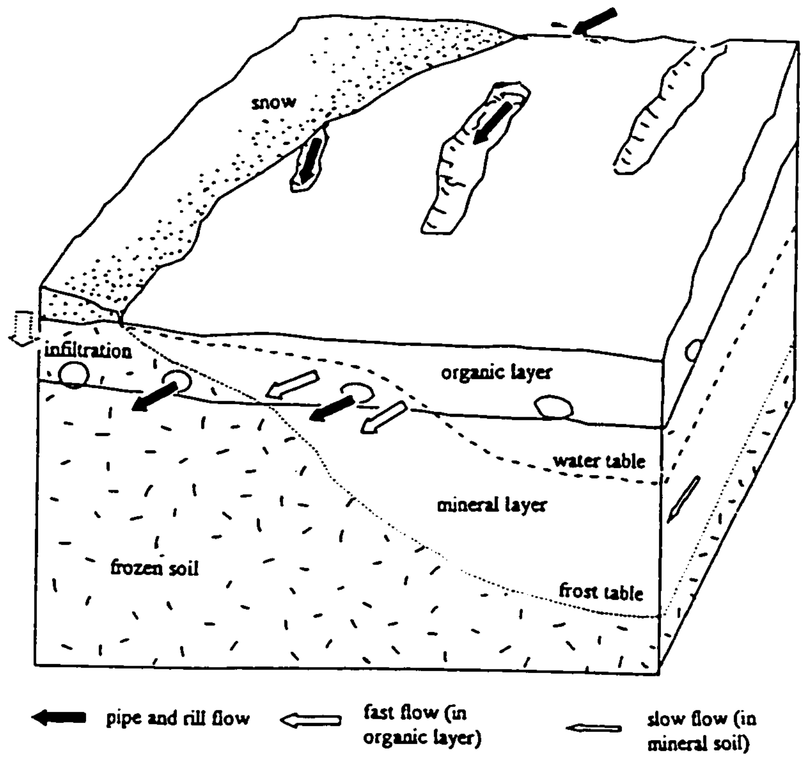

Figure A10. Conceptualization of occurrence of pipeflow, rill flow and matrix flows in a subarctic slope underlain by permafrost.

APPENDIX B

RUNOFF CALCULATIONS AND VARIABLE COMPARISONS

B1.0 RUNOFF CALCULATIONS

In Chapters 4 and 5, runoff was both measured and calculated as a residual of the water balance. Tables B1 and B2 compare measured runoff with runoff calculated as a residual in the water balance equation (Equation 4.6). For the North Slope, evapotranspiration was calculated using the Bowen Ratio method (Equation 4.2) whereas on the West Slope, the Priesteley and Taylor method was used (Equation 4.4).

Table B1. Comparison of measured and calculated runoff for the North slope, 1997

North Slope

Storm Date, 1997	*Runoff - measured (mm)*	*Runoff - calculated (mm)*
June 4 to 6	8.2	6.8
June 17/18	5.5	4.6
June 28/29	1.7	3.2
July 3	2.7	3.3
July 7	1.2	2.0
July 22	4.8	3.3
July 26	2.3	1.5
July 31	0.3	1.0
September 5	0.9	1.4

Table B2. Comparison of measured and calculated runoff for the West slope, 1997

West Slope

Storm Date, 1997	*Runoff - measured (mm)*	*Runoff - calculated (mm)*
June 17/18	3.8	4.3
June 28/29	0.9	1.2
July 7	0.8	0.7
July 22	2.2	1.8
July 26	0.8	1.1
September 5	0.4	0.3

B2.0 Variable Comparisons

Tables B3 and B4 summarize the results of t-tests at the 0.05% confidence level to indicate whether environmental variables and water balance components are significantly different between the North and South slopes in 1997 and the East and West slopes in 1999. SD indicates that the difference in the mean values of the two groups is greater than would be expected by chance; there is a statistically significant difference between the input groups. ND indicates that the difference in the mean values of the two groups is not great enough to reject the possibility that the difference is due to random sampling variability. There is not a statistically significant difference between the input groups.

Table B3. T-test comparison of atmospheric variables between the North and South slopes, 1997

Variable	*Result*
Air Temperature	SD
Net Radiation	SD
Incident Short-wave radiation	SD
Relative Humidity	SD
Wind Speed	SD

Table B4. T-test comparison of atmospheric variables between the East and West slopes, 1999

Variable	*Result*
Air Temperature	SD
Net Radiation	ND
Incident Short-wave radiation	ND
Wind Speed	SD