VARIABILITY OF WINTER STREAMFLOW IN SUB-ARCTIC RIVERS

by

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Variability of Winter Streamflow In Sub-Arctic Rivers

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ABSTRACT

The period of winter low flows is critical in terms of aquatic habitat and water quality in the sub-Arctic. However, little is known of winter streamflow variability in sub-Arctic rivers. This research investigated the nature and causes of winter streamflow variability in southern Yukon. Frequent discharge measurements made through the winter of 1994-95 were analyzed, together with water quality, near-stream piezometric, soil temperature, ice-cover temperature, and climatic data.

Four conceptual storage-depletion models were tested for two groundwater-dominated streams. Data during and immediately following the period of active ice formation, during which streamflow dipped below the recession trend, were excluded from the analysis. Three of the four models (the 'non-linear reservoir', 'layered linear reservoir' and 'multiple linear reservoir' models) fit observed discharge for the calibration period within ±10%, whereas the 'linear reservoir' model over-estimated early winter flows and under-estimated late winter flows. The 'layered linear reservoir' model, which plots as two intersecting straight line segments on a semi-logarithmic hydrograph, was most consistent with uncalibrated pre-freeze-up recession flow and with water quality indications of reservoir structure. Large negative deviations from the recession trend at and immediately following freeze-up were caused, at least in part, by transient increases in channel storage. However, at one river, only 30% of the discharge depression could be accounted for as water going into channel storage. At this river, a reversal of near-stream hydraulic gradient at freeze-up may have blocked groundwater

inputs, accounting for the majority of the discharge depression. Deviations from the recession trend during the calibration period appeared to be non-random but could not be accounted for in terms of air temperature. It is hypothesized that if the stage-up response is non-uniform along the length of a channel, then stage disturbances may propagate upstream causing quasi-periodic discharge fluctuations at a given cross-section.

Lake storage depletion accounted for approximately 43% of the winter streamflow at the outlet of Kusawa Lake, with the majority of the storage depletion occurring in early winter, when lake stage was high. Ice formation at the lake outlet caused lake outflow to dip below the recession trend three times during the winter. The lake outlet polynya became partially re-established between each of these events despite sustained sub-freezing weather.

ACKNOWLEDGEMENTS

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CHAPTER ONE

INTRODUCTION

1.1 SIGNIFICANCE AND NATURE OF WINTER STREAMFLOW VARIABILITY IN SUB-ARCTIC RIVERS

Snowmelt runoff in spring and early summer dominates sub-Arctic streamflow volumes, while low flows occur during the winter months. Most previous studies of sub-Arctic hydrology have focused on the spring-summer high flows, primarily due to concerns with flooding and reservoir inflows. Of particular interest have been phenomena related to ice jam formation (e.g., Beltaos et al., 1990; Prowse and Gridley, 1993). As a result of this, less effort has been expended on operational streamflow surveys during winter, as compared to the break-up and open-water periods. For example, Water Survey of Canada (WSC) typically makes only three discharge measurements during the period of winter ice cover, and interpolates daily flows from these three points. The comparable agency in the United States, the U.S. Geological Survey (USGS), uses a similar program of data collection.

More recently, issues related to aquatic habitat and water quality have gained importance (e.g. Whitfield and McNaughton, 1986; Cheng et al., 1993; Power et al., 1993; Whitfield et al., 1995). Discharge affects habitat parameters such as water depth and velocity. Variability in winter discharge can affect winter survival of fish through overcrowding and the freezing of redds and juvenile fish (Power et al., 1993). Protection and enhancement of aquatic habitat for desirable species requires an understanding of the processes governing winter discharge variability. In terms of water quality, dilution of contaminants would be at a minimum during winter and reduced stream velocities under an ice cover would further limit the dispersion of contaminants. These impacts would come at a time when the riverine ecosystem is affected by minimal solar radiation and prolonged freezing temperatures.

The nature of winter streamflow variability is poorly understood because of a principal focus on economic quantities such as maximum flow and total runoff. While new automated methods of determining discharge are being tested under winter conditions, there is currently no technology proven to produce reliable, continuous discharge data through the winter period. During open-water conditions a stage-discharge relation can be established from measurements; and stage data can be collected on a continuous basis to provide continuous discharge data. During the period of ice-affected flow, stage has not proven to be a reliable predictor of discharge (Walker, 1991). Winter discharges are generally determined by interpolation between two or three discharge measurements obtained over a period of time which may be six months or longer. Assumed variability in winter flow due to temperature effects is incorporated using meteorological data to modify the interpolated discharge estimates. An example of the problems that may be created by this methodology is presented in Figure 1.1. The McQuesten River is located in central Yukon, approximately 100 km south of Dawson city. The hydrographs for the two years were produced by two different technologists.

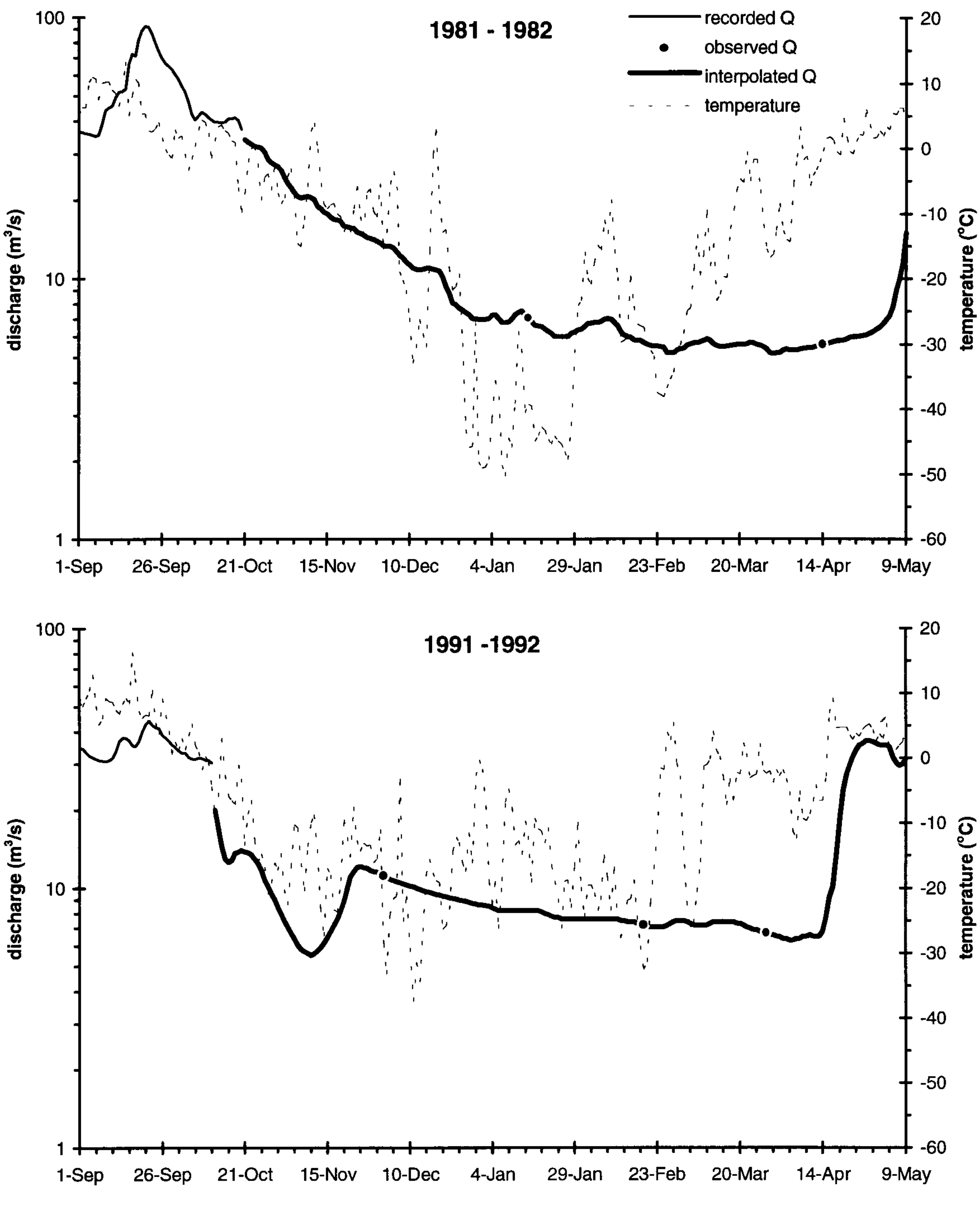


Figure 1.1 Hydrographs for McQuesten River near the mouth for the winters of 1981-82 and 1991-92 showing a difference in interpolation assumptions.

The hydrograph for 1981-82 shows a recession which reflects assumed temperature effects, whereas the hydrograph for 1991-92 demonstrates a smooth recession unaffected by temperature but with an assumed flow depression following freeze-up. Current knowledge of streamflow variability is insufficient to assess which hydrograph more closely resembles the actual discharge.

Although improved temporal resolution of discharge during the winter months would be desirable, economic constraints under which data collection agencies must operate will likely result in a reduction in data collection effort. An alternative to improved data collection is to gain a better understanding of the processes controlling winter streamflow and to use this knowledge to generate better estimates of winter discharge from currently available data.

The next section reviews the processes demonstrated or hypothesized to influence winter streamflow variability. This review provides the basis for the specific objectives of this study.

1.2 CAUSES OF WINTER STREAMFLOW VARIABILITY

Because inputs of rain and meltwater are minimal during winter, depletion of storage from groundwater or surface water reservoirs should dominate winter streamflow variability in the sub-Arctic. There may be one or more reservoirs in each catchment and depletion characteristics may differ among reservoirs. External mechanisms acting on reservoir depletion may cause cumulative or transient effects on winter streamflow. External mechanisms may include temperature effects or mechanical effects such as the frictional effect of an ice cover. These mechanisms may also cause changes in stream-aquifer interactions, disrupting a normal storage depletion process. These processes are reviewed in detail below.

1.2.1 Groundwater storage depletion

Groundwater storage depletion has been reviewed by Hall (1968) and Tallaksen (1995). These reviews show that, despite the theoretical basis for groundwater flow being described over a century ago (Boussinesq, 1877), groundwater storage depletion remains an active area of research. This may be due to difficulties in describing a storage depletion curve for a catchment in which streamflow recession is complicated by factors such as carry-over storage from a prior recharge period, variations in areal pattern of recharge, channel, bank and flood plain storage and evapotranspiration (Nathan and McMahon, 1990). Discharge through a sub-Arctic winter may be a good opportunity to examine storage depletion because the variables identified by Nathon and McMahon should be negligible or non-existent (Whitfield et al., 1995).

Figure 1.2 shows four conceptual models of storage depletion for recession analysis. A recession function for each of the models is also given in Table 1.1 Many variations of these models have been published which are conceptually equivalent to one of the following descriptions.

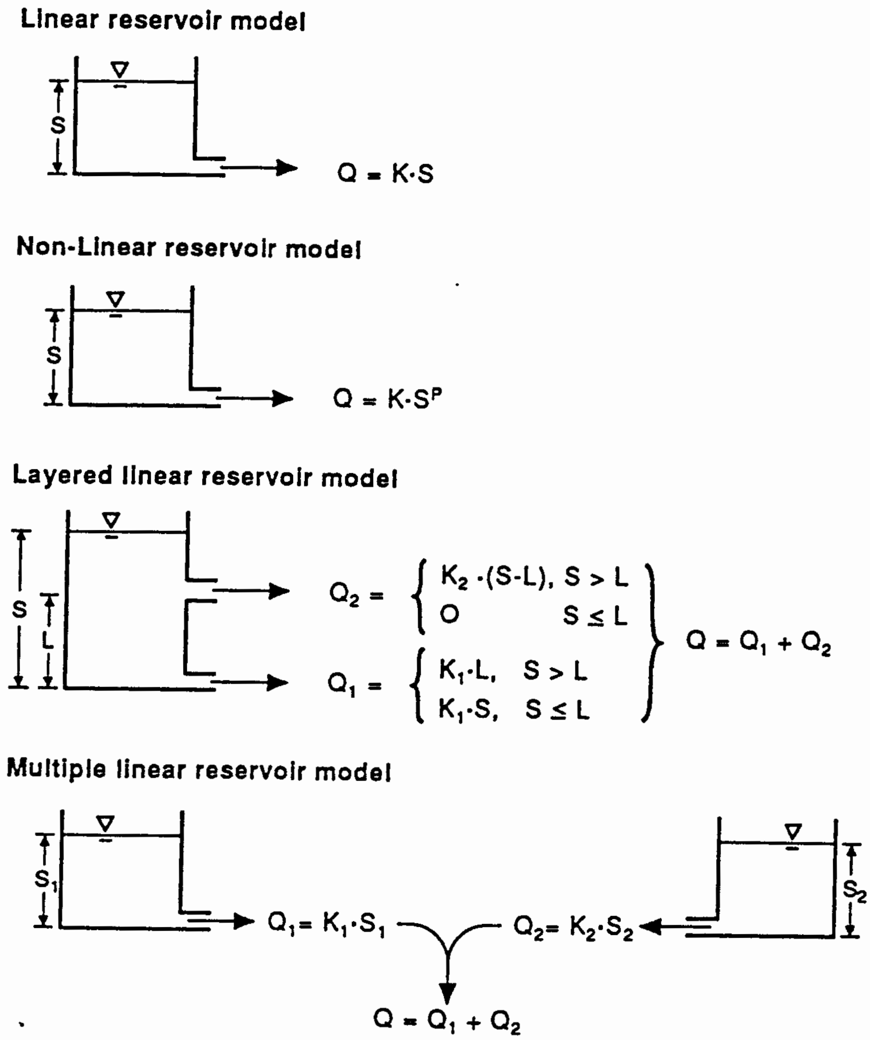


FIGURE 1.2 Four conceptual storage-depletion models.

Table 1.1 Groundwater storage depletion recession models

model	function	equation no.
linear reservoir model	$Q_t = Q_0 \cdot e^{-k \cdot t}$	**(1.1)**
non-linear reservoir model	$Q_t = Q_0 \cdot (1 + \mu \cdot t)^{\beta}$, $\beta = \frac{p}{1-p}$	**(1.2)**
layered linear reservoir model	$Q_t = \begin{cases} Q_0 e^{-k_2 t} & t < t_L \\ Q_0 e^{-k_2 t_L} e^{-k_1 (t - t_L)} & t \geq t_L \end{cases}$	**(1.3)**
multiple linear reservoir model	$Q_t = A \cdot e^{-k_1 t} + B \cdot e^{-k_2 t}$	**(1.4)**

Note: Q_t = discharge at time t,
Q_0 = discharge at t = 0,
μ, β, k_1, k_2, t_L, A and B are parameters

The linear reservoir model (Eq. 1.1) was described by Boussinesq (1877) and remains one of the most widely used models for recession analysis. On a semi-logarithmic plot of discharge against time this relation plots as a straight line. The equation has a strong theoretical basis for groundwater flow for systems where the Depuit assumptions are valid (negligible vertical flow components, and where the effect of capillarity above the water table can be neglected) (Tallaksen, 1995). Model simplicity and estimation of just one model parameter are reasons for the popularity of this model. The model is typically valid for only a limited range of discharge. This limitation has led to the development of the following alternatives.

The non-linear reservoir model (Eq. 1.2) was developed by Boussinesq (1904). The curve plots as a straight line on log-log paper for the variables Q_t and $(1+\mu t)$. This model has been demonstrated to be valid for outflow from unconfined aquifers (Werner and Sundquist, 1951; Hornberger et al., 1970), and for recessions dominated by gravity drainage from soil moisture storage (Hall, 1968).

The layered linear reservoir model (of which Eq. 1.3 is a special case representing two levels) was presented by Bergstrom and Sandberg (1983) to represent drainage from an unconfined aquifer in till with variable porosity. This model plots as intersecting straight line segments on a semi-log plot of discharge against time, where the number of straight line segments represent the number of levels in the reservoir.

Boussinesq (1904) developed the multiple linear reservoir model (of which Eq. 1.4 is a special case representing two reservoirs). The form of this model is non-linear on semi-log paper and may be quite similar in shape to the single non-linear reservoir model. A difficulty in the application of this model is that, because the reservoirs are independent, there is no *a priori* way to determine the relative state of storage in each reservoir at the start of any recession period.

Various combinations of these four models in series (representing cascading flow through several reservoirs) or in parallel (representing independent reservoirs) can accommodate almost any reservoir structure. However, model selection should adhere to the principle of parsimony. Complex models with more than two or three calibrated parameters are difficult to justify given the uncertainty in establishing a storage depletion curve through a wide range of discharge.

1.2.2 Lake storage depletion

Lake discharge can be expressed in terms of the following water balance equation:

$$Q_0 = Q_i - \frac{\Delta S}{\Delta t} \qquad (1.5)$$

where Q_o is discharge at the lake outlet, Q_i is inflow into the lake, S is lake storage and t is time. Lake storage is a function of the water level and surface area of the lake. For sub-Arctic lakes Q_i typically exceeds Q_o during the snowmelt period, resulting in an increase in storage. During the frozen season the reverse is true, with a depletion of storage which would augment winter low flows compared to a similar catchment without lake storage.

Lake discharge is typically determined from a non-linear relation between water level (stage) and discharge as shown in Figure 1.3, and expressed in the form:

$$Q=\beta\cdot(H-H_0)^{\alpha} \qquad \textbf{(1.6)}$$

where α and β are constants, H is stage and H_0 defines the lower limit for the equation. That is, a lake behaves like a non-linear reservoir.

Ice could affect lake discharge by increasing frictional resistance to flow at the lake outlet (Gerard, 1990). Lake outlets are typically ice-free year-round in the southern Yukon and it is has been observed that the presence of these polynyi generally prevents ice from affecting lake discharge (Chin, 1966).

Snowfall on floating lake ice would be expected to have an effect on lake discharge by displacing a volume of water approximately equal to the water equivalent in the snow pack (Gerard, 1990). Likewise during the fall, prior to the development of lake ice cover, snow falling on the lake is immediately available for flow, whereas snow falling on land is not available until the snowmelt season. Hence, precipitation onto the lake can be included in the inflow term in the lake's water balance.

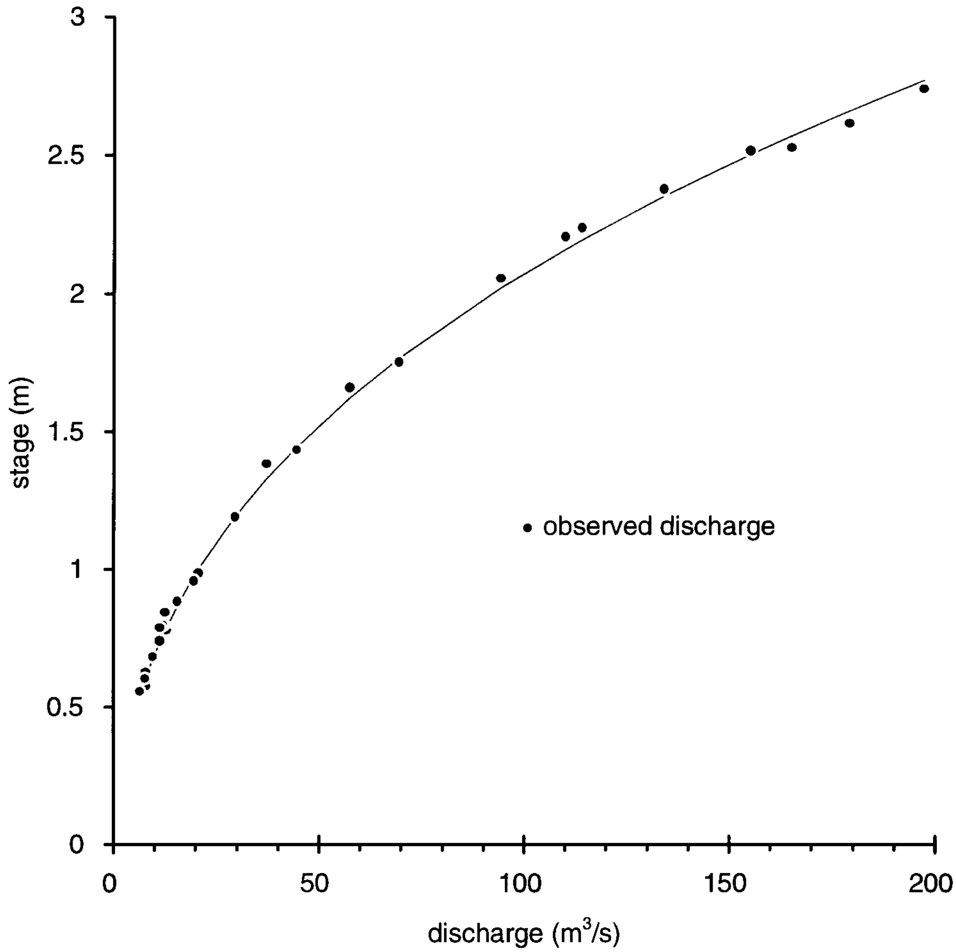


Figure 1.3 Stage-discharge curve for Takhini River at the outlet of Kusawa Lake.

1.2.3 Temperature effects

The assumption of a temperature effect on winter discharge during periods of sub-freezing weather is controversial, and has never been conclusively tested. Most literature references which acknowledge a temperature effect cite Chin (1966). However, the magnitude of the effect observed by Chin was in the order of ± 10%. Though the uncertainty of a discharge measurement under an ice cover has never been explicitly quantified, the error will be greater than for open water discharge measurements (Pelletier, 1989; Pelletier, 1990). It is reasonable to assume that the uncertainty of a measurement under an ice cover may be at least ± 10% given that the uncertainty of an open water discharge measurement under ideal conditions at the 95% confidence level may be as high as ± 6% (Pelletier, 1988).

The need for further research to resolve the issue of a temperature effect has been ignored until this time. Both WSC and the U.S. Geological Survey (USGS) have both adopted the assumption of a temperature effect and air temperature is used as a variable in most currently available methods of estimating winter discharge (Melcher and Walker, 1992). This assumption requires a simple relation between temperature and discharge in order for the effect to be practically useful in predicting streamflow.

That temperatures above 0°C can result in flow increases due to snowmelt is not in dispute. What is uncertain is whether streamflow can vary when temperatures remain below freezing. Even if there is a temperature effect, there is no agreement on the mechanism by which changes in air temperature would cause a change in discharge. Chin (1966) proposed a dynamic relation between frost depth and the water table in discharge zones by which an increase in temperature would result in a thaw-back of the frost horizon, allowing more groundwater discharge into the channel. Alternative hypotheses include in-channel effects and micro-climatic effects on snowmelt runoff. In-channel effects may be due to water moving into and out of channel storage in response to temperature-related changes in the hydraulic resistance of the ice cover and cross-sectional area available for flow (e.g., Wankiewicz, 1984). Micro-climatic effects would be a result of snowmelt runoff occurring during periods when the bulk air temperature remains below freezing but when conditions in sun-exposed, near-channel, micro-climates are conducive to snowmelt.

It is also uncertain how quickly sub-Arctic streams respond to above-freezing temperatures. Snowmelt water may be stored in the snowpack, in frozen or unfrozen ground under the snowpack or in the channel before being released as discharge. This transient storage of meltwater would affect discharge response to periods of above-freezing weather.

1.2.4 Transient storage

The loss of discharge to channel storage at freeze-up is conceptually feasible (Gerard, 1990; Beltaos et al., 1993; Burn, 1993) but is not widely accepted. For example, USGS hydrologists dispute trans-boundary Yukon River data estimated by WSC technologists which show a discharge depression at freeze-up. Discharge depression at freeze-up can be a result of an increase in resistance to flow as an ice cover develops (Alford and Carmack, 1987a; 1987b; 1988). Beltaos et al. (1993) showed that a stage-up increase of 32% of channel depth is predicted by the Manning equation for the increase in hydraulic resistance caused by an ice cover. Water is abstracted from discharge to occupy the increase in cross sectional area caused by stage-up. Inflow from tributaries to a channel may also be reduced due to backwater effects resulting from stage-up. As the stage rises in the main channel, the slope of the lower reach of a tributary would be reduced with a consequent flow reduction. Further work is required to test the theory of freeze-up discharge depression against field data.

1.2.5 Stream-aquifer interactions

The potential loss of water to bank storage in response to stage-up has been hypothesized (Chin, 1966) but not tested by field studies. The actual loss of water may be compounded by hydraulic damming of groundwater inflow by a reversal of the hydraulic gradient at the streambed. Figure 1.4 illustrates stream-aquifer relations before and after freeze-up. Prior to freeze-up there is a positive hydraulic gradient across the streambed with groundwater flow into the channel. Following freeze-up, channel stage is higher than the hydraulic head immediately adjacent to the channel, causing a reversal of hydraulic gradient across the streambed. This reversal of hydraulic gradient would not only result in loss of water from the channel, but would also block groundwater inflow.

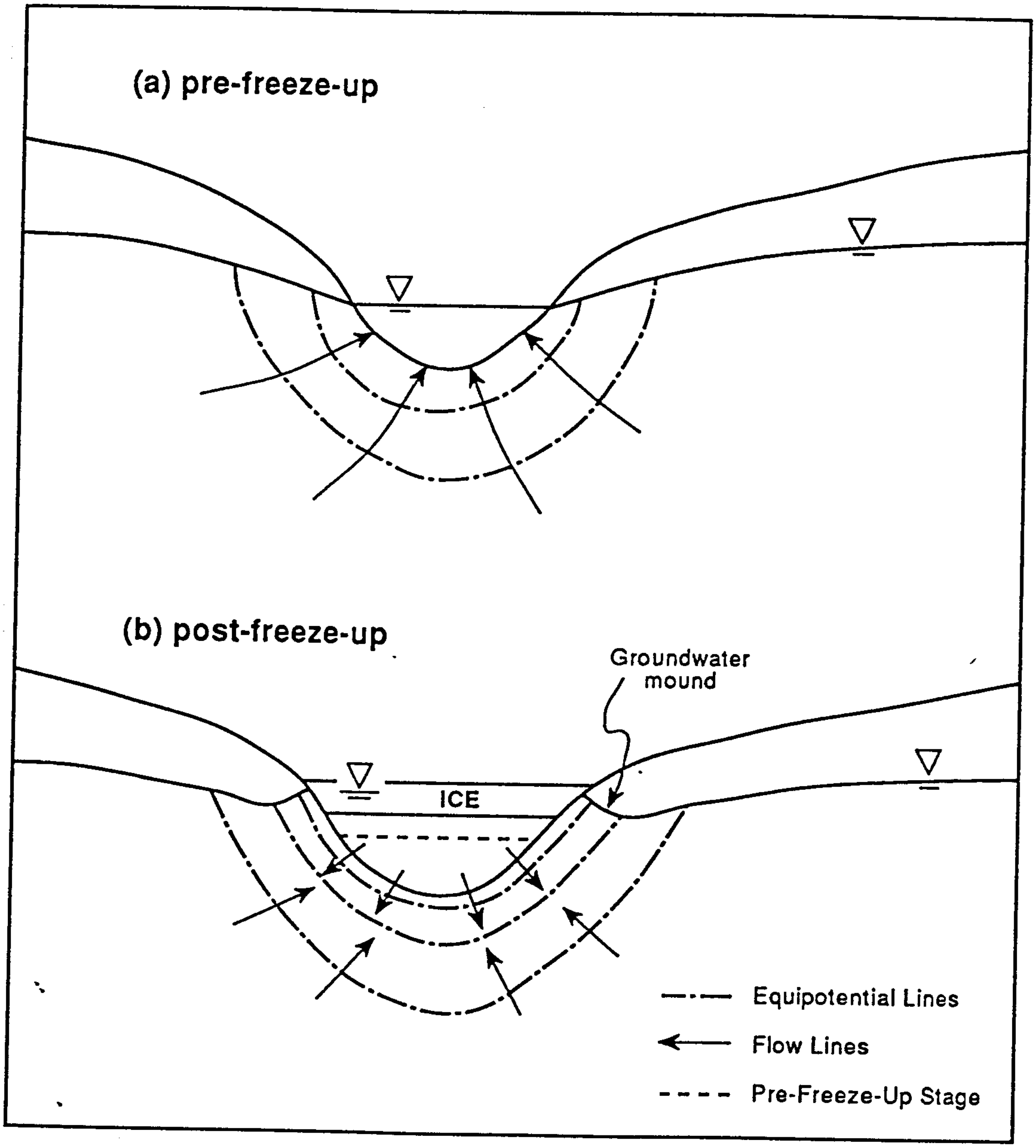


FIGURE 1.4 Sketch illustrating groundwater mounding process

1.3 OBJECTIVES

Existing hydrometric data do not allow for a complete evaluation of the variability in winter discharge. The objective of this thesis was to conduct a field program on three sub-Arctic streams with different catchment characteristics to develop a data set which can be used to evaluate variability in winter streamflow and to identify the processes producing that variability. Four research questions will be addressed in this thesis.

(1) What is the nature of the groundwater storage depletion process?

To answer this question the four proposed storage depletion models (Figure 1.2) will be tested against discharge from two groundwater-dominated streams to determine which conceptual model best represents observed discharge. Water quality data will provide additional evidence for the structure of the proposed storage reservoir(s).

(2) Is there a discharge depression following freeze-up?

- *(a) Can channel storage account for flow depression?*
- *(b) Is there evidence of stream-aquifer interaction?*

Question two can be answered by hydrograph inspection. Part (a) will be answered by reconciling changes in discharge volume with changes in channel storage volume based on estimates of channel length, width and depth. The answer to part (b) may be inferred from the answer to (a), with supplementary evidence provided by observations of near-stream groundwater response. A comparison of measured flow at Takhini River near Whitehorse and the sum of measured upstream flow will be examined for evidence of stage effects on groundwater inflow into a channel. Water quality variables will be used as an indication of groundwater contributions to flow.

(3) Is there a temperature effect on discharge?

Conclusive evidence for a practically significant temperature effect on discharge requires that the effect exceed measurement uncertainty. For the purpose of this study a nominal uncertainty of ± 10% will be assumed for winter discharge measurements. Time series plots of discharge and temperature as well as plots of the residuals from the storage depletion models against temperature will be used to determine whether a correspondence between temperature and discharge variability exists. An attempt will be made to find evidence of a mechanism which could cause temperature effects on discharge.

Assuming that cold ice is less plastic than warm ice and that changes in the plasticity of the ice may affect hydraulic resistance to flow, a correlation should exist between ice temperature and stage. If an increase in plasticity (warm ice condition) reduces hydraulic resistance water should be released from channel storage with an increase in discharge and a decrease in stage. Cold ice would have the opposite effect.

If the frost depth in shallow groundwater discharge zones responds to air temperature, then a correlation should be seen between air temperature and frost depth and discharge. If micro-climatic snowmelt is an important factor, then snowmelt at sun-exposed, near channel sites should correspond with positive residuals from the discharge recession trend.

(4) What factors influence streamflow variability below a lake?

(a) What are the relative roles of lake inflow versus storage depletion, and how do these change through time? In particular, how important is lake storage depletion at the time of seasonal low flow?

(b) What is the effect of snowfall on lake discharge?

(c) Is there evidence of any other atmospheric effects on lake discharge?

A water balance approach using lake stage and discharge at the outlet will be used to estimate lake inflows. Time series plots of lake storage depletion obtained from the difference between estimated inflows and outflows will be used to show the relative importance of inflow and storage depletion. The volume of water displaced by snowfall events will be compared to changes in discharge at the lake outlet. Discharge variability from the lake will be examined for probable causes.

The remainder of this thesis is devoted to developing and answering these research questions. Chapter two describes the physiography and climate of the study area and outlines the methodology used in the collection of data. Chapter three presents the results. Chapter four provides interpretation and discussion of the results, and chapter five summarizes the main conclusions and makes recommendations for future research.

CHAPTER TWO

METHODS

This chapter describes the biophysical characteristics of the Yukon territory in general, and the study catchments in detail. It then presents the methods of data collection and analysis. Finally, the climatic and hydrologic conditions antecedent to and during the study period are described as an aid to placing this study in a longer term context.

2.1 CLIMATE AND PHYSIOGRAPHY OF THE YUKON

The Yukon climate is classified as sub-Arctic and Arctic-continental and is characterized by large annual variation in temperature, low relative humidity and low precipitation (Janowicz, 1991). Mean annual temperature ranges from near 0°C in the south to -10°C in the north. Monthly mean temperature ranges from +16°C to 0°C for July and from -15°C to -35°C for January. Mean annual precipitation ranges from less than 200 mm to 700 mm, though portions of the St. Elias Mountains may receive amounts up to 3500 mm (Janowicz, 1991).

The Yukon Territory is located between 60° and 70°N. Most of the Territory is within the Cordilleran region with physiography ranging from interior rolling uplands to the rugged Coast and St. Elias Mountains. Figure 2.1 shows the location of the study area.

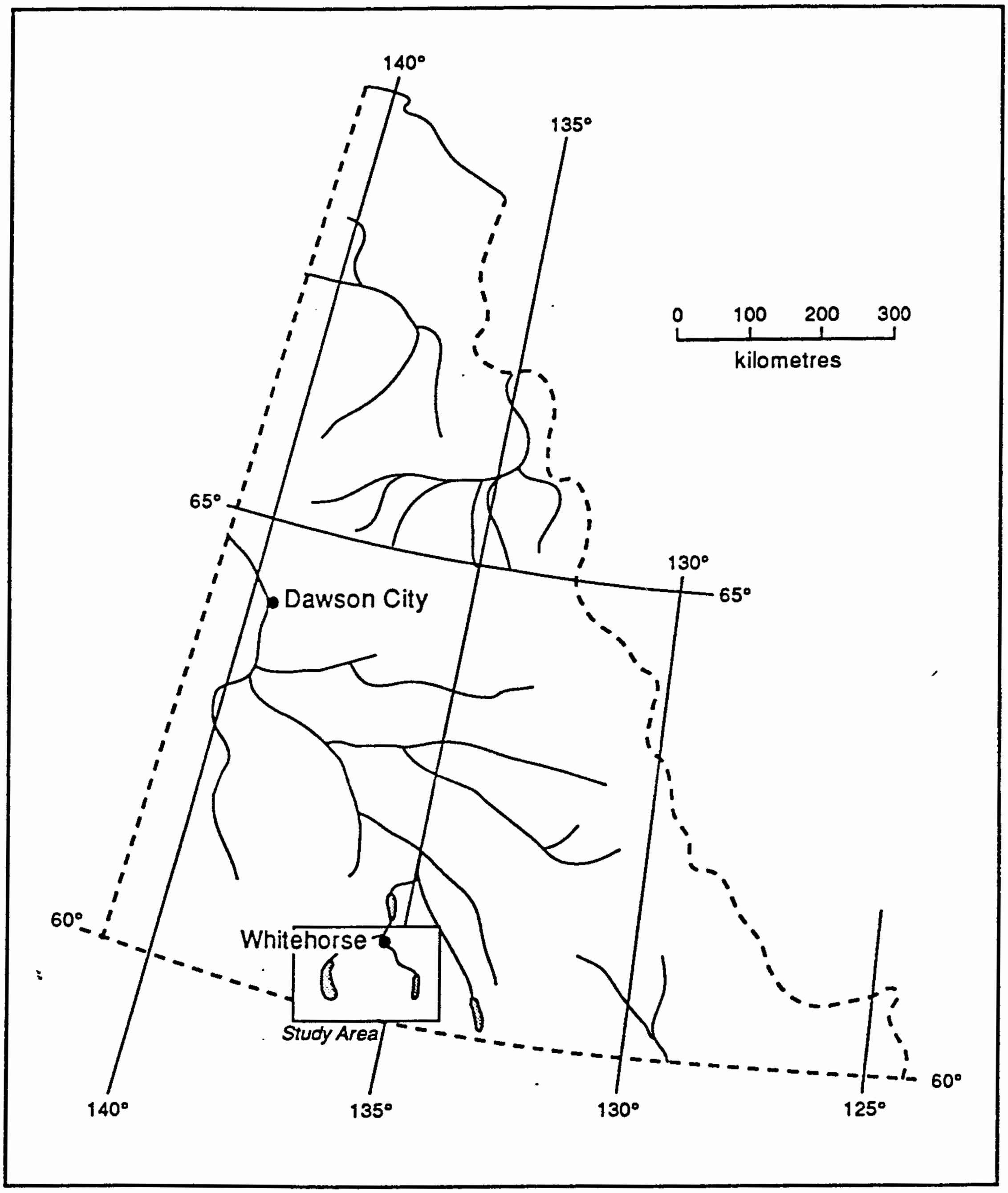


FIGURE 2.1 Map of Yukon Territory showing study area

2.2 FIELD SITES

The study area is located in southern Yukon near Whitehorse, Yukon Territory (Figure 2.1). The climate is predominantly continental, with coastal influences in the upper Takhini River basin. Temperatures as high as 34°C and as low as -52°C have been recorded at the Whitehorse airport with a mean annual temperature of -3°C. During the period 1961 to 1990, Whitehorse airport received an average of 269 mm of total precipitation per year.

Soils in the study area are a sequence of glacial, glaciofluvial and glaciolacustrine deposits that are typical of glaciation and deglaciation in mountainous terrain. Ice is estimated to have reached elevations of 1825 to 1975 m during the McConnell glaciation (Smith and Mouget, 1995), covering all but the highest peaks in the area. Glacial moraines in the area are composed of till which has sandy loam to sandy clay loam texture with a high percentage of angular cobbles and boulders. Glaciofluvial features such as eskers, kames and kame terraces consisting of stratified poorly sorted gravel and sandy gravel are evident in the Ibex and Takhini River valleys. Fine grained silt and clay glaciolacustrine sediments up to 60 m thick are found in the study area. These originate from Glacial Lake Champagne, a large proglacial lake that formed in the latter stages of deglaciation about 9,000 years ago (Smith and Mouget, 1995). Fluvial and eolian processes continue to transport and modify these soils of glacial origin. Fluvial deposits vary from silt to sand in texture and have been redeposited in low terraces, abandoned meanders and oxbow lakes. Localized loess deposits are found throughout the study area.

Vegetation in the study area is dominated by pine and spruce boreal forest. Discontinuous permafrost is found in the study area. Above the treeline at about 1200 m, sub-alpine shrubs such as Salix sp. and Betula sp. diminish to alpine tundra at elevations greater than about 1500 m.

Some characteristics of the study catchments are summarized in Table 2.1. The terrain is mountainous with a range in elevation of approximately 1900 m from valley bottoms to the highest peaks.

2.2.1 M'Clintock River

M'Clintock River is located southeast of Whitehorse, Yukon Territory. The WSC station (09AB008) has been in operation since 1955. The station is located 7 km upstream from Marsh lake in the Yukon river system (Figure 2.2). Length of existing record, easy access and a good winter metering section were primary considerations in selecting this station. Data collected at this site included stage, discharge, hydraulic head of near-stream groundwater and soil temperatures (Figure 2.3).

2.2.2 Takhini River

The Takhini River is a tributary of the Yukon River and is located to the west of Whitehorse, Yukon Territory (Figure 2.2). The Takhini drains from the Coast

Table 2.1 Physiographic characteristics of study catchments

	M'Clintock River near Whitehorse	Ibex River near Whitehorse	Takhini River near Whitehorse	Takhini River at outlet of Kusawa Lake
Drainage area (km^2)	1700	457	6990	4070
Station elevation (m)	655	850	655	670
Lake area (km^2)	8	0	197	184
Maximum elevation (m)	2084	1908	2515	2515
Main channel length (km)	40	25	150	120
Main channel slope (%)	0.24	1.72	0.38	0.46

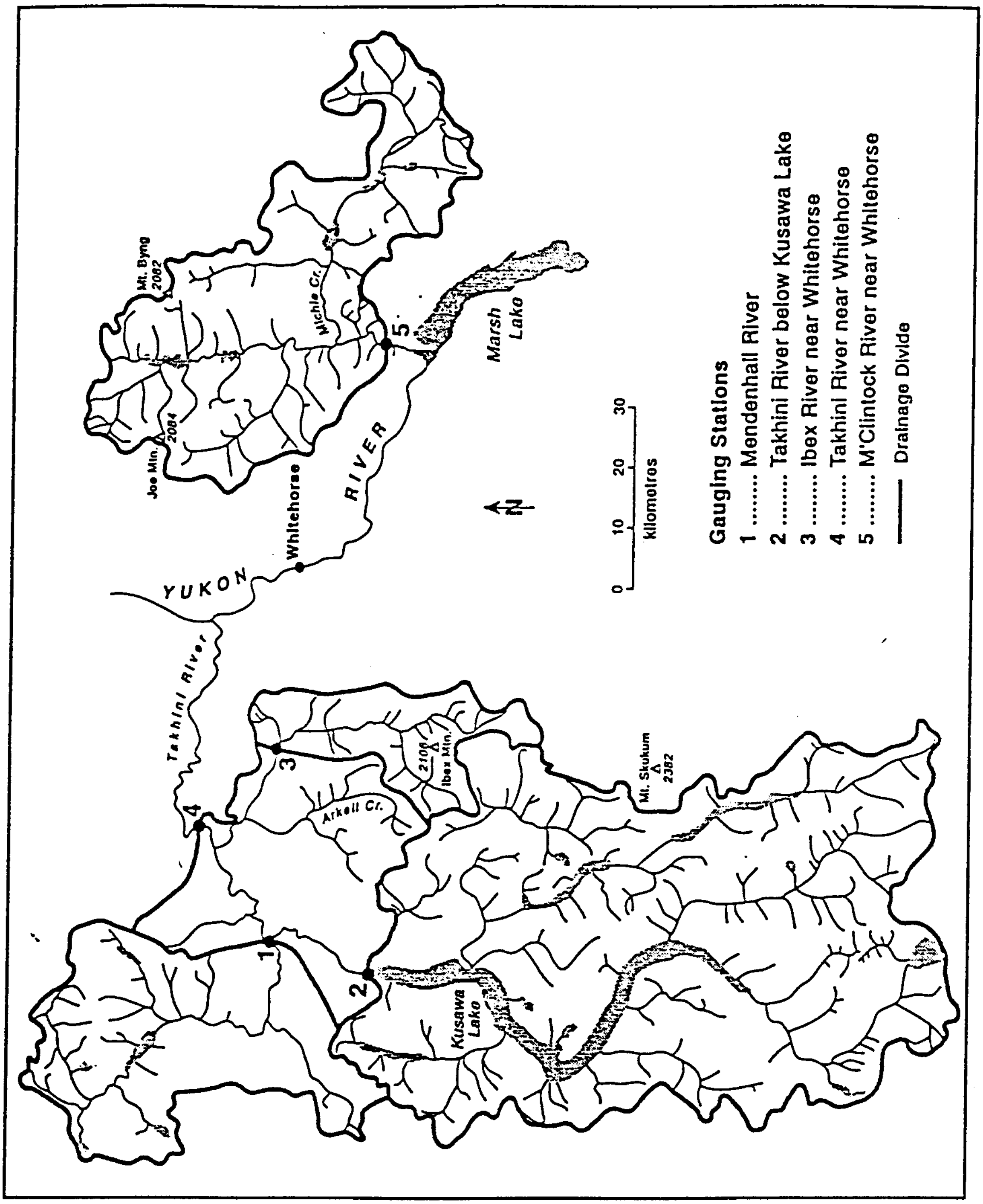


FIGURE 2.2 Map of Whitehorse area showing study catchments

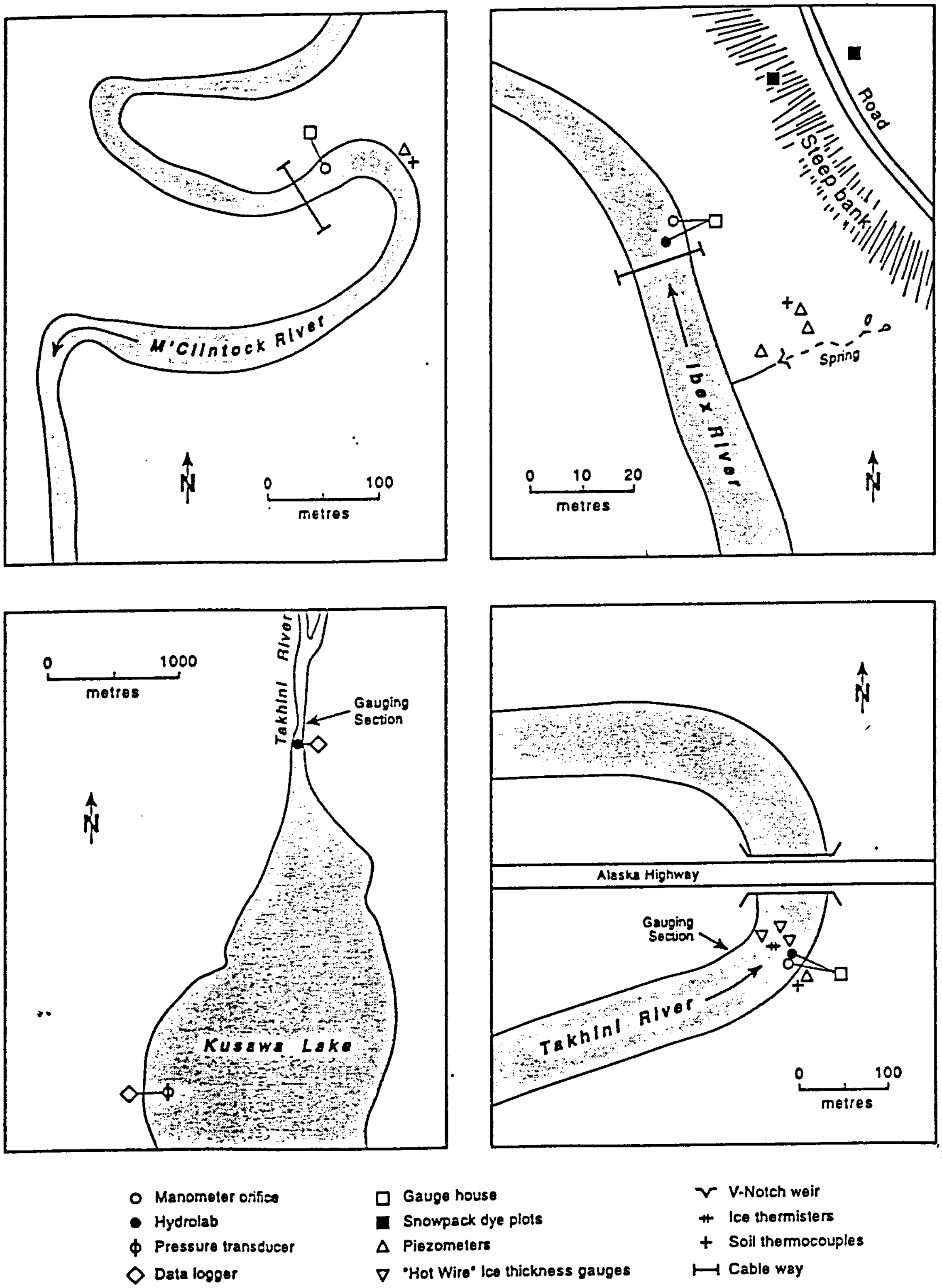


FIGURE 2.3 Sketches showing the four study sites

Mountains in the south and from the Yukon Plateau in the north. Data were collected from four sites in the Takhini River Basin.

2.2.2.1 Takhini River Near Whitehorse

This WSC Station (09AC001) has been in operation since 1948. The station is located at the Alaska Highway bridge km 1523, 40 km downstream from the outlet of Kusawa Lake. This is the same station studied by Chin (1966). The long period of record, easy access and good winter metering conditions were primary considerations in selecting this station. Data collected at this site include stage, discharge, hydraulic head of near-stream groundwater, soil temperatures, ice temperatures, water quality and ice thickness (Figure 2.3).

2.2.2.2 Kusawa Lake

This WSC Station (09AC004) was in operation from 1952 to 1986. The location is at the outlet of Kusawa Lake. Isolating the effects of lake storage on winter discharge was the primary reason for selecting this station. Data collected from this station include lake stage, discharge and water quality (Figure 2.3).

2.2.2.3 Ibex River

This WSC Station (09AC007) has been in operation from 1989 to the present. The station is located 20 km above the confluence with the Takhini River. The primary reason for selecting this station was to isolate groundwater discharge response. There is no significant surface water storage reservoir in this catchment. Data collected at this site included stage, discharge, hydraulic head, groundwater spring discharge, soil temperatures, and water quality (Figure 2.3).

2.2.2.4 Mendenhall River

This site has no history of winter data collection apart from measurements made for the Chin study of 1966. Only discharge data were collected at this site. The purpose of this site was to improve streamflow accounting for balancing inputs to the Takhini River near Whitehorse. This site is sensitive to backwater effects from the Takhini River.

2.3 FIELD MEASUREMENTS

The following descriptions summarize the techniques and equipment used for data collection. Brand names are used only for informational purposes.

2.3.1 Discharge

Stream discharge measurements were made using standards, procedures and equipment supplied by WSC. The following brief descriptions are derived from WSC field and training manuals (e.g. Terzi, 1981).

The general method used for determining discharge is the area-velocity method. A minimum of 20 measurements of depth and velocity across the cross-section were used to determine the total discharge. The instrument used to determine velocity is the Price current meter, which is suspended in the stream by the following sounding and positioning methods.

2.3.1.1 Bridge

Measurements up to and through the freeze-up period at the Takhini River near Whitehorse were made from the Alaska highway bridge. A two-wheel crane was used as a platform for a sounding reel from which the current meter was suspended above a sounding weight. Cross sectional position in the stream was obtained by wheeling the crane across the bridge a measured distance. Depth was measured by lowering the sounding weight from the surface to the stream bed on a steel sounding line. The sounding weight also suspends the current meter at the appropriate depth for the determination of velocity.

2.3.1.2 Wading

Wading measurements were used for the Ibex River, and for late season measurements at Takhini River at the outlet of Kusawa. For this method a nylon or kevlar tagline was stretched across the section for cross-sectional position. Sounding and positioning of the current meter was accomplished using a top-setting wading rod.

2.3.1.3 Cableway

Cableway measurements were used for the pre-freeze-up determination of discharge at the M'Clintock River. For this method a permanent steel cable spanning the river was used for positioning a cablecar to which a sounding reel was attached. The cableway is a part of the permanent installation operated and maintained by WSC at this station. Cross-sectional positioning was obtained from marks on the cable. Sounding and current meter positioning were accomplished using a sounding weight suspended from the sounding reel.

2.3.1.4 Boat

A boat was used for measurements at the outlet of Kusawa Lake. Cross-sectional positioning was obtained from either a steel or a kevlar tagline which was also used to secure the boat in position in the current. Sounding and meter positioning was accomplished using a sounding weight suspended from a sounding reel secured to the boat.

2.3.1.5 Ice

Discharge measurements from an ice cover were used for the Takhini, Mendenhall and M'Clintock Rivers throughout the frozen season. For this method holes were drilled through the ice and the current meter lowered into the water attached to sounding rods which were used to determine ice thickness and depth as well as to position the meter. Cross-sectional positioning was obtained from a nylon tagline stretched across the ice.

2.3.2 Stage

River and lake stage were monitored continuously through the study period for all stations except Mendenhall River. The following sections describe the methods of measurement and recording of stage.

2.3.2.1 Bench marks

Vertical control for river stage was obtained with reference to WSC bench marks. A survey level run was used to obtain water level with each visit. For M'Clintock, Takhini and Kusawa two bench marks were tied in with each level run. For Ibex only one submerged bench mark was used to determine water level and that bench mark was tied in to three other bench marks at the start and end of the study period.

2.3.2.2 Sensors

Mercury manometers were used for continuous measurement of stage at M'Clintock, Takhini and Ibex Rivers. These instruments work on the principle of displacement of fluid in a U-tube to measure change in pressure. In order that the U-tube is a manageable size mercury is used in these instruments. Mercury has 14 times the density of water so the U-tube only needs to be 1/14th as high as if it were filled with water. One end of the tube is open to the atmosphere while the other end is connected to a tube which extends to an orifice fastened to the bottom of the river. A source of compressed air is attached to the orifice line and bubbles continuously purge out of the orifice. The pressure in the line is then regulated by the height of the column of water over the orifice. This pressure displaces mercury in the U-tube which activates a servo-control mechanism which adjusts the pressurized end of the U-tube until the height of mercury is equal on both sides of the U-tube. These adjustments are mechanically converted to shaft rotations which can be recorded either digitally on a datalogger or on an analog chart recorder. A Stevens pressure transducer was used at Kusawa Lake. A Tavis pressure transducer was used for the groundwater spring at Ibex.

2.3.2.3 Loggers

Stevens analog chart recorders were used for back-up data collection and visual inspection of manometer data at Takhini and Ibex rivers. Valcom (VEDAS) data loggers were used at Takhini, Ibex and Kusawa, a Stevens Multilogger was used at Kusawa, a Stevens analog chart recorder was used at M'Clintock River.

2.3.2.4 Stage-discharge relation

A stage-discharge relation was used to convert recorded stage to discharge. The relation was obtained by plotting measured open-water stage against measured discharge (e.g. Figure 1.3). WSC standards require that if measurements do not plot within 5% of the stage-discharge curve a shift correction be used.

Two types of corrections can be applied in processing stage and discharge data. Gauge corrections are applied to account for deviation in the sensor values from measured stage. Gauge corrections may be abrupt in the case of a movement of the sensor, or distributed over time in the case of instrument drift. Shift corrections are adjustments to the stage-discharge relation to accommodate temporary changes in the stage-discharge relation such as occur due to the effects of an ice cover.

Gauge corrections were used at M'Clintock and Ibex rivers and at Kusawa Lake to correct for orifice movements. Shift corrections were used to correct for the backwater effects of ice for producing continuous discharge data at M'Clintock, Ibex and Takhini rivers and at Kusawa Lake. The shifts were determined based on each discharge measurement and linearly distributed over the interval between measurements. An assumption is required that a linear interpolation of shift correction between measurements is valid for ice affected flow. Several variations of this method are widely used for estimating winter flow (Melcher and Walker, 1992).

2.3.3 Groundwater measurements

Hydraulic head was measured in the near-stream area of Takhini, Ibex and M'Clintock stations with nests of piezometers and water table wells. No instruments were used for recording these data and so data are only available when visits to the station were made and up until frost penetrated to the water table, freezing the water in the wells and piezometers.

2.3.3.1 Construction of piezometers

Investigation of near-channel groundwater processes through a sub-arctic winter has not received much attention in the past. I was reluctant to use antifreeze substances in the piezometers (e.g. glycol or diesel) for environmental as well as scientific concerns. The primary environmental concern was the release of foreign liquids into the soils. Scientific concerns are related to the difficulty in interpreting the results when the medium used for measurement has a different density and viscosity than the medium being measured (water). For those reasons the piezometers were installed knowing that it was unlikely that they would all survive the winter without freezing.

The piezometers were constructed of 38 mm I.D. PVC plastic pipe. Holes (6 mm) were drilled in the bottom 15 cm of the pipe, and screened using nylon mesh. Water table wells were constructed of the same materials but holes were drilled along the full length of pipe. Installation was in a 60 cm diameter hole dug with a bucket auger. Coarse sand was packed around the screened area and a layer of Bentonite was used to seal the piezometers, followed by backfilling with parent material.

Protection against early freezing was provided by thermal insulation of the tops of the piezometers with collars and caps of building insulation cut to fit over the ends of the pipes. Plugs of insulation were also pushed into the piezometers with the intent of additional protection from freezing.

2.3.3.2 Placement of piezometers

At M'Clintock River the piezometers were placed in a silt bank on the outside of a bend 100 m downstream of the gauging station (Figure 2.3). This location is at the bottom of a long hill with evidence of saturated soils making this a likely groundwater discharge zone. Three piezometers were installed at the water's edge and three more installed 2 m inshore. The soil was fine silt. The depth of installation was from 0.65 m to 1.3 m below ground surface.

At Ibex River piezometers were installed in the proximity of a groundwater spring (Figure 2.3). The soil was mixed gravels, stones and silt. The depth of installation varied from 0.4 m to 1.0 m below ground surface.

At Takhini River the piezometers were installed in silty soil at the gauging site. Three were installed at the water's edge and three more installed 1.5 m inshore (Figure 2.3). The depth of installation varied from 0.7 m to 1.1 m below ground surface.

2.3.3.3 Measurements

Piezometer measurements were made by observing resistance across two electrical leads attached to a sounding rod. The distance from the top rim of the piezometer to the water surface was taken when the electrical resistance abruptly dropped indicating submergence of the leads. Vertical control was obtained by survey level from WSC benchmarks to the top rim of each piezometer.

2.3.3.4 Spring discharge

A standard design plate metal 90° v-notch weir was installed in an attempt to measure groundwater spring discharge at the Ibex River site. The standard rating for this design could not be used because that rating assumes zero approach velocity, which could not be obtained in the gradient of this spring.

Stage behind the weir was obtained from a gauge plate attached to the weir and continuous stage recorded using a Tavis pressure transducer and bubbler system. Volumetric measurements were used to calibrate the weir. These measurements were obtained by timing the filling of a container of a known volume with water spilling over the weir. However, the recorded stage record was of little value due to ice forming on the weir and affecting the weir rating. For this reason only discharges from actual volumetric measurements were used for interpretation.

2.3.4 Soil temperatures

Soil temperature probes were positioned at 0.3 m above the ground, at ground level, at 0.1 m increments to 0.5 m, and at 1.0 m below the ground surface. At Takhini and M'Clintock thermocouples were read at each visit using a hand held digital display. At Ibex River thermistors recorded temperature every three hours on the VEDAS datalogger. The soil temperature profiles were located adjacent to a water table well at each location (Figure 2.3) to provide a reference between soil temperature and the water table elevation. Vertical control was by survey level to nearby WSC benchmarks for referencing elevations to gauge datum.

Calibration of all thermistors and thermocouples was done with reference to a calibrated Atmospheric Environment Service (AES) mercury thermometer in a water bath, prior to installation. Accuracies of ± 0.2° C were obtained and no sensor corrections were applied for this study.

Frost depth was calculated as the depth at which soil temperature equalled 0° C, determined by linear interpolation between sensors. While recognizing that a linear model may be a simplistic representation of the temperature profile between thermistors, this simplicity is consistent with the objectives of this part of the investigation.

2.3.5 River ice

The distance from the bottom of the ice to the water surface in a hole cut in the ice (float depth) and total thickness of the ice were recorded for each of the 20 or more holes drilled for discharge measurements. A different cross section must be used for each discharge measurement because of the effect trampling the snow and drilling through the ice has on the subsequent ice development. The interpretation of ice thickness data from discharge measurements must therefore take into consideration the effect of sampling a different location each time.

2.3.5.1 Temperature profiles

Ice temperatures were recorded at a location approximately at mid-flow on the Takhini River near Whitehorse (Figure 2.3). Data were recorded on a three-hour time interval on the VEDAS datalogger. Thermistors and thermistor leads for ice temperature measurements were positioned inside a piece of 38 mm I.D. PVC pipe for protection and for ease of retrieval. The PVC pipe flooded with water on installation and froze into place with the thermistors in direct contact with the ice inside the pipe. The plastic pipe and the thermistor leads likely affected the recorded temperatures due to differences in thermal conductivity relative to an undisturbed ice cover. However, given that the primary objective was to observe temperature trends, this effect on absolute accuracy was ignored. The seven thermistors were located at 0.5 m above the ice surface, at the ice surface, and at 0.15 m increments down to 0.75 m below the ice surface. Calibration prior to installation was conducted in a water bath using an AES calibrated mercury thermometer for reference. Observed temperatures were within ± 0.2°C and no sensor corrections were applied.

2.3.5.2 Ice thickness

At the Takhini River near Whitehorse three 'Hot Wire' ice thickness stations were established (Figure 2.3). Ice thickness was recorded at each visit to the station. The measuring device, described by Sherstone et al. (1986), uses an electric current to heat a wire frozen into the ice. The 'hot wire' can then be drawn through the intact ice cover until a weight strikes the under side of the ice. Ice thickness is derived from the length of wire that can be withdrawn less the known length of the wire.

2.3.6 Water quality

Variability in chemical composition of the water was monitored in order to assess whether changes in flow sources could be verified independently of the discharge data. Water quality parameters included both physical and chemical measurements.

2.3.6.1 Sample collection and processing

Water samples were collected using Environment Canada protocols (Environment Surveys Branch, 1991). Open water sampling techniques were used at Takhini River near Whitehorse prior to November 21, at Takhini River at outlet of Kusawa Lake and at Ibex River near Whitehorse. Through-ice sampling technique was used at Takhini River near Whitehorse after November 21. Analysis of the samples was performed at the Environment Canada Conservation and Protection Laboratory (Vancouver B.C.).

2.3.6.2 Hydrolabs

Automatic recording of several water quality variables was accomplished using Hydrolabs. These instruments can be deployed and left in the stream for extended periods. Power to the Hydrolab and data from the Hydrolab are transferred via armoured Serial Data Interface (SDI) cable which connects the Hydrolab to the datalogger on shore. These instruments can be operated in a stand-alone configuration with internal batteries and data logging; however, for this project I elected to have the power, logging, and sample schedule controlled by the datalogger so that the instrument would not need to be retrieved to change any of the above.

Variables measured by the hydrolab for this project included water temperature, pH, specific conductance, total dissolved solids (TDS), dissolved oxygen, turbidity and oxygen-reduction potential (Redox). Hydrolabs were installed at Takhini River, Ibex River and at the outlet of Kusawa Lake. The Hydrolabs were controlled by the VEDAS datalogger and programmed to sample every 3 hours.

The Hydrolabs were calibrated to standards in a laboratory prior to deployment after which they were not touched for the duration of the study period. A fourth Hydrolab was frequently re-calibrated and rotated amongst stations for operation in parallel with the primary Hydrolabs in order to evaluate instrument drift.

2.3.6.3 Environmental isotopes

Weekly samples were collected for isotopic analysis at Ibex river throughout the winter and at M'Clintock river from late winter through to break-up. Analyses were carried out at the University of Waterloo. The results of the analysis are expressed relative to the concentration of the isotope in Standard Mean Ocean Water (SMOW) in units per mil according to the formula

$$\delta^2 H \text{ or } \delta^{18} O = (1000 \text{ per mil}) \cdot [C_{sample} - C_{SMOW}] / C_{SMOW}) \quad \textbf{(2.1)}$$

where C is concentration. One sigma error is ±0.15 per mil for $^{18}0$ and ±3.0 per mil for ^2H. Negative values result from the fact that stream and groundwater samples are depleted in heavier isotopes relative to ocean water because of the fractionation which occurs when sea water evaporates (i.e. lighter isotopes tend to evaporate at a higher rate than heavier isotopes).

The value of isotopic analyses for this study is in the relative stability of isotopic concentration within a population of water. In contrast, water chemistry can be altered in relatively short time scales through contact with the parent material in the aquifer and will vary with contact time and flow path. Therefore, isotopically homogenous water can be considered to originate from a single source and variations in water chemistry within that population can be considered to be due to changes in flow path. Comparisons between isotopic results and water quality results were used to reconcile conceptual reservoir structure with recession modelling results.

2.4 METEOROLOGICAL VARIABLES

Daily temperature and precipitation data were obtained from the Atmospheric Environment Service (AES) station at the Whitehorse airport. This station is probably representative of the climate in the lower elevations of the study catchments.

Cloud cover, wind speed and direction, barometric pressure and temperature were recorded at each station each visit. Cloud cover and wind speed and direction were estimated, temperature was observed on a hand held thermometer and barometric pressure observed on a pocket altimeter.

2.5 SNOWMELT OBSERVATIONS

In late winter (Feb 13) water soluble paint powder was spread over the snow surface at two locations at the Ibex River (Figure 2.3). The paint powder was subsequently covered by snowfall. The two locations were chosen to evaluate micro-climatic effects on snowmelt. One location was on a steep exposed bank near the river and the other location was a sheltered level location 50 m further inland. Throughout the spring, snow pits were dug at each of these sites and photographs taken of the dye layer to identify the time when meltwater started percolating through the snowpack.

2.6 PHOTOGRAPHY

Flights over the catchments were conducted during the freeze-up period. Photographs of stream conditions were taken during these flights to document the status of freeze-up upstream of the study sections.

Photographs were taken at each site in an upstream and a downstream direction to document changes in the ice cover. At Takhini at outlet of Kusawa Lake and Ibex River, photographs were taken every visit. At Takhini and M'Clintock Rivers photographs were taken only during the freeze-up season until a stable ice cover developed.

2.7 DATA ANALYSIS

2.7.1 General strategy

An important assumption underlying the data analysis is that streamflow over the winter should be dominated by storage depletion, which should yield a concave-up trend. The effects of channel storage and stream-aquifer interactions should produce a dip below the recession trend, while temperature effects should produce variations about the trend line. For the M'Clintock and Ibex Rivers, which are dominantly groundwater-fed during the winter, the analysis proceeded as follows:

a. Visually inspect the hydrograph and identify whether a significant dip occurred at and following freeze-up. If so, locate the time at which flow recovered to the recession trend. It is acknowledged that identification of the time of flow recovery is subjective. However, consideration of groundwater response and stage data may prove useful in this task.

b. Fit storage-depletion models to the data, excluding data during a post-freeze-up depression, as well as data at the end of the winter which may be influenced by snowmelt runoff. Evaluate storage depletion models in terms of fit to the observed data, and in terms of water quality trends.

c. Estimate the volume of water involved in the discharge depression (V_d) by integrating the difference between the predicted recession trend and the measured flow.

d. Compare the volume of water involved in the discharge depression to an estimate of the increase in channel storage associated with the stage increase at freeze-up. If the increase in channel storage is significantly less than V_d then it can be inferred that stream-aquifer interactions caused at least part of the discharge depression.

e. Plot residuals from the recession trend, following recovery, against temperature and evaluate the correlation of the plot to determine whether a simple temperature effect exists.

For the Takhini River at the outlet of Kusawa Lake the analysis proceeded as follows:

a. Solve for the storage depletion contribution to discharge on the date of each discharge measurement. Storage-depletion discharge Q_d is calculated as

$$Q_d = \frac{H_{t-10} - H_{t+10} \cdot A}{d \cdot s} \quad \textbf{(2.2)}$$

where H is stage, t is the time of the measurement (days), A is lake area (m^2), d is the length of the period (21 days) and s is the number of seconds in a day (86,400).

b. Solve for inflow (Q_i) at the date of each measurement as

$$Q_i = Q_o - Q_d \quad \textbf{(2.3)}$$

where Q_o is measured outflow discharge.

c. Fit a storage depletion model to calculated inflow.

d. Plot inflow, outflow and storage depletion as time series for visual analysis.

e. Calculate the volume of water displaced by snowfall events. Calculate the increase in discharge associated with an increase in stage due to snowfall. Compare these volumes to the variability in lake outlet discharge.

f. Identify transient discharge depressions and verify these events with discharge from the Takhini River near Whitehorse and with water quality data. Relate these events to observed changes in the ice cover at the lake outlet.

For the Takhini River near Whitehorse the analysis proceeded as follows:

a. Sum known upstream discharges (Ibex River, Mendenhall River and Takhini River at outlet of Kusawa Lake).

b. Compare measured flow at Takhini River near Whitehorse to the sum of known upstream flow.

c. Examine deviations between measured flow and the sum of known upstream flow for evidence of local storage and release effects. If local effects are evident, relate these effects to either channel storage (as indicated by a corresponding stage response, backwater effects at the Mendenhall River, and perhaps by changes in the ice cover) or to stream-aquifer interactions (as indicated by piezometric response and water quality parameters).

d. Plot time series of ice thickness and temperature for comparison with stage and discharge to examine for effects of temperature-related changes in the ice cover on transient channel storage and release.

2.7.2 Model fitting

Parameters in the recession models were estimated by use of a non-linear iterative algorithm as implemented in a commercial statistical package (Systat). The loss function to be minimized was a sum of squared relative errors (SSRE), calculated as

$$SSRE=\sum_{i=1}^{n}[(Q_p-Q_a)/Q_a]^2 \quad \textbf{(2.4)}$$

where Q_a is observed flow on date i, Q_p is predicted flow on date i, and n is the number of data points used for fitting the model. The fractional residuals were used in the loss function, rather than the raw residuals, in order to avoid biasing the model to fitting the higher discharges. The use of fractional residuals is also consistent with the notion that errors in discharge measurements are a fraction of actual value.

Three-week time steps were used for the solution of the water balance equation (1.5). The need for this coarse time resolution is to average out errors in stage measurement. Small errors in stage become greatly exaggerated when multiplied by lake area to determine change in volume.

2.8 OVERVIEW OF STUDY PERIOD

The results of this study can be put in perspective by comparison of this year with the period of record. This context is useful in establishing the broader relevance of the findings of this study.

2.8.1 Meteorological record

Monthly mean temperatures, extremes and total precipitation for the study period are compared to climate normals for the period from 1961 to 1990 in Table 2.2. The following description is intended to provide some interpretation for the data in Table 2.2.

The summer of 1994 was warm and dry in May, July and August with cool and wet weather in June. The fall was mild but wet and snow did not stay on the ground until October 26. November weather was cold with high northwest winds (average 17.2 km/hr) and over double the normal snowfall (51.2 cm compared to the normal of 25.5 cm). In December precipitation was light, average temperature was near normal and the prevailing south winds were strong (average 17.7 km/hr). An El Nino event in the Pacific caused a southerly flow which resulted in above normal January temperatures. Bright sunshine was prevalent in February, resulting in diurnal temperature fluctuations reaching 20°C. March was generally cloudy, cold and snowy. April was warm, dry and sunny. April temperatures were well above normal but not above records set in 1993 and 1994. Accumulated snowfall is slightly above normal at

Table 2.2 Comparison of 1994 1995 monthly means to climate normals (1961-1990) at Whitehorse airport

Month	Temperature (°C)				Total Precipitation (mm)			
	94/95	Normal	Max	Min	94/95	Normal	Max	Min
May	7.6	6.6	9.5	4.4	2.7	14.4	38.4	0.8
June	12.4	11.6	16	9.3	44.5	31.2	90.9	6.0
July	15.5	14.0	17	12.1	29.4	38.5	110	5.6
August	16.8	12.3	17	8.4	11.7	39.3	103	1.3
September	6.7	7.3	11.1	4.7	44.4	35.2	67.0	6.8
October	1.7	0.7	3.7	-3.3	27.5	23.0	50.6	1.0
November	-13	-10.0	-1	-20	33.1	18.9	51.8	2.0
December	-15	-15.9	-4	-28	6.9	18.9	42.4	3.0
January	-15	-18.7	-4.0	-33	9.1	16.9	46.0	1.6
February	-12.0	-13.1	-3	-28.0	15.0	11.9	37.1	0.9
March	-10	-7.2	-2	-15	22.3	12.1	43.9	0.8
April	3.3	0.3	3.5	-5.1	1.6	8.3	37.9	TR

TR = Trace amount

146.7 cm (normal 141.5 cm). Average temperature during the study period (November to April) was -10.4° C, fairly close to the normal of -10.8° C.

Time series plots of precipitation and temperature for the study period are presented in Figure 2.4. The abscissa scale is common to all time series data plots to assist in comparisons.

2.8.2 Hydrometric record

Monthly mean discharge for the study period is compared to monthly mean discharge for the period of record in Table 2.3. No data are presented for Kusawa Lake for the period May to September because the datalogger and transducer were not installed until September 16. The summer prior to the study period was characterized by slightly lower than normal discharges. M'Clintock River flows for the period May to October were 12% less than average, Takhini River flows were 1.7% less than average and Ibex River flows were 33% less than average. Flow picked up with the fall rains and were higher than average in October, flows for the period November 1994 to April 1995 were 25% greater than average at M'Clintock River, 20% greater than average at Takhini River, and 24% greater than average at Takhini River at the outlet of Kusawa Lake. However, Ibex River flows were 8% less than average. The annual means were -2% of average for M'Clintock River, +1.6% of average for Takhini River and -29% of average for Ibex River. A relatively short period of record (1989 to 1993) may account for the apparent differences in Ibex River comparisons.

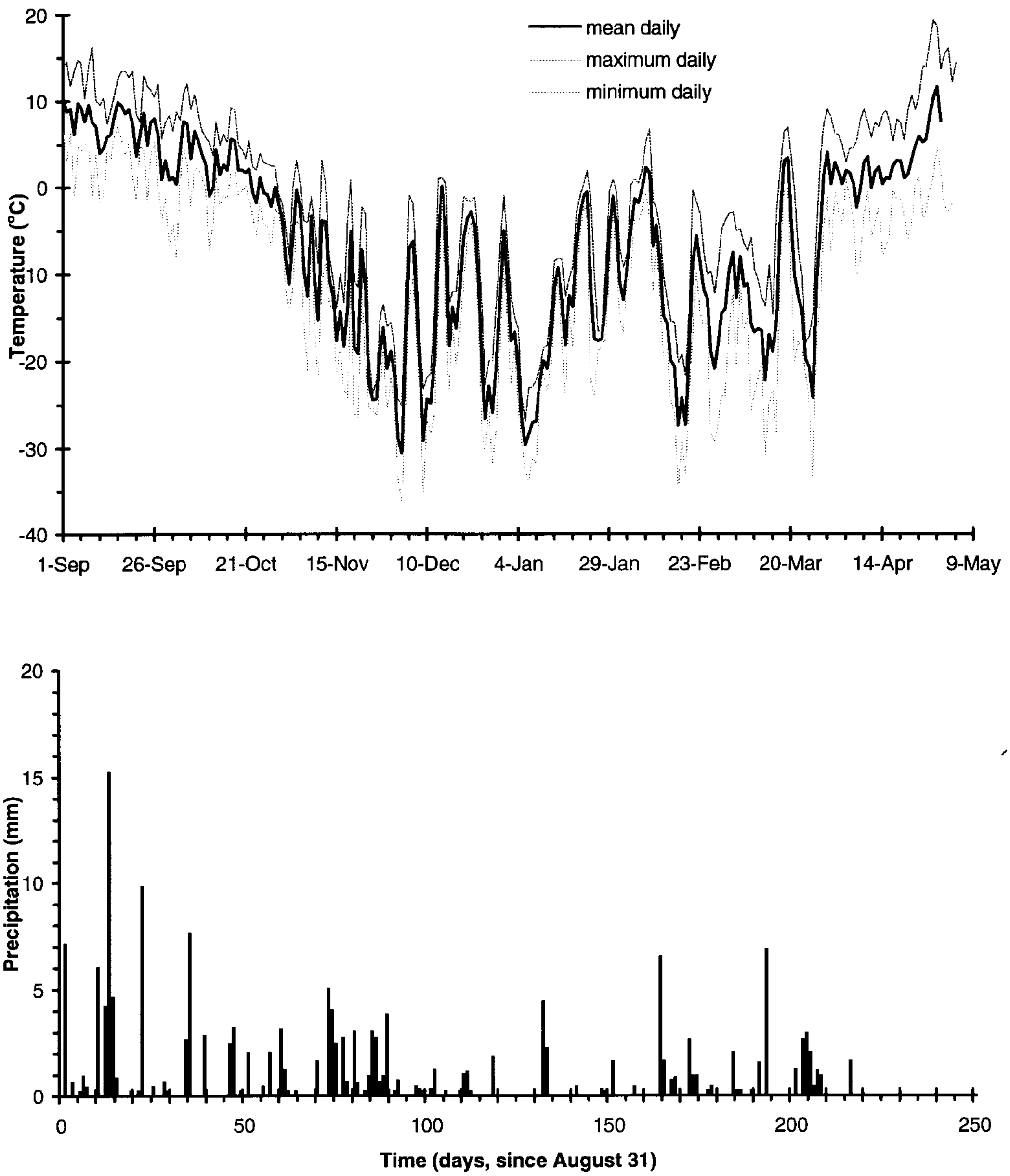


Figure 2.4 Whitehorse airport daily temperatures and precipitation

Table 2.3 Comparison of 1994 - 1995 discharge to period of record

Month	M'Clintock River		Takhini River		Ibex River		Kusawa Lake	
	94/95	Mean	94/95	Mean	94/95	Mean	94/95	Mean
				(m^3/s)				
May	18.6	17.9	23.5	22.2	1.46	2.86	...	...
June	22.2	27.2	95.1	102	4.19	6.61	...	...
July	12.4	15.4	159	184	3.02	5.68	...	...
August	7.31	11.2	161	162	2.27	3.47	...	...
September	8.75	11.1	108	110	2.92	3.66	...	...
October	12.2	9.67	87.8	65.1	2.44	2.18	71.3	53.9
November	6.37	5.76	43.9	34.3	1.26	1.15	33.6	29.9
December	5.12	4.02	21.8	20.3	0.835	0.833	18.8	17.4
January	4.24	3.22	15.4	13.9	0.584	0.670	14.4	11.3
February	3.66	2.89	13.2	11.6	0.485	0.512	12.9	8.87
March	3.40	2.80	11.6	10.3	0.404	0.452	9.61	7.81
April	4.87	3.42	15.7	11.1	0.469	0.790	8.83	7.16
Mean	9.09	9.28	63.0	62.0	1.69	2.400	24.2	19.5

M'Clintock River near Whitehorse mean is for period 1955 to 1993

Ibex River near Whitehorse mean is for period 1989 to 1993

Takhini River at the outlet of Kusawa Lake mean is for period 1952 to 1986

Takhini River near Whitehorse mean is for period 1948 to 1993

CHAPTER THREE

RESULTS

This chapter presents the data collected for this investigation starting with groundwater-related processes as represented by M'Clintock and Ibex Rivers in sections 3.1 and 3.2. The data from Kusawa Lake (section 3.3) pertain to the effect of lake storage on winter streamflow and the section on the Takhini River near Whitehorse (section 3.4) is relevant to the effect of in-channel processes on ice-affected flow. The final section in this chapter compiles the most important results and identifies the key issues that will be addressed in chapter four.

3.1 M'CLINTOCK RIVER

3.1.1 Overview

The development of ice cover was monitored by aerial reconnaissance with flights on November 2, November 6 and November 14. Ice cover at the station was first observed November 2 and by November 14 approximately 90% of the stream surface was ice covered. The lower reaches of the river were the first to freeze followed by tributaries without lake storage. Michie Creek, a tributary fed by Michie Lakes, was the last reach to freeze and was partially open on the last overflight November 14.

Observed discharge (solid circles in Figure 3.1) does not always coincide with recorded discharge (solid line) because the recorded discharge represents daily

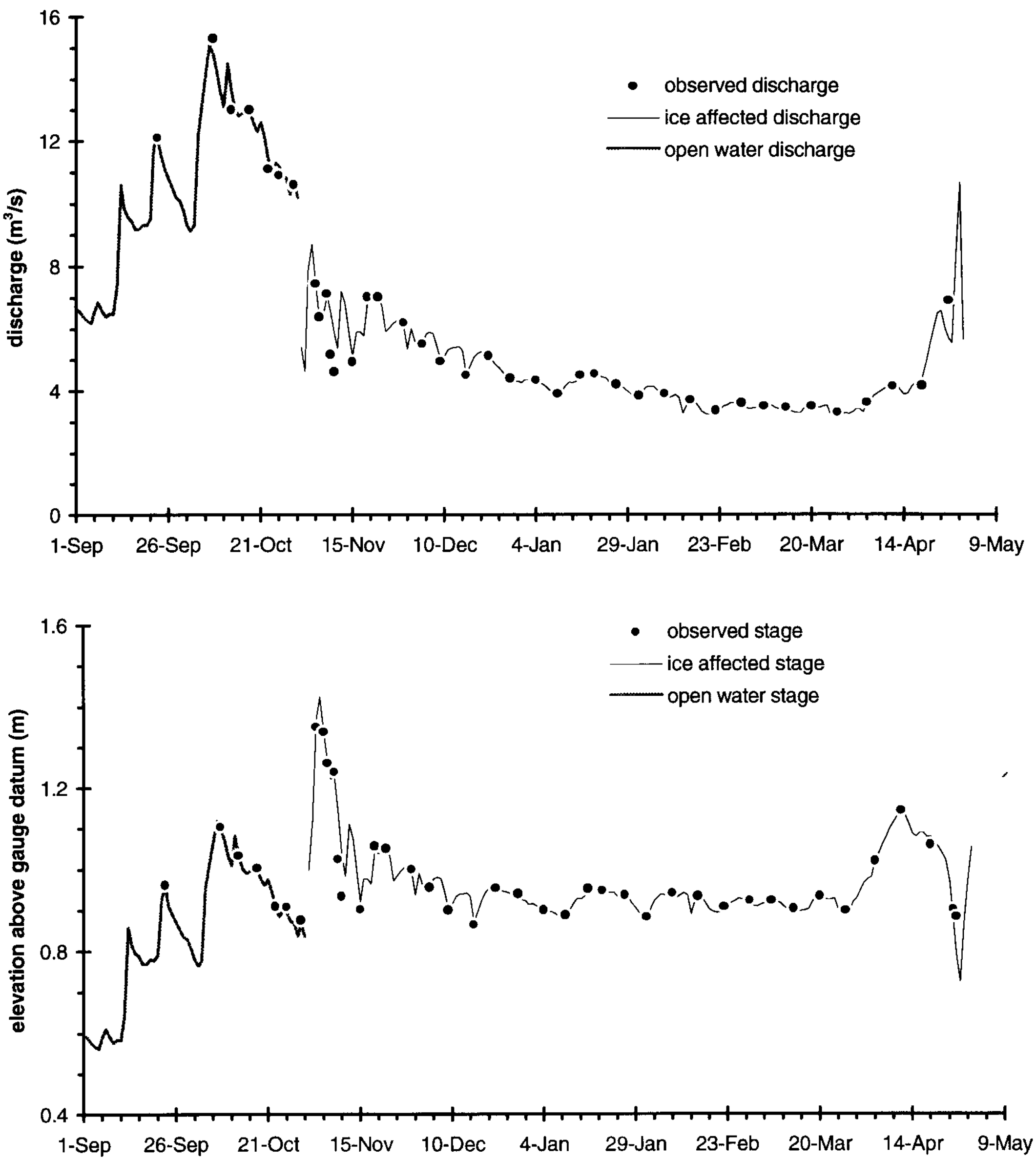


Figure 3.1 M'Clintock River near Whitehorse, stage and discharge.

averages which, during periods of changing discharge, will be different from the discharge measured within a short time span. Because the river was typically measured in the morning it is possible that measured discharges are close to the daily minimum if flows are correlated with air temperature on a daily time scale. Slush ice was present in the measurement section during the freeze-up period and occupied up to 30% of the cross-sectional area. After November 22, the section was free of slush. Detailed observations of slush and channel geometry are reported in Appendix A.

Increase in discharge after April 4 is likely due to snowmelt response. The timing of this increase is consistent with increased flow at Ibex and Mendenhall Rivers and with snowmelt observations at Ibex River. The increase in stage between March 27 and April 11 may be due to an increase in hydraulic resistance of the ice cover caused by thermal decay (Carmack and Alford, 1987a; 1987b; 1988).

3.1.2 Storage depletion

The four recession models (Figure 1.2) were applied using discharge on the last day of open water as Q_0. Measurements prior to November 20 or after April 4 were not used for model calibration to exclude effects due to freeze-up processes or snowmelt. Only measured discharge was used for model calibration. However, results are presented including recorded discharge to provide a more complete picture of discharge variability. Model results plotted in arithmetic scale (Figure 3.2a) show differences in absolute error and the semi-logarithmic scale (Figure 3.2b) shows the

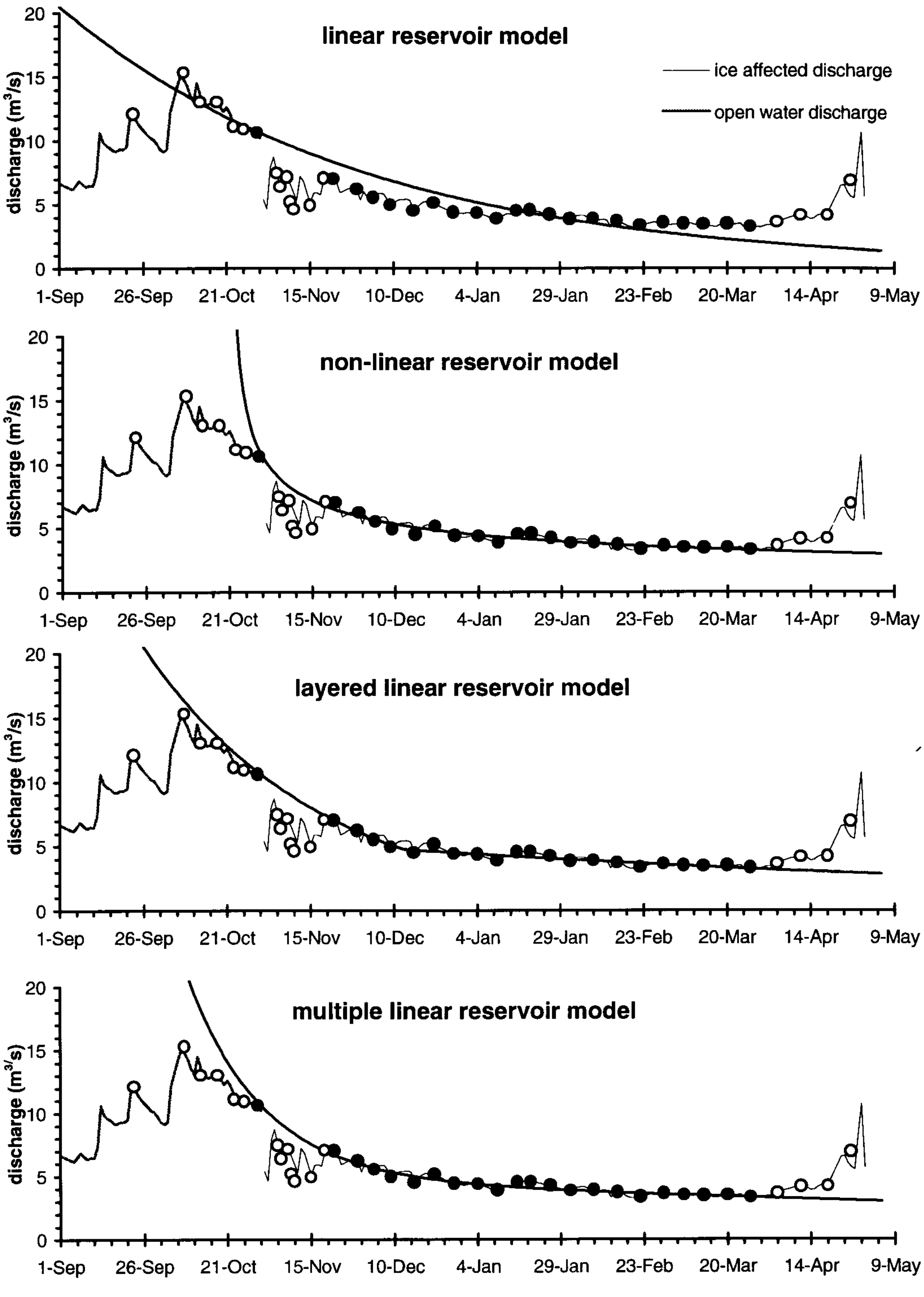


Figure 3.2a M'Clintock River near Whitehorse, recession model results plotted on arithmetic scale. Solid circles indicate measurements used for model calibration.

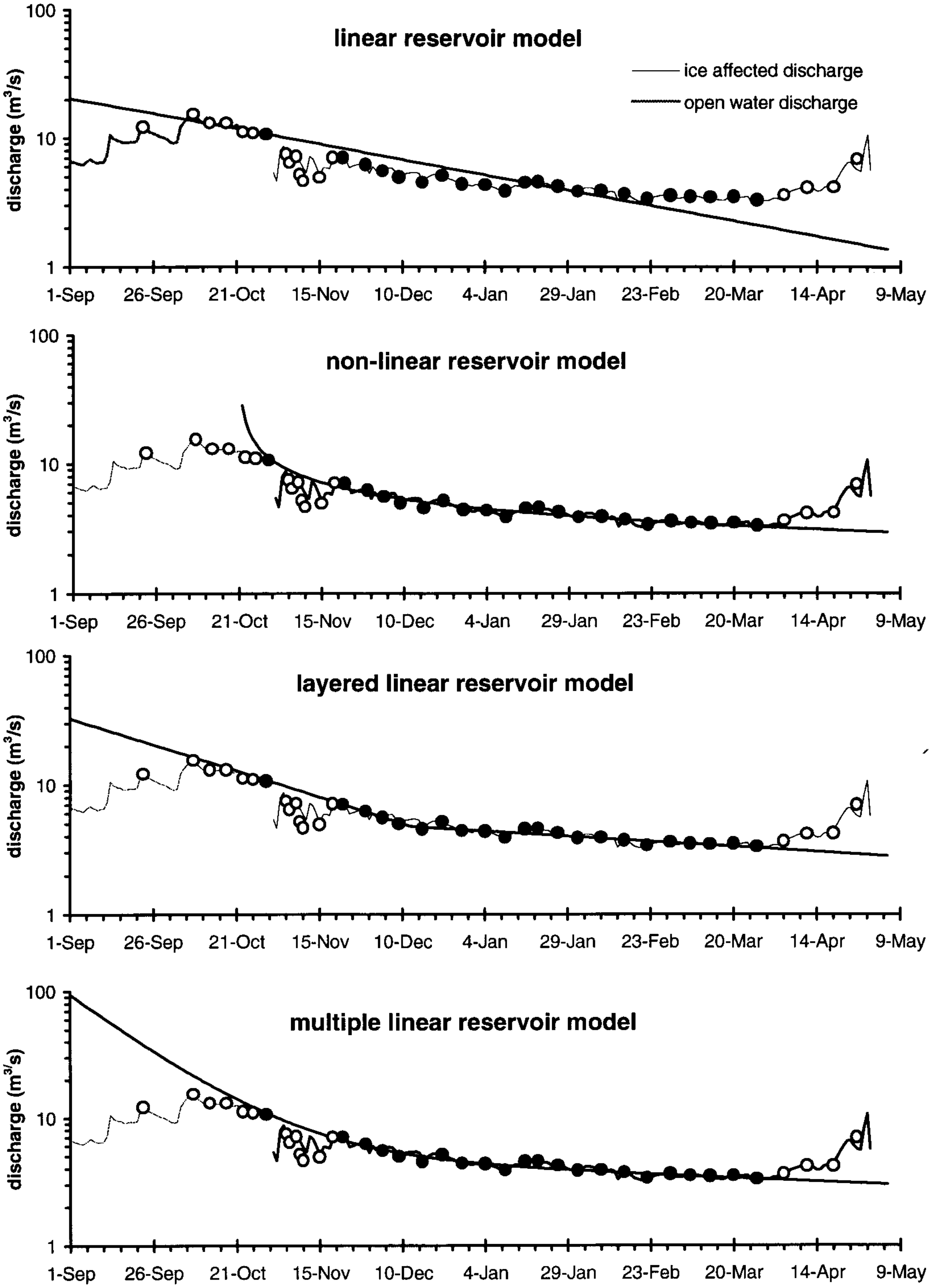


Figure 3.2b M'Clintock River near Whitehorse, recession model results plotted on logarithmic scale. Solid circles indicate measurements used for calibration.

difference in linearity of the models.

All four models over-estimate discharge for the freeze-up period (November 1 to November 20) and under-estimate for the period after April 4 (Figure 3.3). The linear reservoir model (Eq. 1.1) produces good results for the uncalibrated period prior to freeze-up but produces negative residuals for the first half of the winter and positive residuals for last half of the winter. These results indicate that Equation 1.1 is an inappropriate model for winter discharge. The pattern of residuals for the remaining three models are similar for the calibration period (November 20 to April 4) but differ in the fit to uncalibrated data prior to freeze-up. The non-linear reservoir model (Eq. 1.2) residuals are off of the scale for most of the pre-freeze-up period. The layered linear reservoir model (Eq. 1.3) provides good continuity across the freeze-up period. The multiple linear reservoir model (Eq. 1.4) residuals are less than -10% for most of the pre-freeze-up period. The residuals for the calibration period (November 20 to April 4) are within nominal measurement error; however, the distribution of residuals appears to be auto-correlated.

The sum of relative squared error (Table 3.1) shows that the layered linear reservoir model (Eq. 1.3) is slightly better than the multiple linear reservoir model (Eq. 1.4) and the single non-linear reservoir model (Eq. 1.2) and substantially better than the linear reservoir model (Eq. 1.1) in estimates of discharge for the calibration period.

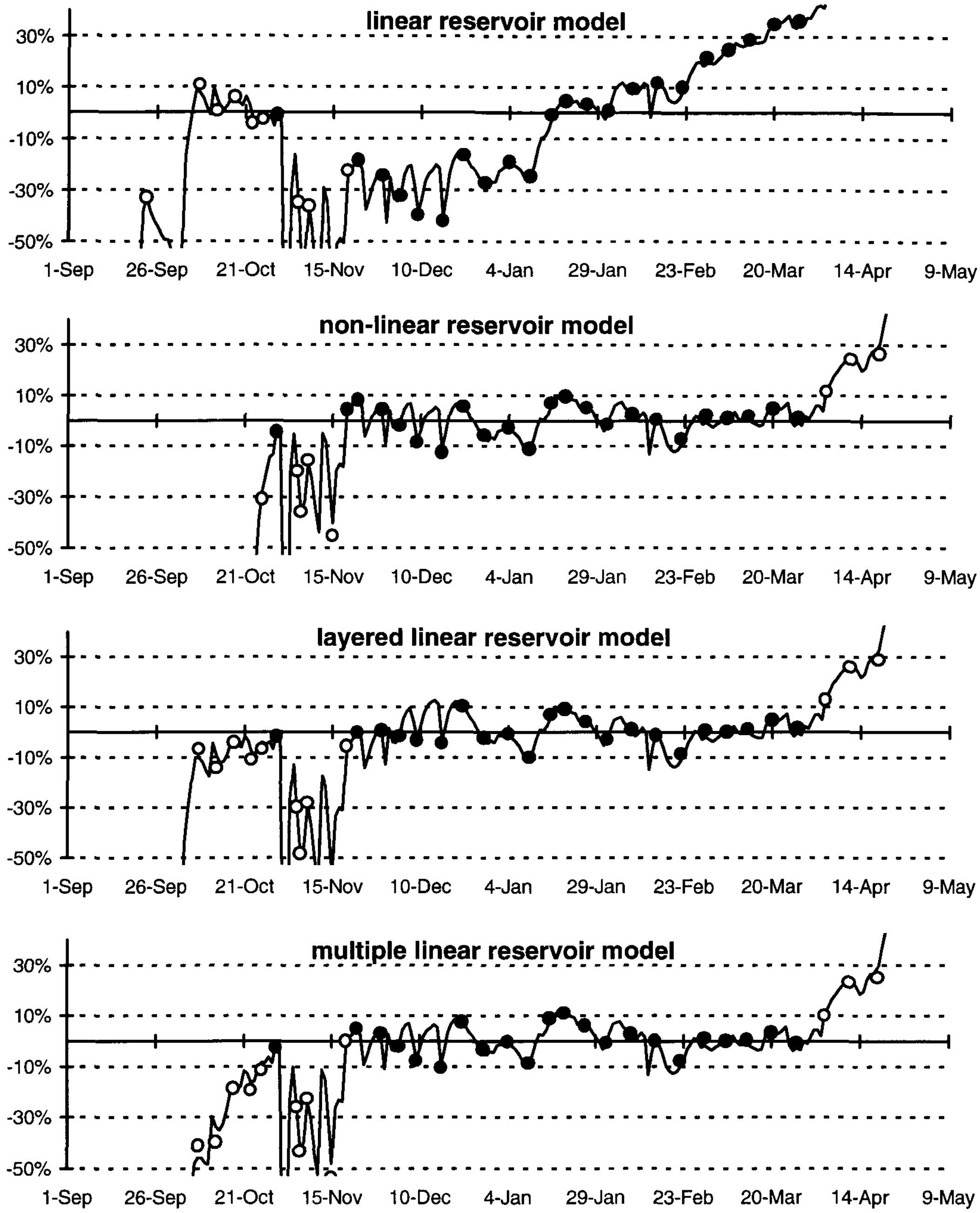


Figure 3.3 M'Clintock River near Whitehorse, residuals from recession models. Solid circles indicate deviation from measurements used for calibration.

Table 3.1 Recession modelling results

Linear reservoir model (Eq. 1.1)

	k	SSRE	n
M'Clintock River	0.011	1.18	22
Ibex River	0.012	0.67	24

Non-linear reservoir Model (Eq. 1.2)

	β	μ	SSRE	n
M'Clintock River	0.429	0.100	0.08	22
Ibex River	1.104	0.020	0.17	24

Layered linear reservoir Model (Eq. 1.3)

	k_1	k_2	t_L	SSRE	n
M'Clintock River	0.019	0.0036	42.1	0.05	22
Ibex River	0.016	0.0063	61.6	0.19	24

Multiple linear reservoir model (Eq. 1.4)

	k_1	k_2	A	B	SSRE	n
M'Clintock River	0.0025	0.045	4.77	6.37	0.07	22
Ibex River	0.0028	0.026	0.536	1.18	0.16	24

n = sample size

SSRE = sum of squared relative error (Eq. 2.4)

3.1.3 Discharge depression following freeze-up

Stage was observed to increase by 0.46 m between October 30 and November 5. This increase is equivalent to 35% of mean channel depth (1.31 m) on October 30. The lowest discharge measured during the freeze-up period was 4.61 m^3/s (November 10). This discharge is almost 30% greater than the late-winter minimum of 3.29 m^3/s (March 27). The volume of the discharge depression (V_d) was estimated as the difference between modelled discharge and recorded discharge through the freeze-up period, and V_d was found to be $2.7 \cdot 10^6$, $3.8 \cdot 10^6$ and $3.3 \cdot 10^6$ m^3 with respect to the non-linear reservoir, layered linear reservoir and multiple linear reservoir models. The duration of the event was approximately 20 days with the minimum flow occurring after 10 days. Flow recovery commenced once stage returned to the pre-freeze-up water level (Figure 3.1).

The volume of channel storage was calculated from estimates of channel length, width and increase in depth. This estimate should be an upper limit given that the width and depth diminish toward the source and that the magnitude of stage-up response is a function of depth. Channel length, including major tributaries, is estimated to be 150 km by map interpretation. The increase in depth, 0.46 m, multiplied by width at the measurement section, 18 m, results in an increase in cross-sectional area of 8.3 m^2. Multiplying the increase in cross-sectional area by channel length, a volume of $1.2 \cdot 10^6$ m^3 is estimated as a maximum value for channel storage, or about 30% of the volume of water in the discharge depression.

The difference in hydraulic head between the near-stream piezometers and the channel was positive prior to freeze-up, indicating groundwater discharge into the stream (Figure 3.4), but became negative after freeze-up indicating a reversal of hydraulic gradient (Figure 1.4).

3.1.4 Temperature effect

Close inspection reveals no consistent pattern relating temperature to discharge (Figures 2.4 and 3.1). However, there may be a trend in the plot of model residuals against temperature (Figure 3.5). The statistical significance of this correlation cannot be rejected outright because it is very close the 95% confidence level (R^2 =0.17, P = 0.061, n = 21). The practical significance of this effect, if it exists, is minimal with only three residuals exceeding the nominal measurement uncertainty of ±10% and 73% of measurements are within ±5% of model predictions.

3.1.5 Oscillations in stage

Stage variability during the period of intact ice cover was expected to be a function of channel hydraulic resistance. After the initial stage-up at freeze-up, stage was expected to decrease gradually in response to a smoothing of the under-side of the ice and decrease in discharge. However, stage dropped dramatically during the freeze-up period, returning to a level close to the pre-freeze-up stage (Figure 3.1). For the remainder of the winter, stage oscillated between the pre-freeze-up stage and a level about 7 cm higher. These oscillations do not consistently correspond to temperature

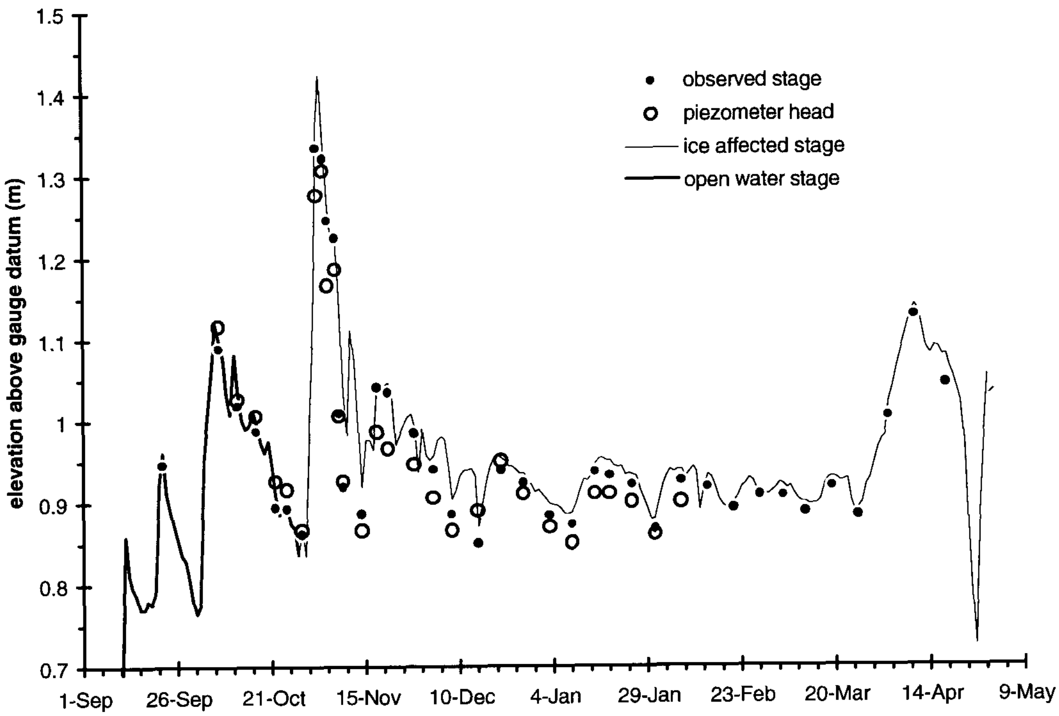


Figure 3.4 M'Clintock River near the mouth, piezometric head and channel stage

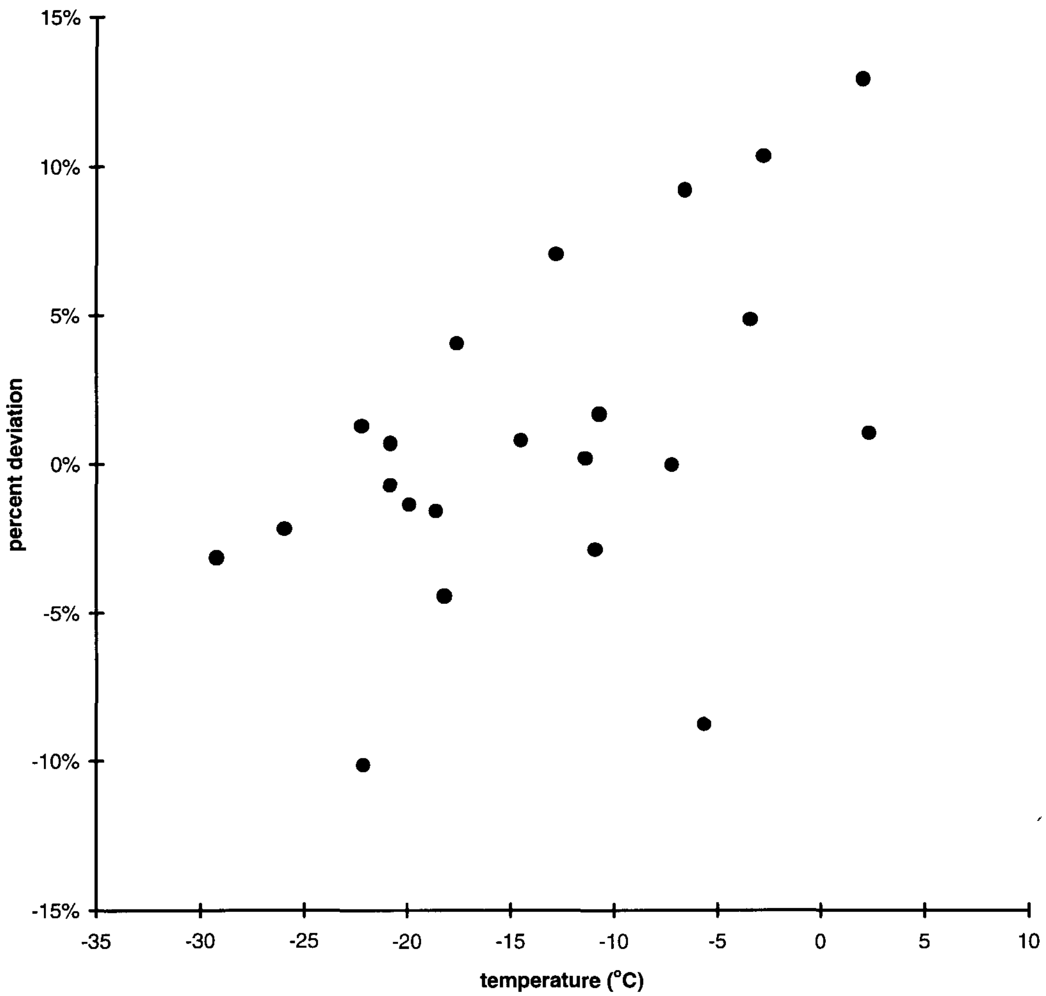


Figure 3.5 M'Clintock River near Whitehorse, residuals from layered linear reservoir model (Eq. 1.3) for the period November 20 to April 4 plotted against Whitehorse airport mean daily temperature

variability. It is unlikely that they are due to measurement error because they are observed independently in the manometer data (which measures changes in pressure above an orifice on the stream bed) and in survey level checks run from bench marks on the shore.

3.1.6 Environmental isotopes

Environmental isotope data (Table 3.2) show that variability in the results are less than analytical precision. This shows that there is no detectable change in isotopic concentration over the time period sampled.

3.1.7 Summary

The most active period of ice formation in this catchment was a three week period starting November 1. Minimum observed flow during this period was 4.6 m^3/s (November 10) which is 40% greater than the seasonal minimum observed flow of 3.29 m^3/s (March 27). During the freeze-up period a discharge depression of about $3.3 \cdot 10^6$ m^3 was observed with respect to discharge predicted by three recession models. The estimate of channel storage was $1.2 \cdot 10^6$ m^3, which is only about a third of that required to satisfy the volume unaccounted for on the discharge hydrograph. A comparison of time series of discharge and temperature is ambiguous in terms of temperature effect on discharge. The plot of model residuals against temperature (Figure 3.5) shows a weak trend. However, the effect, if it exists, is less than nominal measurement error.

Table 3.2 Environmental isotope data for Ibex and M'Clintock Rivers showing little variability over time.

Date	Ibex River		M'Clintock River	
	^{18}O	^{2}H	^{18}O	^{2}H per mil
November 7	-21.69	-168.6	...	...
December 5	-21.75	-168.5	...	...
December 19	-21.85	-167.1	...	...
January 9	-21.95	-166.3	...	...
February 13	-21.97	-166.5	...	...
March 01	...	...	-21.22	-165.9
March 20	-22.07	-166.8	-21.15	-165.2
March 27	...	...	-21.23	-165.2
April 4	...	...	-21.23	-162.5
April 10	-22.04	-166.3	...	...
April 18	-21.92	-167.7	...	...
April 19	...	...	-21.15	-164.1
April 24	-21.87	-168.1	...	...
April 26	...	...	-21.2	-162.4
May 6	-21.87	-170	...	...
min	-22.07	-170	-21.25	-165.9
max	-21.69	-166.3	-21.2	-162.4
mean	-21.90	-167.6	-21.23	-164.2
std. dev.	-0.119	-1.216	-0.019	-1.49
range	0.38	3.7	0.05	3.5

Precision of $^{18}0$ estimate ± 0.15 per mil

Precision of ^{2}H estimate ± 3.0 per mil

3.2 IBEX RIVER

3.2.1 Overview

Ice was first recorded on November 2 in the form of slushy anchor ice forming on the stream bed. Disturbance (such as walking through the ice with waders) would cause the ice to float to the surface and move away downstream. This frazil ice was very 'sticky' and would adhere to any obstacle. Anchor ice could be seen along most of the stream bed during a flight over the catchment November 2. By November 14 the stream was approximately 80% ice covered. Open leads persisted at the measurement section throughout the winter, in part due to the effects of groundwater inflow and in part due to efforts to keep the section ice-free. Only one measurement (February 20) was made from a complete ice cover. Recorded discharge is not continuous through the freeze-up period (Figure 3.6) because of changes in the ice cover which were made at the measurement section to ensure accurate discharge measurements. The manipulations to the ice cover invalidate the assumption of a linear distribution of shift corrections and so stage record could not be used to produce discharge data during this time. Stage and discharge measurements (solid circles) after April 4 plot below recorded discharge (solid line) because of a pronounced diurnal trend in stage during this period. The stream was typically measured in the morning which was close to the daily minimum. Detailed discharge measurement data are presented in Appendix B.

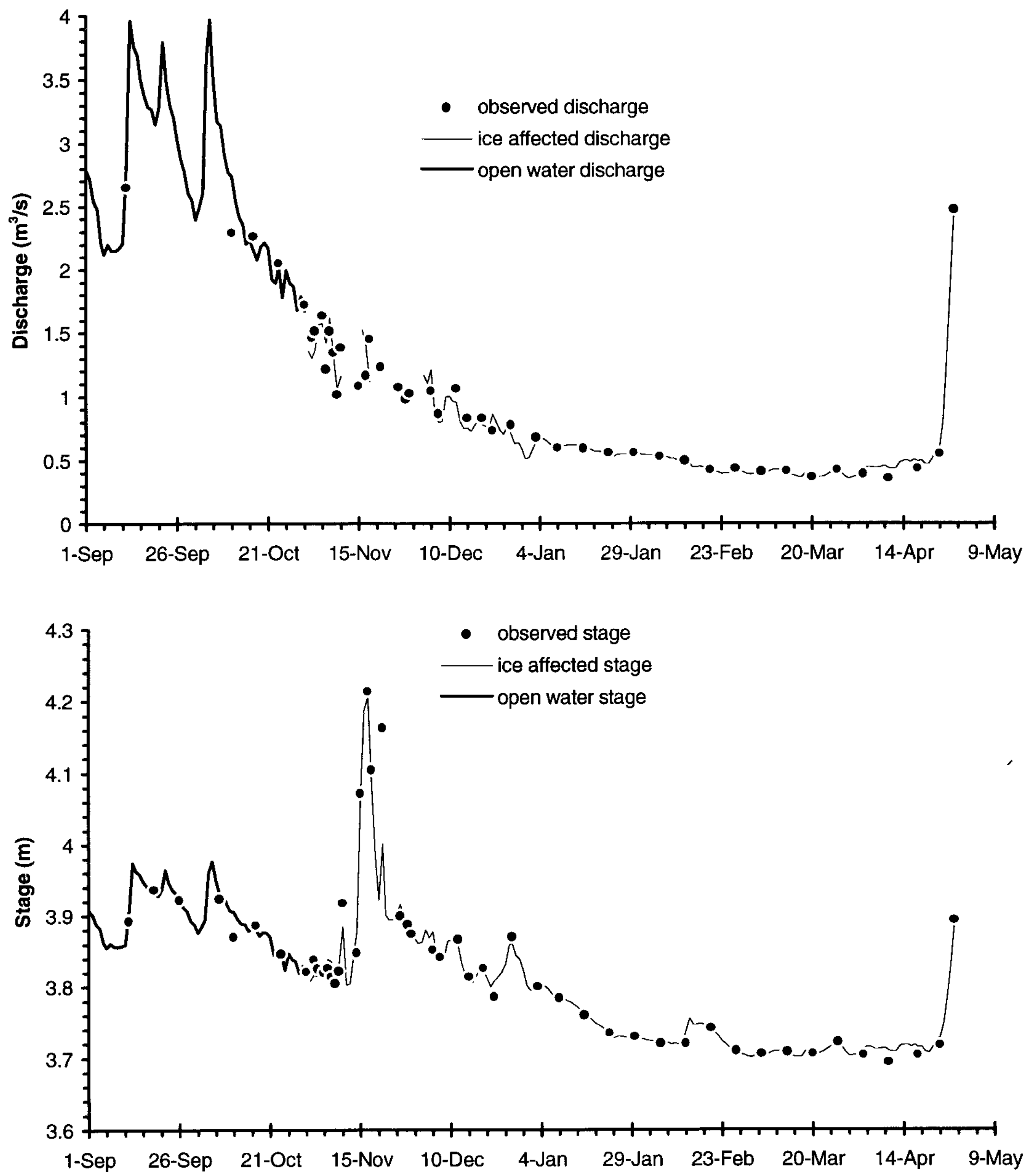


Figure 3.6 Ibex River near Whitehorse, stage and discharge.

3.2.2 Storage depletion

Model results (Figure 3.7a plotted on an arithmetic scale and Figure 3.7b on a semi-logarithmic scale) show that the layered linear and multiple linear reservoir models (Eqs. 1.3 and 1.4) provide the best continuity through the freeze-up period with a good fit to the pre-freeze-up discharge recession. The pattern of residuals is similar for Eqs. 1.2, 1.3 and 1.4 (Figure 3.8) and during the recession period (November 1 to April 4) most of the variability is within nominal measurement error. The residuals from the linear reservoir model (Eq. 1.1) show that this model does not represent the data well, with deviations of -30% during early winter.

Sodium concentration (Figure 3.9) shows a trend consistent with the the concept of a layered reservoir with a curvilinear slope when both levels are active (which would be expected from a diminishing contribution to flow from a relatively dilute upper layer) and an essentially flat slope when only the lower layer is active (after the break point in the model predictions) The plots for specific conductance and calcium lend support for this structure though the changes in slope are not as distinct as the sodium plot.

Discharge from the spring roughly followed a linear recession (Eq. 1.1) with a recession constant (k) of 0.0043 d^{-1} (Figure 3.10). At the start of the winter period (October 31) spring discharge was 0.007 m^3/s or 0.5 % of Ibex River discharge. At the end of the recession (April 3) spring discharge was 0.004 m^3/s or 1% of Ibex River discharge. The recession rate for spring discharge is similar to the slow

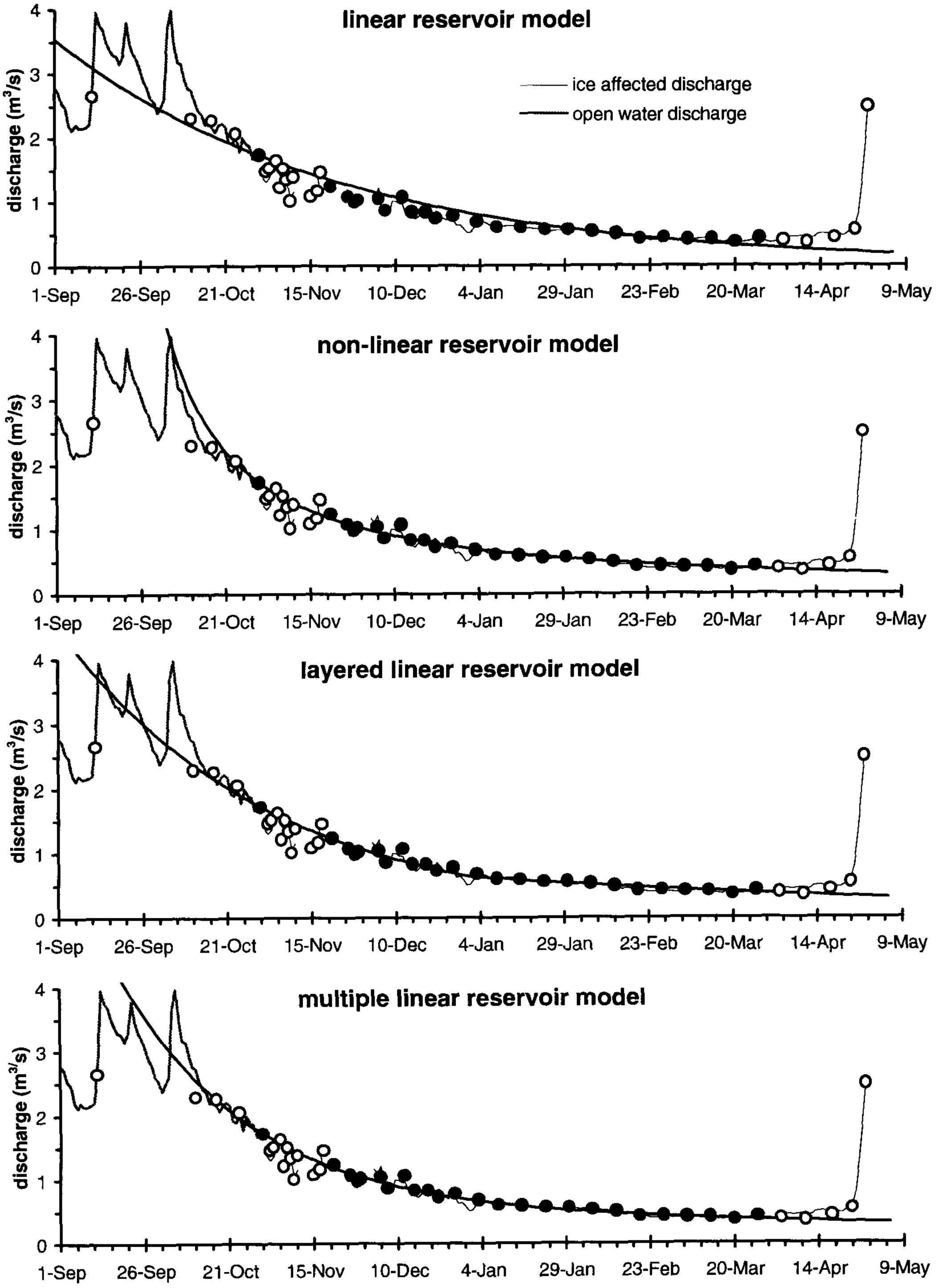


Figure 3.7a Ibex River near Whitehorse, recession model results plotted on arithmetic scale. Solid circles show calibration measurements.

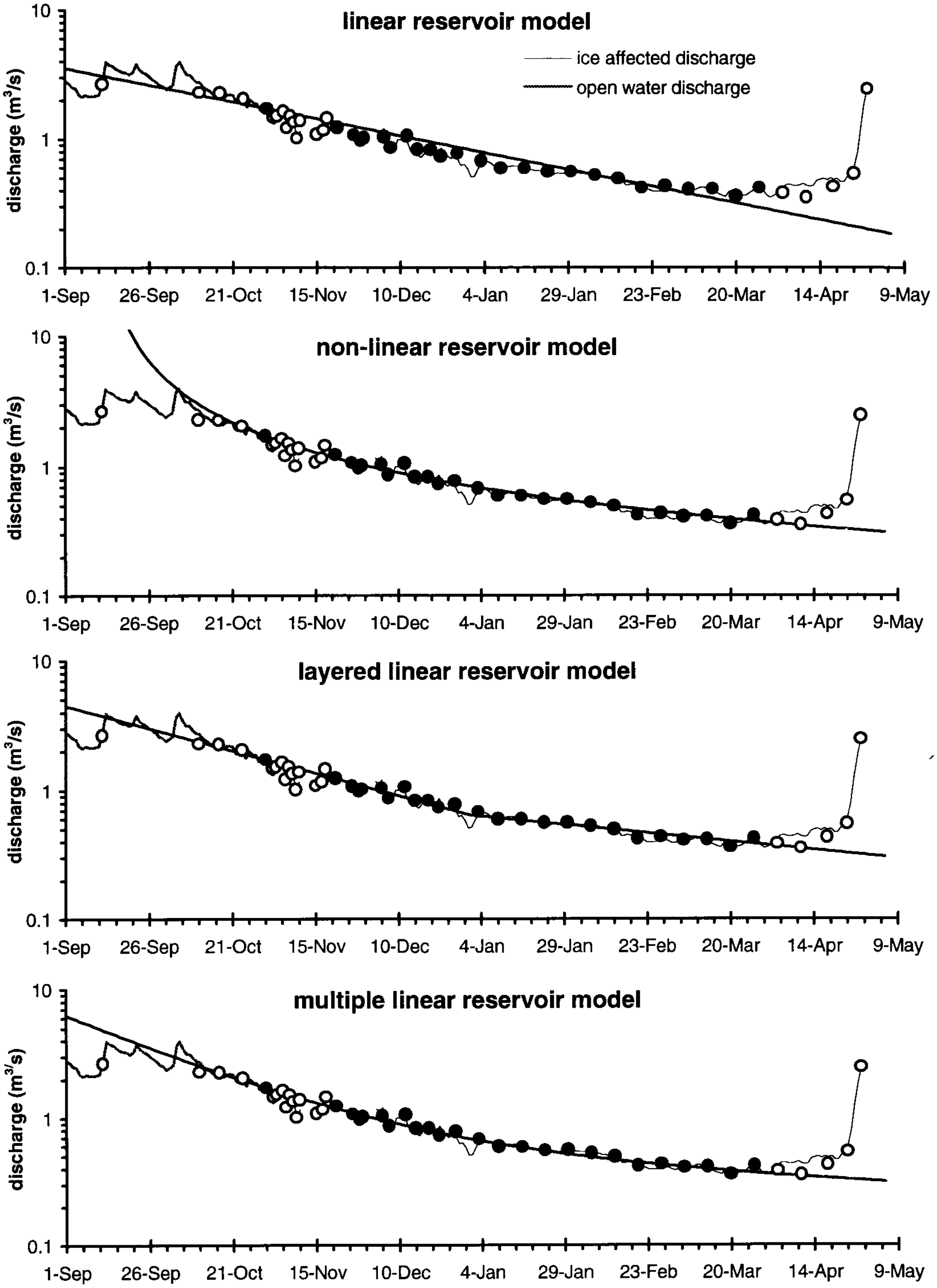


Figure 3.7b Ibex River near Whitehorse, recession model results plotted on logarithmic scale. Solid circles show calibration measurements.

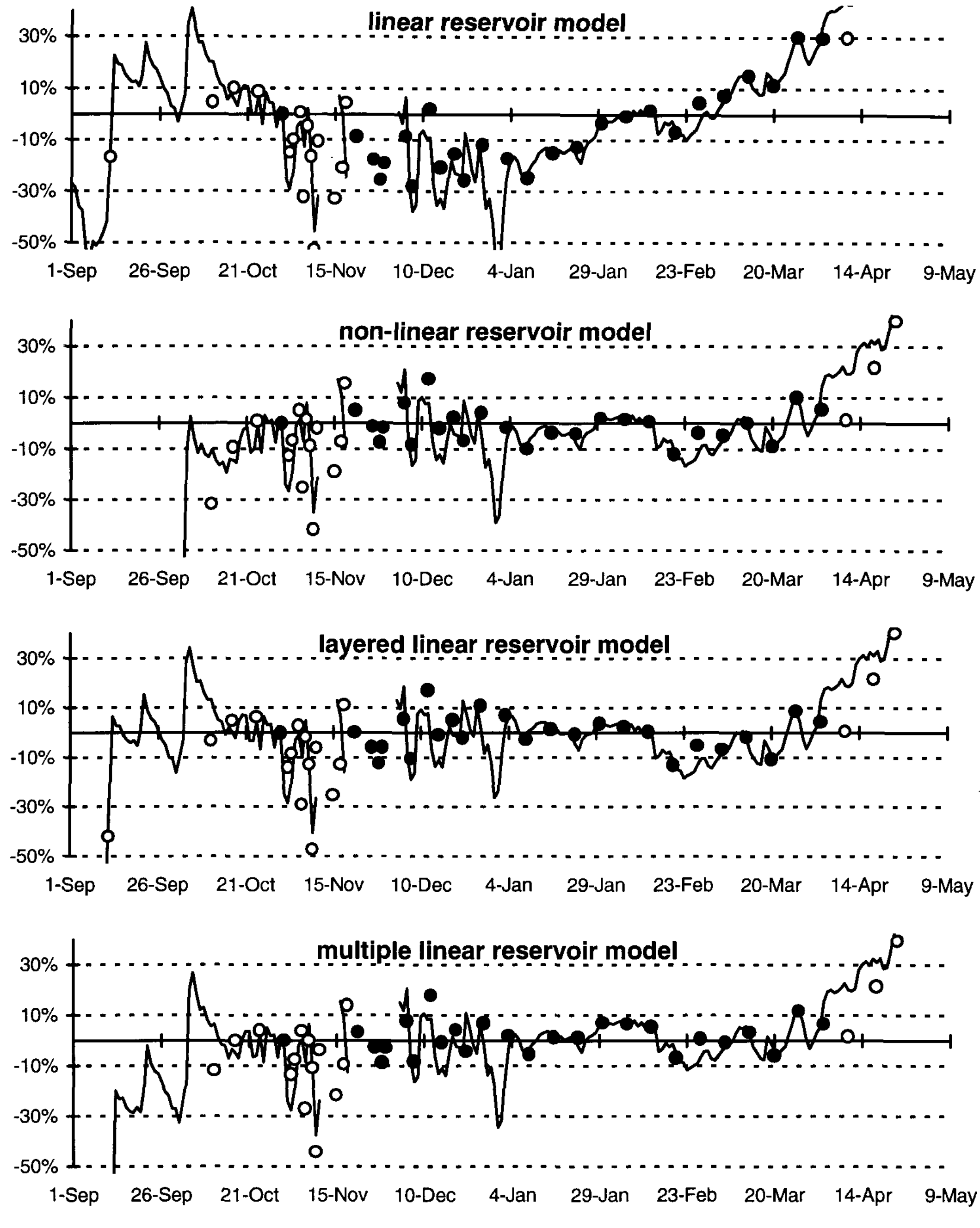


Figure 3.8 Ibex River near Whitehorse, residuals from recession models. Solid circles indicate deviation from measurements used for calibration.

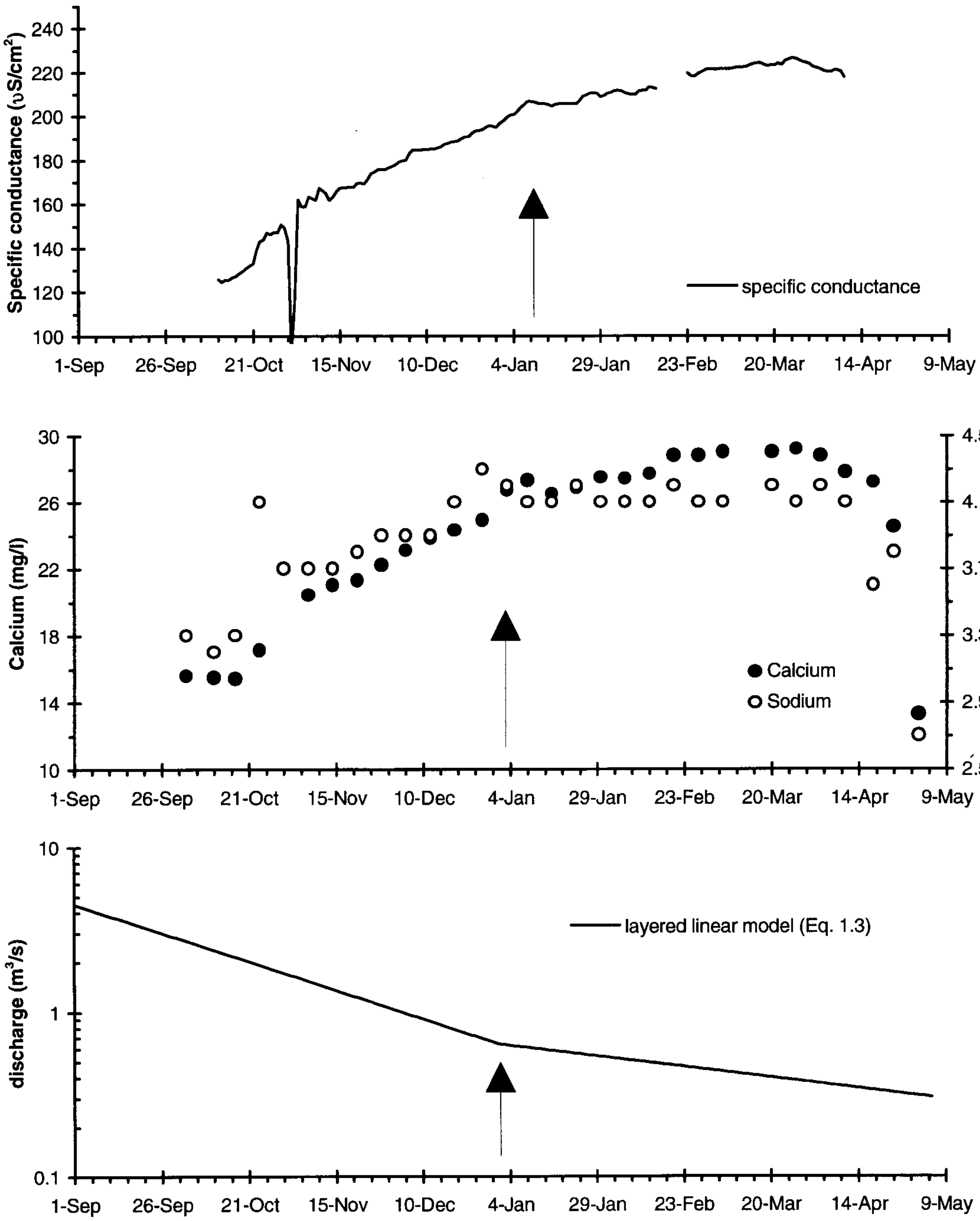


Figure 3.9 Water quality variables for Ibex River near Whitehorse showing a break in trend coincident with the break in slope of the layered model.

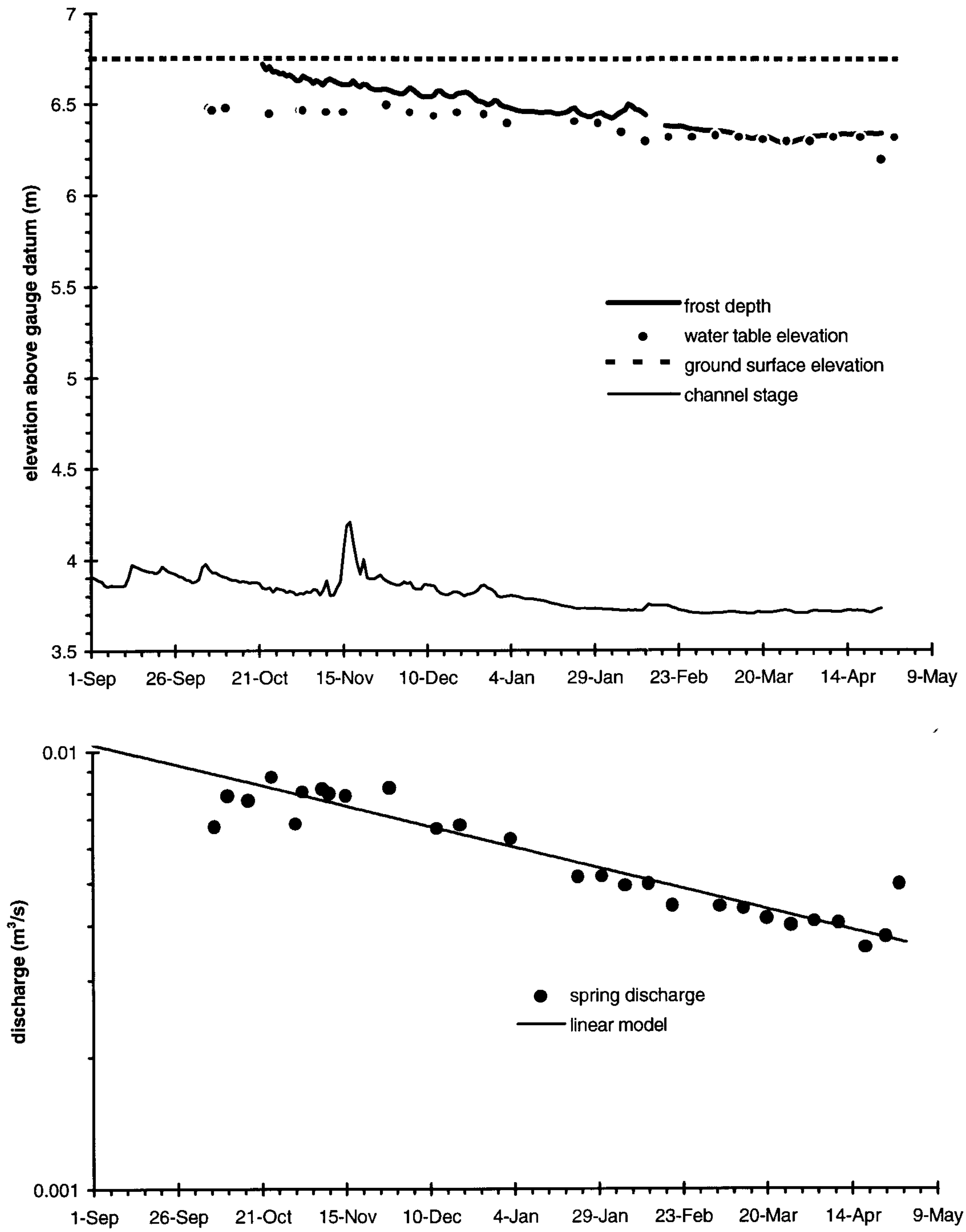


Figure 3.10 Ibex River near Whitehorse, frost depth, water table elevation, and spring discharge.

reservoir of the multiple linear reservoir model (K1 = 0.0028 d^{-1}) but slower than the lower level of the layered linear reservoir model (A2 = 0.0063 d^{-1}).

3.2.3 Discharge depression following freeze-up

The minimum discharge during the freeze-up period was 1.01 m^3/s (November 9), almost three times the late-winter minimum of 0.355 m^3/s (April 10). No calculation of V_d was made for this station because of the lack of reliable, continuous record through the freeze-up period. Any estimate made from the available data would be biased toward discharge values affected by local disturbances of the ice regime.

3.2.4 Temperature effect

The time series of temperature and discharge (Figures 2.4 and 3.6) are inconclusive in establishing correspondence between temperature and discharge. A plot of model residuals against temperature (Figure 3.11) shows no correlation between temperature and discharge variability with respect to the layered linear model. Linear regression results confirm that there is no statistically significant correlation between temperature and model residuals (R^2 = 0.005 and P = 0.722 for n = 25). Most of the measurements (80%) are within nominal measurement error; however, 5 residuals exceed ± 10% of model predictions.

3.2.5 Groundwater observations

The piezometers at the Ibex River were located in a groundwater discharge zone with

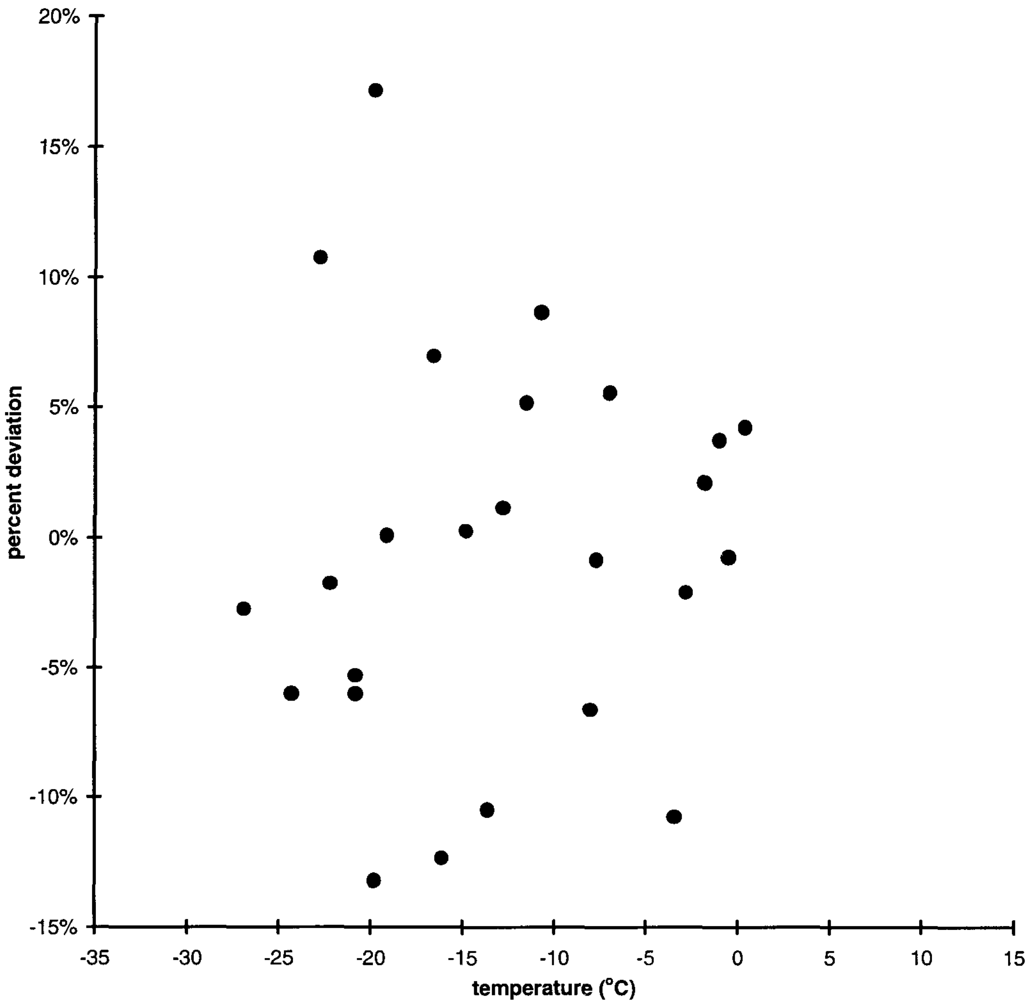


Figure 3.11 Ibex River near Whitehorse, residuals from layered linear reservoir model (Eq. 1.3) for the period November 20 to April 4 plotted against Whitehorse airport mean daily temperature.

a strong hydraulic gradient. The head differential between the water table well adjacent to the soil thermistors and the stream channel was approximately 2.7 m over a horizontal distance of 7 m (Figures 3.10 and 2.3). The Ibex is incised down to a level less than the surrounding water table due to the high gradient of the stream and bank seepage was observed throughout the winter.

Soil temperatures were compared to water table elevations for a site near the groundwater spring. Chin (1966) hypothesized that the movement of the frost horizon in groundwater discharge zones may respond relatively quickly to changes in air temperature and could act much like a valve in regulating the discharge of groundwater into the streams. Figure 3.10 shows a progressive deepening of the frost horizon which intercepts the water table in late winter. During the early winter upward spikes in frost depth correspond to periods of warm temperature (Figures 3.10 and 2.4). However these effects diminish with time, perhaps due to increased snow pack and depth below the surface. Variability in spring discharge shows no pattern consistent with these transient changes in frost depth as would be predicted by the Chin hypothesis.

3.2.6 Snowmelt observations

Snowmelt observation were made on March 20, March 27, April 3 and April 10. On March 20 the dye formed a clearly defined horizon under approximately 12 cm of new snow at the level site. However, on the hillside site there was no clear horizon of dye and the entire snow column was stained red signifying meltwater percolation. The depth of snow at the hillside site was approximately 12 cm compared to 36 cm at the level site. On March 27 there was no noticeable change from March 20 indicating no further melt during that week. On April 3 there was no snow left on the hillside site, and only 12 cm of snow was left at the level site, with staining of the entire snow column evident. On April 10 there was no snow left at the level site, though approximately 12 cm of snow depth remained over the general area. The dyed snow apparently melted more rapidly than the nearby undyed snow once the protecting surface snow cover melted, exposing the dye to the sun. Dilution of water quality variables, which is expected when meltwater reaches the stream, starts April 03 (Figure 3.9).

A 16% increase in discharge between March 20 and 27 may be due to snowmelt. However, Whitehorse airport mean daily temperature was above freezing only on March 18 and 19 (3.1 and 3.5 °C) and by March 27 the mean temperature was down to -10.7 °C. If the increase in discharge was due to snowmelt, the meltwater was likely delayed by transient storage en-route to the stream.

3.2.7 Environmental isotopes

The results of isotopic analyses (Table 3.2) show that there is little difference in the isotopic characteristics of the water over the course of the winter. This may mean that there is no change in source water over time such as would be the case with the single-reservoir models. Changes in water chemistry over the same time period as shown in Figure 3.9 indicate a change in flow path coincident with the layered linear reservoir model.

3.2.8 Summary

The most active period of ice formation occurred over a three week period starting November 1. Streamflow during this period was highly variable due to local storage and release events, particularly associated with anchor ice formation. Minimum flow during this period was 1.01 m^3/s (November 9) almost three times the seasonal minimum of 0.355 m^3/s (April 10).

A plot of residuals from the layered linear reservoir model (Eq. 1.3) against temperature shows no correlation between temperature and discharge variability (Figure 3.11). No evidence could be found from soil temperature profiles to support the Chin hypothesis of a temperature effect on groundwater discharge. Meltwater runoff, with routing delays may explain one of the deviations from the recession trend.

3.3 TAKHINI RIVER AT OUTLET OF KUSAWA LAKE

3.3.1 Overview

Ice cover was first observed on the south end of Kusawa Lake on November 30. The outlet polynya did not extend into the lake and some shore ice was forming along the channel edges. By December 10, the outlet polynya had narrowed to less than 10% of the channel width for a distance of 300 m downstream of the lake. The river was open bank to bank below 300 m from the lake due to a riffle which kept this section ice free all winter. An open lead persisted in the ice cover on the 300 m section of the river from the lake to the riffle through most of the winter and varied in width depending on temperature and wind speed. Time series plots of stage and discharge are presented in Figure 3.12 and detailed discharge measurement data are presented in Appendix C.

3.3.2 Lake storage and lake outflow

The magnitude of lake storage depletion was estimated by multiplying the difference in lake stage between the beginning and end of the recession period by lake area. Lake storage depletion over the recession period is estimated to be $1.6 \cdot 10^8$ m^3, which accounts for 43% of the total recession flow of $3.7 \cdot 10^8$ m^3 at the lake outlet. A more detailed analysis of lake storage depletion is shown in Figure 3.13. The hydrograph at the top of the page (Figure 3.13) shows storage depletion flow (Q_d) calculated using Eq. 2.2 and inflow (Q_i) calculated using Eq. 1.5 plotted along with observed and recorded discharge and modelled inflow. The layered linear reservoir model (Eq. 1.3) was used to represent Q_i based on visual inspection of the data points and results from Ibex and M'Clintock Rivers which support the validity of this model.

The lower plot in Figure 3.13 is a time series plot showing the proportion of lake outflow which can be attributed to lake storage depletion. The variability in this plot,

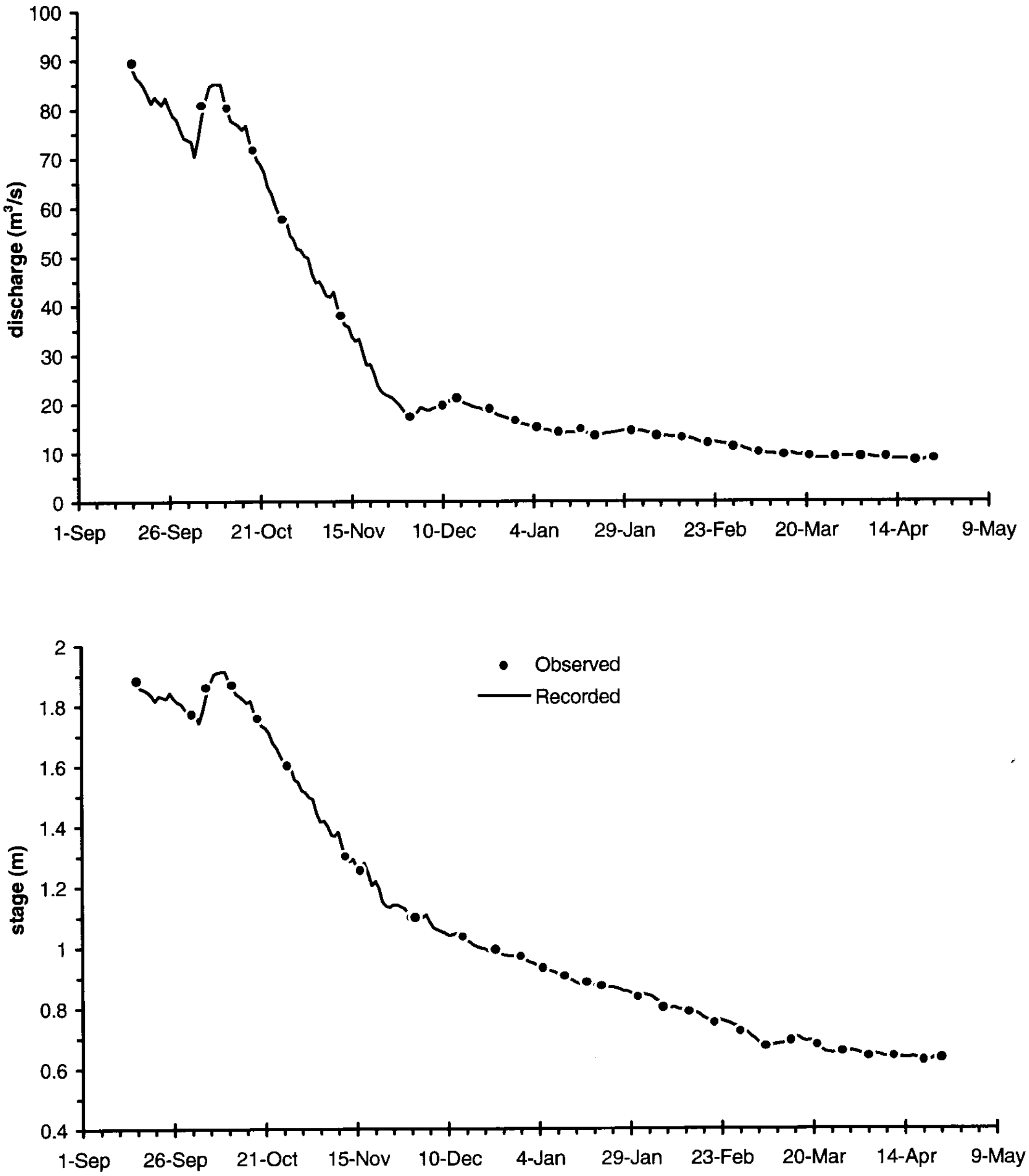


Figure 3.12 Takhini River at outlet of Kusawa Lake, stage and discharge.

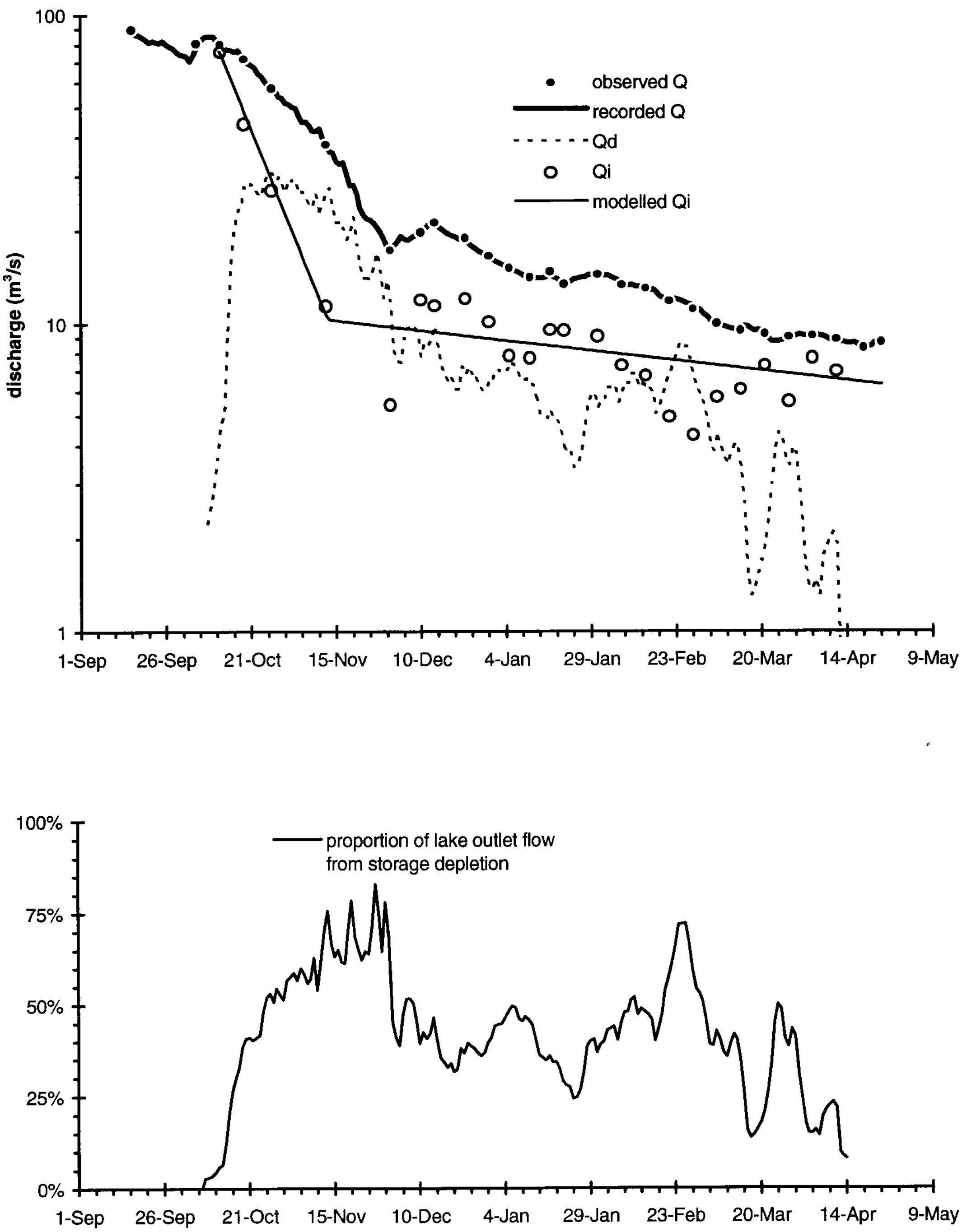


Figure 3.13 Inflow and outflow hydrographs for Takhini River at outlet of Kusawa Lake

particularly in late winter when discharge is low, may be due to measurement uncertainty. Eq. 2.2 is highly sensitive to accuracy in the measurement of stage because any error is multiplied by lake area which is a relatively large number. However, the trend which can be seen from this plot shows that Q_d is highest in early winter (~70%) and lowest in late winter (~30%).

3.3.3 Effect of snowfall on lake outflow

A comparison of precipitation and lake discharge (Figures 3.12 and 2.4) shows no apparent correspondence between precipitation events and discharge. The major precipitation events deposited snow water equivalents in excess of 5 mm. This amount is sufficient to displace 700,000 m^3 of water ($0.005\ m \cdot 1.4 \cdot 10^8\ m^2$ surface area). This effect can be calculated from the stage-discharge equation (Eq. 1.6). An increase of 5 mm to a stage of 0.75 m results in an increase in discharge of 0.2 m^3/s for a flow increase of 1.7%, which is less than measurement accuracy.

3.3.4 Effects of ice formation and decay on lake outflow

A discharge depression was evident at the outlet of the lake on December 1, caused by ice formation at the lake outlet. No obvious change in lake stage was associated with this event (Figure 3.12). The discharge depression is evident in comparison with discharge from the Takhini River near Whitehorse (Figure 3.14) and in a concurrent upward spike in specific conductance at Takhini River near Whitehorse (Figure 3.14). An increase in conductance indicates a decrease in dilute lake water, which lends

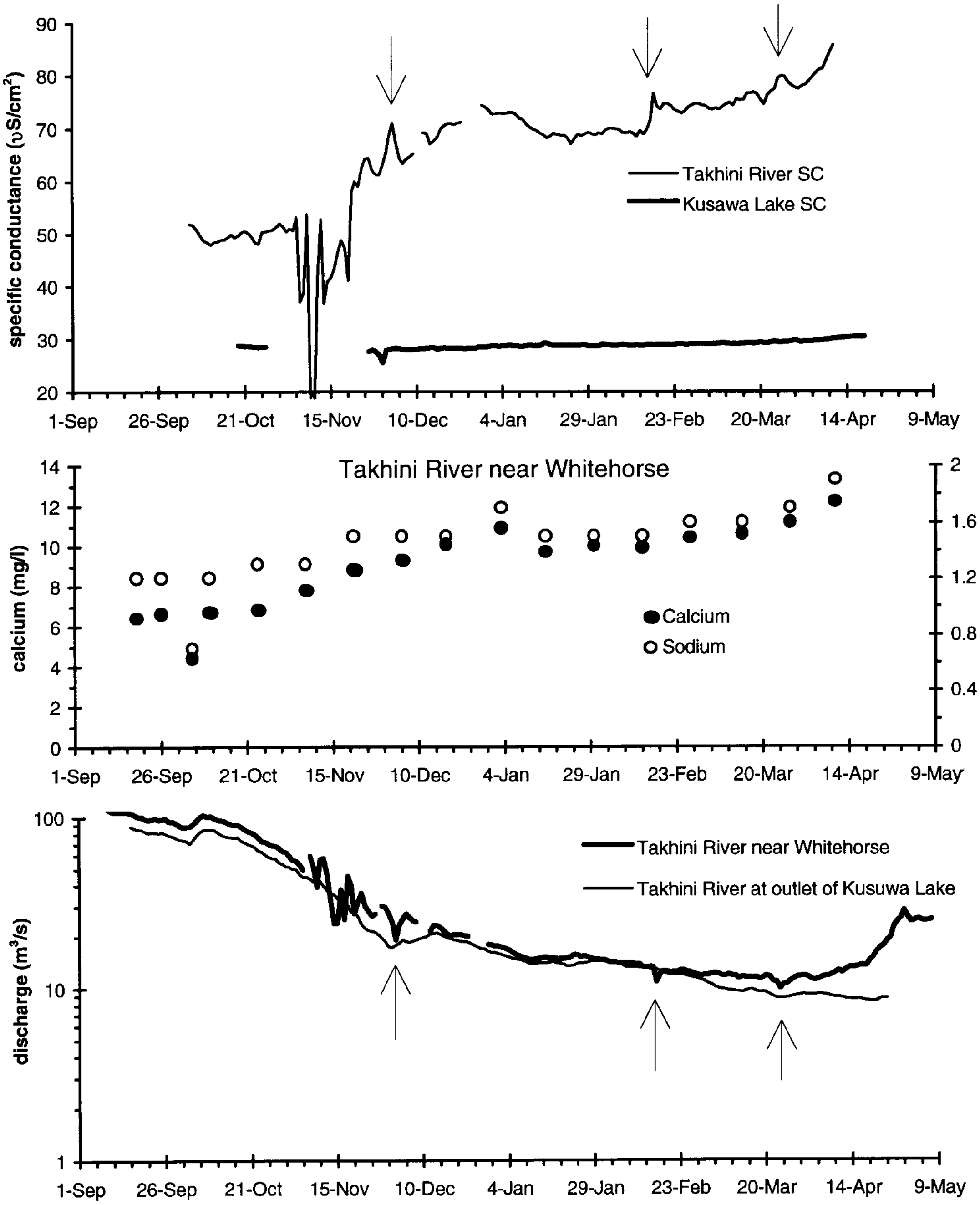


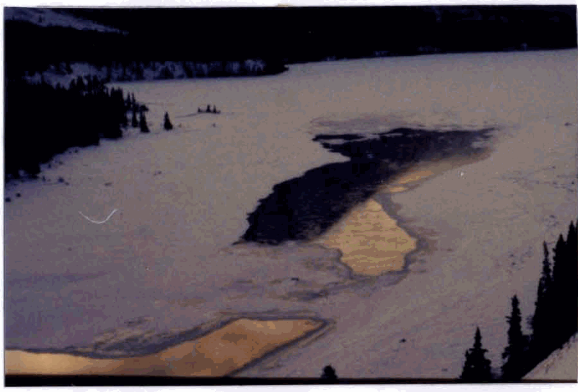
Figure 3.14 Selected water quality variables for Takhini River near Whitehorse and Takhini River at outlet of Kusawa Lake.
Arrows point to events associated with Kusawa Lake discharge depressions

support to the conclusion that this discharge depression originated from Kusawa Lake. The downward spike in conductance at the lake outlet (Dec 01) is likely due to frazil ice accumulating on the measurement probe.

There is evidence for two smaller lake outlet discharge depressions though not from the lake outlet discharge data, which are calculated from lake stage. Specific conductance spikes concurrent with discharge spikes at Takhini River near Whitehorse occurred on February 16 and March 24. These events are associated with very cold night-time temperatures (Figure 2.4). Figure 3.15 shows three pairs of photographs of the lake outlet associated with the three events. The first pair shows the outlet on November 12 prior to ice formation and on December 10 after the discharge depression event of December 1. The second pair of photographs were taken on February 14 and February 21 showing that the lake outlet polynya was substantially re-established by February 14 but substantially reduced after the second discharge depression event. The third pair of photographs taken on March 21 and March 28 are less conclusive in establishing a link between the size of the polynya and discharge depression. However, it is likely that ice that may have formed March 24 under clear skies with 11 km/hr NNW winds and an overnight low of -34°C, may have decayed by March 28, which was overcast with 21 km/hr SSE winds and a daytime high temperature of +7.3°C.

November 12 1994

December 10 1994



February 14 1995

February 21 1995

March 21 1995

March 28 1995

Figure 3.15 Photographs of the outlet of Kusawa Lake before and after three discharge depressions

3.3.5 Summary

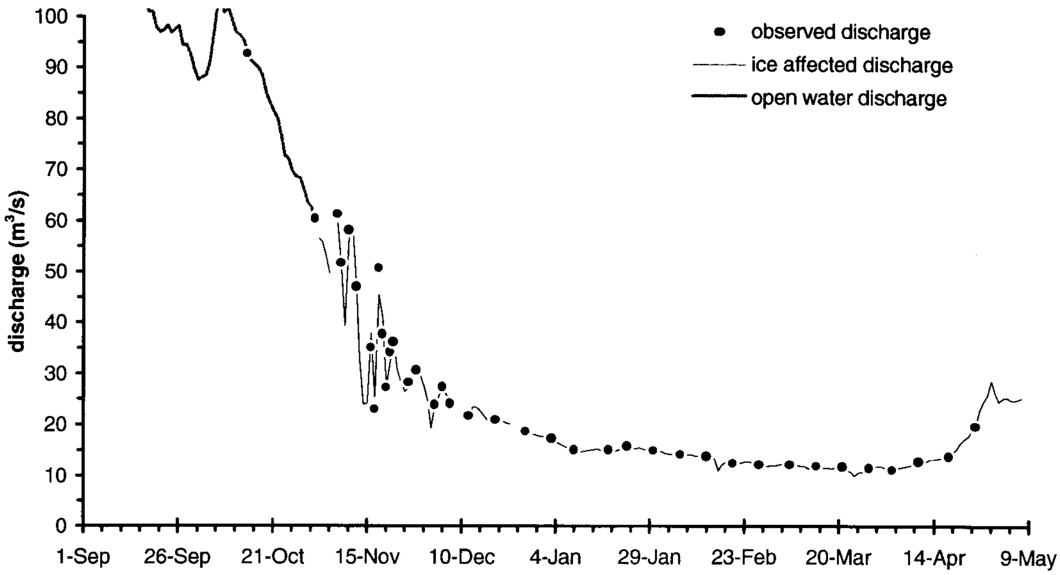
Lake storage depletion was found to account for 43% of winter recession streamflow at the lake outlet over the winter period. Lake storage contribution to flow was greatest in early winter when lake stage was high. Depletion of lake storage is controlled by the stage-discharge relation at the lake outlet and is highest during the early winter when lake stage is high. Lake discharge does not respond rapidly to precipitation events even though the volume of water added to the lake may be substantial. Ice formation at the lake outlet can obstruct discharge, causing discharge depressions to occur. These events may occur at any time of the year in response to atmospheric conditions due to the effect the lake outlet polynya has on controlling streamflow.

3.4 TAKHINI RIVER NEAR WHITEHORSE

3.4.1 Overview

The development of ice cover on the Takhini River occurred over a 3 week period. Ice was first observed in the Mendenhall River tributary on October 27 and on November 2 ice was observed on the lower Takhini. There was no ice from the highway bridge up to the lake, but frazil was forming in-channel and the river was bank full of frazil at the confluence with the Yukon River. On November 6 the river was 80% full of frazil ice at the Yukon River and 5 to 10% frazil ice at the highway bridge. On November 7 frazil ice accumulated on a 1" diameter steel cable which was anchoring electrical cables to transducers at the study site. The volume of ice was sufficient to float the cable off of the streambed in over 2 m of depth and rip the transducers out of the streambed. Stage response to the displacement of water by frazil ice was evident (Figure 3.16). On November 9 frazil ice at the station occupied 50% of the surface area and water turbidity was noticeably greater, with visibility reduced to less than 0.3 m from over 1 m (as indicated by the depth of submergence of the current meter before it could no longer be seen). On November 10 the water was very turbid, and the frazil ice was full of stones and gravel indicating that it had formed on the stream bed and lifted. There was no slush ice on November 12. On November 15, 70% of the surface area was slush ice and hard pans of ice. On November 16 the river was 95% ice covered. By November 21 the turbidity had cleared and the current meter could be seen to a depth of 1.5 m when it was lowered into the water.

Stage increased approximately 0.8 m at freeze-up (Figure 3.16) which is more than 50% of the pre-freeze-up mean channel depth (1.5 m). No clearly defined discharge depression was associated with this stage-up response. Much of the variability in discharge during the freeze-up period was due to runs of frazil ice. During periods of anchor ice formation, water is abstracted from flow to satisfy channel storage and then released when the anchor ice lifts and flows downstream. A typical pattern would be for anchor ice to form in cold, clear weather and to run during warmer, cloudy weather. After April 4 stage first increased and then decreased substantially. Discharge response during this time was inconsistent with stage change. The most



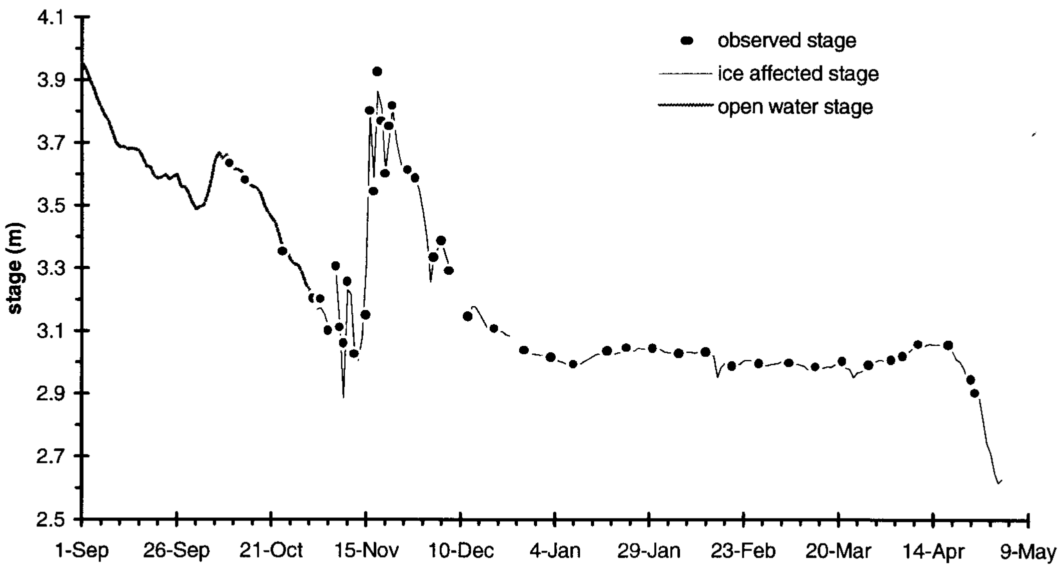


Figure 3.16 Takhini River near Whitehorse, stage and discharge

likely cause of the stage increase is a change in channel storage associated with an increase in the hydraulic resistance of the ice cover due to thermal decay. The decrease in stage after April 10 is likely associated with a release of channel storage as open leads increase in size. By April 24 the channel was 30% open at the study section. Discharge measurement data for Takhini River near Whitehorse are listed in appendix D and discharge measurement data for Mendenhall River are listed in appendix E.

3.4.2 Streamflow accounting

An estimate of the increase in channel storage due to stage-up was made for the Takhini River. Channel length (40 km) multiplied by width (70 m) multiplied by the increase in stage (0.9 m) gives a volume of $2.5 \cdot 10^6$ m^3. Stage-up occurred between November 15 and 18. Flow from Kusawa Lake was not affected and was 35 m^3/s. At that rate, channel storage could be satisfied in 20 hours. The combination of a relatively short reach of river affected by stage-up and an ample supply of flow result in a discharge depression at the Takhini River near Whitehorse that is difficult to distinguish from events related to frazil flow.

Stage records were not collected for Mendenhall River and so reconciliation of streamflow can be done only for those days on which discharge measurements were conducted at Mendenhall River. Figure 3.17 summarizes the streamflow accounting. Prior to February 28 known upstream flow was consistently greater than measured

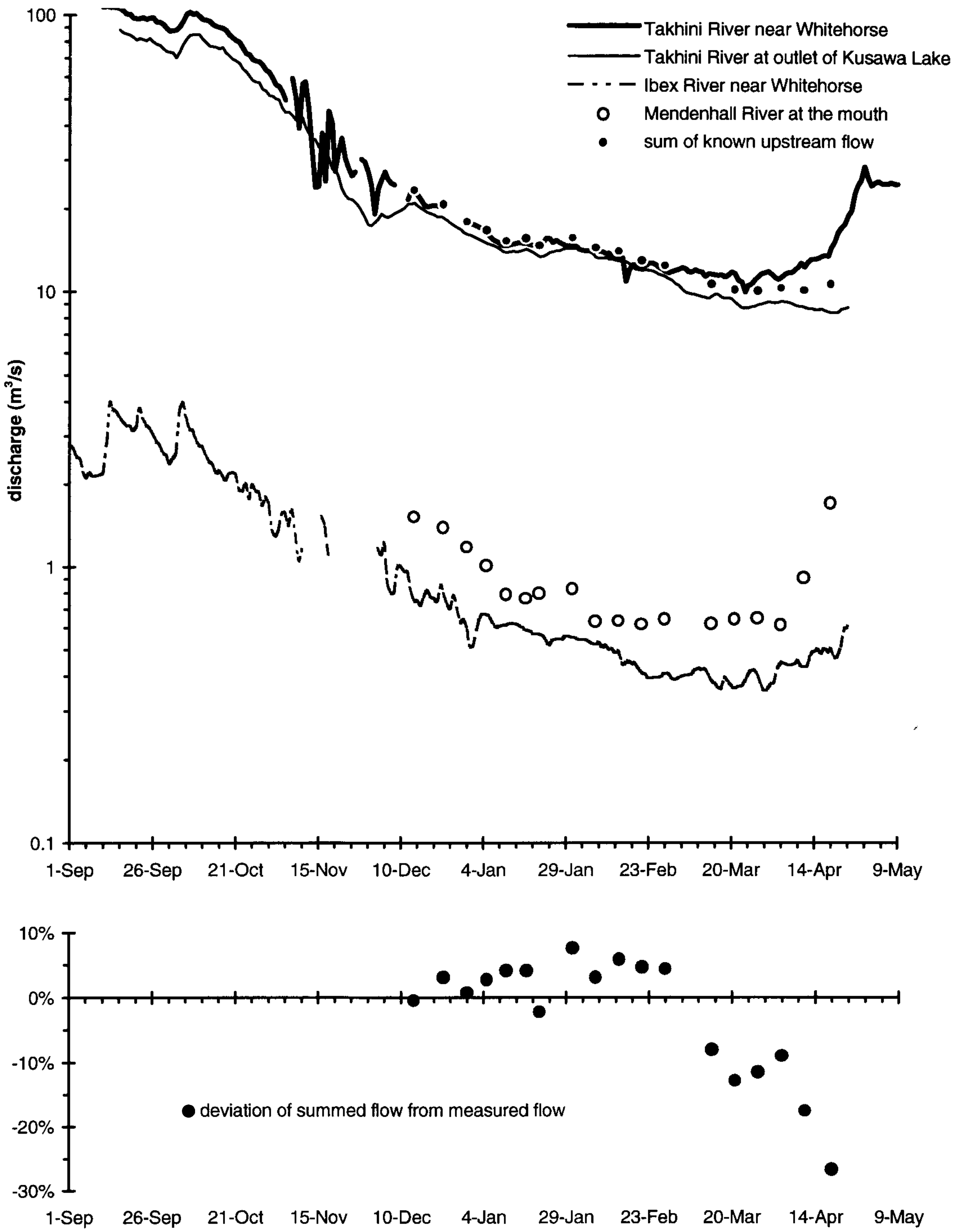


Figure 3.17 Takhini River system, streamflow accounting

flow and after February 28 known upstream flow was consistently less than measured flow (Figure 3.17, lower graph). One would expect the sum of known upstream flow to be slightly less than that measured at Takhini River near Whitehorse due to unmeasured flow contributions from the lower Ibex River, Arkell Creek and the reach of Takhini River between the lake outlet and the highway bridge (Figure 2.2). A random distribution would be expected if measurement error were the cause of deviations of summed flow from measured flow. Low elevation snowmelt is a likely cause of the deviation of the last two points (April 11 and April 19) in this plot. However, the four points during the month of March cannot be explained in terms of meltwater due to cold temperatures prevalent during March (Figure 2.4). An alternative explanation of the deviations may be that changes in channel storage and/or groundwater inputs may affect the volume of water unaccounted for in the unmeasured reach of river. The results in the following sections on water quality, stage and discharge and ice thickness and temperature will be evaluated with respect to these unexplained deviations.

3.4.3 Water quality

Specific conductance at Takhini River near Whitehorse (Figure 3.14) increased through the period when deviations of summed flow from measured flow were negative (February 28 to April 19), indicating an increase in the ratio of groundwater inflow to lake water flow.

There was a slight depression in specific conductance and in calcium and sodium concentrations (Figure 3.14) during much of the period when deviations of summed flow from measured flow were positive (January 5 to February 28), indicating that groundwater inputs were reduced relative to lake outflow during that time. These results indicate that groundwater inflow can account for at least some of the unexplained deviations. However, water quality data are insensitive to changes in channel storage and an examination of stage and discharge is required to determine in-channel effects on discharge.

3.4.4 Stage and discharge

If the unexplained deviations were a result of changes in channel storage in the unmeasured reach then channel stage should have been decreasing (indicating water coming out of storage) during the period when deviations were negative. This explanation is not supported by stage data (Appendix D) which show stage to be only slightly less on March 28 (2.989 m) than on February 27 (2.994 m). The magnitude of the deviations during March indicate that approximately 10% of the Takhini River discharge (1.2 m^3/s) is unaccounted for. Based on a channel length of 30 km and width of 45 m, a change of approximately 2.3 m in stage would be required to account for the flow during this period, much greater than the observed stage decrease of 5 mm.

Mendenhall River discharge increased from 0.616 m^3/s on February 21 to 0.650 m^3/s

on March 28. This observation is not supported by snowmelt observations and so may be due to a reduction in backwater effects from the Takhini River. Flow from the Mendenhall would have been reduced due to stage-up on the Takhini during and following freeze-up, but as Takhini River stage dropped, channel storage from the lower reaches of the Mendenhall River would be released. While the increase in discharge at Mendenhall is accounted for in the calculation of discharge deviations it may be indicative of the process occurring in the lower reaches of the Ibex River which would similarly be affected by backwater from the Takhini River.

The effect of main-channel stage on tributary inflow is similar to the process of hydraulic damming and groundwater inflow would likely be affected in a similar way as supported by the water quality data. Figure 3.18 shows channel stage and near-stream hydraulic head from piezometer 2. This plot is similar to Figure 3.4 for the M'Clintock River in that a reversal in hydraulic gradient occurs associated with freeze-up, providing further evidence that hydraulic damming may be occurring on the Takhini River.

3.4.5 Ice thickness and temperature

Figure 3.19 shows a time series plot of ice thickness observations and mean ice cover temperature. Changes in plasticity or thickness of the ice cover may affect channel storage and release and hence, discharge. A correlation should exist between changes in channel stage (Figure 3.18) and either ice temperature (if temperature is a suitable

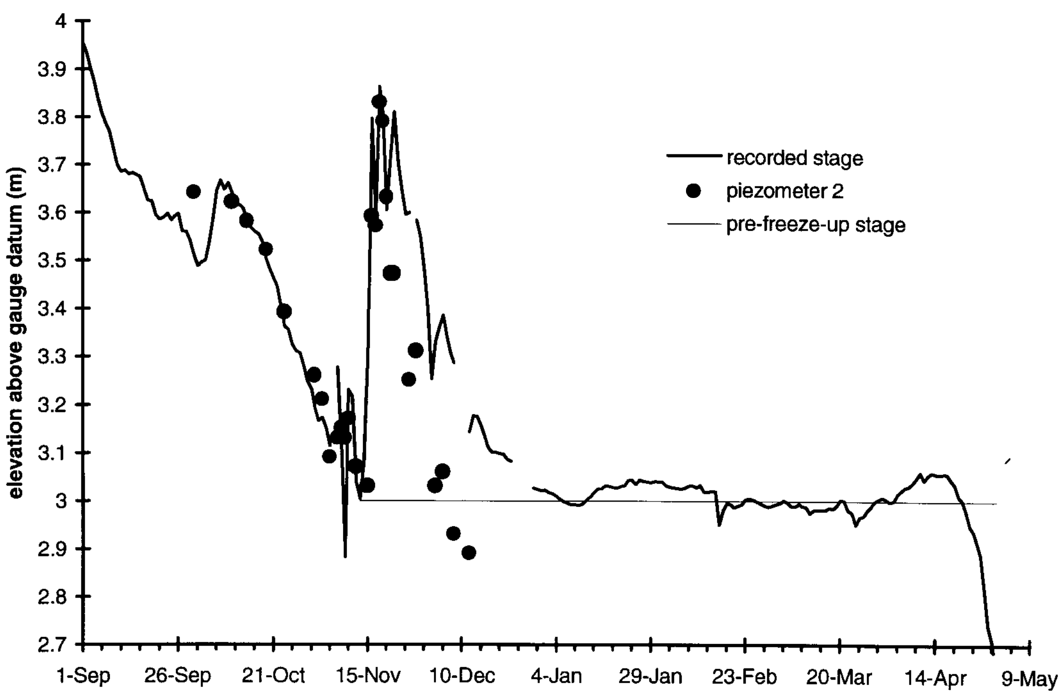


Figure 3.18 Takhini River near Whitehorse, channel stage and piezometric head

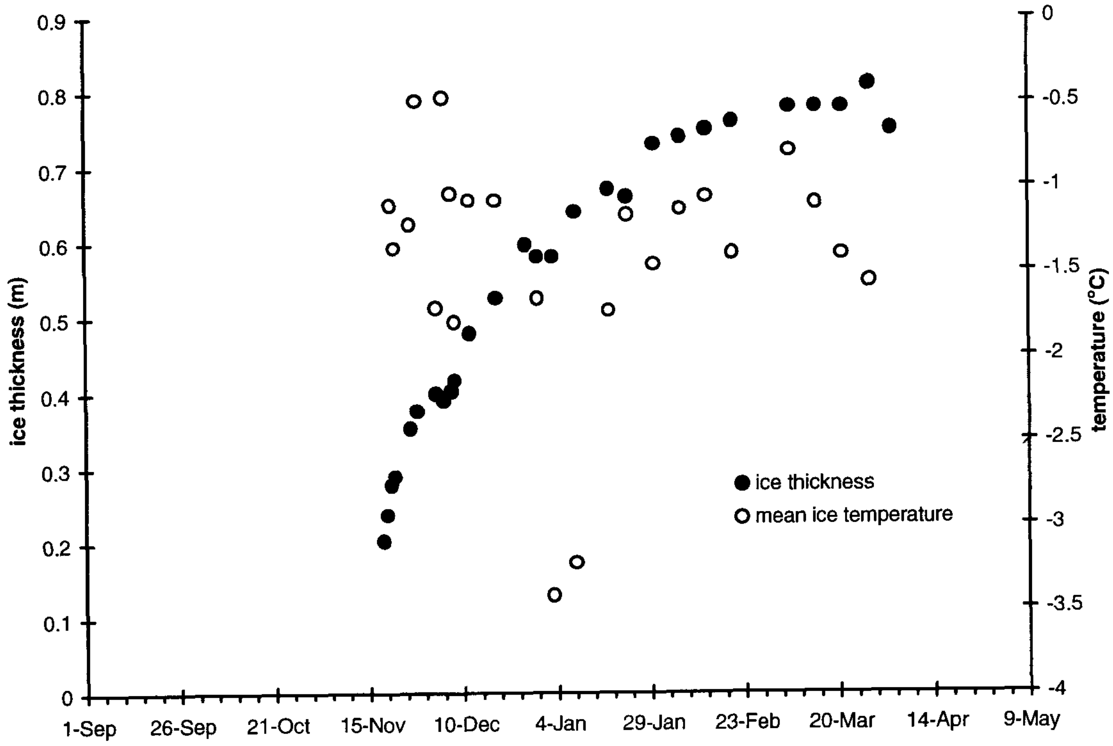


Figure 3.19 Takhini River near Whitehorse ice cover temperature and thickness

surrogate measure of the plasticity of ice) or ice thickness if this effect is significant on a daily time scale. There seems to be little variation in ice thickness from a relatively smooth concave downward curve. Daily mean temperature of the ice cover did vary; however, this variability is not consistent with either stage or discharge variability.

3.4.6 Summary

A stage-up response was observed on the Takhini River near Whitehorse without a coincident discharge depression due to the dominance of lake outflow. Frazil ice flow during freeze-up resulted in rapid changes in discharge during the period of ice formation. Reconciliation of discharge from the Takhini River near Whitehorse with the sum of known upstream flow showed that through mid-winter measured discharge was less than expected and during late winter measured discharge was greater than expected. Channel stage and near-stream hydraulic head measurements provide evidence that hydraulic damming occurred associated with stage-up. These observations are supported by water quality variables which show that groundwater inputs were reduced during the mid-winter period relative to the late-winter period. Groundwater flow reductions, along with reduced tributary flow during the period of stage-up, contribute to a subsequent drop in stage which is necessary for flow from these sources to resume.

3.5 CHAPTER SUMMARY

Hydrographs for the winter period from two groundwater dominated streams (M'Clintock and Ibex) can be represented with three different recession models (Eqs. 1.2, 1.3 and 1.4) with similar measures of goodness of fit. The layered linear reservoir model (Eq. 1.3) was found to be most consistent with uncalibrated pre-freeze-up discharge and with water quality indications of reservoir structure. Deviations from the model predictions occurred during the freeze-up period, indicating freeze-up induced discharge depression. The magnitude of the M'Clintock River discharge depression is three times as great as the volume of water that can reasonably be accounted for by channel storage. Measured hydraulic gradients between the stream and the near-stream aquifer support the hypothesis of hydraulic damming of groundwater inputs.

Evidence of a temperature effect on discharge is ambiguous. Neither the groundwater nor in-channel hypotheses could be supported with results from this study. There is evidence for micro-climatic snowmelt associated with an interval of above-freezing weather. However, routing delays obscure a clear correlation of this meltwater input with temperature.

Lake storage depletion was found to account for approximately 43% of total winter discharge at the outlet of Kusawa Lake. Discharge depressions related to ice formation can occur throughout the winter period as atmospheric conditions affect ice conditions at the outlet polynya. Snow on ice events can displace water but the release of this water is controlled by the stage-discharge relation at the lake outlet and is distributed over a long period of time without immediate measurable impact on discharge from Kusawa Lake.

Variability in stage and discharge were observed at M'Clintock River (Figure 3.1). These variations seem to be non-random but not related to temperature. The magnitude of the discharge deviations were mostly within nominal measurement error but are not likely due to measurement error because the same pattern is evident in the stage data, which is an independent measurement.

Chapter four will offer answers to the research questions posed in chapter one with respect to the data presented in this chapter. The interaction between stage and discharge will be examined in more detail and a hypothesis will be presented to explain the variability in stage and discharge observed at M'Clintock River.

CHAPTER FOUR

DISCUSSION

The research questions posed in chapter one are discussed in the first four sections of this chapter. Section five addresses stage-discharge interactions, with specific attention to proposing a hypothesis which may explain stage oscillations observed at M'Clintock River. The last section examines the broader context of these findings.

4.1 GROUNDWATER STORAGE DEPLETION PROCESSES

Winter streamflow from sub-Arctic, groundwater-dominated streams can be represented within measurement error by three storage depletion models (Eqs. 1.2, 1.3 and 1.4). That three conceptually different models all produce similar model performance statistics is an indication that all three models have sufficient degrees of freedom to fit the simple shape of the recession. The linear reservoir model (Eq. 1.1) does not represent observed discharge recession when calibrated to the entire winter period. That the linear reservoir model fails over such a long time period is consistent with the observation of previous authors (e.g. Hall, 1968; Tallaksen, 1995) that this model is valid for only a limited range of discharge.

The layered linear reservoir model (Eq. 1.3) represents groundwater storage depletion processes the best of the four models based on fit to uncalibrated pre-freeze-up discharge and water quality data. The multiple linear reservoir model (Eq. 1.4) closely approximates the performance of the layered linear reservoir model in this study. However, the multiple linear reservoir model may be more conceptually appropriate for a case where there are two or more geologically distinct tributaries contributing flow or for a case where an upper aquifer is separated from a lower confined aquifer by an aquitard or aquiclude (e.g. Clausen, 1992). A potential application of Eqs. 1.2 and 1.3 is the forecasting of winter flows at the time of freeze-up. Such forecasting would be feasible if the parameters of the recession functions did not vary from year to year. Further research should address the consistency of storage depletion functions.

4.2 FREEZE-UP RELATED DISCHARGE DEPRESSION

Discharge depressions associated with freeze-up were observed at M'Clintock River as well as Takhini River at the outlet of Kusawa Lake. The magnitude of the stage-up response at M'Clintock River was 35% of mean channel depth which is fairly close to the prediction of 32% (Beltaos et al. 1993). No sustained discharge depression was observed at either Ibex River or Takhini River near Whitehorse. The observed stage-up response at the Takhini River near Whitehorse was greater than predicted at 50% of mean channel depth. The difference between theoretical and observed stage-up response is likely due to the presence of frazil and anchor ice in the channel at the Takhini River. Stage at Ibex River during freeze-up was affected by attempts the keep the cross-section free of ice for discharge measurement purposes and therefore is not representative of natural processes.

Minimum flow associated with freeze-up discharge depression was found to be substantially greater than late winter low flow. By comparison, Gray and Prowse (1993) showed freeze-up related minimum flow which is substantially less than late-winter low flows for the Liard and Mackenzie Rivers. Burn (1993) also reported freeze-up related low flows for the Mackenzie River which are substantially lower than late winter low flows.

The magnitude of the freeze-up discharge depression depends on the nature of the ice cover and on channel geometry. Generally, initial ice conditions on large northern rivers are very rough and large volumes of frazil ice are usually prevalent. These conditions would increase the stage required to pass a given volume of flow and therefore would require proportionately greater channel storage than the streams examined in this study. Large rivers also tend to have relatively large plan areas with respect to smaller rivers and so would require relatively more volume per unit of stage increase. Comparisons are further complicated by differences in effects occurring in near-by channels. For example, the Liard River discharge depression (Gray and Prowse, 1993) may be exacerbated by stage-up on the Mackenzie River causing a slope change at the confluence, resulting in an additional backwater effect. The Mackenzie River discharge depression was observed in East Channel where stage-up not only causes increased channel storage but can also result in inflow into some Mackenzie Delta lakes (Burn, 1993).

Channel storage alone cannot explain the difference between observed discharge and discharge predicted by any of the storage-depletion models during the freeze-up period at M'Clintock River. Reversal of the hydraulic gradient across the stream bed associated with stage-up was observed at both M'Clintock River and at Takhini River near Whitehorse. This observation is consistent with the hypothesis that bank storage and groundwater mounding occur, blocking, or at least reducing, groundwater discharge into the channel (Chin, 1966). As a result, discharge rapidly diminishes, resulting in a drop in stage. When stage drops to the pre-freeze-up water level the blockage of inflow is eliminated, allowing discharge to recover to the recession trend.

4.3 TEMPERATURE EFFECTS ON WINTER STREAMFLOW

It is acknowledged that some of the observed variability in winter streamflow may be a result of synoptic processes such as an effect of temperature which is masked by a lag in discharge response. However, no convincing evidence was found in this study to link residuals from the groundwater storage-depletion model to temperature during periods of sub-freezing weather. Temperature effects on groundwater discharge or on in-channel processes could not be supported. If temperature effects exist, they would appear to be less than measurement uncertainty and therefore not easily detectable.

Discharge from Kusawa Lake was observed to respond to cold temperatures as a result of ice forming at the lake outlet with a subsequent recovery of flow with the decay of lake outlet ice. This observation may explain Chin's (1966) conclusions of a temperature effect on groundwater discharge given that he was working without a complete set of data from the lake outlet and assumed monotonically-decreasing lake outflow. Chin therefore attributed variations in discharge at Takhini River near Whitehorse to variations in groundwater contributions to flow downstream of the outlet.

The persistence of the concept of temperature effects on winter discharge may also be due to an assumption that all winter streamflow measurements are equal in quality. However, there are many sources of error in winter discharge measurements (Pelletier, 1989; Pelletier 1990) and temperature may have a large effect on measurement quality. The effects of cold temperature on metering equipment and on the technician are likely biased toward underestimating flow. For example, exposure of the current meter to extremely cold air will cool the meter below freezing, which will cause a thin layer of ice to form on the meter when it is submerged, resulting in an under-estimation of velocity. The diligence of the technician in ensuring that the meter is performing properly may also be affected in a negative way by extreme temperatures.

4.4 STREAMFLOW VARIABILITY BELOW A LAKE

Lake inflow is conceptually independent of the lake storage reservoir. However, at the start or end of any recession period (i.e. during the interval when lake stage is neither rising or falling) $Q_i = Q_o$ (from Eq. 1.5). From the stage-discharge equation (Eq. 1.6) we know that for any discharge there is a unique stage associated with that discharge. Any stage value is related to lake storage by the lake area. Therefore, there is a unique volume of storage associated with any inflow at the start of a recession. From the mass balance equation (Eq. 1.5) the most appropriate model to represent discharge recession at the outlet of a lake is one in which a model of lake storage depletion (which can be derived from the stage-discharge relation) is coupled in series with a model of lake inflow. For practical purposes the two models can be lumped together in one empirically determined model because time and initial discharge are common variables to the two models. Empirical studies in Norway have shown that hyperbolic recession equations are generally suitable for catchments which are characterized by a high lake percentage (Tallaksen, 1995).

During the winter period several variables affect lake discharge recession. Ice changes the stage-discharge relation by affecting hydraulic roughness at the lake outlet, by affecting the stage-area relation and by affecting channel slope. The result of ice formation at the lake outlet is to reduce the rate of lake storage depletion. Conversely, decay of lake outlet ice will restore the open-water stage-discharge relation, causing an increase in discharge. Variables which affect the formation of ice include initial water temperature, air temperature, depth of water, heat influx from the bed, heat influx from solar radiation, incoming long wave radiation, wind speed and water vapour pressure (Efremova, 1972). Hence, discharge below a lake may respond to changing meteorological conditions.

Snowfall increases lake storage by the amount of water equivalent in the snowfall multiplied by lake area. The rate of release of this increased storage volume is controlled by the stage-discharge relation at the lake outlet. The effect of snowfall-related stage increments on discharge is a slight increase in discharge over an extended period of time. As a result of overlapping effects from sequential snowfall events the cumulative effect is increased discharge in late winter. For semi-arid sub-Arctic regions such a southern Yukon the effect of snowfall on winter low flow may not be important. For example, total monthly precipitation during the late winter at Whitehorse is typically about 12 mm. A 12 mm increase in late winter Kusawa Lake stage results in only a 3.2% increase in discharge when distributed over one month's time. This result contrasts with those in other studies (e.g. Kuusisto, 1984) wherein snowfall onto lakes accounted for a substantial portion of winter flow.

The 7-day low flow for Takhini River occurred from March 28 to April 03. During this period lake storage depletion is estimated to have contributed approximately 30% of the flow at Takhini River near Whitehorse. Janowicz (1991) found that a surface water storage index improved multiple regression results only slightly for models of 7-day low flows in Yukon Territory. The effect of lake storage would be expected to be minimal during the late winter as lake discharge decreases to a quantity near that of lake inflow. That the lake storage component of low flow was so high for the Takhini River, when only a slight effect was anticipated, may be due to the large discharge depression in early December or to the higher than normal discharge at the start of the recession period.

4.5 INTERACTION BETWEEN STAGE AND DISCHARGE

4.5.1 Backwater effects on tributary inflow

An abrupt increase in channel stage at a stream confluence would have the effect of reducing channel slope in the lower reaches of the tributary stream. This reduction in slope would reduce stream velocity with a consequent reduction in discharge. The flow restriction at the tributary mouth would cause flow from upstream sources to accumulate in the lower reach causing a rise in stage. The reduction in tributary discharge would have a proportional effect on main channel flow which could be substantial enough to result in a drop in stage in the main channel which would also help to restore inflow.

4.5.2 Longitudinal slope variability

The effect of an ice cover on discharge is more complex than simple abstraction of flow to satisfy channel requirements of stage-up. The magnitude of stage-up depends on local conditions (Beltaos et al., 1993) and is therefore unlikely to be uniform along the length of a stream. It is possible that this variability can result in stage and discharge instability as observed on the M'Clintock River. The effect a given amount of stage-up will have on discharge will also vary between reaches along the length of a river depending on channel geometry (which determines the volume of water required for channel storage) and the nature and type of inflow into the reach.

A schematic example of a non-uniform stage-up response is shown in Figure 4.1 which represents a longitudinal cross-section of a river where the stage-up in reach A is 30% of channel depth, reach B is 60% of channel depth and reach C is 40% of channel depth. A consequence of a non-uniform stage response is that channel slope anomalies would occur with associated velocity anomalies. The effect of these velocity anomalies would be that the stage anomaly would propagate upstream as a result of water draining from the downstream end (due to increased velocity) and accumulating in the upstream end (due to reduced velocity). As this stage anomaly propagates upstream it may pass a groundwater discharge zone or a tributary stream causing a blockage of inflow.

The stage and discharge response at any given cross-section would therefore depend on the magnitude and duration of these stage anomalies propagating upstream. The duration of these events would be determined by the volume of water required to satisfy changes in channel storage requirements divided by discharge entering the reach from upstream. The change in channel storage requirement would be a function of the length and width of the reach and the magnitude of the stage disparity. The magnitude of the stage disparity would be due to differences in hydraulic conveyance between the reach in question and the upstream and/or downstream reaches. Discharge from upstream would then fluctuate on the time scale of the speed of propagation of the stage waves divided by the length of channel between inflow reaches.

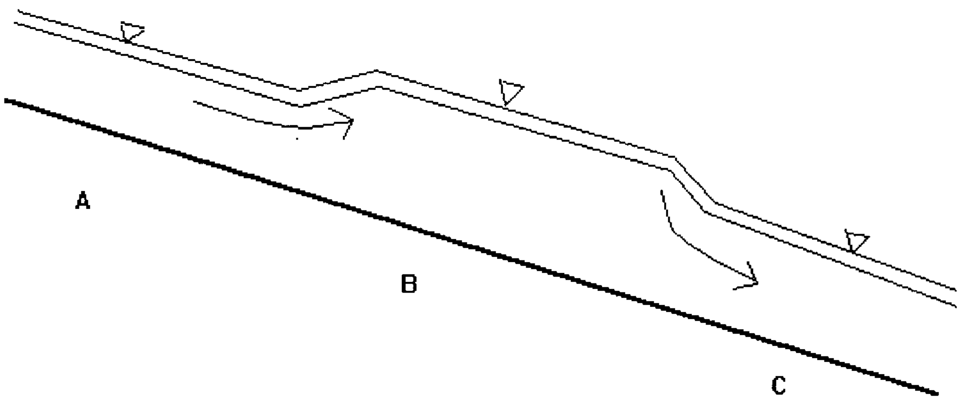


Figure 4.1 Sketch illustrating the change in channel slope between consecutive channel reaches with non-uniform stage-up response.

Longitudinal variations in stage, causing discharge variations over time, could complicate the definition of winter stage-discharge curves. An implicit assumption in the use of stage-discharge curves is that channel slope is either constant or varies in a predictable way with stage. If channel slope at a given cross-section varies over time, as outlined above, the assumption of a unique stage-discharge relation is invalid. Shift corrections to stage-discharge relations determined under open water conditions should only be used with caution under ice conditions because the low flows encountered during the winter period may be well below the calibrated portion of the curve.

4.6 Summary

The dominant process controlling discharge in sub-Arctic streams is storage depletion. The most appropriate storage depletion model for the two groundwater-dominated streams studied was found to be the layered linear reservoir model (Eq. 1.3). Deviations from the model occurred during the freeze-up period in the form of a discharge depression due to the formation of an ice cover. Water was abstracted from flow during this period to occupy channel storage. This effect was exacerbated by groundwater mounding, which blocked groundwater inputs, and backwater effects which blocked tributary inputs. Periodic variability in the model residuals throughout the winter may be due to variability in channel slope caused by non-uniform stage-up response along the length of a channel.

In a lake-dominated stream, two recession models in series, with one of the models representing lake storage depletion, and the other representing inflow, was most appropriate. These two reservoirs are linked but could be lumped together in an empirically determined function. Environmental variables significantly influence storage depletion from a lake but not from groundwater reservoirs. The dominant variables are those that contribute to ice formation (e.g. temperature, wind speed and solar radiation) and snowfall. The persistence of the lake outlet polynya throughout the winter means that ice effects can be transient and manifest at any time during the winter when atmospheric conditions result in ice formation. The effect of snowfall is to increase lake storage causing higher flows in late winter than would be otherwise expected, although this was a minor factor in this study.

CHAPTER FIVE

CONCLUSION

5.1 SUMMARY OF FINDINGS

5.1.1 Groundwater storage depletion models

The layered linear reservoir model (Eq. 1.3) provided the best fit to observed discharge over the winter period at the two groundwater-dominated streams. The non-linear reservoir model (Eq. 1.2) and the multiple linear reservoir model (Eq. 1.4) also provided a good fit to the calibration data, but the layered linear model was better at providing good continuity with uncalibrated pre-freeze-up discharge and was also consistent with water quality data.

Because of the simplicity of the shape of the recession curve, any number of variations of these models may be found which could provide reasonable statistical measures of goodness of fit. This may be part of the reason why recession modelling continues to be an active field of discussion and research in hydrology (e.g. Tallaksen, 1995). Isolating a recession curve from open-water data is a subjective process which can introduce errors greater than the variability in the actual storage depletion process. Working within the range of introduced error, different investigators may come to different conclusions about the storage depletion process for any given catchment. The sub-Arctic winter offers an opportunity to examine the storage depletion process in the absence of effects from hydrological variables which confuse recession analysis during the open-water season.

5.1.2 Freeze-up related discharge depression

Discharge depression associated with freeze-up was observed at the M'Clintock River near Whitehorse and at the lake outlet (Takhini River at the outlet of Kusawa Lake). Sustained discharge depression was not evident at Takhini River near Whitehorse, likely because the volume of flow from Kusawa Lake was sufficient to rapidly satisfy channel storage requirements. The discharge depression in the M'Clintock River is not as great as observed in investigations of the Liard and Mackenzie Rivers relative to the scale of the rivers (e.g. Gray and Prowse, 1993; Burn, 1993). This is likely due to differences in the volume of channel and bank storage requirements and the supply of flow to satisfy that storage. In addition, the Liard River may be affected by backwater effects due to stage-up on the Mackenzie.

5.1.3 Atmospheric controls on winter discharge

Temperature was found to have no measurable effect on groundwater-dominated discharge. Snowmelt runoff may follow even a brief interval of above freezing weather subject to some routing delay. Snowfall on lake ice is equivalent to inflow and therefore affects discharge downstream from a lake. However the release of snowfall-displaced water from the lake is controlled by the stage-discharge relation at the lake outlet. As a result, no immediate measurable response results from typical snowfall events. Atmospheric variables which affect the formation or decay of ice such as temperature, wind speed and sky condition can affect discharge downstream of a lake through changes in ice conditions at the outlet polynya.

5.1.4 Lake storage effects on winter discharge

The volume of water stored as surface water in lakes can contribute substantially to total winter discharge below a lake. The majority of this water will be released in early winter because the rate of discharge from a lake at high stage will typically be much greater than inflow into the lake whereas, as lake stage decreases, lake outlet discharge will diminish to a rate approaching that of lake inflow. As a result, lake storage would be expected to have minimal effects on late winter low flows. Ice formation at the lake outlet can delay lake storage release and therefore increase late-winter low flows. Snow accumulation on lake ice can also increase late-winter low flows downstream of a lake.

5.1.5 Stage-discharge relation under an ice cover

The relation between stage and discharge under an ice cover is complicated by the effects ice conditions have on channel shape, channel slope, and hydraulic resistance. Generally stage has a positive relation with discharge, at a given cross-section, as an increase in stage is associated with an increase in area available for flow. However, during the stage-up process stage, over a reach of river, has a negative relation with discharge as water is abstracted from flow to satisfy channel storage requirements. Stage can also directly affect groundwater inflow through the process of hydraulic damming and tributary inflow by affecting channel slope at the tributary confluence. These processes can also cause a negative relation between stage and discharge.

The complexity of the stage-discharge relation under an ice cover may cause flow instability, as observed in the stage record and model residuals for M'Clintock River. The instability may be the result of a sequence of events triggered by an increase in stage at a given reach of river which causes a change in slope at that reach. The slope upstream would decrease while the downstream slope would increase. As a result, flow out of the reach is increased while flow into the reach is reduced. This would cause the stage to increase in the upstream reach and the process would be propagated upstream. Stage-up response is unlikely to be uniform along the length of the stream. Ice cover is not formed instantaneously along the length and breadth of the stream, which means that different stream reaches would be in a different phase of filling or draining channel storage at any given time. Upstream reaches may diminish or exacerbate the response at any given reach by controlling the rate at which flow is available to fill channel storage requirements.

5.2 RECOMMENDATIONS FOR FUTURE RESEARCH

The sub-Arctic winter, by virtue of the length of the period of sustained sub-freezing temperatures, presents an opportunity which is rare in the natural sciences, specifically, an extended period of time during which a natural process can be studied in the absence of most confounding variables. Cold weather effectively eliminates recharge (precipitation falls in a solid form which stays in place), transpiration (terrestrial plant life is dormant) and evaporation (water surfaces are sealed under a layer of ice). This opportunity could be used to control for comparisons of recession analysis techniques commonly applied to open-water data. By comparing a recession model determined objectively during the frozen season to results obtained using open-water techniques (e.g. Brutsaert and Nieber, 1977; Anderson and Burt, 1980; Petras, 1986; Nathan and McMahon, 1990) the validity of those techniques can be tested.

There is a pressing need for the development of a model for predicting winter discharge given the importance of low flows to environmental concerns. The development of an accurate model for predicting winter discharge will require more study of freeze-up related discharge depressions with attention given to developing an ability to predict the magnitude and duration of the discharge depression. The variability of recession model parameters from year to year will also need to be investigated.

Further work is required to test the hypothesis presented in this thesis for explaining the model residuals observed at M'Clintock River. This should include piezometric studies to confirm that groundwater mounding occurs causing hydraulic damming. A longitudinal study of stage and discharge along the length of a river should also be conducted. The purpose of the longitudinal study would be to look at variability in channel slope along the river, and the effect changes in slope have on discharge through affected reaches of the river.

The apparent contradiction between the importance of lake storage in low flow discharge in this study to the statistical analyses of Janowicz (1991) should be examined. Further water balance studies are needed to assess the relevance of intitial discharge and lake outlet discharge depressions to late winter low flows below a lake.

Appendix A M'Clintock River near Whitehorse, discharge measurement data

Date	Width	Area	Velocity	Stage	Discharge	Ice Thickness	Slush
	(m)	(m^2)	(m/s)	(m)	(m^3/s)	(m)	(m^2)
23-Sep	18.2	21.5	0.565	0.961	12.1		
08-Oct	18.5	24.8	0.617	1.104	15.3		
13-Oct	18.5	24.8	0.523	1.033	13.0		
18-Oct	18.3	24.4	0.532	1.002	13.0		
23-Oct	18.3	22.6	0.491	0.909	11.1		
26-Oct	18.5	23.6	0.462	0.907	10.9		
30-Oct	18.0	23.6	0.447	0.875	10.6		
05-Nov	18.6	20.2	0.368	1.336	7.43		
06-Nov	18.5	19.1	0.334	1.260	6.38	0.14	7.8
08-Nov	18.5	18.9	0.377	1.239	7.12	0.17	7.2
09-Nov	18.0	15.1	0.342	1.024	5.16	0.23	6.6
10-Nov	17.5	13.8	0.334	0.934	4.61	0.25	5.9
15-Nov	17.0	13.9	0.356	0.902	4.92	0.30	5.3
19-Nov	17.0	17.3	0.406	1.056	7.02	0.30	2.2
22-Nov	17.5	18.1	0.388	1.050	7.01	0.36	1.1
29-Nov	17.0	18.0	0.344	1.000	6.19	0.41	
04-Dec	17.0	17.6	0.313	0.955	5.51	0.38	
09-Dec	17.0	15.8	0.313	0.900	4.94	0.45	
16-Dec	16.0	14.4	0.312	0.865	4.50	0.45	
22-Dec	17.0	14.8	0.347	0.954	5.13	0.48	
28-Dec	16.5	13.7	0.320	0.939	4.40	0.53	
04-Jan	16.0	12.3	0.353	0.899	4.35	0.58	
10-Jan	16.0	11.6	0.335	0.888	3.89	0.59	
16-Jan	15.5	12.1	0.374	0.952	4.51	0.58	
20-Jan	16.0	12.7	0.358	0.945	4.55	0.57	

Continued

Appendix A, M'Clintock River near Whitehorse discharge measurement data
page 2

Date	Width (m)	Area (m^2)	Velocity (m/s)	Stage (m)	Discharge (m^3/s)	Ice Thickness (m)	Slush (m^2)
26-Jan	16.0	11.8	0.355	0.936	4.21	0.58	
01-Feb	16.0	11.9	0.323	0.882	3.84	0.57	
08-Feb	17.0	12.2	0.320	0.941	3.89	0.57	
Feb 15	19.0	13.0	0.284	0.933	3.70	0.47	
22-Feb	19.0	12.4	0.271	0.907	3.36	0.50	
Mar 1	20.0	12.8	0.281	0.923	3.59	0.46	
06-Mar	20.0	12.8	0.272	0.921	3.49	0.49	
13-Mar	20.5	12.6	0.274	0.903	3.45	0.49	
20-Mar	20.0	12.5	0.278	0.933	3.49	0.47	
27-Mar	20.0	12.1	0.273	0.898	3.29	0.48	
04-Apr	21.0	12.7	0.286	1.019	3.61	0.47	
11-Apr	21.5	14.6	0.283	1.143	4.14	0.44	
19-Apr	22.5	15.7	0.266	1.059	4.17	0.50	
26-Apr	16.0	20.3	0.341	0.883	6.92		

Appendix B   Ibex River near Whitehorse, discharge measurement data

Date	Width (m)	Area (m^2)	Velocity (m/s)	Stage (m)	Discharge (m^3/s)
12-Sep	10.2	3.69	0.720	3.895	2.66
11-Oct	10.5	3.52	0.650	3.870	2.29
17-Oct	10.6	3.73	0.606	3.886	2.26
17-Oct	10.6	3.70	0.594	3.886	2.20
24-Oct	11.3	3.19	0.643	3.846	2.05
31-Oct	10.9	2.97	0.578	3.821	1.72
02-Nov	10.8	3.18	0.460	3.838	1.46
03-Nov	10.6	2.85	0.530	3.825	1.51
05-Nov	10.6	2.95	0.551	3.820	1.65
06-Nov	10.6	2.95	0.411	3.826	1.21
07-Nov	10.4	2.75	0.551	3.814	1.51
08-Nov	10.4	2.64	0.505	3.806	1.34
09-Nov	10.4	2.77	0.363	3.822	1.01
10-Nov	10.4	3.70	0.372	3.917	1.38
15-Nov	8.80	4.87	0.222	4.071	1.08
17-Nov	5.15	3.76	0.308	4.213	1.26
18-Nov	10.6	5.08	0.286	4.104	1.45
21-Nov	10.2	6.13	0.201	4.162	1.23
26-Nov	10.0	3.54	0.302	3.899	1.07
28-Nov	9.50	3.13	0.313	3.887	0.978
29-Nov	9.50	3.13	0.326	3.874	1.02
05-Dec	9.50	3.30	0.314	3.852	1.04
07-Dec	9.50	2.54	0.339	3.842	0.861
12-Dec	9.25	2.90	0.365	3.867	1.06
15-Dec	9.50	2.31	0.360	3.814	0.830
19-Dec	9.00	2.32	0.356	3.826	0.828

Continued

Appendix B, Ibex River near Whitehorse, discharge measurement data page 2.

Date	Width (m)	Area (m^2)	Velocity (m/s)	Stage (m)	Discharge (m^3/s)
22-Dec	8.50	1.97	0.371	3.786	0.733
27-Dec	7.50	2.01	0.385	3.870	0.774
03-Jan	5.00	1.36	0.500	3.805	0.680
09-Jan	4.00	1.14	0.521	3.784	0.594
16-Jan	3.30	0.934	0.634	3.760	0.592
23-Jan	3.60	0.959	0.581	3.735	0.557
30-Jan	3.50	0.933	0.599	3.730	0.559
06-Feb	4.00	0.992	0.531	3.720	0.527
13-Feb	7.00	3.00	0.137	3.720	0.496
20-Feb	7.00	1.77	0.237	3.742	0.419
27-Feb	6.80	1.64	0.271	3.710	0.432
06-Mar	6.20	1.71	0.239	3.706	0.409
13-Mar	6.80	1.72	0.240	3.708	0.411
20-Mar	7.00	1.51	0.240	3.706	0.362
27-Mar	7.00	1.79	0.235	3.722	0.421
03-Apr	7.40	1.30	0.295	3.704	0.385
10-Apr	7.20	1.49	0.238	3.694	0.355
18-Apr	6.90	1.31	0.327	3.704	0.430
24-Apr	7.50	1.53	0.356	3.718	0.545
28-Apr	9.75	3.34	0.740	3.893	2.48

Appendix C Takhini River at outlet of Kusawa Lake, discharge measurement data.

Date	Width (m)	Area (m^2)	Velocity (m/s)	Stage (m)	Discharge (m^3/s)
16-Sep	74.0	134	0.666	1.881	89.4
05-Oct	72.0	121	0.668	1.860	80.7
12-Oct	72.0	123	0.649	1.867	80.2
19-Oct	69.0	119	0.603	1.757	71.7
27-Oct	66.2	107	0.540	1.600	57.6
12-Nov	59.0	89.1	0.426	1.301	37.9
01-Dec	39.0	32.2	0.538	1.098	17.3
10-Dec	33.0	24.8	0.792	1.037	19.7
14-Dec	34.0	23.8	0.888	1.035	21.2
23-Dec	34.0	22.5	0.842	0.991	18.9
30-Dec	32.0	20.5	0.807	0.969	16.5
05-Jan	30.0	20.2	0.750	0.930	15.1
11-Jan	31.0	19.5	0.722	0.904	14.1
17-Jan	30.0	19.7	0.748	0.884	14.7
21-Jan	30.0	18.4	0.728	0.871	13.4
31-Jan	30.0	19.2	0.748	0.835	14.4
07-Feb	30.0	17.8	0.747	0.801	13.3
12-Feb	30.0	17.8	0.729	0.785	13.0
21-Feb	31.0	17.0	0.688	0.749	11.8
28-Feb	30.0	16.9	0.696	0.720	11.1
07-Mar	31.0	16.0	0.623	0.672	9.97
14-Mar	31.0	15.5	0.613	0.689	9.48
21-Mar	31.0	15.9	0.579	0.676	9.22
28-Mar	30.5	14.5	0.621	0.656	9.03
04-Apr	32.0	14.8	0.617	0.639	9.10

Continued

Appendix C, Takhini River at outlet of Kusawa Lake discharge measurement data, page 2

Date	Width (m)	Area (m^2)	Velocity (m/s)	Stage (m)	Discharge (m^3/s)
11-Apr	31.5	14.8	0.603	0.639	8.90
19-Apr	31.2	14.1	0.588	0.624	8.31
24-Apr	31.0	14.8	0.587	0.632	8.70

Appendix D Takhini River near Whitehorse, discharge measurement data

Date	Width	Area	Velocity	Stage	Discharge	Ice Thickness	Slush Area
	(m)	(m^2)	(m/s)	(m)	(m^3/s)	(m)	(m^2)
13-Sep	73.0	130	0.862	3.681	112		
14-Oct	54.8	109	0.851	3.580	92.7		
25-Oct	52.3	96.1	0.752	3.356	72.3		
01-Nov	50.3	89.2	0.677	3.202	60.3		
07-Nov	50.8	92.2	0.663	3.305	61.2		
08-Nov	49.8	81.5	0.635	3.111	51.7		
10-Nov	48.8	86.5	0.672	3.255	58.1		
12-Nov	48.8	75.8	0.620	3.025	47.0		
16-Nov	53.6	112	0.313	3.800	35.0		
17-Nov	52.6	101	0.228	3.543	22.9		
18-Nov	56.6	117	0.433	3.925	50.6		
19-Nov	73.0	102	0.369	3.767	37.7	0.20	5.4
20-Nov	71.0	84.0	0.324	3.600	27.2	0.24	14.0
21-Nov	71.5	89.5	0.381	3.750	34.1	0.28	14.8
22-Nov	71.5	97.1	0.372	3.816	36.2	0.29	9.9
26-Nov	71.5	76.4	0.369	3.612	28.2	0.35	11.4
28-Nov	71.0	77.3	0.397	3.586	30.6	0.38	9.5
03-Dec	69.0	64.1	0.372	3.333	23.8	0.40	0.7
05-Dec	68.0	68.1	0.402	3.385	27.3	0.39	
07-Dec	70.0	61.7	0.391	3.290	24.1	0.40	
08-Dec	70.0	57.4	0.418	3.270	24.0	0.42	
12-Dec	66.0	46.4	0.467	3.144	21.7	0.48	
19-Dec	65.0	44.3	0.472	3.106	20.9	0.53	
27-Dec	61.0	34.7	0.540	3.037	18.7	0.60	
03-Jan	58.0	33.7	0.509	3.014	17.2	0.58	

Continued

Appendix D, Takhini River near Whitehorse, discharge measurement data page 2

Date	Width (m)	Area (m²)	Velocity (m/s)	Stage (m)	Discharge (m^3/s)	Ice Thickness (m)	Slush Area (m²)
09-Jan	66.0	29.6	0.506	2.992	15.0	0.64	
18-Jan	61.0	28.5	0.528	3.034	15.0	0.67	
23-Jan	61.0	29.6	0.532	3.042	15.8	0.66	
30-Jan	58.5	28.2	0.526	3.043	14.8	0.71	
06-Feb	57.0	27.5	0.513	3.026	14.1	0.69	
13-Feb	60.0	28.3	0.483	3.031	13.7	0.68	
20-Feb	43.0	25.7	0.482	2.985	12.4	0.69	
27-Feb	41.0	24.5	0.493	2.994	12.1	0.69	
07-Mar	41.0	24.0	0.506	2.995	12.1	0.66	
14-Mar	45.0	25.2	0.468	2.984	11.8	0.67	
21-Mar	45.0	24.7	0.475	3.002	11.7	0.67	
28-Mar	40.0	24.2	0.473	2.989	11.4	0.68	
03-Apr	40.0	26.5	0.421	3.007	11.1	0.67	
10-Apr	67.0	33.6	0.377	3.058	12.7	0.57	
Apr 18	67.5	42.6	0.321	3.055	13.7	0.38	
Apr 24	...	...	...	2.944	...	...	
Apr 25	43.0	57.1	0.345	2.903	19.7	...	

Appendix E Mendenhall River near the mouth, discharge measurement data

Date	Width (m)	Area (m^2)	Velocity (m/s)	Discharge (m^3/s)	Ice Thickness (m)
14-Dec	18.0	9.86	0.154	1.52	0.38
23-Dec	19.0	8.55	0.162	1.39	0.45
30-Dec	19.0	7.59	0.155	1.18	0.49
05-Jan	19.0	7.04	0.144	1.01	0.58
11-Jan	19.0	7.37	0.107	0.792	0.62
17-Jan	18.5	7.89	0.097	0.764	0.63
21-Jan	18.0	7.93	0.101	0.799	0.65
31-Jan	17.5	7.62	0.109	0.827	0.76
07-Feb	17.5	4.82	0.131	0.632	0.77
14-Feb	19.0	5.45	0.116	0.633	0.75
21-Feb	18.5	5.20	0.118	0.616	0.71
28-Feb	18.0	5.34	0.121	0.644	0.71
14-Mar	18.0	5.01	0.124	0.619	0.74
21-Mar	17.0	5.01	0.128	0.642	0.71
28-Mar	16.0	4.54	0.143	0.650	0.74
04-Apr	15.8	4.77	0.129	0.614	0.74
11-Apr	16.8	5.32	0.171	0.911	0.68
19-Apr	17.5	6.49	0.263	1.71	0.55

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