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Water quality dynamics in meltwaters draining Peyto Glacier, Alberta

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WATER QUALITY DYNAMICS IN MELTWATERS
DRAINING PEYTO GLACIER, ALBERTA

By

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THESIS

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Abstract

Variations in the quantity and quality of meltwaters draining Peyto Glacier, Alberta, are assessed to examine to what extent meltwater characteristics might be indicative of any distinctive source and routing of water through a glacierised drainage basin.

Continuous monitoring of solute concentrations of the principle melt-stream, Peyto Creek, was undertaken for significant periods in 1982, 1984 and 1987, coupled with measurements of suspended sediment concentration at the glacier terminus. Discharge of meltwater was determined for periods in 1982 and 1987. Supplementary monitoring of hydrogen-ion variation and total solute concentration was undertaken at additional sites within the basin, at a stream draining the accumulation area, and at a groundwater spring in the lower basin.

Determination of quantities of deuterium and tritium in different meltwater types in 1989 was used to supplement more detailed records of deuterium content collected in 1978. Records indicate different sources of water for groundwater and bulk meltwater flow, and variations in deuterium content at different elevation bands over the glacier.

Examination of short-term variations in solute concentration reveal distinctive behaviour of the glacial drainage net at times of maximum and minimum water pressure, when subglacially enriched pockets of water flow may become integrated with total flow.

Differences in the manner of water flow through the accumulation area in the course of the summer are indicated by contrasting water quality signatures of the upper stream between mid and late August 1987.

Dynamics of flow variations and suspended sediment concentrations appear stable during the principle periods of study in August 1982 and 1987, suggesting that at Peyto interesting results might be obtained from a concentration of research effort throughout the summer months following initiation of snow-melt.

Acknowledgements

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Results of fieldwork at Peyto in 1978, 1982 and 1984, which are discussed here, were funded by grants awarded to Dr. Collins by the National Hydrology Research Institute at Saskatoon. The Institute also provided excellent facilities at the Peyto base camp and unpublished discharge and temperature data. A summer at Peyto in 1989, which enabled samples to be taken for determination of isotope content was financed by an NSERC award to Dr. Gordon Young. Dr. Mike English provided funds for isotope analysis at the University of Waterloo. Dr. Scott Munro made available thermohydrograph charts for the summer of 1987.

I am very grateful for the guidance throughout this study provided by Dr. English and the contributions of Dr. Young and Dr. Houston Saunderson. Dr. Collins provided some insight into glacier hydrology and patiently indicated the value of adopting a versatile system for organisation of data during a visit to Manchester. I would also like to thank Dr. Peter Johnson for serving as outside examiner at short notice.

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Table of Contents

List of Figures

List of Tables

Chapter 1

Introduction

1.1 Background

Glacierised drainage basins have an important hydrological role in determining flow levels in many continental river systems. The majority of runoff is temporally confined to a short period over the summer which is made possible through seasonal storage of precipitation within the snowpack. Melting of the winter accumulation of snow from the whole basin and of glacier ice exposed by rise of the snowline through the summer produce seasonal flow cycles. On a shorter time scale rivers draining glaciers exhibit strong diurnal variations which reflect levels of radiation input. Drainage of subglacial and ice marginal bodies of water provide an additional source of runoff modification.

At the present time 10 per cent of the Earth's land area is permanently covered with ice, while a further 20 per cent is subject to seasonal snow coverage. This represents significant storage of fresh water within glacierised areas, drainage from which is a vital component in the world's hydrological budget. Although the majority of water stored is held within the polar icecaps and the Greenland ice sheet part of the recent rise in sea level has been attributed to an increased contribution of waters derived from the melting of alpine glaciers (Meier, 1984). At a time of concern over possible implications of climatic change and a global rise in temperatures it is essential that mechanisms of runoff variation from alpine glaciers be properly understood.

Glaciers tend to disguise minor climatic oscillations such that patterns of runoff variation may possess little obvious relation to precipitation levels (Meier and Tangborn, 1961). The presence of only a small glacier may alter the flow regime considerably (Krimmel and Tangborn, 1974). The proportion of the ice-melt contribution to river flows is greatest during dry summers, so compensating for any reduction in rainfall over the basin (Fountain and Tangborn, 1985). When summers are cool and wet, more precipitation falls as snow and ablation of glacier ice is reduced as the glacier is covered by a snow layer with high albedo. At the same time quantities of rainfall in the lower elevations of the basin will be higher. Changes in the extent of glacier cover may therefore produce changes in the flow regime of river systems (Johnson and David, 1987). Winter discharge from glacierised watersheds is negligible, consisting of water released from storage within the glacier and from groundwater. In spring discharge increases with snowmelt from the lower basin, storage in the snowpack and glacier remaining significant. These supplies of stored water are released progressively through the summer supplementing water derived from ablation of an increased area of glacier exposed by the rise in the snowline. The position of the snowline is determined by the amount and distribution of snow at the end of the winter and the temperature and precipitation record during the summer.

Critical timing of river flows and the proximity of glacierised areas to population centres in Europe, Asia, and North and South America has encouraged use of water from the melting of snow and ice for a variety of purposes, including irrigation and for industrial and domestic purposes (Meier and Roots, 1982). The high relief associated with young fold mountains has also permitted development of capital-intensive hydro-electric schemes in parts of Europe which has entailed diversion of glacial meltwater close to glacier portals and construction of adduction galleries to access subglacial sources of water (Bezinge, 1987). These schemes benefit from the regulatory manner of glacier flow cycles, which balance between wet and dry years, and the large differences in elevation which occur over small horizontal distances in high mountain areas, which provide ample elevation drops for water to drive turbines for electrical power generation (Tangborn, 1984). In order to increase utilisation of this resource it is necessary to understand thoroughly the particular character, timing and hazards associated with glacial streamflow as the demands placed upon the water resources provided by snow and ice melt are likely to increase through time.

Potential development in the high mountain environment is limited by the high frequency of active geomorphic processes as demonstrated by measurements of denudation rates. Continuous monitoring of discharge, suspended sediment and solute transport close to glacier portals, over short time scales, permit estimations of denudation rates before rivers have access to Quaternary deposits downstream. Measurements indicate the importance of subglacial erosion whereby waters flowing at the glacier bed provide efficient evacuation of the products of glacier abrasion (eg Collins, 1979a).

Sediment yield in rivers draining glacierised areas is thus much higher than in rivers with a non-glacierised catchment of comparable size (Walling and Kleo, 1979). Total daily suspended sediment transported by waters draining the 23 km^2 Peyto catchment, in the Canadian Rockies, in 1981 was estimated at 68,000 tonnes (2956.5 tonnes/km^2/yr) (Binda *et al.* 1985). While at Hilda Glacier, in the Canadian Rockies, area 2.24 km^2, annual suspended sediment transport was estimated at between 244 and 360 tonnes/km^2/yr (Hammer and Smith, 1983), illustrating how differing sediment supplies to meltwater systems, and different sampling strategies may produce widely varying results.

1.2 Objectives of Study

Investigation of water flow through glaciers is complicated by the inaccessibility of the glacier-bed which restricts the opportunity for study of 'within-glacier' meltwater characteristics. Study of variations in the chemistry of glacial meltwater, as it leaves the glacier, permits separation of meltwater on the basis of differing sources and pathways through the basin (Collins, 1979b).

These methods are applied to the basin of Peyto Glacier, a heavily glacierised basin in the Rocky Mountains of Alberta, where sedimentological and hydrochemical data have been collected since 1978, much of which remains to be analysed.

Studies of variations in the quantity and quality of meltwaters flowing at different sites within the drainage basin in different years are considered in an attempt to clarify the manner of water flow through the basin. The following questions are addressed:

1. The degree of change in hydrological characteristics of the drainage network among different years, and the extent of any differences in sediment supply to Peyto Creek.

2. Detailed examination of conductivity charts are undertaken to assess the role of short-term variations in water source and routing, as evidenced by records of total solute content, which have not been previously identified in the literature.

3. The extent of any differences in structure of the glacial drainage system over the area of the glacier and between the accumulation and ablation areas, in addition to any variation in response to similar meteorological conditions over the whole basin.

4. The practicality of using variations in the natural isotope Deuterium (^2H), and the radio-active isotope Tritium (^3H), as a tracer to isolate different sources of meltwater within the basin.

1.3 The Study Area: Peyto Glacier

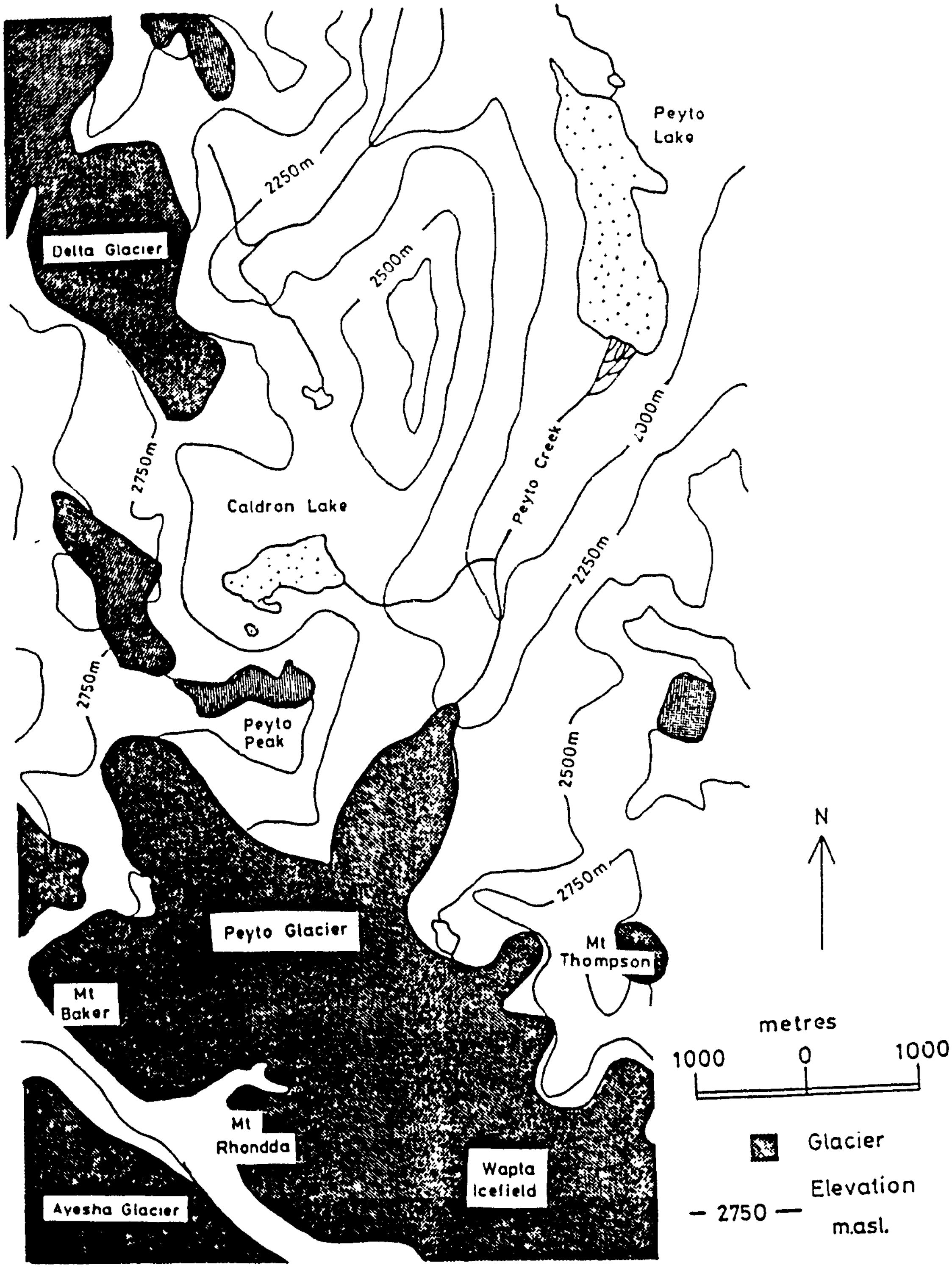
Peyto Glacier lies on the eastern side of the Canadian continental divide in the Waputik Range of the Rocky Mountains, Alberta. At latitude 51°40' North, longitude 116°34' West it is the most northerly outlet of the Wapta Icefield, 37 km North of Lake Louise (Figure 1.1).

The drainage basin containing Peyto Glacier covers an area of 21.9 km^2 above the location of a former Water Survey of Canada gauging station (No. 05DA008). Surveying of the glacier edge in the summer of 1989 indicated that the contiguous area of Peyto Glacier is 12.09 km^2. Additional isolated patches of ice within the basin cover an area of 0.45 km^2 giving a total ice coverage of 12.54 km^2, representing 57 per cent of the total basin area.

The lower ablation area of Peyto, 1.17 km^2 in area, is separated from the upper glacier by an ice fall over a dolomite rock outcrop at approximately 2400m.a.s.l. The ablation area descends from this point, at a gentle slope of approximately 6°, and currently terminates at an elevation of 2135m.a.s.l. The lower portion of the glacier is distinguished by a well-defined medial moraine which rises sharply from the glacier surface and which can be traced up-glacier to the slopes of Mt. Rhondda. Adjacent slopes consist of extensive ice-cored moraine which are partly maintained by debris received from frost shattering of local valley sides.

Above the ice-fall there is a broad, relatively flat, expanse of ice, with significant crevassing along the eastern margin and in the north-west. The glacier rises from this surface to a height of 3172m.a.s.l. at Mt. Baker, 3070m.a.s.l. at Mt. Rhondda, and 2970.5m.a.s.l. at Trapper Peak. These mountains conveniently divide the accumulation area into three separate basins; a broad col below Mt. Rhondda; the area between Mts. Rhondda and Baker; and between Peyto and Trapper Peaks. Small lateral moraines can be traced to the east and west above the icefall while several snow patches can persist for much of the year in small hollows and in areas above the position of the snowline at any point in time.

Figure 1.1 Location map of Peyto Glacier



To the east of the icefall, at a height of 2482m.a.s.l., there is a lake covering $0.043km^2$. The lake receives water from a small ice-marginal stream and melting of patches of stagnant ice in the vicinity. Water level is kept constant through drainage of water over a bedrock step, from which the water flows under the glacier to join the main meltwater channel. Peyto Glacier is drained by one stream, Peyto Creek, which flows into Peyto Lake, 3 km downstream of the glacier portal. The stream descends sharply 60m from the glacier snout over a resistant band of bedrock, it then flows for 100m beneath a collapsed ice-cored moraine before entering a deeply cut gorge where it is joined by a stream draining Caldron Lake, itself fed mainly by ice-melt. From the gorge the valley opens out and the Creek assumes a braided profile, it then passes through a dolomite outcrop before entering the outwash plain of Peyto Lake. Peyto Creek also receives contributions to flow from two small glaciers, Dragonback Glacier, an ice-drift glacier below Mt. Thompson, and a cirque below Mt. Jimmy Simpson. Isolated patches of perennial ice and snow also persist above the average position of the snowline, which is approximately 2600m.a.s.l.

Geologically the basin is dominated by sedimentary rocks, principally gently folded shale, which is interbedded with limestone, sandstone and dolomite of Cambrian age. These structures are the western remnants of a large eroded anticline which is centred on the Mistaya Valley. Sedgewick (1966) described how the basin can be divided into four distinct areas, each separated by an outcrop of resistant rock. Thus under Peyto Lake and the outwash fan bedrock appears to be Pre-Cambrian Shale. Above there is a riegel of Lower Cambrian bedrock which separates the next valley segment of St. Piran quartzite and some shales and limestones. There is then a steep dolomite-limestone ridge followed by a large outcrop of massive dolomites and limestone, above which the glacier occupies a wider portion of the valley, the bedrock again being interbedded shales and limestones.

Vegetation is generally sparse within the Peyto basin, most surfaces consisting of bare rock and scree with recent glacial and fluvioglacial deposits mantling the valley floor. Below the 2000m treeline a forest cover dominated by Engelman Spruce occupies the valley sides in the vicinity of the Lake. Significant vegetation communities are found in only two areas: in the meadows below Dragonback Glacier; and in the area around Caldron Lake where grasses and mosses cover the surface.

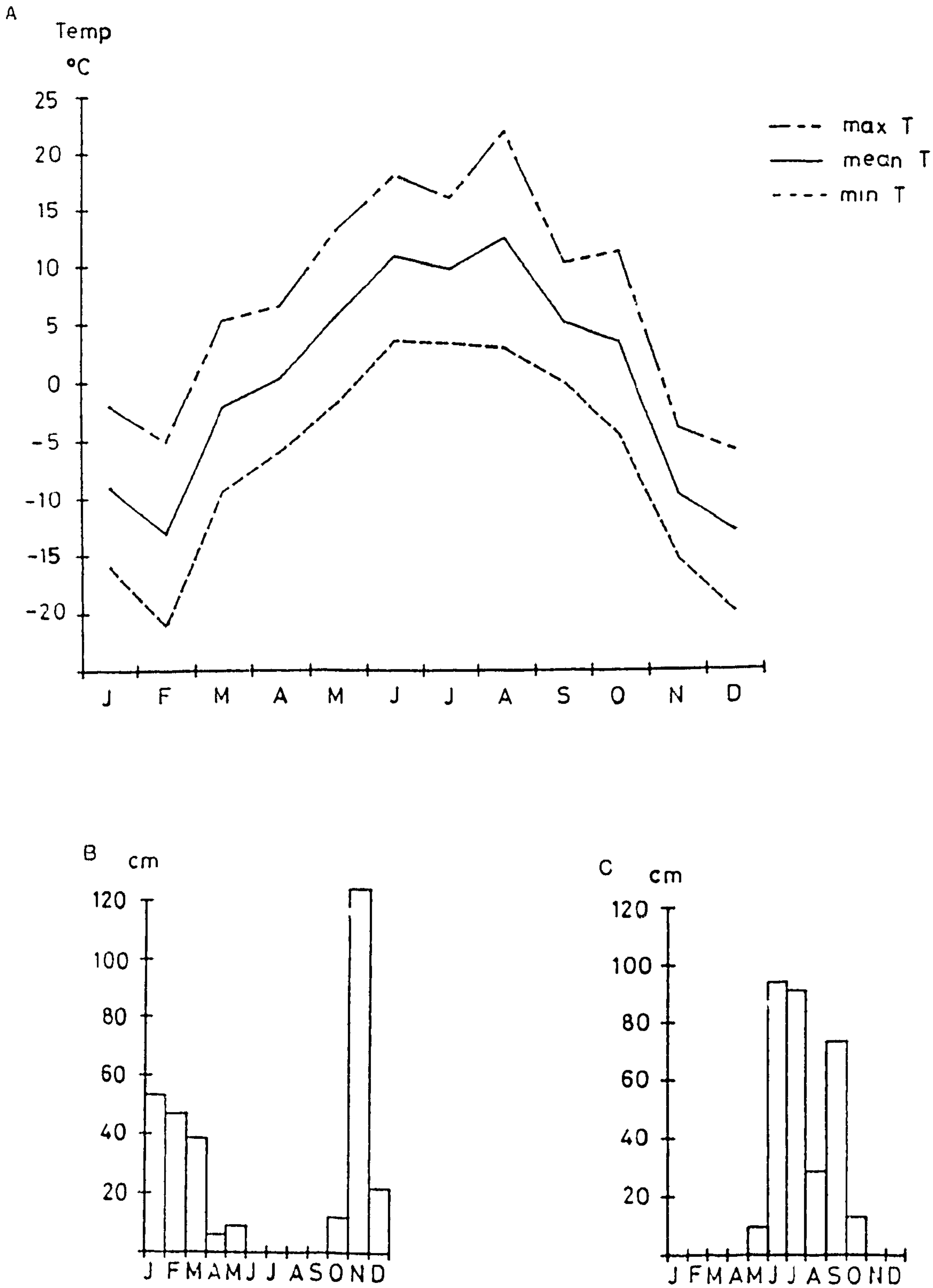
Peyto Glacier is situated 700 km from the Pacific Ocean, consequently climate is predominantly continental in nature. Topographic factors contribute to produce weather conditions which are very variable over short horizontal and vertical distances. The Coast Ranges and Columbia Mountains restrict the incidence of westerly air flows into the Peyto area, consequently there are large ranges in annual temperature together with a strong diurnal cycle. Precipitation is generally low but snow may fall at any time of year, depending on prevailing weather conditions.

Characteristics of the region's climate can be seen in Figure 1.2 with data for 1986 from the Atmospheric Environment Service meteorological station at Lake Louise (elevation 1534m.a.s.l.), the closest station to Peyto. Extremes of temperature range from 21°C in August to a minimum of -21°C in February. Mean temperatures remain above freezing from late April to late October, minimum temperatures were above freezing from mid-May to mid September.

Precipitation is well distributed throughout the year, part of the total falling as snow between January and March and then from October onwards. Rainfall was concentrated in the months May to October. Greatest monthly precipitation occurred in November, 1204mm, followed by high rainfall totals in June, July and September.

Although no meteorological data have been collected at Peyto throughout an entire year it is possible to obtain an approximate idea of weather conditions at this altitude. Given an adiabatic lapse rate of approximately 0.55°C per 100m in the region (Gardner, 1968), temperatures over the Peyto basin, where the glacier terminates at 2135m, 600m above the Lake Louise station, will be at least 3°C cooler. This disregards any particular local climatic effects that may arise due to the proximity of a glacier, such as katabatic winds.

Figure 1.2 Climate record from Lake Louise in 1986.
a) monthly maximum, mean and minimum temperature;
b) monthly snowfall; c) monthly rainfall.



Chapter 2

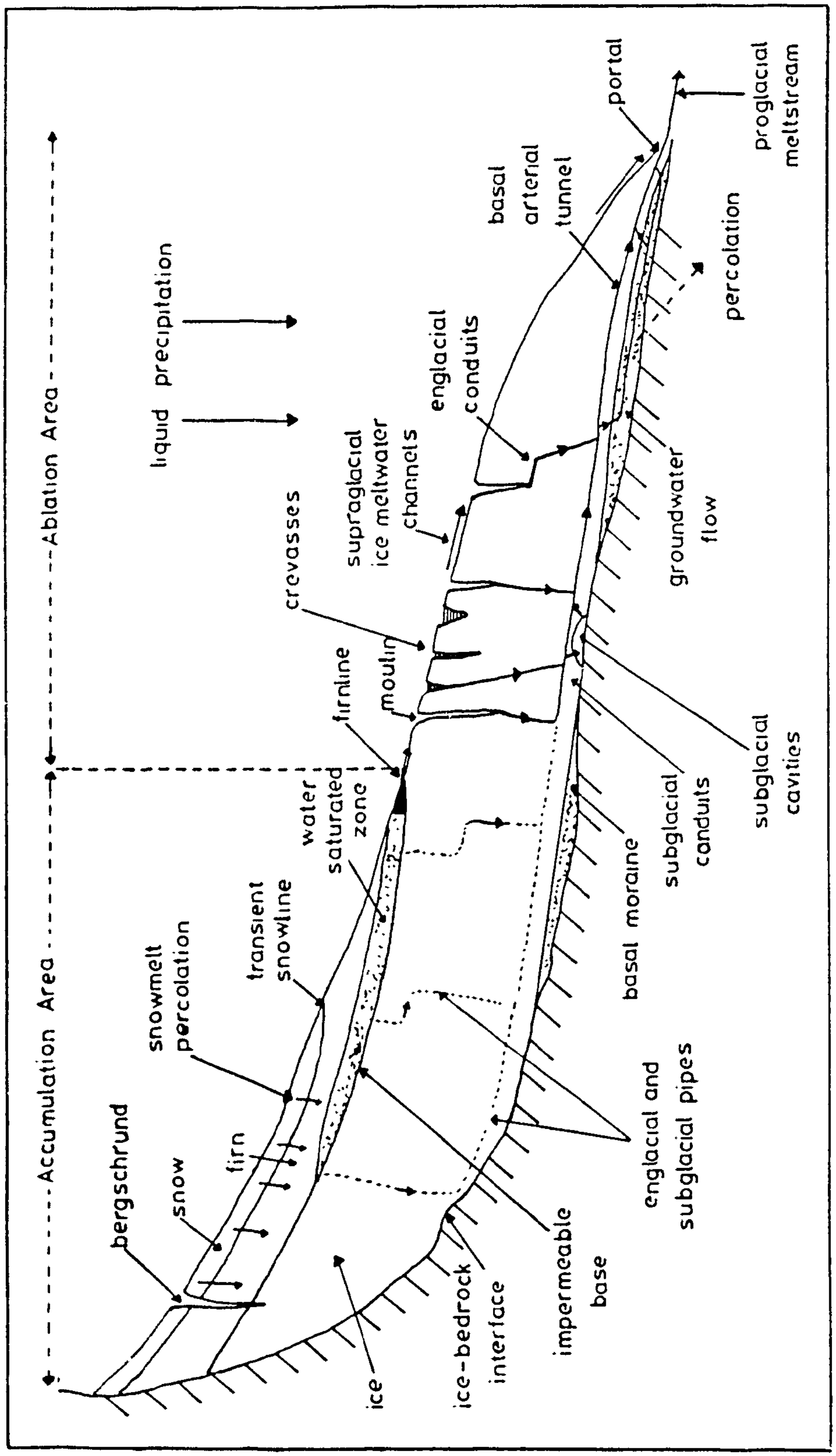
Theoretical Review

2.1 Water flow through glaciers

Discharge records of rivers draining glacierised basins show marked changes through time. Examining patterns of discharge variation helps to quantify the characteristic form of a river regime which arises as a consequence of the presence of a glacier within its catchment. Aside from a seasonal concentration of discharge within the summer months, river discharge from glacierised basins has a strong diurnal cycle during times of high ice-melt. This is a reflection of levels of incident solar radiation and air temperature within the basin. A strong daily peak in discharge occurs within a few hours of peak temperatures, which is superimposed on a slightly varying base-flow. Base-flow is composed of contributions from any groundwater sources within the basin, releases of water stored for various periods either within the ice, at the glacier-bed, or in the firn area, in addition to any sudden discharges arising from lake drainage (Röthlisberger and Lang, 1987).

A schematic cross-section of a glacier showing typical hydrological and glaciological features is presented in Figure 2.1. In alpine glacierised basins the majority of water leaving the basin originates at the glacier surface, derived either from rainfall or

Figure 2.1 Theoretical glacial cross-section (after Collins, 1988).



melting of glacier ice. Some additional water is produced at the base of the glacier where frictional and geothermal heating contribute to help basal ice-melt. The proportions of total flow derived from these two distinct areas of meltwater generation differ considerably, as illustrated by figures given by Röthlisberger and Lang (1987 p.233). Surface melt contributes in the order of 0.1 to 10 m yr^{-1}, while water from frictional and geothermal melting produces a total of approximately 10^{-2} m yr^{-1}.

The passage of meltwater through a glacier varies spatially over the glacier surface and temporally over the summer, depending on the location of a water source within the glacier's long profile and characteristics of the glacier itself (see Figure 2.1). Significant quantities of meltwater may be stored within the firn layer in the accumulation area which then follow different pathways through the glacier. Where the surface consists of snow and firn, at high elevations over the glacier, water may percolate slowly as it would through any unsaturated porous medium, obeying Darcy's Law, Q = kiA, where Q is discharge, k is the coefficient of permeability, i is hydraulic gradient, and A is cross-sectional area (Meier, 1973). Ice lenses and distinctive layers and fractures within the snow concentrate the flow direction of percolating water so that eventually only a few large channels reach the glacier-bed. In this respect crevasses intersecting the firn will be of importance, enabling a rapid routing of meltwater from the glacier surface. This is reflected in observations of water table variations in the vicinity of small crevasses (Fountain, 1989). The volume of water travelling through a glacier independently of the principal channels in this way may be considerable. Theoretical studies by Nye and Frank (1973) suggest that water may percolate along intergranular veins and capillaries at rates as high as 1 m^3 per m^2 of the ice surface per year. However the importance of Darcian flow is limited due to the presence of air bubbles within the ice which restrict meltwater percolation (Raymond and Harrison, 1975).

The majority of surface melt soon becomes concentrated in a supraglacial drainage network of meandering channels. Where the glacier surface is heavily crevassed an intricate network of streams may not develop. Water channels will descend crevasses or moulins thereby taking a fast route to the glacier-bed. At lower elevations of the glacier, distant from any crevasse fields, and in areas of high melt, surface channels may transport much of the total water volume. The distinctive meandering channel form assumed by supraglacial drainage networks has been studied for comparison with alluvial river systems (Ferguson, 1973). Mathematical treatment of the meandering process is helped considerably by the ability of supraglacial streams to modify their profile within the ice continually as flow conditions vary.

A classification of possible glacial drainage pathways, as either englacial or subglacial water routes, has been outlined (Shreve, 1972). Englacial channels are envisaged as being ice-walled with a low sediment and solute load, giving an opportunity for uninterrupted passage of water quickly through the glacial drainage system. In contrast, subglacial drainage of water occurs at much slower rates as meltwater percolates through the glacier in intimate contact with till and bedrock, sediment is available and there are opportunities for solute uptake.

A distinction has also been drawn concerning the characteristics of the principal conduits. Major channels at the bed may either be incised into bedrock (termed Nye channels after Nye, 1973) or cut into the overlying ice. In the latter case frictional melting generated from running water compensates any tendency to tunnel closure by deformation of the weight of overlying ice (R-channels after Röthlisberger, 1972). The precise form of the main glacial channels has important implications for the potential of flowing waters to access fresh sediment supplies.

Current concepts of glacier hydrology involve bringing together ideas of subglacial and englacial drainage routes in an integrated model of drainage pathways. A mixed cavity-conduit system at the glacier bed is envisaged as feeding into a few large arterial tunnels. This network of interconnected conduits and cavities at the glacier-bed permits more intimate contact with basal sediments and so is characterised as subglacial drainage. Arterial tunnels, even if at the glacier-bed, have only limited contact with sediment and provide fast routing of meltwater through the glacier, thereby being more characteristic of englacial drainage.

The presence of cavities at the glacier-bed has been discussed theoretically by Lliboutry (1968). Cavities are envisaged as forming on the down-glacier side of perturbations in the glacier bed where compressive forces from the weight of overlying ice are reduced. If water from the glacier surface enters these cavities, considerable fluctuations in basal water pressures are likely, reflecting variations in water supply from the glacier meltwater system (Mathews, 1964; Iken *et al.* 1983). Erosion of basal sediment and melting of overlying ice surfaces as water accumulates can further enlarge cavities at the glacier-bed. Iken and Bindschadler (1986) correlated fluctuations in subglacial water pressure with glacier movement. Pressures were determined by measurement of water level in bore-holes which made contact with the subglacial drainage system, while glacier surface velocities were obtained by frequent measurement of profiles of survey poles positioned across the glacier surface. This has led to clarification of mechanisms associated with what has been termed the alpine spring melt event.

Principal basal arterial tunnels close under ice overburden pressure through the winter when water flow is negligible. In spring resumption of ice-melt produces quantities of water which exceed the carrying capacity of ice-enclosed channels within the glacier. Inability of the main channels to transport water encourages water accumulation within basal cavities which produces an increase in subglacial water pressures. Ice velocities, as measured at the glacier surface, were strongly associated with times of greatest increase in water pressure (Iken and Bindschadler, 1986). Rapid glacier movement by these means is thought to be responsible for the phenomena of glacier surges. Only a minority of glaciers are thought to exhibit all features associated with glacier surges. The question of whether a glacier surges seems to depend upon the stability of the cavity drainage system in the face of a tendency for flow to become concentrated within major conduits. Kamb (1987) derived an empirical equation to determine whether a glacier is likely to surge. The equation was developed through consideration of the physical characteristics of the orifices separating adjacent cavities using a stability parameter ψ, which Kamb defined according to the equation:

$$\psi = a_w^{3/2} h^{7/6} (\eta/v)^{1/2} (P_i - P_w)^{-1/2} M^{-1}$$

Where a_w is hydraulic gradient, h is orifice step height, η is ice viscosity, v is sliding velocity, P_i is ice overburden pressure, P_w is subglacial water pressure and M is the Manning roughness. Kamb suggests that a cavity drainage system is stable against perturbations for $\psi < 0.8$. The value of ψ, at different times through the hydrological year seems to determine the nature of any transition between the surging and quiescent phases of a glacier.

Theoretical considerations therefore suggest the configuration of the drainage system at the glacier-bed is a vital component in glacier hydrology. Unfortunately difficulty of access to the glacier-bed has meant there are limited opportunities for measurement of processes in action. Advances in understanding of glacier hydrology have therefore depended heavily on the adoption of several complimentary techniques. In particular empirical models describing meltwater behaviour have been developed. Practical developments have relied upon the examination of areas of bedrock which until recently had been ice-covered, and the study of output series of variations in quantity and quality of water immediately upon leaving glacierised areas.

Theoretical analysis of the pressure distribution around major tunnels suggest that only water from a small area around the margins of a major channel will flow into the channel (Weertman, 1972). This has implications for the major water source supplying a glacier's principal arterial channels. The persistence of a tunnel system at the glacier-bed therefore depends upon reception of water which descends to the bed in quantity at specific locations. Where the channel persists at the same location in consecutive years a concentration of fluvio-glacial erosion will promote the development of Nye-Channels incised into the rock surface. Elsewhere in the glacier, water that is basally-produced may remain in a thin subglacial water film, perhaps 10 μm thick, which has been termed a 'Weertman film' after Weertman (1966).

Walder and Hallet (1979), Hallet and Anderson (1980) have studied deglaciated bedrock at Blackfoot Glacier, Montana and Castleguard Glacier, Alberta. Their descriptions of the selective erosion of these bedrock surfaces by glacier meltwater have provided further evidence of drainage pathways. At Blackfoot Glacier a drainage network of channels incised into bedrock links a series of leeside cavities which cover 20-30 per cent of the glacier bed. Significant quantities of water may be stored within the cavity system at the glacier-bed, storage beneath Castleguard Glacier was thought to be equivalent to a water layer 27mm thick over the whole glacier (Hallet and Anderson, 1980). Holmlund (1988) has explored the glacial drainage system by studying changes through time in the moulin system at Storglaciären, Sweden. A total of seventy moulins were studied, and seven descents down moulins were completed in winter when water flow was a minimum. The geometry of the moulins was found to be influenced by the amount of water passing through and the depth and width of any connecting crevasse.

More attention has been devoted to study of fluvial processes where glacier hydrology has benefitted from application of techniques which have been successfully used elsewhere within the hydrological sciences. Using recession curves Elliston (1973) determined that much of the meltwater from the Gornergletscher catchment in the Swiss Alps had a residence time of at least one day. In this catchment the possibility of any groundwater contribution was discounted given an impervious bedrock. Patterns of suspended sediment variation (Østrem *et al.* 1967; Østrem, 1975) and conductivity (eg. Collins, 1979b) have also been studied in attempts to clarify patterns of interaction between sediment and flowing water at the glacier bed. These methods have been expanded with the introduction of dye-tracing and measurement of isotope variation through time (eg. Behrens *et al.* 1971). Refinement of such techniques has permitted clarification of the intricacies of glacier hydrology which still need to be examined in greater detail.

The use of different investigative techniques provides support for theories of variable drainage pathways. Some water may be routed quickly through glaciers as evidenced by dye-tracing (Krimmel *et al.* 1973) while in certain situations meltwater may have a high residence time taking a slower route through the glacier. Meltwater may remain within the glacier over the winter, or drain over a period of days if the input of meltwater temporarily ceases (Stenborg, 1969; Elliston, 1973).

2.2 Sedimentological and hydrochemical characteristics of glacial waters

2.2.1 Suspended sediment variation.

Process based studies of glacial meltwater can be traced back to the work of Liestøl (1967), and Østrem *et al.* (1967) which were partly concerned with investigations of suspended sediment transport within an integrated study of the glacial environment. Suspended sediment is the portion of a stream's sediment load that is carried by upward directed velocity vectors in turbulent eddies. Transportation of particles will occur when shear stress, velocity or stream power are sufficient to initiate movement and overcome particle inertia. The distribution of suspended sediment is not uniform throughout the profile. Where flow is laminar the velocity of water is logarithmically distributed in the vertical, mean velocity occurring at a fraction, 0.6, of the depth (Richards, 1982 p.71). The distribution of suspended sediment in a vertical cross-section is then given by:

$$C_y = C_o \times \frac{y_o (d-y)^{w/kv*}}{y (d-y_o)}$$

where C_y is concentration at depth y (mg l^{-1}); C_o is concentration at reference depth o (mg l^{-1}); d is flow depth (m); w is particle fall velocity (m^2s^{-1}); k is Karman's constant (proportional to the depth of eddy penetration); and v* is shear velocity (m^2s^{-1}) (Richards, 1982 p.101,102).

Although derivation of this equation will not be given, consideration of the components clarifies variation of suspended sediment in a glacial context.

Strong vertical gradients are found for large particles (with a large w) and in conditions of weak turbulence (small v*). Glacial rivers have a strong tendency towards turbulent conditions. High velocities result from strong pressure gradients and little friction in ice-walled channels, consequently shear velocity may be high with a uniform distribution (eg Østrem, 1975). Significant levels of sediment transport within glacial rivers will also influence the value of k. Lateral variations in suspended sediment may still occur in glacial rivers. Metcalf (1979) noted variation of between two and eight times in suspended sediment within thirty seconds at a certain stream cross-section at Nisqually Glacier, Washington.

Further variation is introduced due to the high viscosity of glacial meltwater, which is 1.8 $mNsm^{-2}$ at 0°C compared with 0.8 $mNsm^{-2}$ at 30°C. This reduces particle fall velocity (w) and increases suspended sediment concentration as suggested by the above equation (Drewry, 1986 p.64).

Suspended sediment transport in glacial rivers follows a diurnal pattern similar to discharge variation after sustained ablation (eg. Østrem *et al.* 1967). Sediment levels in glacial rivers reflect the interaction between the fluvial system, which can both transport and erode deposits, and another sediment producing process. The relationship of suspended sediment transport and discharge is therefore not constant through time which limits the value of applied simple regression models. Short-term fluctuations in sediment transport, unrelated to discharge, have been described by Gurnell (1982). The glacial environment is typically one of high energy exchange and is in a perpetual state of change. Any increase in discharge, due to an outburst from an englacial, subglacial or ice-marginal water body may contribute to diurnal sediment peaks unrelated to daily discharge fluctuations (as in Collins, 1979a). Variations in sediment transport have thus been variously attributed to outbursts from glacially-dammed lakes, severe storms, and reorganisation of subglacial drainage channels (Collins, 1989). Simultaneous observation of sediment peaks and precise measurements of ice surface velocities have permitted quantification of characteristics of the route taken by meltwater through Variegated Glacier, Alaska (Humphrey *et al.* 1986).

Simple consideration of constraints on glacial systems suggest that quantities of sediment transport are determined firstly by the rate of sediment production, and secondly by the ability of the glacial drainage system to remove accumulated sediment. Sediment transport levels have thus been qualitatively related to the rate of change of discharge and the length of time elapsed since occurrence of a comparable discharge (Liestøl, 1967).

The importance of different sediment contributing events may vary considerably between years for the same glacier. The adjacent ice-free area is also a distinct part of the glacial sedimentary system. Working on two outlet glaciers from the Jostedalsbre ice cap in west Norway, Bogen (1980) considered runoff from moraines and slopes to be negligible. However at Storbregrova, in Norway, Richards (1984) described rainfall induced maxima in suspended sediment which were derived from ice-free areas. Gurnell (1982) found that the relative proportion of sediment contributed from areas currently glacierised decreased downstream where river waters could access old glacial deposits.

Given the interaction of the above factors the value of empirical rating curves of the form, $S_s = aQ^b$ (where S_s is suspended sediment concentration; Q is discharge; and a and b are best-fit parameters) is severely limited. However such equations have been applied to glacial rivers with varying degrees of success. A wide amount of scatter has been observed around the rating line. Such features have been observed in non-glacial catchments where factors such as rainfall direction, relative to basin aspect, and antecedent discharge are important (Walling, 1974). Discontinuous sediment supply resulting from a glacier's presence encourages hysteresis effects at several timescales (Bogen, 1980). Variable discharge in broad shallow channels enables temporary storage of sediment within the channel system. Clockwise hysteresis in sediment levels occur as a result of higher concentrations of suspended sediments on the rising limb of the diurnal hydrograph. At such times flowing waters are able to mobilise easily entrained sediments deposited by falling stage flows of the previous day.

Attempts to apply rating equations have continued due, partly, to the desire of hydro-electric companies to predict times of peak sediment concentration. This is essential for efficient control of water flow and minimisation of turbine damage from high sediment loads (Fenn *et al.* 1985; Bezinge *et al.* 1989; Fenn, 1989). Improvements on the simple rating equation have been possible with separation of the data series into distinct sections. These have been characterised on the basis of either periods of sustained ablation, or as times of falling stage (Østrem, 1975). On a different time-scale separate rating curves for the rising and falling limbs of the hydrograph have been used (Collins, 1979a).

The combination of a varying relationship between discharge and suspended sediment, and the likelihood of higher sediment content on the rising limb of the hydrograph, is likely to produce a regression model with residuals exhibiting high serial autocorrelation. To counter this Gurnell and Fenn (1984a) derived a Box-Jenkins transfer function, transforming the suspended sediment data logarithmically and introducing a one hour lag. This provided some improvement in forecasts of suspended sediment transport when compared with models obtained using other methods. A further model of suspended sediment transport was developed by Gurnell and Fenn (1984b), who applied the Collins (1979b) mixing model to separate total discharge into fast and slow routed components. The relationship of suspended sediment concentration to these different portions of total flow could then be assessed. Results of the application of these statistical techniques have still to be related to the causal processes in glacier hydrology.

2.2.2 Hydrochemistry of fluvio-glacial river systems.

Glacierised drainage basins are useful sites for chemical studies as limited regolith development in such areas restricts the production of solutes from biological and pedological activity. Given a knowledge of underlying geology and an understanding of chemical processes at the glacier bed, identification of different chemical species can help to identify solute sources within a basin. It should also be possible to determine the route taken by water through the basin. Initial solute concentrations are low and can be represented by typical atmospheric levels, water acquires additional solute on passage through the glacier as it contacts reactive particle surfaces. Any karstic groundwater system within a glacierised basin provides a further chemical environment which has not had to be considered in basins with impervious bedrock. Studies have tended to view any groundwater flow as an extension of the subglacial drainage network, characterised by waters having high solute loads, made possible by water having sufficient residence time to reach a saturation level in chemical content (Collins and Young, 1979a).

Uptake of solutes is determined by the availability of free hydrogen ions resulting from equilibration of waters with atmospheric carbon dioxide and oxidation of sulphide minerals. Further limits are provided by the supply of suitable debris and sufficient time of contact. Greatest rates of solute uptake are provided where water, saturated with hydrogen ions, has prolonged contact with reactive surfaces, freshly ground by glacier abrasion or rock shattering (Slatt, 1972).

Models of solute acquisition have been described by Raiswell (1984) who distinguished between open and closed systems applying concepts of differing drainage routes. Open systems remain in equilibrium with atmospheric carbon dioxide, waters pass through the glacier rapidly, retaining some free hydrogen ions. Solute levels are restricted by limited time of contact in englacial passages (Thomas and Raiswell, 1984). Closed system characteristics describe the chemical behaviour of slow routed waters, the number of free hydrogen ions within the water provides a limit on solute content in an environment where there is prolonged contact between water and rock surfaces. Support for the existence of closed system water at the glacier bed, which has a long residence time and high chemical loads, has been provided by observations of deposits of calcite and silica on areas of formerly ice-covered bedrock (Hallet, 1975; 1976). These are thought to be a product of regelation at the glacier bed which may induce supersaturation of basal waters and hence precipitation of these deposits.

Knowledge of hydrogen-ion, or proton, concentration is thus critical to understand the chemistry of snow and ice meltwaters. Few continuous measurements of pH have been attempted on glacial meltwater. Probes measuring pH are not designed to survive the turbulent flows and high sediment content found in glacial rivers. It is, however, essential to measure hydrogen-ion concentration at the time of sampling as changes in solutes can occur during storage (Slatt, 1972). Marked temporal differences in pH are likely, Metcalf (1984a; 1984b) observed that meltwaters draining the Gornergletscher in the initial spring event were oversaturated with carbon dioxide, pH increased rapidly with outgassing of carbon dioxide. In contrast, summer ice-melt was depleted in carbon dioxide, pH fell to equilibrium levels on encountering the atmosphere. Such differences in carbon dioxide appear to arise partly because of an increased solubility of the gas at the higher pressures which are likely within the glacier during accumulation of spring meltwater.

Identification of individual chemical ions in meltwater provides further clarification of subglacial processes. Sodium and potassium ions are derived partly from atmospheric sources but, being monovalent and having weak attractive forces, they can also be obtained through ion exchange at the surface of clay particles. Cations in water at the glacier-bed are retarded by electrochemical forces which promotes chemical sorting and solute enrichment of water (Souchez and Lorrain, 1975). Supporting evidence for this process was obtained by Lemmens and Roger (1978) who noted a rapid acquisition of sodium and potassium ions by meltwater flowing over glacial moraine at Tsidjiore Nouve Glacier. In contrast calcium and magnesium ions are divalent, the strong bonds can thus only be broken slowly through processes of acid hydrolysis and limestone dissolution. The metal cation is dissolved by a fixed quantity of hydrogen ions according to the equation:

$$CaCO_3\ (s) + H^+\ (aq) \rightarrow Ca^{2+}\ (aq) + HCO_3^-\ (aq)$$

The proportion of divalent ions is therefore likely to be greater in slow routed subglacial waters, where water has sufficient residence time for a slow acting process such as acid hydrolysis to be effective. Determination of the relative quantities of monovalent and divalent ions (ie. the ratio (Na^{++}+K^{++}/Ca^++Mg^+)) should be proportional to the different contributions of subglacial and englacial water within the river.

It seems that not all available cations may be taken up as dissolved load. Lorrain and Souchez (1972) noted that some cations were adsorbed on the surface of suspended sediment particles. Where significant quantities of suspended sediment are transported the proportion of cations adsorbed onto their surfaces may be considerable.

Study of variation in individual ionic species is expensive and requires a time-consuming sampling strategy if a continuous record is to be obtained. Consequently measurements of electrical conductivity have been used as an indicator of total solutes. This term describes the conductance of ions in Siemens per centimetre (Scm^{-1}). Characterisation of different waters by particular conductance values has enabled application of mixing models to describe the relative importance of different components of flow:

$$Q_s = Q_t \times \frac{(C_t - C_e)}{(C_s - C_e)} \qquad \text{(Collins,1979b)}$$

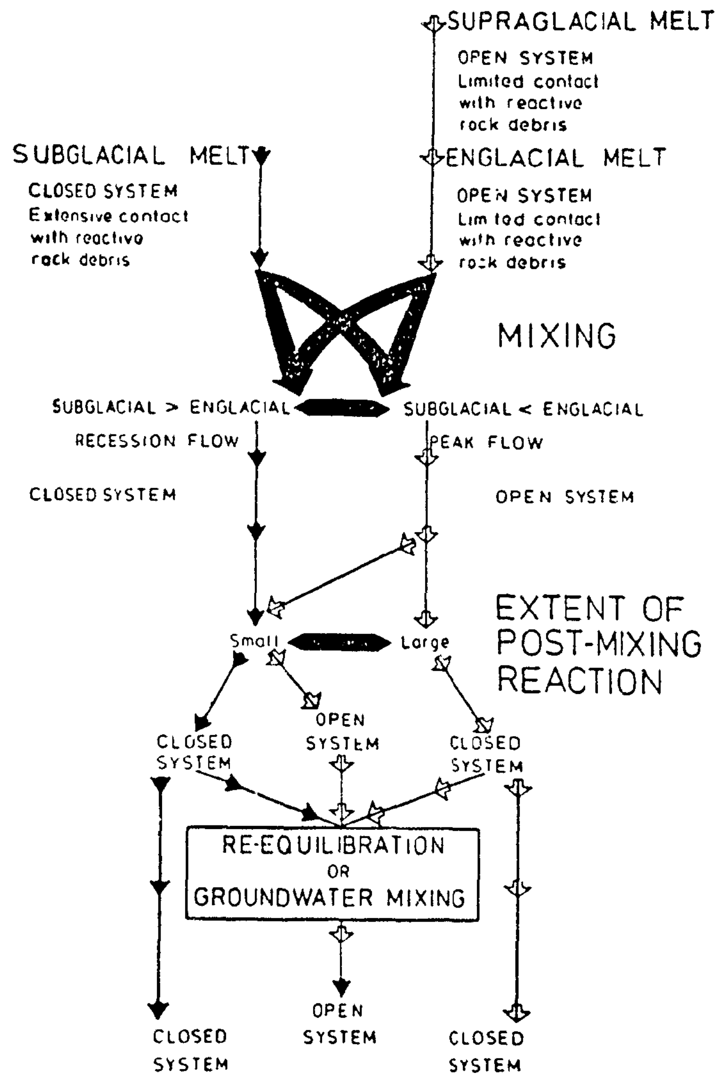
where Q_s and Q_t are subglacial and total discharges; C_t, C_e, and C_s are total, englacial and subglacial conductivities.

Separation of waters on the basis of electrical conductivity has permitted identification of differing drainage pathways beneath glaciers (Collins, 1979b). The englacial component of discharge, routed quickly through the glacier, provides between 50 and 80 per cent of total discharge and varies in phase with the characteristic diurnal discharge pattern. The relationship of slow-routed subglacial flows is more complex and can be either in phase or out of phase with the englacial contribution depending on the particular configuration of the basal drainage network (Collins, 1979b). At Findelengletscher, in the Swiss Alps, englacial and subglacial flows vary in phase, while at the adjacent Gornergletscher Collins observed an out of phase relationship. This seems to depend on variations in water present in the conduit and cavity systems, and the degree of connection between the two systems. At Findelengletscher there appears to be widespread connection between subglacial and englacial systems. At Gornergletscher limited connections result in storage of meltwater at high flows when water pressures are high, which are then released during daily minimum discharges. Some support for the results obtained from the mixing model approach has been obtained using different methods at Findelengletscher (Iken and Bindschadler, 1986).

Use of electrical conductivity (EC) does have limitations and for comparative purposes EC should be standardised to a reference temperature of 25°C. As EC is an aggregate measure of solute content it may not describe the fluctuation of individual ions satisfactorily (Thomas, 1986). Levels of electrical conductivity vary depending on source of solute, and with changes in flow routing of meltwater through the glacier (Collins, 1979b; Gurnell and Fenn, 1985). Meltwater from glacier ice is poor in solutes, subsequent EC of flowing waters depend upon the amount of contact with reactive surfaces. Temporal variation of EC reflects the varying contribution, through time, of water from different sources which have taken various routes through the glacier. Rating equations of the form $C_d = aQ^{-b}$ are consequently inadequate with the occurrence of similar multiple hysteresis effects to those observed for suspended sediment.

An essential part of the mixing model approach using electrical conductivity is the requirement that when water that has taken different pathways through the glacier meets, mixing is conservative. Thus waters derived from open and closed meltwater systems will maintain their integrity so that distinct values for conductance may be applied. The precise nature of any post-mixing reaction has still to be established. Raiswell (1984) has discussed the implications of the assumption of conservative mixing, the various possible combinations are summarised in Figure 2.2. Changing proportions of subglacial and englacial flow are found during recession flows and at times of sustained ablation. Mixing should produce bulk meltwaters with open system characteristics enabling further solute acquisition from suspended sediment and/or moraine material in the channel margins. Post-mixing reaction may therefore produce meltwater which tends towards closed system characteristics. However water will re-equilibrate with atmospheric CO_2, the increased supply of hydrogen ions will enable further dissolution of material. In addition supplementation of bulk meltwaters with further open-system water is likely downstream which will modify the characteristics of the total stream.

Figure 2.2 Schematic diagram showing possible variations of mixing behaviour in meltwater systems (after Raiswell, 1984).



2.3 Isotope studies in glacial hydrology

Variations in the level of natural and radioactive isotopes in waters within glacierised basins have been successfully used to isolate waters derived from snow and ice-melt, groundwater and summer precipitation.

Quantities of the radioactive isotope tritium (^3H) depend upon atmospheric fall-out on the glacier surface (Ambach *et al.* 1968). Significant increases in the tritium content of precipitation occurred following the first atmospheric thermonuclear tests in 1952, tritium contents subsequently decreased following cessation of atmospheric testing, and have remained approximately constant from 1972 (Behrens *et al.* 1979). Determination of tritium content enables calculation of the proportion of runoff derived from melting of glacier ice, formed prior to 1952, which is entirely free of tritium (Behrens *et al.* 1971). Absolute quantities of tritium in meltwater exhibit marked temporal variations as the water source varies, either daily or seasonally. Higher tritium contents occur in the summer months than in the remainder of the year. Tritium levels therefore remain low after the winter until rainfall or melt from summer snowfall becomes the dominant sources of water. Complications in this simple relationship have been observed as tritium content varies between different precipitation events and presents marked winter peaks, as shown by Behrens *et al.* (1979) for the years 1970-1977. The residence time of groundwater can be estimated through measurements of tritium variation. Ambach *et al.* (1973) noted that springwater near the Kesselwandferner basin in the Ötztal Alps of Austria had a winter peak in tritium, when precipitation has minimum tritium content, suggesting groundwater had a six month residence time.

Flow separation on the basis of the stable isotopes of oxygen and hydrogen, ^{18}O and ^{2}H (oxygen-18 and deuterium) relies on a fall in concentration of the heavy isotope with an increase in altitude and as condensation temperatures decrease (Moser and Stichler, 1975). Enrichment takes place through fractionation at phase changes during melting, freezing, evaporation, or sublimation, although Moser and Stichler (1980) recorded little isotopic fractionation, and hence increase in the heavy isotope, during melting or sublimation of glacier ice. Selective enrichment of certain isotopes occur due to a higher vapour pressure in water molecules which contain the heavy isotopes.

The two stable isotopes of hydrogen (^{1}H and ^{2}H) occur naturally in the relative proportions of 99.984 per cent and 0.016 per cent respectively, deviations from this ratio are considered with reference to a sample of standard mean ocean water. Higher levels of deuterium have been recorded in ice meltwaters and in summer rain and snow when compared with winter precipitation. As the summer proceeds deuterium levels in meltwater streams increase as the proportion of runoff due to ice-melt and summer precipitation increases. Groundwater systems seem to have deuterium signatures lower than that of ice meltwater due to fractionation during ice formation and the supplementation of the groundwater system by meltwater from winter snowfall with low deuterium concentrations (Ambach *et al.* 1976).

Combined analysis of tritium, deuterium and oxygen-18 has enabled description of temporal variations in runoff components (Moser and Ambach, 1977). Residence times of meltwater can be determined from seasonal changes in isotopic content of precipitation with a decrease in deuterium in late autumn and an increase in tritium over the winter.

Behrens *et al.* (1979) applied these methods and concluded there were two reservoirs of water within the Rofenache catchment (Ötztal Alps) draining over 100 days and between 3.6 and 4.8 years.

2.4 Peyto Glacier - Research in the I.H.D. and beyond

Peyto Glacier has been one of the most intensely studied Canadian glaciers. Early research is summarised by Ommanney (1972). From 1965 Peyto has been the subject of a detailed research programme as a representative glacierised basin within Canada's contribution to UNESCO's International Hydrological Decade Programme. Investment in base camp facilities at Peyto Glacier has enabled completion of a variety of research projects in addition to routine data collection.

The importance of glaciers in the Rocky Mountains as sources of streamflow for several Canadian continental river systems has already been described (Collier, 1958). Measurements of glacier mass balance at Peyto Glacier have been collected since 1964. These are described most recently by Young (1981) for the period 1965 to 1978 in which time an annual loss of 0.21 m water equivalence was recorded. The mass balance record has been correlated with meteorological conditions (Föhn, 1973; Young, 1977a). Letréguilly (1988) compared the Peyto mass balance record with adjacent weather stations and mass balance data from Place and Sentinel Glaciers. Place and Sentinel are located further west than the Peyto basin and hence are subject to a greater maritime climate influence. Letréguilly found that of the three glaciers the mass balance series at Peyto correlated most satisfactorily with summer temperatures at the Lake Louise weather station. Observations of differential ice ablation under debris cover were modelled by Nakawo and Young (1981), while a shortwave radiation model for the Peyto basin was developed by Munro and Young (1982).

Some work has been completed on isotope variations both within the Peyto basin and in the nearby river systems. Prantl and Loijens (1977) examined variations in the isotope tritium in the headwaters of the Mistaya and North Saskatchewan Rivers to demonstrate the importance of glacier-melt towards total stream-flow. Snow pits in the accumulation area of Peyto Glacier have been dug for analysis of the variation of deuterium and oxygen-18 in the annual accumulation layers (Hislop, 1974; Krouse *et al.* 1975). The considerable variation in isotope content at Peyto has been attributed to the effects of the Chinook wind in the vicinity (Krouse, 1974).

Records of summer streamflow in Peyto Creek were maintained from 1967 until 1982, characteristics of which have been described in Young (1977b; 1980). Loijens (1974) examined how meltwater flow from different sub-catchments within the 250km^2 Mistaya basin are integrated to produce total flow. In certain years more comprehensive hydrological programmes were undertaken at Peyto although work has been curtailed recently due to instabilities in an ice-cored moraine above Peyto Creek. This has produced a series of flood events one of which, in June 1983, has been described by Johnson and Power (1985). Recording instruments were destroyed in this event and again in a similar event in 1984.

Collins and Young (1979a; 1979b) applied the Collins (1979b) mixing model to Peyto Glacier observing an in-phase variation of subglacial and englacial runoff components at Peyto. This is similar to observations at Findelengletscher in the Swiss Alps, but contrasts with an out-of-phase relationship at the adjacent Gornergletscher. This suggests greater interlinkage with an underlying groundwater system at Peyto, compared with the Gornergletscher. Collins (1982) monitored the dispersion pattern of Rhodamine dye injected into a moulin on the glacier surface, observing a discharge-dependant diurnal pattern of flow. Measurements of suspended sediment transport and total chemistry in Peyto Creek have been described by Binda *et al.* (1985).

The high proportion of glacier cover in the Peyto basin has enabled studies of the effects of glaciers on streamflow through comparison with the adjacent unglacierised catchment of the Amiskwi River in the Yoho National Park (Collins and Young, 1981). Different regimes of flow for ice and snow-fed rivers could thus be described using hydrochemical records from the two rivers.

The complex inter-relationships in runoff generation observed in glacierised basins present the hydrologist with a considerable modelling problem when attempting to forecast flow. In view of the dependence of the Canadian Prairies on water sources in the Canadian Rockies, the combination of high glacial cover, the existence of stream discharge records and summer meteorological data from the Peyto basin, attempts have been made to develop predictive models of discharge. Derikx (1973) simulated discharge by describing water flow through the glacier by analogy with groundwater flow using a partial differential equation based on hydraulic conductivity and rate of change of the water table relative to an impermeable layer above a given reference level (Boussinesq's equation). Power and Young (1979) examined the suitability of the UBC watershed model for forecasting flow from glacierised basins by using data from the Peyto basin. Fortunately this model considers specific elevation bands within a basin, which is important in snow and ice hydrology as the position of the snowline at any time is an important determinant of meltwater processes. The model also includes a glacier-melt component which was modified to account for the actual proportion of glacier cover within the Peyto basin.

A further model to predict runoff from glacierised high mountain basins has been described by Gottlieb (1980). In this case the different heat exchanges are derived using a degree-day approach, having distinguished between ice-covered and ice-free portions of the basins. Water is envisaged as passing through a series of different temporary storages as it flows through the ice-free portion of the basin, whilst in areas of ice-cover routing of meltwater occurs through one linear reservoir.

Peyto Glacier has thus been the subject of a wide range of research projects in recent years. It represents an ideal site for scientific study, being readily accessible, possessing wide expanses of crevasse-free surface, and from a hydrological perspective, the geology is responsible for widespread presence of groundwater springs. The latter provide a third component to any model of glacier hydrology and a useful addition to the characterisation of meltwater flows as either subglacial and englacial in nature.

Chapter 3

Measurement Programme

3.1 Hydrological Procedures.

The work described below was undertaken as a supplement to routine collection of standard meteorological and mass balance data for the Peyto basin. In 1982, 1984 and 1987 a range of hydrochemical and sedimentological data were collected at sites throughout the catchment. The location of measuring sites is illustrated in Figure 3.1. A summary of data records obtained from the three years is presented in Table 3.1. Unfortunately records from 1984 are severely limited due to destruction of the main gauge in August 1984.

Records of river stage were collected at a site approximately 1km downstream of the glacier terminus at an elevation of 1960 m.a.s.l. (A on Figure 3.1). After Peyto Creek emerges from a collapsed ice-cored moraine, and above the confluence of Peyto Creek with the stream draining Caldron Lake, a natural stream cross-section occurs. A degree of stability of the channel margins is provided with bedrock on the left bank and boulder-filled gabion on the right. A stilling well was constructed here which was maintained as a Water Survey of Canada gauge and summer records of discharge were collected from 1967 to 1983 when the gauge was destroyed. Over this period stage was measured using

Figure 3.1 Location of study sites.

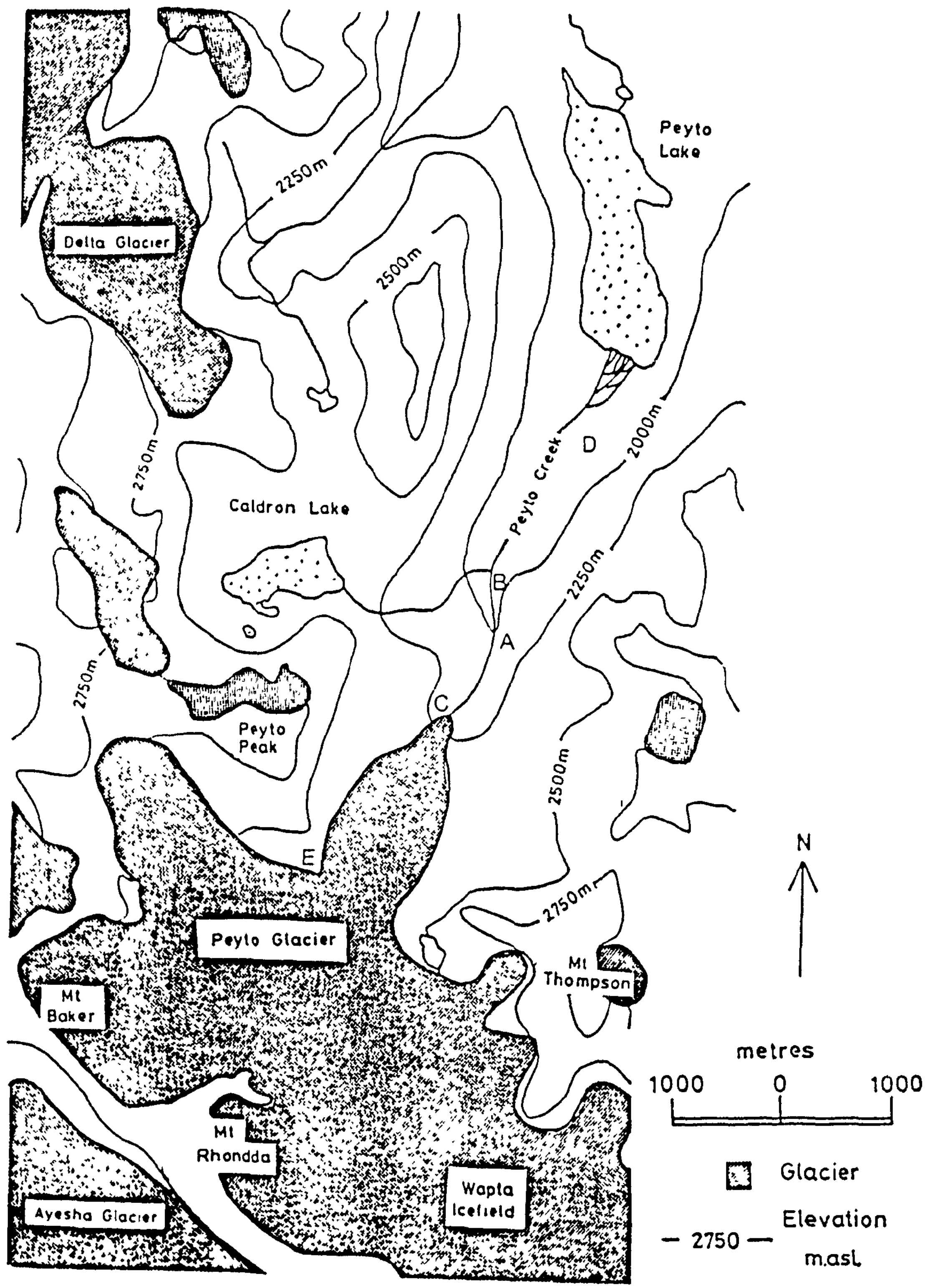


Table 3.1 Summary of measurement programme.

	Site	1982	1984	1987
	WSC Gauge.			
A.	Discharge	7/6 - 11/9	--------	---------
A.	Electrical Conductivity	21/4 - 16/10	16/9 - 16/10	---------
A.	pH	19/8 - 6/9	16/9 - 15/11	---------
	Gauge No. 2.			
B.	Stage (pressure transducer)	------------	--------	15/8 - 6/9
B.	Electrical Conductivity	-----------	---------	10/8 - 6/9
	Glacier terminus.			
C.	Suspended sediment	10 - 26/8	---------	10/8 - 30/8
	Groundwater spring.			
D.	Electrical conductivity	-----------	---------	10/8 - 30/8***
D.	pH	-----------	---------	9/8 - 31/8
	Waterfall site.			
E.	Elecfical conductivity	13 - 16/8	18/8 - 13/9	10-14/8; 22-29/8
E.	pH	------------	8/9 - 13/9	8 - 22/8

*** - Discontinuous readings

a Stevens Water Level Recorder. A stage-discharge rating curve was maintained with periodic recalibration using different methods which included salt dilution (Østrem, 1964) and dye-dilution (Church and Kellerhals, 1970). A gulp injection of salt solution was applied to Peyto Creek above the gauge before water descended a 15m high rock step in its course towards the gauge. The passage of the flood wave was detected through continuous measurement of electrical conductivity.

By 1987 changes in the channel configuration had occurred and it was decided to install a stage recording device approximately 200m downstream of site A (B on Figure 3.1). The channel margins in the vicinity appeared stable and there was sufficient protection to store a data logger within 3m of the Creek. Water stage was recorded at 30 minute intervals by monitoring the electrical current, in mA, produced by a Druck pressure transducer which had been secured within the channel. The electrical current generated was directly proportional to the pressure of overlying water which was itself assumed to be proportional to the head of water. The transducer had a compensating device to account for changes in atmospheric pressure. On two occasions in 1987 falling discharges during recession flow necessitated relocation of the pressure transducer. The probe was maintained within the same cross-section but moved down a certain distance in the vertical. This distance was measured precisely. A stage-discharge rating equation was not produced in 1987.

At both gauging sites, A and B, the electrical conductivity of waters within Peyto Creek was continuously recorded. Sproule electrolytic dip cells of carbon electrodes in resin probes were immersed in the turbulent flow of the stream. Conductivity was monitored by attaching a Walden Precision Apparatus CM25 conductivity meter to the cell. Measured values were recorded on a 6V Rustrak Chart Recorder. Conductivity cells were not calibrated with solutions of known conductivity. Time checks at frequent intervals helped to improve accuracy of the time series. When processing data subsequently compensation could be introduced to account for any changes in the rate of chart advance on the Rustrak Recorder. At this time also the conductivity cells were removed to ensure no sediment had become lodged in the probe.

Electrical conductivity measurements were not standardised to a reference temperature of 25°C. No measurements of water temperature were attempted, however in 1978 temperatures were recorded at three different streams within the basin (Collins and Young, 1979b). Water temperatures in Peyto Creek remained within the range 0.4-0.5°C. Glacial waters do not tend to exhibit marked fluctuations in temperature, consequently it should be possible to compare values of conductivity between different years within the Peyto basin.

Water leaving the glacier snout was sampled as close as possible to the terminus to avoid inclusion of water and sediment from the proglacial area. Effectively choice of site reflected a balance between the desire to obtain representative samples and the need to ensure equipment safety.

In all three years a Northants Automated Liquid Sampler was installed within 75m of the glacier terminus (C on Figure 3.1). It was programmed to collect hourly samples of meltwater, between 100 and 300ml in volume, through a fixed perspex tube. The range in volume arose through variation in the ability of sample bottles to maintain a vacuum. The intake was secured 1m from the stream bank at least 0.20m above the bed to restrict inclusion of saltating bed load. Distances were determined approximately in the field. Height of water above the intake varied between 0.1m and 0.4m. The sampler hose was removed periodically for cleaning purposes and any sediment lodged in the tube was removed at this time. Some error might arise due to the fixed orifice position and variations in suspended sediment through the river cross-section. The significance of this error was minimised by maintaining the sampler hose at the same point throughout the measuring period.

Sample volumes were accurately determined in the field using a measuring cylinder, and were then filtered in the field. Samples were passed through individually numbered and pre-weighed Whatman 542 hardened cellulose filter papers under pressure from a hand pump. The filter papers were stored in numbered, sealed polythene bags and were returned to the laboratory for weighing. Having air-dried the sediment samples, a Mettler balance, with infra-red heating attachment, was used to bake the samples until loss of weight due to evaporation became negligible whereupon weight was measured. Dry sediment weight was determined by subtraction of clean filter paper weight. Suspended sediment concentration, in mg l^{-1}, was obtained by dividing dry sediment weight by filtrate volume. No measurements were made of grain-size variation in the sediment samples.

Additional measurements of water quality were completed at other sites in the hope of identifying characteristic indicators of different meltwater types. Thus at a distance of 2km from the glacier terminus the electrical conductivity of a small groundwater spring was monitored in 1982 (D on Figure 3.1). The spring came to the surface in the flood-plain above Peyto Lake, on the eastern side of Peyto Creek. Conductivity was measured as before at Peyto Creek, the values were recorded on a Tinylog data logger. pH was monitored using a Russel gel-filled electrode which was standardised using reference solutions of pH 4 and 7. Values of pH were recorded on a 6V Rustrak chart recorder which was time-checked at regular intervals.

Similarly a further monitoring station was established in 1982 above a waterfall on the western side of the Peyto Glacier icefall at an elevation of 2349 m.a.s.l. (E on Figure 3.1). Measurements of electrical conductivity were attempted in 1982 and 1987, pH was also monitored here in 1987. The waterfall stream is composed of waters draining the western basin of Peyto lying between Trapper Peak and Peyto Peak. A stream emerges from the ice and flows for a distance of about 50m before descending a rock step as a waterfall. It then flows beneath the glacier and joins water which leaves the glacier via the main terminal stream. Experimental procedure was as above, although here values of both electrical conductivity and pH were recorded using a Rustrak data logger.

Subsequent data manipulation involved downloading data from the Tinylog recorders to obtain data in digital format which could then be analysed easily. Where data storage had been on Rustrak chart recorders it proved necessary to scan the charts manually so that analogue data could be transformed to a digital format. Readings were taken at " intervals on the charts which corresponded to a 30 minute time interval. Some compensation was introduced where declining voltages of the Rustrak batteries affected the rate of the chart advance. The amount of compensation required was determined by the space between successive time-checks which had been marked on the chart. In some instances choice of a " interval between readings severely limited observations of water quality variation at smaller time-scales. It was therefore decided to re-examine some charts, readings were averaged to produce a 15-minute timing between readings and selected values were modified where they diverged from the average. Where conductivities above 100 μScm^{-1} occurred it proved necessary to adopt a less sensitive scale on the conductivity meter so that conductivity values jumped in increments of five. This gave a step-like appearance to the resulting chemograph which was not seen in the original chart.

Values for the stage record from 1982 were read manually from the original NHRI chart. A rating curve for 1982 was unavailable and consequently the equation derived from stream gauging at intervals during the previous year was used. It is to be expected that whilst early season discharges, as obtained in this way, should be obtained accurately, the calculated discharges will become progressively more inaccurate as high flows in the course of the summer may change the precise form of the channel cross-section. Given the highly turbulent flows through the gauge it proved impossible to quantify any changes that might have occurred in this way. Peak discharges recorded by the NHRI during recalibration of the gauge in 1979 were of a similar magnitude to the peak flows obtained in 1982 suggesting a satisfactory transformation of the stage record.

The 30 minute records of stage in 1987 were averaged to give hourly values, in an attempt to limit the effect of standing waves and turbulence within the stream. It also proved necessary to adjust the stage record to account for relocation of the pressure transducer at times of falling water level as discussed above. Measurement of the distance the probe was moved enabled appropriate compensation to be introduced. As a rating curve was not produced in 1987 discharges were not obtained in this year: all results are thus presented on the basis of water stage.

Some meteorological data have been collected in the Peyto drainage basin as part of the routine measurement programme. The principal results of this work have been discussed elsewhere (eg. Foessel, 1974). To clarify the changing patterns of hydrometeorological conditions through the study period records of daily temperature and precipitation in 1982, and continuous thermohydrograph charts in 1987 have been analysed. These measurements were made at the main Peyto base camp (site F in Figure 3.1). The instruments were maintained in a Stevenson screen at an altitude of 2241 m.a.s.l., approximately 100 m above the position of the glacier terminus in 1989.

3.2 Isotope Procedures.

In an attempt to compare ideas of glacier hydrology as determined by records of water quality a variety of water samples from different hydrological environments were collected for determination of deuterium content. The environments were specifically chosen in the hope of identifying major flow components of water leaving the basin via Peyto Creek. Thus samples of snow and ice were obtained at different locations in the basin, in addition to precipitation at the Peyto base camp. Sampling of meltwaters in two supraglacial streams, of Peyto Creek at the glacier terminus and at the former gauge at site B, in addition to water in the groundwater spring (D) should help clarify patterns of progressive changes in contributions of different flow components.

Field procedures in 1989 depended heavily on work undertaken in 1978 by Collins (unpubl.). In July and August 1978 samples of water from different environments were collected for determination of deuterium content. A total of one hundred and forty seven samples were analysed by atomic absorption spectrometry at the University of Calgary. The mean percentage error was estimated at 1.8 per cent, the error ranging from 0.1 - 4.3 per cent. Calculated deuterium concentrations varied from a minimum of -173.7‰ in Peyto Creek during the ablation season, to a maximum value of -147.7‰ in snowmelt water. A summary of results from this study is given in Table 5.1. No samples of groundwater were collected and consequently it was not possible to separate different components of flow. In July and August 1989 samples of water were collected from a similar range of hydrological environments with the addition of the groundwater spring (site D). Samples were collected in 15ml scintillation vials and were returned to Waterloo where they were analysed by mass spectrometry for deuterium content at the Earth Science Isotope Laboratory at the University of Waterloo.

Deuterium levels were derived from :

$$\delta D = \frac{R_{sample} - R_{standard}}{R_{standard}} \cdot 1000‰$$

where R_{sample} and $R_{standard}$ are the ratios of deuterium to hydrogen in the sample and in a standard. The standard is taken to be mean ocean water (SMOW). This is a universally accepted standard, based on a US National Bureau of Standards distilled water sample (Craig, 1961).

3.3 Summary of data coverage and rationale for study

As Table 3.1 indicates a variety of different sources of data have been brought together for this work. While the quality of the records varies it is possible to examine variations in water characteristics and in the pattern of recorded discharges in 1982, 1984 and 1987. In attempting to explain any of the observed differences between the three different field-seasons it should also be possible to determine how the glacial drainage net might change in different years.

The 1982 season is especially valuable in extending the temporal coverage of electrical conductivity from April to October with pH also recorded upto mid-November in 1984. This should improve knowledge of water pathways through the Peyto basin at a time when the ice-melt component of runoff does not dominate discharge. Monitoring water quality at such times provides the parameters necessary in any mixing model designed to sub-divide flows on the basis of water source or pathway.

Supplementing the results derived from the hydrochemical studies with suspended sediment data in Peyto Creek and measurements of deuterium quantities in different environments within the Peyto Basin should further improve understanding of the hydrology of one glacierised basin. Daily variation of suspended sediment should indicate the manner of interaction of the glacial sediment producing process with meltwater flow through the glacier which completes sediment removal. Sampling of waters for natural and radio-isotope content represent an attempt to determine to what extent conclusions drawn from the variation of water quality parameters may conform with results from other techniques. Isotope methods appear ideally suited to glacier hydrology where fractionation effects at phase changes between snow and ice are likely to produce distinct isotope signatures for different sources of runoff. Application of radio-isotopes may also be invaluable in determining residence time of different waters within a glacierised basin. Such attempts are still necessary if the 'black box' approach of the hydrological modeller is ever to be improved.

Chapter 4

Flow dynamics in the Peyto Basin

4.1 Seasonal variation in meltwater quality and quantity

Discharge of water from a glacierised basin varies distinctly throughout the year. In Figure 4.1 mean monthly discharges recorded at the Water Survey of Canada gauge on the Mistaya River (No. 05DA007) are plotted for the years 1977-1986. The total catchment area, which includes the Peyto basin, is approximately $250km^2$, of which less than 13 per cent is glacierised. River flows exhibit a regular pattern of annual variation with considerable range in monthly discharges. Highest discharges occur in the summer months following initiation of snow-melt in May, discharges remain high until September as melting of glacier ice contributes to flow. This produces a distinctive annual flow regime which is typical of glacierised drainage basins. Peak mean monthly discharges range from a maximum of $17.4m^3s^{-1}$ in August 1977 to a maximum of $24.4m^3s^{-1}$ in July of the following year. Winter discharges are almost constant in the different years, the lowest monthly discharge of $0.303m^3s^{-1}$, occurring in March 1978.

Variations in water quality over the course of a year are strongly related to any changes in the quantity of water supplied. This can be illustrated with reference to Figure 4.2, where records of temperature, precipitation, discharge and electrical conductivity (EC)

Figure 4.1 Monthly mean discharge of the Mistaya River, 1976-1986.

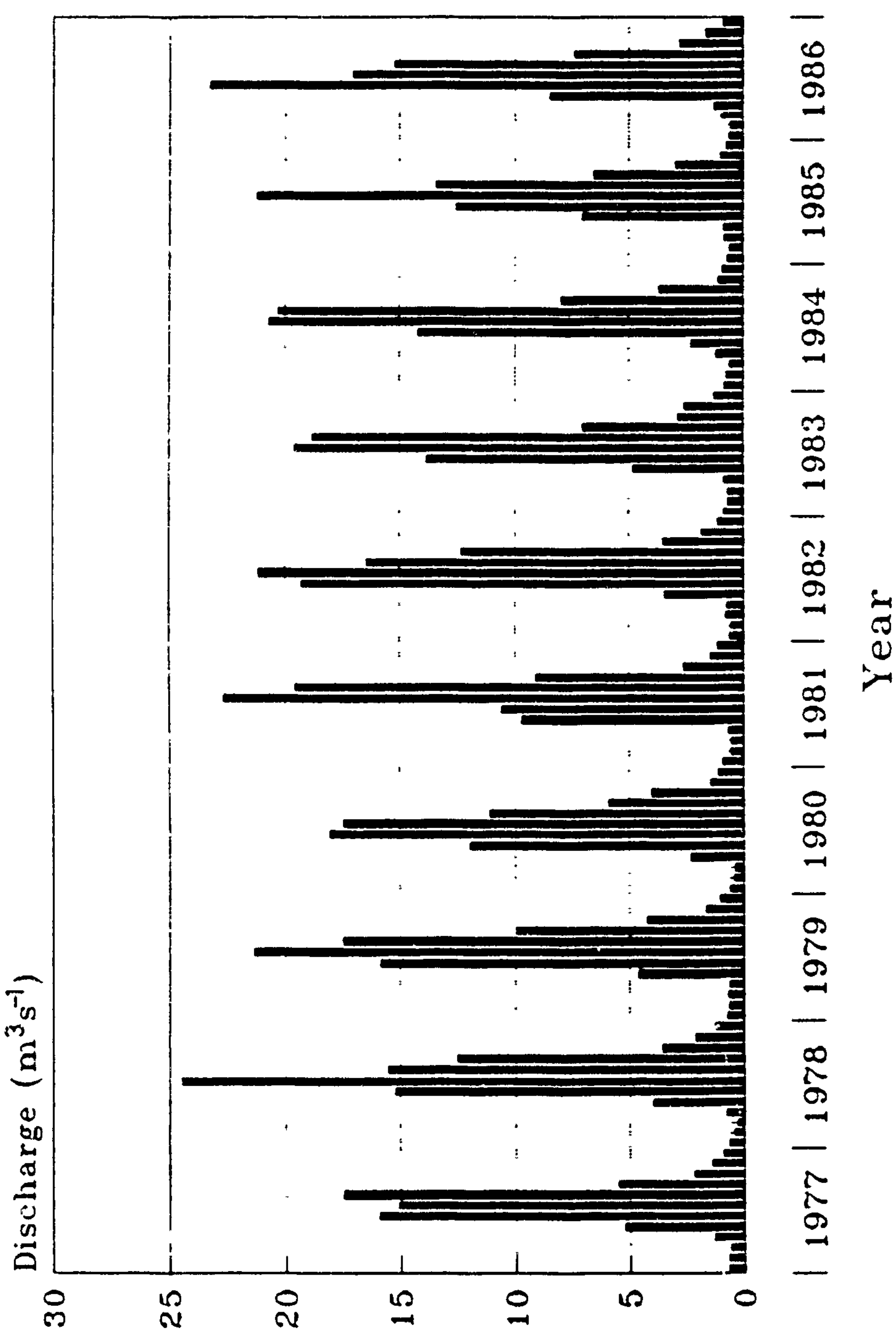
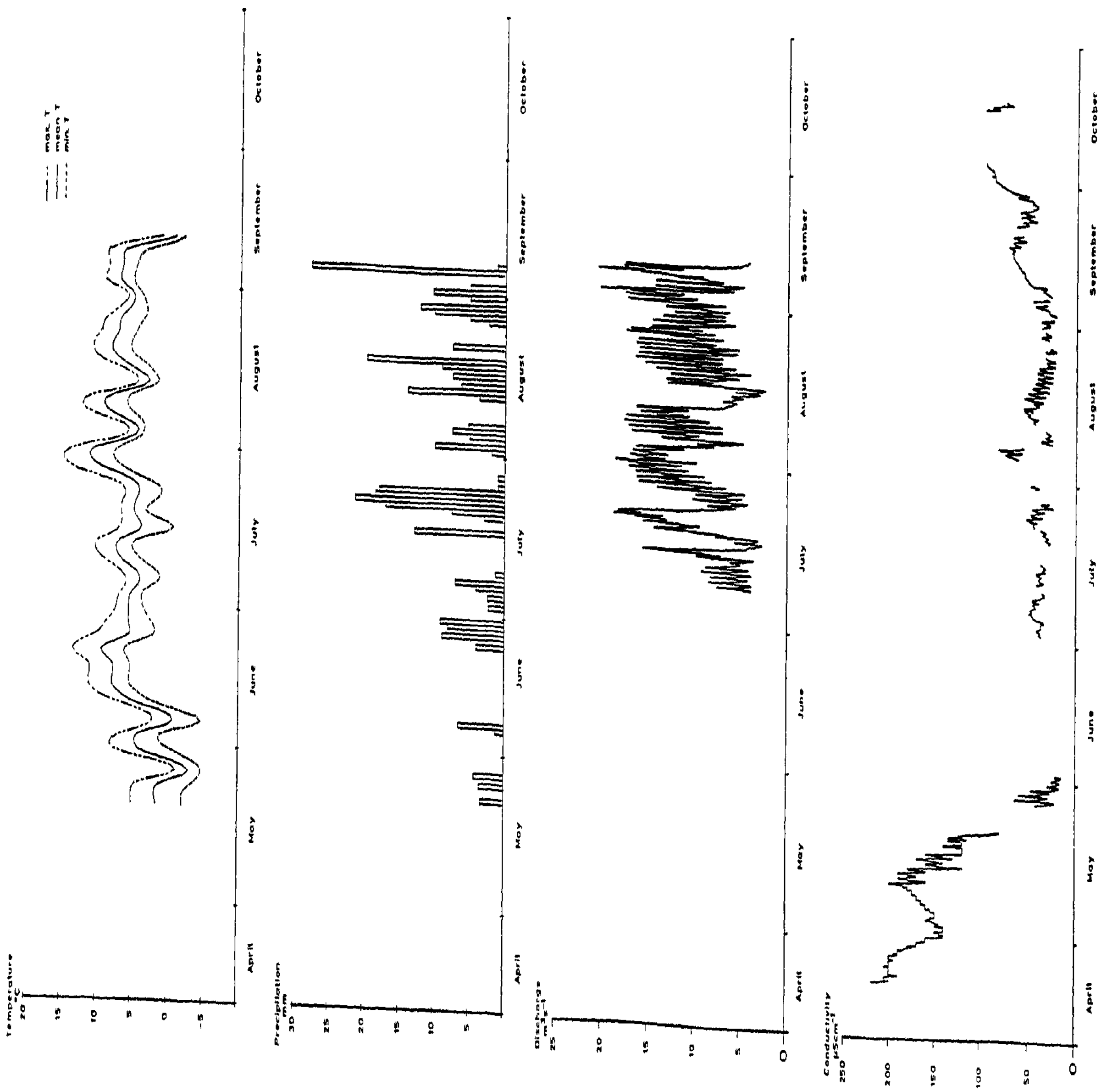


Figure 4.2 Temperature, precipitation, discharge and electrical conductivity records from Peyto, summer 1982.



from the summer of 1982 are presented. The longest series of records exist for EC, where coverage extends from April to October 1982. Breaks in the chemograph occur on several occasions when sediment had accumulated in the conductivity cell, or where the cell had become exposed by falling water levels in Peyto Creek. In June instrument malfunction led to loss of data from 3 June until the next visit on 30 June.

General levels of EC are inversely proportional to water supply. Maximum spring conductivities, which reached 220 μScm^{-1} on 21 April 1982, were recorded under conditions of very low discharge in Peyto Creek before melting of the snowpack had commenced. At this time solutes would have been contributed mainly by groundwater flows within the basin. Melting of the snowpack appears to have begun on 27 April when EC started to fall from 190 μScm^{-1} to a minimum of 145 μScm^{-1} between 2 May and 5 May. These values of conductivity, which were still relatively high when compared with solute content later in the summer, partly represent dilution of chemically enriched groundwater flow. In addition the first snow-meltwater leaving the basin in a year is likely to have higher solute concentrations due to rejection of solutes held within the snow-pack during changes of phase. Some water appears to flow within Peyto Creek throughout the year. The low discharges, before the main snow-melt period commences in May, will be composed of waters which would have had considerable potential for acquisition of solutes. This water might either be groundwater or old glacier meltwater which drains slowly from storage sites within the glacier and underlying sedimentary deposits.

A short period of cooler weather from 5 May onwards temporarily halted snow-melt and produced an increase in EC to a maximum of 200 μScm^{-1} on 11 May. The subsequent return of higher temperatures brought a resumption in snow-melt and hence a sharp fall in EC over the next ten days to 80 μScm^{-1} on May 22. Values then fell to a minimum of 15.5 μScm^{-1} as chemically dilute snow-melt waters left the basin from 27 May to 3 June. There was considerable diurnal variation in EC at the end of May which reached a daily peak of 64.5 μScm^{-1} on 28 May.

In July, when records resume, an increasing area of the glacier surface became exposed as the winter snow-cover melted. This increased the contribution of ice-melt towards the total discharge. Electrical conductivities were slightly higher than in early June, some diurnal fluctuation in solute content being evident through to mid-September.

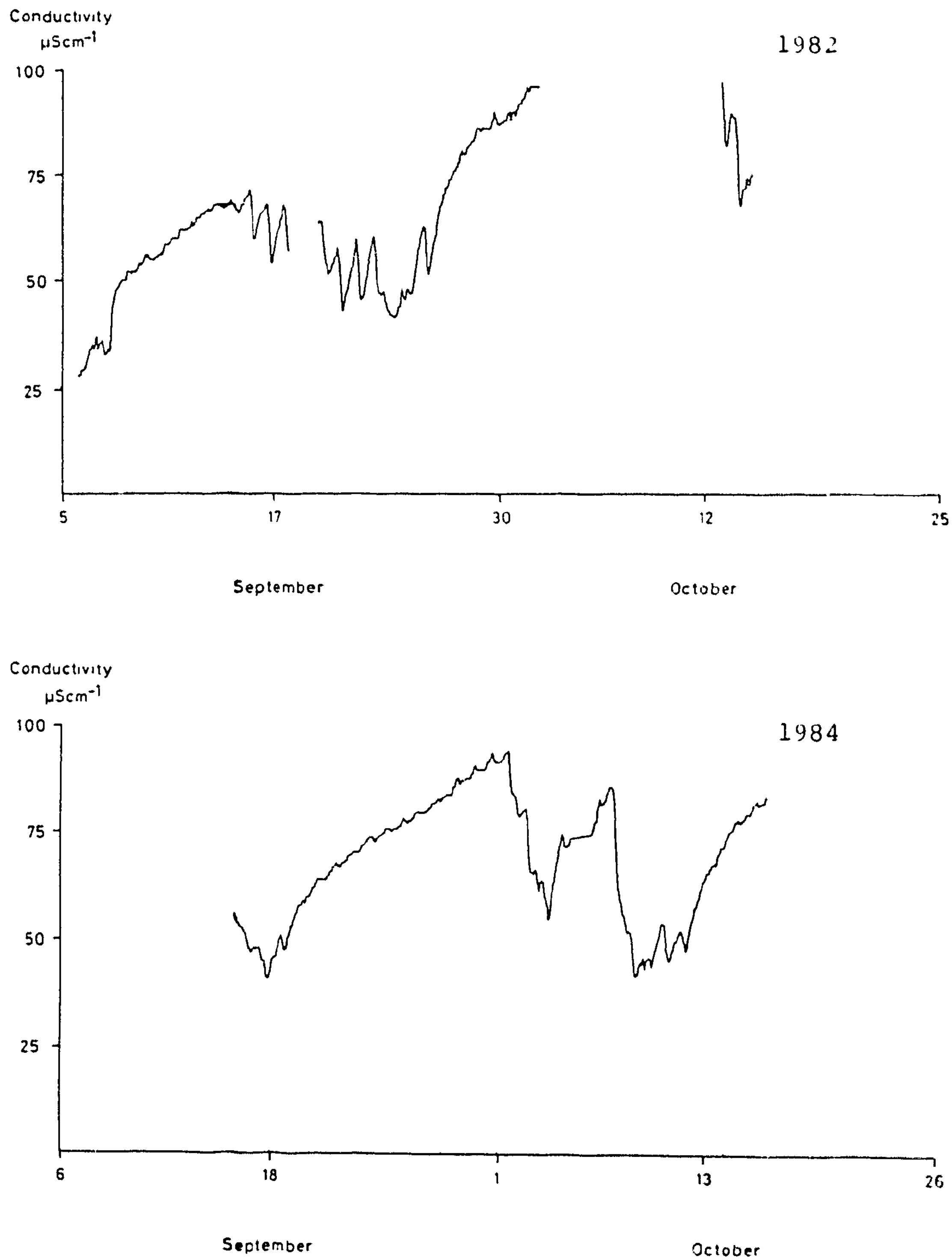
Solute concentrations over the summer are dominated by changes in the proportion of meltwater which passed quickly through the glacier and remained chemically dilute as opposed to meltwater passing slowly through the basin. The latter proportion of flow was stored temporarily within the basin, either in the firn aquifer, or in cavities at the glacier-bed. It is probable that a certain quantity of meltwater also passed through a groundwater drainage system within the catchment, in vadose passages extending over much of the basin area. Water which has a long residence time within the basin has greater potential to acquire solute, when compared with the daily totals of ice-meltwater which may pass quickly through the glacierised area.

During times of high ablation the time series of EC has a regular periodic form which remains out of phase with discharge. Whenever cooler weather conditions or snowfall limit melting of ice on the glacier EC rises as waters drain from the firn reservoir. Decreasing discharges during recession flows also limit dilution of the groundwater contribution towards total flow. Thus conductivity in Peyto Creek increased sharply, to 79 μScm^{-1}, in early August when recession flows, which followed six days of precipitation, enabled drainage of water which had been stored at different locations within the glacier. Rising temperatures at this time permitted the maintenance of a diurnal rhythm in solute content through a continuation in contributions from ice-melt.

From mid-September onwards recession flows, arising from snowfall on the glacier, produced a drop in discharge and an increase in solute concentration with no apparent daily fluctuation. Warmer conditions in the week from 16 September interrupted the period and EC again fell to a minimum of 41 μScm^{-1} on 25 September, with a daily maximum of approximately 60 μScm^{-1}. EC then rose steadily from 27 September to concentrations above 100 μScm^{-1}, which were too high to be recorded on the Rustrak chart. In mid-October the return of warm weather, which melted some of the recent snowfall, produced a fall in conductivity to a minimum of 68 μScm^{-1}. One daily cycle in solute concentration can be identified before the chart finished.

In Figure 4.3 EC of Peyto Creek in September 1984 is presented for comparison with measurement results observed over the same period in 1982. A similar pattern of an increase in EC during recession events as discharges decrease is apparent on two occasions in 1984. Higher temperatures around 3 October, and the consequent melting of snow at lower elevations, contributed to a fall in conductivity from a peak of 95 μScm^{-1} on 2 October to a minimum of 55 μScm^{-1} on 4 October. Reduced flows in Peyto

Figure 4.3 Electrical conductivity of Peyto Creek, September-October 1982 and 1984.



Creek from 4 October through to 9 October produced a temporary increase in EC to a peak of 86 μScm^{-1} on 8 October. A return to warmer temperatures encouraged the melting of recent snow falls, resulting in electrical conductivities which fell within the range 43-54 μScm^{-1} and exhibited some diurnal fluctuation . Cooler weather from 12 October led to a steady increase in EC to values above 80 μScm^{-1}.

Viewing variations in solute concentration at a smaller time-scale illustrates that minor fluctuations in quantity of meltwater flows still occur in September over a daily cycle. Slight daily oscillations in conductivity are evident throughout the period of record, albeit on a rising base level.

Examining patterns in the discharge and temperature records shown in Figure 4.2 provide evidence of several hydrometeorological cycles over the 1982 study period. These can be divided into times of sustained ice ablation which are followed by precipitation within the basin. Precipitation temporarily produces a sharp increase in discharge as rain falls at lower elevations within the catchment. Snowfall on the glacier's upper slopes inhibits ice-melt for several days as the ice-surface becomes covered by fresh snow. This increases albedo from a typical value of 0.6 for bare ice, to one of 0.9, for fresh snow. The correspondingly greater reflection of incident radiation limits the amount of energy available for melting snow on the glacier surface. Subsequent improvement of weather conditions produces a resumption of ice-melt and enables a return to the 'normal' diurnal discharge variations dominated by daily receipts of chemically dilute ice-meltwater. The amplitude of these diurnal fluctuations in discharge increases as meltwater from a greater area of glacier ice contributes to total discharge.

The role of temperature in determining discharge levels can be seen through comparison of the graphs of temperature and discharge in Figure 4.2. The start of the discharge record on 7 July corresponded with generally warm temperatures, which rose to a local peak of 10°C on 10/11 July. This is reflected by a daily increase in discharge to approximately $7m^3s^{-1}$, above a base flow level of $5m^3s^{-1}$. On 12 July temperatures fell sharply, almost entirely suppressing the normal rise in discharge. An increase in flow, to $16m^3s^{-1}$, over the following two days occurred as a result of 13mm heavy precipitation on 14 July. Assuming uniform distribution of precipitation over the basin the quantity of water arising from 13mm of precipitation could approach $284{,}700m^3$ which would produce a rapid increase in the Peyto Creek hydrograph. Discharges arising from ice-melt remained low, however four days of precipitation, which totalled 75mm from 17 to 21 July, contributed to an increase in discharge to almost $20m^3s^{-1}$. Only a small proportion of these maximum flows arose through large diurnal increases in discharge. Base-flows steadily increased over this time and a seasonal maximum overnight flow of $16m^3s^{-1}$ occurred on 19 July. After rainfall ceased discharge fell sharply but immediately began to increase as temperatures reached a seasonal peak of 18°C on 29 July. The increase in discharge consisted of steady daily variations superimposed on a base flow (here describing the overnight daily minimum discharge) which increased from under $5m^3s^{-1}$ to values above $10m^3s^{-1}$ as meltwater from a steadily increasing area of ice began to contribute to flow.

Several days of rain and snow in mid-August produced a further recession event and for three days discharge barely rose during the day. A resumption in ice-melt accounted for the increase in discharge to previous levels but the amplitude of daily variations increased considerably. This was a consequence of a steady decrease in the hours of daylight towards the end of the summer which temporally concentrated the period of peak ice-melt at the glacier surface. This, when combined with a more efficient glacier drainage net at the end of the ablation season, accentuated the amplitude of daily increases in discharge.

4.2 Short-term hydrochemical dynamics

The intricacies of glacier hydrology and the implications of variable source and routing of glacial meltwaters can be appreciated through consideration of variations in meltwater quality on short time-scales during periods of sustained ice-ablation. At such times regular pulses of meltwater, which have relatively consistent chemical content and which travel at a fast rate through the glacier, provide a constant modification to the chemograph. These meltwaters mix with water from different sources, such as that derived from basal melting, which do not form runoff immediately and may remain at the glacier-bed for a considerable time.

4.2.1. 1982 hydrochemical record from Peyto Creek

Electrical conductivity is plotted at a 15-minute time interval, together with hourly discharge, during a time of high rates of ice ablation from 7 August to 27 August 1982 in Figure 4.4.

In the graph of discharge in August 1982 the regular 24-hour cycle of water variation dominates much of the record. Discharges typically peak at about 16:00 hrs, with a daily minimum flow occurring at 10:00 hrs, although several features interrupt the regular cycle. Snowfall on the 10 August produced a recession event in which discharge decreased sharply to a minimum of $2.5 m^3 s^{-1}$ on the morning of 15 August. Some daily increase in flow occurred on 12, 13, and 14 August although on a falling base-flow. Distinct peaks in discharge can also be identified which fall outside the normal diurnal oscillation. These apparently arise due to rainstorms within the basin. Precipitation in the lower basin can flow rapidly over debris-covered surfaces and thus may be identified by an immediate rise in the hydrograph. Such local peaks in flow occurred on 11, and 12 August during the recession event, due to precipitation and again on 16, 25, 26 August.

A series of peaks in discharge occurred on 9 August, a day on which only traces of precipitation were recorded at the NHRI rain gauge. Discharge rose sharply in the early morning of the 9th, subsequently falling before increasing to produce a double peak at about 16:00 hrs, the time of the normal daily peak discharge. A further peak occurred overnight on the 9th which interrupted the falling limb of the hydrograph. Temperatures on 8 and 9 August 1982 were among the highest of the year. On both days the mean

Figure 4.4 Electrical Conductivity and Discharge of Peyto Creek, August 1982.

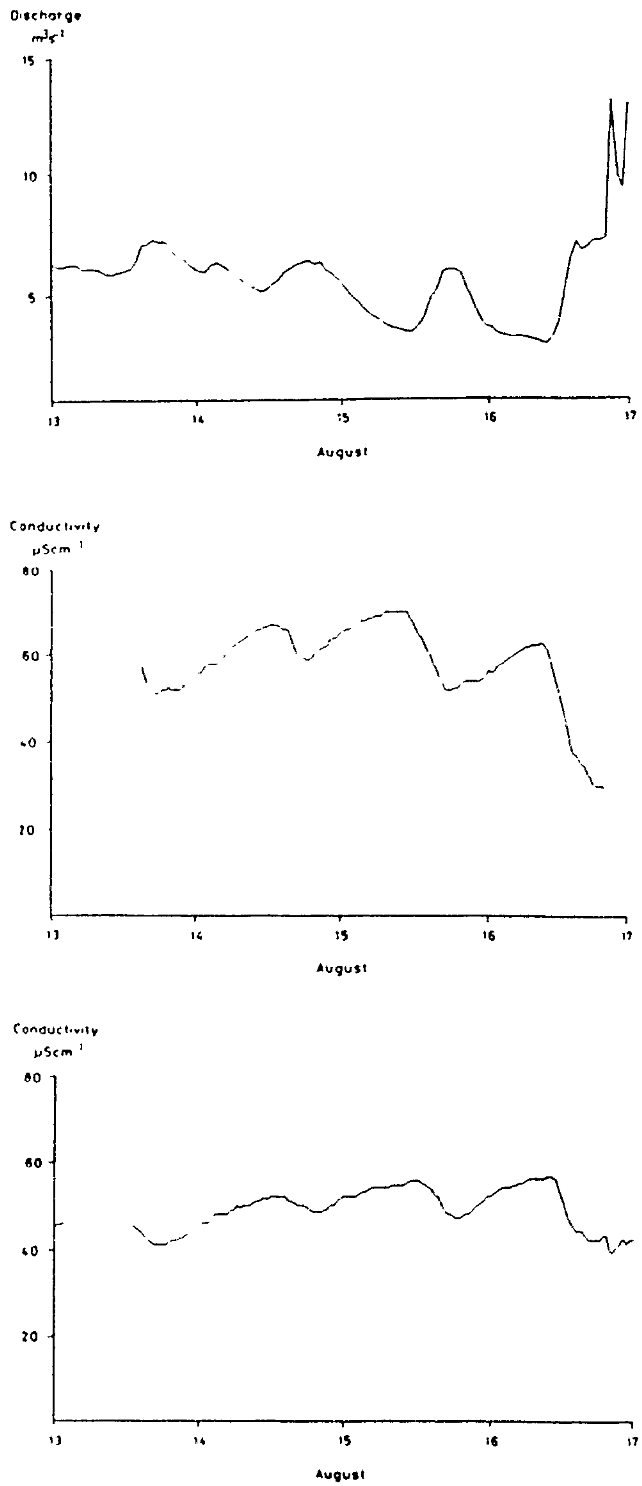
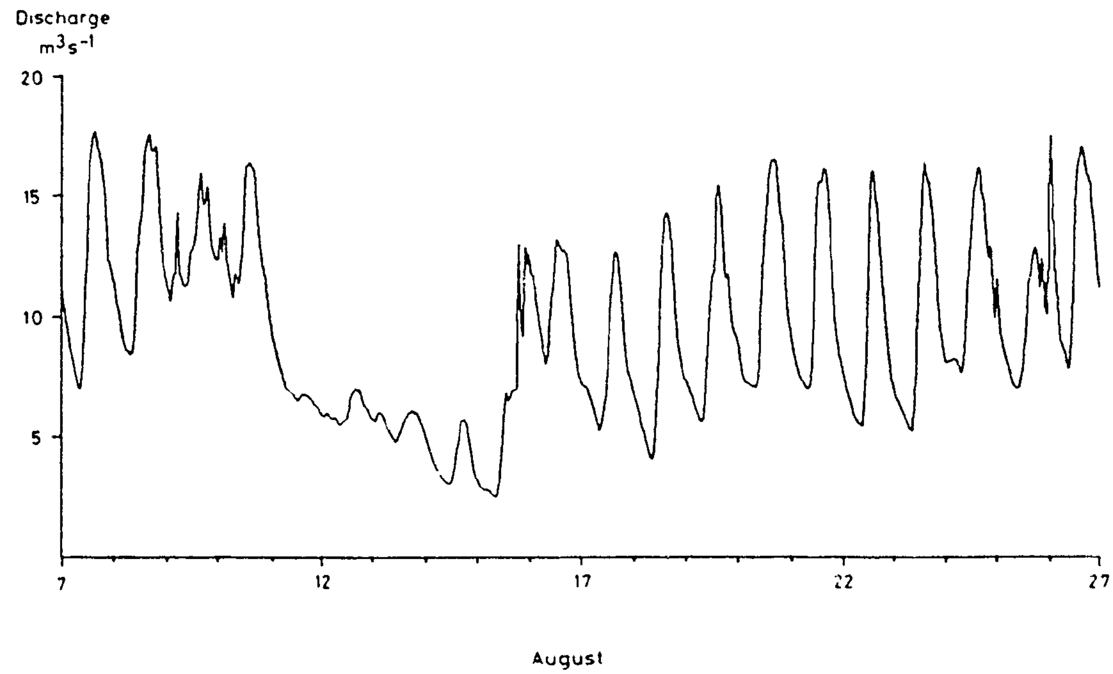
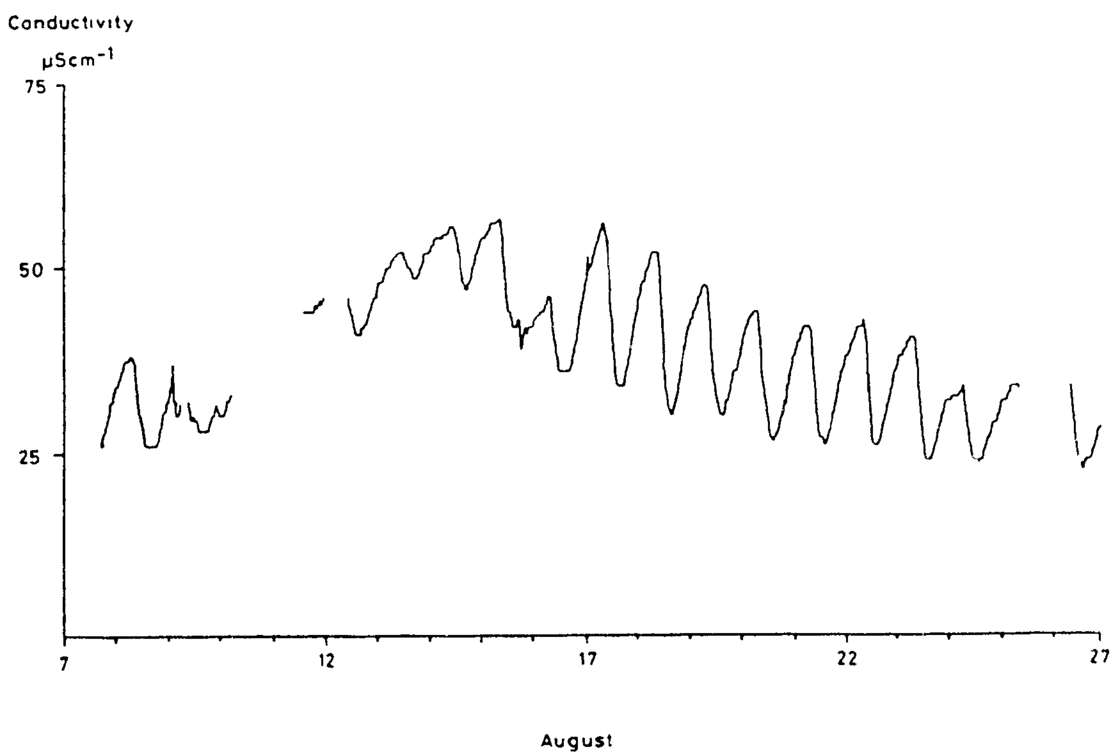


Figure 4.4 Electrical Conductivity and Discharge of Peyto Creek, August 1982.





temperature was above 11°C, maximum temperature between 14 and 15°C, and most importantly, minimum temperatures were 10°C. This maintenance of high temperatures overnight without a comparable increase during the day seems to account for the high mean discharge and may explain the series of discharge peaks over these two days. Examination of the conductivity record suggests that the increase in flow cannot be explained by the release of subglacially stored waters which have become chemically enriched at the glacier-bed since there is no appreciable rise in the chemograph at this time. It is possible that the incidence of warm temperatures, perhaps associated with the Chinook phenomena, may have maintained ablation through the night. Alternatively there may have been a sudden release of a large body of water, perhaps from the small lake situated to the East of the ice-fall. Further consideration of hourly meteorological records over this period would be desirable to determine the most likely cause.

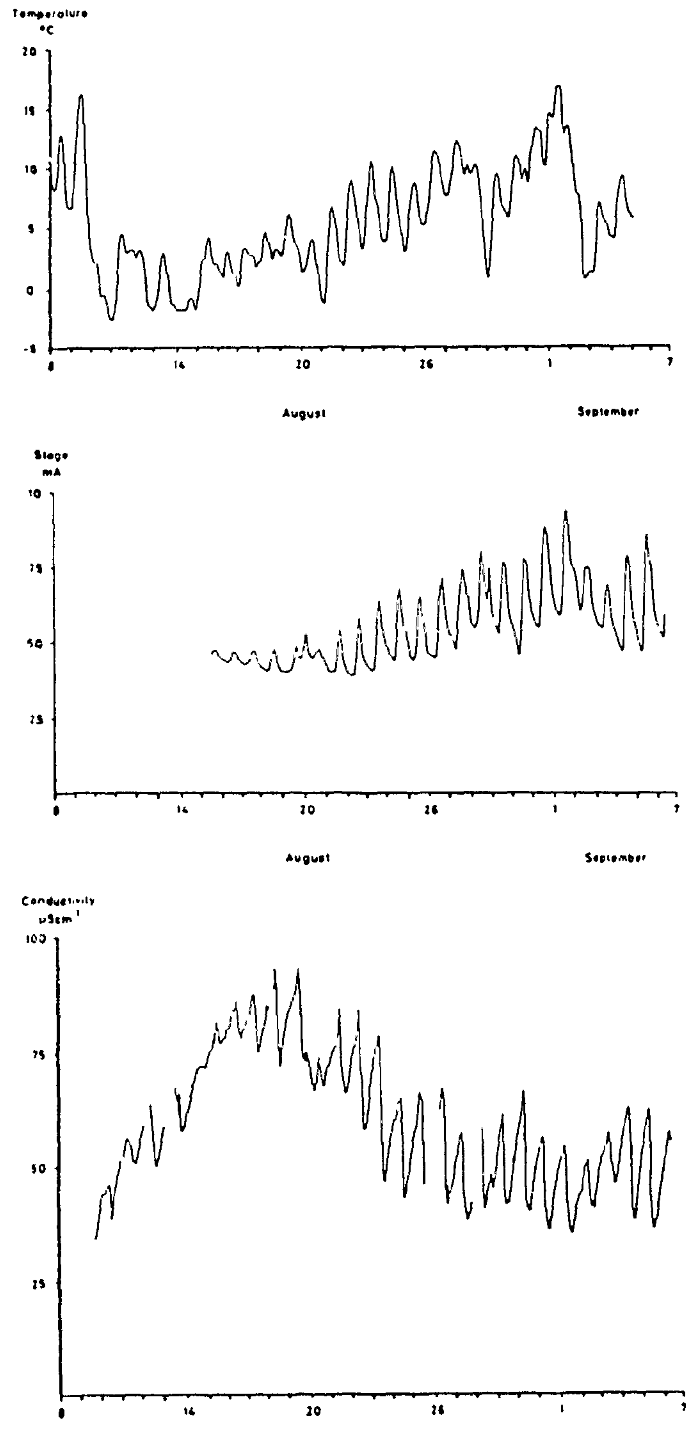
The pattern of variation of electrical conductivity in Peyto Creek has a similar diurnal form to the quantity of water exiting the basin. The dominant trend is that occurring from 16 August with a daily cycle where peak solute concentrations coincide with times of minimum flow, waters being most dilute when discharge is high. Steadily increasing peak daily discharges from 16 August produce a corresponding fall in daily peak EC. Although rising water flows provide more potential for solute evacuation as the area of contact with basal surfaces increases the greater volume of dilute ice-meltwater reduces average solute concentrations. An unusually high discharge in the morning of 24 August, which reflects a high minimum temperature of 8°C overnight, suppressed the early morning peak in electrical conductivity. The sharp peak in EC in the morning of the 9 August reflects an interruption of the daily minimum discharge, EC begins to increase then drops sharply as flow rises through precipitation. A subsequent increase in EC at the time of peak discharge can be attributed to the acquisition of solutes by surface runoff from precipitation flowing over bare rock surfaces adjacent to Peyto Creek.

4.2.2. 1987 hydrochemical record from Peyto Creek

In Figure 4.5 results from the 1987 field season are presented which include a comparable period to the 1982 data discussed above. The data cover the period 8 August through to 7 September. Whilst electrical conductivity was monitored continuously, apart from interruptions due to electrical faults on 13 and 27 August, problems with the Tinylog data recorder prevented measurement of stage until 15 August. Stage is plotted in mA reflecting the current produced by a pressure transducer, itself proportional to the overlying head of water. The hydrological data is supplemented by continuous air temperature data as recorded by a thermohydrograph at the Peyto base camp.

The period of study can be suitably divided into three distinct sections, each characterised by particular hydrometeorological conditions. Initially temperatures are high, attaining a local maximum of 16°C on 9 August. Temperatures then fall markedly, not rising above 0°C on 14 August and only starting to increase after 20 August. A peak in temperature was recorded on 1 September, at 17°C, but sharp overnight decreases in temperature occurred on 28 August and 2 September.

Figure 4.5 Electrical Conductivity and Stage, Peyto Creek with temperature (top) August-September 1987.



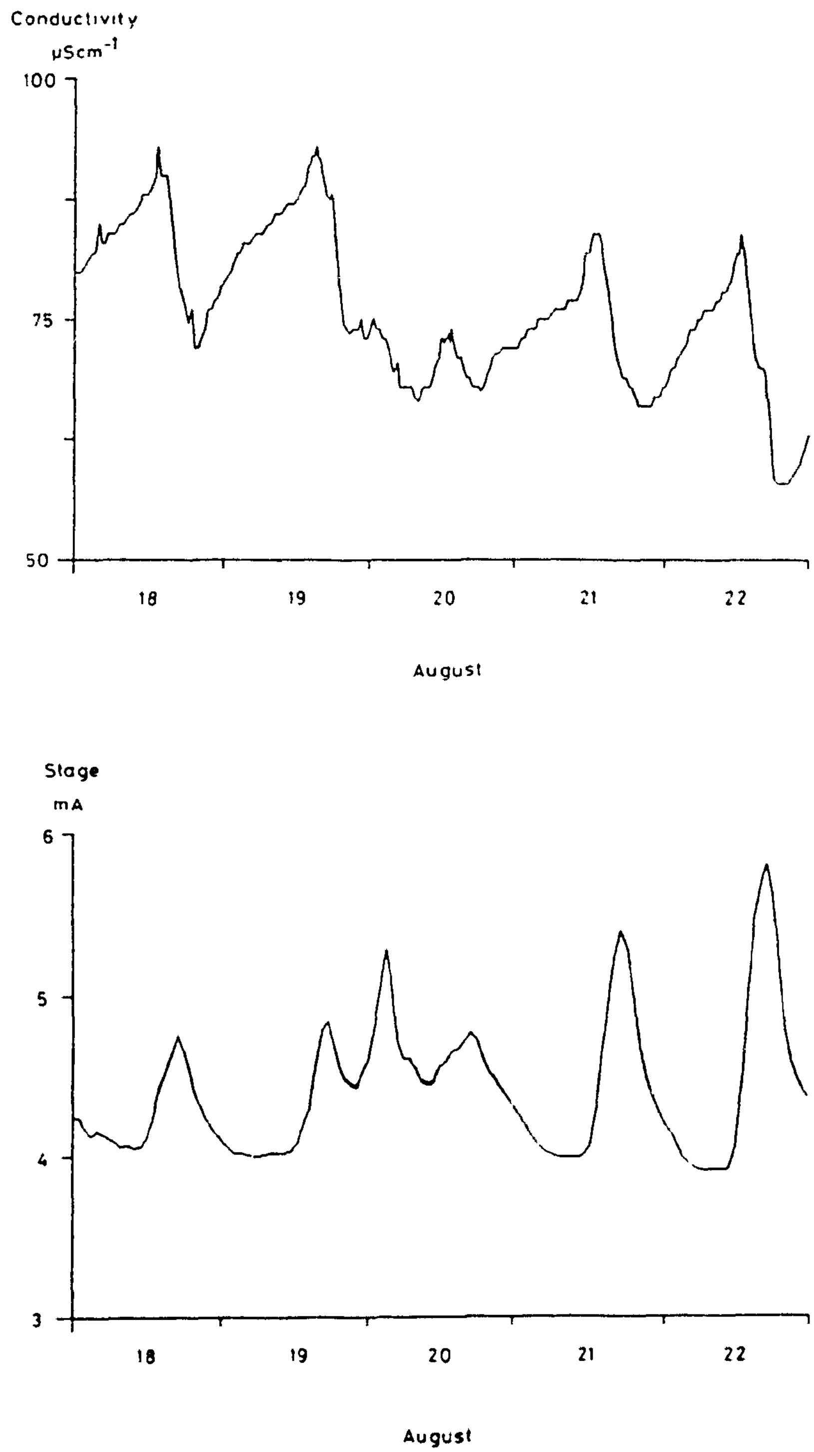
In response to the initially warm conditions electrical conductivities begin the period below 50 μScm^{-1} and then show a pattern of rising solute concentrations which are superimposed on the daily cycle of the chemograph. Cooler weather reduces the quantity of ice-meltwater flow, limiting any dilution of groundwater, as water drains from storage locations at the glacier-bed or in the firn area. Any diurnal increase in conductivity on 15 August is prevented by the absence of a temperature rise on the preceding day, thereafter the return of the daily temperature cycle brings with it the resumption of a diurnal cycle in the chemograph. Initially the range in conductivity is between 95 and 74 μScm^{-1} and discharges are low. The proportion of groundwater within total flow is thus high whilst snow-melt during the day is able to provide some dilution of the enriched proportion of flow. A continuation in the improvement in weather conditions and higher rates of meltwater production steadily increases daily peak discharges, while the resumption in ice-melt reduces solute levels. The maintenance of a daily oscillation in electrical conductivity reflects variable routing of the main water source over a daily period with consequent differences in the potential for meltwaters to acquire solute.

The chemograph shown in Figure 4.5 has several other features of interest in addition to the regular diurnal variation in electrical conductivity. Solute concentrations rise very sharply as a result of precipitation, which accounts for the short term rise in EC of 10 μScm^{-1} on the falling limb of the chemograph on 14 August, and similarly on the rising limb on 18 August. Two major storms, on the morning of 20 August and 28 August, produce rapid runoff which distorts the normal asymmetrical shape of the hydrograph. On these two days overnight rainstorms produce distinct flood peaks which limit the expected rise in electrical conductivity the following morning by continuing the typical dilution effect performed by ice-meltwater during the day.

Close examination of the 1987 electrical conductivity chart from Peyto Creek in Figure 4.5, and Figure 4.6 on which five days of stage and EC records are presented, reveals sharp increases in EC on certain days at times of maximum or minimum solute concentration. Thus maximum electrical conductivities exhibit sharp peaks on 18, 19, 20, 21, 22, 23, 24 August and 2, 3, 5 September, while the usual smooth minimum points on the chemograph are distinguished by isolated short-term increases in conductivity on 18, 26 August and 2 September.

Such increases represent a markedly consistent disruption to an otherwise regular time series and seem to indicate some instability in the basal hydrological system at a time of minimum and maximum water pressure. Rising discharges from 18 August onwards, after a prolonged recession flow, produce quantities of water which are in excess of the capability of the tunnel system to transport rapidly through the glacier. Pressure of the overlying ice produces gradual closure of the tunnel system, frictional melting from water flow compensates this and maintains free passage for meltwater flow. During recession flow drainage routes through the glacier will adapt to reflect the characteristics of meltwater supply. Sudden change in quantities of ice and snow-melt will produce a range of discharges which are different to those that the glacial drainage net has become adapted to transport. Consequently there may be instabilities in the basal drainage net which might be identifiable through examination of the precise form of time-series of different indices of water quality and quantity. Such change in glacial drainage networks

Figure 4.6 Electrical Conductivity and Stage, Peyto Creek, 18-22 August 1987.



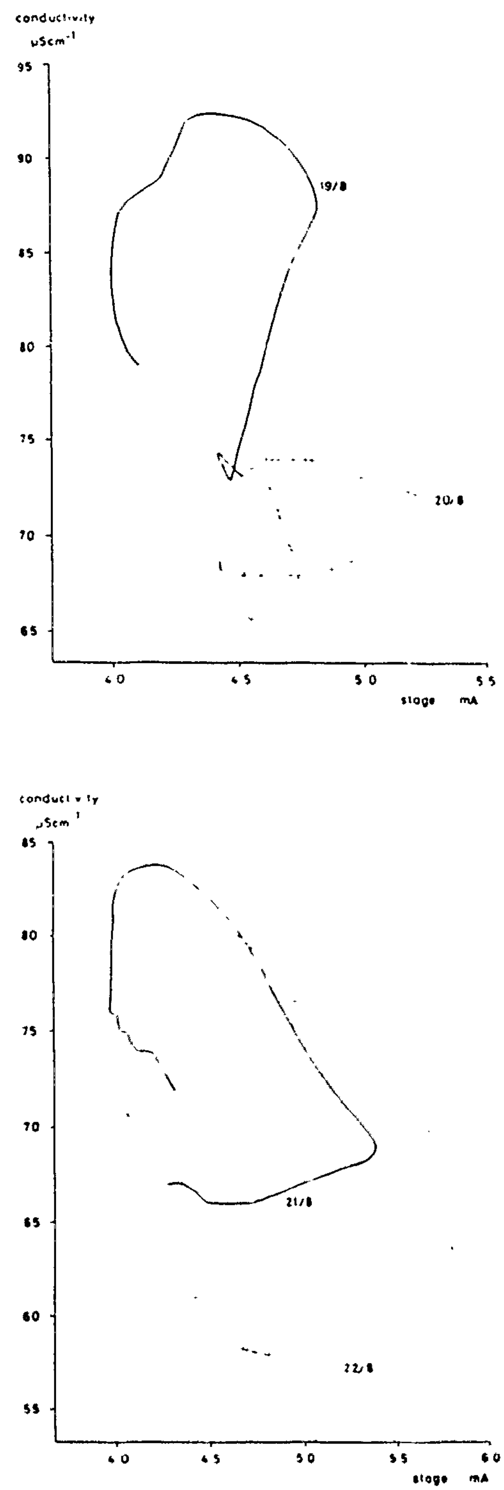
is most likely to be apparent in the period immediately following a recession event or during similar sudden variations in discharge over a short time interval.

Extension of the drainage net at high discharges when water pressures are also high will integrate chemically enriched pockets of subglacial meltwater with the total flow. This will produce a short-term increase in recorded solute concentration when conductivity is normally low.

Similarly as flows and water pressures increase, to the point of exceeding the capacity of a tunnel drainage system, some water may follow a different route through the glacier. Pockets of water may drain into cavities or a subglacial till layer at the glacier-bed. As discharges decrease on the falling limb of the hydrograph these parcels of water will be able to rejoin the conduit and flow from the glacier basin through the main arterial channel. During temporary storage at the glacier-bed water has access to abundant supplies of debris, the products of glacier abrasion, which may be dissolved by flowing meltwaters, subject to availability of sufficient hydrogen ions. The water thus acquires 'closed system' characteristics as further dissolution of minerals is prevented by a lack of free hydrogen ions for exchange. Drainage of this water produces further increases in solute concentration during the daily flow minimum which can only be seen at a short time-scale.

In Figure 4.7 electrical conductivity is plotted against stage for four consecutive days in August 1987. The general pattern is one of clockwise hysteresis whereby the same stage is associated with higher solute concentrations on the rising limb of the

Figure 4.7 Conductivity Hysteresis Loops, Peyto Creek, 18,19,20,21 August 1987.



hydrograph. Water that forms part of the initial daily increases in flow is able to access considerably more sources of fresh solute than water at the same stage on the falling limb of the hydrograph.

The form of the curve is quite regular, however precipitation on 20 August disrupts the relationship for that day. Increased discharges overnight restrict the typical increase in EC, however as flows are higher than on preceding days, greater quantities of solute can be transported, thus limiting the daily solute minima. Evacuation of solute, and the maintenance of high flows in the evening of 20 August, contribute to production of the final small sub-loop of the electrical conductivity curve on that day. The return of sustained ablation from 21 August produced a symmetrical solute - stage relationship for the following two days with clockwise hysteresis.

While the plotted hysteresis curves may be representative of hydrochemical dynamics at times of high ablation, when there is a well-defined contrast between source and routing of meltwaters, at high and low flow the relationships may change as hydrometeorological conditions vary. This is illustrated by additional hysteresis plots for five dates in 1982 which are shown in Figures 4.8 and 4.9. In both graphs there is an abrupt transition between high and low flow which produces a distinctive type of hysteresis. In Figure 4.8 the hydrograph for 12, 13, 14 July demonstrates how on the 12th the typical sharp flow increase is absent, only rising to $6.5m^3s^{-1}$, following a recession event, melting of ice resumes on the 13th producing an increase in discharge from $3.5m^3s^{-1}$ in the early morning to $10m^3s^{-1}$ in the afternoon, while on the 14th

Figure 4.8 Conductivity Hysteresis, Peyto Creek, 12,13,14 July 1982.

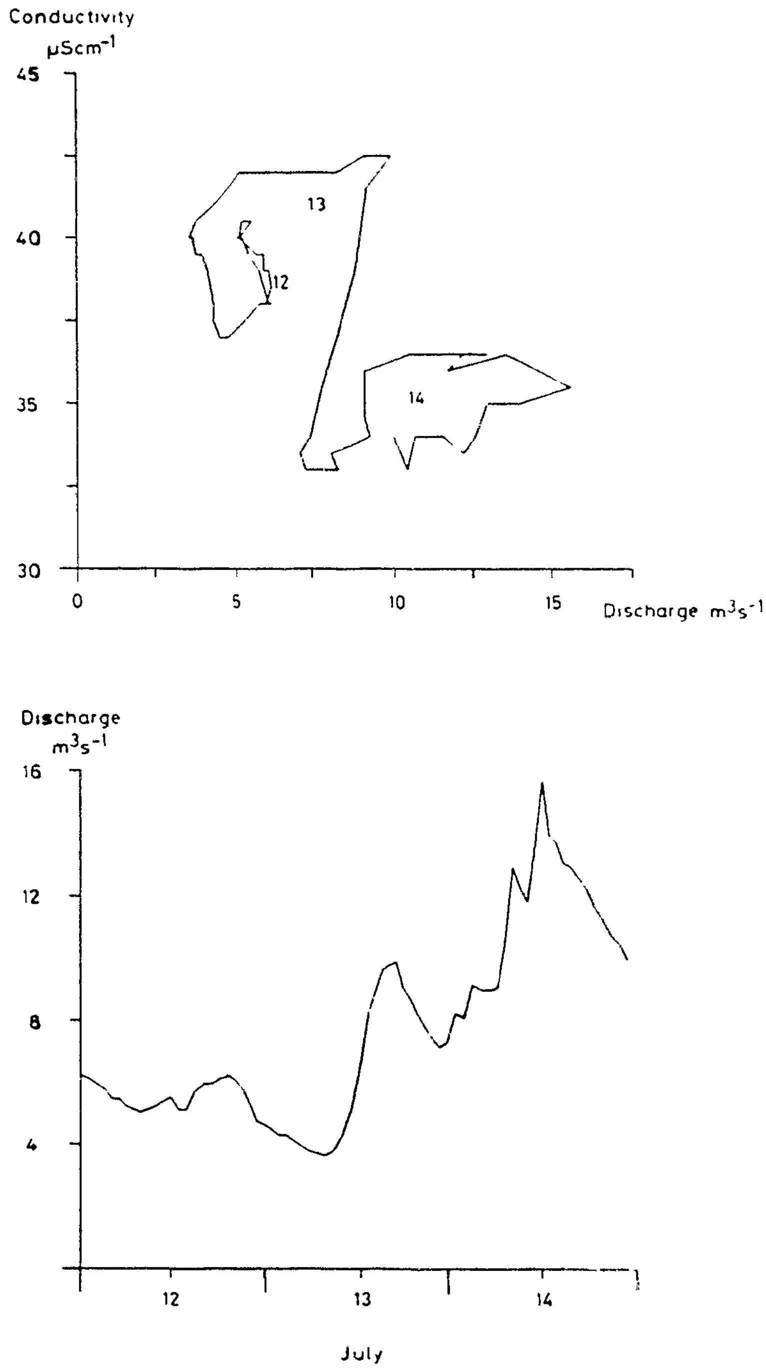
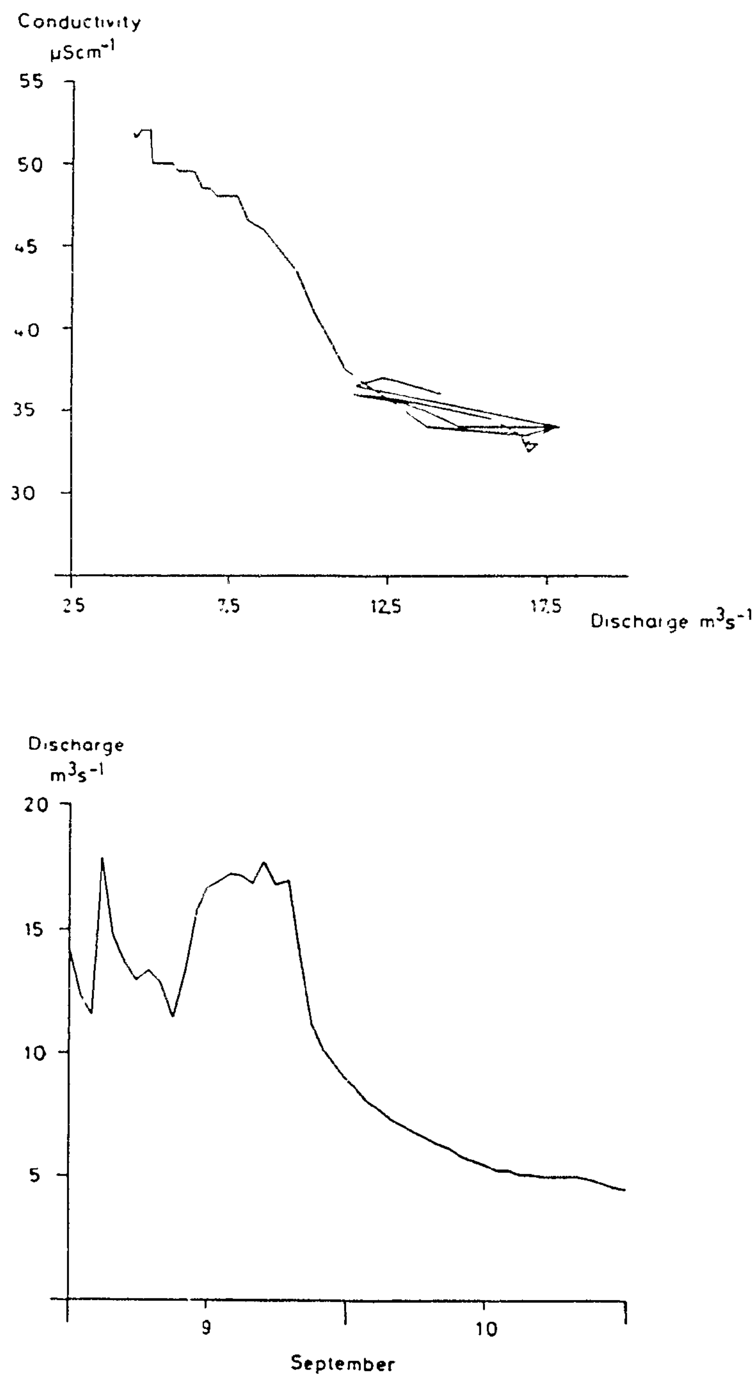


Figure 4.9 Conductivity Hysteresis, Peyto Creek, 9,10 September 1982.



discharge peaks at just under $16m^3s^{-1}$. These changes produce an interesting series of hysteresis curves. On 12 July the dilution effect caused by increasing quantities of ice-meltwater in the afternoon is absent and there is little difference between solute levels on rising and falling limbs of the hydrograph. The dilution effect reestablishes itself on 13 July, as conductivities fall from a maximum of 42.5 μScm^{-1} to a minimum of 33 μScm^{-1}. On this particular day discharge and electrical conductivity reach a peak at the same time demonstrating that flows have been low for a considerable period to the extent that rising water flows in channels beneath the glacier are able to access further sources of solute. On 13 July greatly increased discharge accentuate meltwater dilution such that solute concentration drops 1.5 μScm^{-1} before discharge peaks at $15.5m^3s^{-1}$.

A different hysteresis curve is shown in Figure 4.9 for 10 and 11 September 1982. As the hydrograph indicates, the pattern of variation of discharge is quite distinct from the diurnal fluctuations observed at the height of the ablation season. A storm in the early morning of the 10th interrupts the daily flow decrease with a sharp rainfall-induced discharge peak of $17.5m^3s^{-1}$. Discharge then falls before increasing to a peak of $17m^3s^{-1}$ at 16:00hrs on the same day. Flows then decrease exponentially as snowfall prevents further ablation and the hydrograph reflects a typical recession event. The hysteresis curve corresponding to this hydrograph reflects the interaction of these different processes. Discharges are initially high, and solute concentrations low, but there are two separate loops in the plot, one corresponding to the early morning storm, the other to the usual afternoon increase in discharge. Given the small time interval between the two events electrical conductivities are essentially comparable. On the following day as discharges decrease EC increases steadily as meltwater drains from different locations within the drainage basin. Diurnal variations in both conductivity and discharge are absent at this time, consequently there are no hysteresis loops on this day.

The relationship between conductivity and discharge is therefore one that deserves further investigation. While hysteresis effects limit the potential of applied regression models, study of continuous time series of electrical conductivity seems to provide some indication of how the dominant pathway taken by meltwater through a glacierised basin can vary on several time-scales, and most noticeably on the diurnal scale.

4.2.3. Peyto Waterfall

Comparison of a chemograph from the measuring site at Peyto Waterfall with the electrical conductivity record from Peyto Creek illustrates how meltwater hydrochemistry varies depending on meltwater provenance. The location of the waterfall site is described in Chapter 3, and is illustrated in Figure 3.1 (site E). Differences in the characteristics of the catchments drained by the waterfall stream and Peyto Creek can be identified by reference to Table 4.1.

The area drained by the waterfall stream cannot be delimited accurately. Consideration of the direction of surface streams as indicated by the map of Peyto Glacier produced by the Glaciology Division (1975) suggests that the waterfall stream drains the basin between Peyto and Trapper Peaks. Although the proportion of the basin which is glacier-covered is not significantly different from the entire basin the sub-catchment

Table 4.1. Summary characteristics of the Peyto basin and the Waterfall sub-catchment.

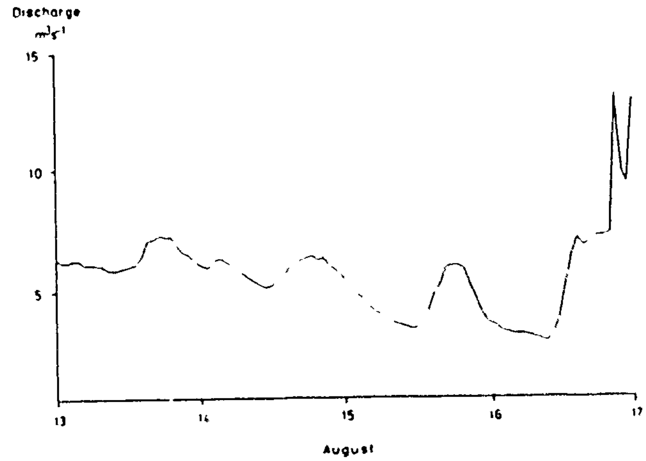
Characteristic		Peyto Creek	Waterfall Stream
Catchment Area	km^2	21.90	3.08
Glacierised Area	km^2	12.54	1.78
Per cent glacierisation		57.0	57.8
Catchment elevation range		1960-3172	2349-3016
Glacier elevation range		2135-3172	2355-3116

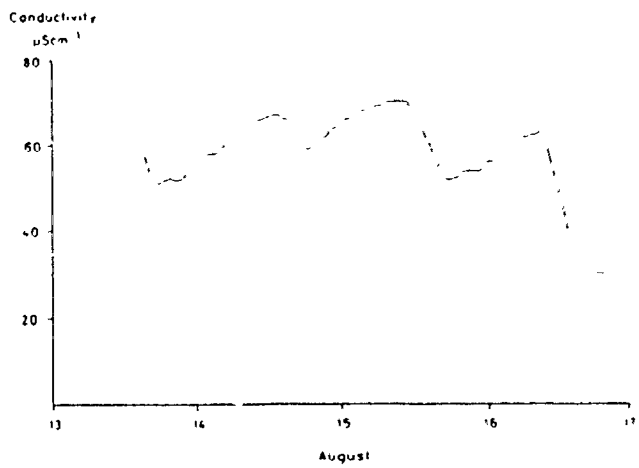
would be expected to possess a greater percentage of firn. The area above the waterfall also contains several perennial snow-patches which cannot be determined satisfactorily and which would increase the proportion of the basin which is covered by snow and ice. The quantity of total flow which is derived from groundwater sources is also likely to be less in the waterfall catchment when compared with the remainder of the Peyto basin. The ice-free portion of the basin above the waterfall consists of the non-glacierised South facing slopes of Peyto Peak which are composed of small lateral moraines, above which are some solifluction deposits and screes supplied by frost-shattering.

Figures 4.10, 4.11, and 4.12 present all electrical conductivity results that were obtained from the waterfall site in 1982, 1984 and 1987. Although some records are incomplete, considerable variation occurred in mean EC observed in the three years, which is summarised in Table 4.2.

As the table indicates the lowest mean electrical conductivity, 17.8 μScm^{-1}, occurred in the waterfall stream in late summer 1987, while the highest mean EC, 57 μScm^{-1}, was also at the waterfall in 1982. Absolute differences in solute concentrations will reflect variations in the contribution of dilute ice-meltwater. Water draining from the upper basin is likely to have a varying relationship with the bulk meltwaters as measured at Peyto Creek. In some situations, when the snowline is at a particularly high elevation, significant quantities of ablation may occur in the upper western basin. However, after a snowfall, ablation of ice may be restricted to a small area of the lower glacier tongue, in which case solute loads in the waterfall stream will be higher than waters within Peyto

Figure 4.10 Waterfall stream conductivity August 1982 (middle), with Peyto Creek discharge (top) and conductivity.





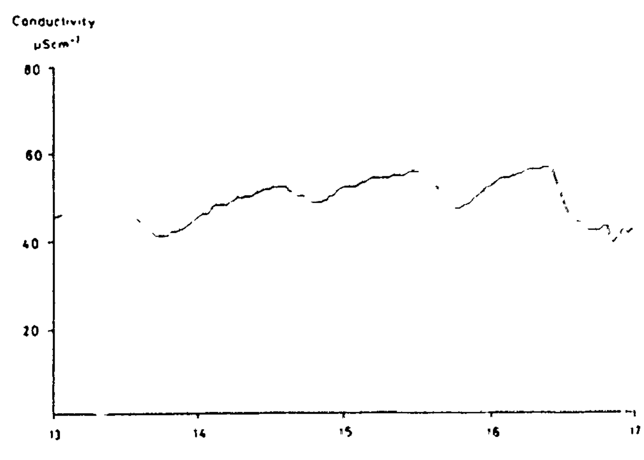


Figure 4.11 Waterfall stream conductivity August-September 1984.

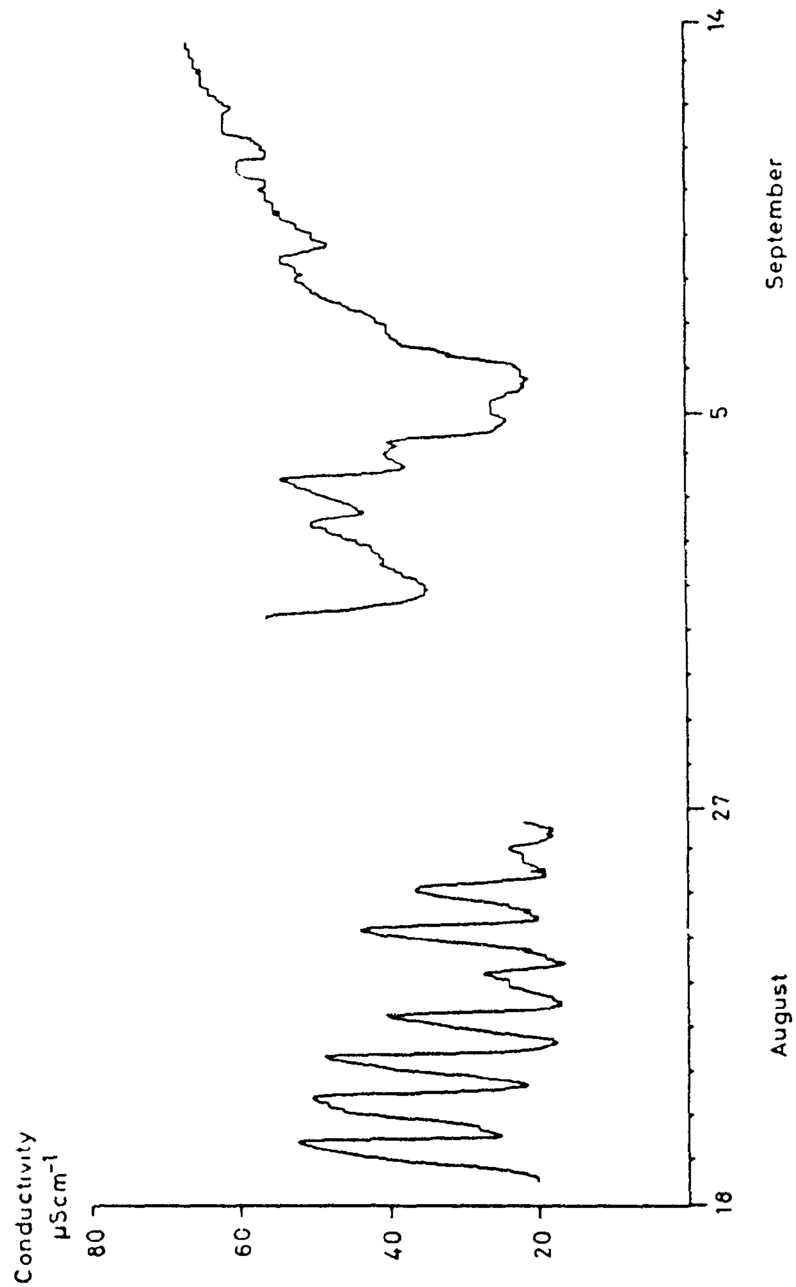


Figure 4.12 Waterfall stream conductivity August 1987 with Peyto Creek stage and conductivity (top and bottom).

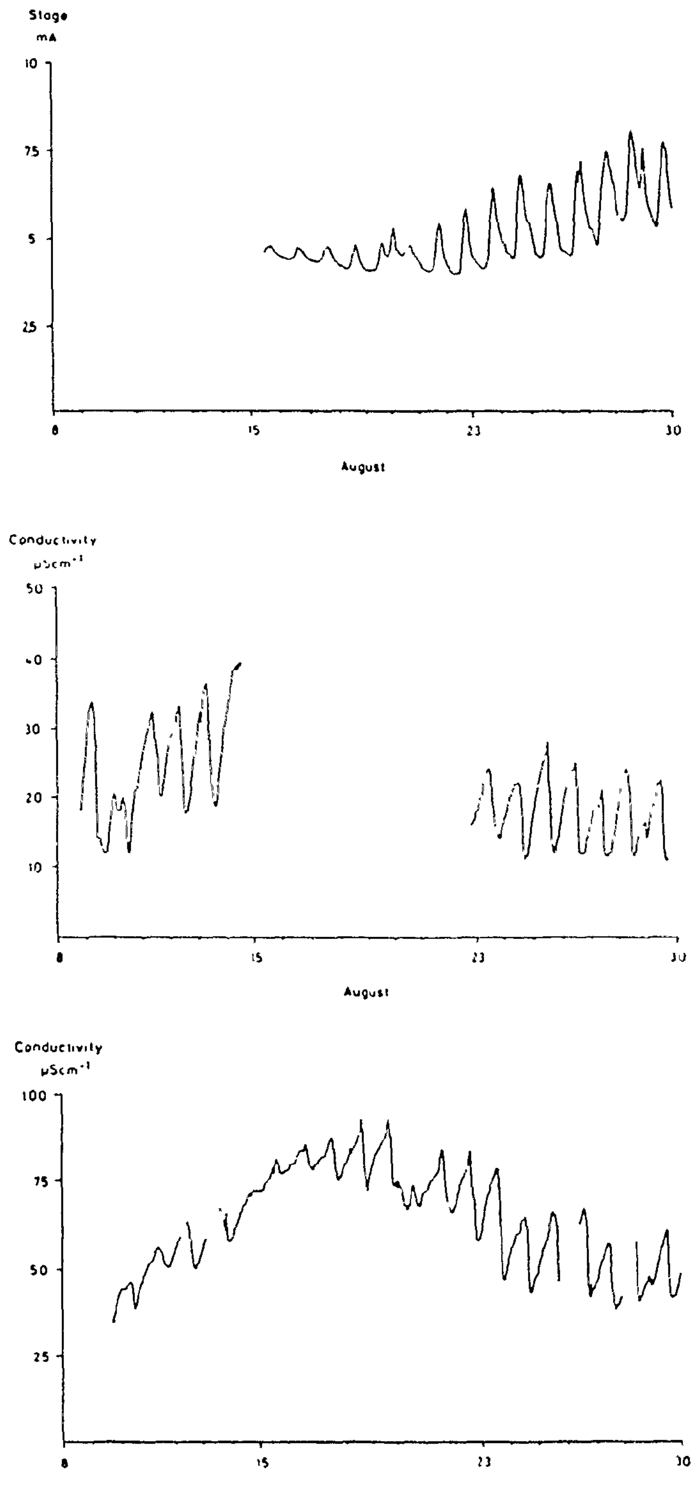


Table 4.2 Summary characteristics of solute concentrations in Peyto Creek and Waterfall sites, 1982, 1984, 1987.

Year	Site	Range in EC μScm^{-1}	Mean EC μScm^{-1}	St. deviation μScm^{-1}
1982	Creek	40-55	49.7	4.6
1982	Waterfall	30-70	57.8	9.6
1984				
August	Waterfall	19-52	29.1	10.0
September	Waterfall	20-65	46.8	12.5
1987				
early Aug	Creek	30-73	53.0	8.4
early Aug	Waterfall	12-42	25.7	6.9
1987				
late Aug	Creek	45-84	54.8	9.9
late Aug	Waterfall	14-30	17.8	4.1

Creek. Water may drain from storage locations in the firn area, and may have access to some sources of solute on first encountering reactive particle surfaces.

Detailed examination of the electrical conductivity time-series reveals slight differences in timing of melting in the western basin drained by the waterfall, when compared with the entire basin. Consideration of the exact timing of maximum and minimum solute concentrations in the waterfall stream and in Peyto Creek reveals that melting in the upper basin tends to precede melting in the remainder of the basin by 45 minutes. This may arise through the aspect of the Peyto basin, the area between Mt. Baker and Peyto Peak is the first part of the basin to receive daily radiation, at which time the ablation area below the icefall is still shaded from the rising sun by Mt. Thompson. Slight differences in routing of meltwater in the two streams and superimposition of flows in Peyto Creek may account for part of the variation.

Without discharge measurements at the waterfall site, rates of solute evacuation cannot be determined. Although the measuring instruments were installed at the same location in the three years it is possible that changes in the hydrological network beneath the glacier altered the contributing area of the glacier which supplied meltwater to the waterfall stream. In the period 1987-1989 the glacier margin in the vicinity of the waterfall did not appear to have changed, however in 1989 the stream no longer followed its former course. It seems that waters draining from the western basin assumed a different route entirely beneath the glacier throughout the summer melt-season.

Diurnal variations in solute concentrations in the waterfall stream demonstrate that meltwater from a variety of sources contribute to streamflow. Maximum elevation of the snowline varies between years, in 1987 and 1989 it reached an elevation of 2600m.a.s.l., 250m above the altitude of the waterfall site (2349m). During recession flows following summer snowfalls the snowline will descend below the waterfall so that quantities of water derived from ice-melt are likely to be less at the waterfall site when compared with Peyto Creek where some ice-meltwater may be received from the lower ablation area.

The 1982 data series in Figure 4.10 only covers the period 13-16 August however the chemograph provides evidence of changes in water quality in response to significant variations in flow totals. Initial flows are very low with a daily peak discharge below $7m^3s^{-1}$ as the glacier remained snow-covered. An improvement in weather from 15 August onwards produces a rapid rise in the snowline and with rainfall in the afternoon of 16 August a sharp increase in flow to $13m^3s^{-1}$ is apparent. Electrical conductivities in both the waterfall stream and Peyto Creek display only limited diurnal variation before the rainfall peak on 16 August. A greater diurnal range in the waterfall stream EC probably reflects a simplified drainage net above the waterfall site. The large basin area above Peyto Creek, consisting of three distinct accumulation basins, would have been sufficient to support a better distributed drainage system which under low flow conditions would account for much of the total water transport. A rapid fall in electrical conductivity on 16 August occurred as water quality responds to a resumption of ice-melt in the area immediately above the waterfall. A more gentle decrease in EC in Peyto Creek occurs through integration of dilute meltwater flows from areas of exposed ice over the whole basin which follow different routes to reach the gauge site.

Changes in routing of channels above the waterfall will expose different sources of solutes which may then be dissolved by flowing meltwaters. Higher solute concentrations in the waterfall stream than in Peyto Creek in 1982 indicate a significant source of solute in the upper western basin.

Electrical conductivities in the waterfall stream in August and September 1984 are presented in Figure 4.11. Solute concentrations display a range of 30 μScm^{-1} at the beginning of the record in mid-August around a mean of 29.1 μScm^{-1}. A limited overnight increase in maximum of EC, 40 μScm^{-1}, on 21/22 August reflects warmer conditions overnight which restricted the typical daily conductivity increase which normally occurs when discharges decrease. The same occurred on 24/25 August when EC reached a maximum of 27 μScm^{-1}. A distinct short-term increase in conductivity on 25 August reflects either precipitation or expansion of the drainage net which provides an opportunity for isolated pockets of water to drain from storage sites within the accumulation area.

Solute levels in the waterfall stream in early September are similar to conductivities recorded in Peyto Creek in the autumn of 1982 and 1984. Daily fluctuations in solutes are evident at first, a descending snowline rapidly halts ablation of ice increasing measured electrical conductivities as waters drain from subglacial and englacial storage locations within the basin.

In 1987 electrical conductivities in the waterfall stream and in Peyto Creek, which are shown in Figure 4.12, exhibit distinct similarities. Weather preceding the record was ideal for sustained ice-ablation which produced a large diurnal range in EC on 9 August at the waterfall. The daily rise in EC on 10 August was subdued at both Peyto Creek and

2 OF/DE 2

1.0 2.8 2.5 3.2 2.2 3.6 2.0 4.0 1.1 1.8 1.25 1.4 1.6

the waterfall. At the former site a minor increase in EC occurred in the early morning before higher discharges contributed to the regular fall in solute concentration. In the waterfall stream a different pattern through time was observed and two isolated peaks in EC are noticeable. This could be due to increasing water flows at a time when discharges are normally a minimum. A difference in timing of the flood-peak reaching the two measuring sites would account for the slightly different form of the two chemographs.

On the following day maximum solute levels show a much greater increase as the weather deteriorates and discharges fall. Electrical conductivities in Peyto Creek appear to be affected to a greater extent by these factors than solute concentrations in the waterfall stream. At the latter site EC reached a maximum of 40 μScm^{-1} on 14 August when snow-cover over the glacier surface severely limited ice-ablation although EC on the previous three days presented a regular diurnal form. In contrast at Peyto Creek flow-routing of meltwaters produces a rising daily time-series of conductivity to over 70 μScm^{-1} on 14 August. The greater distance that flowing meltwaters have to travel to reach the Creek gauge site permits greater acquisition of solute when compared with the waterfall site where lower values of conductivity were recorded in 1987. Differences in the relative magnitudes of Peyto Creek and Waterfall solute concentrations in 1982 and 1987 can be explained partly in terms of a difference in the proportion of the waterfall stream which was routed at the glacier surface in the supraglacial drainage system. These waters would have limited potential to acquire dissolved material. Also important will be the position of the snowline at any time, and hence the proportion of flow contributed by snow-melt. Insufficient data are available to compare the importance of this factor between different years.

The waterfall stream does not show any evidence of the precipitation event on the evening of 13 August 1987 at Peyto Creek on the falling limb of the chemograph. This produced a distinct increase in EC at Peyto Creek from 62 μScm^{-1} at 20:00 to 66 μScm^{-1} at 20:15 which then fell to 61 μScm^{-1} at 20:30. This increase would reflect surface runoff over moraines on the slopes above Peyto Creek. Precipitation above the waterfall, where the elevation is 400m higher than at the main gauging site, would be more likely to have fallen as snow, which would account for the absence of any short-term rise in solute concentration.

The electrical conductivity record from 23 August 1987 also displays marked differences between the chemographs of the Waterfall stream and Peyto Creek. A greater range in solute concentrations occurred in Peyto Creek where EC fell from 80 μScm^{-1} to 48 μScm^{-1} over the course of 23 August, at the Waterfall the corresponding range was 25 μScm^{-1} to 14 μScm^{-1}. The sequence of increasing or decreasing maximum electrical conductivities over subsequent days is also different between the two stations. At the waterfall maximum EC after 23 August was recorded on 25 August, the peak daily value then fell until 27 August. In contrast daily maximum EC in Peyto Creek presents a different temporal pattern, having a high value on 23 August daily maximum EC was almost constant between 24 and 26 August, then falling on 27 August and exhibiting a rising form for the next four days. The diurnal range of EC in the Waterfall stream is also much less in late September when compared with early to mid August, mean EC decreased from 25.7 to 17.8 μScm^{-1}, while the standard deviation fell from 6.9 to 4.1

μScm^{-1} (see Table 4.2). In contrast at Peyto Creek over the same interval mean conductivities increased from 53 to 54.8 μScm^{-1}, and the standard deviation rose from 8.4 to 9.9 μScm^{-1}.

Such differences in water chemistry between the two sites provide some indication of changes in drainage of meltwater through the glacier towards the end of the summer melt-season. Coolei temperatures in the accumulation area in late summer, as total daily radiation input decreases with a reduction in daylight hours, restrict meltwater percolation through the firn area. Meltwater therefore freezes within the firn whereas earlier in the summer it would have continued passing through the glacier overnight and would register a high conductivity on the chemograph. This process is spatially confined to the accumulation area because of differences in the hydrological characteristics of firn and glacier ice. Quantities of water percolating at grain boundaries in the ablation area are insignificant in comparison with meltwater percolation through snow and firn in the accumulation area.

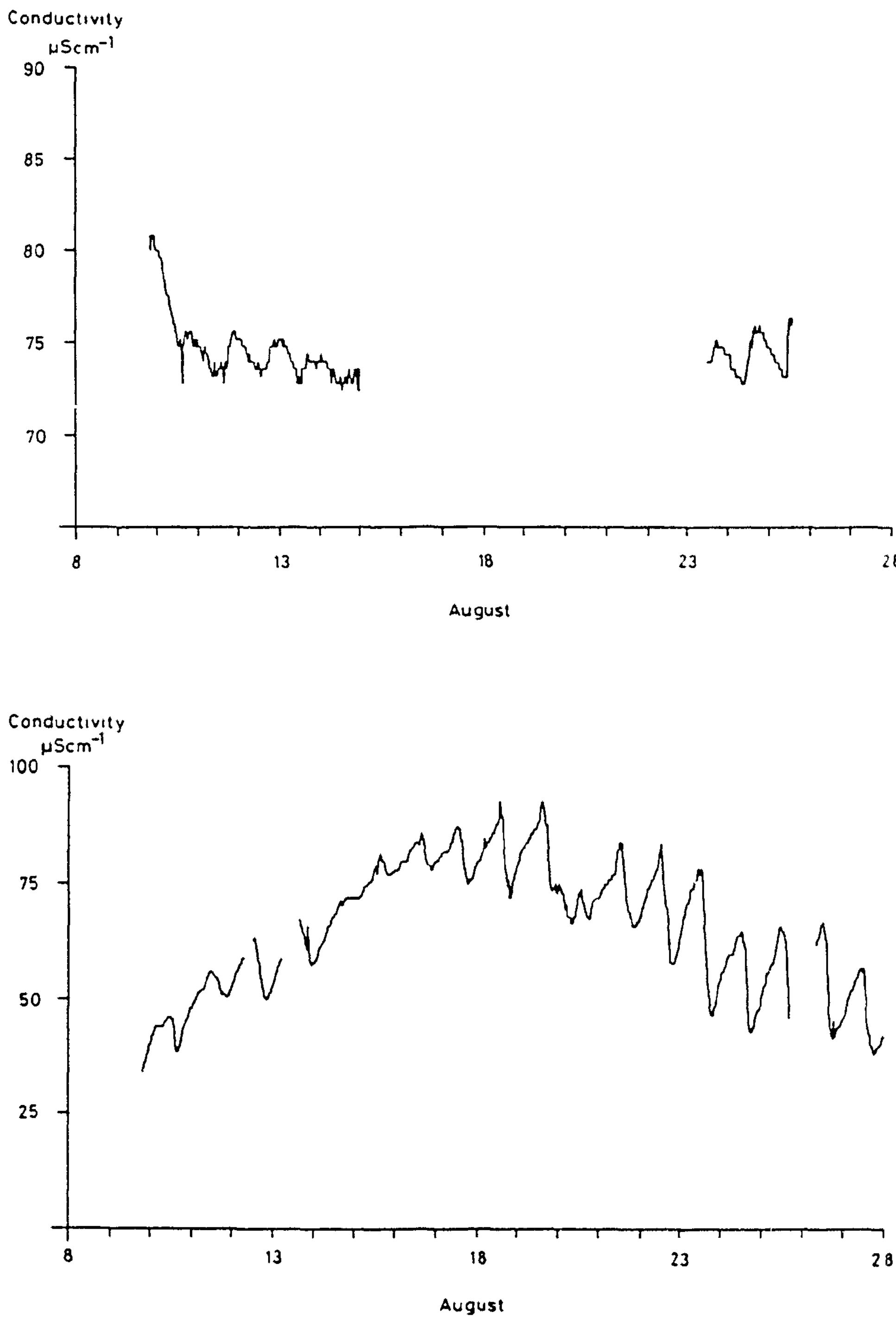
These results are in accordance with results described at South Cascade Glacier by Fountain (1989). He noted an increasing impermeability of glacier firn in September and October which was explained in terms of firn compaction. Water tables in the accumulation area of South Cascade Glacier reached a maximum level in late summer after the principal period of ice-melt had passed.

4.2.4 Groundwater contribution to streamflow

Electrical conductivity of a groundwater spring was successfully recorded for several days in 1987 and is presented in Figure 4.13 together with conductivity in Peyto Creek. An electrical fault prevented continuous measurements throughout the study period, however six days of data from 10 to 15 August yielded values of conductivity of between 73 and 81 μScm^{-1}, which presented a slight diurnal fluctuation of approximately 2 μScm^{-1}. Timing of the daily conductivity variations in the spring was out of phase with EC fluctuations recorded in Peyto Creek. Solute levels in the spring would appear to be related to the absolute quantity of water routed through the groundwater network which does not remain constant through time. Slow passage of water through a karst system would present abundant opportunities for acquisition of solute. The geometry of drainage routes through the ground will be such as to permit increased solute uptake as an increased wetted perimeter will produce a greater area of contact with rock surfaces as discharges increase.

The initial sharp fall in EC from 81 μScm^{-1} to below 75 μScm^{-1} on August 9 and 10 arose as meteorological conditions changed, temperatures fell considerably from a peak of 17°C on August 8 to one of 3°C on August 9. This halted processes of snow and ice-melt, although some rainfall would have provided a dilute source of water to the groundwater system. Flow through the groundwater system would thus appear not to be constant but to reflect variations in meltwater supply, probably derived from the melting of several perennial snow patches and glacierets on slopes to the east of Peyto Creek.

Figure 4.13 Groundwater conductivity (top) with Peyto Creek conductivity, August 1987.



Solute loads in the groundwater seem to display distinct trends as hydrometeorological conditions vary. Electrical conductivities of groundwater in 1987 are lower than the 98-100 μScm^{-1} recorded by Collins and Young (1979b) at the same site in 1978, suggesting greater supply of dilute meltwater to the groundwater system in the summer of 1987. Changes in the quantities of water passing through the groundwater network in addition to absolute changes in the amount of solute available in the course of a year are likely. This may account for different electrical conductivities in Peyto Creek in April 1982, which were twice the solute concentrations observed in September and October 1982 and 1984. Although flow in Peyto Creek was probably higher in the autumn than in the period immediately before the first snow-melt, greater solute evacuation would be likely early in the year as slowly percolating groundwater has access to rock surfaces for the first time that year.

Solute concentrations of the groundwater stream do not increase significantly during the subsequent recession event in mid-August as rainfall within the catchment at lower elevations maintains groundwater flow. Higher electrical conductivities at the measuring site at Peyto Creek thus probably indicate greater availability of solutes in the immediate vicinity of the glacier where processes of glacial abrasion and frost shattering of adjacent moraine surfaces can continually generate new surfaces suitable for dissolution. Equivalent processes are not found in the distinct hydrochemical environment which waters flowing in the groundwater spring can access.

The question remains as to what spatial variation in groundwater solute content might be recorded within a glacierised basin. It is most likely that significant quantities of groundwater, which have been derived from a karstic groundwater system beneath Peyto Glacier, will contribute unseen to discharge in Peyto Creek. These meltwaters may have higher initial solute concentrations, having slowly passed through the portion of the basin which is glacier-covered and thereby acquiring further dissolved material.

The addition of groundwater to flow in Peyto Creek may account for a progressive increase in conductivities as distance increases from the terminus of Peyto Glacier. Comparing Figures 4.4 and 4.5, conductivities in 1982 during times of high ablation were approximately 10 μScm^{-1} lower than during similar hydrological conditions in 1987. The main monitoring site in 1987 was some 150m downstream of the Water Survey of Canada site used in 1982 and although there was no visible inflow of water between the two sites differences in the relative contribution of groundwater at the two sites would be possible.

4.3 Hydrogen-ion concentration of snow and ice meltwater

Study of time series of electrical conductivity helps to determine how flowing meltwaters may take different routes over varying time scales in their passage through a glacierised basin. However EC is a measure of total dissolved ions and in some instances it may not provide an accurate indication of the behaviour of individual ion behaviour (Thomas, 1986). Measurement of the variation through time of an individual ion may therefore help to identify the shortcomings of a research design which is based upon conductivity. Continuous monitoring of pH, an inverse logarithmic indicator of hydrogen potential, would appear suitable here as the ability of meltwater to acquire solute is dependent upon availability of sufficient free hydrogen ions for exchange at mineral surfaces. Accurate determination of hydrogen ion concentration should therefore indicate maximum possible solute concentrations in glacial meltwater, given sufficient supply of soluble material which water may contact during passage through the basin. Differences in hydrogen ion content in the bulk meltwaters of Peyto Creek would therefore be expected through time as the potential of different water types to acquire an initial amount of hydrogen ions will vary. Initial sources of hydrogen ions are mainly atmospheric, through gas transfer of carbon dioxide and dissolution of carbonic acid (H_2CO_3) and sulphuric acid (H_2SO_4). High suspended sediment concentrations in meltwater will also influence recorded values of pH. Metcalf (1984a) noted that an increase in pH of between 0.1 and 0.3 might arise in this way and suggested filtering of meltwater to minimise the problem.

In Figure 4.14 pH of Peyto Creek is plotted for a short period in August and

September 1982, together with discharge and conductivity. Discharge and pH vary in phase, and are inversely proportional to electrical conductivity. Hydrogen ion concentration thus decreases as discharge increases and as such the ion behaves in a similar way through time to total solute concentration as determined by monitoring of conductivity. Significantly the two highest hydrogen ion concentrations occur on August 20 and 24 at pH's of 8.75 when overnight discharges were the lowest during the period of overlapping coverage of pH and discharge (i.e. 12:00 on 19 August to 3 September).

Ice-melt appears to be depleted in hydrogen ions by comparison with water from other sources. Processes of firnification involve progressive rejection of solutes, so that initial hydrogen ion concentrations will be progressively diminished during the formation of ice. Percolating snow-meltwater may be able to access these ions increasing both the hydrogen ion and the total ion concentration of the flowing meltwater, especially in comparison with ice-meltwater. Variation in the contributions of water derived from snow and ice-melt, which vary distinctly in their solute contents, would account for the distinctive diurnal cycle in the pH plot.

In Figure 4.15 the pH of the waterfall stream is plotted for a short period in August 1987. While the rest of the pH record collected at this site in 1987 is circumspect, because of sedimentation in the collecting bottle used to shield the pH probe, the graph illustrates the effect of precipitation in the accumulation area in the morning of 10 August, when pH fell from 9.3 to a minimum of 5.8. Precipitation over a glacier surface which has no snow cover will quickly form surface runoff and, in the absence of

Figure 4.14 Peyto Creek pH, August/September 1982, with discharge (top) and conductivity (bottom).

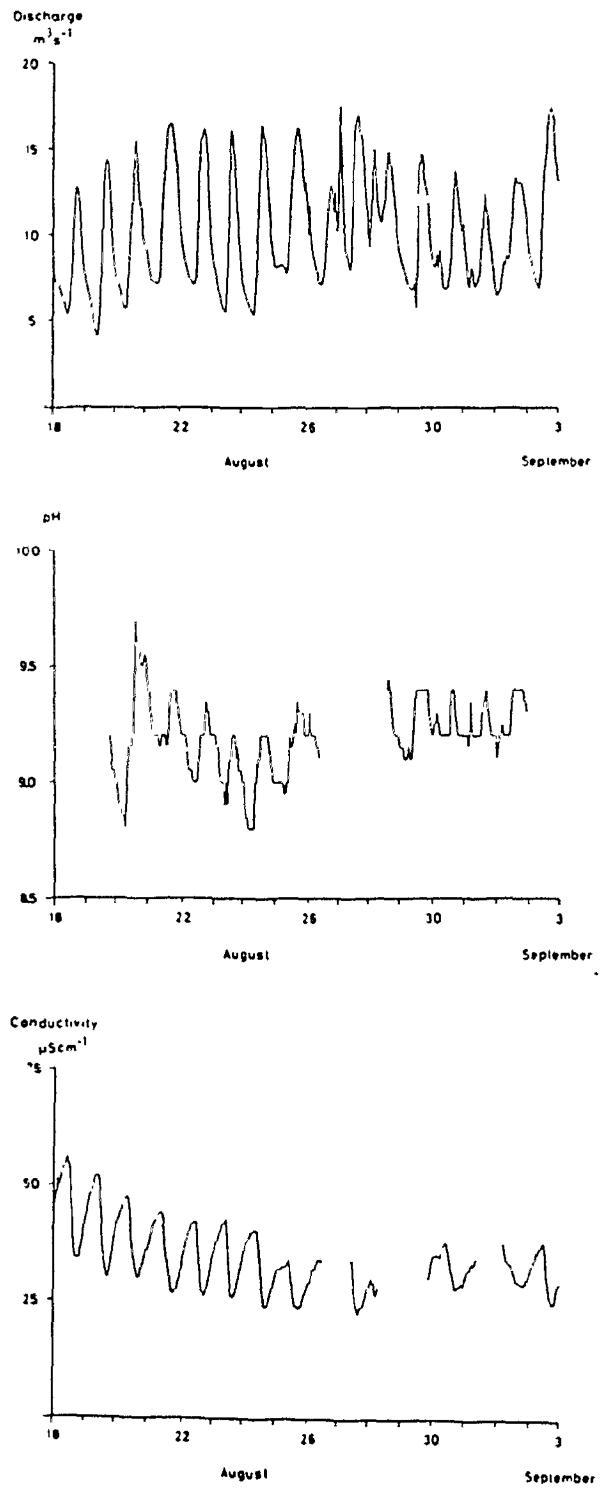
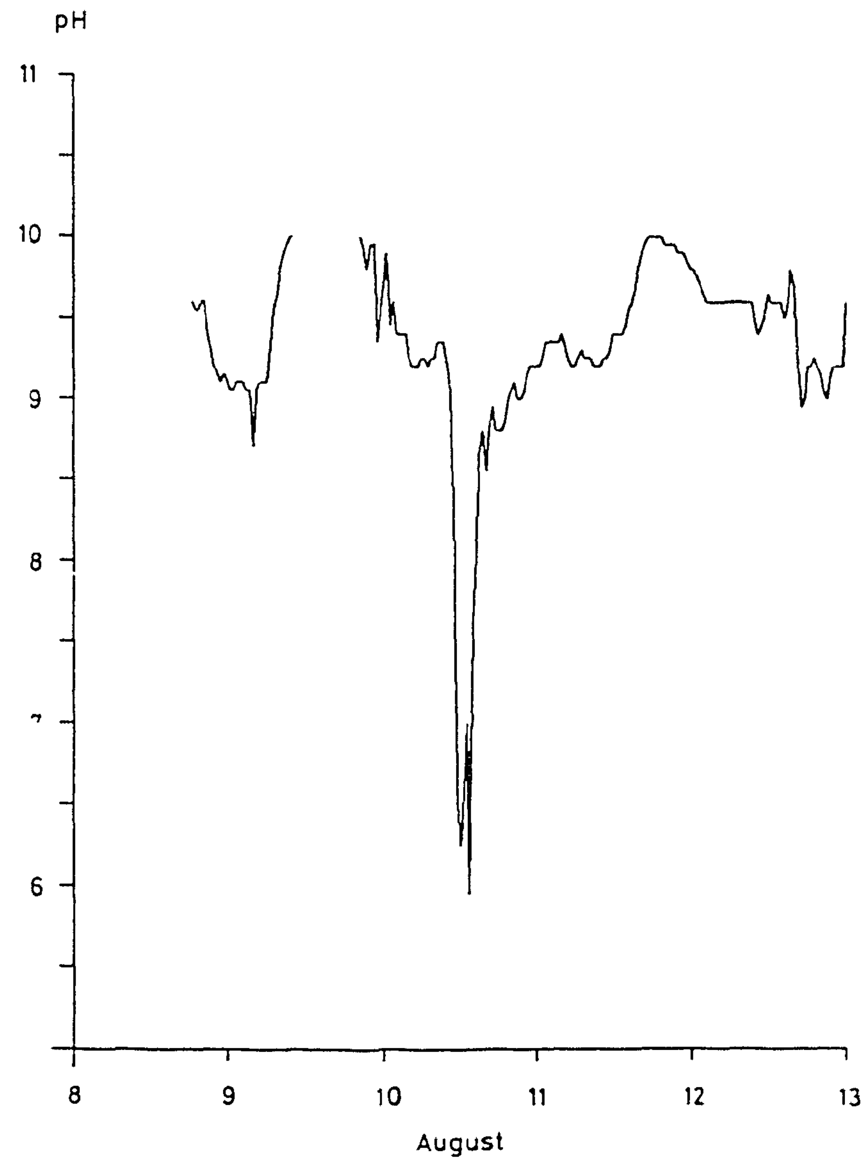


Figure 4.15 Waterfall stream pH, August 1987.



supraglacial debris, may become channel flow without undergoing significant chemical modification. At this time the chemistry of the waterfall stream is thus a reflection of the characteristics of the precipitation event as the bare ice surfaces are not able to buffer the stream waters against an increase in pH. In contrast the measuring site on Peyto Creek is surrounded by an extensive area of ice-cored moraine, the surfaces of which are able to serve as a source of ions for exchange with free hydrogen ions in the rainfall thereby buffering the waters of Peyto Creek against a decrease in pH.

4.4 Fluvio-glacial sedimentary processes

Rates of erosion are likely to be significant in mountain areas where there is a combination of processes of glacial erosion with an efficient removal of deposits through regular meltwater flow. Determining sediment content in glacial meltwaters should improve estimates of denudation rates and help clarify how fluvio-glacial sediment transport relates to the distinct diurnal discharge cycle observed in glacierised basins.

Sediment concentrations at the terminus of Peyto Glacier were determined in 1982 and 1987. Results are given in Figures 4.16 and 4.17. Records indicate a strong dependence of sediment content on discharge. Maximum sediment concentrations tend to occur on the rising limb of the daily hydrograph when the rate of increase of discharge is greatest. Concentrations then fall rapidly on the falling limb when reduced discharges and a correspondingly lower stream competence encourages deposition of sediment within the channel cross-section.

Figure 4.16 Suspended sediment concentrat:on and discharge, Peyto Creek, August 1982.

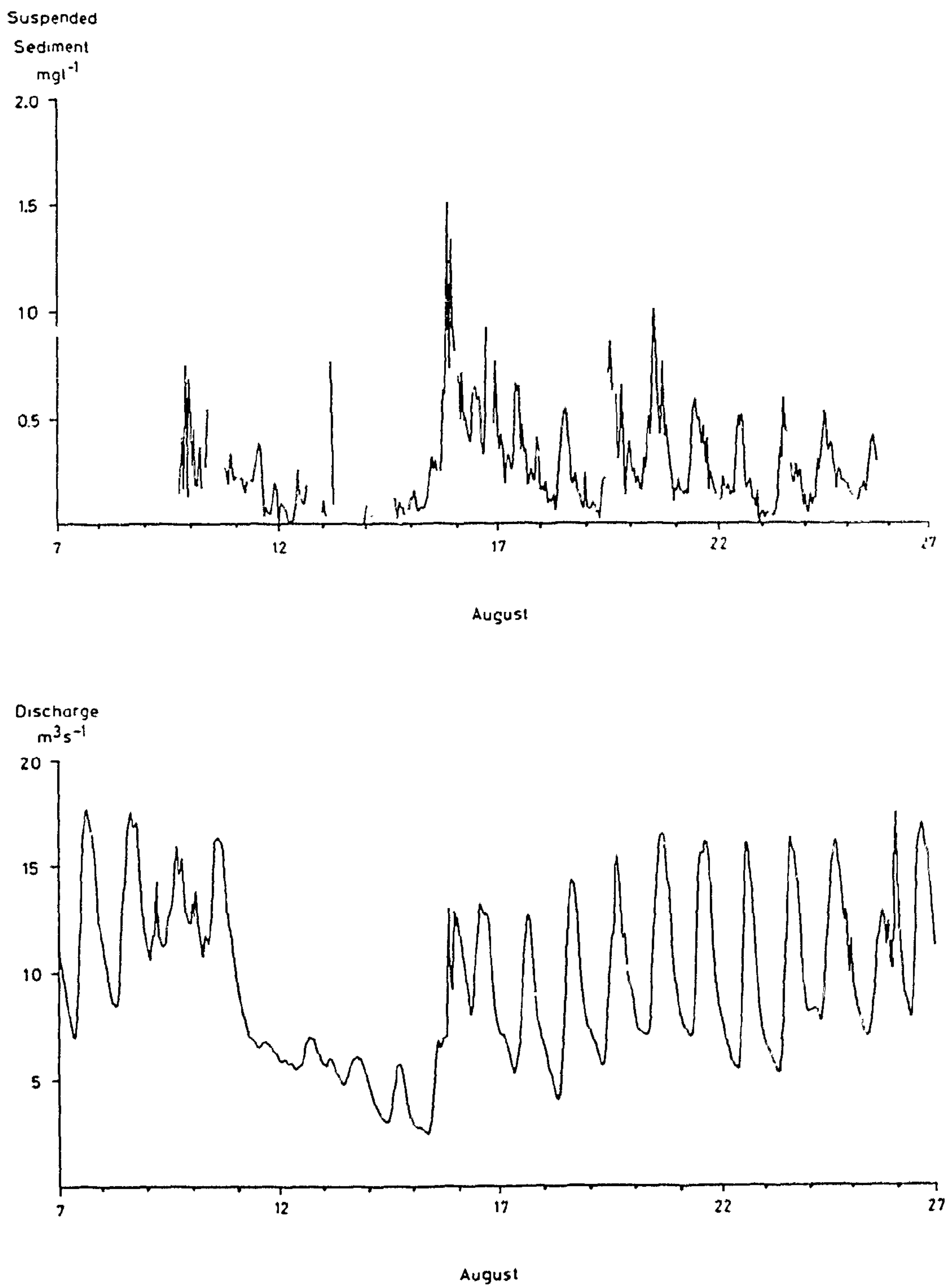
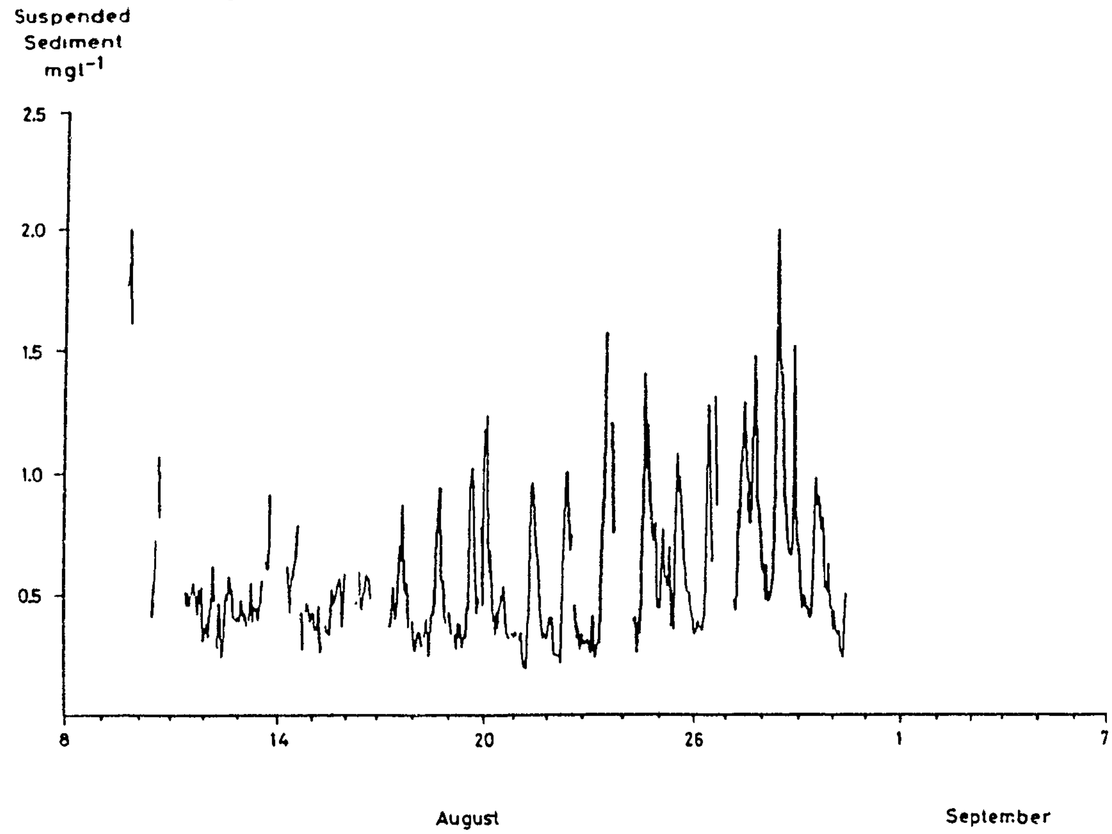
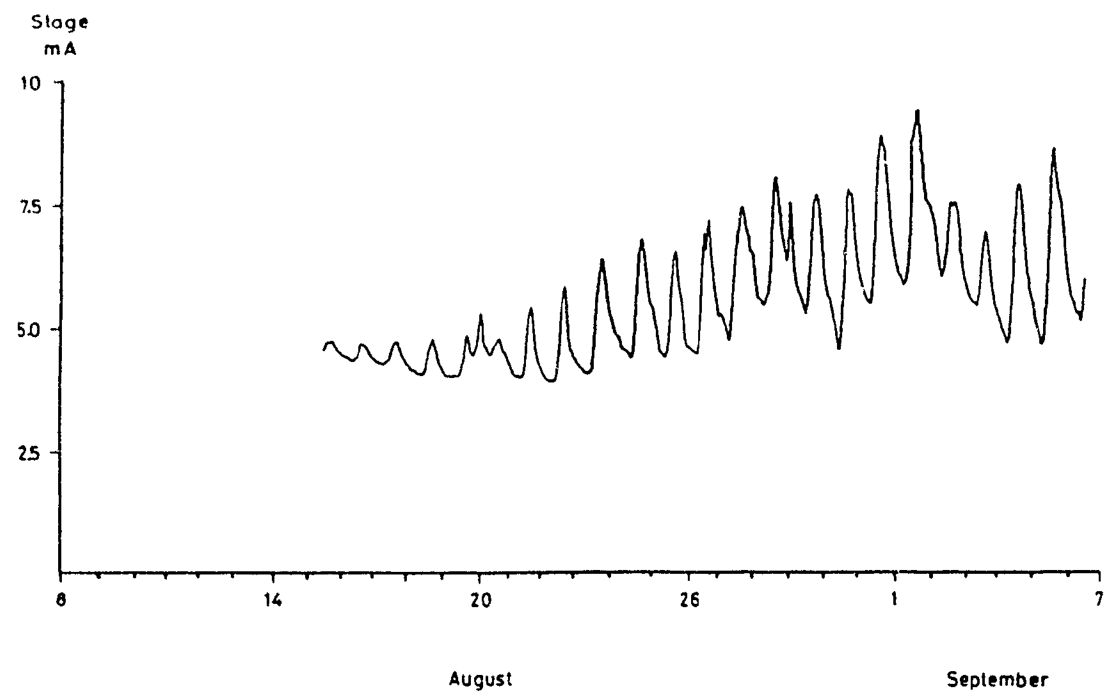


Figure 4.17 Suspended sediment concentration and stage, Peyto Creek, August-September 1987.





Sediment concentrations in 1982 are dominated by a regular cycle from 16 August which appears to have a consistent relationship with the diurnal variation of discharge. Four consecutive days of low flows produced low sediment concentrations in the terminus stream. The return of high discharges through ice-ablation from 16 August produced meltwater flows which had access to considerably greater sedimentary material derived from glacier abrasion and crushing, which is reflected in the high peak in suspended sediment concentration of 1500 mg l^{-1} on that day. Following this peak the diurnal maximum sediment concentration fell until 18 August when three days of steadily increasing discharges were able to account for increased sediment evacuation. Instability in sediment supply is indicated by twin peaks in sediment concentration on 16, 19, 20 August. On these occasions the secondary peak occurs after the diurnal flood-peak had subsided, possibly indicating slumping of material at the margin of subglacial channels after the passage of the daily flood wave had weakened the sedimentary deposits.

In both 1982 and 1987 initial sediment concentrations were low. In 1982 sediment concentrations remained below 500 mg l^{-1} from 10 - 17 August, with little discernible diurnal fluctuation. Discharge was also low at this time. From then onwards sediment concentrations progressively increased, reflecting parallel increases in stage through sustained ablation. On 29 August sediment levels were much reduced although stage was not significantly lower, possibly indicating temporary exhaustion of sediment supply following a storm event the previous day.

To examine the importance of minor variations in suspended sediment and the degree of variance around the hourly sample points the Northants sampler was programmed to collect samples at fifteen minute intervals between 08:00 and 20:00 hrs on 23, 25 and 27 August 1987. The results are given in Figure 4.18, the hourly samples collected under the routine measurement programme being circled and plotted separately to the right of the main diagrams. In some cases variation around hourly samples was considerable. On 21 August in particular sediment reaches a peak concentration of 1400 mg l^{-1} at 14:45hrs, while the concentration at 15:00hrs was 900 mg l^{-1}, the plot of hourly values thus fails to account for this large increase. In general the hourly samples give an accurate representation of sediment concentrations at lower stages and on the falling limb of the hydrograph. Greatest misrepresentation seems to occur at times of maximum flow when highest suspended sediment concentrations are expected. The diurnal peak sediment concentrations on 21 and 25 August at 14:45 and 13:00 therefore do not appear clearly. Similarly on 27 August the peak in sediment concentration between 13:00 and 14:00 is not apparent on the hourly plot, consequently the temporary decrease in sediment concentrations between 13:30 and 15:00 cannot be identified at the larger time-scale. Such behaviour may be a genuine feature reflecting rapid evacuation of accumulated sediment, followed by temporary exhaustion of sediment supply and increased dilution of average sediment concentrations during passage of the diurnal flood peak. Alternatively there could be significant slumping of marginal sediment in basal channels as discussed above.

Plotting suspended sediment concentration with stage in mA in Figure 4.19 over the same four day period as for conductivity in Figure 4.7 illustrates that sediment concentration has a similar relationship with discharge. Clockwise hysteresis was

Figure 4.18 15 minute variation in suspended sediment, 21 August (top), 25 August (middle), 27 August (bottom) 1987. Normal hourly values are circled and plotted on right.

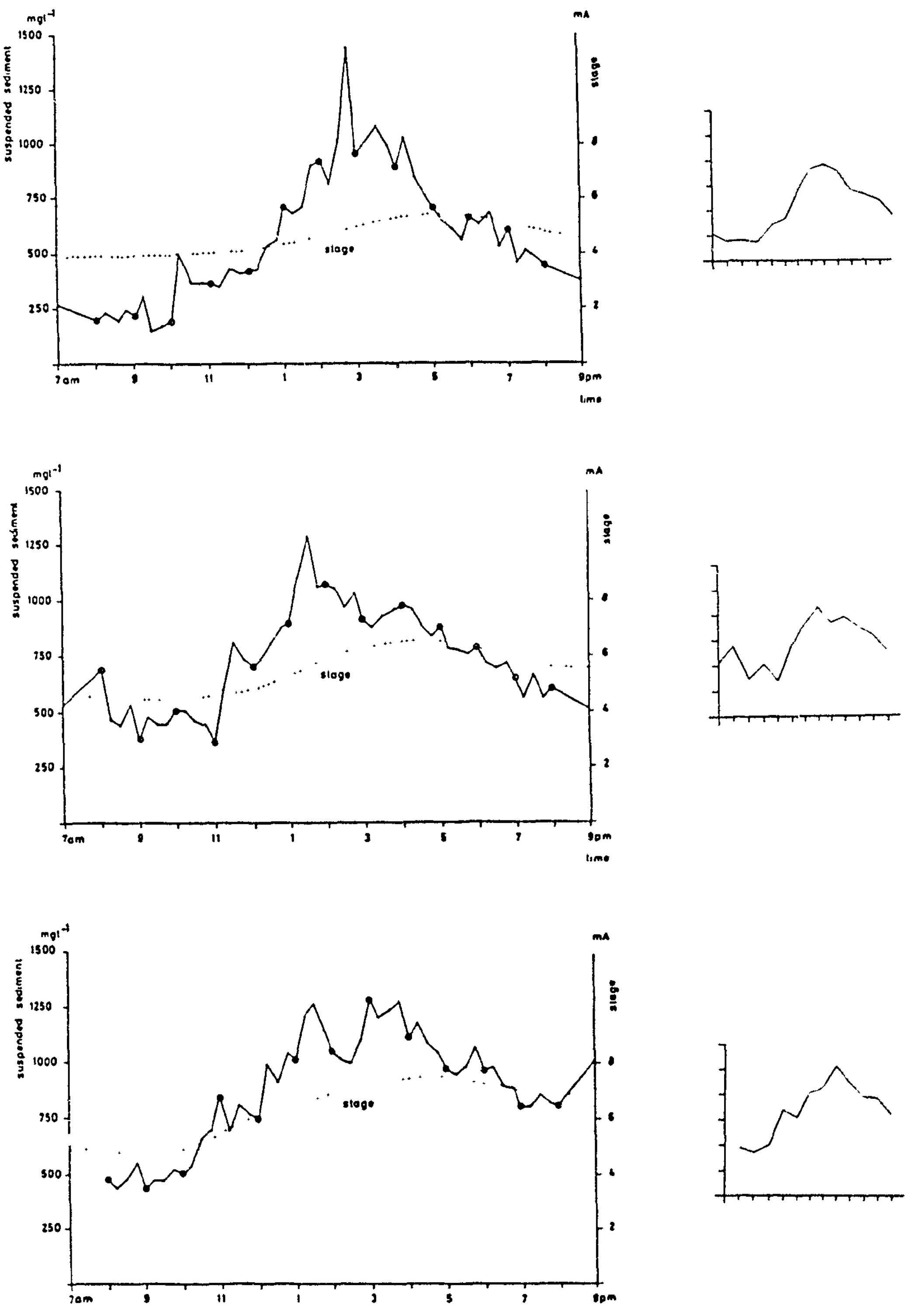
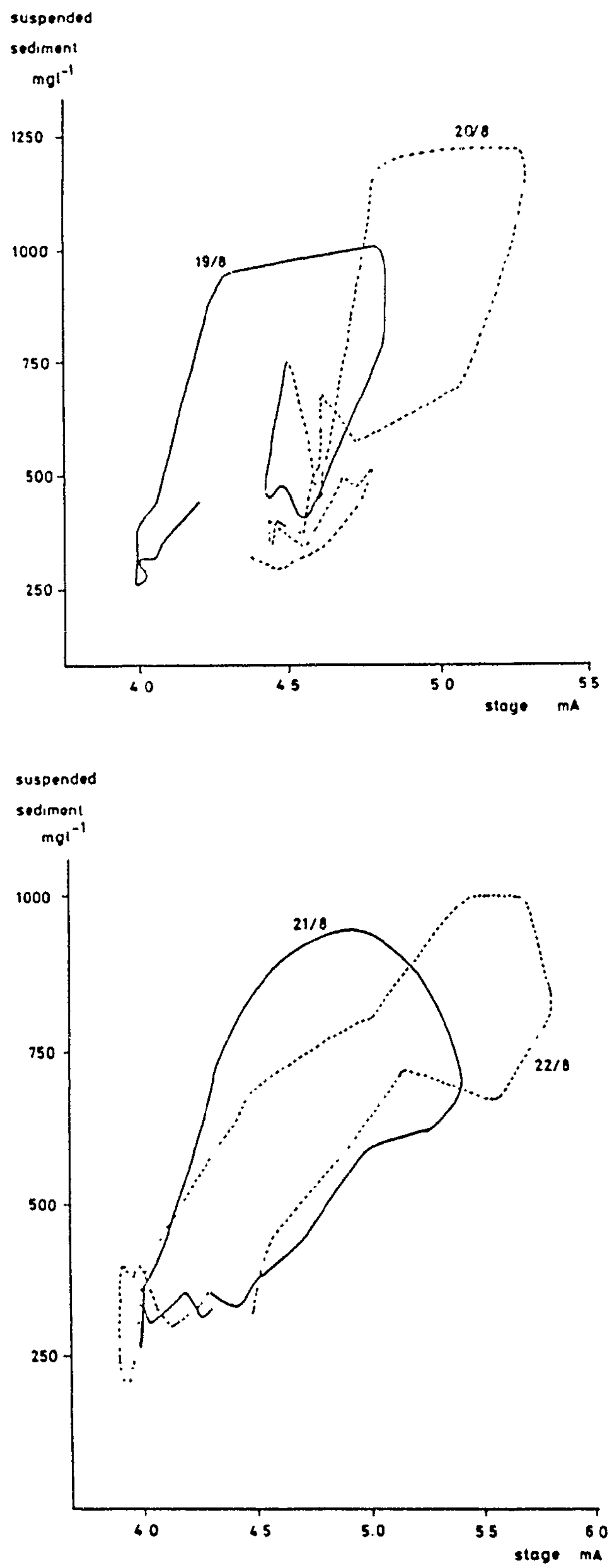


Figure 4.19 Suspended sediment hysteresis, 19,20,21,22 August 1987.



observed with sediment concentrations higher on the hydrograph's rising limb. Overnight on the 19/20 August a small sub-loop occurred, associated with a storm event that night. On 20 August there is a further large separate loop as high water stages encounter basal sediment. On 21, 22 August the sediment concentration loops regain a regular format although higher stages on 22 August were not associated with appreciably greater sediment concentrations.

Study of the time-variation of suspended sediment concentration indicates that diurnal fluctuations are again important. However interpretations of hydrological pathways using this evidence are subject to a few problems. Sediment concentrations given here were obtained using hourly samples at one point in the stream cross-section. As the experiment with a 15-minute sampling time indicated, variation around these values may be considerable, reflecting differences in sediment supply and the degree to which flowing waters have access to sediment sources. Random inclusion of large particles, and the sampling of different size fractions of the stream's load, as the position of the nozzle of the water sampler hose varies with respect to water stage, are likely to influence the results. In addition, despite location of the water sampler as close to the glacier portal as possible, a gully system in the terminal area could have accounted for significant sedimentary inputs into Peyto Creek. Such high increases may occur at times outside the normal diurnal fluctuation of suspended sediment when slope failure occurs on the moraine above Peyto Creek. This would introduce additional 'noise' into the time series.

In view of all the above complications which might be expected to influence recorded values of suspended sediment, the results of the sampling undertaken in both 1982 and 1987 display a very regular form that appears closely related to discharge. In both years sampling was completed late in the ablation season, much higher sediment loads would be expected earlier in the year since the first increase in meltwater will have access to sediment accumulated over the preceding winter. Thus Binda *et al.* (1985) noted a maximum suspended sediment concentration of 3379 mg l^{-1} early in the ablation season at Peyto in June 1981. Changes in the predominance from a glacial to a fluvial regime in sediment discharge are suggested by differences in the hysteresis curves at Peyto. The regular hysteresis curves observed at Peyto in 1987 are highly unusual for a glacierised basin. In 1981 Binda (1984) observed several complicated hysteresis loops and sub-loops at Peyto which were similar to those mentioned by Collins (1979a) at the Gornergletscher in Switzerland. Such features seem to be indicative of reorganisation of basal channels which suddenly make available new sources of sediment. The lack of such changes in sediment load of Peyto Creek suggests greater stability of the main hydrological channels towards the end of the summer.

Chapter 5

Stable and Radio-isotope Analysis

Sampling of water from different locations within the Peyto basin was undertaken in 1978 and 1989 for determination of deuterium and tritium content. The intention was to examine in what way distinctive isotopic signatures of meltwater can be representative of various hydrochemical environments. Total flow within Peyto Creek is composed of waters derived from a variety of sources. While the daily discharge peak is dominated by water derived from melting of ice at the glacier surface, differing quantities of slowly-routed basally produced meltwater, meltwater from winter and summer snowfalls, summer rainfall and groundwater may all contribute to total flow. Relative contributions from the above sources will vary through time producing distinctive annual, seasonal and diurnal flow regimes in glacial rivers. Measurement of the variation in the isotopes, deuterium and tritium, in samples taken from different meltwater sources, together with samples taken from the bulk meltwater in Peyto Creek may permit separation of individual flow components through labelling different meltwater types by characteristic isotope contents.

5.1 Description of Results

Results from 1989 and 1978 are given in Tables 5.1 and 5.2. In 1989 replicate determinations on two samples gave an error of ±4.4‰ and ±1.8‰. This is equivalent to a percentage error of 2.6 and 1.3 per cent, an average of 1.95 per cent. Some difference between samples can be discerned. In 1989 the deuterium content of samples ranged from a maximum of -71.77‰ during a thunderstorm on 1 August to a minimum of -169.4‰ for a fresh snow sample collected on 15 August. All other samples displayed deuterium samples that fell within the range -143.9‰ to -157.2‰.

Samples of water collected from Peyto Creek at the glacier terminus and at the 1987 gauge site (B on Figure 3.1) were within 2.5‰ of each other, the five samples averaging -154.48‰. In contrast the samples of groundwater were isotopically heavier, at -150.8‰ and -151.5‰, although this is barely significant given the error in sample determination. A sample collected from a supraglacial stream displayed a similar value, -151.1‰. Greater range was observed in deuterium content of ice samples collected at the glacier surface. A sample collected in the vicinity of Stake 40, elevation 2217 m.asl. had a deuterium concentration of c. -145‰. This was significantly heavier than two other samples obtained at 2245 m.asl. and 2287 m.asl., where the deuterium contents were -159.4‰ and -157.2‰ respectively.

Four samples from 1989 were submitted for determination of the content of the radio-active isotope, tritium (^3H). Results are given in Table 5.1. Tritium content was low with a substantial error of ±8 T.U., however considerably higher tritium values were recorded in the groundwater samples.

Table 5.1. Isotopic results, July/August 1989

Type of water	Date	Deuterium δD	Tritium T.U.
Precipitation (thunderstorm)	1.08.89	-71.77‰	
(snow)	15.08.89	-169.4‰;-165.0‰	
Glacier terminus stream:			
11:00	1.08.89	-154.1‰	6 ± 8 T.U.
11:00	17.08.89	-155.7‰	
17:00	17.08.89	-155.3‰	
13:40	4.09.89	-153.2‰	
Peyto Creek gauge site:	4.09.89	-154.1‰	
Groundwater:	18.08.89	-150.8‰	
	4.09.89	-151.5‰	14 ± 8 T.U.
Supraglacial stream:	12.08.89	-151.1‰	
Ice (stake 80) 2287masl	3.09.89	-157.2‰	6 ± 8 T.U.
Ice (stake 50) 2245masl	3.09.89	-159.4‰	
Ice (stake 40) 2217masl	3.09.89	-145.7‰; -143.9‰	6 ± 8 T.U.

Table 5.2. Summary of deuterium content of samples, 1978 (from Collins, Unpubl.).

Type of water	Site	Date	Range of deuterium content δD
Snow-melt (seasonal and firn)	Surface of Peyto Glacier	24-28 July	-147.7 - -154.4
Ice-melt	Surface of Peyto Glacier	28 July	-164.5 - -165.0
Snow-melt (seasonal)	Alpine meadow	5 August	-148.9 - -154.4
Runoff from ice-free area	Caldron Lake outflow	5 August	-164.5 - -165.4
Glacial meltwater stream	Peyto Creek	14-28 July	-160.7 - -173.7
		29 July - 12 August	-160.1 - -171.3
		ablation season	-160.1 - -173.7

Deuterium results from 1978 are taken from Collins (unpubl.) and are given in Table 5.2. Mean percentage error in measurement of isotope content was 1.8 per cent, with a range of 0.1 to 4.3 per cent. Samples taken from Peyto Creek in 1978 are isotopically lighter by comparison with the 1989 samples, ranging between a deuterium concentration of -160.1‰ and -173.7‰. Samples of ice-melt collected in 1978 were also lighter, the values falling between -164.5‰ and -165.0‰. Snow-melt had deuterium concentrations between -147.7‰ and -154.4‰.

5.2 Tritium Results

Quantities of tritium are very low, water in Peyto Creek and two ice samples yielded a tritium concentration of only 6±8 T.U. A sample of groundwater had a slightly higher tritium concentration, 14±8 T.U. This contrasts with tritium measurements described in Ambach *et al.* (1976) who found that maximum tritium contents in winter runoff from the Hintereisferner and Rofanche in 1973 and 1974 were between 200 and 250 T.U. Prantl and Loijens (1977) determined tritium content of water within the Mistaya basin between 1969 and 1972, peak concentrations were 572 T.U. and 728 T.U. in summer precipitation between 1970 and 1971. In groundwater they found tritium concentrations of between 353 and 364 T.U. in the North Saskatchewan and Mistaya basins.

Although data are limited, these first tritium results suggest that a reevaluation is needed of current potential applications of this radioactive isotope in glacierised drainage basins. High levels of tritium occurred in the period 1952-1972 before cessation of atmospheric testing of nuclear weapons, although some tritium remained in the atmosphere as the isotope decayed according to its half-life of 12.5 years. It would seem that the low levels of tritium in the samples from 1989 arise solely from natural production of the isotope through cosmic radiation with no anthropogenic input. Measurement of the current tritium content of individual precipitation events at Peyto throughout a year would be desirable. Prantl and Loijens (1977) found highest concentrations of tritium in summer precipitation over the period 1969-1972, which is consistent with seasonal variation in global tritium transport. A similar temporal pattern was also observed in measurements of precipitation between 1965 and 1969 at Vienna (Ambach *et al.* 1973). Higher tritium concentrations in groundwater than in samples of ice-melt and total discharge can therefore be attributed to summer precipitation replenishing the groundwater system, a conclusion in accordance with that of Prantl and Loijens (1977). When discharge in Peyto Creek is dominated by ice-melt, tritium concentrations will be correspondingly low as the tritium concentrations in ice have been depleted through radioactive decay and probably predate the period of atmospheric thermonuclear testing.

It is also possible that groundwater may contain some percentage of old water which has retained some residual tritium, after over a decade of radioactive decay. Further sampling of different water sources for subsequent measurement of tritium content would therefore be desirable, in particular it would be useful to determine if any water sources within the basin retain evidence of the atmospheric nuclear testing of the 1960s and 1970s.

5.3 Deuterium Results

Deuterium concentrations in individual precipitation events may vary considerably. Although only two samples of precipitation were collected in 1989 their deuterium concentrations had a range between -71.77‰ for rainfall, to -169.4‰ for snow. The observations of isotopic content of the different meltwater sources over July and August 1989 are therefore very consistent. Lightest samples tend to be the snow-melt proportion of flow as phase changes during snow metamorphosis selectively enrich samples in the heavy isotope (^2H). Higher concentrations of deuterium were therefore recorded in glacier ice. The differences in deuterium content of samples taken at three different altitudes on the glacier arise as a consequence of the continuing enrichment of ice in the heavy isotope with ablation on the glacier surface. The deuterium gradient at Peyto appears greater than the natural isotope altitude effect of -3‰ per 100 m quoted by Ambach *et al.* (1976) in the Ötztal Alps.

In the absence of more detailed measurements and comparative meteorological data for 1978 and 1989 it is only possible to speculate on possible reasons for the differences in isotopic content in the two years. Samples collected for analysis in 1978 appear to have deuterium concentrations which are lighter by between 5 and 15‰ for similar hydrochemical environments. Lower concentrations in 1978 may be a result of lower temperatures in that year. In 1989 there may have been greater quantities of isotopically heavy water contributing to water flow in Peyto Creek. This may have been derived from greater summer snowfalls in 1989, or an increased proportion of runoff derived from rainfall in this year. Precipitation might either have run straight over bare rock surfaces, thereby immediately contributing to runoff, or have served to recharge the groundwater system, increasing the groundwater contribution to total flow.

One possible reason for the absolute difference in deuterium concentration between 1978 and 1989 may have been a change in the elevation of the main areas of the glacier surface which contribute towards stream discharge. Survey of Peyto Glacier during the summer of 1989 revealed that since a previous map had been produced in 1966, the mean elevation of points in the lower ablation area below the ice-fall had dropped by 60m. The glacier terminus had also retreated by some 100m at a rate of 10m per year. A lower glacier surface would have the effect of increasing the concentration of isotopes as precipitation would occur at lower elevations and warmer temperatures would be expected. Variation in the deuterium content of the three samples of ice collected at different elevations in 1989 indicated that altitudinal effects may be significant although more samples are needed to quantify the relationship for the Peyto basin.

5.4 Summary

The preliminary results outlined above indicate the potential value of collection of isotope data over a full calender year at Peyto. Determination of the variation of natural isotopes can usefully distinguish between certain sources of meltwater flow.

In contrast to deuterium results from Rofenache in the Ötztal Alps groundwater isotopic contents are not significantly lower than ice-meltwater. At Rofenache Ambach *et al.* (1976) explained this in terms of groundwater recharge over the winter by precipitation with a low deuterium content. The tritium results suggest this process does not take place at Peyto, colder temperatures through the winter probably restrict melting and consequently limit the addition of meltwater to the groundwater system over the winter. This conclusion is in accordance with the limited tritium results which indicate distinct differences between groundwater and samples of ice and water from Peyto Creek.

The initial isotope results suggest there is considerable scope for further studies of a similar nature in the Peyto basin. This would ideally entail a fully integrated investigation, including collection of isotope data from snow pits in the accumulation area as described in Krouse (1974) and Krouse *et al.* (1975), and sampling of precipitation events throughout a year. Studies of glacier hydrology have all too frequently depended upon measurement of certain meltwater characteristics for limited periods during the ice-melt season, contributions of different meltwater sources vary in distinct ways throughout the year, consequently it would seem most useful to measure meltwater characteristics over a similar period. Isotope studies have the potential advantage that through sampling of different layers in snowpacks in the accumulation area average precipitation characteristics may be determined, initial metamorphosis removing some of the variation in individual precipitation events.

Chapter 6

Discussion and Conclusions

Determination of the manner of water flow through glacierised drainage basins presents a considerable hydrological problem in how one is to interpret variations in water quality and quantity in a way which might clarify processes of glacier hydrology. In this dissertation an attempt has been made to describe how indices of water quality can be used to determine temporal variations in the route taken by meltwater through a glacier. Drainage of meltwater through snow and ice varies through space and time, some water may rapidly form runoff from the basin, travelling through the glacier in large arterial channels. In other situations water which has been produced by basal melting, or which is draining from storage sites within the basin, may flow slowly in a discrete cavity system at the glacier-bed. Chemical characteristics of snow and ice meltwater will reflect the source and routing of meltwater within a glacier. Initial water quality will reflect atmospheric solute inputs, additional solute may be obtained through reactions with sediment that flowing meltwaters may come into contact with during passage through a glacier. Chemical characteristics of meltwater will therefore depend in part on the degree to which waters are able to access solute sources, which will be a product of the residence time of water within the basin and where in the glacier meltwater is routed, whether at the glacier bed or in ice-walled channels within the glacier. Variation in total chemistry thus provides some indication of switches in the character of flow between high flows, which are dominated by the discharge of fast-routed, chemically-dilute, meltwater, and low flows when slow-routed, chemically-rich, meltwater predominates. These correspond to the englacial and subglacial categories of meltwater described by Shreve (1972).

Simple examination of variations in water quality and quantity should therefore prove to be an invaluable indicator of hydrological processes beneath glacierised basins. The extremely regular diurnal cycle of electrical conductivity in Peyto Creek, during the ablation season in 1982 and 1987, demonstrates how there is a regular fluctuation in areas contributing to total streamflow. Pulses of dilute meltwater provide a constant input to glacial drainage channels, the exact form of the chemograph reflects the quantities of meltwater which may dilute a solute-rich base-flow. As such the electrical conductivity graph will reflect predominant hydrometeorological conditions and any changes which may be induced by passage through a glacial drainage net. Rainfall produces a distinct hydrograph with a characteristic peak of short duration, which is also reflected by a parallel rise in the chemograph.

Examination of electrical conductivity charts at short time scales also provide evidence of releases of small amounts of chemically enriched water at times of maximum and minimum water pressure. Interlinkage between arterial conduits and adjacent cavities is suggested as occurring at times of high and low water pressure. At high water pressures extension of the drainage net will integrate isolated pockets of water which have been stored at subglacial locations and which will have become saturated in solute. The contribution of these waters to total flow will produce a temporary peak in EC at maximum discharge when solute concentrations are a minimum. Similar inter-connection between cavities and conduits at times of minimum discharge will entail drainage of subglacially enriched water into the principal channel. Meltwater may be temporarily stored in the channel margins during high flow, as flows recede drainage of this water will temporarily increase EC.

Measurement of electrical conductivity at a site near the accumulation area enables comparison of meltwater quality at different locations within the glacierised basin. Differences in timing were found in response to the same hydrometeorological conditions as at Peyto Creek while differences in solute concentration reflect spatial variation of the principle sources of solute and changes in the contribution of snow-melt between the study periods. Divergence between EC in a stream draining the accumulation area and Peyto Creek, draining the whole basin, were especially marked in late August when reduced daylight hours inhibited slow percolation of meltwater at higher elevations in the basin. The early morning maximum in EC was therefore more subdued in the upper stream as the slow-routed proportion of flow froze within the accumulation area thereby increasing average densities and helping the gradual metamorphosis of snow to ice.

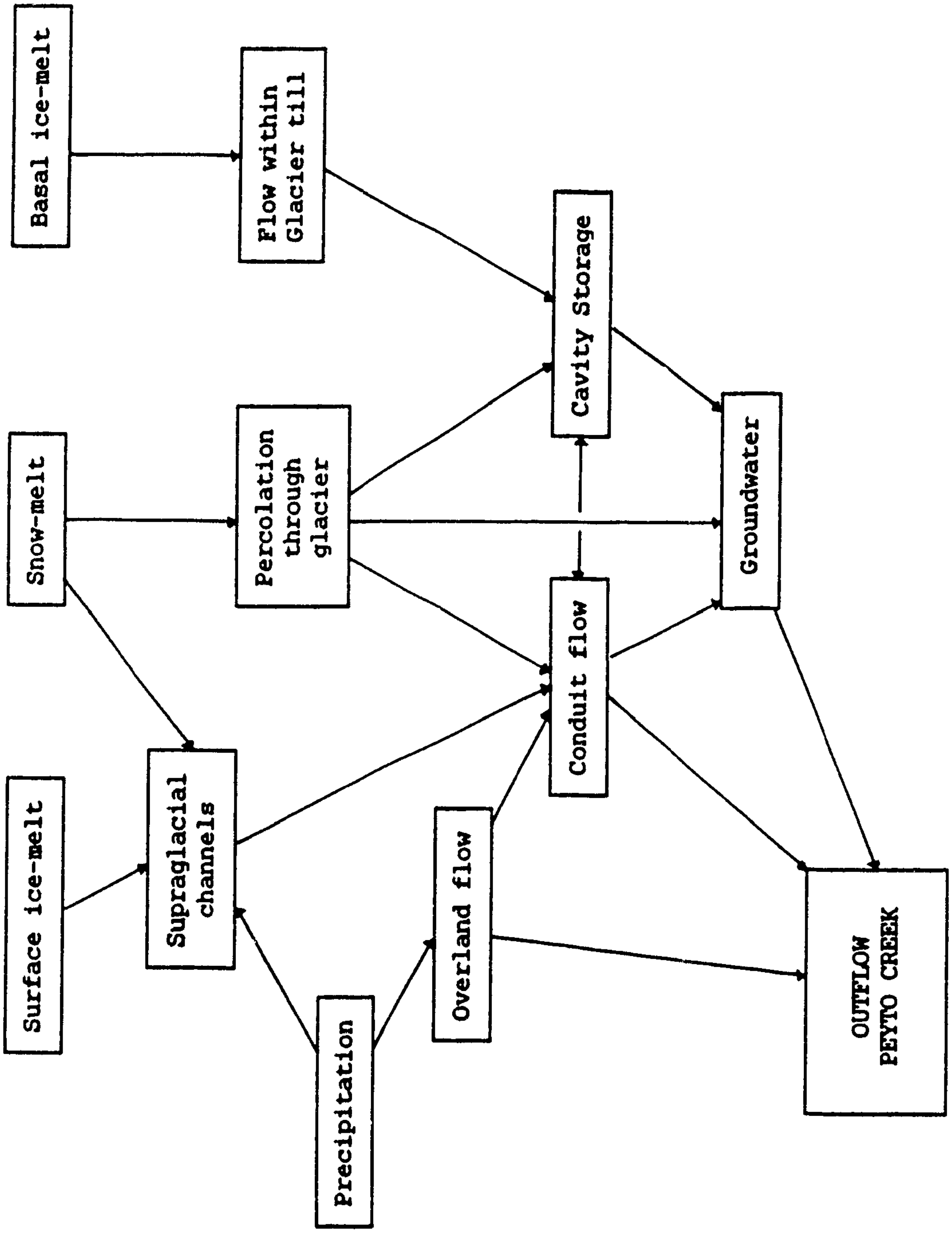
Continuous measurement of suspended sediment concentration over August in 1982 and 1987 demonstrates how sediment levels reflect the diurnal flow variation. At this late stage in the ablation season the principal conduits are well-established, and most sediment is derived from the immediate channel margins. Plotting of hysteresis curves illustrates how sediment concentrations are higher on the rising limb of the hydrograph as sediment deposited at times of falling stage the previous day can be transported as water stage increases. Sampling of meltwater at a 15-minute time interval demonstrates how there may be considerable micro-variation around the simple diurnal trend. While some error will be introduced by the sampling procedures and random inclusion of large particles, some of the changes in concentration may be a genuine feature, for example reflecting slumping of marginal deposits after the diurnal peak flow has subsided. Continuous monitoring of sediment transport would therefore be desirable, in which respect, despite calibration problems, measurement of the attenuation of a light beam emitted by a photo-electric cell immersed in the turbulent flow of the river could be attempted at Peyto. This would initially enable qualitative interpretation of diurnal variations in suspended sediment, while comparison of continuous time series of conductivity with the record from a photo-electric cell would permit examination of any interdependence between the two parameters.

Sampling of water from different hydrochemical environments within the basin for determination of deuterium and tritium concentrations provides some evidence for variations in water quality characteristics. Significant differences were observed in the tritium contents of groundwater and ice-meltwater, suggesting recharge of the groundwater system by summer precipitation, which has a higher tritium content than precipitation at other times in the year. Variation in deuterium content is less significant, although differences in content of the isotope at various elevations on the glacier surface may be attributable to progressive enrichment of the heavy isotope with ablation at the glacier surface. Differences in deuterium content between 1978 and 1989, were significant, corresponding samples being approximately 10‰ lighter in the former year. This may reflect a variety of factors including differences in average temperatures, quantities of snowfall, or a progressive change in the elevation of the glacier surface over the eleven years which elapsed between sample periods.

Characteristics of meltwaters emitted from glacierised areas thus reflect the particular combination of source and routing of water through the basin. An attempt has been made in Figure 6.1 to summarise the different routeways that meltwaters may conceivably follow through the Peyto Basin before joining Peyto Creek. The illustration is necessarily a simplification, as is true of much related to glacier hydrology, and some analysis therefore consists of descriptions of probable routing of water and the likely consequences in terms of suspended sediment and solute load. Other combinations of flow through the Basin may occur in addition to those illustrated in the diagram. On meeting these different water types may also combine in a non-conservative manner thereby altering the characteritics of the bulk meltwater and complicating any attempt at resolving flow into distinct components.

The manner of water flow through glaciers thus still presents certain unanswered questions which will not be resolved until techniques, other than those described here are applied to Glacier Hydrology. Studies in specific drainage basins have still to predict conclusively patterns of runoff variation from glacierised basins and more still needs to be learnt concerning the nature of water channels within glaciers. Electrical conductivity has been routinely used in glacierised basins where the advantage of cost-efficient continuous monitoring of meltwaters has proved very attractive. Monitoring of EC should

Figure 6.1 Possible routing of meltwater through the Peyto Catchment.



continue for comparison of future data with existing records, however additional, complimentary, techniques are essential within a unifying framework. Determination of isotope content provides some indication of source of meltwater, as opposed to a combination of source and routing. Additional stable isotopes may also prove useful within glacierised basins. The use of dye-tracing techniques can potentially reveal the time taken for meltwater to flow from a known point on the glacier to a gauging site. While the use of these techniques have been described in individual basins their simultaneous application in one drainage basin has yet to be attempted. The work described here also illustrates that measurements throughout the ablation season are essential. The glacial drainage net appears to have opened and assumed at Peyto before detailed measurements commenced in August 1982 and 1987. Increasing the period of coverage of solute concentrations to before initiation of snow and ice-melt provides additional information on the chemical characteristics of water routed slowly through the drainage basin, and could usefully be undertaken at other sites.

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