VARIATIONS IN THE RELATION BETWEEN SUSPENDED SEDIMENT AND SOLUTE DELIVERY IN GLACIAL MELTWATERS, MAXWELL GLACIER, YUKON TERRITORY.

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ABSTRACT

The purpose of this thesis is to investigate the interactions between meltwaters and sedimentary material at the base of an alpine glacier which are indicated by intra- and interseasonal variations of discharge, solute and sediment outflow from the terminus. During the 1989 and 1990 ablation seasons, discharge, suspended sediment concentration and dissolved solids draining from Maxwell Glacier, St. Elias Mountains, Yukon Territory, were obtained at hourly intervals. The discharge time series of Maxwell Creek reveals that flow varies at a period of one to two weeks, at a diurnal level and over one to three hours. Over all three time scales, electrical conductivity is inversely proportional to discharge suggesting that during high diurnal flows, outwash water is derived primarily from surface melt while at low nocturnal discharge, water originates from subglacial and englacial stores. Variations in suspended sediment concentrations are proportional to discharge on a weekly time scale but become less well defined on a diurnal and shorter time period. Examination of short-term variations in solute concentrations reveal distinctive behaviour of the glacial drainage network at times of minimum and maximum water pressure, when subglacially enriched pockets of water may become integrated with the total outflow.

In spring, when supraglacial meltwater begins to access the subglacial drainage system the basal water pressure is raised. Large areas previously hydraulically isolated are integrated releasing stored basal sediment. During periods of increasing discharge in the early summer hydrological events resulting from gradual expansion and/or isolation of subglacial cavities, temporary blockage of subglacial conduits or from input from adjacent gravity deposits occur. As flow increases, conduit diameters increase and water gradually spreads out over the glacier bed as a diffuse network of linked cavities. Throughout the ablation season, different areas of the sole are integrated with flow. Towards the end of the ablation season, meltwater supply declines resulting in the reduction of flow in subglacial passageways and reduced sediment and solute supply.

Future progress in the understanding of the character of the glacial drainage system will only be achieved with increased field observations. Continuous high frequency observations are necessary to understand the glacial drainage system because they are not restricted to specific conduit geometries and behaviours. Field data will provide the best evidence available to constrain theoretical models. Increased understanding of glaciological processes can then be incorporated into development strategies for areas sensitive to glacial discharge-sediment regimes.

RÉSUMÉ

Le but de cette recherche était d'examiner les interactions entre les eaux de fonte et les sources de sédiments à la base d'un glacier alpin reflétées dans les variations intra- et intersaisonniers de la concentration en soluté et en sédiments des eaux de fonte au terminus de ce glacier. Pendant les saisons d'ablation de 1989 et 1990, le taux d'écoulement, la concentration des sédiments en suspension, et le total des solutés dissous provenant du glacier Maxwell, Monts St. Elias, Territoire du Yukon, ont été obtenus à chaque heure. Les séquences d'évènements hydraulogiques du Maxwell Creek ont révélé que l'écoulement varie sur une période d'une à deux semaine, à une échelle journalière et à une échelle d'une à trois heures. À chacune de ces trois échelles, la tension ionique est inversement proportionnel à l'écoulement. Ceci suggère que les eaux de fonte s'écoulant d'un glacier, durant des hauts taux d'écoulement diurne, provienne primitivement en surface du glacier alors que les bas taux d'écoulement nocturne sont d'origine de réserves sous-glaciares et intraglaciares. Les variations dans la concentration de sédiment en suspension sont proportionnel à un écoulement hebdomadaire mais deviennent moins bien définies à une échelle de temps diurne ou plus courte. L'examen des variations à court terme des concentrations en soluté des eaux de fonte ont révélé un comportement distinct du sytème d'écoulement glaciaire lorsque la pression d'eau est à un minimum et à un maximum. À ce temps, des cavités subglaciares, enrichies d'eau, s'intègrent graduellement au débit.

Au printemps, lorsque les eaux de fonte supraglaciare accèdent le système d'écoulement subglaciare, la pression d'eau de base s'élève. De grande aires de la plaque fondamentale du glacier, hydrauliquement isolées auparavant, s'intègrent au système pour ainsi approvisionner les eaux de fonte subglaciare en débris. Au début de l'été, quand le débit s'accroit; les évènements hydrauliques qui résultent de l'agrandissement ou l'isolement de conduits subglaciares ou de la contribution de dépôts de gravité ont lieu. Lorsque le débit s'accroît, le diamètre des cavités s'accroît et l'eau s'étend progressivement au-dessus de la plaque fondamentale du glacier sous l'aspect d'un réseau de cavité raccordées par des conduits. Au cours de la saison d'ablation, différentes parties de la plaque fondamentale du glacier s'intègrent au débit. Plus tard dans la saison, l'alimentation des eaux de fonte diminue. Conséquemment, l'écoulement dans les conduits subglaciaires et l'apport de sédiments et de solutés sont réduits.

Une augmentation d'observations sur le terrain peut assurer l'approfondissement des connaissances sur le caractère de système de vidange glaciale. Des observations fréquentes et continues sont nécessaires pour comprendre le système d'écoulement glaciare puisque celui-ci n'est pas seulement déterminé par la géometrie et le comportement de conduits spécifiques. Les données obtenues sur le terrain fournissent les évidences les plus utiles à contraidre les modèles théoriques. Une meilleure compréhension de processus glaciaires pourra d'abord être incorporée aux stratégies de développement des régions sensibles aux régimes d'écoulements glaciaires chargés en sédiments.

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CHAPTER 1

INTRODUCTION

'The Alpine and high-mountain areas of the world play an extremely important and distinctive role in the hydrological processes of the planet, and in the regional hydrology of all continents. It is in the alpine regions where meteorological, glaciological, periglacial and hydrological phenomena have most intimate and complex interaction and variability on short space scales and short time scales, yet the results of these interactions have a profound effect on hydrological regions over much greater distances and much longer time scales.'

(Roots and Glen, 1982, p. V)

1.1 PREAMBLE

The supply of glacial meltwater is an important non-renewable resource for hydroelectric power generation, water supply, and irrigation in all industrialised and industrialising mountain areas of the world (Meier and Roots, 1982). The timing and magnitude of water storage and release have critical implications for the development of economic activities. For example, glacier-fed rivers provide much of the water supply for agriculture in regions such as the Canadian prairies, Pakistan and India (Fisheries and Environment Canada, 1978; Butz, 1989), along with hydroelectric power production in France, Norway and Switzerland (Meier, 1969; Wold and Østrem, 1979; Tangborn, 1984; Hooke *et al.*, 1985). The Grande Dixence Hydroelectric Power Company, Switzerland, operates an innovative and complex network of reservoirs designed to collect glacial meltwaters for the generation of hydroelectricity from 35 drainage basins ranging in area from 1 to 80km^2 (Lang, 1975; Bezinge, 1987). Not only do large-scale industrial operations benefit from a glacial river regime, but smaller mountain settlements may also rely totally on meltwater for irrigation of agricultural lands (Butz, 1989). The diversion of glacial meltwater in arid mountain zones provides necessary water for irrigation to settlements where land would otherwise remain unproductive (Price, 1981). The presence of glaciers has a positive effect on these operations by reducing the variability from year to year in comparison with ice-free basins (Krimmel and Tangborn, 1974). Although glaciers do reduce streamflow variability over a number of years, sudden flood events can occur (Østrem, 1975; Clarke, 1982) causing short-term downstream flooding.

Hence, it is necessary to understand the changing character of glacio-fluvial hydrology and the hazards it poses. This may allow for more efficient utilisation of the resource as future demands possibly increase. Variations in the quantity of snow and ice stored and released from glacierised basins strongly influences natural hydrological regimes. Unlike other types of streamflow, glacier runoff is characterised by large diurnal fluctuations and maximum flow during summer at a time when water is otherwise scarce. Glaciers affect agricultural, municipal and industrial projects by their manner of storage and rapid release of water, by increasing sediment and solute loads of streams, by providing a constant source of streamflow and by resembling naturally regulated reservoirs (Tangborn, 1984).

Glacier runoff depends on the climate at the glacier surface and the physical mechanisms controlling the flow of water within the glacier and at the bed. Meteorological data, specific to an individual glacier, changes with time of year and between years (Stenborg, 1970; Fountain and Tangborn, 1985). In comparison with ice-free basins, the presence of glaciers in drainage basins affects the reliability of runoff by reducing the variability from year-to-year (Krimmel and Tangborn, 1974; David, 1989). Glaciers are continually in the process of slowly adjusting to relatively rapid climate changes by the release and/or storage of precipitation. Ice temperatures largely control the behaviour of a glacier. Heat derived from three sources: the atmosphere, the geothermal heat flow and friction within the glacier and at the bed produces ice temperatures within glaciers. Climatic regimes at the glacier surface, together with local surface topography, control local mass balances. Solar radiation and associated air temperature variations primarily control the diurnal and annual cycles of glacial runoff within the basins. Since the greatest proportion of meltwater produced within glacierised basins is a result of melting ice, annual meltwater yields relate directly to high summer energy availability (Collins, 1984). Hence, annual fluctuations in climate directly control the storage and retention of ice and water in a glacierised basin. The temporary storage and aperiodic release of water are typical features of glacier hydrology and runoff that varies at time scales from an hour through to a complete ablation season. However, glaciological controls on hydrological regimes are important moderators of intraseasonal runoff and timing of peak discharge (Collins, 1984).

Since, glacier meltwaters form an important resource in all high mountain areas of the world (Meier and Roots, 1982), an understanding of the production, storage, and transport mechanisms of glacial meltwaters is important. Glaciers are not only stores of water, but are also powerful erosional tools. The most commonly recognised processes of glacial erosion are: plucking or quarrying, crushing and shearing, abrasion, and entrainment (Fenn, 1987). Measurements indicate the importance of subglacial erosion whereby waters at the glacier bed result in the removal of large quantities of fine sediments (Østrem, 1975; Collins, 1979a). The initial loading of suspended sediment in glacierised basins provides the basis for the sediment regime of the river. The glacier, gravity deposits and glacial deposits are three hydrological sub-systems that provide the initial loading of suspended sediment glacial meltwaters (Johnson and Kruszynski, 1990). Streams highly charged with suspended sediment can create problems for hydroelectricity power plants and irrigation projects unless sediment is accounted for in their design (Tangborn, 1984). Sudden incursions of vast amounts of poorly-sorted sediment can reduce the effectiveness of water reservoirs and power turbines. Even though storage and release of water have been observed in almost all mountain areas of the world, the production and transport mechanisms of englacial and subglacial waters are not well understood (Paterson, 1983). However, the temporally changing characteristics of sediment concentrations and solute loads provide insight into the role of glacier dynamics in the production and transport of sediments in a glacierised environment.

Meltwaters emerging from glacier termini acquire their water quality characteristics during the passage through conduits, cavities, and tunnels that comprise the subglacial hydrological system (Collins, 1987). Meltwaters interacting with unconsolidated sediment at the base of the glacier produce variations in water and sediment outflow from glacier termini (Collins, 1979b). The variation in water and sediment outflow from glacier termini provides means of estimating glacier erosion (Østrem, 1975) and chemical weathering (Reynolds and Johnson, 1972). These variations of cation and suspended sediment concentrations in meltwater emerging from glacier termini allow the understanding of the form, development and stability of subglacial drainage systems (Collins, 1979c, 1987, 1989). Patterns of variation throughout two consecutive ablation seasons with contrasting hydro-meteorological conditions have been used to help determine periods of change and stability of the subglacial drainage network during spring and summer cycles of development (Kruszynski and Johnson, 1993). Available energy for entrainment of sediments in a stream increases with increased flow (Morisawa, 1968). However, although closely related to discharge on an annual and seasonal basis, over diurnal and shorter periods, transport and concentration of fluvioglacial sediments are not well correlated with flow. "Exceptional water discharge events" as described by Hooke *et al.* (1984), Østrem (1975) and Mathews (1973) can be interpreted as indicators of changes in the internal drainage of glaciers. Johnson (1991a) has shown that pulses in discharge occur frequently and are concentrated in the spring and early summer when linked conduit and channel drainage networks respond to the increased basal water pressure from the spring melt. Periods of prolonged ice-melt and heavy rainfall result in peaks of sediment concentration in glacial meltwaters (Østrem, 1975). During the early development of the drainage network, events involving high basal water pressure and enhanced sliding also contribute to fluctuations of sediment and solutes (Iken *et al.*, 1983; Iken and Bindschadler, 1986) but those have yet to be directly correlated to sediment pulses. Østrem (1975) described incursions of vast amounts of sediment comprising substantial fractions of the annual total sediment load for several glaciers and may point to abrupt dislocations of subglacial flow pathways at the glacier sole, access to new pockets of sediment at the sole or sediment derived from adjacent gravity deposits (Collins, 1989; Johnson, 1991a). The diversion of conduits and creation and closure of cavities comprising the subglacial hydrological network produce variations in sediment and solute concentrations (Raiswell, 1984) in the form of short-time frame pulses to meltwaters. It is therefore possible to gain an understanding of the glaciological and hydrological processes governing hydrological regimes in glacierised basins by the measure and analysis of meltwater quality characteristics.

The purpose of this thesis is to examine the characteristics of a brief period of discrete field observations of discharge, suspended sediment concentration and solute load to ascertain the interactions between meltwater and sediment at the base of an alpine glacier. Patterns of such variation recorded from bulk meltwaters in peak and recession flow conditions for two consecutive ablation seasons are deemed necessary to determine periods of development in the subglacial drainage system. The intensive data collection possible during the sampling period permits some useful conclusions to be drawn concerning temporal and spatial controls of suspended sediment transport in this glacierised catchment.

1.2 Objectives

The objectives of the research are three-fold. The first reflects a need to investigate fluvioglacial characteristics and processes of meltwater drainage from an alpine glacier. The second is to assess the use of electrical conductivity as a surrogate measure for cation concentration at the sole of an alpine glacier. The third deals with the investigation of the intra- and interseasonal development of an alpine glacier's subglacial drainage system in order to establish the varying magnitudes of characteristics in glacier-fed streams. In detail, the objectives of this study are:

1. to measure variability of streamflow, suspended sediment concentration, electrical conductivity, and dissolved solids content in meltwaters draining from a glacier-fed tributary stream of Maxwell Glacier[1] during the 1989 and 1990 ablation seasons and deduce hydrological and glaciological processes occurring beneath the glacier;

2. to examine the importance of temporal variations in major cation species in glacial meltwaters versus electrical conductivity as an indicator of source and routing of glacial meltwater flow through the subglacial drainage network;

3. to assess the form, stability and development of the subglacial drainage system with a view to determine periods of change and stability of the drainage network during the ablation season.

[1] The toponyms 'Maxwell Glacier', 'Maxwell Valley', and 'Maxwell Creek' are not recognized by the Canadian Gazetteer for this site; rather they are officially identified as the adjacent glacier, valley, and creek to the east of the current site. These toponyms are used solely for convenience by the author.

CHAPTER 2

THEORETICAL CONSIDERATIONS

2.1 INTRODUCTION

Within the alpine glacier system, water is an important medium and a primary mechanism for movement and transport of weathered products from within and beneath the glacier. Chorley *et al.* (1984) defined erosion as not only the breakdown of in-situ material but also the transport of the weathered product away from its source. Although glaciers are efficient erosive agents, this efficiency is dependent on the interdependency of processes, differential erosive efficiency and spatial and temporal selectivity in sediment production (Fenn, 1987). However, the ability of ice as an erosive agent is secondary to its ability to store sediment. Drewry (1987) demonstrated that glaciers behave more as sediment stores for they transfer sediments at a very slow rate ranging from 100-1000 years. Unlike glacier ice, glacier meltwaters are the primary mechanism able to erode large quantities of weathered sediment from englacial and subglacial stores within short periods of time. Meltwater flow with sufficient turbulence to maintain fine particles in suspension is temporally limited to the ablation season and spatially limited to the subglacial drainage network (Collins, 1989). Thus the changing temporal and spatial characteristics of the englacial and subglacial drainage system is ultimately reflected in outwash records. Analysis of temporal variations in water quality and sediment load characteristics provides information regarding the interaction of meltwater and sediment at base of an alpine glacier

(Østrem *et al.*, 1967; Hooke *et al.*, 1985; Collins, 1989). A model for the subglacial drainage system and the mechanisms controlling the weathering and transport of material at the glacier bed can be deduced from theoretical models and field observations of sediment and solute acquisition.

This chapter is intended to provide a summary of sediment and solute transfer processes within glacierised alpine basins. Attention is focused on fluvioglacial components of sediment transfer and considers the importance of erosion and transport of sediment at the glacier bed and through the proglacial zone. Three recent comprehensive reviews containing considerably more detail, and covering topics considered to go beyond the scope of the present paper, are given by Gurnell and Clark (1987), Röthlisberger and Lang (1987) and Kelly (1990).

2.2 GLACIAL HYDROLOGY

2.2.1 Components of Runoff

Melting at the glacier surface is usually the most important source of water on a temperate alpine glacier (Shreve, 1972; Theakstone and Knudsen, 1981). Meltwater produced internally by the dissipation of mechanical energy in the flow of ice or water, frictional melting at or very near the glacier sole, geothermal heat flow at the contact between the ice and the substratum and by liquid precipitation events also produce meltwater in glaciers (Röthlisberger and Lang, 1987). For a typical alpine glacier, surface melting during the spring and summer ablation seasons provides the most important source of meltwater (Shreve, 1972). Röthlisberger and Lang (1987) have illustrated that surface melting represents the greatest proportion of meltwater for a temperate glacier. Surface melt rates vary between 0.1 and 10m yr^{-1} depending on altitude, while water from frictional and geothermal melting produces rates orders of magnitude lower (~10^{-2}m yr^{-1}). Therefore, within the context of glacial hydrology, emphasis should be placed on the role of melting processes at the glacier surface.

The melting process at the glacier surface can be explained by understanding the energy balance (heat budget) at the glacier surface. The energy available (Q_M) for melt on a horizontal snow or ice surface of unit area and over a unit of time is described by:

$$Q_M = Q_{NR} + Q_S + Q_L + Q_P + Q_G$$

where

Q_{NR}	is net radiation;
Q_S	is sensible heat;
Q_L	is latent heat of condensation or evaporation;
Q_P	is heat provided from precipitation;
Q_G	is heat from conduction in the snow pack;
Q_M	is heat used for melt or gained from refreezing of meltwater.

The melted water equivalent of snow or ice per unit area per unit time (melt rate) can then be determined by:

$$M = Q_M / S$$

where S is the latent heat of melting (333.7 Jg^{-1} at 0°C).
(Röthlisberger and Lang, 1987)

Results from heat balance measurements at the Aletschgletscher at two different altitudes: accumulation area and ablation area; has shown that net radiation is important in the ablation process at the glacier surface (Röthlisberger and Lang, 1987). Maximum melt rates were found at times of high total net radiation and when sensible and latent heat fluxes were high. The net radiation contributes 71% of the total energy balance followed by sensible heat (21%) and latent heat (8%) (Lang *et al.*, 1977; in Röthlisberger and Lang, 1987). The importance of net radiation balance at the glacier surface becomes clear when discussing the impact of snow coverage on the glacier's heat balance. Snow-cover is an important component of the glacier's mass balance and hydrology governing daily and seasonal variations in the melt rate. For a snow-covered surface the mean daily albedo for that surface will be approximately 0.80, whereas a snow-free glacier surface has a lower albedo of approximately 0.27 (Henderson-Sellers and Robinson, 1986; Bezinge, 1987). Therefore, under the same conditions of incident radiation, much of the available energy is reflected rather than absorbed on the snow-covered surface retarding melt and runoff during the ablation season.

During the ablation season, meltwater runoff from a glacierised basin arises from two sub-catchments, the glacier and the ice-free areas as defined by:

$$Q_t = Q_i + Q_s + Q_p + Q_g$$

(Collins and Young, 1981)

where Q_i is the glacial component of discharge, which includes icemelt, firnmelt and snowmelt from the ice covered areas. From the ice-free areas the discharge has been broken down into Q_s, meltwater originating from snowpack depletion in early spring; Q_p, liquid precipitation and Q_g, from groundwater comprise the summer runoff component.

During the winter, snowfall on the glacier surface increases the surface albedo and hence reduces the amount of melt and runoff. In spring as the daily mean temperature and amount of incoming radiation increase, the snow-cover provides much of the initial source of meltwater. The distribution and amount of snow coverage are two important factors in the storage and release of meltwater over and into the glacier (Meier and Tangborn, 1961). As the spring progresses into summer, the increased amounts of net radiation, sensible heat and latent heat provide energy to melt and gradually reduce the extent of snow coverage. Although snowmelt contributes to the outwash record, surface melt of the glacier will be delayed until the snow-cover has been appreciably reduced allowing for increased melt and runoff (Krimmel and Tangborn, 1974). As the ablation season continues into summer, greater areas of the glacier become exposed and gradually increase glacier meltwater flows. However, if winter snowfall accumulations are small, meltwater runoff will begin soon after spring melt begins. Low snowfall accumulation does not permit the storage of meltwater within the snowpack and hence delaying runoff. Retention of meltwater occurs not only in the snow-cover but also occurs seasonally in the firn aquifers. Tracer experiments by Oerter and Moser (1982), at the Vernagtferner, Oetztal Alps, Austria, notes that recharge occurs in the firn aquifer during the ablation season. Firn recharge results in the seasonal storage and release of meltwater. Mid-summer snowfall events have also been shown to reduce glacier runoff due to the ability of snow to reflect incoming solar radiation rather than be absorbed by the ice (Collins, 1977). In general, a thick snow-cover will store rainfall and meltwater, whereas a bare ice surface permits rapid runoff. Therefore, the beginning of meltwater flow into the glacier may occur well after the beginning of snowmelt as a result of thermal and hydraulic retention capacity of snow, firn, englacial and subglacial drainage systems.

Production of meltwater can also occur at the ice-bedrock interface. At the base of a temperate glacier, a meltwater layer between the sole of the glacier and its underlying bed is produced and maintained by the sliding and creeping action of the glacier under the force of gravity. Glacier creep is the way in which glaciers move as a result of ice deformation. Gravitational forces continually develop strains greater than the critical limit of ice. Both the force of gravity and the polycrystalline nature of the ice lower the critical stress point at which plastic flow is dominant and results in the movement of the glacier under Glen's Flow Law (Glen, 1955). Glen's Flow Law is a result of laboratory studies on the rate of deformation or strain rate of ice. The behaviour of ice under stress can be approximated by the power function,

$$\varepsilon = A\,\tau^{n}$$

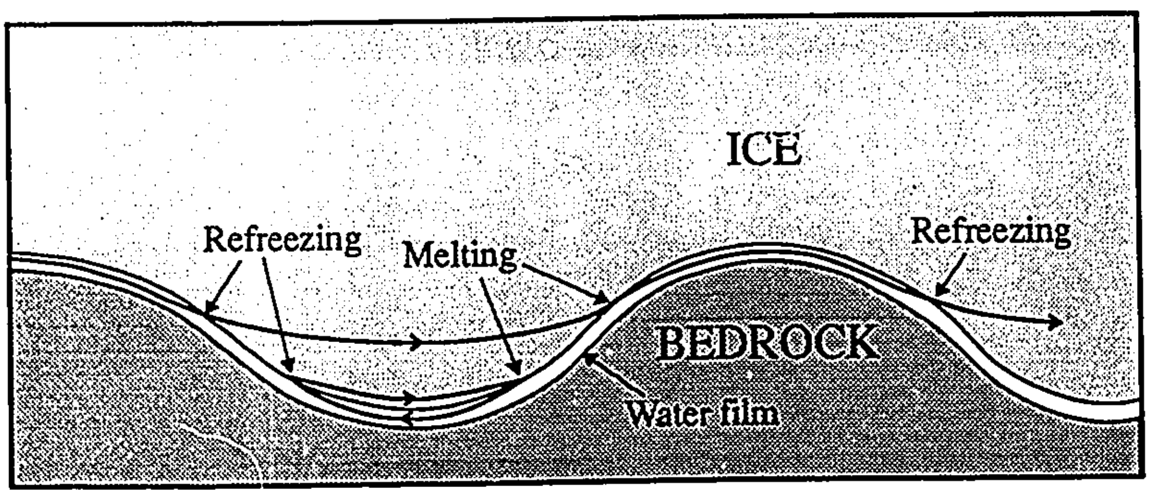
where ε represents the strain rate, A is a constant depending on ice temperature, τ is the effective shear stress and n is an exponent with a mean value of 3. The significance of this relationship is that it demonstrates that the rate of deformation is highly sensitive to shear stress. As a result of the strain in the ice, enough heat is produced to cause melting of the ice through the vertical profile. Application of Glen's Law gives important insights into glacier behaviour. It explains why most deformation takes place in the basal layers of a glacier where shear stresses are highest and it explains why glacier ice is unable to withstand high shear stresses without deforming. However, it also explains why there is creep within cold based glaciers that are unable to slide. Creep in warm based glaciers is probably found only to occur during the winter when ablation is minimal and water volumes at the glacier bed are insufficient to allow sliding. Drewry (1987) has calculated the contribution of meltwater from creep is low at approximately 0.01 to 0.02m yr^{-1}.

Weertman (1957) described two forms of sliding, enhanced deformation and regelation. Both processes involve two important factors; 1) bedrock protrusions and 2) the force exerted by a column of ice. Stresses on the ice result in the dissipation of mechanical heat to the ice producing meltwater for the drainage system. Temperate glaciers move across their bed by basal sliding and regelation (Kamb, 1970 and Weertman, 1979) producing greater dynamic pressures on the stoss side of small bedrock protuberances causing the ice to melt (Hallet, 1979b; Lliboutry, 1979). The meltwater produced on the stoss side (up-glacier side) of the bump then migrates to an area of lower pressure on the lee side and refreezes (Figure 2.1). The processes of sliding and regelation are intrinsically linked to the availability of meltwater. It has been demonstrated that the sliding of the ice over bedrock protuberances lowers basal shear stress allowing sliding velocities to increase (Weertman, 1957, 1964, 1986). Prolonged exposure to high dynamic pressure results in increased regelation and meltwater production. The efficiency of regelation and basal sliding not only depends on the availability of water but also the size of the protuberance.

Theoretically, if the bedrock protuberance is greater than 0.1m regelation slip is least effective (Nye, 1973a). If during the ablation season bumps are larger than this size water may accumulate on the lee side forming water filled cavities where the ice has become separated from its bed due to plastic flow (Walder and Hallet, 1979). Water may eventually migrate under changing flow conditions to larger conduits and contribute to the outwash record (Kamb, 1970).

Depending on local geological conditions, geothermal heating may also be a source of water from a glacier. Paterson (1983) has suggested that geothermal sources may contribute an annual melt rate of 6mm per year at the base of a glacier at pressure melting point. The impact of water derived from geothermal sources to the outflow record is low because it continues throughout the year with little variability.

Figure 2.1: Melting-refreezing process of glacier ice over bedrock of rough topography (after Lliboutry, 1979).



Groundwater flow and precipitation events are other sources of water to the glacier system. Water derived from snow melt and storm runoff from valley sides may enter the glacier hydrological system from medial streams or via groundwater flow (Johnson and Kruszynski, 1990). Aperiodic variations of outflow records can be attributed to extreme precipitation events occurring during the ablation season. Extreme flood events may contribute to the hydrograph record, however, compared with glacier melt its effect is usually minimal.

2.2.2 Supra- and Englacial Water Routing

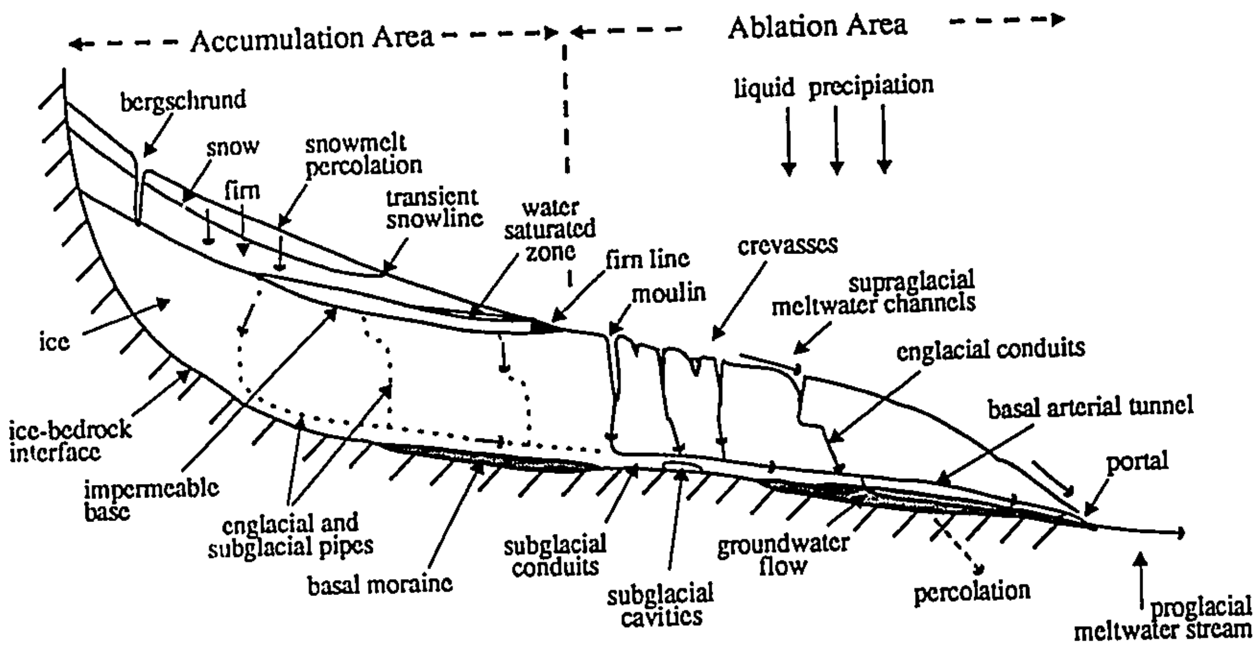
The movement of meltwater over and through a glacier varies temporally and spatially. The manner in which meltwater enters into the glacier's hydrological system varies according to the glacier surface conditions and source of meltwater. The sources and drainage paths that meltwater may follow through a glacier are shown in Figure 2.2 (after Collins, 1977).

Within the accumulation zone of a temperate glacier the surface layer consists of snow and firn. This layer of snow and firn acts as an unsaturated porous medium through which meltwater may percolate (Meier, 1973; Ambach *et al.*, 1981) until it reaches a layer of considerably lower permeability, such as glacier ice. This layer of snow and firn forms an aquifer in which water is stored and flows under gravity according to Darcy's Law,

$$Q = k\, i\, A$$

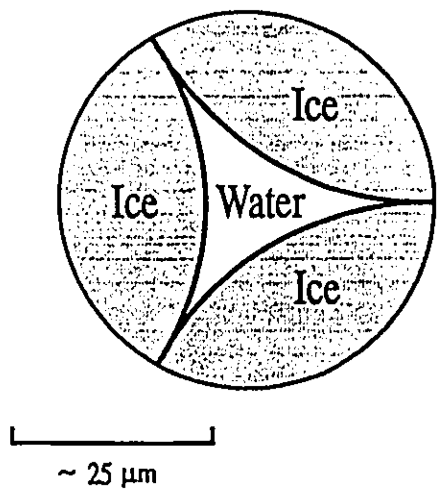
where Q is discharge, k is the coefficient of permeability, i is hydraulic gradient and A is cross-sectional area (Meier, 1973). Due to greater volumes of meltwater flow with increasing depth meltwaters become channelled as they percolate through the ice and snow. Meltwaters reaching the ice-firn transition then create a water saturated zone due to the lower permeability of ice below. Significant quantities of meltwater may be stored within the firn depending on the season (Fountain, 1989). The drainage of stored meltwater through the firn from the above snow pack and saturated zone can drain from the firn aquifer in three ways: outflow at the firn line into channels on the glacier surface, drainage into crevasses and seepage through the glacier ice (Röthlisberger and Lang, 1987).

Figure 2.2: Diagrammatic cross section of the internal drainage system of a glacier (after Collins, 1987).



Seepage of meltwater through the firn and into englacial conduits is attributed to the presence of veins. The term 'vein' is used by many authors to describe capillaries situated along three-grain intersections and that these veins should join together at four-grain intersections to form a network of capillary sized tubes along which water can move (Nye and Frank, 1973). These veins are triangular in cross section (Figure 2.3) and have been observed in samples collected from depths up to 60m in Blue Glacier, Washington (Raymond and Harrison, 1975). Nye and Frank (1973) concluded that the ice is permeable and that water may percolate through the fine grained ice at rates as high as 1m yr^{-1} from the surface. Lliboutry (1971) questioned Nye and Frank's arguments regarding permeable ice stating that permeable ice would disappear by frictional melting and water pressure, but Nye (1976) pointed out that there was not enough water supply for this to happen. Raymond and Harrison have shown that the contribution of frictional melting is limited in coarse-grained ice due to the presence of air bubbles that they felt would result in a significant reduction of permeability.

Figure 2.3: Cross section of a vein, where three grain boundaries meet (after Nye and Frank, 1973).



The enlargement of veins and passageways due to frictional melting is an important process in the development of englacial pipes and conduits. As intergranular flow increases the meltwater descends and seeps into cracks and fissures allowing the development of the englacial drainage system. Shreve (1972) showed that as the flow increased in the larger passage more heat per unit wall area would be generated and would allow larger passages to gradually increase in size at the expense of smaller one. Raymond and Harrison (1975) observed such tubes in samples collected from the Blue Glacier. Thus, there is an arborescent network of tubes in the ice extending upstream within the glacier.

Meltwate... lowing on the surface in the ablation area of the glacier, whether derived from the accumulation area or the ablation area, quickly flow either into large surface streams or disappear into crevasses or moulins supplying water to englacial conduits. On the glacier surface meltwater flow becomes concentrated in supraglacial channels. Meltwaters tend to exploit structural weaknesses in the ice and enlarge their cross-sectional area by the addition of mechanical heat energy along the wetted perimeter of the channel. These channels may eventually reach the terminus of the glacier, however, they also tend to quickly find their way into the glacier via moulins or crevasses.

Crevasses are formed as a result of increased tensile stresses exceeding the tensile strength of the ice when there is a sudden steepening in the glacier surface. Surface meltwaters flowing into the crevasse descend rapidly and may enlarge the passage and propagate downward until it reaches the glacier bed. If the crevasses are not too deep the meltwater may fill the crevasse and overflow into channels that lead further down glacier. The motion of the glacier over its bed also results in the closure of crevasses. Meltwater flowing into crevasses can maintain the connection with the deeper part of the englacial system by the dissipation of mechanical heat of flow that results in the formation of moulins. Holmlund's (1988) study on the geometry of moulins on Storglaciären, Sweden, has found that moulins are generally 30-40m deep and may connect with other moulins originally located along the same crevasse joining and flow into englacial and subglacial conduits.

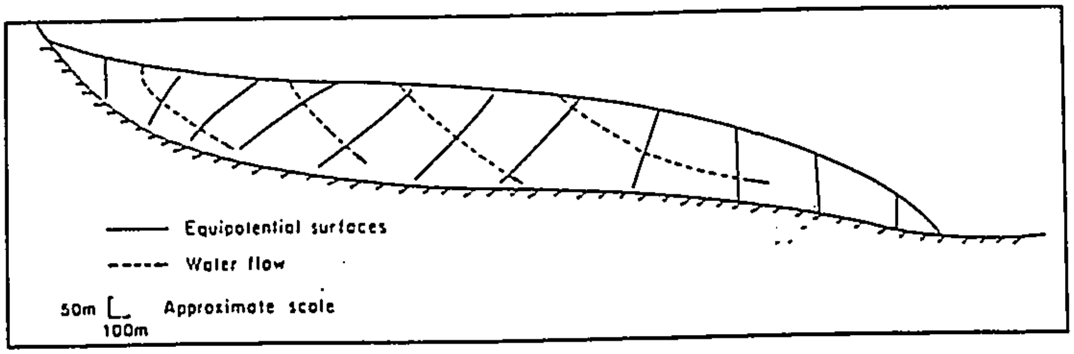
Englacial conduits are described as being circular in cross-section and full of water (Röthlisberger, 1972). The dimensions of the conduits are maintained by an equilibrium between inward ice pressures and by the dissipation of heat at the ice-water interface. Ice pressures tend to close the conduit through ice deformation because of overburden ice pressure greater than the water pressure. Water flow through conduits has the effect of melting the ice by the dissipation of kinetic energy (Nye, 1973b). Meltwater in englacial conduits flows in the downstream direction of the glacier perpendicular to the equipotential surfaces as defined by Shreve (1972):

$$\Phi = \Phi_o + \rho_w g Z + \rho_i g(H - Z) + \rho(\tau)$$

where Φ is the potential, ρ_w and ρ_i are the densities of water and ice, respectively, g is the acceleration due to gravity, H and Z are the elevations of the glacier surface and of a point within or on the bed of the glacier, respectively, τ is the rate of closure of passages by plastic flow, and $\rho(\tau)$ denotes a contribution to the potential that is a function of τ. The first term to the right of the equation is a reference potential, the second term is the potential energy of the water due to its height above sea level, the third is the pressure of the water due to ice overburden pressure, and last part of the equation is a pressure difference between the water and the ice that results in the opening or closure of the conduit by plastic flow. If the horizontal and vertical values within the equation are increased to allow for increasing distance from the terminus, assuming that the gradient in $\rho(\tau)$ is smaller than all the other elements, and the equation is set to zero, it is shown that the equipotential surfaces within a glacier dip up-glacier at approximately 11 times the slope of the glacier surface (Figure 2.4). Thus, englacial conduits will be aligned horizontally towards the terminus but with a subtle downward component.

However, Hooke (1984) calculated that the conduits are not filled by water and are at atmospheric pressure. He demonstrated that once the channels reach a diameter of 3 to 4m, the mechanical energy dissipated by the descending water would be enough to melt the walls faster than the walls could close by ice deformation. This would result in a conduit rapidly increasing in diameter and tending to become vertical as melt rates increase. Hooke (1984) however notes that this would not occur if constrictions in the conduit produce backwater effects.

Figure 2.4: Longitudinal cross-section of a glacier showing equipotential surfaces dipping up-glacier and theoretical flow directions of englacial water routing (after Hooke, 1989).



2.2.3 Water Routing Beneath a Glacier

Meltwater produced at the glacier surface or internally via the dissipation of heat energy reaches the glacier bed through a system of intergranular veins, moulins and conduits. Theoretical models of subglacial hydraulic conditions assume steady state conditions at a glacier bed that is impermeable and rigid (Röthlisberger, 1972; Weertman, 1972; and Walder 1982). However, geological evidence given by Clarke *et al.*(1984) has suggested that regions underlain by easily erodible material (example, Iceland) and regions of recent uplift and intense glacier erosion (for example: St. Elias Mountains, Himalayas and Andes) may provide the geological factors necessary for a permeable and deformable glacier subsole (Clarke *et al.*, 1984). Therefore, if the glacier bed is permeable, water may flow along the ice-bed interface or seep into the bed. The water that cannot drain through the bed must then flow along it. The distribution of water at the glacier bed probably does not remain constant but changes with time (Clarke *et al.*, 1984; Collins, 1987).

Much work has been focused on the role of water flowing at the bed of a glacier and how it affects glacier sliding (Iken, 1981; Mathews, 1964a). Accumulating evidence favours a model for the subglacial drainage network as a winding system of linked-cavities transected by a few large, broad conduits (Hooke, 1989). The average flow at the glacier bed is described as being generally not normal to equipotential surfaces and largely controlled by ice overburden pressure and bed topography (Drewry, 1987). Theoretical papers presented by Lliboutry (1968), Shreve (1972), Röthlisberger (1972), Weertman (1972), Nye (1973b, 1976), Iken (1981), Walder and Hallet (1979), Walder (1982) and Weertman and Birchfield (1983a, b) have all presented arguments for the flow of water either in sheets, channels or cavities under a glacier.

Röthlisberger (1972) and Nyse (1973b) proposed two distinct methods in which water may flow at the glacier bed. Weertman (1972) describes these variations in conduit flow as Röthlisberger channels (R-channels) and Nye channels (N-channels). R-channels are formed by the flow of water at the glacier bed incising upward into the ice and the latter by flowing water incising downward into the bedrock. Nye (1973b) proposed that if a meltwater channel remains at the same location in consecutive years a concentration of fluvio-glacial erosion will promote erosion into the bedrock. N-channels are found to parallel the direction of glacier flow and have a depth greater than the amplitude of the largest undulation in the bed. Water produced elsewhere along the ice-bed interface may remain in the form of a thin water film. Within the deglaciated zone of the Blackfoot Glacier, Montana, Walder and Hallet (1979) have found channels parallel to the former direction of ice flow. Drewry (1987) suggested that the channels may have been formed by large volumes of meltwater flowing at the glacier bed. Clarke and Mathews (1981) have also suggested that large volumes of meltwater released from the drainage of an ice-dammed lake cut N-channels into the bedrock of the eastern edge of Donjek Valley, St. Elias Mountains.

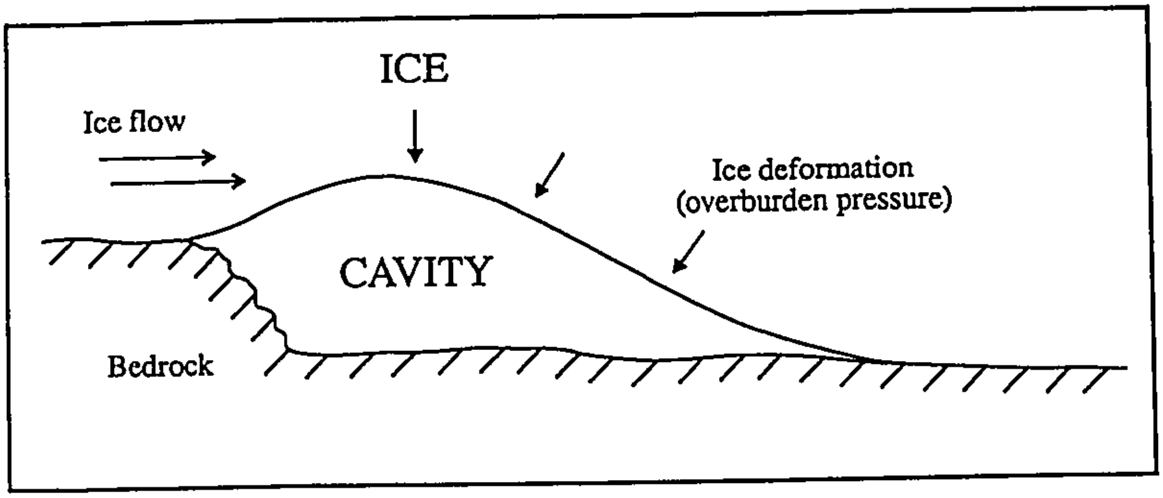
Unlike Shreve's (1972) paper dealing with the influence of ice pressure on the direction of water flow through and under glaciers, Röthlisberger presents a theory regarding the character of the flow in the hydraulic system. Röthlisberger's hydraulic-systems approach predicts that the melt rate within subglacial tunnels will equal that of the closure rate. The energy expenditure that determines the melt rates on conduit walls is linearly dependent on discharge and conduit closure rates are linearly dependent on conduit radius. Increasing the discharge two-fold with the overburden pressure remaining the same results in an increase of conduit diameter. However, if the closure rate is greater than the melt rate the conduit will close.

Reducing the conduit diameter and constricting the flow of meltwater increases the potential gradient and leads to higher water pressures throughout the conduit system. The closure of the conduit and increased water pressure has been observed indirectly by Kamb *et al.* (1985) and Hooke *et al.* (1989) by direct field measurements of water pressure during the autumn and winter. Increased water pressures in the winter indicate that after a period of readjustment to reduced meltwater flow in autumn the conduits begin to close (Kamb *et al.*, 1985; Hooke *et al.*, 1989). Weertman (1972) suggest that as pressure continues to increase within conduits the water may be forced out beneath the glacier forming a thin film of water. Although it is uncertain whether flow will remain constricted to conduits or spread out into a thin film of water, it is known that increased water pressures vary directly with increased ice velocities (Iken and Bindschadler, 1986).

Weertman (1972) proposed a model for a thin film of water less than 4mm thick flowing along an impermeable glacier bed. He argued that discrete channels would be unable to collect meltwater as overburden pressures would drive the water away from the conduits under the ice forming a relatively widespread and thick sheet of water. Under steady-state conditions with water pressure near or equal to that of the ice overburden pressure water would flow freely at the base (Weertman, 1972). Weertman's water sheet will be present provided that the water transported through Nye channels is only a minor part of the total flow (Weertman, 1986). Weertman's (1986) analysis, based on a very simplified glacier bed, assumes that the glacier rests on an impermeable bedrock surface that is irregular in both the transverse and longitudinal directions. However, some topography at the base of a glacier is complicated by small-scale protuberances on the bed where bed slope angles exceed those at the glacier surface. When ice flows over a bedrock protuberance the pressure on the stoss side of the bump is higher than the overburden pressure of the ice, and the pressure is lower on the lee side. If the ice flow is fast enough or the bump large enough the ice may separate from the bed forming a cavity. Water produced on the stoss side of the bump may find its way into the cavity and if the water is under pressure the cavity will increase in size. The stability of sheets of water at the glacier bed has been investigated by Walder (1982). Walder (1982) suggests that sheet flow at the base of a temperate glacier is nearly always unstable on planar glacier beds except for films of water (up to 4mm thick) overlying a rough bed producing quasi-stable sheet flow. Weertman (1986) however notes that even though a thick water film may be unstable and eventually flow into R-channels (Walder, 1982; Weertman and Birchfield, 1983b), R-channels themselves may become unstable (Weertman and Birchfield, 1983b) and transform back into a film of water.

However, Kamb (1987) noted that when ice flows over protuberances in the glacier bed, the ice becomes separated from the bed (Figure 2.5). If meltwater accumulates within the cavity and if the water is under pressure, the size of the cavity will increase. Lliboutry (1983) has shown that when the ice again meets the bed on the lee side of the bump, the cavity or channel becomes constricted or even pinches out. Hooke *et al.*(1985) suggested that constriction would increase water pressures and result in the outflow of water from R-channels to a thin film under the ice forming sheet flow. However, water flowing at the glacier bed will avoid regions of high cryostatic pressure and remain concentrated in channels (Walder, 1986). Hence, if the supply of meltwater is adequate, channels between cavities will remain open leading to the formation of a linked-cavity drainage system (Iken and Bindschadler, 1986; Kamb, 1987).

Figure 2.5: Vertical cross section of a cavity parallel to the basal sliding direction (after Kamb, 1987).



The linked-cavity configuration is formed by the lateral spreading of pressurised water flowing perpendicular to equipotential surfaces. In some areas flow may parallel the contours (Kamb, 1987). Unlike Röthlisberger's (1972) conduit theory, in which water pressure decreases with discharge increase, the water pressure in Kamb's linked-cavity system varies directly with discharge. In Kamb's linked-cavity configuration the closure rate of the cavity is of less importance than that of the amount of water flow. Although a 50% decrease in water flow would result in the same decrease of melt rate, this would result in only a small amount of cavity closure. Contrary to Röthlisberger's arborescent network of larger conduits growing at the expense of smaller ones, the linked-cavity system can remain linked to other cavities and flow out from different areas of the glacier (Collins, 1989).

Evidence for bed separation beneath a valley glacier on deglaciated bedrock surfaces as well as evidence for the connecting channels has been described by Walder and Hallet

(1987) during their examination of recently deglaciated limestone formerly underlying the Blackfoot Glacier, Montana. Thin deposits of calcite and silica in cavities and channels scoured into the bedrock are thought to be formed by expulsion from the regelation water when it refreezes on the downstream side of the bump (Walder and Hallet, 1979).

Numerous dye tracer experiments conducted during summer months have shown that velocity of water flow in subglacial drainage networks does not reflect that of a tortuous linked-cavity system (Stenborg, 1969; Behrens *et al.*, 1975; Burkimsher, 1983; and Iken and Bindschadler, 1986). Flow velocities of 0.2 to 0.7 ms^{-1} were often recorded in experiments in which tracers have been put into moulins or boreholes on glaciers indicating a transfer time far higher than that would be possible in a tortuous linked-cavity system. Burkimsher (1983) observed in his dye tracer experiments at Pasterzengletscher, Austria, repeated peaks on the time versus concentration curve suggesting that water from many moulins flowed into a number of discrete lateral drainage courses. Iken and Bindschadler (1986) reported, however, that they observed multiple peaks in the concentration versus time curves. They proposed that the flow of water passed through a subglacial drainage system with many subdivisions and storage cavities. Hooke (1989) suggests that this accumulating evidence favours a model for the subglacial drainage network as a winding system of linked cavities transected by a few large, broad conduits.

The form of the drainage system is not only linked to glacier's sliding movement and separation of ice from the bed in the lee of some bumps. The ability of such a system to retain its form during metamorphism is a question to be addressed. Fowler (1987) suggested that equilibration of water pressures in conduits and cavities will favour a stable linked-cavity model. The stability of the linked-cavities between seasons, when discharges are low, conduits shrinking and water pressure rising, is unknown. However, Kamb (1987) and Fowler (1987) have stated that if the previous years tunnel system does collapse new tunnels will form by growth and coalescence of links. The form and stability of the linked-cavity drainage are thus dependent on the sliding velocity, the bed geometry, the water pressure, and the water discharge in the conduits (Hooke, 1989). Seasonal variations in all the aforementioned support the model for a linked-cavity network transected by a few broad low conduits.

2.2.4 Temporal Variations of Glacier Discharge

The storage and retention of ice and water in a glacierised basin are primarily controlled by the annual and diurnal fluctuations in climate (Lang, 1973) and the physical mechanisms controlling flow of water within the glacier and at the bed. Solar radiation and associated air temperature variations primarily control the diurnal and annual cycles of glacial runoff within alpine glacier basins. Glacier runoff is largely characterised by a baseflow composed of several components: groundwater flow; englacial storage; and the drainage of the firn aquifer. Diurnal components of runoff superimposed on diurnal baseflow are meltwaters from the lowest part of the glacier basin, draining supra- and subglacially to proglacial streams; and meltwater from the ablation area flowing directly into englacial and subglacial drainage systems (Röthlisberger and Lang, 1987). Interactions between hydro-meteorological conditions at the glacier surface and the flow of water through englacial and subglacial drainage networks produce variations in glacial discharge (Johnson, 1991a). Patterns of discharge variation throughout consecutive ablation seasons are considered to provide insight into the development and form of the subglacial drainage system (Collins, 1984, 1989).

Discharge of glacial meltwater streams fluctuates greatly during seasons, days and hours. On a short time scale there are fluctuations lasting a matter of minutes or seconds thought to be a result of the changing orientation of the drainage system and release of stored water and sediments (Röthlisberger and Lang, 1987; Johnson, 1991a). The seasonal development of the subglacial drainage may best be reflected in the diurnal variations in runoff at the gauge station at Pegelstation Vernagtbach, Oetztal Alps, Austria in Figure 2.6. During the period of 1974-80, measurement of the variable runoff at the Vernagtbach suggests a progressive development of the drainage system of an alpine glacier in the early part of the ablation season and its significant increase in extent during the remainder of the ablation season (Lang, 1973; Oerter *et al.*, 1981). Elliston (1973) made the important point that, however characteristic the fluctuation, it is superimposed on a baseflow accounting for the bulk of the discharge. The discharge record of a proglacial stream over the ablation season is, therefore, composed of a chain of repeating hydrographs that vary in size and shape over the ablation season (Elliston, 1973). However, in addition to the diurnal baseflow fluctuations, glacier discharge is characterised by extremely variable flow conditions within the ablation season. These aperiodic and extreme variations in discharge provide further insight into the stochastic elements of alpine hydro-meteorology (Figure 2.7) and the abrupt changes in the form and stability of the subglacial drainage network (Hooke *et al.*, 1984; Johnson, 1991a).

Temporal variations in the stability and form of the linked-cavity network are to be expected (Röthlisberger and Lang, 1987) as in the case of loss of flow through subglacial intakes of glacier d'Argentière (Hantz and Lliboutry, 1983). During the ablation season changing flow conditions and ice and water pressures slowly result in the expansion or constriction of cavities and conduits (Röthlisberger, 1972; Shreve, 1972; and Hooke, 1989).

During the late fall and winter, meltwater production decreases until it is non-existent. As a result of decreased meltwater flow, reduced melt rates within englacial and subglacial conduits results in the closure of conduits and cavities due to ice overburden pressures. In the spring, the drainage network is not yet fully developed causing meltwater to be stored within the glacier and at its bed. Water pressure builds within these cavities and conduits

Figure 2.6: Diurnal Variation of runoff of Vernagtbach, Austrian Alps, 1974-1980 (data source: Oerter *et al.*, 1981).

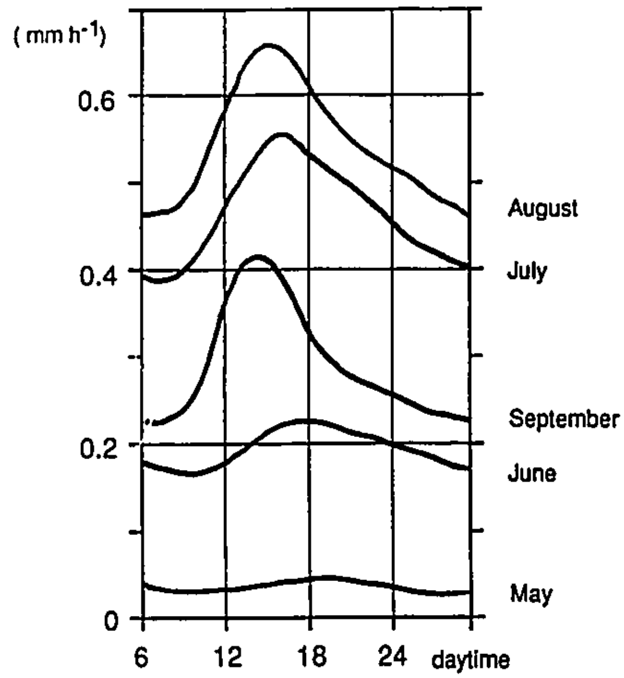
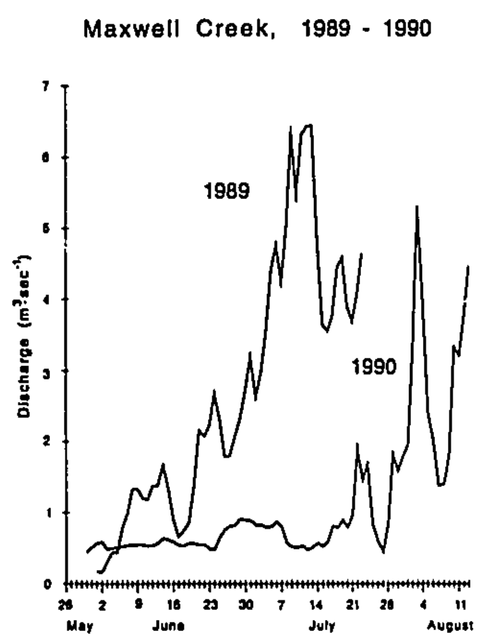


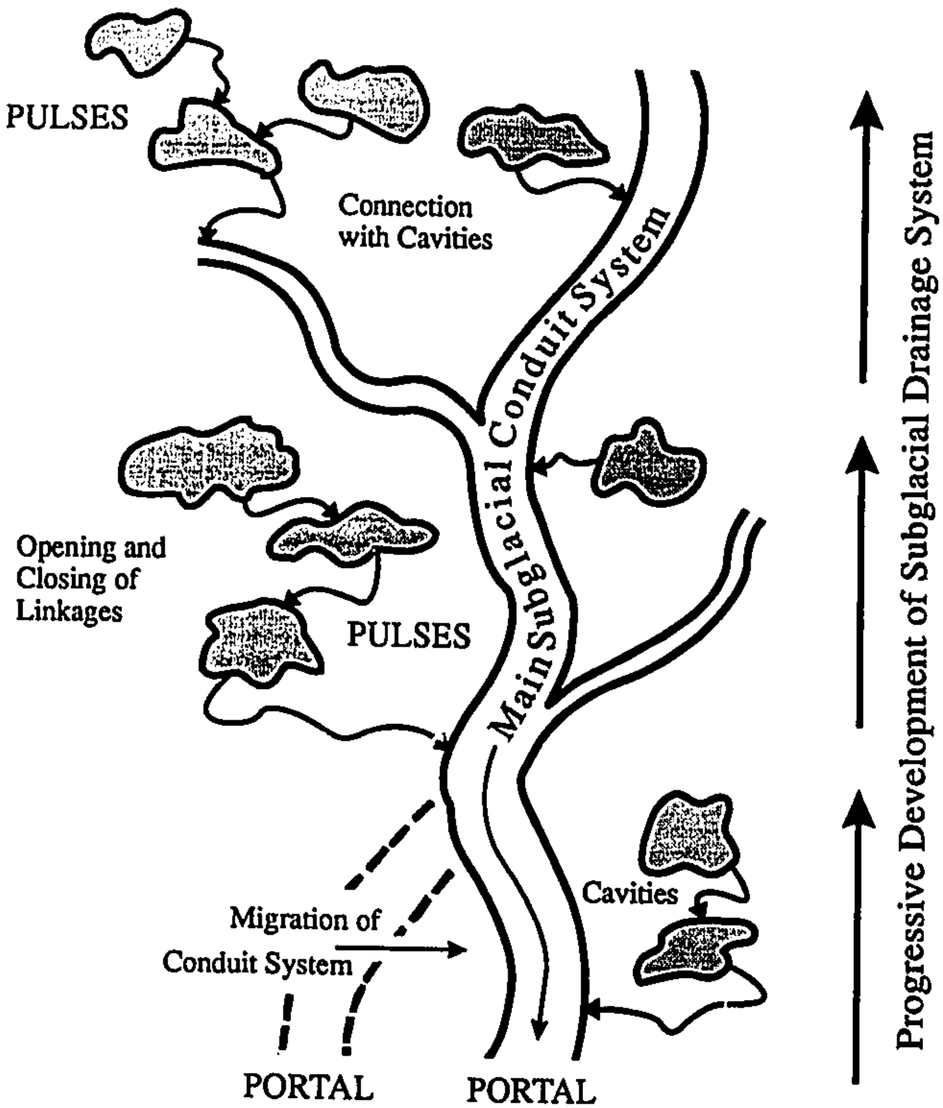
Figure 2.7: Aperiodic variations of runoff exemplified by mean daily flows of an alpine glacier, St. Elias Mountains (modified after Kruszynski and Johnson, 1993).



and slowly forces water to spread out over the bed. The spread of meltwater over the glacier bed suggests that the drainage network begins in spring as a diffuse network of linked cavities rather than a few large conduits (Collins, 1989). Discharge levels in meltwater streams are most subject to variations in early spring due to the changes in the drainage network (Johnson, 1991b). Johnson (1991a) interpreted pulses in discharge into four hydrograph patterns , similar to "extraordinary" events reported by Hooke (1989) and Röthlisberger and Lang (1987), as indicators of changes in the internal drainage system of glaciers. Increased water pressure within cavities during the spring also results in the expansion of cavities and enhancement of basal sliding. Expansion of the drainage system continues during the ablation season creating a linked-cavity conduit system. As flow levels increase in conduits that experienced closure during the winter they quickly enlarge to their approximate previous summers size. During the mid to the end of the ablation season previously isolated pockets of stored water may also become integrated into the drainage system and may produce peaks in the flow hydrograph. Figure 2.8 is a schematic diagram of the internal drainage system of glaciers as outlined by Kamb (1987) and modified by Johnson (1991a). The occurrence of discharge pulses throughout the ablation season are common occurrences in glacierised basins and can achieve magnitudes that may produce significant geomorphological changes in the basin (Johnson, 1991a) as a result of complex relationship between climatological and glaciological elements.

Other runoff phenomena are related to the processes of water flow through the glacier and at the bed of the glacier. Analysis of seasonal diurnal variations in hydrochemical characteristics of meltwaters provides information regarding the interaction of meltwater with sediments at the base of a glacier and the development of basal structures during an ablation season (Collins, 1979b). During passage through a linked-cavity conduit system (Kamb, 1987) meltwater acquires chemical and sedimentological characteristics indicative of source and routing. Changes in routing and water source result in temporal variations of

Figure 2.8: Model for the englacial/subglacial drainage system of an alpine glacier. (modified after Johnson, 1991a).

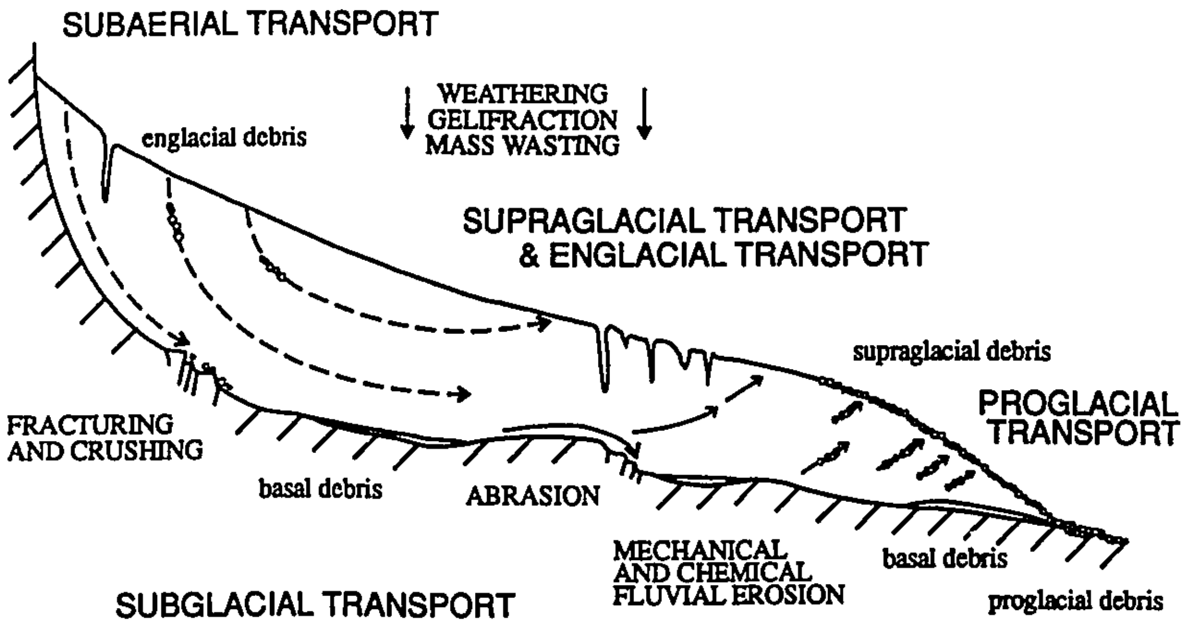


solute and sediment composition and concentration during ablation seasons. Peaks in discharge, suspended sediment concentration and total dissolved solutes are associated with the hydrological mechanisms at the base of an alpine glacier (Lemmens and Rogers, 1978; Collins, 1979a,b,c, 1989). Modelling the highly variable nature and function of the linked-cavities and conduits of the subglacial hydrological network remains subject to discussion (Collins, 1989).

2.3 SEDIMENT SOURCE, PRODUCTION AND TRANSPORT

Within glacierised basins sediment is continually produced by periglacial, fluvial and glacial erosion processes. Figure 2.9 schematically represents the glacierised basin sediment transfer systems. Essentially the alpine suspended sediment system is part of a complex sediment system that incorporates three hydrological sub-systems; the glacier, gravity deposits and glacial deposits (Johnson and Kruszynski, 1990). Throughout the year sediment is produced by glacial erosion processes across the glacier bed. During the ablation season when large volumes of water reach the glacier bed sediment is entrained in the flow and transported to the glacier portal. Sediment concentrations are low in the early spring increasing to high but variable in late spring and early summer, with lower flow during the late summer interrupted by aperiodic peaks (Collins, 1987; Kruszynski and Johnson, 1992). In spring, the flow of meltwater at the glacier bed results in the removal of sediment accumulated since the previous erosional period during the previous summer or earlier. During mid to late summer when flow conditions exhibit a diurnal rhythm, small increases of measured sediment concentration reflect material derived from ongoing glacial erosion at the glacier bed. Thus, the delivery of sediments to the glacier snout is limited temporally to the ablation season and spatially to those cavities and conduits comprising the

Figure 2.9: Sediment source, production and transport processes acting on and within an alpine glacier.



subglacial drainage system (Collins, 1989). The addition of sediment to the outflow regime from gravity deposits occurs primarily during the spring with snowmelt. The increased saturation of adjacent slopes results in an increase of sediment deposited on the glacier from subaerial and slope erosion mechanisms. Summers may also produce pulses in sediment when surface runoff occurs (Johnson and Kruszynski, 1990). Hence, the processes of glacial erosion, production of sediment and availability of water at the glacier bed are important when interpreting outflow records of glacier discharge and suspended sediment concentrations.

This following section examines the processes of sediment production and transfer within glacierised alpine basins. This section is not intended to be a complete review of details relating to particular erosive processes; such information has been presented in detailed individual research papers (e.g. Boulton, 1974; Hallet, 1979a) and in general texts (e.g. Embleton and Thornes, 1979; Sugden and John, 1984; and Drewry, 1987). Rather, this section will demonstrate the importance of understanding glacier erosion mechanisms when investigating the behaviour of a subglacial drainage system and its development.

2.3.1 Supra- and Englacial Sediment Supply and Transport

The glacial sediment transport system has three main paths: supraglacial, englacial and subglacial (Figure 2.9). Debris from either supraglacial or subglacial sources is incorporated into the system via sedimentary layering in the accumulation zone, open crevasses and by the transference from the glacier bed by thrust faulting. The transfer of sediments upwards from a debris-rich basal layer can occur in particular situations (Small, 1987); however, the greater proportion of the englacial sediment load is of supraglacial origin (Fenn, 1987).

The supraglacial sediment system is a passive system that accumulates material released from gravity deposits on adjacent valley slopes in the form of mud flows, rockfalls and debris flows. Debris accumulation on the glacier surface in the form of supraglacial lateral moraines and medial moraines undergoes relatively little modification in clast size and shape during its passive transport down glacier (Boulton, 1978). Although the supraglacial sediment system is passive; supraglacial debris along glacier margins, in the lee of nunataks and along the centreline provides much of the sediment that is incorporated into the englacial sediment system.

Supraglacial sediments are incorporated into englacial transport system via a variety of mechanisms: sedimentary layering, slippage into crevasses and the upward transference of material from the glacier bed (Small, 1987). Within the accumulation zone of alpine glaciers, debris material is preserved in the winter's snow accumulation and later incorporated into glacial debris layers and bands through the formation of firn. These debris layers or bands are transferred down glacier by flow to be later exposed nearer to the terminus.

Debris from mass movements along valley slopes accumulates and fills crevasses along glacier margins and near the headwall (Drewry, 1987). Over time these deposits are buried in snow, firn and in turn, glacier ice forming a deep debris layer within the glacier. Debris material that accumulates within the ablation zone, however, is exposed at the glacier surface either by closure of crevasses and expulsion of material or through lowering of the glacier surface by ablation. Near the terminus of alpine glaciers, debris material may be transferred upward from the debris-rich basal layer to the surface. Flowing ice near the glacier snout is compressed as it overrides slower or stagnant ice (Boulton, 1978). Thrust planes form at the boundary between the stagnant ice and the faster flowing ice carrying debris upward. With the exception of material thrust upward from the glacier bed, the original characteristics of the debris indicate only minor modification during englacial transport. Englacial debris layers and bands are not exposed to intense crushing, sliding and fluvial transport, such as debris material at the glacier bed. The englacial transport system, like the supraglacial system, is a passive mode of transport for debris material.

2.3.2 Subglacial Sediment Supply, Process and Transport

Glaciers move by creeping over and sliding across their beds (Paterson, 1983) continually producing sediment at the bed and along the margins. Sediments transferred through supraglacial, englacial and subglacial routes are produced through supraglacial or subglacial erosion; englacial sediment is the result of either incorporation from above or from below (Boulton, 1974; Small, 1987). The most commonly recognised processes of subglacial erosion are: friction cracking, fracture of bedrock, quarrying and plucking;

abrasion; and mechanical and chemical fluvial erosion. At any one site any combination of these processes may be active. However, processes operating on rock debris at the base of a glacier and on bedrock surfaces are difficult to observe directly. Although Walder and Hallet (1979) and Bogen (1989) have directly observed the magnitude of subglacial erosion processes, their observations are limited to specific times of year, underlying surface topography and geology. Although work by Hallet (1979a, 1981) and Boulton (1979) have made important contributions to the development of subglacial erosion models, the models are still primarily derived from field observations.

Abrasion is the process by which bedrock is scored and ground into small particles by contact with other rock particles held in the basal ice layer. There are three elements necessary for abrasion to occur at the glacier bed: basal debris, sliding of basal ice, and movement of debris towards bedrock. Hallet's (1979) widely accepted theoretical model of glacial abrasion suggests that the most critical factors affecting the efficiency of glacier abrasion are the flux of particles in contact with the bed, the force with which these particles are pressed against the bed and the relative hardness of the bed and particles. The rate of abrasion will also be intricately associated with the relative hardness of rock particles and bedrock, particle size and shape, debris concentration (Hallet, 1979), effective normal stress (Boulton, 1979), and the efficient removal of fine sediments by meltwater. In Hallet's (1981) paper he revised his theory on glacier abrasion by taking into account the significant impeding effect of debris concentration above a certain value that can reduce the abrasion rate by reducing the rate of sliding. Röthlisberger (1979) suggested that if the hardness of the clast is similar to the bedrock, that the clast would be able to erode some of the bedrock before becoming completely fragmented. Accumulation of debris at the bed would result in the sliding of the glacier over a bed of clasts sliding and rolling over one another. This multiplying effect is readily observed in proglacial streams where the majority of sediment is of smaller particles (e.g. glacier flour).

Plucking is usually the result of two important mechanisms: fracture of bedrock by overlying ice and the incorporation of fragments into the glacier by freezing-on (Drewry, 1987). Large blocks held in the ice and forced against smaller bedrock protuberances may cause the fracture of bedrock. It is the concentration of stress onto a small area, which permits fracturing of the bedrock and the formation of features such as chattermarks. The chattermarks may have an orientation up or down the direction of glacier movement largely controlled by the angle of contact between the pointed block and the fracturing surface. The process of increased pressure of overriding rock particles can in a way be regarded as the same process as abrasion, except that the rock fractures are larger and can follow joints in the bedrock. The contrast between abrasion and plucking is reflected in the high confining pressures, associated with abrasion, and low confining pressures associated with plucking. Theakstone (1967) and Bennet (1968) have suggested that the, seasonal freezing near the ice margin plays a significant role in glacial plucking, a process observed by Anderson *et al.* (1982) at the Grinnel Glacier. Anderson *et al.* (1982) observed that quarrying and plucking at the glacier base does not require a great ice thickness, rather the process of freezing-on (plucking) is closely associated with the availability of subglacial water. The removal of weakened and loosened rock is associated with rising basal water pressures in basal cavities; in such situations the high water pressures support more of the weight of the glacier and allows freezing in the vicinity where the ice is in contact with the bed (Röthlisberger, 1981). The increased basal slip associated with increased basal water pressures is an effective means of removing rock fragments that have fallen into the cavity. Considering the forces required to remove blocks by plucking, it is likely that high bedrock pore pressures will promote plucking and fracturing.

The flow of glacial meltwater also has important erosional effects beneath a glacier. At the smallest scale the water film or channel film beneath an alpine glacier allows both physical and chemical rock removal and alteration. Meltwater confined to conduits and/or channels transports debris lying on the bed and, because of the transport of such debris mainly in suspension, is able to 'carve' the glacier bed. The erosion of bedrock by the abrasion of water at the bed is determined by an assortment of elements which contribute to the rate of abrasion, included in the list are: clast size, hardness and concentration.

Cavitation of bedrock surfaces is another mechanism of subglacial fluvial erosion. Turbulent, well-aerated water flowes from areas of low pressure into areas of higher pressures generating molecular instabilities and pressure changes in flow (Theakstone, 1980). The rapid expansion and explosion of the air bubbles near to the bedrock surface causes physical erosion of channel surfaces. Walder and Hallet (1979) noted the presence of large-scale subglacial meltwater channels in a formally glaciated valley in Montana resulting from physical and chemical erosion. Shreve (1972) has shown how meltwater flowing at base of a glacier will tend to increase its velocity as it flows over a convexity in the underlying surface favouring the physical erosion of bedrock surfaces. Flowing meltwater at the ice-bedrock interface not only allows continued excavation of debris produced by subglacial erosion and continued physical erosion but also allows the continued chemical modification of the rock surface.

Solution, hydrolysis and carbonation are a few of the important mechanisms in subglacial chemical weathering. Chemical weathering of bedrock is the decomposition of minerals into their ionic constituents. Solution is when water acts as a solvent by dissociating and removing soluble ions from minerals. Hydrolysis is when water acts as a reactant not a solvent in a manner where hydrogen dissociated from water replaces mineral cations, leading to possible expansion and decomposition of mineral structures. The displaced mineral ions may combine with the free hydroxides to form salts. Solution and hydrolysis are most active where dilute meltwaters first encounters reactive rock debris; where fresh ground particles are abundant, where progressive solution is not inhibited by the formation of insoluble residues and where flushing and/or leaching of ionically saturated solutions enables subsequent chemical activity (Thomas and Raiswell, 1984). The production of the hydrogen ions required for acid hydrolysis may be dissociated in two steps and is derived from prolonged contact of meltwaters with atmospheric CO_2:

$$CO_2\,(g) + H_20\,(l) \Leftrightarrow CO_2\,(aq) \Leftrightarrow H_2CO_3\,(aq) \Leftrightarrow$$
$$H^+\,(aq) + HCO_3^-\,(aq) \Leftrightarrow 2H^+\,(aq) + CO_3^{-2}\,(aq).$$

The dissociation of the neutral compound (H_2CO_3) gives off the hydronium cation (H^+) and the bicarbonate anion (HCO_3^-). Carbonation is thus the most important chemical mechanism controlling the dissolution of limestones and feldspars and influences the chemical composition of glacial meltwaters. Carbonation is the reaction between a mineral and the carbonate and bicarbonate ions producing soluble products and insoluble residues contributing them to water by the absorption of atmospheric CO_2. The mineralogical composition of the bedrock is the primary control in the chemical weathering process. Walder and Hallet (1976) noted how calcium carbonate ($CaCO_3$) is derived during pressure solution on the stoss side of bumps and precipitated by refreezing. Hallet (1976) noted that meltwaters that did not refreeze carried the solute load out of the glacier system. This chemical erosion of bedrock was indirectly observed by Collins (1977) in his study of the electrical conductivity in glacial meltwaters. Collins determined that the ionic enrichment of proglacial meltwaters was of glacial source. Hence, when investigating subglacial erosion processes by sediment and water quality characteristics, it is important to consider both the physical and chemical weathering processes.

2.3.3 Glacio-Fluvial Sediment Delivery

Suspended sediment is that part of a river's load that is carried in suspension. It consists of the finest solid particles that can be supported by the water without recourse to saltation, bedload or dissolved load. Transportation of the particles occurs when the shear stress is sufficient to initiate movement and overcome particle inertia. Within glacierised basins the initial sediment load is primarily derived from glaciers and proglacial areas, particularly in unvegetated areas. Kelly (1990) has suggested that subglacial sediments are perhaps, the most important debris source compared with sediments from englacial and subglacial regions. He proposes that the sediment yield is a function of the physical parameters of the basin: area of the basin, percentage of glacierised area and total runoff, but does not consider the function of glacier dynamics or the structure of moraines (Johnson and Kruszynski, 1990). Johnson and Kruszynski (1990) have proposed however, that the suspended sediment within glacierised basins originates from three hydrological sub-systems; the glacier, gravity deposits and glacial deposits. Each component of the system contributes to the highly variable sediment regime of a glacierised basin. A glacial hydrological systems approach emphasises the interaction between glacier mechanics and the hydrological regimes of the non-glacierised components.

The suspended sediment regime of a glacierised basin depends on the relative contribution from each hydrological subsystem. The hydrological systems of the three components interact to produce distinct discharge and suspended regimes (Johnson and Kruszynski, 1990). The gravity deposit system of debris transferred from valley slopes provides input to the glacier and the glacial deposit systems as well as direct input into the stream. The glacier and glacial deposit systems also interact to produce fluctuations in the suspended sediment regime.

The glacier subsystem is the active ice component within the basin that actively produces and conveys sediment within the basin. Sediment is primarily produced and stored at the glacier bed, to a lesser extent within the ice, due to a variety of transfer processes from the surface or bed. The transport of sediment from the glacier bed to the glacier portal depends on the routing of water at the bed. During the early spring the majority of meltwater is of supraglacial origin and has low suspended sediment concentrations with little diurnal fluctuation. Supraglacial runoff from late spring snow melt does not produce an increase in suspended sediment delivery except where there is sediment available to the stream in the proglacial zone. During early summer supraglacial meltwater begins to open englacial and subglacial conduits allowing meltwater to reach the glacier bed, weathered sediments that have accumulated since the last ablation season and are easily entrained in the flow. The initial supply of meltwater to the glacier bed effectively flushes out accumulated sediment stores reducing the availability of debris for meltwater removal (Collins, 1987). The availability of debris and the quantity of debris available at the bed reflects the rate of erosion by the glacier and the period of time between consecutive removals. Hence, during the late spring and early summer, large accumulations of sediment are readily entrained into the flow (Hooke *et al.*, 1985). During the summer discharge and suspended sediment regimes become variable due to sudden re-routing of subglacial conduits introducing new sediment sources (Collins, 1979a). Channel migration and tapping of new sediment sources, unrelated to discharge, produces peaks in proglacial suspended sediment concentrations (Johnson, 1991a; Gurnell, 1982). Later in the summer, once the drainage system is fully established, the summer variability and high concentrations of suspended sediment decrease and a diurnal regime is established. However, unique events producing sharp peaks of up to 17,000 mgL^{-1} have been recorded with discharges ranging from $10m^3sec^{-1}$ to $250m^3sec^{-1}$ in the Kaskawulsh and Maxwell Creek Basins later in the summer (Johnson and Kruszynski, 1990). Thus, 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 discharge fluctuations (Collins, 1979a, 1986). Variations in suspended transport have been variously attributed to outbursts from glacially dammed lakes (Collins, 1987), severe storms (Johnson and Power, 1985), and reorganisation of subglacial drainage channels (Collins, 1989). Therefore, in high mountain areas with some glacierised component, both the glacier-dominated river regime and the potential non-glacial regime need to be investigated to determine the characteristics of the larger proglacial basin.

Sediment input from glacial and gravity deposits contribute high sediment loads periodically through the ablation season and complicates the modelling of erosion and transport processes in glacierised basins (Johnson and Kruszynski, 1990). The hydrological regimes of the glacial and gravity deposits are both dominated by snow melt and precipitation events that provide external inputs to the suspended sediment regime. Increased water content of adjacent valley slopes results in the occurrence of avalanches and slush flows during the spring. Mass movements such as slush flows, mud flows and debris flows introduce abundant fine sediment that adds significantly to suspended loads in the hydrological system (Johnson and Kruszynski, 1990). Hence, the specific timing and rate of snow melt and the spatial and temporal variations in rainfall produce, together with the glacier hydrologic system, a highly variable suspended sediment regime, with irregular peaks.

Suspended sediment concentrations in rivers and streams reflect the interaction between the fluvial system, which can both transport and erode deposits, and the glacial sediment producing system. The relationship between suspended sediment concentration and discharge has been the foci of current research into fluvioglacial sediment transfer (Binda *et al.*, 1985; Bogen, 1980; Collins, 1979a; Østrem, 1975) but these have frequently failed because simple linear relationships do not occur and the geometry of the glacio-fluvial system is continually evolving and changing throughout and between ablation seasons.

Østrem (1964, 1975) observed that in general, suspended sediment transport in glacial rivers follows a diurnal pattern similar to discharge variation after sustained ablation. Østrem (1975) noted that sediment transport in glacier-fed streams increases with increased flows and declines in lower flows. The application of linear regression analysis has been applied to describe the relationship between the varying concentrations of suspended sediment and discharge for different glacier-fed streams in order to produce rating curves (Østrem, 1975; Collins, 1979a). When employing a regression model the variable should be transformed to provide scatter in the observations indicating a linear relationship between suspended sediment concentration and discharge with residuals in the dependent variable (suspended sediment concentration) that are random, have zero mean, are normally distributed with constant variance and are serially uncorrelated (Gurnell and Fenn, 1984). However, Fenn *et al.* (1985) applied linear regression analysis to data from monitoring periods of varying duration and observed that regression analysis often resulted in the serial autocorrelation of residuals and that this autocorrelation biased the rating curve and any estimates of suspended sediment concentration derived from it. Walling (1977) showed how predicted concentrations are underestimates of reality due to estimations from log transformed data. Finally, Fenn *et al.* (1985) concluded that it is impossible to estimate an unbiased simple regression model relating suspended sediment to discharge for the glacier de Tsidjiore Nouve.

The application of rating curves for glacier-fed streams are due partly for the need to predict the timing of peak sediment concentration for industrialised applications. This is essential for efficient control of water resources and minimisation of equipment damage due to high suspended sediment concentrations (Fenn *et al,* 1985; Bezinge *et al.*, 1989; Fenn, 1989). Østrem (1975) noted however, that is was possible to estimate sediment rating curves that fitted the observations of suspended sediment concentration and discharge extremely well for some time periods, but that curves providing a good fit were unusual. Hammer and Smith (1983) found it necessary to develop rating curves for early and late season periods to produce reliable and useful curves. These observations show that for different time periods at the same site, rating curves varied substantially between and within years producing biased predictions.

Collins (1979a) established suspended sediment concentration - discharge rating curves for rising and falling hydrograph limbs in waters draining Gornergletscher to compensate for hysteresis. The hysteresis, a looped relationship between two variables, occurs on diurnal to annual cycles. Hammer and Smith (1983) determined it was necessary to calculate a regression equation for suspended sediment concentration and discharge depending on whether data were collected earlier or later in the ablation season to compensate for the hysteresis. The hysteresis effect is a result of the greater availability of transportable sediment on the rising versus falling limb of the hydrograph. Thus, the maximum concentration of sediment coincides with the maximum rates of increase in flow (Richards, 1984). Also, the exhaustion of accumulated sediments occurs before the maximum discharge producing a steeper falling limb than the rate of increase. Hence, although the supply of sediment to glacial meltwaters is unlimited, the transport of suspended sediment material exhibits exhaustion effects during periods when the current discharge level has recently been exceeded.

The relationship of suspended sediment transport and discharge is therefore not constant through time limiting the value of applied linear regression analysis. Proglacial streams are highly variable and characterised by greatly varying suspended sediment concentrations. Short-term fluctuations in sediment transport, unrelated to discharge, are common to proglacial streams. This variability is the major attribute of the system but is treated as anomalous (Johnson and Kruszynski, 1990). The release of water at the glacier bed may entrain new sediments, may have no access to sediment or may entrain new sediments without any change in discharge. External, proglacial inputs, with surface flow, near surface flow or groundwater percolation, generate mudflows and debris flows which also contribute to temporally irregular pulses. The interactions of the three hydrological sub-systems contribute to the character of the suspended sediment regime of the headwater basin making quantification rather more complex than can be characterised by regression analysis. For it accounts for neither accounts for the proglacial sediment sources nor the historical influences of the hydrograph apparent in sediment delivery.

Linear regression analysis applied to suspended sediment regimes in glacierised basins has proved to be inadequate in both long- and short-term modelling. Gurnell and Fenn (1984) have derived a Box-Jenkins transfer function to apply to data from the glacier de Tsidjiore Nouve, transforming the suspended sediment data logarithmically and introducing a one hour lag (Gurnell, 1987). This method of applying transfer functions, leads, lags and rate of change has provided some improvement in forecasting the sediment concentrations in proglacial streams. However, this method of time series modelling relies on long, continuous data sets that are not always available and are not always based on physical processes (Kelly, 1990). These intricate statistical relationships can cost predictive value since the type of modelling technique is susceptible to side-tracking into certain limiting aspects of methodology that do not improve overall predictions (Klemeš, 1982). Hence, the results of the application of statistical techniques have still to be related to the causal process in glacier hydrology and its integral components.

2.4 Meltwater Hydrochemistry

Meltwaters emerging from glacier termini acquire much of their compositional characteristics during their passage through the conduits, cavities, and tunnels that comprise the subglacial hydrological system (Collins, 1987). Meltwaters interacting with debris above the glacier, at the base of the glacier, and in front of the glacier produce hydrochemical variations in meltwater outflow from glacier termini (Collins, 1979b). These variations in water chemistry provide a means of estimating subglacial chemical weathering (Reynolds and Johnson, 1972) and allows the interpretation of the form, development and stability of subglacial drainage systems (Collins, 1979c, 1987, 1989).

The aim of this section is to consider the characteristics of the alpine glacier environment that may contribute to the hydrochemical characteristics of proglacial alpine waters. The discussion emphasizes the nature and controls on solute load in proglacial streams including the relationship between stream discharge and solute load. The final part considers the uses, limitations and contribution of electrical conductivity studies in glacier hydrochemistry and glacier hydrology.

2.4.1 Solute Content of Glacial Meltwaters

The chemical characteristics of meltwater issuing from the terminus of an alpine glacier are the result of the mountainous environment and the glacier drainage system. The chemical composition of natural waters is affected by the soluble products of weathering and erosion. Recently deglaciated alpine basins are useful sites for hydrochemical studies. Alpine glacier development usually occurs between 2000 and 4000m a.s.l., situated above the tree line with only scant vegetation on surrounding slopes and valley bottoms (Souchez and Lorrain, 1987). The lack of organic decomposition by soil micro-organisms or by root respiration reduces the contribution of solutes and restricts the contribution of CO_2 to the meltwaters reducing the amount of chemical alteration to proglacial meltwaters. Solutes present in glacial meltwaters are derived from three sources: precipitation, atmosphere and through the passage of waters in englacial and subglacial channels (Trudgil, 1986). Water may also be released in large quantities either because of snow and ice melting during the summer or, occasionally, by the release of stored water from englacial or subglacial cavities. Given a knowledge of underlying geology and the chemical processes at the glacier bed, the identification of individual ionic species can help to identify their sources and transport pathways within a glacierised basin.

To understand the mechanisms of chemical weathering by means of a study of solutes in meltwater, it is essential to delineate the different environments of solute origin. The greatest proportion of meltwater within a glacier hydrological system originates from snow or surface ice-melt. Hence, when determining the chemical characteristics of meltwater streams it is useful to consider the atmospheric input to the solute content of meltwaters. The atmospheric contribution consists of the ionic composition of the snow and rain that falls on the glacier surface and surrounding valley sides throughout the year. As indicated by the very dilute character of supraglacial streams, the atmospheric contribution to the meltwater system is minimal (Lemmens and Roger, 1978). Moreover, Thomas and Raiswell (1984) have also examined the anthropogenic contribution of dissolved sulphates in North-Western and central European proglacial streams. They observed that the anthropogenic contribution to the total sulphate was low and is likely to have little influence on the total solute load of meltwater streams. Various studies on alpine glaciers, Rainwater and Guy (1961), Behrens *et al.* (1971), Elliston (1973), Church (1974), Collins (1979a,b,c), and Thomas and Raiswell (1984) have suggested a two-component model of total flow; water passing through subglacial channels with prolonged contact between water and rock and meltwaters running off rapidly via englacial channels, without undergoing chemical change. The chemical enrichment of supraglacial surfaces is suggested to contribute typically less than that on subglacial surfaces (Lemmens and Rogers, 1978; Gurnell and Fenn, 1985). Church (1972) observed that high concentrations of dissolved material derived from the subglacial conduits is mainly because of intense glaciation of the catchments.

The process by which major ions are released into solution at the water-bed interface and distributed in the solution is called ion exchange (Carroll, 1959). Ion exchange appears to be an essential mechanism in the explanation of the dissolved cation content of meltwaters in the frontal zones of glaciers. This phenomena is favoured in this zone due to the presence of fresh material with a high abundance of clay sized particles. This material is scoured from basal layers of ice in the lower part of the ablation zone of the glacier and, for the proglacial area, of a recent glacier retreat, exposing morainic debris. Due to the very dilute character of supra- and englacial meltwater, contact with these particles results in rapid cation exchange and water enrichment (Lemmens and Roger, 1978). Negatively charged clay minerals, surrounded by positively charged cations, allow the major ions including Na, K, Ca, and Mg to be exchanged with the surrounding water. Experiments conducted by Lemmens and Roger (1978), on the rates of diffusion of different cations, demonstrate the rapidity of ion exchange in proglacial meltwaters. This rapid enrichment of dilute supraglacial waters in sub- and proglacial zones indicates the importance of sub- and proglacial ion exchange in glacial meltwaters (Souchez, *et al.*, 1978).

2.4.2 Subglacial Solute Acquisition

Chemical weathering processes, either under or in front of a glacier, are selectively controlled by the amount of contact between materials and where there is maximum availability of mobilised ions. Within the alpine glacier environment, physical erosion by glacier abrasion is rapid and constantly exposing fresh mineral grains to flowing water so that the initial alteration of mineral surfaces has a significant influence on water chemistry (Slatt, 1972; Souchez *et al.*, 1973, 1978). Bezinge *et al.* (1973) deduced that glacial meltwaters flowing through subglacial channels become chemically enriched by interacting with basal morainic material. The concentration of solutes carried in proglacial meltwaters acquired during flow through the linked-cavities, conduits and channels of the subglacial drainage system, is controlled by carbonate equilibria dependent on the availability of hydrogen ions supplied from dissolution of atmospheric CO_2 and oxidation of sulphides (Raiswell, 1984). Chemical models for the evolution of bulk meltwaters (Raiswell, 1984) have been constructed on the assumption that the hydrogen ions used in weathering rock minerals are mainly derived from a reaction involving CO_2 known as carbonation (Raiswell and Thomas, 1984a). The hydrogen ion is responsible for the alteration of mineral lattices by acid hydrolysis (Raiswell and Thomas, 1984). The process of carbonation is thus the most important chemical mechanism, controlling the composition of meltwaters (Thomas and Raiswell, 1984; Souchez and Lorrain, 1987). The production of the hydrogen ions required for acid hydrolysis may be dissociated in two steps and is derived from prolonged contact of meltwaters with atmospheric CO_2:

$$CO_2\,(g) + H_2O\,(l) \Leftrightarrow CO_2\,(aq) \Leftrightarrow H_2CO_3\,(aq) \Leftrightarrow$$
$$H^+\,(aq) + HCO_3^-\,(aq) \Leftrightarrow 2H^+\,(aq) + CO_3^{-2}\,(aq).$$

A second mechanism producing hydrogen ions responsible for weathering is the oxidation of sulphides (commonly pyrites). The reaction can be written in the following way:

$$4FeS_2 + 15O_2 + 8H_2O \Rightarrow 2FeS_2O_3 + 16H + 8SO_4^{2-}$$

Substantial quantities of sulphates in glacial meltstreams may indicate that this reaction is going on in the basin. The hydrogen ions produced can react with the carbonates, silicates and aluminosilicates in the following forms:

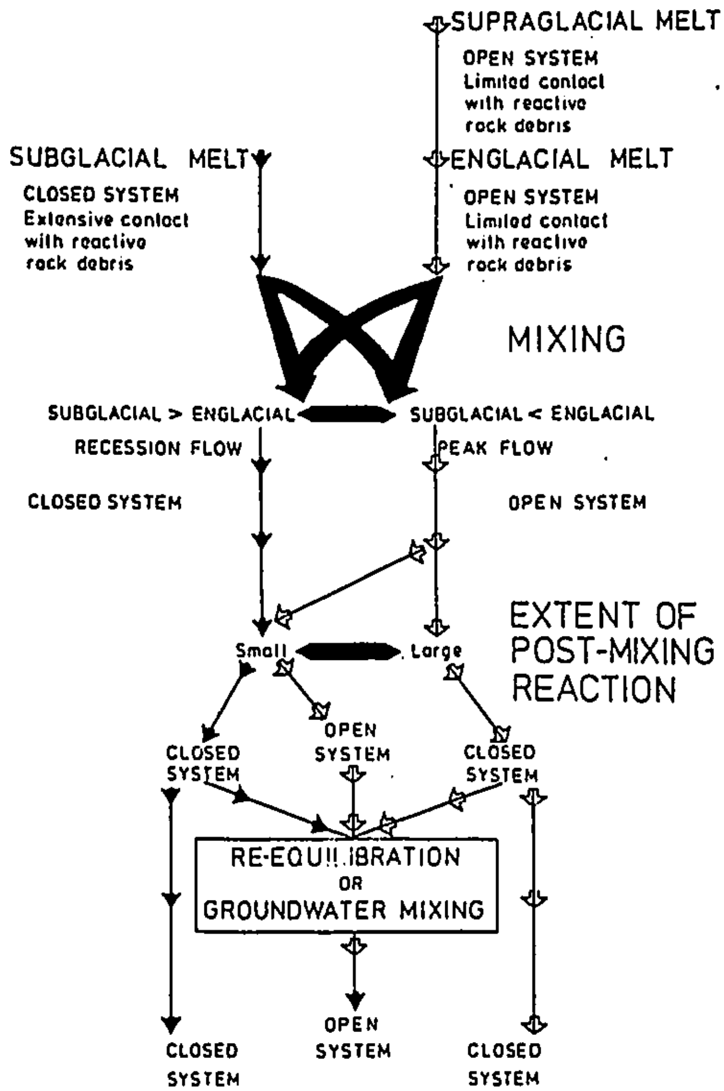
a) for carbonates:
$CaCO_3\,(s) + H^+ \rightarrow Ca^{2+}\,(aq) + HCO_3^-$
b) for silicates:
$Mg_2SiO_4 + 4H^+ \rightarrow 2Mg^{2+} + H_4Si_4$
c) for aluminosilicates:
$2NaAlSi_3O_8 + 2H^+ + 9H_2O \rightarrow 2Na^+ + 4H_4SiO_4 + Al_2Si_2O_5(OH)_4$

The dissolution of carbonates is more rapid than silicate or aluminosilicate breakdown and explains why Ca^{2+} is generally the most abundant cation in glacial meltwaters (Slatt, 1972). On the other hand, feldspars containing calcium are also among the most aluminosilicates.

This explains why calcium is in general dominant cation in glacial meltstreams even if the basin contains very small amounts of carbonate rocks.

Models for the evolution of meltwaters have been described as open and closed system models by Raiswell (1984) (Figure 2.10). The closed system is defined by the requirement that there is no dissolution of gaseous CO_2 to replace the dissolved CO_2 used in weathering. By contrast, open systems remain in equilibrium with atmospheric CO_2 and oxidation of sulphide minerals. In this situation, the rate of H^+ supply dependent upon CO_2 dissolution, approaches that of consumption by weathering and the supraglacial meltwater should maintain open system characteristics. This occurs with poorly reactive minerals and when rates of open system characterised waters, which will flush away products, are sufficiently high. Short residence times of surface meltwaters quickly flowing through the englacial system result in the limited capacity for mineral-water reactions to occur ensuring that these waters maintain open-system characteristics (Thomas and Raiswell, 1984). By contrast the subglacial environment, closed system, is the opposite of both the supra- and englacial environments. Within the subglacial environment the residence time of water may be several days since subglacial recession flows can be maintained in the absence of ablation and after the drainage of englacial reservoirs (Collins, 1977, 1979b). Solute efficacy would appear to be favoured in subglacial cavities and channels where there is intimate solute-solvent contact, adequate residence time and continuous production of freshly ground mineral grains. However, if there is inadequate replenishment of meltwater, the concentration of dissolved hydronium ions (H^+) will limit the solute content until the supply of atmospheric CO_2 recharges the concentration of H^+. Thus, rapid weathering occurs within the subglacial environment chemically enriching meltwaters to a greater extent than water flowing through englacial passages.

Figure 2.10: Schematic diagram showing possible variations of mixing behaviour in meltwater systems (after Raiswell, 1984).

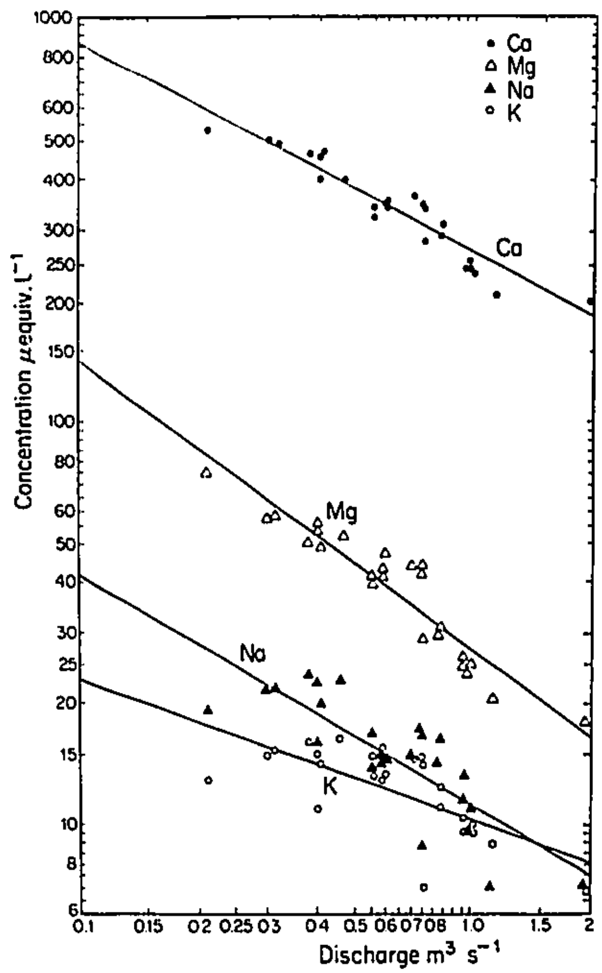


Ion exchange may also occur at the glacier sole where particle layers of mud act as semi-permeable membranes enhancing solute concentrations (Souchez and Lorrain, 1975). Interstitial water above the mud layer is forced through the mud under a pressure gradient and then refreezes below the layer in a cavity. The mud layer is frozen close to the melting point, so that a thin water film will be present at the base of a temperate glacier with high solute concentrations. Over saturation of the chemically enriched pockets or cavities may result in the precipitation of calcite or silica in the lee of bedrock protuberance (Hallet, 1976). Such basal precipitates will eventually become linked with basal meltwater flow producing variation in meltwater hydrochemistry.

2.4.3 Variations in Solute Concentrations

The relationships between discharge and solute load in glacial meltwaters are sometimes defined by a straight line on a bilogarithmic graph, as illustrated in figure 2.11 (Lemmens and Rogers, 1978). The relationship between the major cations exhibits an inverse relationship of the type $C=kQ^{-1}$, where C is the concentration of solute, Q is the discharge, k a constant and n an exponent having, for example, a value of 0.6 in the Chamberlain glacier area, Alaska (Rainwater and Guy, 1961). The inverse relationship can be explained by mixing of waters from two different sources. Collins (1977, 1979c) recognised two main components of glacial discharge: 1) a dilute surface meltwater that rapidly flows with little chemical alteration through englacial channels and 2) solute-rich waters flowing more slowly through the subglacial system. Lemmens and Rogers (1978) proposed that during periods of higher discharge the hydraulic radius of the subglacial channels increases, reducing the residence time and surface contact between water and sediments, resulting in a lower concentration of solutes in meltwaters. Lemmens and

Figure 2.11: Relationship between the concentration in the four major cations and discharge for the Glacier de Tsidjiore Nouve meltstream (after Lemmens and Rogers, 1978).



Rogers (1978) also suggested that the during low flow conditions the water and sediment had greater residence time and approached chemical equilibria, resulting in higher solute concentrations. The different slopes of the concentration-discharge regression lines for the four major cations are controlled by the different rates of diffusion for the different cations. Lemmens and Rogers (1978) experiment on diffusion rates support their results from the glacier de Tsidjiore, where the slope sequence of Na, K and Mg corresponds exactly to that of the experimental diffusion coefficients. Supporting the hypothesis that the character of glacier ice meltwater is defined by the rapid cation exchange and chemical enrichment in the subglacial frontal zone and proglacial zone of alpine glaciers.

However, the study of variations in ionic species is expensive and requires a time consuming sampling statergy if a representative record is to be obtained. Consequently, electrical conductivity measurements have been used as an economical and representative indicator of total dissolved solids.

2.4.4 Application of Electrical Conductivity

Electrical conductivity (EC) measurements used within the study of glacial hydrochemistry aid in distinguishing meltwaters according to their sources and pathways through the glacier hydrological system (Collins, 1977, 1979c; Thomas and Raiswell, 1984). Nakamura (1971) and Behrens *et al.* (1971) first discussed and established that EC is a useful parameter for runoff analysis in glacial river courses. Glacial meltwaters are enriched with solutes during their passage from the supraglacial through to the proglacial system. This solute enrichment is explained by ion exchange and/or direct solution of solutes and their transport, whether in fluid or attached to sediment, through the subglacial drainage network (Rainwater and Guy, 1961; Lemmens and Rogers, 1978; Collins, 1977, 1981).

The use of EC in glacier hydrology is essentially based on two principles: 1) its ability to be used as a surrogate measure of total dissolved solids and 2) the ease, portability and economy in which it may be measured (Fenn, 1987). Fenn's (1987) exhaustive review of five major papers discussing the use and limitations of electrical conductivity (EC) as a parameter of water quality for natural waters, soil solutions (e.g. Hem, 1970; Tanji and Biggar, 1972; Gloterman *et al.*, 1978; Foster *et al.*, 1981) and glacial meltwaters (e.g. Behrens *et al.*, 1971; Collins, 1977) provides a thorough summary of the important qualifications and limitations of EC studies. Fenn (1987) emphasizes that "the ease of measurement is a necessary but not a sufficient condition for the suitability of monitoring EC as a meltwater quality parameter: its electrochemical bases must also be acceptable". In spite of the bases and limitations of EC as water quality parameter, Fenn (1987) invariably recommends EC a useful as a surrogate measure for total solute concentrations in glacierised basins.

The source and pathways of meltwaters can be determined by examining their hydrochemical characteristics (Collins, 1979b, 1981). Variations in isotopic and ionic characteristics of snowmelt, icemelt and ground waters are used to distinguish their different sources (e.g. Behrens *et al.*, 1971; Zeman and Slaymaker, 1975; Eyles *et al.*, 1982). Route specific discriminations have been based on the differing electrical conductivity of quick-flow and delayed flow meltwaters (e.g. Collins, 1977, 1979c; Oerter *et al.*, 1981; Gurnell and Fenn, 1984). Collins (1977, 1979b) separation of meltwaters on the basis of EC permitted the identification of different drainage pathways within a glacier. He showed that a simple mixing model of the form:

$$Q_s = [(\mathrm{EC_t} - \mathrm{EC_e}) / (\mathrm{EC_s} - \mathrm{EC_e})]\, Q_t$$

where Q and EC are defined as before and the subscripts t, e and s represent total, englacial and subglacial components, respectively; may be used to separate englacially routed waters from subglacially routed waters. Collins (1979b) suggested that the Gornera glacier stream may be fed by waters derived from these hydrochemically different routes in which subglacial waters are chemically enriched and englacial waters are diluted. Collins (1979b) observed that englacial component of discharge flowed quickly through the glacier contributing between 50-80 per cent of total discharge and was in phase with the characteristic discharge hydrograph. Moreover, the slow subglacial component of the hydrograph may either be in or out of phase with the englacial contribution depending on the form and development of the subglacial drainage system. Collins (1979b) model for the Findelengletscher, in contrast, involves widespread interconnection between Q_c and Q_s pathways, with diurnal water pressure fluctuations in-phase. Collins (1981) found that the bulk of the annual solute load is discharged during the summer. During this period the high meltwater discharge efficiently reworks chemical components from subglacial locations. For the winter period low discharge is responsible for only a minor flux of solutes despite higher conductivities. Iken and Bindschadler (1986) have obtained similar results using different methods supporting Collins (1979b) mixing model approach.

The access to previously hydrologically isolated water at the glacier bed has been considered by Collins (1981) to account for the variations in meltwater chemistry. This stored water found in cavities or within basal sediments where prolonged contact with debris may give rise to considerable equilibrium enrichment. Therefore, EC-based flow separation studies enable characteristics flow routing to be identified for a given glacier and compared with those of a different glacier and interpreted.

CHAPTER 3

STUDY AREA AND SITE

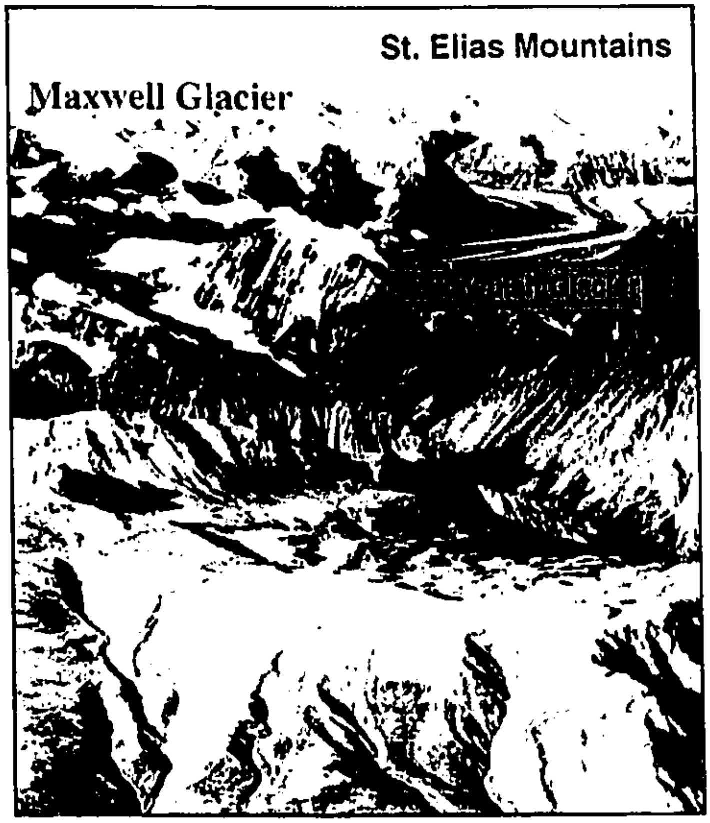
3.1 REGIONAL CONTEXT

3.1.1 Preamble

The St. Elias Mountains, located within the Canadian Cordillera at the edge of the American lithospheric plate, are the westernmost of the major physiographic and geological regions of Canada (Clague, 1989). This region is of diverse topography, geology, climate, and vegetation. The general form of the Cordillera resembles that of a great wall flanking the Interior Plains and consisting of an elevated platform rimmed by mountain battlements (Bostock, 1948). The western battlement consisting of high mountains, valley glaciers, and ice caps of the St. Elias Mountains (Figure 3.1).

The St. Elias Mountains border the Gulf of Alaska astride the international boundary between the United States and Canada. They contain various mountain ranges including the Icefield Range, Centennial Range, and Kluane Range. The highest peak in the Canadian Cordillera, Mt. Logan, has recently been recorded at 5959 •3m. This massif ranges amongst the highest peaks of North America. Since the St. Elias Mountains rise from the Pacific Ocean and transect the extreme south-west Yukon Territory, Wahrhaftig (1965) qualified them as "the most spectacular mountains of North America".

Figure 3.1: Oblique photograph of the St. Elias Mountains, including study area.



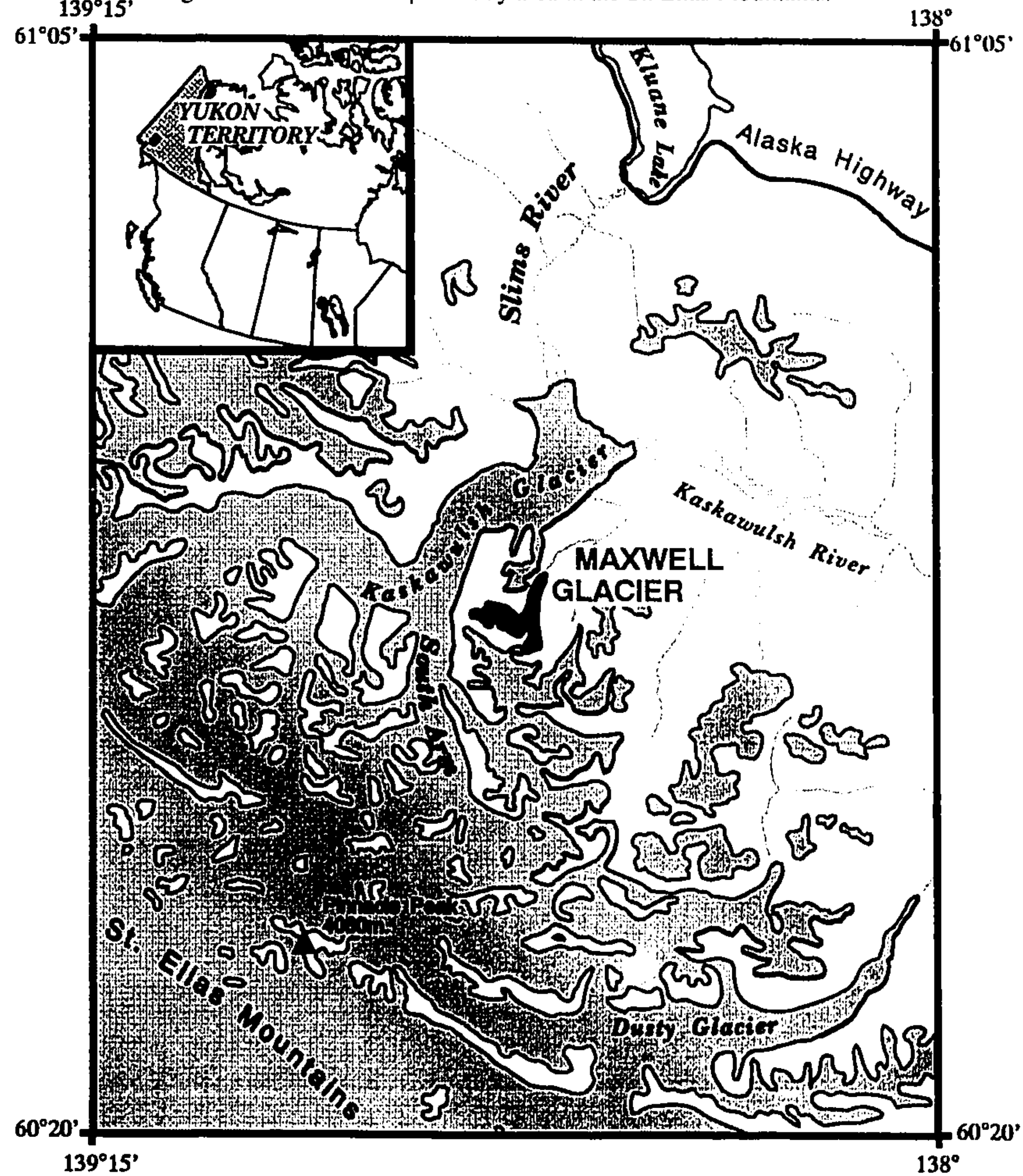
3.1.2 Boundaries of the St. Elias Mountains

Various interpretations exist for the boundaries between these mountain groups due to the diverse assemblage of those concerned with such aspects as geologic structure, hydrologic basins, and common or historical usage. Generally speaking, the St. Elias Mountains are bound by longitudes 138° - 143° west and 60°30'- 62° north (Figure 3.2); including both the main range, known as the Icefield Range, and various subsections, such as the Fairweather Range, and neighbouring mountain groups (Bostock, 1948, 1952). The axis of these mountains, as defined above, is about 750km from the south-east to north-west, with a maximum width of the glacierised area of about 180km measured in south-west-north-east direction. These mountains are an effective orographic barrier to moist maritime air into the interior Yukon.

3.1.3 Climatic Conditions

The St. Elias Mountain Ranges have a complex high alpine climate with a strong maritime influence from the west and a strong continental influence from the east. During winter, the Aleutian Low migrates into the Gulf of Alaska bringing heavy precipitation and clouds to the coastal mountains. Along the continental interior, the cold and dry Mackenzie High influences the frontal ranges of the St. Elias Mountains (Kendrew and Kerr, 1955). The summer patterns are particularly influenced by the Northern cell in the spring bringing major storms and precipitation. Later in the summer the Pacific High brings warm, dry weather to the region (Taylor-Barge, 1969; and Marcus, 1974). Thus, the extreme range in mountain altitudes is a very effective barrier to the intrusion of moist maritime air; resulting

Figure 3.2: Location map of study area in the St. Elias Mountains.

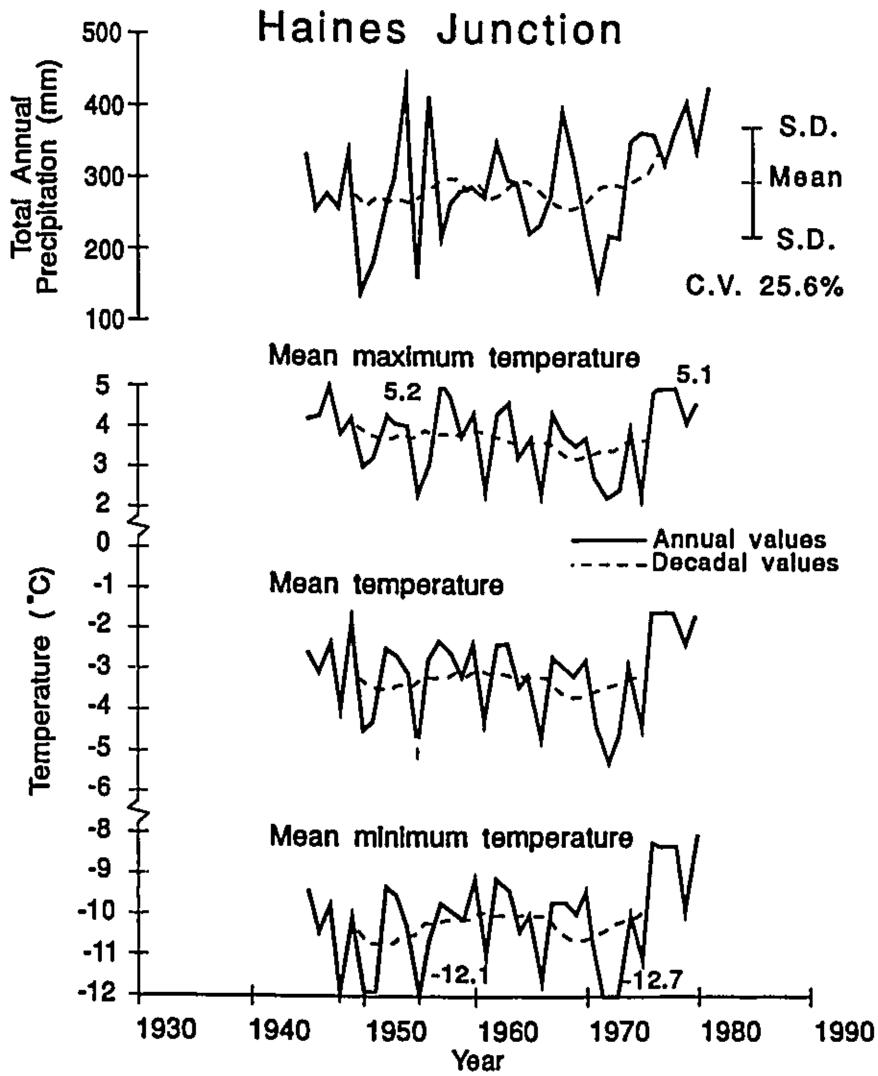


in a strong transition from a wet maritime climate on the coast to a dry continental climate north-east of the barrier (Wahl *et al.*, 1987).

The topographic divide of the St. Elias Mountains, approximately 160km inland from the Gulf of Alaska, acts as climatic divide separating maritime and continental climates (Taylor-Barge, 1969). Within the Icefield Range the mean annual temperature above 2500m. a.s.l. level drops to the $^-10$°C to $^-15$°C range; whereas the mean annual temperature changes from 0°C on the coast to $^-5$°C on the north-east side of the divide (Marcus, 1974). Annual total precipitation varies greatly from over 4000mm on the Gulf of Alaska coast to under 300mm in the Shakwak Trench. Rainfall occurs commonly at the lower elevations during mid-May through to September with major snow accumulations occurring between 1500-3000m. a.s.l. annually (Wahl *et al.*, 1987). The abundant snowfall at high altitudes is primarily the reason for the extensive network of glaciers in the Icefield Range. Thus, effects of latitude, topography, elevation, and atmospheric motions have resulted in the formation of the largest non-polar ice field in the world.

The Yukon climate trends during the last two to three decades have been presented by Wahl *et al.* (1987). It is apparent from the data for Haines Junction and Burwash Airport, the closest stations to the St. Elias Mountains, that there have been considerable variations and no discernible trends in the data (Figure 3.3, from Wahl *et al.*, 1987). Wahl *et al.* (1987) calculated a high coefficient of variation, 25.6 and 19.2%, for the year-to-year total annual precipitation variation and ten year running means for Haines Junction and Burwash respectively. During July 1990, rainfall measured at Haines Junction and Burwash Airport was 263% and 209% of the normal values. These figures correspond to 32% and 42% respectively of the average yearly totals for these stations. These marked annual variations elucidate the contrasts between years in the hydrological responses.

Figure 3.3: Yukon Territory Climate Trends (after Wahl *et al.*, 1987).



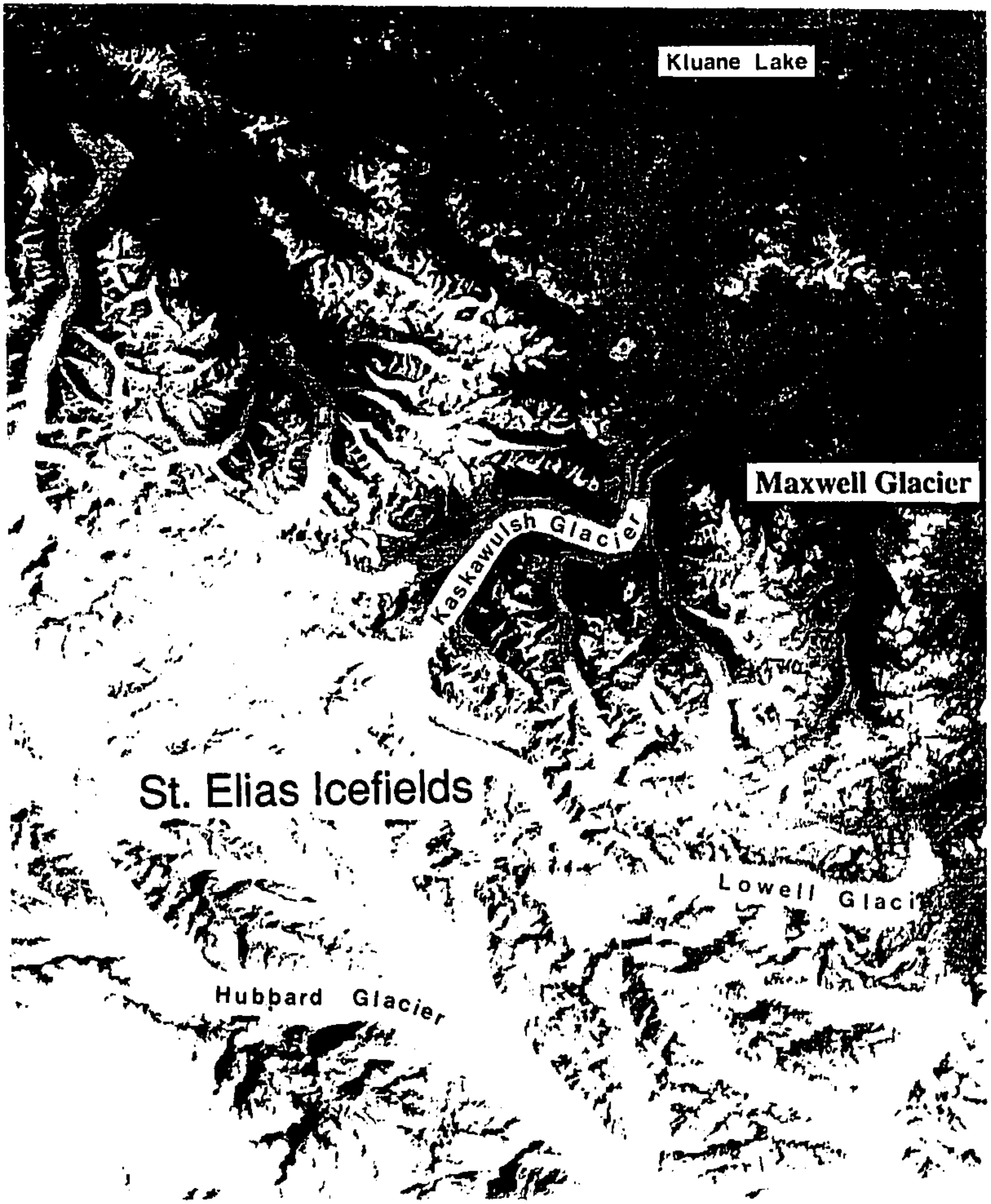
3.1.4 Extent of Glaciation

The severe alpine climate maintains the St. Elias Icefield and some of the largest valley glaciers outside the polar region. Within the central area of the St. Elias Icefield a plateau-like area at 2100-2750m. a.s.l. forms the divide between drainage to the Pacific Ocean to the south-west and to the Yukon River and Bering Sea to the north (Figure 3.4). "From these areas, the ice fields slope outward, gently at first and then more steeply as they separate into defined valley glaciers creeping out of the ranges" (Bostock, 1948). These glaciers are several kilometres wide, tens of kilometres long, and up to 1000m thick ending in the semi-arid Eastern Icefield ranges and the western massive piedmont and coastal glaciers of the Gulf of Alaska coast.

Repeated glaciations have occurred during the past several million years in the St. Elias Mountains. Denton and Armstrong (1969), describe evidence for 12 separate glacial advances at upper White River, from 10 to 2.7 million years ago. In recent geologic time three glaciations have occurred (Denton and Stuiver, 1967), transforming the landscape by continued glacial and fluvioglacial activity. The earliest of these ended more than 46, 400 years ago and the second glaciation ended by 30,000 years ago. The most recent major glaciation, termed the 'Macauley-Kluane' by Denton (1974), began 29,600 and ended by 9,400 years ago. Denton and Stuiver (1967) indicate that a period of climatic warming, Hypsithermal climatic period (Flint, 1971), and glacial recession followed the Macauley-Kluane episode in the south-west Yukon and lasted from about 9,800 to 1,280 years ago. Today, much of the area consists of oversteepened slopes produced by glacier erosion, stream erosion and mass wasting. Late Pleistocene glacier erosion and meltwater activity has produced valleys of glacially scoured bedrock and outwash plains (Rampton, 1981). Active valley sandurs with braided channels are found in valley bottoms surrounded by ice-cored moraines, talus fans and rock glaciers.

Figure 3.4: Extent of Glaciation within the St. Elias Mountains, Yukon Territory.

A Landsat MSS image of the St. Elias Mountains region with the study area (as in Figure 3.2). This imagery was gathered on 3 September 1977 using Bands 1,2,4. Scale is approximately 1:625000.



3.1.5 Discharge Variability within the St. Elias Mountains

The topography of the St. Elias Mountains not only affects the climate of the region but also produces variable local climate. The linkage between climate and topography is important for the glaciology and glacial geomorphology in the St. Elias Mountains. Determinate orographic controls to variable solar radiation intensities on differing slope angles and aspects, account for the local climates within the St. Elias Mountains which vary from one valley to the next. The variability within the St. Elias Mountain Range is a product of minor climatic variations in individual sub-basins that respond to larger changes in climate on different temporal scales and in different forms. From valley to valley, different responses to changing glaciological, meteorological and climatological conditions occur. Investigations of runoff in south-west Yukon glacierised basins have demonstrated extreme variability (a) at the glacier terminus, (b) in the proglacial area, and (c) in the non-glacierised component of basins (Kruszynski and Johnson, 1993). The existence of glaciers within low to intermediate sized mountain basins compensates for changes in the annual input of runoff to the larger basin, reducing annual coefficients of variations but promoting seasonal variability within the hydrological regimes (Fountain and Tangborn, 1985). The hydrological regimes of the Slims River in 1983, an ice-melt dominated year, and 1988 a summer dominated by rain on the non-glacierised component of the basin demonstrate extreme variability. This variability is a product of glaciological conditions, snow-cover, progression and rate of spring snowmelt, and occurrence and timing of summer storms (Kruszynski and Johnson, 1993). Discharge volumes and regimes, rate and timing of melt, rain on snow or ice, rain on saturated or unsaturated basin materials and the effect of glaciological conditions are complex within the St. Elias Range. Natural variability in the St. Elias Mountain basins is of a magnitude that precludes isolation of small scale changes caused by climate change. The Maxwell and Kaskawulsh glacierised drainage basins are both minor components of a larger system that needs to be understood. Each component is an integral part of the dynamic changing alpine hydrologic system. The variability exhibited within sub-basins of a larger basin illustrate that glaciological and hydrological responses to climate are individual. Such complexity of a system requires the continued and expanded research into the function of the basin as well as the products of such basins to isolate long-term climate change induced trends in runoff variability.

3.2 FIELD DATA COLLECTION SITE

3.2.1 Situation: Maxwell Glacier

The study site is a tributary basin of Maxwell Creek, St. Elias Mountains at longitude 138°40' west and latitude 60°42' north (Figure 3.2). The basin has an area of 22.3km^2 of which 7.4km^2 (33%) is glacierised and has an elevation range from 1740m. a.s.l. extending to 2760m. a.s.l. The glacier terminus is approximately 1780m. a.s.l. and rises to an elevation of 2680m a.s.l (Figure 3.5). The glacier trends south-west to north-east with an approximate length of 6km and an average width of 1km. The glacier is composed of two ice lobes originating in two cirques at the head of the valley. The average gradient of the glacier is less than 21%. The glacier is divided into areas separated by an icefall on the south side of the glacier and steepening ice surface to the north at approximately 2360m. The upper tier is mostly composed of the accumulation area beginning approximately at 2480m. Above the ice-fall is a broad, relatively flat, expanse of ice abruptly terminating at the headwall.

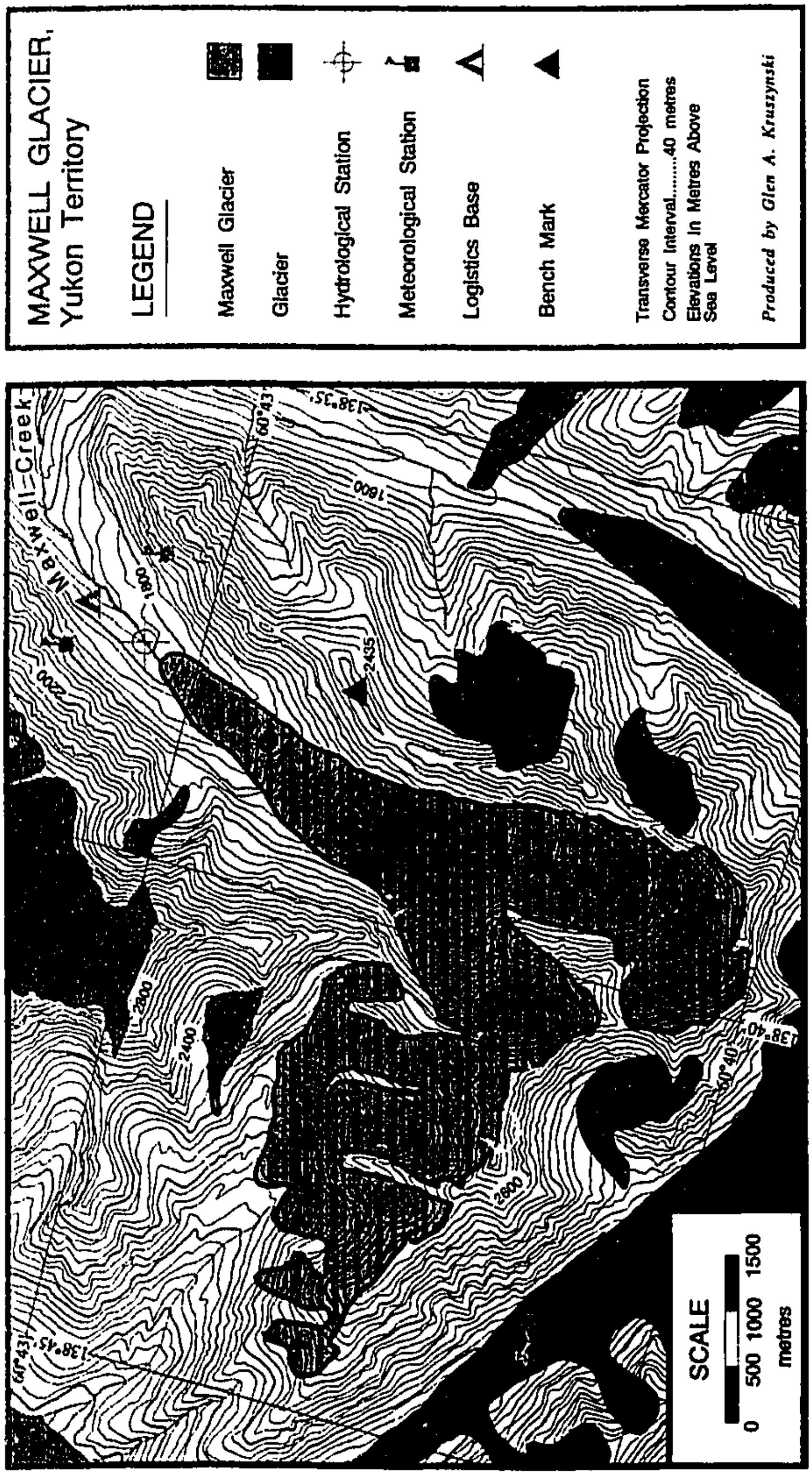


Figure 3.5: Location of Maxwell Creek, St. Elias Mountains, Yukon Territory.

Although some surface drainage on the east side of the glacier tongue persists as a marginal stream, the glacier is primarily drained by one subglacial outlet. The glacier portal is a broad and low conduit on the north-west side of the glacier terminus. Small streams off the glacial deposits and talus augment discharge and sediment load through the non-glacierised component of the basin. Adjacent slopes consist of extensive ice-cored moraines partly maintained by rock avalanche material from valley sides.

3.2.2 Local Climate and Vegetation

The topography of the St. Elias Mountains not only affects the climate of the region but also produces variable local climate. It is the linkage between climate and topography that is important for the glaciology and glacial geomorphology in the St. Elias Mountains. In response to different orographic controls and variable solar radiation intensities on differing slope angles and aspects, the local climates within the St. Elias Mountains vary from one valley to the next.

The climate of the study area is classified as highland sub-arctic. Total annual precipitation is estimated as 800mm with 200mm from the summer and 600mm from the winter (Gray, 1985). Temperature is moderate and variable, with a mean daily temperature of 8°C in July and August.

Winds in the area are strongly influenced by the topography, elevation and thermal contrasts due to the glacier and immediate ice fields. 'Glacier' and katabatic winds prevail throughout the year; but these winds are best developed in summer when greatest contrast in surface cover exists. 'Glacier winds' are predominantly down glacier from the south-west. Wind speeds vary throughout the day, reaching maximum velocities during summer afternoons when the ice-air temperature difference is greatest (Gray, 1985).

Glaciology, climate, geology and geomorphology form an intricate environment suitable for the growth of various plants. The immediate glacierised area is subject to glacial advance and retreat, glacio-fluvial sedimentation and erosion hindering the growth and succession of vegetation. Some common flora observed in the alpine environment of Maxwell Glacier are *Epilobium latifolium, Salix arctica, Phippsia algida, Ranunculus pygmaeus, Saxifragaceae* sp., and the moss heath *Cassiope stelleriana.*

3.2.3 Geomorphology and Geology

Maxwell Creek is an area of active glaciation with numerous glacial land-forms present. Ice cored moraines, braided channels, talus slopes, and valley sandur are few features located within the proglacial zone adjacent to the active ice margin (Figure 3.6). Maxwell Glacier is characterised with features associated with the an active ice margin; including, supraglacial streams, crevasses, moulins, and medial and lateral moraines. Krigström (1962) described how the valley sandur is characterised by having 'only one main water channel, which, however, branches out and creates a large network of smaller, secondary channels does not fill up the whole valley continuously' (Krigström 1962).

The valley slopes and glacierised head of the valley provide sediment and discharge to the valley sandur. Within the valley, the river flows along a constricted flood plain that is subject to many inputs of water and sediment from many different sources. The valley sides are steep bare rock faces with avalanche chutes, gullies, debris flow channels, talus deposits and channel gravel in the valley floor. The mass movements from valley sides have a marked effect on the form of the valley bottom and the route or pattern of the creek in the form of blockages or diversions that cause changes in the local base level for the fluvial system.

Figure 3.6: Geomorphic elements in the vicinity of Maxwell Glacier, St. Elias Mountains, Yukon Territory.

Palaeozoic Alexander Terrane composes the bedrock of the area (Figure 3.7). It is comprised of late Precambrian to Triassic strata including schist and gneiss derived from feldspathic sediments, felsic to mafic volcanic rocks, pelite and carbonate rocks. A detailed description and delineation of lithology and structure the area is provided in Dodds (1982a,b,c). The bedrock is exposed in higher elevations, along ridges, cliffs and peaks and in the valley floor. Three geological zones are identified: 1) Psa; 2) JKgd; 3) UPp' (Dodds, 1982a,b,c). Psa is of the "Lower Icefield Ranges Clastic Assemblage" composed of Lower Palaeozoic and younger moderately calcareous siltstone and sandstone. UPp' is Permian, Triassic or older black graphitic argillite, commonly siliceous, and dark quartz-rich argillaceous siltstone. JKgd is a unit of the "St. Elias Intrusions": mainly granodiorite and local hornblendite (Dodds, 1982a,b,c).

3.2.4 Hydrology

Maxwell Creek is a glacier-fed tributary of the Kaskawulsh River system. Incoming solar radiation and associated pattern of air temperature variations primarily control the strong diurnal and seasonal cycles of discharge. The main runoff component of Maxwell Creek originates from snowmelt in May through June supplemented in July through to the end of the ablation season by the ice-melt component. Small streams originating from extensive glacial and talus deposits augment runoff and sediment from the glacier through the glacierised and non-glacierised component of the basin. The nature of groundwater flows which join the subglacial system is unknown. Thus, the relative contribution to total annual flow by extraneous groundwater flow is also unknown just as the release of summer waters in winter is also unknown. During summer it is assumed that snow and ice meltwater flow dominates over groundwater runoff components.

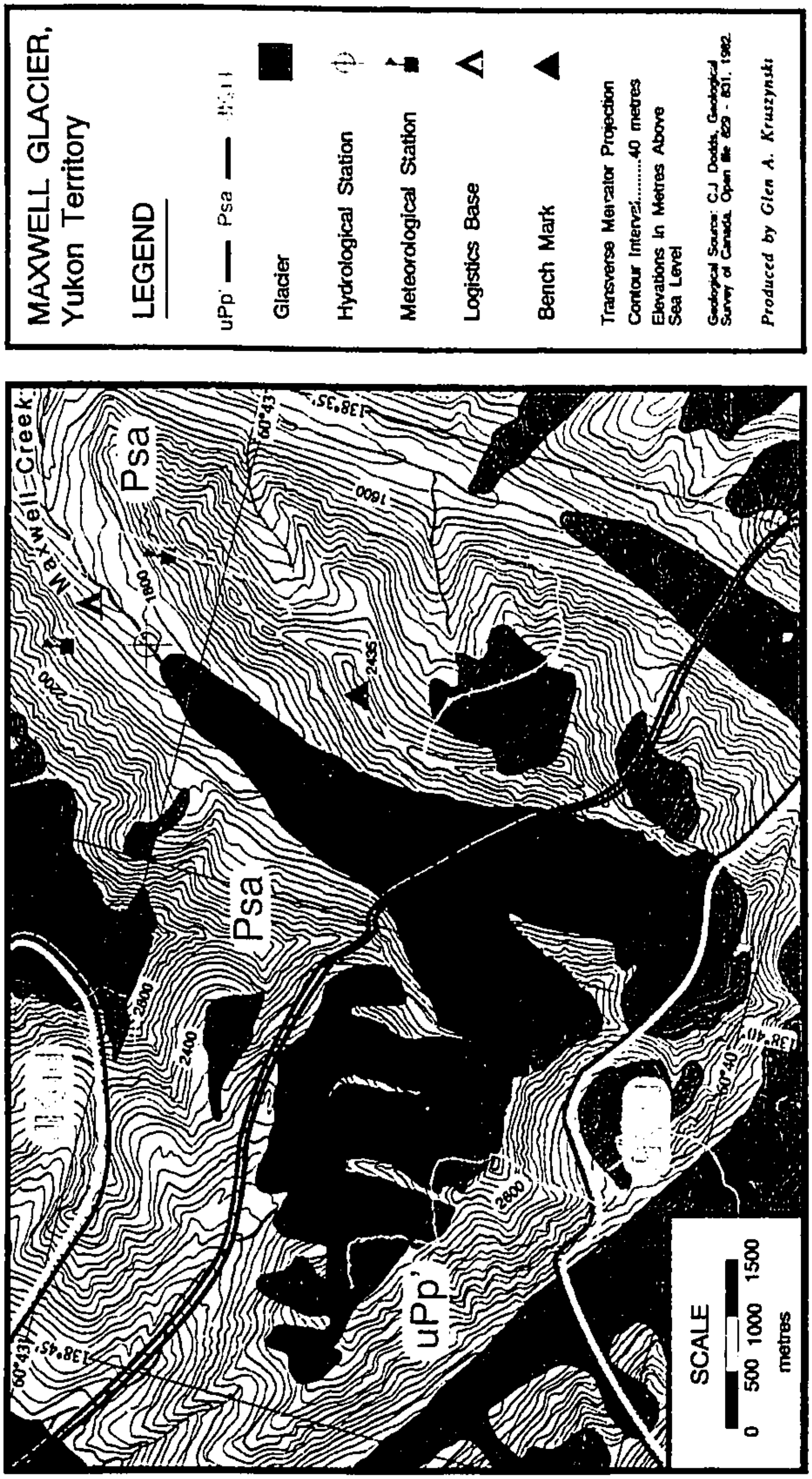


Figure 3.7: Geological Elements in the vicinity of Maxwell Glacier, St. Elias Mountains, Yukon Territory.

CHAPTER 4

METHODS AND TECHNIQUES

4.1 EXPERIMENTAL DESIGN

Modelling the drainage and transport of both sediments and solutes beneath a glacier necessitates monitoring the changing meteorological and glaciological conditions which act upon the system and the products of such a system. Inferences made from the relationships between meteorological (short-wave radiation, air temperature) and hydro-glaciological parameters (discharge, sediment concentration, and hydrochemical characteristics) are used to derive plausible scenarios for sediment acquisition and transport within an alpine glacial drainage system (Lang, 1973; Collins, 1984; Young, 1985; Röthlisberger and Lang, 1987).

A comprehensive field programme undertaken during the 1989 and 1990 ablation seasons allowed the retrieval of daily suspended sediment and meltwater samples, electrical conductivity measurements and meteorological data from Maxwell Creek and Valley.

The present investigation concentrates on once daily observations of suspended sediment and solute concentrations to provide information regarding the interrelationships between subglacial sediment sources and meltwater flow during an ablation season. Seasonal variations in suspended sediment concentration and meltwater hydrochemistry are used to assess the interactions of meltwater with subglacial sediments during the seasonal development of the subglacial drainage system. The identification and analysis of individual minor cation species will aid in the clarification of the subglacial processes (Bradley, 1990). Thence, water quality characteristics are used to asses the seasonal development in the subglacial drainage system (Liestøl, 1967; Collins, 1987; Kamb, 1987). Field work for the project was carried out from 1 June to 23 July 1989 and from 27 May to 13 August 1990. A nearly continuous series of water and sediment samples were collected during these periods of ablation. The 1989 field season resulted in the collection of 1272 suspended sediment samples and a nearly continuous collection of electrical conductivity and discharge measurements During the 1990 field season a total of 1848 hourly sediment and 142 daily water samples were collected: as well as, continuous measurements of discharge, 1848 hourly measurements of electrical conductivity (EC) and continuous measurements of meteorological data. The sampling programme consisted of taking one meltwater and suspended sediment sample at 12:00hrs every day, as well as, three days, at the beginning, middle and end of the field season, where hourly sampling continued from 12:00hrs to 11:00hrs the following day.

4.2 Sample Collection and Processing

4.2.1 Climatic Data

The timing and magnitude of meltwater flow and sediment delivery to a glacierised basin directly depends on its local climatic conditions. The presence of small glacier in a drainage basin has a marked effect on the reliability of flow by reducing the variability of runoff from year to year by comparison with ice-free basins (Krimmel and Tangborn, 1974). The direct control climate exerts on the water released from a glacierised basin has an important regulatory effect on the regimes of glacier-fed rivers (Tangborn, 1984).

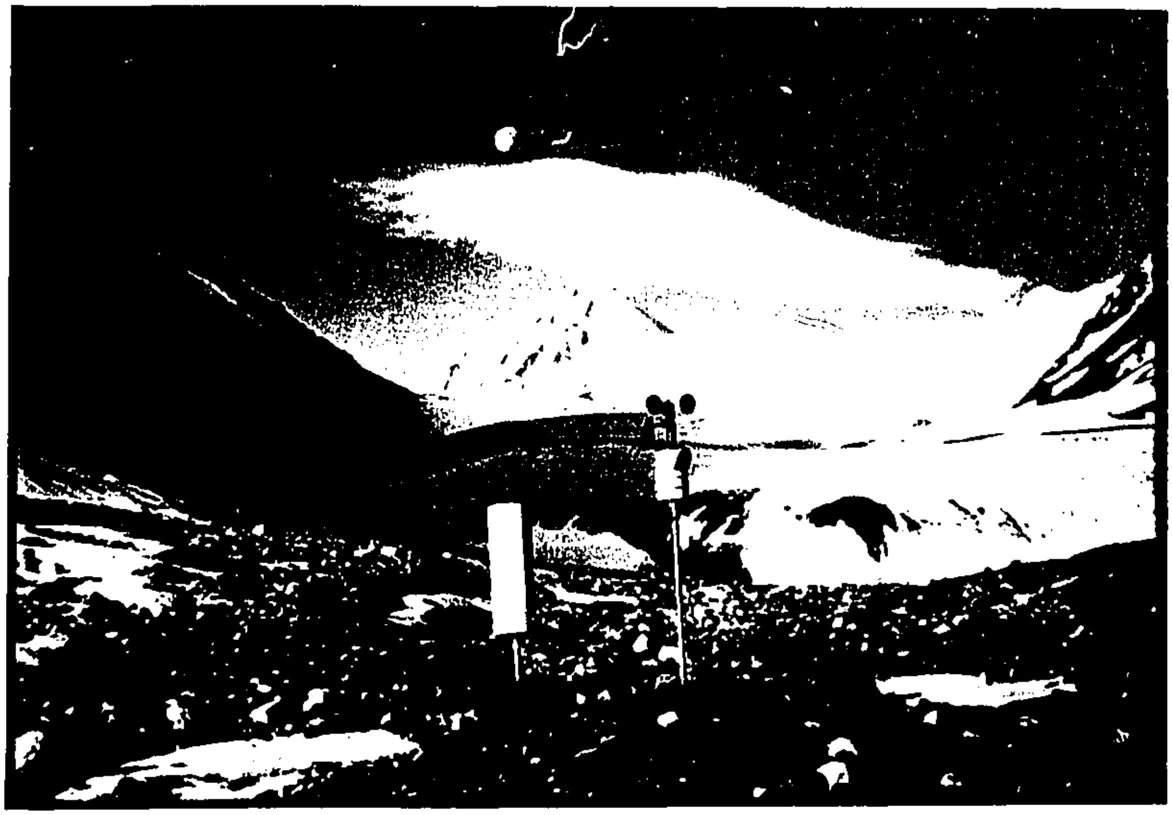
However, significant differences in summer flow (Collins, 1982; Kruszynski and Johnson, 1993) and total annual runoff from glacierised basins (Rasmussen and Tangborn, 1976) does nevertheless occur, and reflect variations in climatic conditions between years and consequential changes in the mass balance of glaciers.

The influence of local climate is assessed using climatological data from three Meteorological Research Incorporated (M.R.I.) recording stations (Figure 3.5). The tripartite M.R.I. recording stations continuously recorded temperature, precipitation, wind speed and direction (Figure 4.1). Precipitation, air temperature and short-wave radiation measurements gathered throughout the season provide hydrometeorological information relevant to Maxwell Valley, the stations being located at the valley bottom and on either valley side A Weather Measure pyranometer provided a measure of available energy for snow and ice-melt.

4.2.2 Stage - Discharge Measurements

Stage and discharge measurements were obtained from the stream during the period 1 June to 23 July 1989 and 1 June to 13 August 1990, at the gauging station 100 m from the glacier terminus (Figure 4.2). With an average gradient of 21%, the glacier rises to an elevation of 2680m. a.s.l. with its terminus approximately 1780m. a.s.l. (Figure 3.5). The close proximity to the glacier portal, well-defined stream channel and favourable hydraulic gradient permitted the use of the same site for both years, 1989 and 1990. A large boulder on the north side of the channel and a sloping bank of large angular debris on the south side helps maintain the position of the channel.

Figure 4.1: Situation of the principal M.R.I. recording station.



Stage measurements were recorded on an A. Ott Kempen mechanical stage recorder with a 7 day recording capacity. Instantaneous discharge measurements were taken on every fourth day at low, medium, and high flow conditions using the product of average flow velocities measured at regular intervals across the stream profile and the segmental area of the channel (Shaw, 1984). Channel area was determined by direct levelling. In this method the difference in elevation between a known elevation and the height of instrument, and then the difference in elevation from the height of instrument to an unknown point, are determined by measuring the vertical distance with a precise level and levelling rods. Velocities were measured with a portable A. Ott current meter, at approximately 0.6x the water depth, in half-metre intervals across the stream channel (Østrem, 1964; Church and Kellerhals, 1970; Gray, 1972).

4.2.3 Suspended Sediment

During the 1989 and 1990 field seasons, suspended sediment samples were collected at hourly intervals at the gauging site in order to minimise the effect of input from the proglacial channel. One problem with the site was that in-channel sediments within the proglacial zone were not always composed of glacial eroded sediments. Some sediments were derived from re-worked fluvioglacial sediment deposits adjacent to the stream and from gravity deposits. However, compared with sub- and englacially derived material, this source of debris was assumed to be minimised due to proximal location to the glacier portal and that gravity deposits occur primarily during the spring with the basin snowmelt (Johnson and Kruszynski, 1990). An America Sigma liquid sampler collected samples from 1 June to 23 July, 1989 and from 1 June to 13 July, 1990 until the central-processing-unit malfunctioned. Commencing 14 July 1990, a Cygnus sampler collected sediment samples

Figure 4.2: Maxwell Glacier hydrological gauging station.

through to August 13, 1990. The sampler nozzle was connected to the stilling well so as to remain below the water at the lowest stages of flow. The nozzle varied in depth from 0.1 to 0.5 m beneath the water level. Due to the highly turbulent flow, it was possible to collect 1272 and 1848 well mixed suspended sediment samples from the same fixed position in the cross-section for the periods of 1989 and 1990. Vertical adjustments were made in the position of the nozzle during the season due to the extremely variable flow. Meltwater sample volumes ranged between 200ml and 1090ml, with an average sample volume of 740ml. The volume was measured and the sample filtered through pressure filters of the type designed by the Norwegian Hydro Electric Board using Whatman 44 ashless filter papers. Samples collected were then stored in clean polyethylene bags. Although the filter papers have a retention capacity of 98%, some sediment is lost during the process due to possible minor leakage around the seal when too much pressure is applied.

Sediment bearing filter papers were placed in a 455°-510°C Pyradia oven to eliminate the filter paper through combustion. The quantity of sediment in each sample was then determined gravimetrically to a precision of 1/100mg.

4.2.4 Electrical Conductivity

Electrical conductivity (EC), used as a surrogate measure of total solute concentration, is measured electrometrically rather cheaply (i.e. by immersing a conductivity cell within the fluid to be measured) with a portable YSI Model 33 Salinity, Conductivity and Temperature meter. The uses and limitations of EC as a measure of water quality have been examined with respect to glacial meltwaters by Behrens *et al.* (1971) and Collins (1977). Electrical conductivity of the meltwater was measured hourly during the period 1 June to 23 July, 1989 and 1 June to 13 August, 1990. Standardisation of conductivities to

25°C (Jackson, 1958) from temperatures near freezing introduces high percentage of errors which exceed measurement errors (Østrem, 1964; Collins, 1977). Hence, conversion of electrical conductivity to 25°C was deemed unnecessary due to low meltwater temperatures.

4.2.5 Hydrochemical Analysis

Meltwater collected at the gauging site from 1 June to 13 August, 1990 was used for determination of hydrochemical characteristics during an ablation season according to the method of sampling designed by Lorrain and Souchez (1972). The meltwater samples were taken from the America Sigma and Cygnus sampler. A 250ml filtered aliquot was stored at cool temperatures within clean polyethylene bottles until returned to the laboratory. To avoid gaseous exchange, sample bottles were filled to the top and sealed to avoid leakage.

Before proceeding with any chemical analysis; each meltwater aliquot was acidified to 4% nitric acid by volume, shaken for 5 minutes, and filtered through Whatman glass microfibre filters (model: 934-AH). A 40ml filtrate aliquot was collected and analysed with the use of the ICP, under standard conditions. Determinations of fifteen cations were made for about 150 samples, those discussed here include the four major cations: Calcium (Ca), Magnesium (Mg), Potassium (K) and Sodium (Na). Although the other eleven cation species are not considered in this work, they may in future studies be used to reflect chemical weathering of underlying bedrock and meltwater routing.

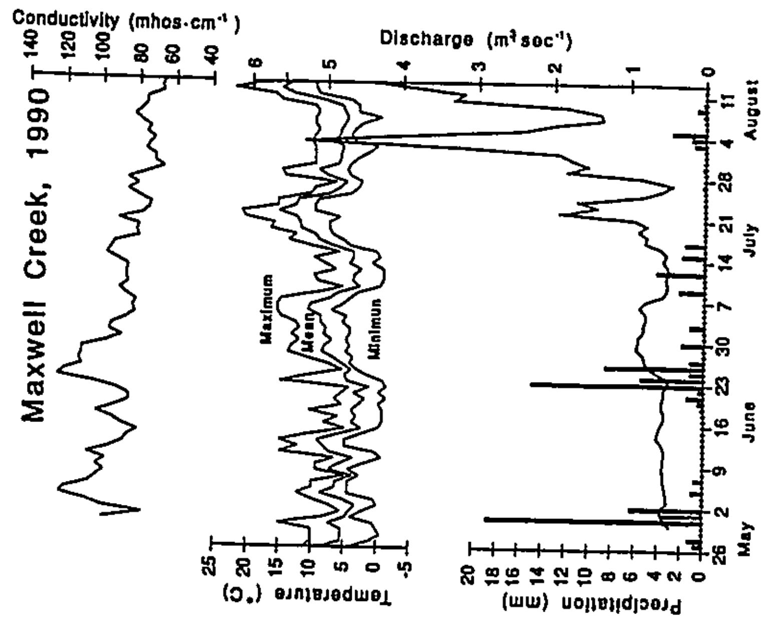
CHAPTER 5

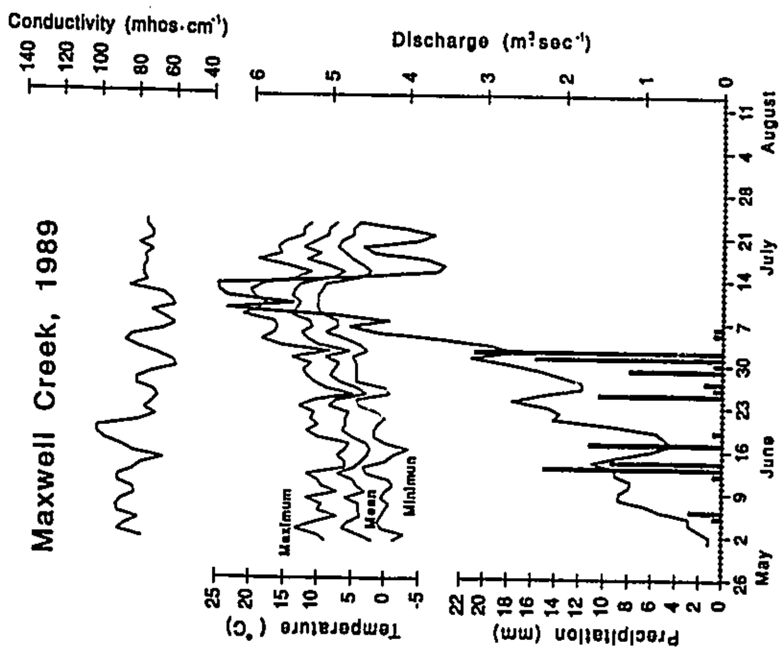
RESULTS, ANALYSIS AND SUMMARY

5.1 DISCHARGE REGIMES

The climatic conditions of the St. Elias Mountains dominate Maxwell Glacier's hydrology and mass balance (Kruszynski and Johnson, 1993). Incoming solar radiation and associated pattern of air temperature variations primarily control the strong diurnal and seasonal cycles of discharge from the glacier. Discharge, air temperature, precipitation and conductivity for Maxwell Creek from 1 June to 23 July 1989 and 27 May to 12 August 1990 are shown in Figure 5.1. The main runoff component of the stream in May though mid-July originates from snowmelt supplemented in late-July through August by glacier ice melt. During the ablation season the stream is turbulent, fast flowing and highly charged with both suspended sediment and bedload material. The runoff and sediment load of the stream is augmented by small streams originating from extensive glacial and talus deposits within the glacierised and non-glacierised component of the region (Johnson and Kruszynski, 1990).

Figure 5.1: Temperature, precipitation, conductivity and discharge records for Maxwell Creek, 1989 and 1990.

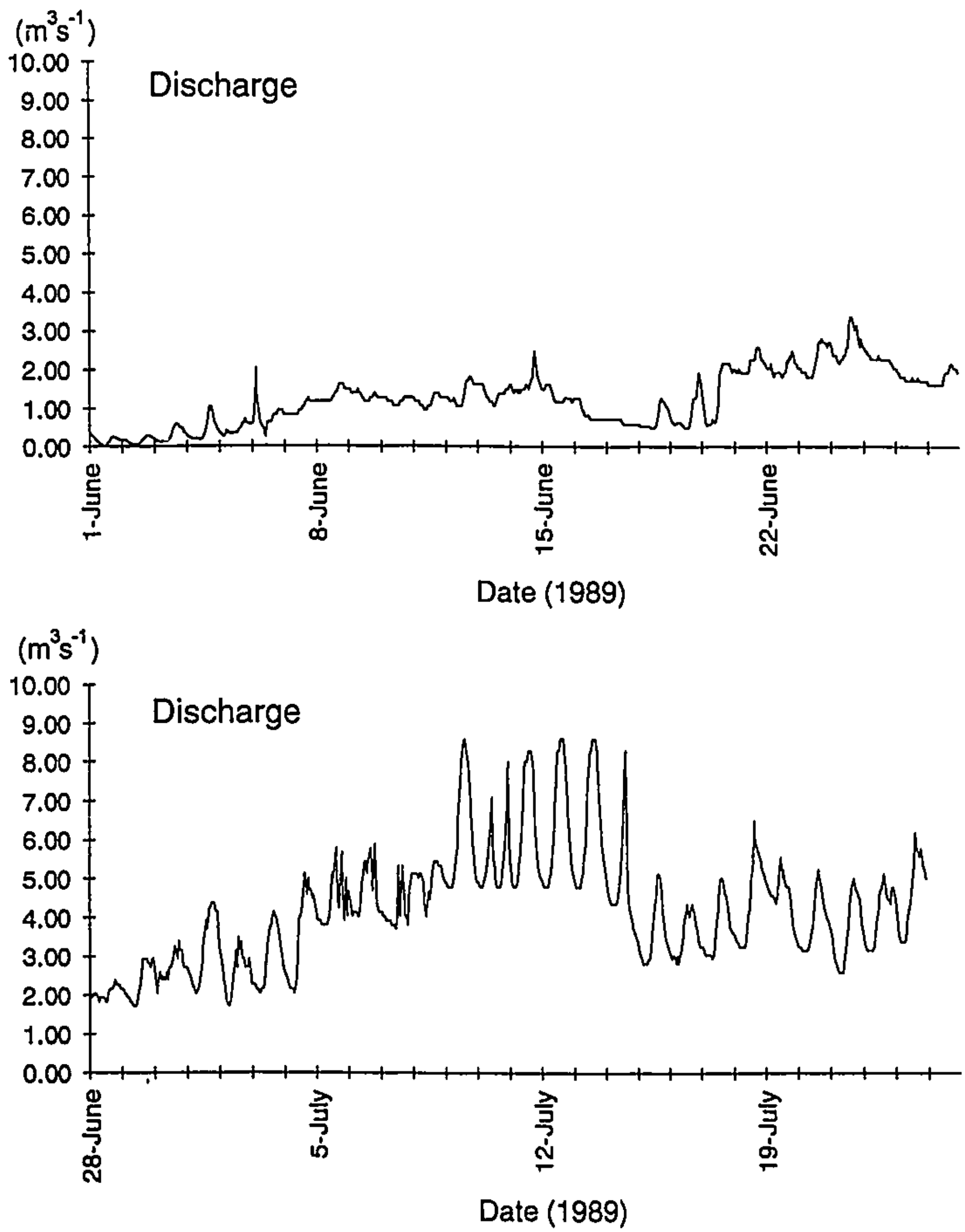




5.1.1 1989 Discharge Regime

Temporal variations in discharge for Maxwell Creek from 1 June to 24 July 1989 are plotted in Figure 5.2. The characteristic diurnal variation in discharge recognised elsewhere (e.g. Rainwater and Guy, 1961; Collins, 1977) is also apparent at Maxwell Glacier despite what some may refer to as aperiodic runoff variations (Röthlisberger and Lang, 1987) occurring on 6 June, 2, 5, 6, 7, 8 and 22 July 1989. Such characteristic diurnal variations generally reflect changes in the amount of supraglacial meltwater produced in response to daily variations in net radiation. The discharge regime of Maxwell Creek during the ablation season of 1989 (Figure 5.2) was influenced by warm dry conditions throughout early-June followed by a warm July infrequently interrupted by snow and rainfall events not lasting more than a few hours. The base flow component of discharge was marked by increasing discharge from 1 June to 6 June when a precipitation event triggered a sudden rise of discharge to a flow value of 2.05 m^3s^{-1}. In response to cooler and overcast conditions during the remainder of 6 June the flow decreased to a lower flow than the previous days recorded lowest flow. Rates of increase of discharge are relatively not affected by this short precipitation event. During the next six days the rate of increase of discharge is increased because of high rates of incoming solar radiation increasing the rate of snow melt over the entire glacier surface. On 13 and 14 June precipitation totalling more than 24.5mm produced a short rise in discharge followed by rapid decrease in the flow of discharge commencing 15 June until 18 June. Discharge levels during this three day period are lower because of reduced incoming solar radiation due to cloud cover. Discharge sharply increases over the next two days during the high flow conditions and rapidly drops off during the evenings and early morning. Commencing 21 June till 26 June discharge increases rapidly to levels exceeding the previous maximum flow of 6 June to level of 3.36 m^3s^{-1}. The period beginning 27 June until 14 July is characterised by increasing flow levels

Figure 5.2: Instantaneous discharge in Maxwell Creek, 1989.



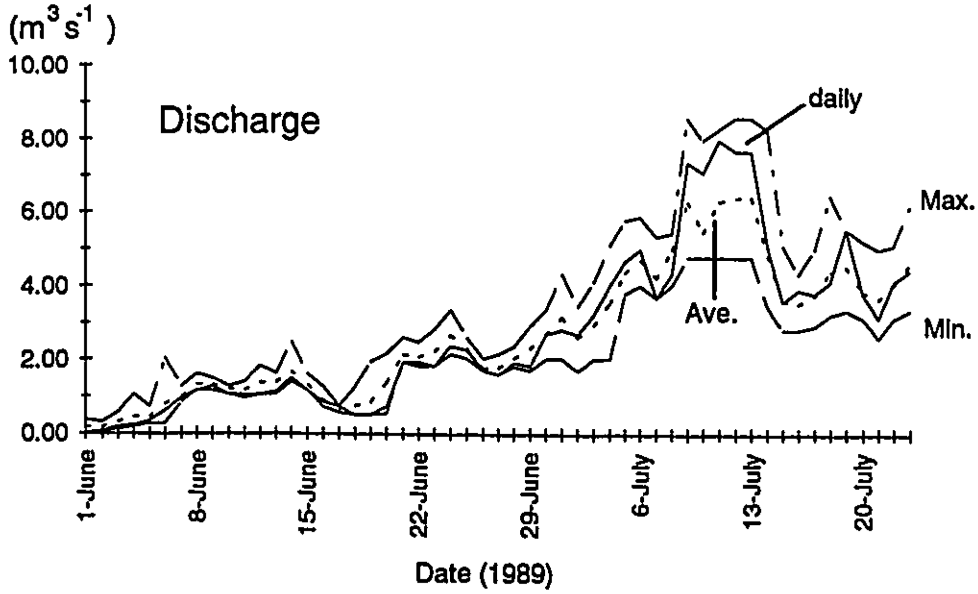
as a response to increasing levels of incoming solar radiation and reduction of the supraglacial snowpack. A discharge peak on 31 July occurs after a period of rainfall commencing at approximately 21:00hrs. 30 June. During the next two days, discharge levels are lower as a consequence of increased cloud cover. In addition to the normal diurnal fluctuations, some pulse type events were recorded on the stage charts during the period of 5 July to 8 July in which relatively little precipitation occurred and the cloud cover was mostly sunny. On 9 July discharge rapidly begins to increase until 14 July when rapidly increasing flow continues during these six days to produce a peak discharge for the measurement period of 8.6 m^3s^{-1}at 16:00hrs on 12-13 July. During this period, radiation inputs were probably at their maximum since there were no clouds present. Subsequent flow during the next two days recede to a minima of 2.82 m^3s^{-1}at 07:00hrs on 15 July before increasing to 6.5 m^3s^{-1}at 16:00hrs on 18 July. The remaining five days of the monitoring period are characterised by slowly increasing discharge to a level equal to that of the period 6-7 July.

Figure 5.3 is composed of two graphs showing the daily minimum, mean and maximum discharge values for the monitoring period, as well as the once daily measurement recorded at 12:00hrs. Base flow characteristics are defined by two periods which very in duration. The first period is defined as beginning 1 June to 28 June and the second commencing 29 June until the end of the measurement period. Unfortunately, since the record does not extend to the end of the ablation season, inferences about the duration of the second cycle are speculative. The range in discharge between diurnal minimum and maximum remains constant throughout each period respectively with mean flows approximately mid-way between minima and maxima for most days. Throughout the monitoring period the 12:00hr once daily discharge measurement approximately follows the general trend of the mean discharge.

The seasonal base flow component of discharge is characterised by the superimposition of diurnal fluctuations caused by the diurnal temporal regime over the glacier surface. During the first period mean discharge is 1.27 $m^3s^{-1}$with a minima of 0.02 m^3s^{-1} and a maxima of 3.37 m^3s^{-1}. The second period is characterised with a higher mean discharge value of 4.3 $m^3s^{-1}$with a minima and maxima of 1.72 and 8.6 m^3s^{-1} respectively. During both periods of the monitoring period the daily discharge data tends to be normally distributed. Mean discharge for the whole period is 2.71 m^3s^{-1} with a minima and maxima discharge of 0.02 and 8.60 m^3s^{-1}. Within the first period of the hydrograph, discharge minima occurs between 08:00-13:00hrs and maxima flows occur between 19:00-01:00hrs. During the second period of the measurement period, the lag between air temperature and minima or maxima discharge is reduced. The minima discharge occurs between 06:00-09:00hrs with the maxima discharge occurring between 16:00-19:00hrs. However, it seems unlikely that there could be an instantaneous relationship between the two variables, because of the finite time taken to induce melting and transfer water from the glacier surface to the snout, through the internal hydrological system. The diurnal hydrograph is generally described as having a steeper rising limb than falling limb indicating that the increase of flow for each cycle is greater than the recession. However, apart from the hydrographs from 1 June to 14 June and 9 July to 21 July, most of the cycles are interrupted by frequent occurrences of pulses as described by Johnson (1991). The pulses are concentrated in late spring and early summer when melt rates are increasing producing greater quantities of melt water. The pulses occur throughout the day suggesting independence of any diurnal regime.

The summer of 1989 discharge records show identifiable pulses throughout the mid-June to mid-July period. Although discharge variations during these events differ in detail, there are many similarities. Four major pulse events can be identified on Figure 5.2 as occurring on 6 June, 2, 4 and 6 July throughout the day. Contrary to Johnson's (1991) findings that the pulses he observed occurred primarily on the rising limb or at the peak of

Figure 5.3: Daily mean, minimum and maximum discharge in Maxwell Creek, 1989.



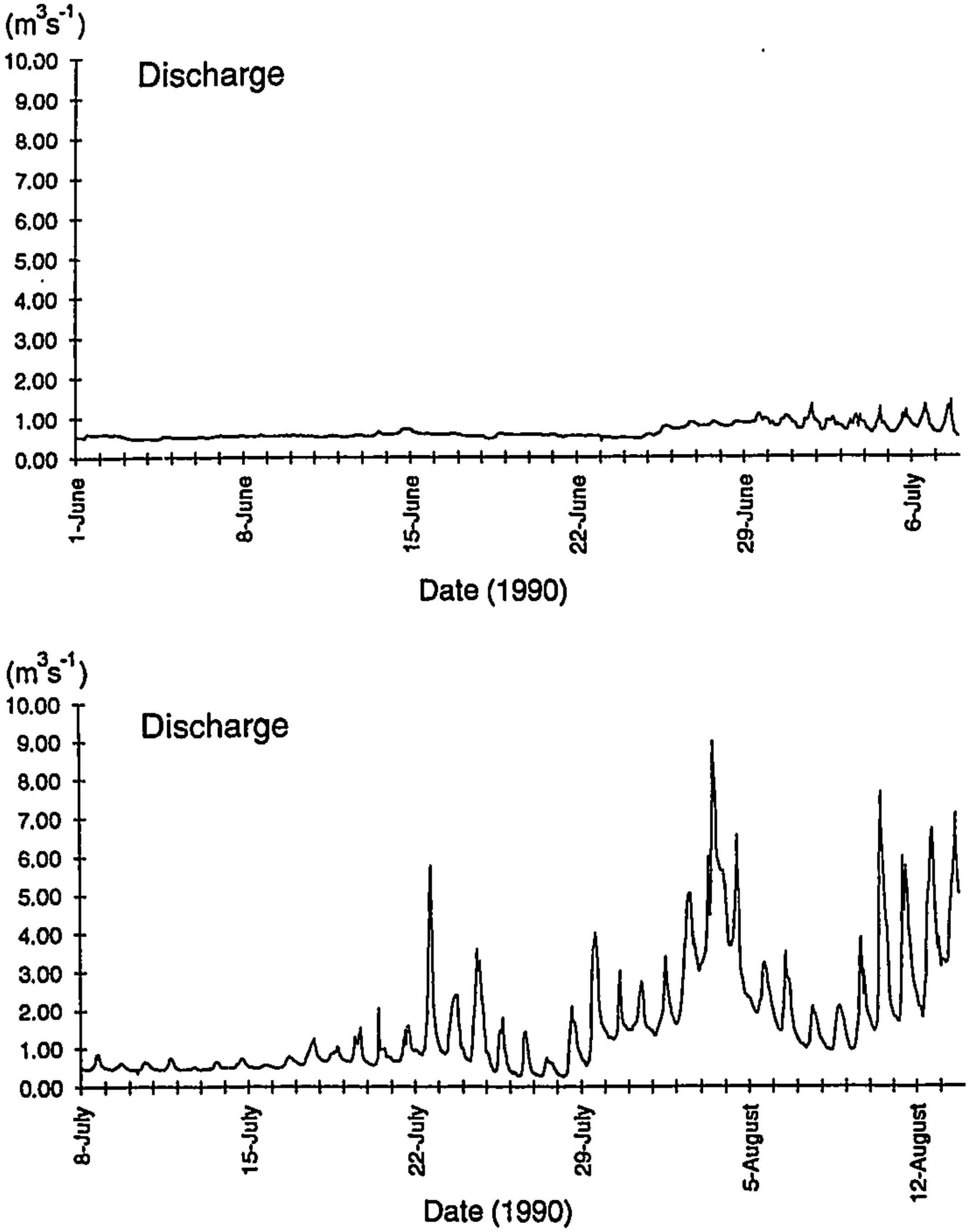
the hydrograph, pulses observed at Maxwell Creek during the 1989 ablation season generally occur throughout the day. On June 6 the hydrograph experienced a type "c" pulse (Johnson, 1991) in which there is a pulse from baseflow conditions without any preceding drop in discharge. This event has been attributed to the release of a pocket of stored water within the englacial or subglacial hydrological system (Johnson, 1991). In the period from 30 June to 9 July the frequency of pulses was 3.3 per day. On 3 July at approximately 14:00hrs a type "b" pulse having a sharp rise in discharge which was then maintained for a short duration. Johnson (1991) hypothesised that this type of pulses may be due to the opening of a previously closed section of the glacier system. Many of these pulses as seen on Figure 5.2 occurred during periods of increasing discharge, as with the recorded pulses on 5 July. The events began with a sharp drop in discharge to a rather low level, as if a major conduit had become blocked. This was followed with several minutes by a rapid increase in discharge to a value above that of the preceding drop. The excess of water released during the rise was estimated as being approximately equal to the deficit resulting from the preceding fall, again suggesting a temporary damming of a conduit as defined by a type "a" pulse (Johnson, 1991). The observed pulses during the late spring early summer are of short duration and usually low-magnitude events. They are however common occurrences in seasonal hydrograph. Further interpretation of diurnal and seasonal hydrograph record is difficult without the analysis of the electrical conductivity and suspended sediment concentration records.

5.1.2 1990 Discharge Regime

Continuous discharge measurements of Maxwell Creek between 1 June and 13 August, 1990 are shown in Figure 5.4. The 1990 hydrograph has lower discharge values in spring and summer compared with 1989. A cool period with overcast resulted in precipitation and reduction of total incoming energy availability. The low values of incoming solar radiation resulted in the slow increase of discharge until 20 July when a period of sunny conditions began to increase the melt rate resulting in a peak of discharge of 22 July of 5.76 m^3s^{-1}, an increase of 4.94 m^3s^{-1} compared with the previous days minima value of 0.82 m^3s^{-1}. Heavy smoke from forest fires in the southern Yukon retarded ablation during the period commencing 7 July to 16 July, 1990. On 16 July the heavy smoke from forest fires had dissipated allowing greater melt at the glacier surface producing a gradual increase of discharge. The period from 16 July to 20 July experienced minor cloud cover generating limited glacier surface melt. Over the following five days maximum incoming solar radiation was incident at the glacier surface over its entire length since cloud cover was minimal. Increased cloud cover and cooler temperatures on 15 July to 28 July resulted in a sharp decrease in discharge. On 28 July discharge was recorded at 07:00hrs to be 0.28 m^3s^{-1} , its lowest level since the beginning of the monitoring period. Discharge levels rapidly increased again on 29 July as sunny warmer conditions prevaile·'. The period from 29 July to 4 August was marked by decreasing cloud cover and warmer temperatures producing an almost instantaneous rise in discharge to a monitoring period peak of 8.98 m^3s^{-1} on 3 August 1990. The remainder of the monitoring period was marked by a short period of precipitation followed by two days of cloud cover and a 4 day sunny period. During the remainder of the monitoring period, discharge decreased with the presence of cloudy periods before and during precipitation events, after which daily discharge returned to slightly lower than previously recorded levels.

Seasonal patterns of discharge in the Maxwell Creek during the 1990 monitoring period are indicated by the daily mean, minimum and maximum discharge levels in Figure 5.4. Figure 5.5 illustrates the daily minimum, mean and maximum discharge values for the monitoring period, as well as the once daily measurement recorded at 12:00hrs. The data

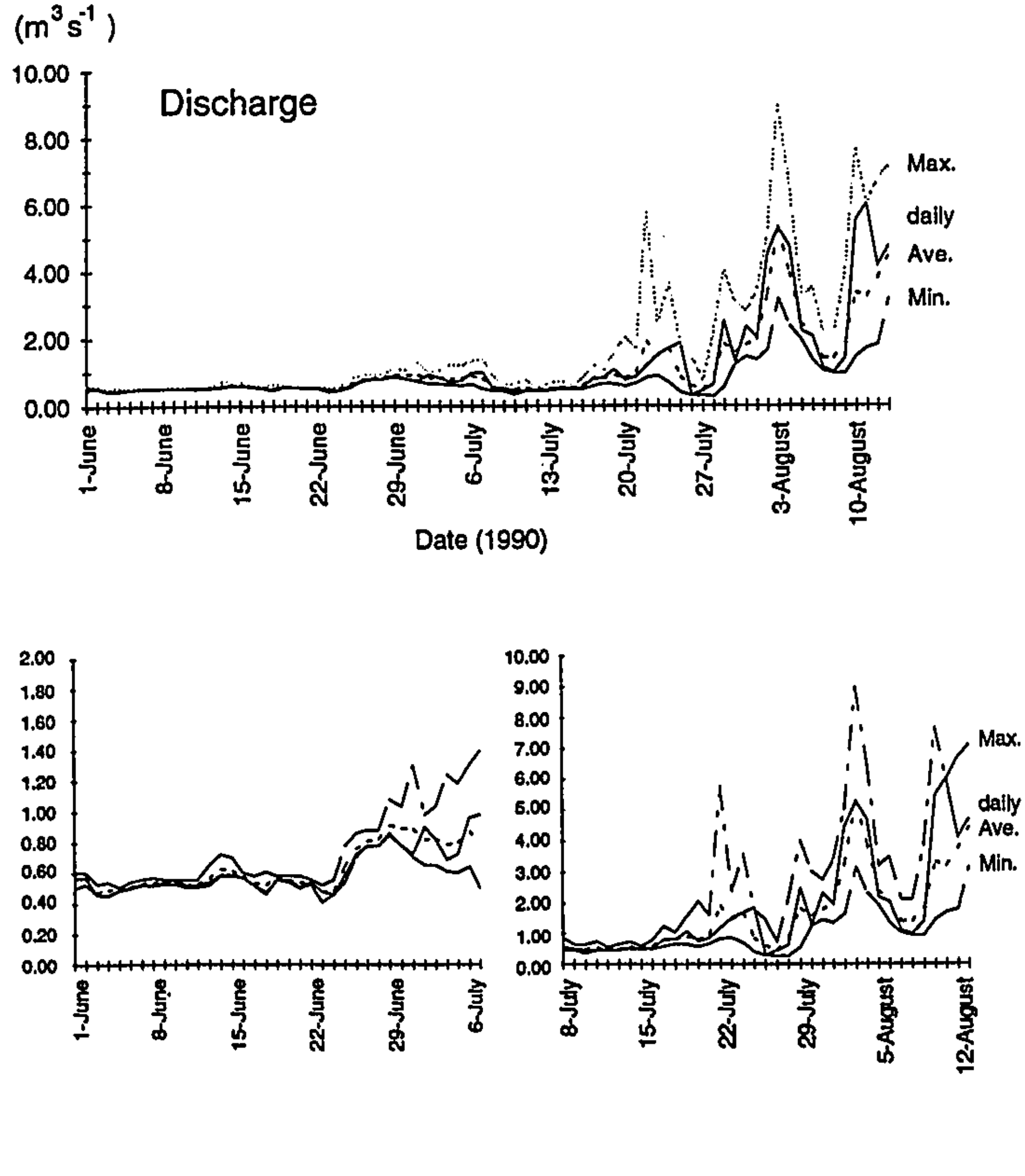
Figure 5.4: Instantaneous discharge in Maxwell Creek, 1990.



presented in the lower half of Figures 5.4 and 5.5 is divided into two periods. The scale of the first period is exaggerated 5x that of the left scale to better illustrate the amount of diurnal variation. Base flow characteristics are defined by two periods which very in duration. The first period is defined as beginning 1 June to 15 July and the second commencing 16 July until the end of the measurement period. Unfortunately, since the record does not extend to the end of the ablation season, inferences about the duration of the second cycle are speculative. Discharge levels are contained within a broad range of 8.72 m^3s^{-1} with a mean value of 1.12 m^3s^{-1}, minima of 0.26 m^3s^{-1} and a maxima of 8.98 m^3s^{-1} recorded on 28 July and 4 August. The range in discharge between diurnal minimum and maximum remains approximately constant throughout each period respectively with mean flows almost mid-way between minima and maxima for most days. Throughout the monitoring period the 12:00hr once daily discharge measurement approximately follows the general trend of the mean discharge. From 18 July onwards, the diurnal fluctuations of the stream become marked. It is well documented that the amplitude of the diurnal fluctuations tends to increase towards late summer when ablation rates are highest (Sugden and John, 1984). A 6.22 m^3s^{-1} difference in discharge between low flow conditions (06:00hrs, 1.44 m^3s^{-1}) and peak flow conditions (14:00hrs, 7.66 m^3s^{-1}) was the greatest recorded on 10 August.

In 1990 a cool June and early July was followed by a warm mid-July which lead into variable conditions throughout the remainder of the monitoring period. The precipitation regimes for 1990 is in contrast with 1989 where June and early part of July 1990 was beset by cloudy days, occasional showers and heavy snowfall on 23 June. Warm, clear weather during the remainder of July 1990 raised flow to its highest levels. During both years, cloudy periods before and during precipitation events were associated with a decrease in discharge, after which daily discharge returned to higher than previously recorded levels.

Figure 5.5: Daily mean, minimum and maximum discharge in Maxwell Creek, 1990.



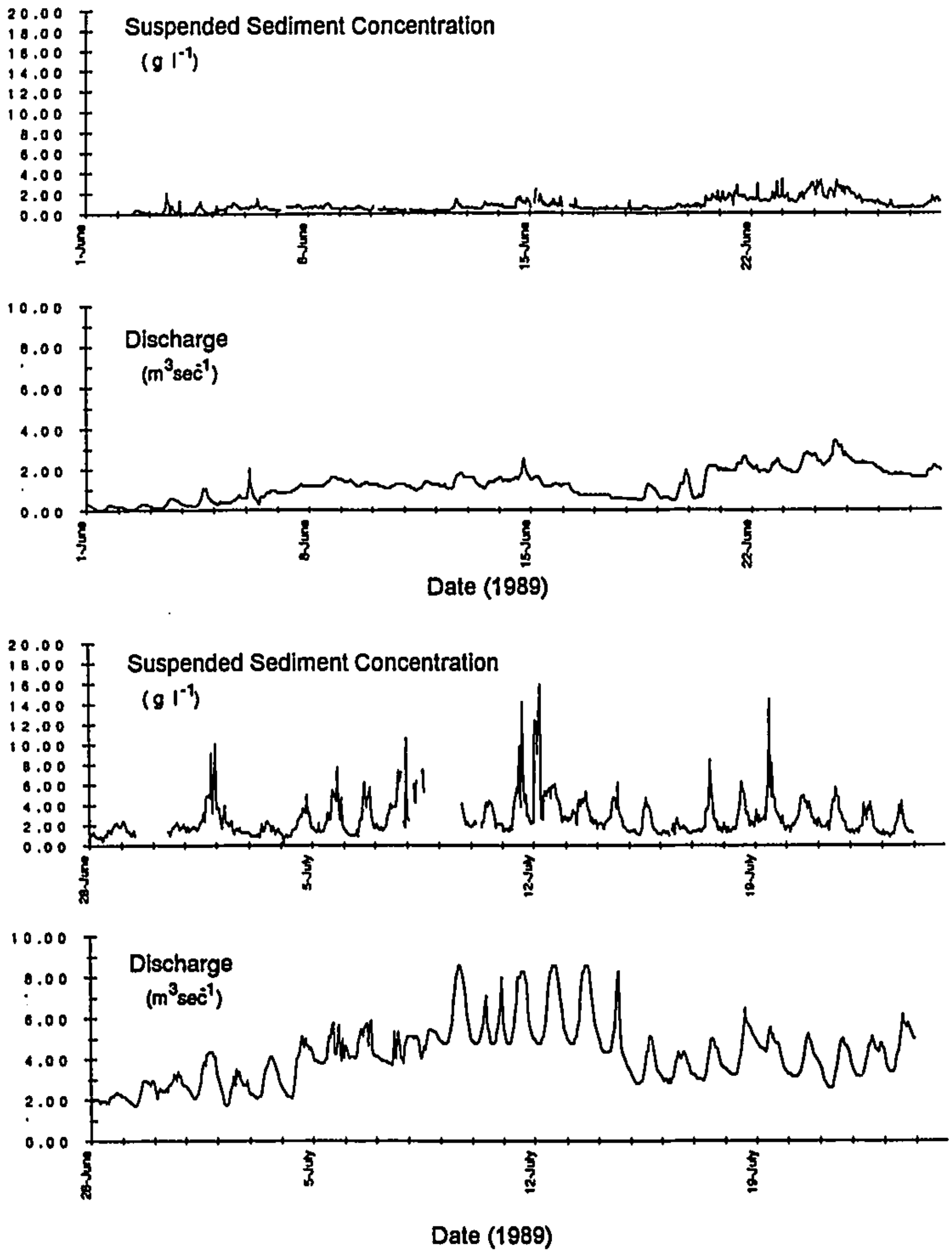
5.2 Suspended Sediment Regime

5.2.1 1989 - Suspended Sediment Regime

Temporal variations in the transport of suspended sediment material in Maxwell Creek from 1 June to 24 July, 1989 are shown in Figure 5.6. This nearly continuous record of suspended sediment concentration (SSC) makes it possible for detailed analysis and interpretation of both diurnal and seasonal trends in the data. Suspended sediment concentrations are characterised by a daily flow component of diurnal cyclic variations with short term fluctuations lasting between a quarter of an hour to two to three hours. Records indicate a strong dependence of sediment concentration on discharge except in the presence of short-term fluctuations superimposed on longer term fluctuations. The characteristic diurnal variations in SSC recognised elsewhere (Østrem, 1975; Richards, 1984) is generally in phase with the discharge data despite the presence of characteristic peaks and troughs in SSC occurring during entire monitoring period (i.e. 6, 21 June and 8, 11, 12, 17 and 19 July). Maximum sediment concentrations tend to occur on the rising limb of the hydrograph when the rate of increase of discharge is greatest. Suspended sediment concentrations then rapidly fall on the falling limb when reduced discharges and correspondingly lower stream competence allows deposition within the stream channel. The short-term fluctuations, of the order of several hours, disguise what otherwise may be a simple diurnal series.

Suspended sediment concentrations (SSC) during the monitoring period are within a range between 0.04 gL^{-1} on 3 June (09:00hrs) and 15.85 gL^{-1} on 12 July (06:00hrs). The mean SSC for the observation period is 1.72 gL^{-1} with a standard deviation of 1.72 gL^{-1}. Suspended sediment concentrations for the entire observation form a bimodal distribution which is separated into two well defined temporal periods. The first period is defined as beginning 1 June to 28 June and the second commencing 29 June until the end of the

Figure 5.6: Instantaneous suspended sediment concentration and discharge in Maxwell Creek, 1989.

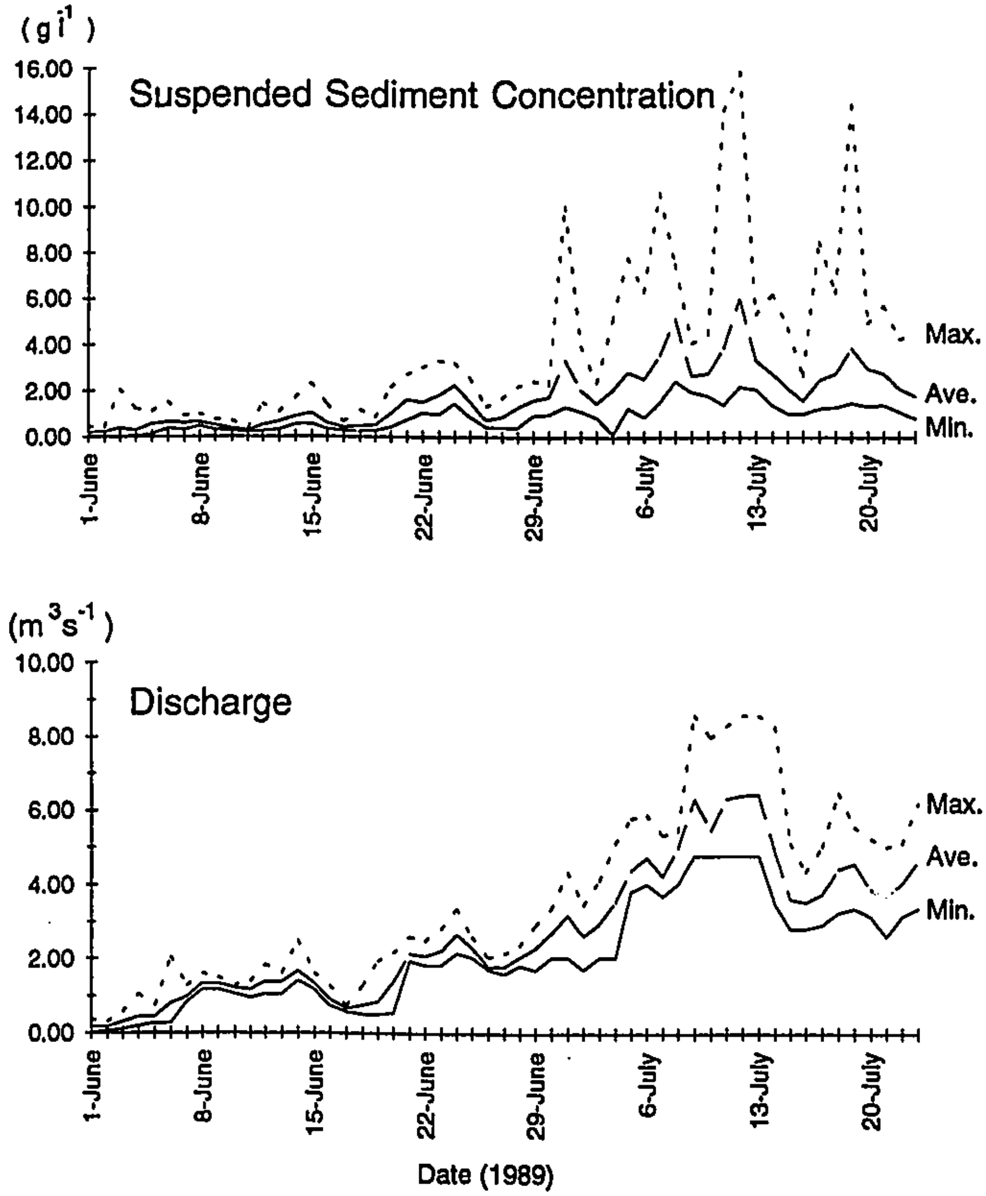


measurement period. The range in SSC between diurnal minimum and maximum remains approximately constant throughout each period respectively with mean flows approximately mid-way between minima and maxima for most days (Figure 5.7). During the first period, mean SSC is 0.80 gL^{-1} with a minima of 0.03 gL^{-1} and a maxima of 3.29 gL^{-1}. The second period is characterised with a higher mean discharge value of 2.72 gL^{-1} with .. minima and maxima value of 0.40 and 15.87 gL^{-1} respectively. During both periods of the monitoring period the SSC data tends to be normally distributed except during the second period. During the second period short-term fluctuations in SSC occur producing extraneous values. The range in SSC for the both periods is 3.26 gL^{-1} and 15.47 gL^{-1} respectively. During the period from 29 June till 14 July the SSC become increasingly variable as discharge levels increased.

Examination of diurnal changing SSC illustrates that the diurnal relationship between discharge and SSC is not as well defined as others have indicated (i.e. (Østrem, 1975; Richards, 1984). Østrem (1975) and Richards (1984) suggest that SSC diurnal increase with rising flows reaching a peak at times of greatest rate of increase in discharge before starting to decline at peak discharge and continuing to decrease as flow discharge declines. However, as illustrated in Figure 5.6 and reported by Binda *et al.* (1985) and Collins (1979a), SSC peaks may occur both before and after discharge maxima or have no pattern at all (Figure 5.8).

The diurnal and seasonal variations in suspended sediment concentration and discharge during four different sub-periods in Maxwell Creek, 1989 is shown by the plot in Figure 5.8. Suspended sediment concentrations during the period of 2 June to 11 June were marked by very low concentrations with a weakly defined trend to follow discharge levels. Sediment concentrations during this period peak between 2-3 hours before maxima discharge levels followed and preceded by short-term variations in SSC. During this period the diurnal variation of discharge and sediment concentration is minimal. The average daily

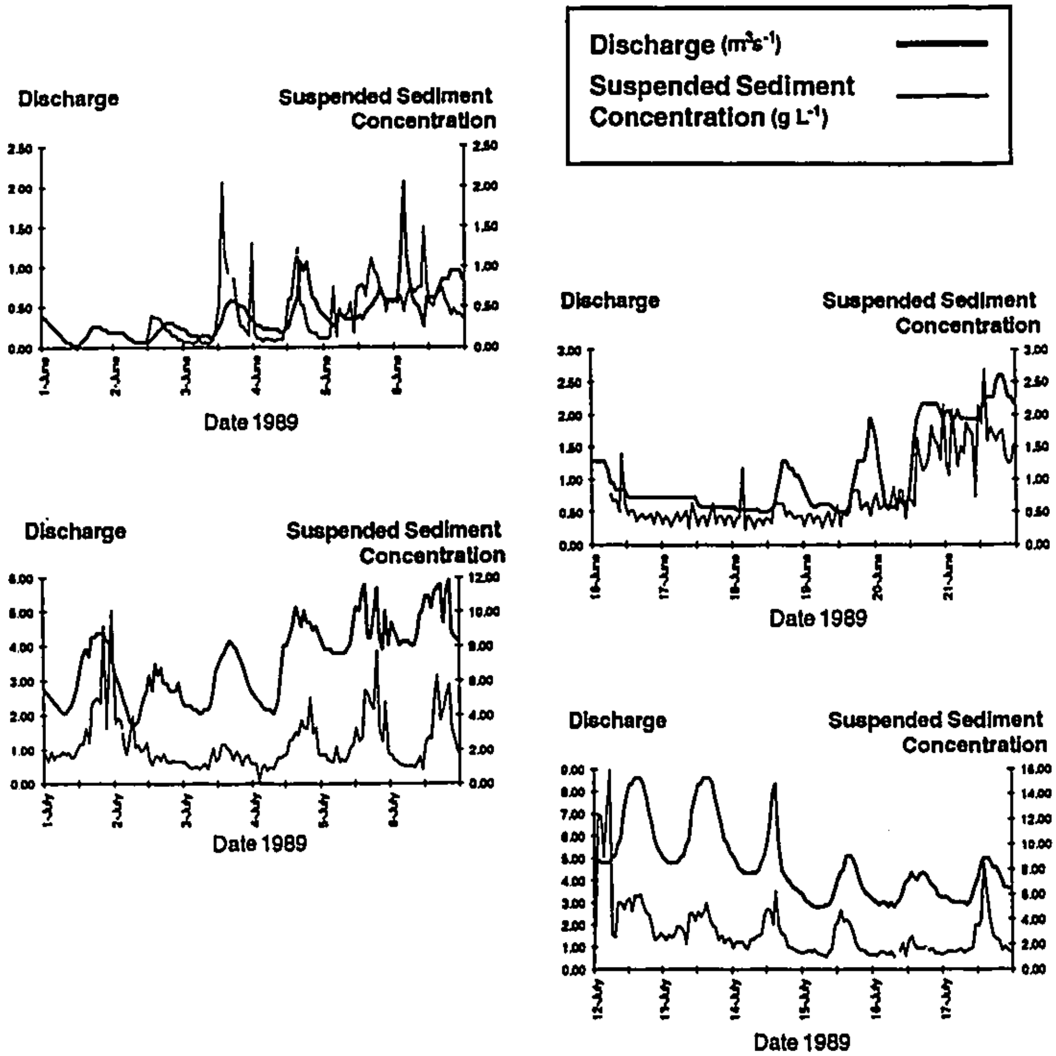
Figure 5.7: Daily mean, minimum and maximum suspended sediment concentration and discharge in Maxwell Creek, 1989.



range in SSC and discharge is 1.21 gL^{-1} and 0.99 m^3s^{-1}. On 6 June, peak concentrations occur 6 hours after peak discharge is possibly in response to acquisition of stored sediment at the base of the glacier. From 6 June onwards, hourly detail in SSC time series is less defined and inversely proportional to discharge. During the evening of June 12, SSC values coincide with the maximum rate of increase in flow. A drop in electrical conductivity (Figure 5.1) may represent derivation from increased supraglacial melt. Increased meltwater from supraglacial stores may have been able to rework sediments deposited during the preceding decline in diurnal flow. Although discharge levels remain approximately the same as the period from 7 June till 11 June, SSC increases rapidly during the next 4 days (until 16 June) to levels exceeding the previous maximum concentration values of 1 June till 11 June. A rapid decrease in the flow of discharge commencing 15 June until 18 June results in a sharp and lasting decline in sediment concentrations until 20 June. From 18 June onwards till 20 June the discharge regime is marked by the characteristic diurnal increase and decrease of flow. However, during this period SSC is depressed and characterised by short pulse events possibly caused by sporadic opening and closing of cavities during high pressure events at the base of the glacier. These variations in SSC continues to persist for the next 2 days until 23 June. The period beginning 23 June until 29 June is characterised by diurnal fluctuations in discharge with corresponding fluctuations in SSC. This segment of the time series is characterised as decreasing by decreasing flow and sediment concentrations. When the daily flow component of the hydrograph is declining the SSC tend to follow due to declining levels in the effective transport of sediments in subglacial stores.

The segment of the time series from 30 June to 15 July contains the peak concentration in SSC for the entire monitoring period. During low flow conditions on 12 July, a concentration maximum of 15.87 gL^{-1} was recorded at the monitoring station at 06:00hrs. This concentration maximum is the culmination of 15 days of increasing daily

Figure 5.8: Suspended sediment concentration and discharge during four different sub-periods in Maxwell Creek, 1989.



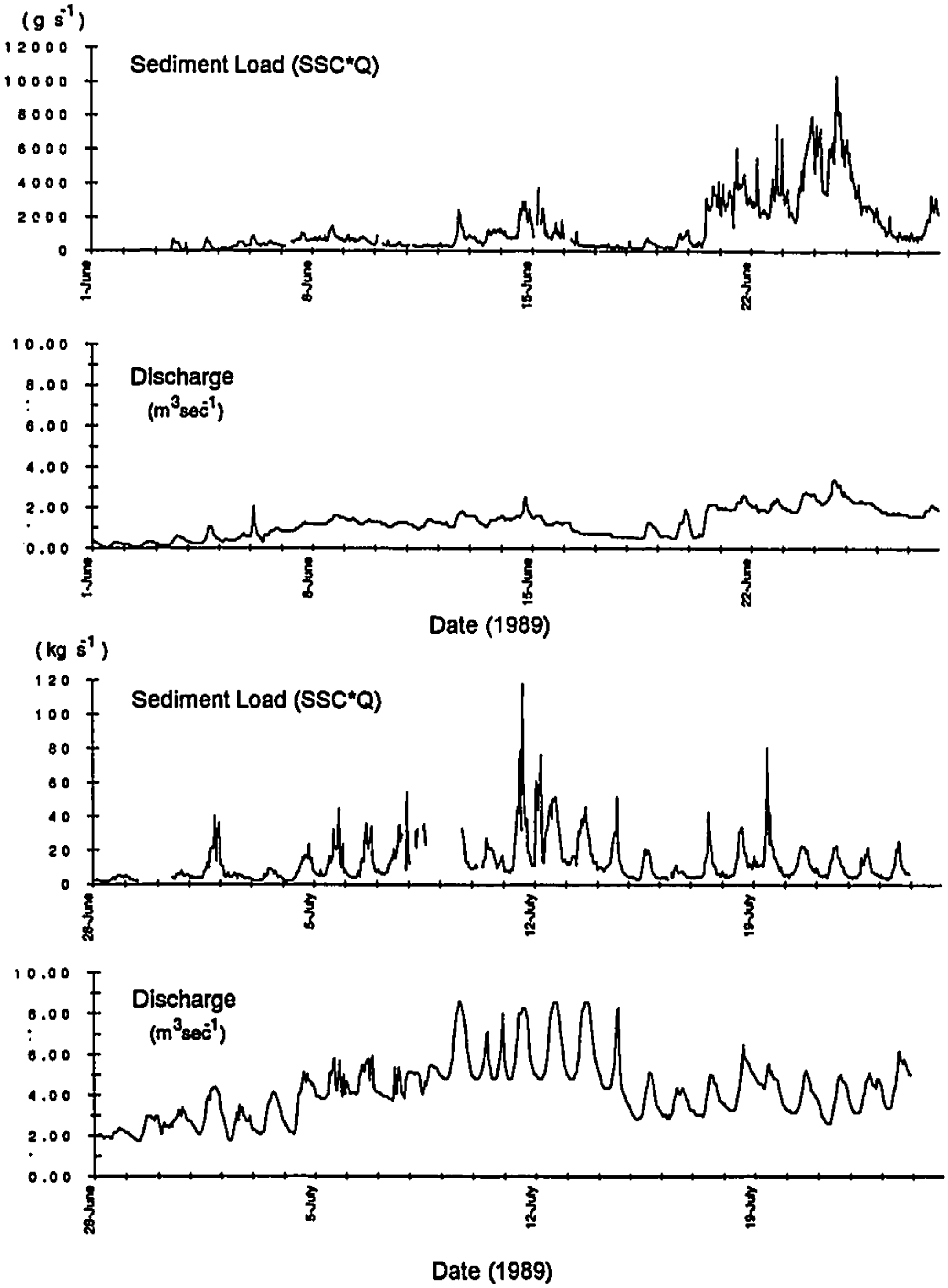
flow and hydrostatic conditions at the base of the glacier. The period leading up to the SSC maximum on 12 July SSC exhibits a diurnal fluctuation in-phase with that of the flow conditions while exhibiting continued hourly fluctuations. On 1 July, a major sediment-transport event occurred (9.80 gL^{-1}), in a period when each day the maximum instantaneous flow exceeded that of the previous day. On 2 July, sediment concentrations decline with discharge to a minimum of 0.40 gL^{-1} at 04:00hrs on 4 July. This does not coincide with the rising daily hydrograph over this period. In the morning and afternoon of 4 July, rising daily flows are accompanied by increasing sediment concentrations with maxima SSC occurring 1-3 hours before and after maximum flow until the evening of 6 July. On 7 July, sediment concentrations rapidly increased during increasing flow conditions and continued to increase after maximum flow to a concentration of 10.60 gL^{-1} at 23:00hrs. Suspended sediment concentrations in the Maxwell River from the morning of 8 July to the afternoon of 10 July are shown in Figure 5.6. Unfortunately this period contains many missing values as a result of problems with the field data collection program. Thus it is not possible to relate flow variations with SSC, detailed analysis and interpretation of diurnal variations over this period are not possible. From 8 July onwards to 14 July, flow levels reached their highest level recorded during the 1989 monitoring period. On 10 July, the diurnal flow regime was interrupted by increasing cloudy conditions suppressing surface melt and discharge levels from 13:00hrs to 19:00hrs when flow began to increase. Suspended sediment concentration during this time reached a daily maximum at 15:00hrs (4.44 gL^{-1}), 3 hours after maximum discharge. Sediment concentrations remained low for the next 9 hours during continued increase and then decrease of discharge throughout the evening and early morning of 11 July.

The segment of the time series from 11 July to 14 July contains the peak concentration in SSC for the entire monitoring period. On July 11, peak flow at 16:00hrs produces the largest SSC to-date with 14.15 gL^{-1}. During the hour prior to maximum flow the SSC dropped markedly low to 3.82 gL^{-1} from 9.82 gL^{-1} with a return to 14.15 gL^{-1} without any decrease in discharge. During the early morning of 12 July (01:00hrs), discharge levels continued to decline reaching a minima of 4.8 m^3s^{-1} at 03:00hrs until 07:00 when a slight increase of 0.07 m^3s^{-1} was measured. During this period of the time series, sediment concentrations markedly increased to 12.38 gL^{-1} while flow levels continued to decline. During the next 4 hours SSC varied between 9.00 gL^{-1} and 15.87 gL^{-1}, only to then rapidly decline to concentrations equal to that of 5, 6, 7 and 10 July.

From 15 July onwards, the diurnal changing patterns of SSC are increasingly more discernible than the previous segments of the time series. Suspended sediment concentrations from 15 July until the end of the monitoring period progressively increased, reflecting parallel increases in discharge through sustained ablation. On 17 and 19 July marked fluctuations in sediment concentrations occurred although there was not a major change in discharge, possibly indicating temporary access and exhaustion of sediment supply due to increasing basal water pressures. For the remaining segment of the time series despite gradually increasing discharge, suspended sediment concentrations continue to decline with relatively well defined diurnal fluctuations. During daily receding flows, sediment concentrations continue to exhibit minor fluctuations in concentrations while SSC decline throughout the remainder of the period.

The pattern of the seasonal interaction between the development of the basal drainage network and the delivery of suspended sediment in Maxwell Valley in the 1989 measurement period is shown by the suspended sediment load time series plot in Figure 5.9. Suspended sediment load is a product of fine sediment being washed out from basal zones in contact with flowing water. The product of discharge and SSC, a surrogate of sediment flux, is dependent on the areal integration of basal sediment and basal water flow (Collins, 1989). Suspended sediment load first increased on 12 June, when maximum hourly mean instantaneous discharge (1.72 m^3s^{-1}) exceeded that of 11 June by 0.33 m^3s^{-1}. Further

Figure 5.9: Instantaneous suspended sediment load and discharge in Maxwell Creek, 1989.



increased flow on 20 June raised instantaneous sediment load to a maxima of 3180gs^{-1}. Sediment load continued to increase until 24 June with discharge remaining high but not greatly exceeding those flows of 21, 22 and 23 June. Cool, overcast weather from 25 June to 30 June lowered flow and sediment levels. Sediment load increased throughout the rising flow of 1 July, reaching a maximum instantaneous sediment load of 40.02 Kgs^{-1} at 21:00hrs. A short-term suspended sediment pulse occurred 3 hours later at 24:00hrs, producing an instantaneous sediment load of 36.04 Kgs^{-1}. Debris supply is reduced on 2 July as sediment loads rapidly decline. However, discharge levels quickly regain their former flow levels on 5 July exceeding the previous maxima level recorded on 1 July by 3.40 m^{3}s^{-1}. Increasing discharge levels augmented by warm, clear weather from 5 July to 8 July produced a parallel rise in sediment load. Suspended sediment load in the Maxwell Creek from the morning of 8 July to the afternoon of 10 July are shown in Figure 5.9. Unfortunately this period contains many missing values as a result of problems with the field data collection program. Thus it is not possible to relate flow variations with sediment load, detailed analysis and interpretation of diurnal variations over this period are not possible. From 8 July onwards to 14 July, sediment load reached its highest level recorded during the 1989 monitoring period. Season maximum daily transport was achieved on 11 July at 16:00hrs (117.45 Kgs^{-1}), one day before maximum suspended sediment concentration. For the remaining segment of data, sediment load remains parallel to discharge. Increased solar radiation over the basin on 18 July produced yet another instantaneous sediment evacuation (33.61 Kgs^{-1}), exceeding again during the sunny day of 19 July (80.49 Kgs^{-1}), after which event the rate of transport of suspended sediment material remained less than 25.36 Kgs^{-1} up to the end of the monitoring season.

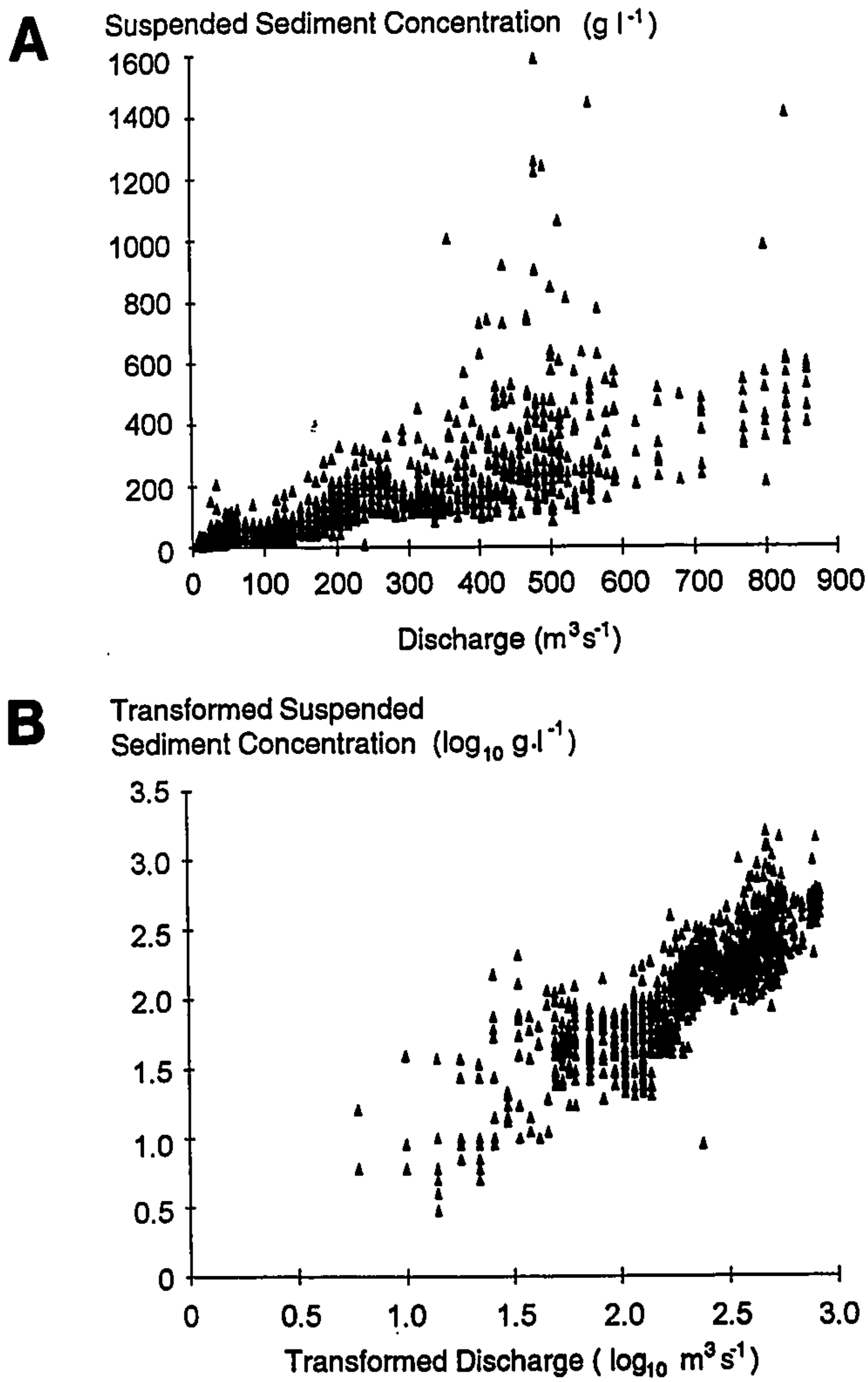
The association between suspended sediment concentration and discharge has often been described and estimated by way of sediment rating curves. Sediment rating curves are derived by applying linear regression analysis to concurrent SSC and discharge observations.

Since the flow of water within a stream transports the sediment, it would seem that a good association between SSC and discharge would be expected. Within a glacierised basin there would also seem to be an ample supply of rock debris that would reduce the regulating effect of sediment supply on the SSC - discharge relationship. Several authors (Mathews, 1964b; Østrem, 1975; Collins, 1979a; Bogen ,1980) have applied rating curves to paired observations taken from glacierised basins over diurnal and longer time scales. However, the suspended sediment regime in glacierised basins is often complicated by short, frequently occurring sediment and/or discharge concentrations and flows.

Linear regression analysis demands that the variables be randomly distributed around a zero mean with constant variance and serially uncorrelated (Gurnell and Fenn, 1984). Figure 5.10 displays the observations of SSC and discharge for Maxwell Creek, 1989. It is clear in Figure 5.10A that such assumptions are not met; a strong linear trend does not emerge. The scatter in the plot can be reduced by transforming the axes to a logarithmic scale (Figure 5.10B). The logarithmic transformation appears to increase the linearity between the variables. However, examination of the residuals indicates a high degree of heteroscedasticity within the data. Hence, it is impossible to estimate a simple regression model relating SSC and discharge for Maxwell Creek.

Østrem (1975) noted however that it is possible to estimate sediment rating curves for a glacierised basin if very short periods of time are considered. Collins (1979a) separated observations into individual rising and falling limbs of diurnal discharge hydrographs and found that hysteresis is a major factor influencing the linear relationship between the SSC and discharge data. Daily rating curves, as defined by Collins (1979a) are produced by SSC peaks on rising and falling limbs producing clockwise and anticlockwise hysteresis loops. Clockwise hysteresis loops indicate greater SSC peaks prior to flow maximum at times of maximum increasing discharge than at similar flow levels on recessional limbs. Anticlockwise curves however describe diurnal SSC peaks at times of declining flow. These

Figure 5.10: Scatter plot of all paired discharge and suspended sediment concentration; as well as all paired log-transformed discharge and suspended sediment concentration in Maxwell Creek, 1989.

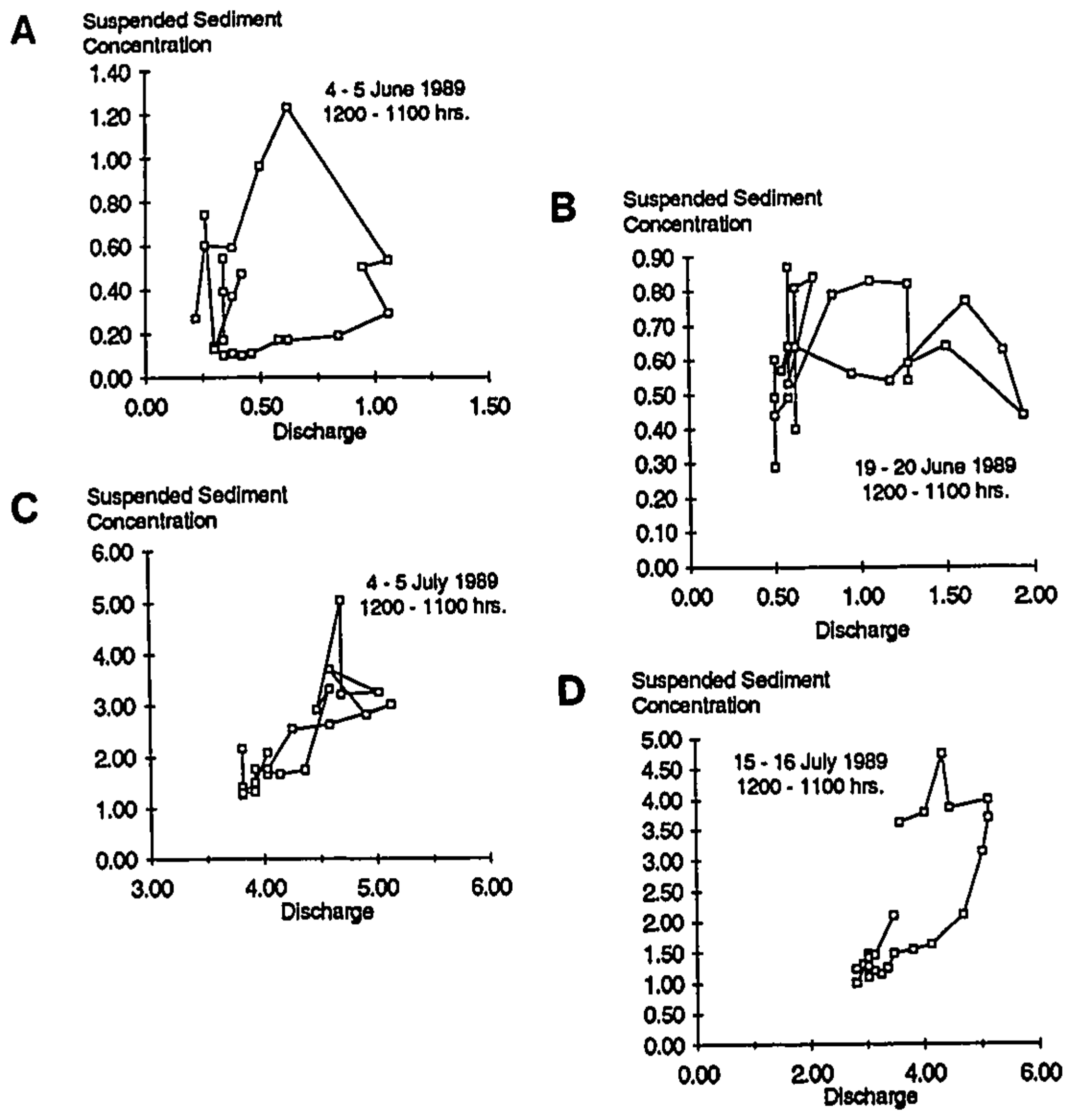


diurnal rating loops may continue through a period of days (Østrem, 1975) and may be apparent over an entire ablation season (Lemmens, 1978). In Figure 5.11 suspended sediment concentration is plotted against discharge for four days during the monitoring period. Each plot in Figure 5.11 consists of paired SSC and discharge data sequentially plotted from the rising limb on one day to the rising limb of the next.

Figure 5.11A shows clockwise hysteresis between 12:00hrs on 4 June to 11:00hrs on 5 June. The looped plot follows a pattern of clockwise hysteresis whereby the same discharge is associated with rising sediment concentrations on the rising limb of the hydrograph. Whereby the water that composes part of the initial daily increase in flow has access to considerably more sources of fresh sediment than water at the same stage on the falling limb of the hydrograph. Figure 5.11B shows connected concurrent discharge and SSC data for the 24 hour period beginning 12:00hrs on 19 June. Two sediment pulses during this period occur on the rising and falling limbs of the 19-20 June hydrograph. This produces both clockwise and anticlockwise hysteresis loops in the plot. Minor fluctuations on both SSC and discharge result in involution's in the plot during low and high flow conditions. Figure 5.11C shows that the relationship between SSC and discharge from 12:00hrs to 11:00hrs on 4 to 5 July respectively is complex with many involutions in the data. During this period, both clockwise and anticlockwise hysteresis loops are produced together with involutions. These involutions and shifting between clockwise and anticlockwise loops are caused by sudden pulse events in concentration record which may not be attributed to changing rates of flow. The form of the plot in Figure 5.11D is one of a quite regular clockwise hysteresis loop. Greater sediment concentrations occur prior to discharge maxima at times of increasing discharge than at similar discharge levels on the recessional limb.

The plot of hysteresis between SSC and discharge data continually changes during the monitoring period as shown in the variety of different looped patterns in Figure 5.11. There

Figure 5.11: Suspended sediment concentration hysteresis in Maxwell Creek, 4, 19 June and 4, 15 July, 1989.



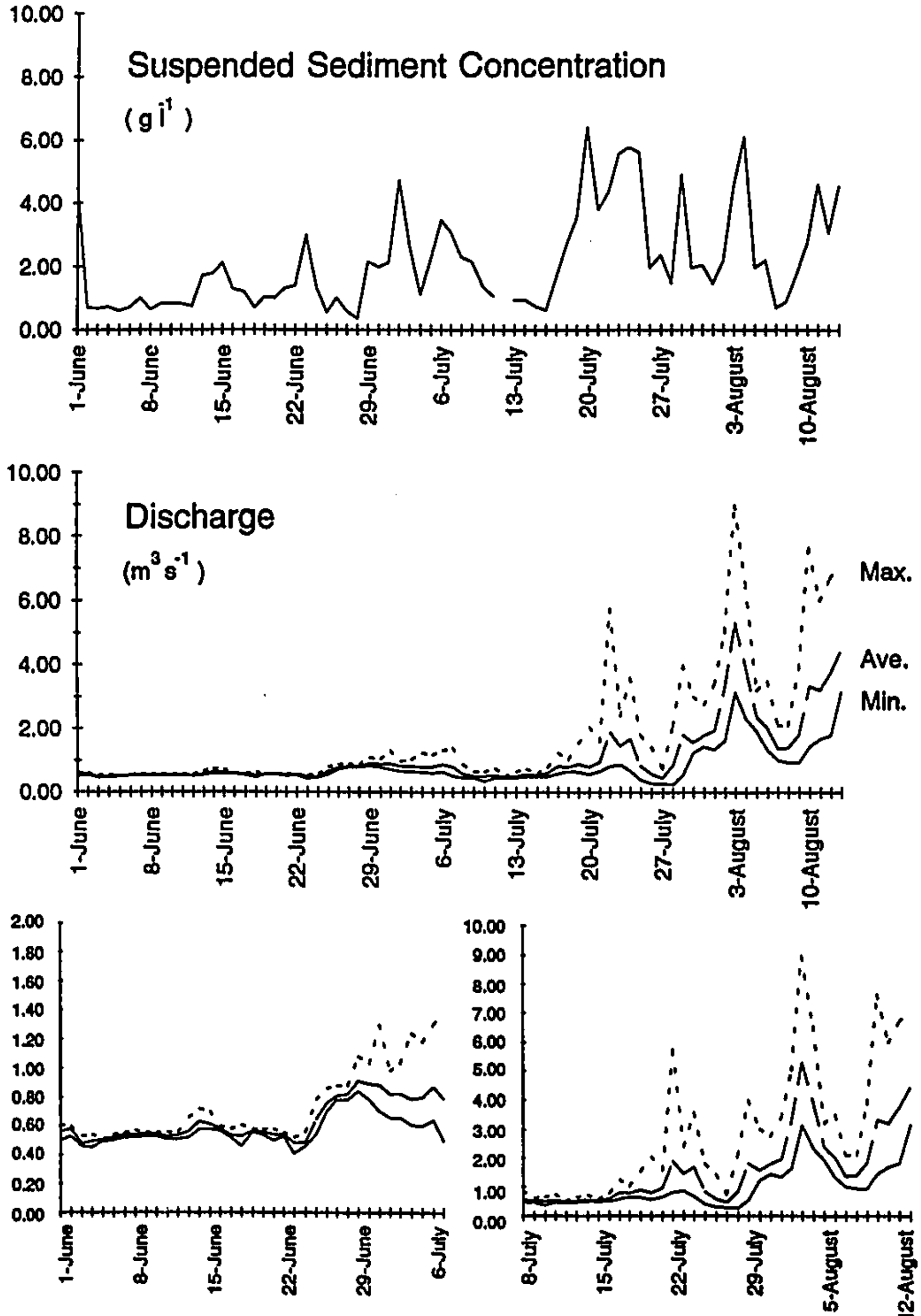
are two important factors influencing the changing nature of the looped pattern relationships between SSC and discharge. First, the exhaustion of sediment supply occurs during periods when the current discharge level has been exceeded. Second, access to stored sediments at the base of the glacier by water results in sudden outbursts of sediment. Such outbursts may or may not be accompanied by discharge variations (Figure 5.6) and may be generated by an assortment of glaciological and hydrological processes. Hence, not only must the present discharge and sediment data be examined but the recent past discharge record is also important to understand to what degree it has influenced the present sediment record.

5.2.2 1990 - Suspended Sediment Regime

Suspended sediment concentration and discharge data in the Maxwell Creek between 1 June and 13 August, 1990, is shown in Figure 5.12. Apart from the short break in the record between 11 and 13 July, the time series is continuous. The observations represent a once daily sediment sample taken at 12:00hrs each day throughout the monitoring period. Consequently, it is not possible to interpret conclusively detailed daily SSC characteristics for the 1990 period although it is possible to relate general trends in the data.

Sediment concentrations are contained within a broad range between 0.34 gL^{-1} and 6.38 gL^{-1} (Figure 5.12). A minimum SSC of 0.34 gL^{-1} was sampled at 12:00hrs on 28 June. The mean concentration for the monitoring period is 2.14 gL^{-1} with a standard deviation of 1.56 gL^{-1}. The maximum SSC for the entire period was 6.38 gL^{-1} at 12:00hrs on 20 July. Although SSC data was sampled once per day, calculations using all available daily discharge data for 1989 and 1990 indicate that the 12:00hr samples lie approximately mid-way between minimum and maximum daily values suggesting a near normal distribution of data. However, the variability observed in the 1989 data restricts the interpretation of the

Figure 5.12: Daily suspended sediment concentration and discharge in Maxwell Creek, 1990.



1990 data to long-term seasonal trend analysis.

Suspended sediment concentration data in the Maxwell Creek are well related with changes in discharge levels. Suspended sediment concentrations for the entire observation form a bimodal distribution which is separated into two well defined temporal periods. The first period is defined as beginning 1 June to 15 July and the second commencing 16 July until the end of the measurement period. During the first period, mean SSC is 1.50 gL^{-1} with a minima of 0.34 gL^{-1} and a maxima of 4.70 gL^{-1}. The second period is characterised with a higher mean discharge value of 3.13 gL^{-1} with a minima and maxima value of 0.61 and 6.38 gL^{-1} respectively. Plotted available data indicate that sediment concentrations during the first period exhibit sharply increasing and decreasing trends during the period which exceed the variation of discharge. The SSC data during the second period of the monitoring period tends to be normally distributed except during the period of 16 July to 20 July. During the second period, the variation in daily SSC closely follows the increasingly more variable discharge values. The range in SSC for the both periods is 4.36 gL^{-1} and 5.77 gL^{-1} respectively. The plotted data for the first period suggests that data are characteristic of highly fluctuating SSC in proglacial streams.

The seasonal pattern of SSC (Figure 5.12) illustrates the development of the basal drainage network and the delivery of suspended sediment in Maxwell Valley during the 1990 measurement period. Suspended sediment concentrations first increased at the beginning of the measurement period (1 June, 4.43 gs^{-1}) when flow was 0.57 m^3s^{-1}. A moderate increase in flow on 12 June raised SSC to a maxima of 1.4 gs^{-1} on 15 June at a peak discharge of 0.59 m^3s^{-1}. Sediment concentrations continued to closely follow the rising and declining trend in discharge data for the period from 16 June to 15 July. Daily fluctuations in SSC continue to exhibit a greater amount of variation between days compared to daily fluctuation in discharge. Sediment concentration during the period of 23 June to 28 June exhibited a decline in concentration despite rising discharge levels till 7 July.

This slight reduction in sediment transport represents a partial depletion of sediment stores due to prior exhaustion during the period of 12 June to 17 June and 19 June to 22 June. Forest fires in southwest Yukon during the period of 6 July to 15 July produced an abundant amount of smoke in the area. The presence of smoke in the valley resulted in decreased surface melt rate accounting for the decline in discharge and SSC levels in the stream. From 18 July onwards, the diurnal fluctuations of the stream become marked. A 6.22 m^3s^{-1} difference in discharge between low flow conditions (06:00hrs, 1.44 m^3s^{-1}) and peak flow conditions (14:00hrs, 7.66 m^3s^{-1}) was the greatest recorded on 10 August (Figure 5.4). Unlike the maximum discharge level recorded on 10 August, sediment concentration maxima is recorded on 20 July (6.38 gL^{-1}) three weeks prior to maximum flow conditions. Suspended sediment concentrations during the period of 18 July to 30 July continue to fluctuate greater than the fluctuations in discharge. During the period of 31 July to the end of the measurement period, the variation in SSC and discharge data is reduced. While sediment concentrations maintain a high amount of fluctuation between samples the amount of difference is reduced between sediment and discharge. Despite declining variation in SSC, sediment concentrations remain high with relatively well defined pairing with discharge values. With prior removal of sediment (20 July to 26 July) from basal stores the varying magnitude of sediment removal from the Maxwell Glacier is shown by the sediment - discharge plot in Figure 5.12. Increasing discharge levels and stable sediment concentrations suggest that new sediment supplies are not be incorporated into subglacial flow towards the end of the measurement period.

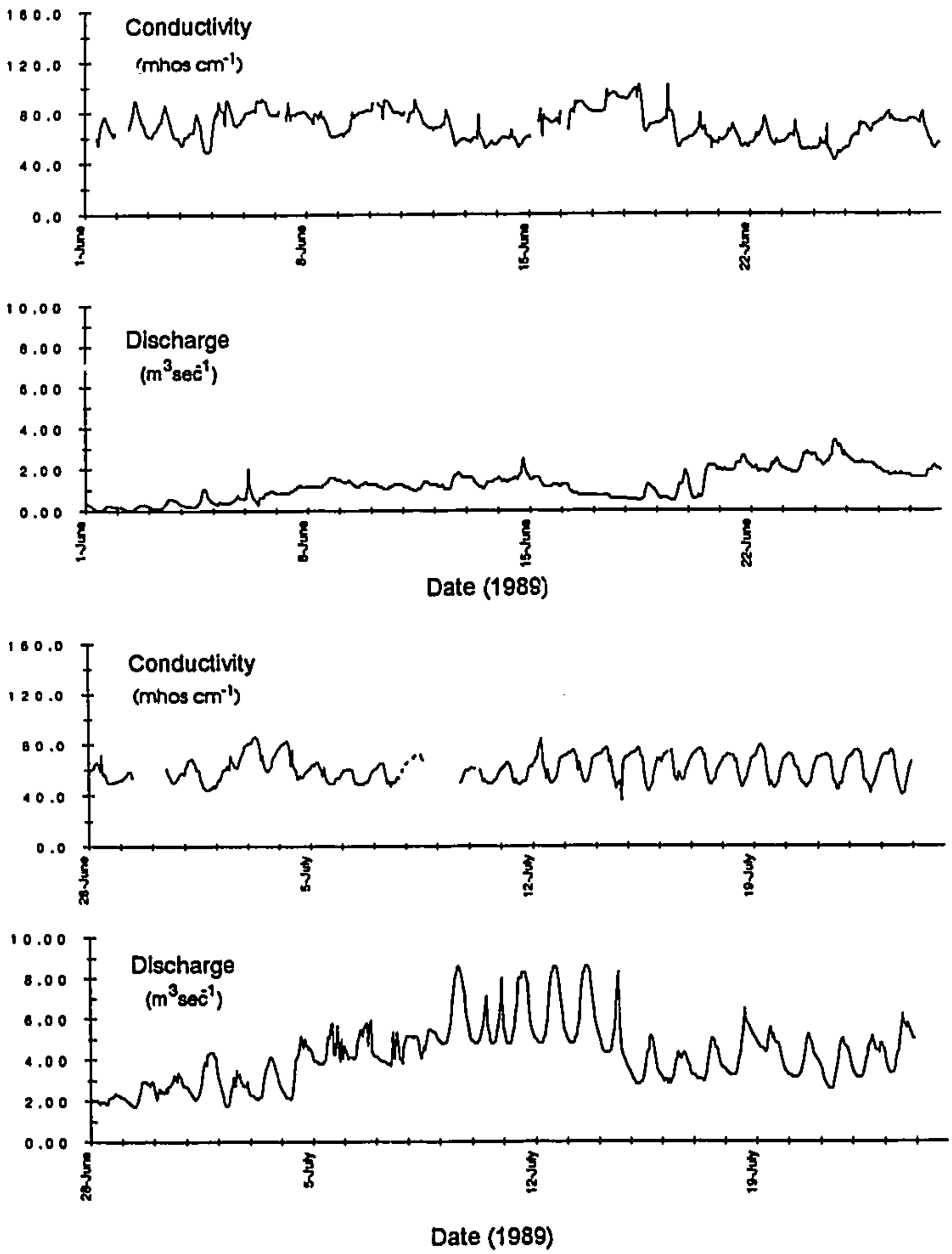
5.3 ELECTRICAL CONDUCTIVITY

5.3.1 1989 Electrical Conductivity

Continuous records of electrical conductivity (EC), a surrogate measure of solute concentration, of meltwaters in Maxwell Creek, the only portal stream draining from the Maxwell Glacier are plotted in Figure 5.13. The record of EC is broken during two periods for an extended amount of time, from 29 June (12:00hrs) to 30 June (11:00) and 8 July (09:00hrs) to 9 July (17:00hrs) when the automated sampler was inoperative. The EC and discharge data in Figure 5.13 illustrates that electrical conductivity is, in general, inversely proportional to discharge. During the majority of the monitoring period electrical conductivity is maximal when discharge levels are minimal.

Conductivity data are contained within a range of 40.0 to 101.0 mhos cm^{-1} with a standard deviation of 12.0 mhos cm^{-1} during the monitoring season. Mean EC in 1989 is 65.3 mhos cm^{-1} with a maximum value of 101.0 mhos cm^{-1} recorded on 18 and 19 June (10:00hrs and 12:00hrs). Conductivity data are marked by a general decrease in value as discharge levels gradually increase from 1 June to 24 June. The maximum EC value coincides with a dramatic decrease in discharge due to poor weather conditions. The period prior to 16 June, from 8 June to 15 June, exhibits an increase in EC values with gradually rising discharge levels. These values seem to suggest increased surface meltwaters had accessed subglacial water sources which have undergone chemical enrichment through contact with basal sediments. Minimum EC was measured at 17:00hrs on 23 August when a value of 40.0 mhos cm^{-1} was recorded. Theoretically this value should coincide with peak discharge when water is most dilute although in this case it does not. It is proposed that beginning 15 July to the end of the measurement period (23 August) the subglacial drainage network is well established, basal water pressures have decreased as a result of declining

Figure 5.13: Instantaneous electrical conductivity and discharge in Maxwell Creek, 1989.

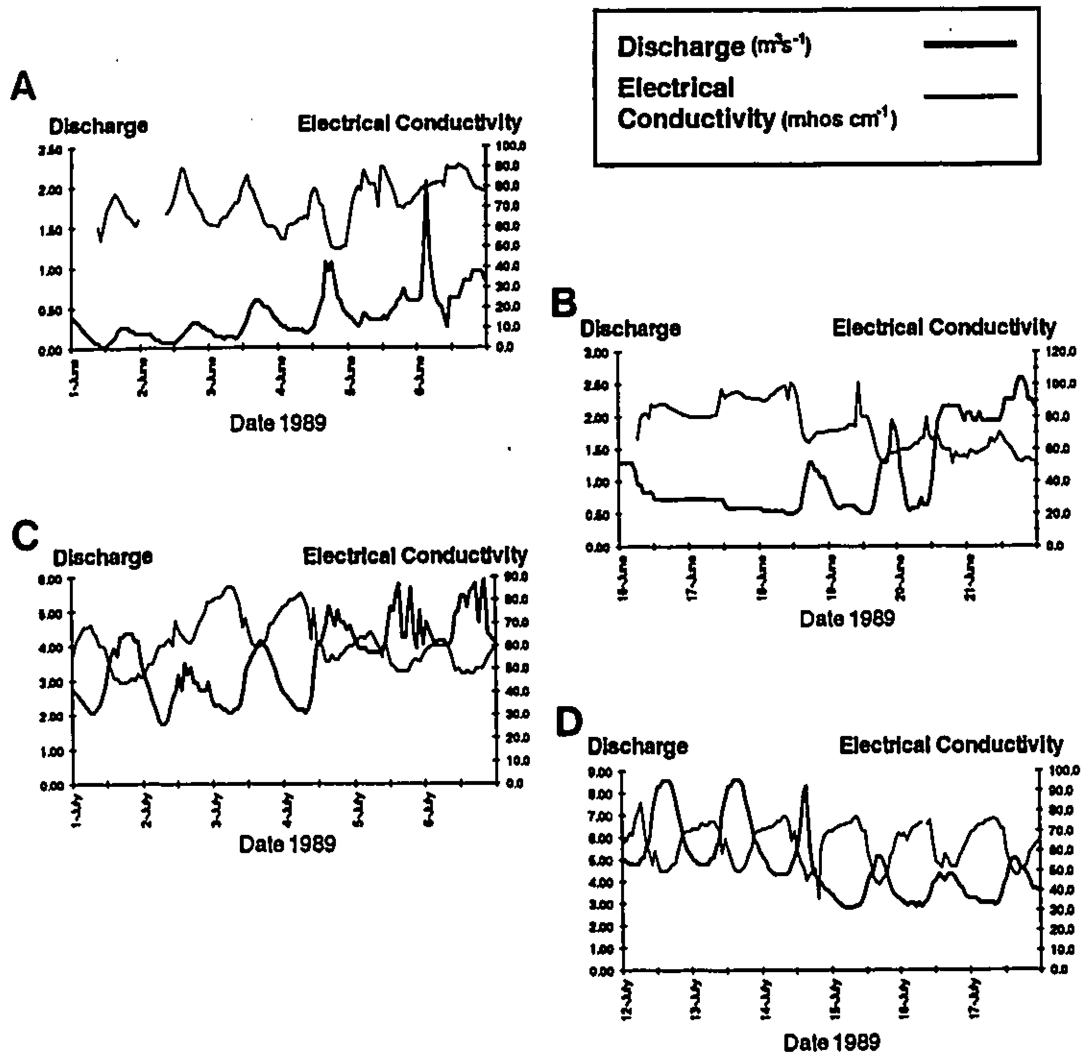


flow and decreased contact time that meltwaters have with the bedrock or sediments is minimal allowing for chemically dilute proglacial meltwaters.

Figure 5.14 shows diurnal and seasonal changes in the relationship between EC and discharge during the measurement period. During the measurement period EC and discharge remain well paired during diurnal fluctuations. As the relative contribution of water sources change from solute-rich, subglacially derived waters in early morning to solute-free, surface ice melt in early evening (Raiswell, 1984) EC and discharge maintain an inverse relationship. Short-term fluctuations of discharge and EC do, however interrupt the hydrograph aiding in the interpretation of the discharge record.

Figure 5.14a shows concurrent EC and discharge data for the period beginning 1 June and ending 6 June. During this 6 day period electrical conductivity values are inversely proportional to discharge with a lag 3-5 hours between maximums. As discharge flows begin to increase, conductivities drop and remain within a relatively constant range between 49 and 91 mhos cm^{-1}. Diurnal fluctuations in EC and discharge remain paired during this period of rising discharge only to be interrupted on a few occasions On 6 June a sudden increase in discharge did not result in the associated decline in EC as would be predicted. Electrical conductivity during 6 and 7 June was however, in phase with discharge. This implies that subglacial solute sources are slowly being accessed as meltwater flow increases. The period prior to 16 June, from 8 June to 15 June, exhibits an increase in EC values with gradually rising discharge levels. The culmination of this period is a maximum EC (101.0 mhos cm^{-1}) recorded on 18 and 19 June (10:00hrs and 12:00hrs)(Figure 5.14b). During the two previous days poor weather resulted in the decline of discharge levels with relatively no diurnal fluctuation, either in discharge or EC. The flushing of solutes on 18 and 19 June seem to suggest increased surface meltwaters had accessed subglacial water sources which have undergone chemical enrichment through contact with basal sediments. During this period as discharge levels gradually increase, increased basal flows under increased water

Figure 5.14: Electrical conductivity and discharge during four different sub-periods in Maxwell Creek, 1989.



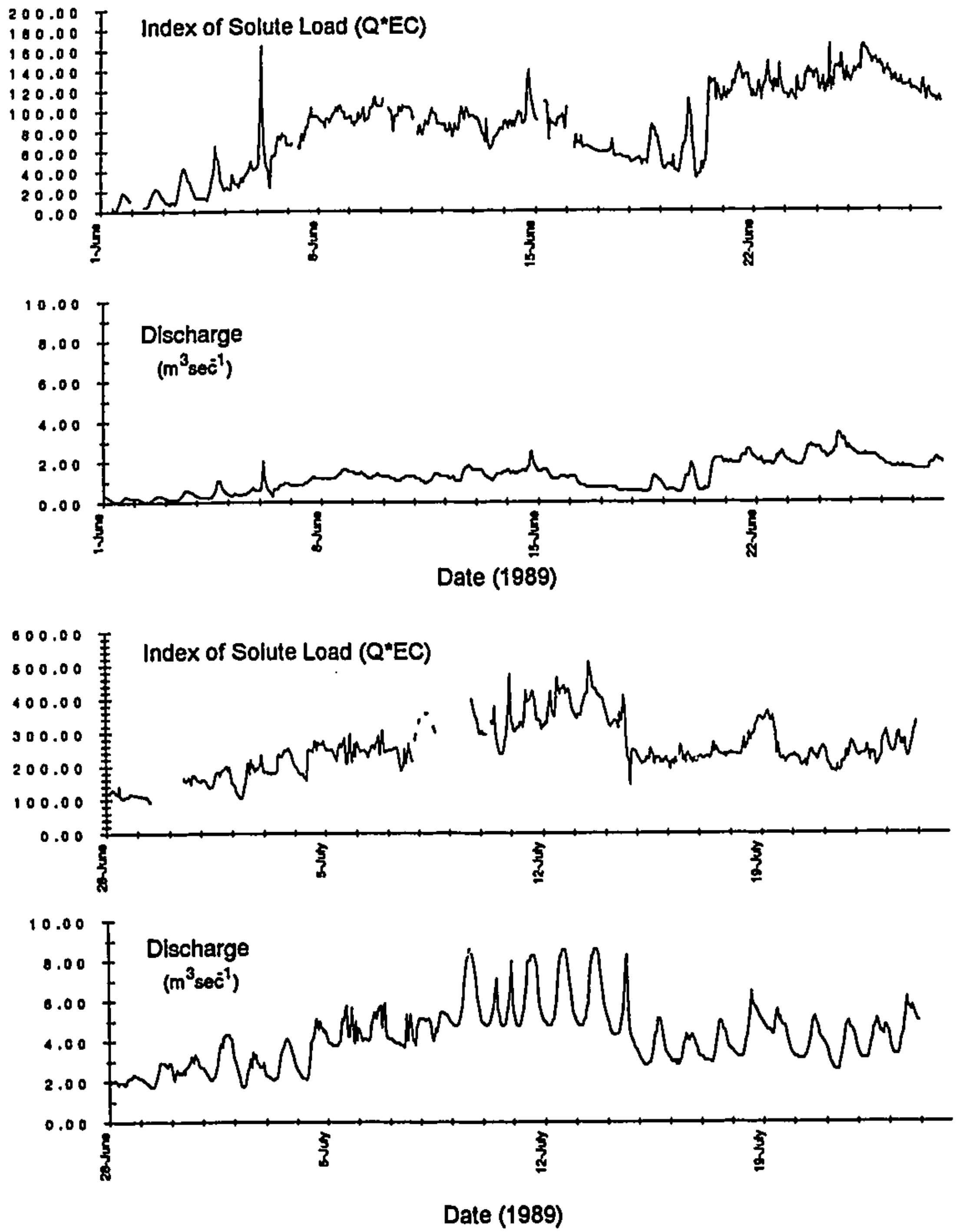
pressure, gain access to hydraulically isolated solute stores that have undergone chemical enrichment through contact with basal sediments and bedrock..

Conductivity data are marked by a general decrease in value as discharge levels gradually increase from 28 June to 9 July. The gradual reduction of the winter snow-cover on the glacier surface increased the contribution of ice-melt towards the total discharge resulting in the dilution of solute concentrations. The increasing discharge regime of 1 July to 4 July is characterised by an increase in average conductivity values during the same period. This indicated that the subglacial drainage system is again expanding as meltwater flow increases rapidly and becomes more variable. A maximum EC value of 86 mhos cm^{-1} (3 July, 07:00hrs) indicates the expansion of the drainage network further up glacier entraining solute rich meltwaters. During the following 3 days the average EC is relatively less (average ~ 5.58 mhos cm^{-1}) for similar discharge regimes. This indicates the short and rapid exhaustion of the solutes supply on the preceding 2 days.

During times of peak discharge EC (Figure 5.13 and 5.14d) electrical conductivity has a regular diurnal fluctuation which is in phase with discharge. The remainder of the season is characterised by fluctuating but ever decreasing levels of solutes in declining meltwaters due to the decreased enrichment of meltwaters within the subglacial system. The minimum EC value is not recorded during peak discharge rather during a period of declining flow at the end of the monitoring period (40.0 mhos cm^{-1}, 23 August, 17:00hrs). This implies that the residency time of meltwaters in subglacial stores has been significantly reduced and the enlargement or expansion of the subglacial drainage system has been halted. Figure 5.15 shows more clearly the exhaustion of sediment supply and rapid decline in drainage system advancement.

Figure 5.15 shows the index of solute load record for the measurement period. The index of solute load is not an indicator of total solute load for it not an absolute value. However, the index does show the varying rate of solute transport from a glacier. The

Figure 5.15: Index of solute load and discharge in Maxwell Creek, 1989.



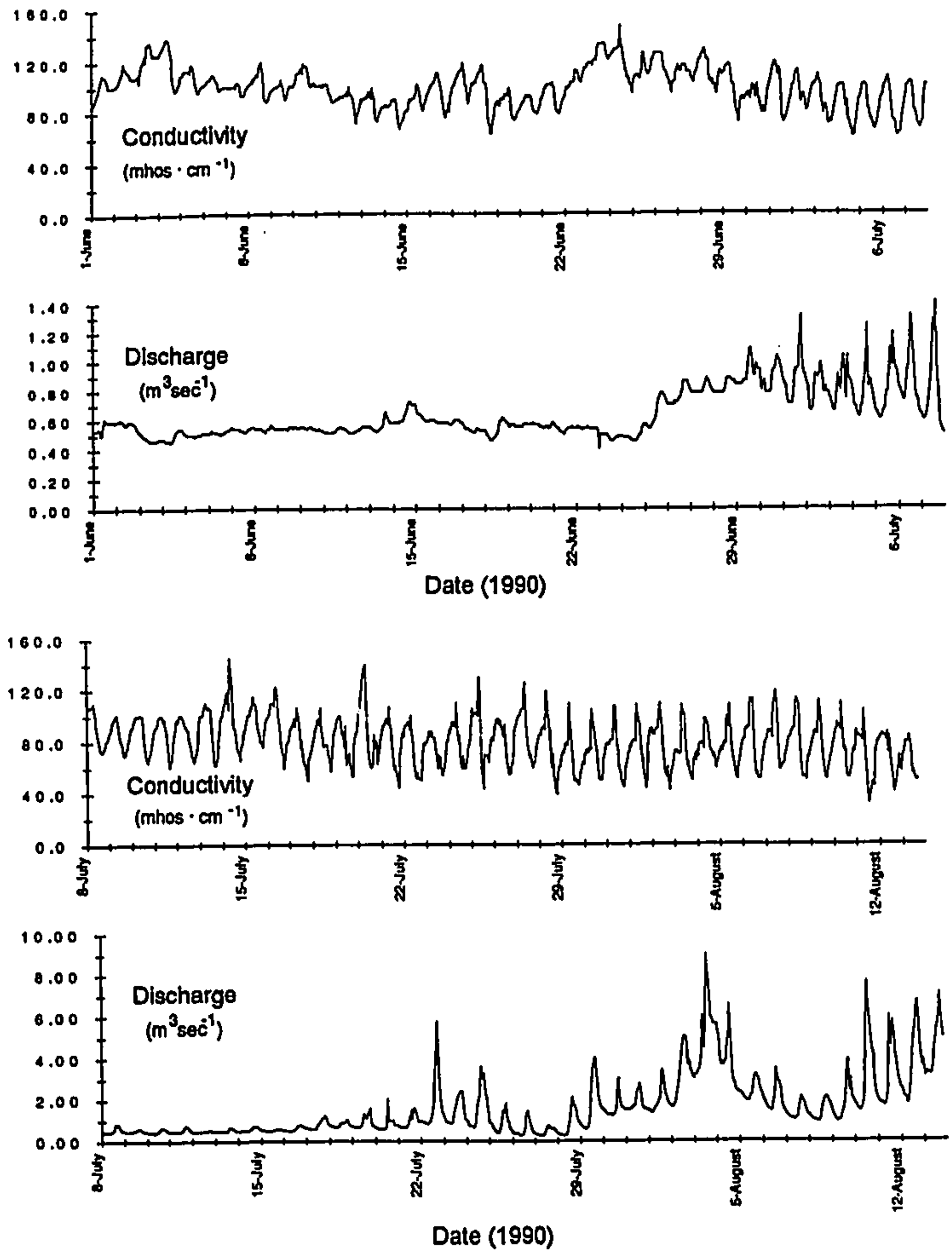
index of solute load is predominantly influenced by the base flow characteristics of discharge levels. Compared to electrical conductivity the base flow component of the index of solute load is more prominent. Increased solute removal occurs at times when discharge levels are increasing. Increasing discharge levels in the glacier drainage system results in the removal of solutes, and minimum solute levels occur at times of declining discharge levels. Maximum rates of solute transport occurring on 6, 14, 25 June, 13 and 19 July are followed by reductions in diurnal solute load peaks providing evidence for the flushing-exhaustion processes in glacier meltwater flow.

5.3.2 1990 Electrical Conductivity

Variations in total dissolved solids in the Maxwell Creek during the 1990 monitoring, are indicated by the continuous electrical conductivity time series in Figure 5.16. The conductivity trace, divided into periods, is constrained by two different water samplers that were used during the monitoring period. The first period, 1 June to 14 July, conductivity data were obtained from water samples taken an America Sigma automated sampler. On 15 July the central processing unit of the sampler failed resulting in its replacement with a Cygnus vacuum sampler. The replacement of one sampler with the other did not however alter the integrity of the trace of EC. During the 1990 monitoring period, 3 - 24hr. periods of 1hr. interval hydrochemical analysis was performed. Throughout each of these three periods, at the beginning, in the middle at the end of the season, hourly samples of meltwater were obtained and hydrochemical analysis performed. The results of this analysis are to be discussed later in section 5.4 *'Variations in Ionic Composition'*.

The EC and discharge traces in Figure 5.16 are inversely proportional during the monitoring period. The entire data set is characterised by maximal electrical conductivity

Figure 5.16: Instantaneous electrical conductivity and discharge in Maxwell Creek, 1990.



occurring when discharge levels are minimal. Solute concentrations are contained within a broad conductivity range of 32.0 to 146.0 mhos cm^{-1}. The average electrical conductivity of all data is 89.43 mhos cm^{-1} with a standard deviation of 19.43 mhos cm^{-1}. On 24 June (12:00hrs), a seasonal maximum solute concentration of 146 mhos cm^{-1} was synchronous with low flow conditions. A seasonal minima value of 32 mhos cm^{-1} at 14:00hrs was recorded on 11 August. Data appear to be relatively normally distributed on a daily basis. The EC record exhibits a relatively high quality time series with no missing data.

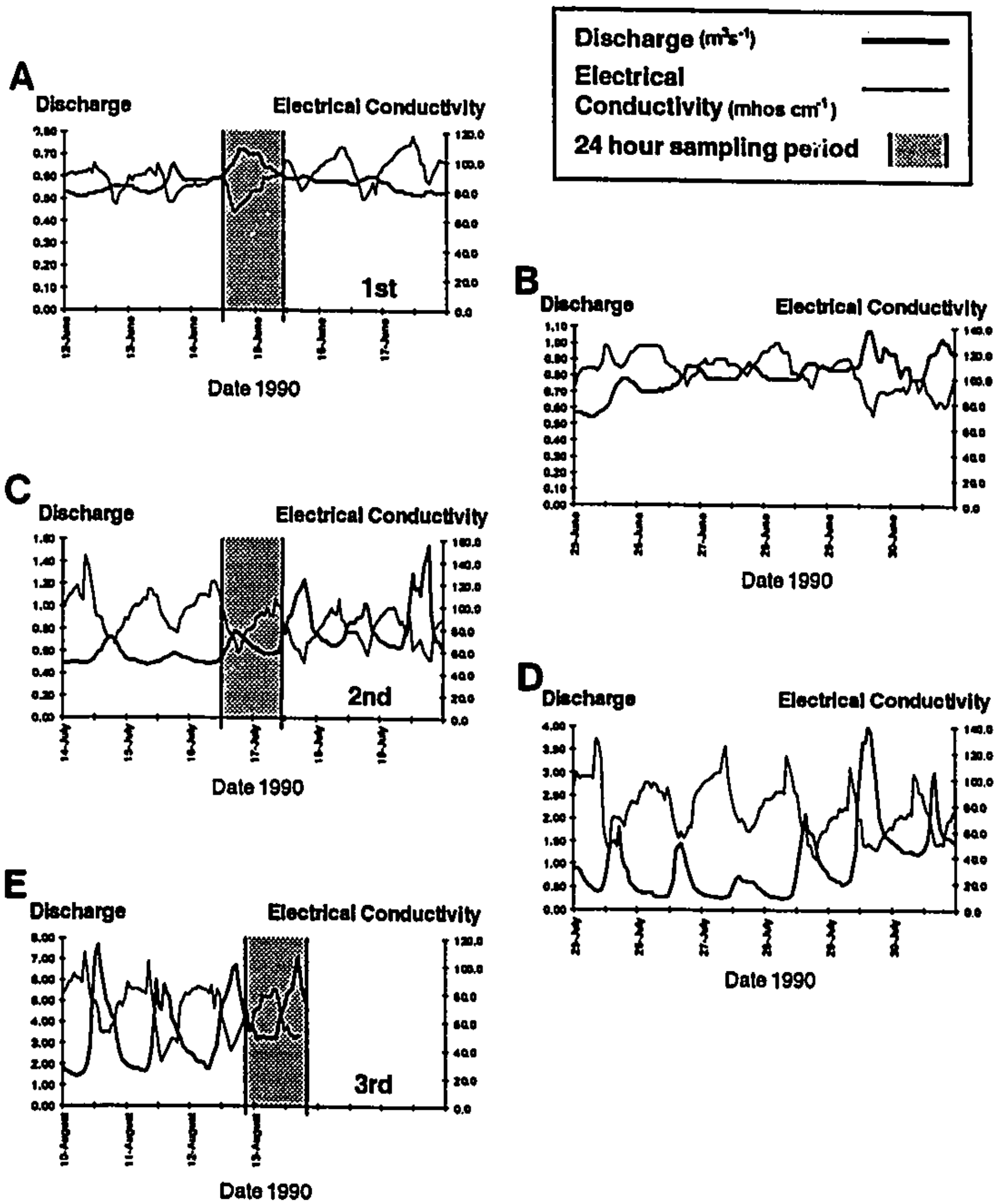
Examination of data shows that the daily trend in EC is generally inversely proportional with discharge data. The reason for this is that minimum concentrations are associated with rising flows of fast routed, relatively solute-free surface meltwaters. The diurnal fluctuations in the conductivity record transform from being symmetrical from the beginning of the measurement period gradually appearing asymmetrical on approximately 30 June with relatively gently rising and sharply falling limbs. The reason for the symmetrical trace in EC during the beginning of the measurement period is that following discharge maxima, flows gradually decline producing a slow build up of concentration in meltwaters. The symmetry in the trace is a result of meltwater predominantly derived from supraglacial snow-pack and firn melt. Commencing approximately on 30 June, an increased contribution and routing of meltwaters from the surface into englacial and subglacial passageways access englacial sediment rich bands and subglacial sediment stores. Conductivity is maximal when meltwaters are predominantly derived from relatively slow flowing subglacial flows which have sufficient time to dissolve bedrock and sediments at the ice-rock interface. At maximum incoming radiation, melting of surface ice contributes solute poor meltwaters which are quickly drained through the Maxwell Glacier into the proglacial stream. Hence, the subglacial meltwaters, which are in equilibrium with mineral particles, are mixed with solute-poor supraglacial meltwaters resulting in lower proglacial solute concentrations in meltwaters.

Figure 5.17 is a plot of EC and discharge variation at a smaller time scale illustrating the seasonal development and influence of hydrometeorological conditions over the 1990 monitoring period. Viewing the variations in Figure 5.17 illustrates that minor fluctuations in solute concentration occur throughout the monitoring with varying proportions to discharge. For example, on the rising limb of 25 June, when flow is rising to its peak at 19:00hrs, conductivity reaches its peak 6 hours before and then reaches a minimum at time of peak discharge. This spike on the rising limb of the hydrograph is possibly associated to an inclusion of subglacially stored water.

The record of solute concentration on the discharge recession limb on the morning of 14 July provides another example where the record of solute concentration aids in the interpretation of discharge time series. On the recession limb, during the morning of 14 July, a sudden decline in EC occurs with no noticeable change in discharge. The pulse in EC is followed by a large increase in concentration to levels exceeding the previous maximum. It is possible that this sudden decline and peak, which is accompanied by no change in discharge, is caused by the inclusion of a hydraulically isolated cavity of solute-rich meltwater suddenly incorporated into the subglacial drainage system.

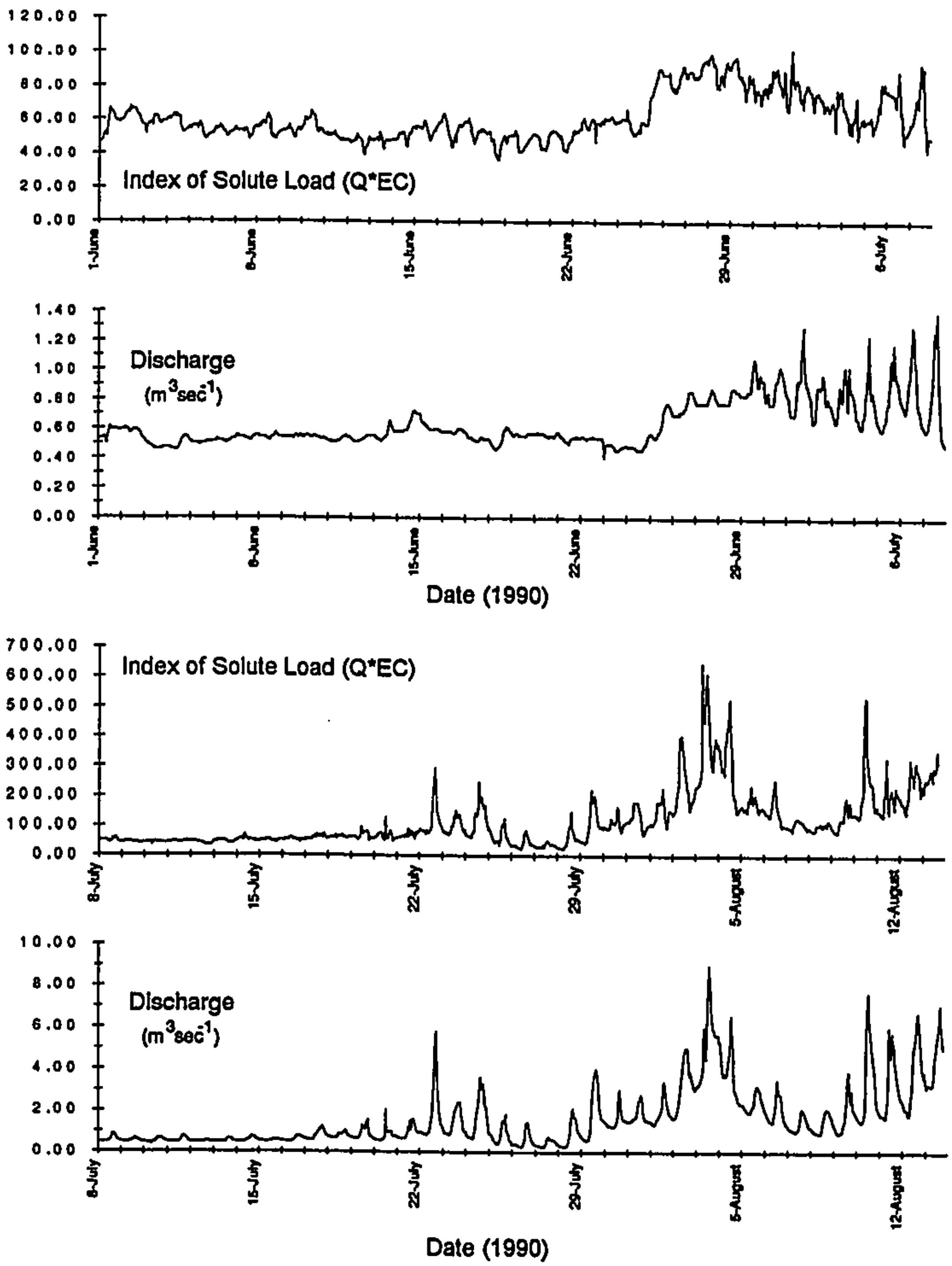
Although minor fluctuations in the 1990 EC trace do infrequently occur, diurnal variations in solute concentration are closely related to changing discharges. As the 1990 monitoring season progressed the average EC value declined with increasing flows. Collins (1977, 1979b) indicated that a poor relationship between discharge and EC is a result of solute-rich meltwaters being progressively accessed over consecutive hydrographs. Hence, solute sources become exhausted during the ablation season unless new stores are continually being accessed. The plot of EC and discharge, however, do not concur with this hypothesis. A plot of the index of solute load and discharge in Maxwell Creek, 1990 (Figure 5.18), more clearly shows such flushing and exhaustion of solute-rich meltwater supplies.

Figure 5.17: Electrical conductivity and discharge during five different sub-periods in Maxwell Creek, 1990.



The varying rate of solute transport in the Maxwell Creek during the measurement period is shown by the index of solute load time series plot in Figure 5.18. The 1990 index of solute load clearly indicates periods of varying duration and form of solute transport in the proglacial stream. Maximum rates of transport on 5, 6 , 7 and 30 July, 3, 12 and 14 August are clearly followed by reductions in diurnal solute load peaks on subsequent days. Therefore, the same flushing-exhaustion mechanisms operative in 1989 are also operative in 1990. The 1990 index of solute trace also reflects changes in the form and stability of the drainage network. The segments in the time series from 5 to 7 July, 30 July to 5 August contain periods of increasing solute load throughout concurrent periods of increasing and declining discharge. Such patterns of index of solute load instability may be the result of increasing basal hydrostatic pressures increasing bed separation and release of stored solutes. During the period of low fall on the morning of 30 July a short pulse in EC occurred with no change in discharge flow. Over the next couple hours the pulse declined to its former level only to increase as discharge declined. The first pulse in the index of solute record is possibly in response to a sudden short blockage of subglacial meltwater flow. As meltwater was temporarily blocked, solutes may have been able to interact with sediments and accumulate in meltwaters to be later released in flow. A similar event occurred on the recession flow limb on 24 July as the drainage system was enlarging and expanding. However, these observations must be treated with care since not only is the relationship between discharge and solute load uncertain by definition, but the role of major and minor ions in the composition of EC is irresolute. The nature and controls of solute acquisition by water-rock interactions in glacial system vary thoughout the system in time and space. Hence, the study of EC and its constituents may contribute to a chemical model of solute acquisition within a glacial system.

Figure 5.18: Index of solute load and discharge in Maxwell Creek, 1990.



5.4 VARIATIONS IN IONIC COMPOSITION

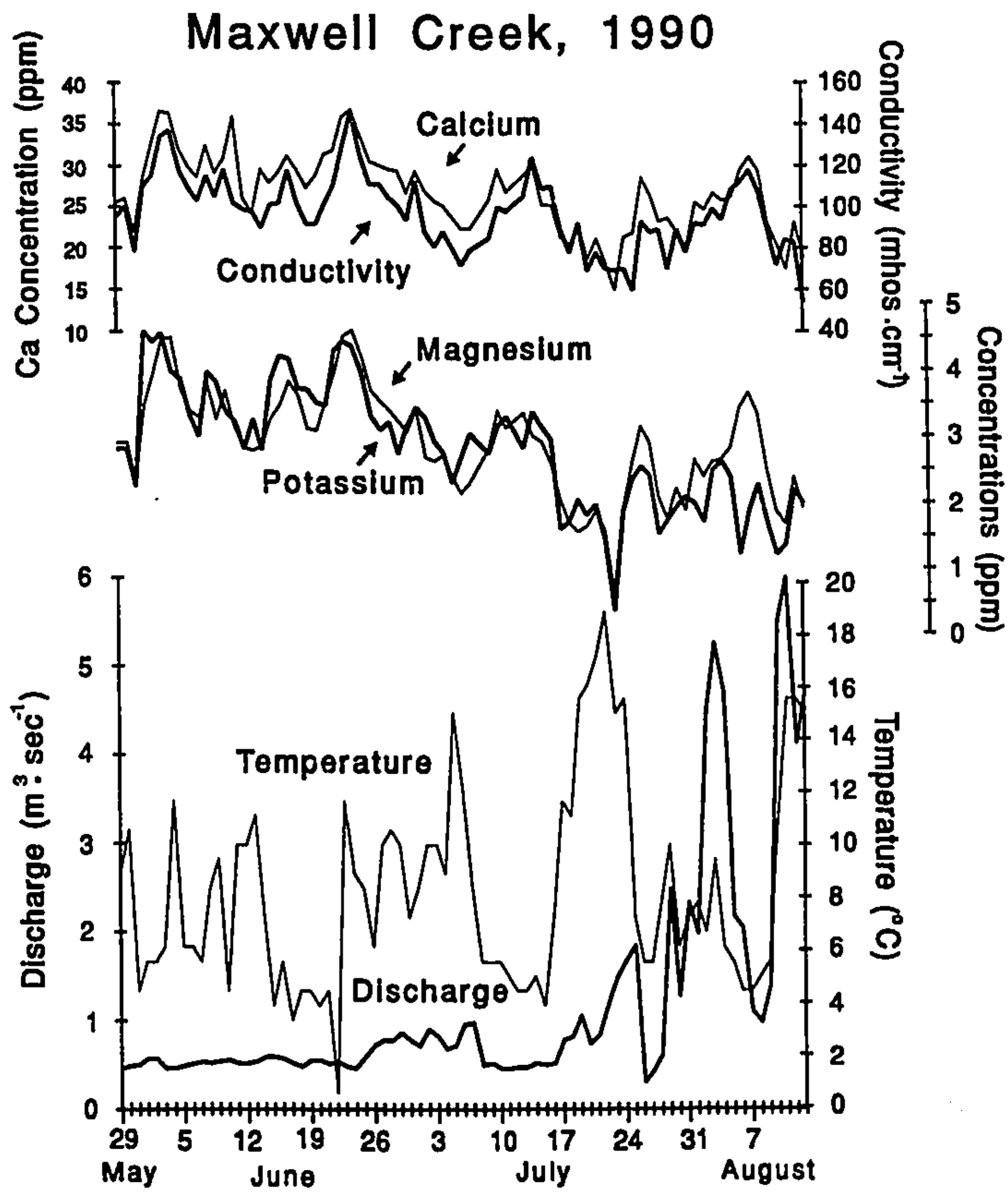
Modelling the drainage and transport of both sediments and solutes beneath a glacier necessitates monitoring the changing meteorological and glaciological conditions which act upon the system and the products of such a system (Østrem, 1975; Collins, 1989). Inferences made from the relationships between meteorological and hydro-glaciological parameters are used to derive scenarios for sediment acquisition and the transport within a glacial system.

During passage through the subglacial drainage system meltwater acquires chemical and sedimentological characteristic's indicative of source and routing. This results from subglacial routing of meltwater providing meltwater access to basal sediments and bedrock. These meltwaters undergo changes in composition as they mix with water from different sources. Hence, identification and analysis of individual cations may provide clarification of subglacial meltwater source, transfer and routing.

5.4.1 Intra-Seasonal Variation in Ionic Composition, 1990

Temporal variations in discharge, temperature and dissolved solute concentrations for Maxwell Creek from 29 May to 13 August 1990 are plotted in Figure 5.19. Existing chemical data permit useful limits to be placed on the range of composition variation in glacial meltwaters. In this study the examination of four major cations; sodium (Na^{+}), potassium (K^{+}), calcium (Ca^{+2}), and magnesium (Mg^{+2}), are discussed. Relationships between the different dissolved species in the meltwaters are expressed in Figures 5.19 through to Figure 5.27. The different meltwaters show relatively little variation in cation composition (Ca^{+2} 71.8-80.4%; K^{+} 3.8-11.3%; Na^{+} (1.9-9.3%; and Mg^{+2} 3.5-9.6%).

Figure 5.19: Instantaneous concentrations of Ca^{+2}, Mg^{+2}, and K^{+}, with paired instantaneous measurements of temperature, discharge and electrical conductivity in Maxwell Creek, 1990.



Variations in cation composition show no consistent relationship with discharge. Higher dissolved loads (measured by the EC and the sum of dissolved cations) occur in June than in August. Calcium is the dominant dissolved cation in the waters examined ranging from 71.8 to 80.4 % of the sum of the four dissolved cations analysed. The relative variations in the four dissolved cations is be shown by the absolute and relative value composition plots 5.20-5.27. Figure 5.20 and Figure 5.21 were computed from samples collected once per day throughout the measurement period. Figure 5.20 is a plot of the absolute values of four dissolved cations with paired discharge (Q), suspended sediment concentration (SSC) and electrical conductivity (EC) showing the seasonal trend in cation variations. Figure 5.21 contains the same absolute Q, SSC and EC data as Figure 5.20; however, the dissolved cation curves show the relative variation in composition of each cation expressed as a percentage of the total cation sum. Raiswell (1984) indicates that there have been chemical studies of the glacial meltwater, but relatively few of them have directed their attention towards identifying the nature and controls of solute acquisition. Figures 5.22 - 5.27 are cation sum and percentage diagrams of three, twenty-four hour periods in which chemical analysis was performed hourly. Each set of figures, 5.22 and 5.23, 5.24 and 5.25, and 5.26 and 5.27 respectively represent the beginning, middle and end of the monitoring period. Each pair of cation diagrams is of the same format as Figures 5.20 and 5.21, with the exception that temperature data has been included. Stratigraphically constrained cluster analysis has been performed on each of the data sets. A dendrogram is attached to each pair of figures illustrating the hierarchical relationships of the adjacent clusters as defined by analysis performed by CONISS, a program of constrained incremental sum of squares cluster analysis, by Eric C. Grimm (1987). Grimm's (1987) sum of squares cluster analysis is designed to minimise the total within-cluster dispersion for geological zones grouped around centroids for each zone. In the constrained analysis, only stratigraphically adjacent clusters are considered. The goal of applying CONISS to the data sets is to delimit temporal zones of the most reasonably similar physical and chemical clusters in within-zone variability or dispersion is minimised.

Examination of the seasonal concentrations of dissolved solutes shows that the trend in solutes is generally inphase and inversely proportional to discharge and SSC. The cation concentration of the meltstream is low since the majority of the water is of supraglacier water. The cationic composition of meltwater from Maxwell Creek is 1.9-9.3% Na^+, 3.8-11.3% K^+, 3.5-9.6% Mg^{+2} and 71.8-80.4% Ca^{+2} (Figure 5.20). Higher discharges are characterised by a lower concentration in the four cations. During the monitoring period Ca^{+2} remains the dominant dissolved cation in meltwaters. Figure 5.20 shows that Ca^{+2} concentrations generally decline as flow increases. Throughout the monitoring period Mg^+, K^+ and Na^+ concentrations remain in phase with Ca^{+2} with the exception that the amount of variation within these cations is considerably less.

Inspection of the temporally constrained dendrogram suggests that data can be divided into four zones of development. The first zone is from 1 June to 30 June, second is 1 July to 14 July, third is 15 July to 4 August and the final zone is from 5 to 13 August. The zonation within the dendrogram is closely related to the trend in increasing flow levels and associated changes in meltwater composition. However, with closer inspection of the relative contribution of each cation to the total cation content it is evident that the relationship is more complicated. Figure 5.21 shows that the relationship between discharge and meltwater composition is complex. The record shows that the range in cations throughout the monitoring period is constant. This may indicate the continued access to solute stores and incorporation into flow. The Ca^{+2} trace indicates that as discharge levels rose throughout the period the amount of Ca^{+2} increased with greater proportion. Analysis of the cation content during the monitoring period is not characterised by simple rising and falling concentrations. During the period of increasing discharge from 21 July to 3 August, the absolute values of cation concentration generally follow a similar pattern to discharge.

Figure 5.20: Instantaneous concentrations of Na^+, K^+, Ca^{+2} and Mg^{+2} with paired instantaneous measurements of discharge, suspended sediment concentration and electrical conductivity in Maxwell Creek, 1990.

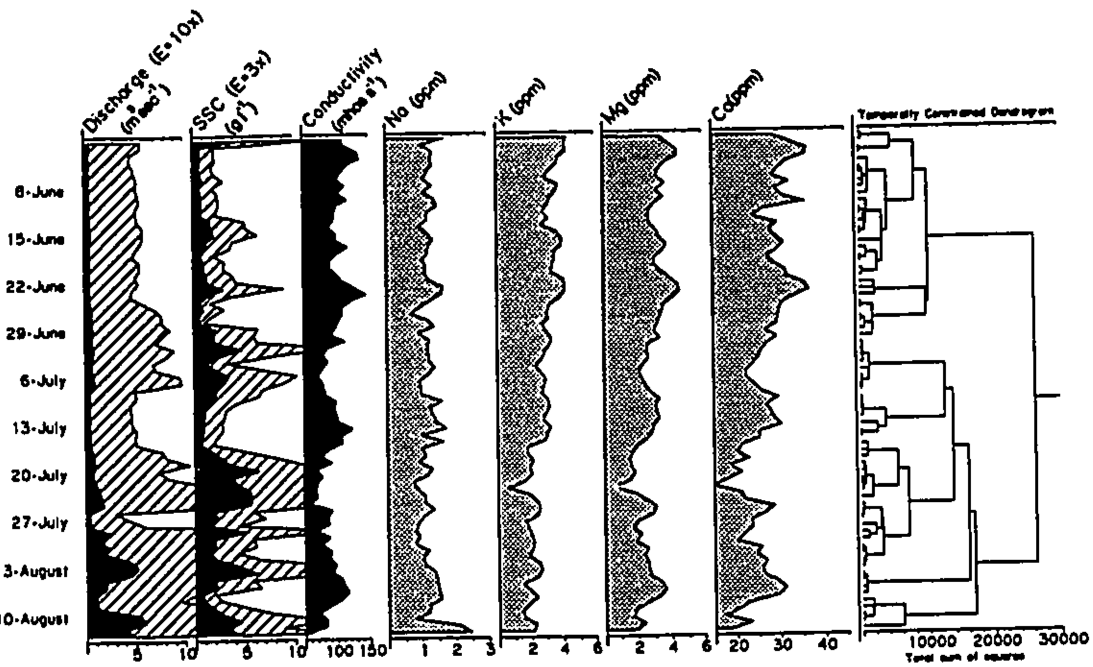
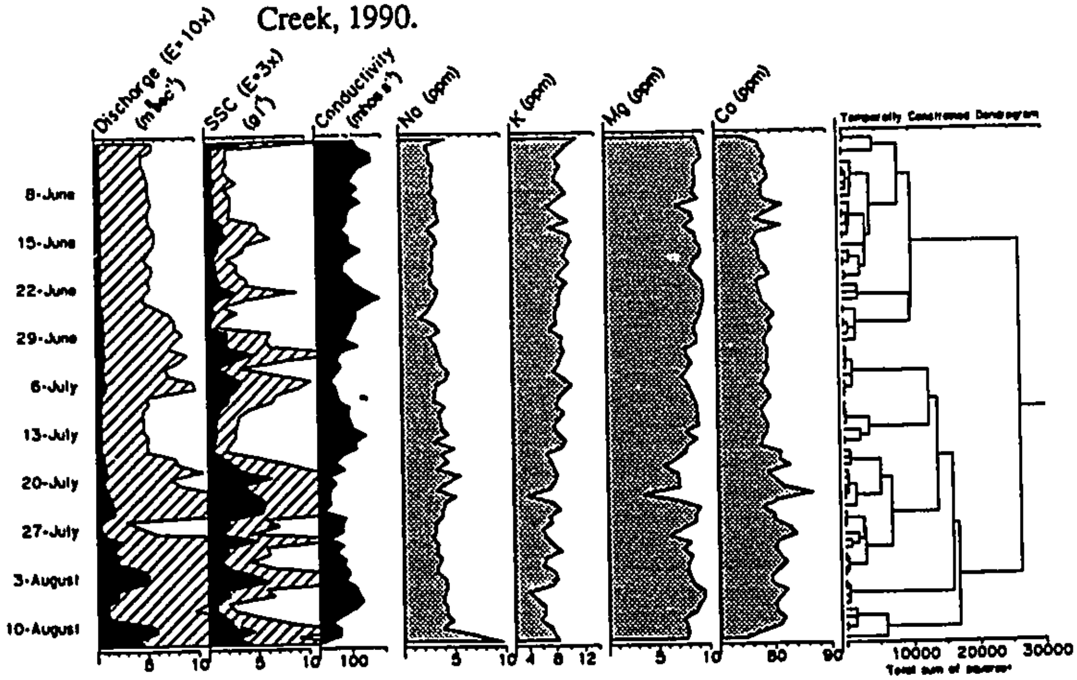


Figure 5.21: Na^+, K^+, Ca^{+2} and Mg^{+2} percentage composition of total solute load with paired instantaneous measurements of discharge, suspended sediment concentration and electrical conductivity in Maxwell Creek, 1990.



Close inspection of the relative composition and contribution of each cation to the total sum indicates that short-term variations in composition are present. It is probable that these variations in cation species may be indicative of a changing source of solute and or duration in transport. This implies that contributions to flow from varying sources throughout the monitoring period may temper or accentuate hydrochemical characteristics of meltwaters draining the Maxwell Glacier. This variation in chemical composition from within the glacier is shown more clearly in the record of the three 24hr hydrochemical periods.

During the 1990 monitoring period three periods, at the beginning, middle and end of the season, of short interval hydrochemical analysis was performed. Figure 5.22 through to and including 5.27 show the relative variations in the four dissolved cations, SSC, Q, temperature and EC during the three hydrochemical sampling periods. These curves indicate dissolved solute contents are closely related to EC and inversely proportional to discharge. The plots of dissolved cation composition reveals more than the paired curves of EC. The diurnal solute load regime is perturbed during both rising and falling limbs of the hydrograph. During the monitoring period it is apparent that in the beginning of the season that the contribution of Ca^{+2} to the composition of total dissolved solids (TDS) is greater and decreases toward the end. The different contribution of Ca^{+2} to the TDS may represent the duration of interaction between meltwater - sediment - bedrock or it may simply reflect a greater abundance of Ca^{+2} bearing soluble source rocks in the area. The other dissolved cations do not show as much variation in composition or contribution to meltwater during the monitoring period. During the monitoring period of 15 June the main variation is in Ca^{+2} (49.4-79.2%), Mg^{+2} (5.5-8.5%), K^{+} (6.2-11.0%) and Na^{+} relatively constant (2.6-4.8%). Variations in cations during the period of 17 July remain relatively constant for each species Ca^{+2} (75.1-79.9%), Mg^{+2} (6.6-8.8%), K^{+} (6.1-8.9%) and Na^{+} (3.5-5.6%). The period of 13 August is similar to the period 17 July with a slight more variation amongst and within species, with Ca^{+2} (71.2-76.4%) and Na^{+} (4.1-9.3%) increasing slightly in variation compared with Mg^{+2} (6.3-9.5%) and K^{+} (6.3-10.1%). In general the variations in cation composition appears to show no consistent relationship with discharge and is demonstrably more complex than the simple EC trace.

Relationships between the different dissolved solutes species in meltwaters are expressed in three correlation matrices (Figure 5.28), which provides an indication of the sources of each element. Calcium, potassium and magnesium are fairly well correlated throughout the three periods as a result of moderately calcareous siltstone and sandstone below the glacier, as they are more prone to chemical weathering.. Sodium is however poorly related to potassium. These relationships are consistent with a dominant carbonate source. However, a combination of limited number of sampling periods and errors involved in analysis (5-15%) is insufficient to fully explain the seasonal deviations in hydrochemical properties of meltwaters. Inorder to fully explain variations in ionic composition it is necessary to have a complete knowledge of the underlying composition of the material that the water is in contact with, and its solubility with respect to numerous chemical and mechanical weathering processes.

Figure 5.22: Instantaneous concentrations of Na^+, K^+, Ca^{+2} and Mg^{+2} with paired instantaneous measurements of temperature, discharge, suspended sediment concentration and electrical conductivity in Maxwell Creek, 15 June, 1990.

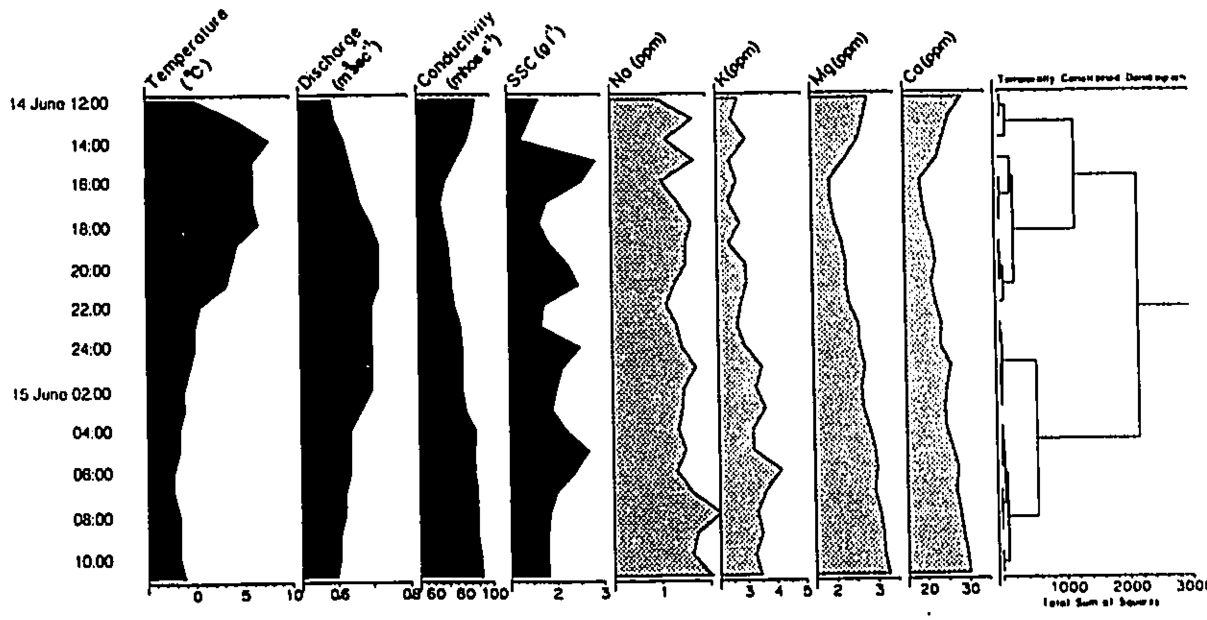


Figure 5.23: Na^+, K^+, Ca^{+2} and Mg^{+2} percentage composition of total solute load with paired instantaneous measurements of temperature, discharge, electrical conductivity and suspended sediment load in Maxwell Creek, 15 June, 1990.

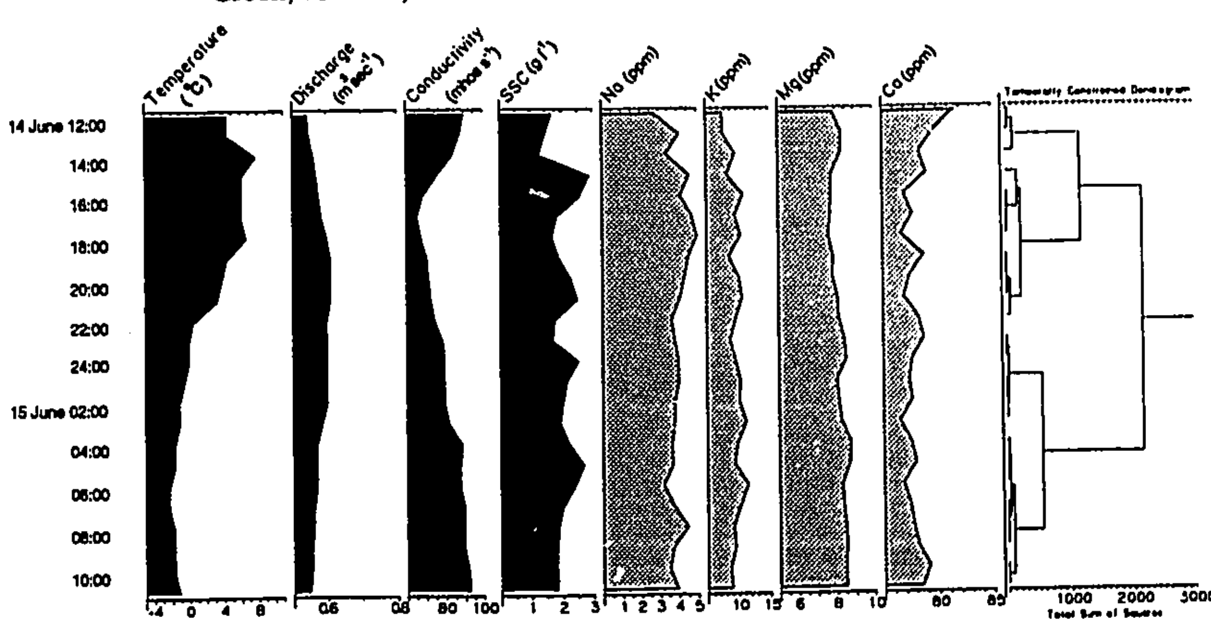


Figure 5.24: Instantaneous concentrations of Na^{+}, K^{+}, Ca^{+2} and Mg^{+2} with paired instantaneous measurements of temperature, discharge, suspended sediment concentration and electrical conductivity in Maxwell Creek, 17 July, 1990.

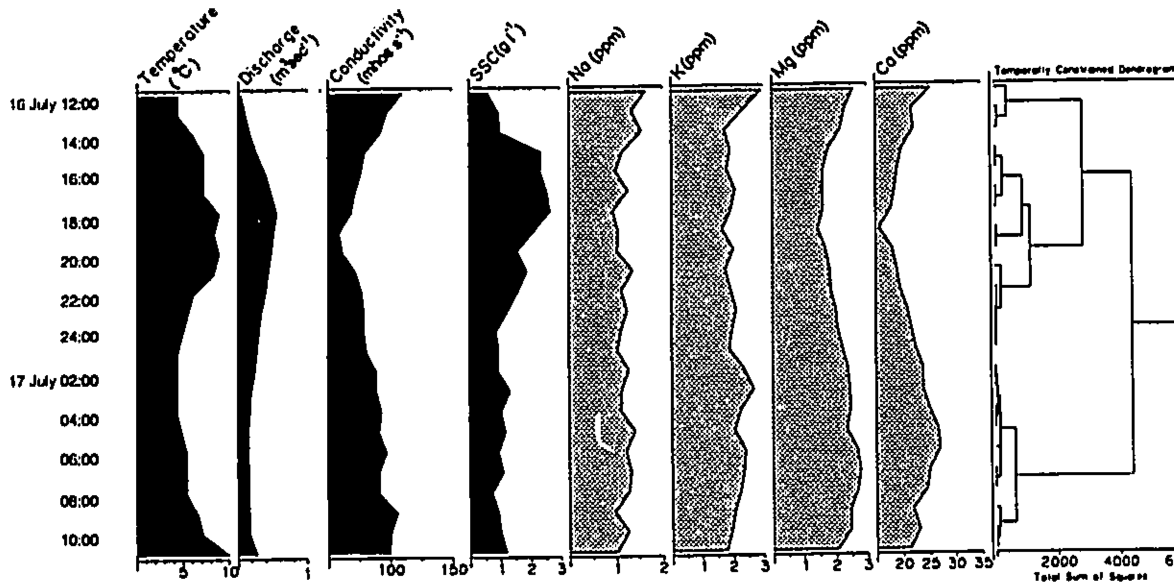


Figure 5.25: Na^{+}, K^{+}, Ca^{+2} and Mg^{+2} percentage composition of total solute load with paired instantaneous measurements of temperature, discharge, electrical conductivity and suspended sediment load in Maxwell Creek, 17 July, 1990.

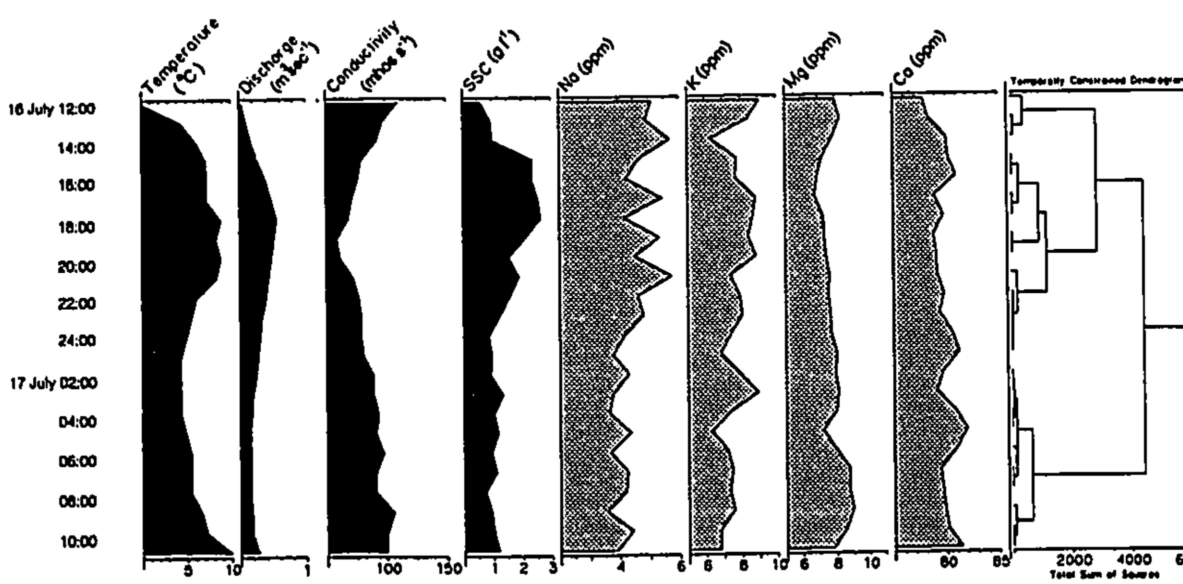


Figure 5.26: Instantaneous concentrations of Na^{+}, K^{+}, Ca^{+2} and Mg^{+2}, with paired instantaneous measurements of temperature, discharge, suspended sediment concentration and electrical conductivity in Maxwell Creek, 13 August, 1990.

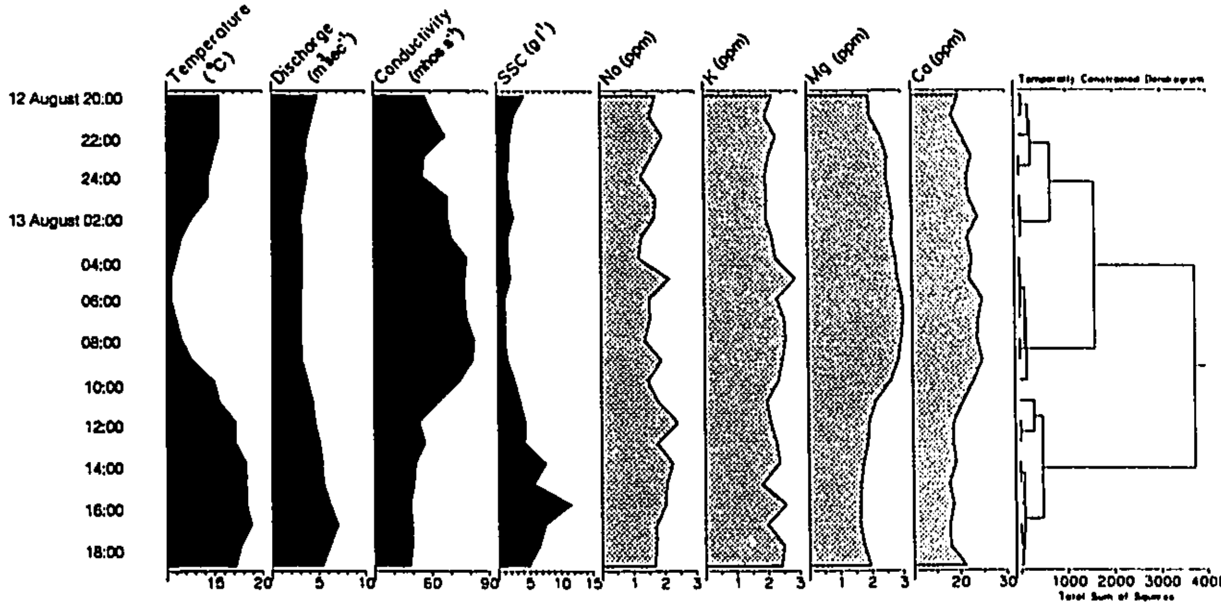


Figure 5.27: Na^{+}, K^{+}, Ca^{+2} and Mg^{+2} percentage composition of total solute load with paired instantaneous measurements of temperature, discharge, electrical conductivity and suspended sediment load in Maxwell Creek, 13 August, 1990.

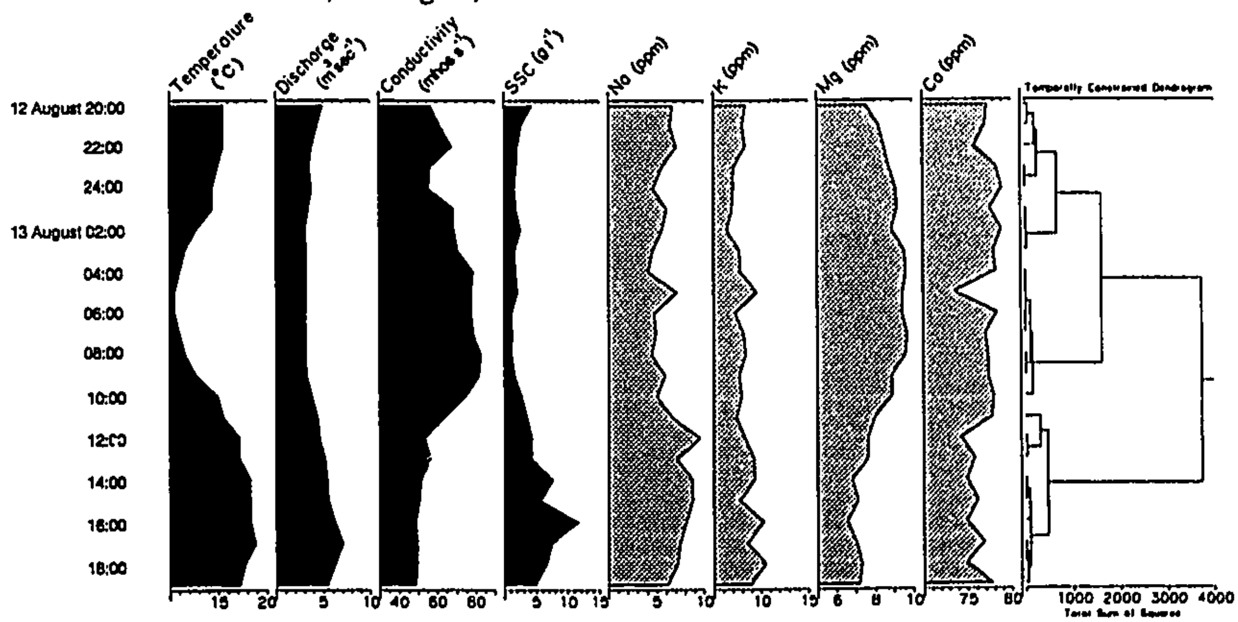


Figure 5.28: Correlation matrices for the 3-24hr. hydrochemical sampling periods for Ca^{+2}, Mg^{+2}, Na^{+} and K^{+} composition.

Figure 5.28a. Correlation coefficients, Maxwell Creek, 15 June, 1990.

	Na	K	Ca	Mg
Na	-	0.22	0.38	0.41
K		-	0.59	0.68
Ca			-	0.96
Mg				-

Figure 5.28b. Correlation coefficients, Maxwell Creek, 17 July, 1990.

	Na	K	Ca	Mg
Na	-	0.41	0.47	0.42
K		-	0.64	0.66
Ca			-	0.92
Mg				-

Figure 5.28c. Correlation coefficients, Maxwell Creek, 13 August, 1990.

	Na	K	Ca	Mg
Na	-	0.13	-0.45	-0.49
K		-	0.27	0.28
Ca			-	0.94
Mg				-

CHAPTER 6

DISCUSSION AND CONCLUSIONS

6.1 DISCUSSION

6.1.1 Seasonal Variations of Meltwater Supply

The hydrographs for 1989 and 1990 illustrate the differences between two hydrographs controlled primarily by contrasting weather conditions. The normal glacierised-basin hydrograph of 1989 has a high annual total discharge because a large area of ice had become exposed to melting early in the season, and remained uncovered by snow throughout the summer. In addition, skies were predominately clear. The 1990 hydrograph had lower flow values in spring and summer due to cool, overcast periods that brought precipitation and reduced energy availability. After the snowmelt peak, the 1989 and 1990 run-off regimes responded to climatic conditions influencing ablation. This is seen by pairing the discharge curves with mean temperature curves for both monitoring periods. Peak discharge occurred a few hours after peak melt, but as summer advanced, the daily rise and fall in discharge became more rapid and the lag time decreased. Total daily discharge reached its maximum in late July early August. When melting was halted by late mid-summer snowfalls, the diurnal variation in discharge declines. When melting began again, base flow takes between 12-24hrs to reach its former level. In general, a snowpack retains rainfall and meltwater in early spring, whereas a bare ice or firn surface may channelise runoff of rainfall producing aperiodic variations in runoff. The low values of incoming solar radiation resulted in the slow increase of discharge until 1 August when a warm and sunny period produced an almost instantaneous rise in discharge to a peak of 8.98 m^3s^{-1} on 3 August 1990.

In 1989 a warm June and July were infrequently interrupted by snow and rainfall. This contrasts with 1990 when a cool June and early July was followed by a warm mid-July which lead into variable conditions throughout the remainder of the field season. The precipitation regimes for the two years greatly contrast. In 1989, a dry and sunny July was preceded by sporadic precipitation events not lasting more than a few hours. In contrast, June and the early part of July 1990 were typified by cloudy days, occasional showers and a heavy snowfall on June 23. Warm, clear weather during the remainder of July 1990 raised flow to their highest levels. During both years, cloudy periods before and during precipitation events were associated with a decrease in discharge, after which daily discharge returned to higher than previously recorded levels. Heavy smoke from forest fires in the southern Yukon retarded ablation during the period commencing July 7 through to 16 July, 1990, resulting in decreased discharge.

6.1.2 Sediment Supply to Proglacial Meltwaters

A comparison of the 1989 and 1990 seasons in the Maxwell tributary provides an illustration of suspended sediment variability in a small glacierised basin. The seasonal variation in suspended sediment flux of the Maxwell Glacier in 1989 and 1990 broadly followed the pattern of fluid discharge, but with marked short-term variation. During the 1989 monitoring period, a high winter snow accumulation with a late spring melt was followed by warm and dry conditions through the late spring and summer. Snowmelt on lateral moraines produced surface runoff, with suspended sediment concentration <1,000 mgL^{-1}, and subsurface and ground water flow which did not initiate mudflow or debris flow activity. By late spring there was little subsurface runoff and, during the dry summer, discharge only occurred as seepage at the base of some of the slopes, producing some localised mudflows. The glacier drainage system opened progressively from late spring through the summer with initial highly variable suspended sediment loadings followed by an early start to the melt season and unsettled weather during the early summer. Initial spring runoff had low sediment loads. Rainfall in the late spring induced rapid runoff with slushflows and mudflows occurring along the moraines as morainic deposits became saturated. Rapid melt, following the 30cm snowstorm of 22 June produced numerous mudflows on the moraines and gravity deposits. The glacier drainage system opened rapidly due to rainfall in the spring and the suspended sediment regimes were not as variable, and peak loads were lower than in 1989. In 1990 slush flows off the glacier margins produced periods of enhanced suspended sediment generation. One slush flow on the south margin of the glacier terminus mobilised a large section of the terminus snow wedge transporting in excess of 10,000 m^3 of snow over 200 m^2 in area. The melt of this slush and snow left a veneer of fine gravel, sand and silt over 5,000 m^2 area of the moraine.

Integration of suspended sediment with flowing meltwaters in subglacial areas was dependent of the rate of meltwater supply and differing subglacial hydrological events. During late spring prior to maximum flow conditions, the sediment record was increasingly more variable with large pulses of sediment. These pulses suggest the continued and progressive development of cavity linkage laterally and up-glacier. The second large pulse of sediment with rising discharge could be associated with the enlargement of a linked cavity system as basal pressures increased with flow spreading laterally across the bed. Areal integration of the glacier bed with flowing meltwater depends on the type of drainage system and how well developed is the system. During periods of declining flow, frictional melting of ice from the basal conduits will decline hence contributing less sediment. During these periods conduits will gradually fill with sediment. Increased discharge will then produce sediment pulses followed by a moderate decline. It is because of these periodic increases in fluid discharge that a condition of continuous sediment supply is unlikely. Suspended sediments continually supplied to meltwaters will be channelised and transported to the portal. Throughout both ablation seasons melting and deformation of the basal drainage system continually introduced new sediments episodically into the flow by allowing meltwater to access previously isolated sediments. Thus the rate of sediment supply to basal meltwaters bears no constant relationship between ablation periods.

Supraglacial runoff from late spring snow melt does not produce an increased suspended sediment delivery except when sediment is available to the streams near the glacier terminus. Englacial and subglacial conduits open in early summer and cavities are connected to the conduits and drained (Hooke, 1989). Fluid discharge and suspended sediment are variable, but are not necessarily correlated. This can be explained by the nature of the conduit/linked cavity system (Johnson, 1990a). Changes in the subglacial drainage network may exploit new sediment sources without any change in discharge. Peaks of up to 17,000 mgL^{-1} have been recorded with discharge ranging from $10m^3sec^{-1}$ to 250 m^3sec^{-1} in the Kaskawulsh and Maxwell Creek basins (Johnson and Kruszynski, 1990). Sediment peaks could be observed at any point in the daily hydrograph. Once the drainage system of the glacier had been established in the summer the variability and high concentrations of sediment decrease and a diurnal regime is established. The late summer sediment concentration range was from 1,000 to 5,000 mgL^{-1}. In 1990 the development of the drainage system was disrupted by a storm in the late spring which resulted in a rapid opening of the glacier conduits and a subsequent damped discharge and suspended sediment regime. During periods of baseflow discharge, the suspended sediment concentrations ranged from 70-8,600 mgL^{-1}.

The importance of lateral moraines to the interpretation of proglacial meltwater characteristics can not be understated for they have similar hydrological systems to the gravity deposits, with the potential for additional effects due to water input from the gravity system and the glacier system. The melt of ice-cores also plays a major role in the hydrological and suspended sediment generation systems. External inputs, with surface flow, near surface flow or groundwater percolation, generate mudflows and debris flows. In Maxwell Creek the mudflow activity has eroded channels which continue to be the courses of similar activity over a number of seasons. These flows either terminate on the valley sides, building up large debris flow fans, or occasionally reach the system. In both cases they inject very high suspended sediment loads over periods of 2-3 hours. In 1990 the Maxwell site, the early season rainfall together with the snowmelt initiated a series of mudflows where gravity system input developed a subsurface channel through the crest of the lateral moraine and emerged into a subaerial channel at the top of a 30-35 degree slope. Slumping and mudflows occurred throughout the middle of June. The melt of the 30cm snowfall on June 22 accelerated this process producing frequent injections of suspended sediment in concentrations in excess of 50,000 mgL^{-1}. The inputs of water from glacier and gravity systems have eroded the deposits overlying ice cores removing sediments by mudflows.

6.1.3 Electrical Conductivity and Ionic Variations

From the 1989 and 1990 records of electrical conductivity (EC) and runoff in Maxwell Creek, the characteristic inverse relationship between solute concentration and the diurnal and intraseasonal variations in discharge occurred throughout the monitoring period with only a few perturbations occurring. As runoff increased, EC decreased and the rate of change in total dissolved solutes was inverse to the rate of flow. Linear correlations between intraseasonal discharge and solutes are low; however, the correlation between flow and total solutes increases when analysed separately during rising and falling limbs of the hydrograph. During both years the maximum value for EC coincides with that of the minimum value of discharge during spring flow conditions. The chemical enrichment of basal waters through contact with basal sediments are represented in EC and fluid discharge pulses in proglacial streams as described by Johnson (1991,b). The fluctuating, but generally decreasing levels of major solutes in meltwater during the latter half of the ablation season are due to the decreased enrichment of meltwaters within the subglacial drainage system.

It has been suggested that chemical characteristics of meltwater will depend on the degree to which water is able to access solute sources, which is the product of water-sediment contact time and routing of meltwater (whether at the bed or in ice-walled channels within the glacier). However, the most important variable is the chemical composition of the material that the water is in contact with, and its solubility with respect to numerous chemical and mechanical weathering processes. Examination of electrical conductivity has been used as a surrogate measure of total dissolved solids primarily for economic purposes. Electrical conductivity records of Maxwell Creek provide evidence of releases of small amounts of chemically enriched water at times of maximum and minimum water pressure. During high water pressures extension of the drainage network integrates hydraulically isolated pockets of solutes which have been stored and become saturated in solutes. Hence, measurements of electrical conductivity provide a general indicator of solute source and transfer within the subglacial network.

Total water chemistry records of Maxwell Creek should prove to be an invaluable indicator of subglacial hydrological processes. The repeated occurrence of short-term fluctuations in cation composition reinforces the idea that total water chemistry aids in understanding the process of ion exchange. It seems prudent to establish a rank order of importance based on 'surrogate' variables rather than compositional.

The nature of the dissolved solute load in meltwaters from Maxwell Glacier is essentially uniform under both autumn high-flow, and spring recession-flow conditions. In both cases Ca^{+2} (71-80%) is the major cation. The magnitude of the dissolved load in meltwaters is considerably higher under recession-flow conditions due to the mixing ratio between the supraglacial and subglacial components. A reason for the high Ca^{+2} composition appears to be the presence of moderately calcareous siltstone and sandstone below the glacier as they are more prone to chemical weathering.

6.2 Conclusions

This study sought to describe, interpret and discuss: 1) the fluvioglacial characteristics and processes of meltwater draining from an alpine glacier; 2) the use of electrical conductivity as surrogate measure for cation activity at the sole of an alpine glacier; 3) the role of intra- and interseasonal subglacial drainage system development in the manifestation of the varying magnitudes of characteristics in glacier-fed streams. To achieve these goals hydrograph and chemograph analysis between two seasons was performed.

This study illustrates the variable nature of hydrochemical characteristics in a glacierised basin. Distinct seasonal changes in discharge regime and climate affect the overall suspended sediment and electrical conductivity regimes. Temporal variability of glacial discharge is a result of different glaciological factors superimposed on climatological influences that modify streamflow on varying temporal scales. The inter- and intraseasonal variability in the quantity and timing of peak flows is observed in Maxwell Creek. Discharge in the spring, caused by snowmelt in the lower basin area, contributes solute poor meltwater to englacial and subglacial stores of solute rich meltwater stores. Hydraulically isolated stores of meltwater in cavities is progressively introduced by a system of meandering conduits and shifting cavities throughout the remainder of the ablation season. Exposed ice at the glacier surface increasingly contributes to the total discharge as the snowline rises. Although a glacierised basin provides a less variable runoff regime than a nival basin, the timing of events is markedly different from one year to the next. Discharge from snowmelt on the glacier, which is retarded compared to snowmelt on the non-glacierised part of the basin is initially supraglacial and contributes little sediment. Opening of the englacial and subglacial conduits produces variable and frequently very high sediment concentrations that may or may not be correlated with the discharge and are generally glaciologically controlled. During the late season, high discharge resulting from ice melt produces lower sediment loadings. Rain or snow at different stages of the snow melt, and rain on the glacier ice produces responses in discharge and suspended sediment concentration ranging from low/low to high/low to high/high. Rain during the period of subglacial conduit development may force the expansion of the system continually changing as the seasonal pattern of climate and meltwater availability dictates.

There is no standard sequence on which sediment-discharge and solutes transport takes place intra- and inter seasonally. Initial solute and/or sediment patterns vary, not only from one year to another, but from one week to the next and from one day to the next. As a result, it is necessary to develop an alternative model for describing the seasonal succession of solute and sediment load. Seasonal succession of initial concentrations can be described by studying discrete time periods of initial development through the season. This represents one view of the seasonal succession of drainage system development by the cumulative effects of the preceding conditions, or by the present effects of water or sediment interactions.

As each season progresses, the exhaustion of basal sediments as well as the enrichment of sediment from pockets at the base of the glacier contributes to the diurnal and intraseasonal variations in runoff. Although the pattern of decreasing solute concentrations tends to repeat itself year-to-year, the timing of the release of solutes varied considerably as a result of changing climatological and glaciological conditions.

The processes which control suspended sediment input are strongly climate related. The amount of winter snowpack and its rate of melt in the spring governs the potential for mudflows and debris flows as well as governing the throughput of water in the glacier system. Glaciological controls, ice movement and structures, generation of sediment and its transfer through the ice mass, control sediment release through the access of conduit and cavity flow to sediment sources. Hysteresis in the relationship between sediment transport and discharge prevents the success of empirical techniques. Even when different time scales were used within the data series to calibrate sediment rating curves, predictions were poor.

Although Ca^{+2}, Mg^{+2}, K^{+}, and Na^{+} are not considered in many hydrochemical studies, it appears they influence or represent changes throughout periods of the seasonal hydrograph. Few studies have included concurrent sampling of both these components in proglacial waters. The extent to which minor species is indicative of hydrological flow and sediment source is unknown. Although many factors are responsible for the development of the drainage system, the appearance of different compositional associations towards the end of the year could be due to subglacial sediment and/or meltwater availability. As sediment and water stores are not all available at the beginning of the season, the succession of events that occur depend on when and how the drainage system develops. The different associations occurring at the end of the year are in response to increased flow and enlargement of the system during peak flow conditions.

The use of cations during the ablation season on an hourly to a quarter-hour interval may provide the necessary information for determining source of both sediment and solute at the glacier bed. Throughout each monitoring period basal water pressure gradually increases early in the melt season as the water flux exceeds the capacity of the conduit system. This increase in water pressure results in ice separation from the bed and hence a high sediment solute flux. This high sediment flux is recorded in the form of short-term spikes and troughs resulting from short-term and seasonal sequence of ice-bed separation. Preliminary results of meltwater dissolved cation composition suggest that electrical conductivity (a surrogate for total dissolved cations) appears to conceal variations in the solute record. Hence, investigation of temporal variations in dissolved solute composition in meltwaters and relationships with discharge provides a useful method of interpreting subglacial processes.

The quality and quantity of glacial meltwaters acquired during passage through the subglacial drainage system reflects the nature and function of the highly variable system. Variations in solute and sediment concentrations relate to contrasting temporal and spatial patterns of sediment and water accessibility. Peaks in discharge, sediment load and total dissolved solutes are associated with the hydrological mechanisms at the base of the glacier. The patterns of sediment and solute quality and quantity can thus be used to assess the impact of hydrological events on the basal drainage system during an ablation season. However, once daily samples of hydrological parameters are not adequate to relate directly to erosion and transport processes of the subglacial drainage system. Changes in the configuration of the drainage network and accessibility to sediments control the short-term fluctuations in observed glacial discharge, sediment and solute concentrations. The Maxwell glacierised drainage basin is a minor component of a larger system of climate and glaciology that needs to be greater understood. Evaluation and prediction of suspended sediment and dissolved solute regimes can not be based on the present incomplete and relatively sparse data set. The variability that is exhibited within this sub-basin of a larger basin illustrates that the glaciological and hydrological responses to climate are individual. Until the ambiguity within the attributes of both the hydrological and glaciological regimes is eliminated, prediction of daily, seasonal and annual sediment-solute regimes for future periods are not possible. Future investigations in this field would benefit from more detailed studies of short-term fluctuations in dissolved cation composition of meltwaters, suspended sediment regimes and the controlling influence of bedrock geology.

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