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**Seasonal and Short-term Periodic Suspended Sediment  
Concentration and Bulk Hydrochemical Variations,  
Slims River 1993 and 1994, Yukon Territory, Canada.**

**By**

**Michael C. Sawada**

**A Thesis Submitted to the School of Graduate Studies  
and Research in Partial Fulfilment of the  
Requirements for the Degree of  
Master of Arts, Geography**

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0-612-20950-4

**Seasonal and Short-term Periodic Suspended Sediment and Bulk Hydrochemical Variations, Slims River 1993 and 1994, Yukon Territory, Canada**

**Mike C. Sawada**

**Abstract**

Peak seasonal discharge takes place after snowmelt in 1994 as meltwater production was amplified by more exposed glacier ice which was indicated by exponentially increasing diurnal discharge amplitude. Air temperature strongly influenced discharge in both years and precipitation was infrequent with limited influence. Discharges in 1994 were under-competent. Diurnal clockwise hysteresis defines the short term relation between suspended sediment concentration and discharge but current explanations fail to explain its frequency. Respectively, the dominant cations are  $\text{Ca}^{2+}$ ,  $\text{Mg}^{2+}$ ,  $\text{K}^+$  and  $\text{Na}^+$ , and each has a strong positive relation with conductivity. Conductivity, and thus individual cation concentrations, decrease over both seasons and are inversely related to discharge. Diurnal conductivity amplitude was greatest with glacier melt and clockwise hysteresis defines the short-term relation between discharge and conductivity.

*To Alice and Joseph Widurski, my great aunt and uncle,  
for the support, inspiration, and dedication that  
made this work possible.*

---

## Acknowledgements

Grateful acknowledgement is given to the Northern Science Training Program (NSTP) for providing the primary funding for this research and, as well to the Natural Sciences and Engineering Research Council of Canada (NSERC) operating grant to Dr.P.G.Johnson. Furthermore, logistical support was given by the Arctic Institute of North America (AINA) and University of Calgary by way of their Kluane Lake Research Station where the research took place.

I would like to thank my supervisor Dr.Peter Johnson for the opportunities he has given me as well as his patience, guidance, and diligence in helping prepare this manuscript. Thanks a million Peter, no make that  $\frac{1}{\text{thanks}^\infty}$  for accepting me as your student and introducing me to the exciting and rewarding field of glacier hydrology.

Thanks Dr. Barry Wellar for the opportunities and co-authorships, I grant you  $ke^{\text{thanks}}$  since you've provided a consistant methodological base for the building block nature of research. Thanks Dr. Daniel Lagarec for the conversations, research training and co-authorships, you receive the *poisson* distribution, pun intentional.

Special thanks to Pat Boyd for giving up a summer in order to help me with my final field season. And, in a related vein, Erinn Boyd for living without him for four months. Consequently, to Pat and Erinn I yield *thanks*<sup>10</sup>. Thanks also to Tony McElrea for your expertise in electronics which is  $\propto RI$  (thanks for the negligible R) and Janice for giving Tony something else to do besides electronics.

Thanks also to Andy Williams for his logistical support and expertise in solving daily minor crises as they would arise. Thanks for the aerial reconnaissance work you provided. Bill wants to fly the new plane in preparation for his next space mission. Thanks also to Jan and Mike Williams for a place to pipe and the ears to listen.

There are a number of individuals who aided in field research but whose contribution is greater than the usual alphabetical gratitude given by many researchers, so I thank you all by way of  $\pi$ , pie is good,

Louisa Vogrig , Monya Felchat, Janet Truchon, Mark Hamelin, Dave Ryan, Stephan Lefebvre, Dave Fitch, Christi na Schoermes, Glen Ffddich, Shan, Joc, Kirsten, Phil Wilson (Zoorpal), Arne Walelet, Louise

Finally, most of all, and beyond formalization, I would thank my family, Mike Sawada Sr., Kathie Sawada, and Raija Sawada for help and encouragement. And of coarse my great aunt Shirley Pelkey for teaching me something of my family history which, in itself, has been a great inspiration.

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# Chapter 1

Rational is being -  
asymptotical dreams  
I am defined in their chase

- *Sawada*

## Introduction

### 1.1 Context

Although glaciers occupy only 0.2% of the earth's land surface outside of polar regions, the headwater tributaries of many great continental rivers are in glacierized alpine zones (Collins, 1983). An understanding of the timing and magnitude of water storage and release from glacierized basins has critical implications for resource management and economic development and has been instrumental in the design and implementation of hydroelectric power production schemes in Switzerland and Norway (Ostrem, 1973; Kelley 1990, p.3). There is a growing need to understand and predict glacier hydrological regimes with respect to water quality and quantity as glacial runoff impacts human, natural and built environments (Davis and Keller 1983). The contribution of ice-melt to river flow is greatest during dry hot summers and so can compensate for lack of rainfall in drought prone or semiarid regions (Fountain and Tangborn, 1985) thereby affecting agricultural activities.

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Glacier-fed streams have high sediment loads as glacier erosional products are transported by seasonal meltwaters (Smith, 1985a). Thus, there is an increasing need for information on sediment transport for the evaluation of various environmental and engineering problems. These information needs cover numerous topics, including, the design of effective hydraulic structures for channel management strategies, soil loss, land degradation, reservoir sedimentation, debris transport and deposition, transport of non-point pollution, as well as factors affecting the conveyance of nutrients and contaminants through terrestrial and aquatic ecosystems. From an environmental viewpoint the discharge of sediment by glacier fed streams can be as important as the water itself.

Sediment movement is central to the discipline of geomorphology and suspended sediment concentrations in glacier fed streams reflect the glacier's erosional capability, the study of which is important to understanding contemporary glacial landscapes (Binda, 1984, p.3). Additionally, glacier hydrological and suspended sediment regimes control glacial lake sedimentation (Ashley, 1985; Bryan, 1974bc), and this can be used for paleoenvironmental reconstruction (Leemann and Niessen, 1994). Because the discharge of water from a glacier can vary from year to year so can sediment transport. The bulk of sediment may be deposited in as little as a few hundred years as ice recedes from an area (Ashley, 1985, p.3). Consequently, past climate change can then be interpreted within the sediment record of a receiving basin as changes in the amount of deposition over time.

## **1.2 Objectives and Approach**

Few researchers, with the exception of Bogen (1980), who have studied the effects of a proglacial area on glacier hydrological parameters over time. Most studies on glacial and proglacial hydrology have been performed in output streams either very close (distances less than 5 km) or very far (distances greater than 100 km) from the glacier terminus (Gurnell, 1982). Non-glacial sources can rapidly disguise the glacier sediment regime as distance from the glacier snout

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increases (Gurnell, 1982), and sediment size and total load decrease downstream from the glacier terminus (Bajewsky, 1985; Borland, 1961). This is especially true in valley trains that are characterized by shallow unstable braided channels undergoing numerous avulsions in response to highly variable seasonal discharges and sediment loads typical of glacial melt (Smith, 1985a).

What has not yet been investigated in detail, in this field, however, is the effect of an expansive proglacial zone on the short-term suspended sediment, solute and discharge regimes found at the glacier terminus. The aim of this research, then, is to describe the streamflow, suspended sediment and hydrochemical dynamics in the Slims River at its input to Kluane Lake, and offer some explanation for the trends and relations discovered. The research is conducted as an element of a larger project that proposes to use Kluane Lake sediment cores for paleoenvironmental reconstruction. To this end specific objectives are:

1. To measure discharge, suspended sediment and dissolved solids variations on short-term (1-10 minute), hourly, daily and monthly scales for parts of two melt seasons;
2. To measure the temporal variations in precipitation, air temperature, short-wave solar radiation and wind over the same time periods;
3. To ascertain the relation between streamflow, suspended sediment and dissolved solids daily and seasonally;
4. To ascertain the relations between hydrological and hydrometeorological variables daily and seasonally;
5. To ascertain and explain relations between the hydrological parameters within and between the two monitoring periods;
6. To suggest factors within the Slims River hydrological regime that are important in controlling sedimentation in Kluane Lake;
7. To suggest actions that may aid future research in determining the exact nature of the relations found in but beyond the scope of the present study.

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## **1.3 Hypotheses**

This exploratory research presents general research hypotheses and sub-hypotheses based on a literature search and review. The general research hypotheses, *Hr<sub>1</sub>*, *Hr<sub>2</sub>*, *Hr<sub>3</sub>*, and associated sub-hypotheses selected for consideration in the present research are as follows;

### **1.3.1 Discharge**

*Hr<sub>1</sub>*: In the early season the flow regime is controlled by snowmelt from proglacial tributary streams below the Kaskawulsh Glacier terminus but as the season progresses Kaskawulsh Glacier icemelt controls the majority of flow, such that,

1. Peak seasonal discharge occurs after snowmelt has ended.
2. Discharge fluctuates diurnally during snowmelt and after snowmelt has ended in response to daily maximum and minimum temperatures.
3. Diurnal discharge amplitude increases as the season progresses in response to increasing meltwater production from the Kaskawulsh Glacier.

### **1.3.2 Suspended Sediment Concentration**

*Hr<sub>2</sub>*: The suspended sediment regime is directly related to proglacial and glacial meltwater production and as the season progresses the Kaskawulsh Glacier hydrological system dominates the suspended sediment concentration regime, such that,

1. Peak seasonal concentrations coincide with the opening of the Kaskawulsh glacier hydrological system without a substantial increase in discharge.
2. Diurnal fluctuations of suspended sediment experience some hysteresis due to supply changes either subglacially or proglacially during both melt seasons.
3. The proportion of washload sized material transported increases as the season progresses as fine products of glacial abrasion are flushed out of the subglacial system.

### **1.3.3 Dissolved Solids**

*Hr<sub>3</sub>*: The dissolved solids regime, as measured by specific conductance, is directly related to meltwater production both proglacially and glacially as the season progresses. Conductance will decrease as the season progresses and Kaskawulsh Glacier melt dominates the flow regime, such that,

1. Bulk hydrochemistry undergoes seasonal and diurnal dilution with increasing snowmelt and/or icemelt.
2. Seasonally, initial peaks in conductivity correspond to snowmelt runoff.

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3. Diurnal fluctuations of conductivity are hysteretic during snowmelt runoff due to non-linear dilution effects within a two component system of snowmelt dilution of proglacially routed meltwater.
  4. Secondary peaks in seasonal conductivities occur with no substantial increase in discharge and are indicative of the opening of the subglacial hydrological system
  5. Hysteresis during glacier melt runoff is explainable in terms of supra-glacial dilution of sub-glacially routed meltwaters.

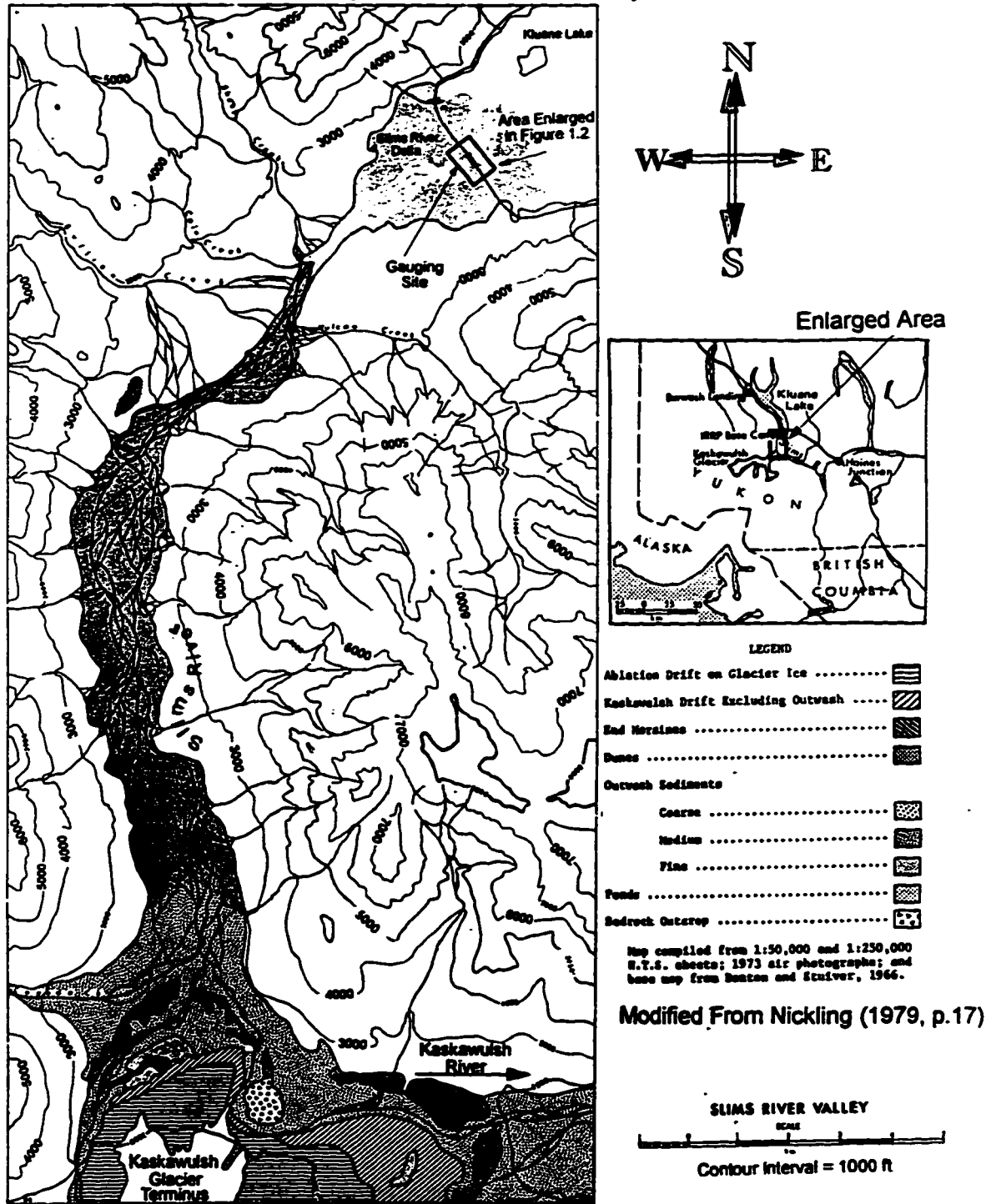
## **1.4 Study Area**

The Slims River valley (Figure 1.1) is situated in the Kluane National Park Reserve, Yukon, Canada. The area lies astride two physiographic regions, the Yukon Plateau to the northeast and the St. Elias Mountains to the southwest. They are separated by the Shawkak Trench, a structural break, and where this break is deepest lies Kluane Lake (Nickling, 1976, p.16). The Slims River runs from the terminus of the Kaskawulsh Glacier northeast (22 km distance) to Kluane Lake (Figure 1.1). The basin is approximately 2,456 km<sup>2</sup>, of which 2,020 km<sup>2</sup> is the Kaskawulsh Glacier basin (Johnson, 1991b). Of the total area 55% is glacierized, 65% in the Kaskawulsh Glacier basin and 9% in the lower basin (Johnson, 1991b).

### **1.4.1 Slims River**

Seventy to ninety percent of the Slims River flow is meltwater from the Kaskawulsh Glacier, with the remainder from the tributary streams below the glacier terminus (Bryan 1974a; Glaciology Division of Canada, Fisheries and Environment 1977) (Figure 1.1). The river source is a number of subglacial resurgences at the Kaskawulsh Glacier terminus that range up to approximately 6 m in diameter (Fahnestock, 1969). Calculated flow rates from these resurgences in 1965 agreed well with stream discharges below the terminus (Fahnestock, 1969). These resurgences have been observed for as long as records have been kept and aerial reconnaissance in 1993 attests to their continued existence, although, changes in their location and/or size are likely to have taken place. The Slims River valley active floodplain is approximately 1 km wide (Figure 1.1) and is

**Figure 1.1**  
**Locational Map of Slims River Study Area**



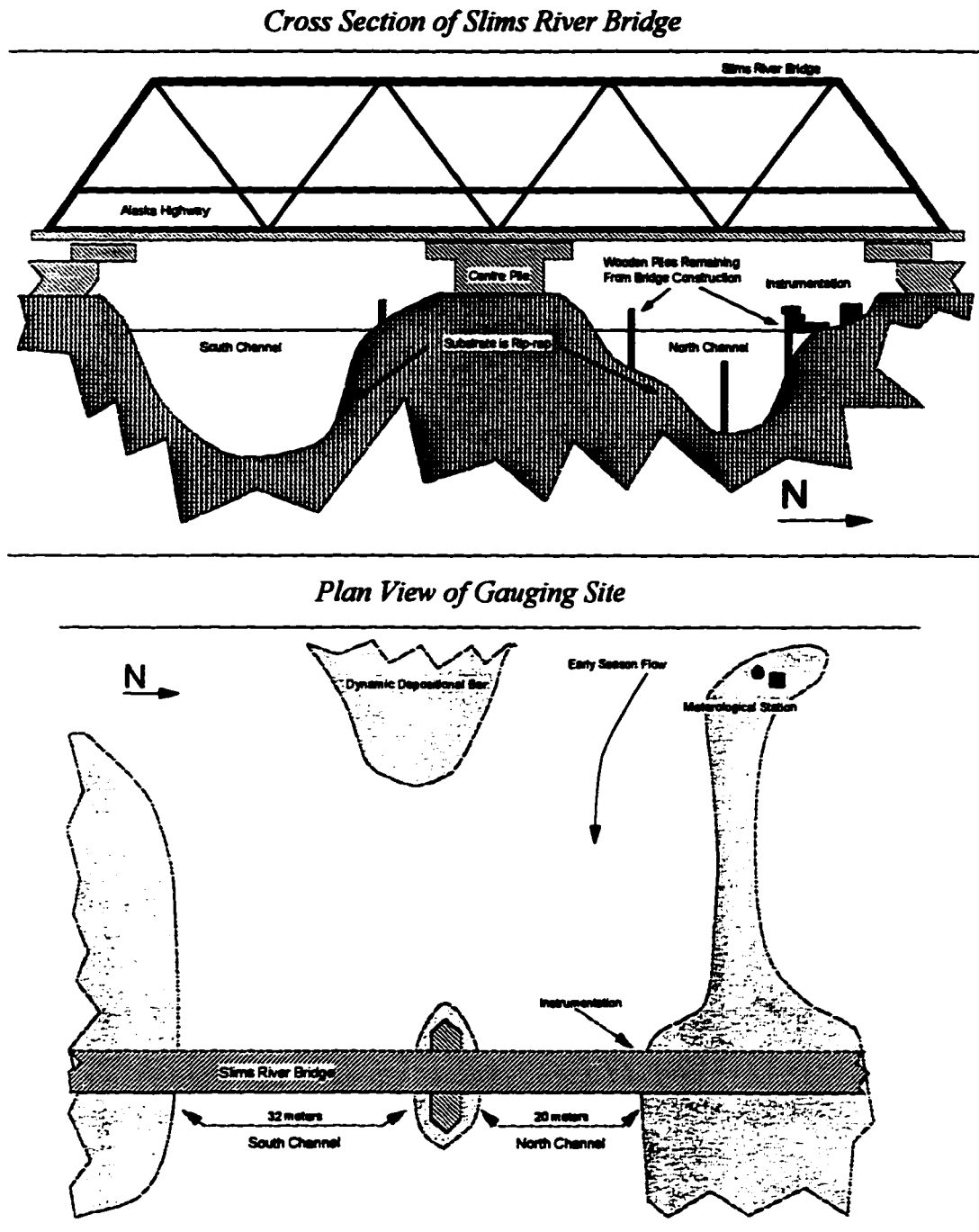
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delineated by a set of terraces approximately 0.5m above the active area (Nickling, 1976, p.18, 29). The river is braided for most of its course and the number of active channels range from two to thirty except where constricted by alluvial fans (Fahnestock, 1969). Before the river enters Kluane Lake it is constricted into two channels, (the north and south channel), that are respectively 20 m and 32 m in width (Figure 1.2).

The flow of the Slims River fluctuates diurnally and seasonally and has peak discharge in late July and early August with little to no flow from February to May (Bryan, 1974abc). Discharge of the Slims River ranged from 0.23 m<sup>3</sup>/sec in February 1962 to 318 m<sup>3</sup>/sec in August 11, 1962, and has been estimated to vary from 0 to 566 m<sup>3</sup>/sec (Fahnestock, 1969). Johnson (1991b) measured a peak discharge of 580 m<sup>3</sup>/sec resulting from quickflow associated with above-average July 1988 rainfall. As meltwater production progresses from snowmelt to icemelt the proportion of Slims River water derived from glacier melt changes (Johnson, 1991b). Contributions to flow in the lower basin come from Sheep Creek, Bullion Creek, Vulcan Creek and Canada Creek (Figure 1.1). Canada, Vulcan and Bullion Creeks are fed by glacial meltwaters and Sheep Creek is fed by high neve fields. These tributaries have high velocities and mainly transport sand and gravel (Nickling, 1976, p.28). Large alluvial fans are created by these tributaries and act to constrict the active floodplain of the Slims River and decrease the number of active channels (Fahnestock, 1969).

Sediments within the Slims River valley train are continually deposited and eroded due to seasonal changes in the river discharge (Nickling, 1976, p.29). Bryan (1974a) suggests that the Kaskawulsh Glacier provides the majority of suspended sediment for the valley train and therefore for the Slims River. Fahnestock (1969) proposes that much of the material going into the construction of the present delta comes from bank cutting along the portion of the valley train below Bullion Creek. The Slims River has well sorted sediments such that the geometric mean

**Figure 1.2**  
**Cross-sectional and Plan Sketch of Gauging Site**



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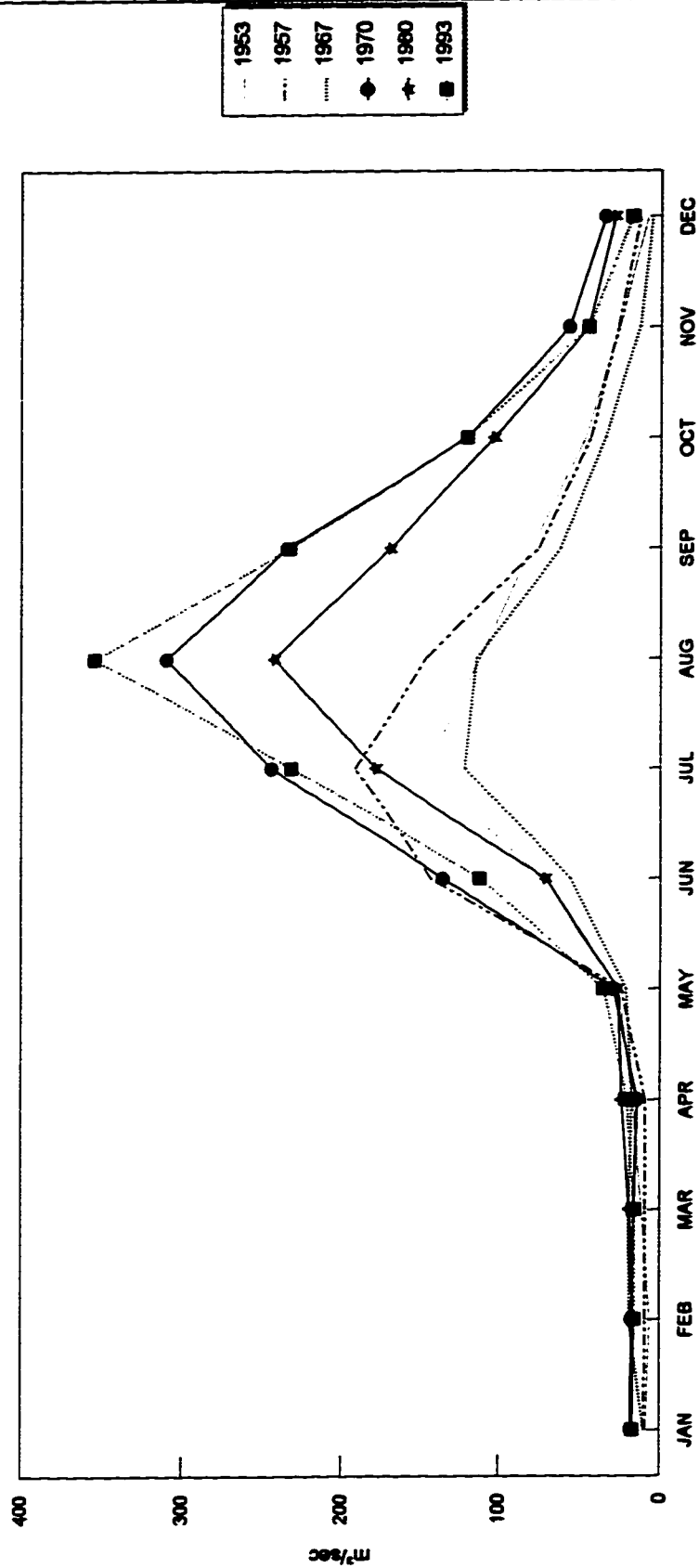
grainsize of valley train material decreases exponentially from the glacier terminus to the delta (Nickling, 1976, p.18). Because of the small distance between the Kaskawulsh Glacier terminus and Kluane Lake, grainsize changes in the valley train are the result of selective transport or hydraulic sorting rather than abrasion or weathering (Fahnestock, 1969). Smith (1985a) suggests this is common in valley trains because coarse material is transported infrequently and only for small distances during high flows whereas fine material is transported further under most flow conditions. Tributary streams below the Kaskawulsh terminus are non-turbid, flow through colluvial and alluvial material and provide little suspended load but they may supply the bulk of the dissolved solids in the Slims River (Bryan, 1974a).

The Kaskawulsh Glacier terminus lies on the hydrological divide between the Alsek River (of which the Kaskawulsh River is a tributary) system and the Yukon River (of which the Slims River is a tributary) system, and streamflow can change between these systems in different years (Johnson, 1991b; Barnett, 1974; Bryan, 1974ab). Barnett (1974) and Bryan (1974ab) have found features in the discharge regime at the Slims River bridge and in Kluane Lake that are indicative of such drainage shifts. This drainage change is detectable in the annual hydrographs for the Kluane River at the outlet of Kluane Lake as a shift in peak discharge from August (Slims River dominant) to July (Kaskawulsh River dominant) in different years (Johnson, 1991b) (Figure 1.3).

### 1.4.2 Kluane Lake

The Slims River is the major source of water for Kluane Lake and the lake level fluctuates with variations in stream discharge (Bryan 1974c). Water input from other streams is of lesser quantitative importance in these fluctuations (Bryan 1974c). Lake levels fluctuated as much as 3.04 m between 1900 and 1945 but recently have not been higher than 1.37 m above 1944 levels (Fahnestock, 1969). Fahnestock (1969) presents evidence for these fluctuations by noting that river silt deposits begin approximately 12.9 km from the Kaskawulsh Glacier terminus -

**Figure 1.3**  
Kluane River Annual Hydrographs



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approximately 0.9 to 1.5 m above present lake levels - which means that the present silt flats forming the upstream delta were related to higher water levels. Moreover, changes in lake level and sedimentological evidence at the lake outlet to the Kluane River suggest that the drainage of Kluane Lake has reversed in the past: During the Hypsithermal the lake drained through the Slims River valley into the Kaskawulsh River (Bostock, 1969).

In summer Kluane Lake temperatures decrease with depth from 10°C in the upper 0.5m to 6°C at 15m and 4-5°C at 40m+ (Bryan, 1974c). The lake possesses a weak thermocline often destroyed by violent mixing during storms (Bryan 1974cd). The fresh lake water does not encourage flocculation of suspended material upon introduction to the lake (Bryan 1974c). The lake is ice covered for 6-7 months and receives sediment from the Slims River over a melt season which is 5-6 months long. This seasonal melt and transport period provides opportunity for efficient monitoring and estimation of the sediment entering Kluane Lake (Bryan 1974c).

### **1.4.3 Geology**

The properties and dynamics of suspended sediment and dissolved load are partially a function of the geologic source material (Symander and Strunk, 1992; Faust and Aly, 1981). Kodybka (1994) provides a detailed geologic summary of the study area including the surrounding St. Elias Mountains. The range consists of a eugeosynclinal assemblage of sedimentary, volcanic and intrusive rocks that fossil date from Devonian to early Tertiary. The dominant geologic feature in the study area is the Kluane Range Intrusion. It is composed of granodiorite, quartz diorite and kiorite with some quartz monzonite containing quartz, feldspar, hornblende and biotite (Kodybka, 1994).

Surrounding the Kaskawulsh Glacier and Slims River, sedimentary and intrusive rocks contribute to hydrochemical and sediment dynamics. Near the terminus and lower part of the Kaskawulsh

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Glacier is the "Lower Icefield Ranges Clastic Assemblage" composed of slightly to moderately calcareous siltstone and sandstone containing mica quartzite, psammitic schist, minor grey phyllite, argillite and schist. The upper reaches contain igneous rocks known as the "Saint Elias Intrusions" and are composed of coarse to medium grained granodiorite containing lesser quartz diorite, minor quartz monziorite and diorite with local hornblende and biotite.

Glacial, alluvial and colluvial materials are found along Kluane Lake, the length of the Slims River valley and around the Kaskawulsh Glacier terminus (Figure 1.1). Alluvial and colluvial materials are products of weathering and erosion of glacially deposited materials after recent deglaciation. Consequently, they are overlying the older glacial diamictons and glacio-fluvial deposits.

## **1.5 Climate**

The climate of the St. Elias Mountains is influenced by the position of the mountains which act as a barrier blocking east air flow in the sub-polar low pressure belt (Nickling, 1976, p.18). Frontal effects over the mountains can reach the Kaskawulsh Glacier but cloud cover and precipitation are usually dissipated by subsidence before reaching the southern end of Kluane Lake (Taylor-Barge, 1969). The Slims River study area is effectively in the lee or rain-shadow of the St. Elias Mountains and mean annual precipitation is relatively low (Nickling, 1976, p.21).

Locally in the Slims River valley, down-glacier winds are a constant feature of the Kaskawulsh glacier and are either katabatic (gravity winds) resulting from night-time surface cooling and downward movement of the air or glacier winds which are achieved by conduction of heat away from the overlying air to the cold ice surface (Taylor-Barge, 1969). These winds cause the Slims River valley to have the greatest amount of eolian sediment transport in the range (Nickling, 1976, p.16). Winds are likely amplified through a venturi effect as channels are formed by the

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steep mountains bordering the north and central arms of the Kaskawulsh Glacier (Taylor-Barge, 1969).

Because of the lack of an on-site permanent meteorological station, only mean monthly temperature and precipitation data can be given as an annual historical record. For annual records the nearest permanent meteorological monitoring stations are at Burwash Landing and Haines Junction, approximately 50 and 60 km away from base camp, respectively (Figure 1.1). Sporadic meteorological data are available from observations taken by the Icefield Ranges Research Project at the Arctic Institute of North America's Kluane Lake Base Camp in 1946 and 1970-71 (Nickling, 1976, p.21), and by Nickling (1976, p.23) during the summer months of 1972-3. Summer data are also available from observations taken for the present investigation for part of the summers of 1993 and 1994. Temperature and precipitation data from the aforementioned sources are given in Tables 1.1 and 1.2.

Data from Burwash Landing and Haines Junction indicate that historically, June, July and August receive the most precipitation over the period of record (Table 1.1). This is also apparent in the data from the Arctic Institute of North America Base Camp record (Table 1.2).

Precipitation measured near the monitoring site shows agrees more closely with the Haines Junction Record than that from Burwash Landing. Precipitation data for 1993 and 1994 measured at Base Camp do not show any outlying events when compared to previous years or the surrounding permanent monitoring stations at Haines Junction and Burwash Landing.

Temperatures vary between the base camp and Slims River meteorological stations, such that, between these two monitoring sites, in both years, the mean temperatures are statistically different at the *alpha* 0.05 level where  $z = -2.627$  and  $z = -11.074$ , for 1993 and 1994 respectively. Trends

**Table 1.1**  
**Temperature and Precipitation Averages for Kluane Lake Area**

**Temperature Averages Haines Junction 1944 to 1985**

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Mean Daily Temperature	-21.5	-15.3	-8.1	-0.4	5.4	10.3	12.6	10.8	6.2	-0.4	-12.5	-21.2
Mean Daily Maximum	-15.8	-8	-1.3	6	12.6	17.9	20	18.6	13.1	4.9	-7.1	-15.5
Mean Daily Minimum	-27	-22.2	-17	-6.9	-1.8	2.5	5.1	3	-0.9	-6.1	-18.1	-28.8
Extreme Maximum	12.2	15	15.6	20.6	29.5	32.8	31.1	31	25.6	20	13.3	10.6
Extreme Minimum	-53.9	-53.9	-43.9	-30.6	-12.2	-6.7	-3.3	-11.1	-21.5	-36	-47.2	-53.9

**Temperature Averages Burwash Landing 1966 to 1980**

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Mean Daily Temperature	-22.3	-17.9	-11.2	-2	5.2	10.3	12.5	10.6	5	-3	-14.5	-20.2
Mean Daily Maximum	-15.9	-10.4	-2.7	4.7	11.8	17.2	19	17.2	11.3	2.2	-6.5	-13.9
Mean Daily Minimum	-28.9	-25.6	-19.9	-8.8	-1.5	3.4	6	3.9	-1.4	-8.4	-20.6	-28.6
Extreme Maximum	6.9	10.6	10.5	20.3	28.7	31.7	28.9	30.5	23.7	15.7	10.6	12.5
Extreme Minimum	-54.4	-55	-48.9	-35.1	-12.8	-5.4	-3.3	-7.8	-22.3	-35.4	-44.7	-51.2

**Precipitation Averages Haines Junction 1944 to 1985**

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Mean Rainfall (mm)	1.1	0.8	0.3	0.8	11.8	28.3	35.2	28.3	33.7	13.3	0.8	0.1
Mean Snowfall (cm)	32.7	19.4	9.9	8.8	4	0	0	0.1	0.4	23.2	31.3	29.8
Total Precipitation (mm)	32.3	18.6	9.2	9	15.7	28.3	35.2	28.5	34.1	35.2	30.5	29
No. Days with Rainfall	*	*	*	9	4	8	9	9	9	4	*	*
No. Days with Snowfall	10	8	5	4	2	0	0	*	*	7	9	11
No. Days with Precipitation	10	8	5	4	5	8	9	9	9	10	9	11

**Precipitation Averages Burwash Landing 1966-1980**

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Mean Rainfall (mm)	0.2	0	0	36	14.4	47	65.5	40.8	19.9	2	0.1	0
Mean Snowfall (cm)	14.2	8.8	11.1	11.8	12.6	1.2	0	1.5	6.4	16.3	14.4	12.3
Total Precipitation (mm)	12.6	7.8	9.9	11.4	26.9	48.4	65.5	42.3	26.1	18.2	12.5	10.3
No. Days with Rainfall	*	*	0	*	6	11	13	10	7	1	*	*
No. Days with Snowfall	8	7	6	5	3	*	0	*	2	8	10	7
No. Days with Precipitation	8	7	5	5	8	11	13	10	8	9	10	7

\*less than 1/2 day

**Table 1.2**  
**Summer Temperature and Precipitation Means**  
**Kluane Lake Research Station (Figure 1) and Slims River Valley**

	Mean °C										Precipitation mm							
	Maximum					Minimum					Slims River Valley	IRRRP Base Camp	IRRRP Base Camp					
	Slims River Valley	IRRRP Base Camp	Slims River Valley	IRRRP Base Camp	Slims River Valley	IRRRP Base Camp	Slims River Valley	IRRRP Base Camp	Days	Total mm				Days	Total mm			
April	1993	12.4 <sup>1</sup>	12.6 <sup>1</sup>	-0.6 <sup>1</sup>	-1.6 <sup>1</sup>	4.9 <sup>1</sup>	5.1 <sup>1</sup>											
May	1971		9.2		-2.4										1.0	1.27		1
	1973	12.1 <sup>1</sup>		0.8 <sup>1</sup>		6.4 <sup>1</sup>												
	1993	15.9	16.3	3.4	1.3	9.1	9.0									6.1		5
	1994	13.5 <sup>1</sup>	15.5	3.1 <sup>1</sup>	1.2	8.0 <sup>1</sup>	8.6									0		0
June	1946		20.3		5.2		12.7											
	1970		14.4		3.1		8.8											
	1971		16.6		3.2		9.9											
	1972	15.9 <sup>1</sup>		5.3 <sup>1</sup>		10.8 <sup>1</sup>												
	1973	15.0		4.9		10.0												
	1983	16.9 <sup>1</sup>	19.4 <sup>1</sup>	7.4 <sup>1</sup>	5.2 <sup>1</sup>	12.4 <sup>1</sup>	12.3 <sup>1</sup>									0		0
1994	16.1 <sup>1</sup>	19.7	6.4 <sup>1</sup>	5.1	10.7 <sup>1</sup>	12.6									21.58		7	
July	1970		16.3		5.5		10.9											
	1971		20.1		6.0		13.1											
	1972	19.0		6.6		12.8												
	1973	19.8 <sup>1</sup>		7.2 <sup>1</sup>		13.4 <sup>1</sup>												
	1994	18.3	22.0	6.1	6.5	11.7	14.5											
August	1970		15.4		3.7		9.6											
	1971		16.9		4.8		10.9											
	1972	17.1 <sup>1</sup>		7.1 <sup>1</sup>		12.1 <sup>1</sup>												
	1994	22.6 <sup>1</sup>	24.8 <sup>1</sup>	6.8 <sup>1</sup>	7.1 <sup>1</sup>	13.7 <sup>1</sup>	16.7 <sup>1</sup>											
~last 20 days		~last 11 days	~first 9 days	~first 4 days	~first 4 days													
~first 23 days		~11 days missing	~first 7 days	~first 10 days	~first 19 days													

<sup>1</sup>First 14 days

<sup>1</sup>First 19 days

<sup>1</sup>First 4 days

<sup>1</sup>First 10 days

<sup>1</sup>First 9 days

<sup>1</sup>First 7 days

<sup>1</sup>First 11 days

<sup>1</sup>11 days missing

<sup>1</sup>First 20 days

<sup>1</sup>First 23 days

Boldface from Nickling (1976, p.26)

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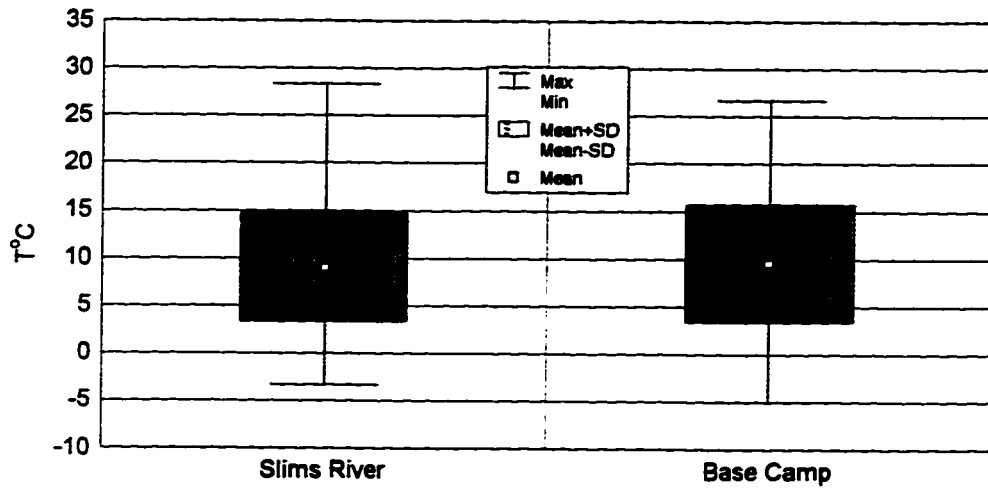
are similar for both seasons, however, the 1993 and 1994 temperatures encountered at the base camp have greater standard deviations and a higher means than those of the Slims station (Figure 1.4ab). Although the ranges and means differ, the two stations show good correlation between the mean daily and hourly time series. Hourly time series values for the monitoring period are well correlated ( $r = 0.93$  for 1993 and  $r = 0.90$  in 1994). In 1994 the mean daily temperatures at the two stations have a correlation of  $r = 0.90$ , for mean daily maximums  $r = 0.85$  and the mean daily minimum's are least correlated with  $r = 0.72$ . Although the Slims River data is on average 2 °C less than the Base Camp site ( $< 1^{\circ}\text{C}$  in 1993) their variations are almost identical as measured by the Pearson correlation coefficient. In comparison with Burwash Landing, Haines Junction and previous monitoring at Base Camp, average temperatures in 1993 and 1994 do not deviate to any noticeable degree. Hence, the Base Camp station will be primarily used in graphical analysis for both seasons as it represents the longer continuous record for both years.

## **1.6 Summary**

Past studies show that the Slims River has large and variable discharges throughout its melt season while transporting a considerable quantity of material, the products of glacial erosion. However, the Slims River short-term hydrometeorological and sedimentological regimes have been largely neglected as explanations have centered on the the fluvial and glacial histories of the area. Testing the hypotheses outlined in this introduction will be a component towards understanding the short-term regimes in the Slims River and may help in the understanding, explanation and prediction of short-term glacio-fluvial regimes in general.

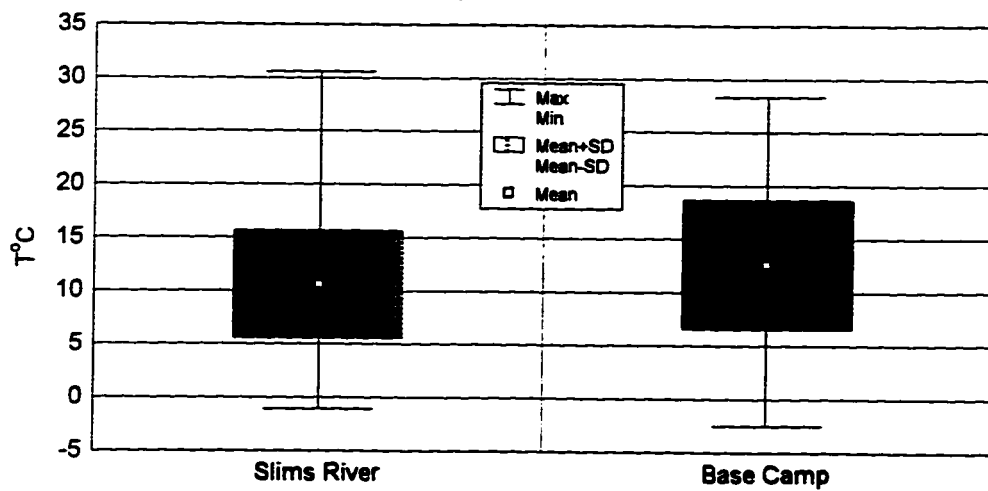
**Figure 1.4a**

Summary Statistics of Air Temperature  
Base Camp and Slims River Bridge, 1993



**Figure 1.4b**

Summary Statistics of Air Temperature  
Base Camp and Slims River, 1994



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## Chapter 2

*Shameful  
these clothes-  
not one stitch mine.*

Lady Sono-jo (1649-1723)<sup>0</sup>

### Background

#### 2.1 Context

The presence of a proglacial valley train between the Kaskawulsh Glacier and the monitoring site (Figure 1.1), affects sediment transport, and the sediment transport regime will in turn reflect the interaction between glacial and fluvial erosion and sediment transport mechanisms. The outcome of these interactions yields a transport/discharge regime that influences sediment deposition in Kluane Lake. Østrem (1973) suggests that since glacier erosion is unlikely to vary greatly from year to year, annual variations in sediment transport are related to other factors, foremost of which is water discharge. Consequently, in glacierized areas, the hydrological and hydrometeorological regimes are the primary controls of sediment deposition in glacial lacustrine environments.

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<sup>0</sup> All Haiku, unless otherwise specified, are from - Stryk, L. and Ikemoto, T. (1981). *The Penguin Book of Zen Poetry*. Penguin Books.

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## **2.2 Glacier Hydrology**

Glaciation has shaped the Slims River valley and provided the material for its valley train and numerous alluvial fans. It presently provides material to Kluane Lake by way of seasonal meltwaters. Moreover, the meltwaters of the Kaskawulsh Glacier provide the majority of the water to the Slims River as the season progresses which affects sediment transport, entriament and deposition in the proglacial zone and accumulation Kluane Lake. Therefore, as a primary influence, glacier hydrology must be explored since this component is paramount to understanding the objectives, hypothesis and present study area as well as in explanation of the relations uncovered in this work.

A number of generalisations regarding glacier-fed streams can be made. From year to year, glacier-fed streams show less annual and monthly variation in discharge than similar non-glacierized streams (Fountain and Tangborn, 1985). In warm dry years, glacier melt contributes considerably to streamflow while in cool wet years lack of icemelt is offset by more runoff from precipitation. Ultimately, mean summer air temperatures rather than total summer precipitation control mean annual runoff from glacierized basins (Leemann and Niessen, 1994; Pertziger, 1990; Ostrem, 1973). Winter runoff is generally negligible and consequently the summer months transport the majority of sediment and water (Leemann and Niessen, 1994; Borland, 1961). The concentration of dissolved solids in winter is generally higher than in summer but the increased runoff volume in summer compensates for lower values by increasing the total amount of dissolved solids transported (Collins, 1983). In spring and summer, discharge is variable and fluctuates diurnally as melt responds to daily maxima and minima temperatures (Kelly, 1990; Hook et al, 1985; Ostrem, 1973). These diurnal fluctuations are strongly influenced by individual weather events which may have dramatic short-term effects on the local regime, for example, a cool-wet period lessening the magnitude of diurnal discharge fluctuations (Bajewsky, 1985).

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With a glacier, there are six main sources of water that contribute to a temperate glaciers discharge (Simmons, 1989), these are,

1. surface ice-melt,
2. ice-melt due to mechanical stress and strain,
3. ice-melt from geothermal heating,
4. groundwater flow,
5. surface runoff, and
6. liquid precipitation.

Drainage routes within the glacier and water from marginal inputs may eventually reach the glacier bed where a linked cavity conduit system converges allowing the water to be discharged from single or multiple points at the terminus (Johnson, 1991a). Because of glacier rheology this subglacial conduit-cavity system is dynamic.

The dynamic nature of the cavity conduit system produces variability in both discharge and suspended sediment concentration at the terminus. With a relatively stable subglacial conduit network variations in subglacial sediment entrainment are measurable at the glacier terminus as diurnal suspended sediment fluctuations following diurnal discharge fluctuations. Alternatively, because of spatial instabilities in the cavity-conduit network and re-routing of water to fresh areas of the glacier bed, sediment flux or pulse events can occur independently of discharge variations (Collins, 1991; Gurnell and Warburton, 1990; Gurnell, 1982). In some glacier streams up to 46% of the total suspended load can be transported in association with pulse events (Gurnell and Warburton, 1990). Variations in discharge may occur as pulses that are due to re-routing of water and tapping of new sources in the dynamic cavity-conduit system beneath the glacier. They can occur in any flow conditions with variable magnitude and frequency, that is, from small fluctuations in the hydrograph to major flood events (Johnson, 1991a).

Diurnal hysteresis between suspended sediment concentration and discharge is common in glacial output streams. Hysteresis occurs when the suspended sediment concentration on the rising limb

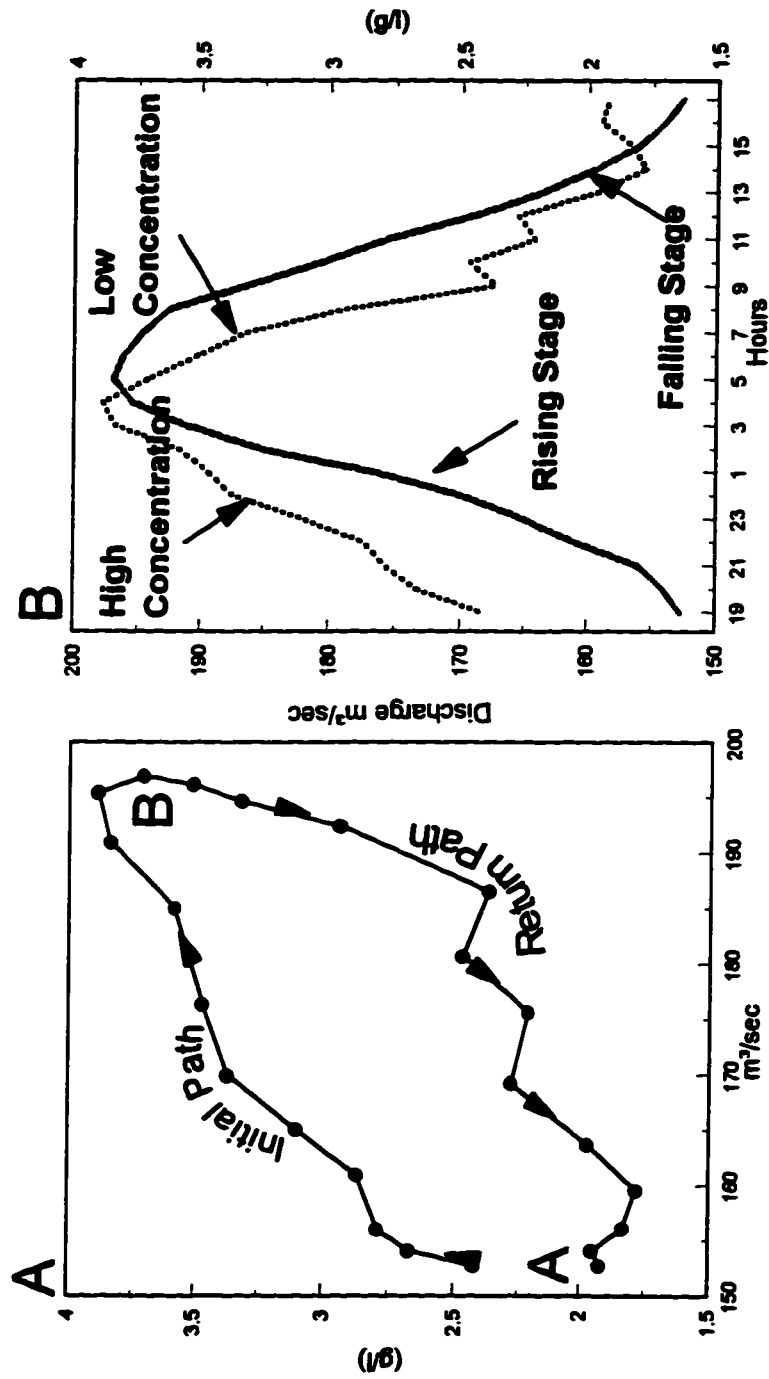
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of the hydrograph is different for a given discharge when compared to the falling limb (Figure 2.1 b). When concentrations are greater on the rising limb a rating plot of the event assumes a clockwise loop (Figure 2.1a). Hysteretic relationships can be either clockwise, counter-clockwise (greater concentration on the falling limb) or involuted (variable concentration on either limb) in glacier streams (Kelly, 1990, p.108; Collins, 1979a) (e.g., Figure 4.16). Ostrem (1973) notes that peak sediment transport preceding that of discharge is "a very common phenomenon" in meltwater streams when samples are taken close to the terminus. Maximum suspended sediment concentration occurs on the rising limb of the hydrograph when the rate of change of discharge is greatest (Kruszynski, 1993, p.105; Bradley, 1990, p.95; Kelly, 1990; Binda, 1984, p.102; Gurnell and Fenn, 1984; Collins, 1979a). Consequently, the rate of increase of discharge volume outstrips the rate of supply of sediment before maximum discharge is attained (Kelly, 1990; Collins, 1979a).

The failure to maintain concentrations at peak discharge may largely be due to greater volume-wetted perimeter ratios offsetting increased shear stress at conduit margins (Collins, 1979a). Involute and counter-clockwise events can be due to increased water pressures at high flow lessening the amount of deformation of basal sediments into conduits and forcing some water into bank storage in basal moraine (Collins, 1979a). Then, during low flow this water is released and aids in the collapse of conduit margins, producing sudden or attenuated peaks in suspended sediment concentration at the glacier outlet (Collins, 1979a).

This non-linear relation between concentration and discharge causes most standard linear or curvilinear rating curves to fail in predicting suspended sediment load from discharge in glacier streams (Kelly, 1990, p.124; Collins, 1979a). In this respect, Gurnell and Fenn (1984) suggest that Box-Jenkins transfer functions are superior to standard rating curves and should be the preferred model to predict suspended sediment load in proglacial streams. As demonstrated in

**Figure 2.1**  
 Hysteresis in Slims River SSC and Q, July 16-17, 94



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section 2.3, it is evident that these processes in the glacial environment that act to produce hysteresis are more complicated when the proglacial zone is taken into account.

In a glacierized basin, dissolved solids regimes are affected by glacial and proglacial water discharge. At the glacier, ionic exchange between the water and sediment is favoured with increasing contact time, leading to solute enrichment of meltwater and higher conductivity values (Collins, 1983; Lemmens and Roger, 1978). The spatial variation in the basal drainage network and seasonal variation in discharge determine the residence time of water in the subglacial network and, hence, to a great extent the dissolved load in the water (Brown and Tranter, 1990). Dissolved solids are also a function of the initial water chemistry and geologic substrate (Lemmens and Roger, 1978). Non-glacial sources of water and meltwater such as groundwater, snowmelt or precipitation all have particular ionic loads that can increase or decrease the concentration of dissolved solids transported (Zeman and Slaymaker, 1975). Seasonally, concentrations are higher in the early season reflecting discharge of enriched water that was in contact with basal sediments followed by a decrease in concentration with the dilution that accompanies sustained ablation (Binda, 1984, p.85; Collins, 1983). This dilution is in response to initial low concentrations in glacier ice that can be represented by typical atmospheric levels (Bradley, 1990, p.25).

Increasing discharge from glacier ice-melt decreases contact time between water and sediments leading to an inverse relation between discharge and dissolved load (Kelley, 1990, p.89; Collins, 1983; Lemmens and Roger, 1978). Bryan (1974c) has found this relation at the Slims River bridge. In close proximity to the glacier snout, the inverse relationship is largely due to relatively fresh glacier water diluting ground or basal water and other extraneous sources of water in the proglacial environment (Bajewsky, 1985; Barnett, 1974). Hysteresis in discharge solute relations in meltwater streams is due to the time delay before solute rich basal meltwaters reach the snout

### *2.3 Open Channel Hydrology and Glacial Interaction*

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(Richards, 1982, p.96). The dilution effect may be amplified as distance from the glacier terminus increases and non-glacial peripheral sources of water enter the system.

In summary, variation in discharge, suspended sediment and dissolved solids at the glacier terminus are due to spatial and temporal modification of the glacier bed via hydraulic integration of isolated areas by escaping meltwaters (Collins, 1990; Collins, 1989; Gurnell, 1982). This leads to hysteresis in both the bulk hydrochemistry and suspended sediment concentration of glacier-fed streams. Moreover, this measurable variability has led researchers to infer that the subglacial/englacial hydrologic system is composed of a network of conduits and cavities whose aerial integration of the glacier bed is in response to changing hydraulic, meteorological and rheologic phenomena over a melt season (Collins, 1991; Johnson, 1990b; Johnson, 1990a; Seaberg et al, 1988; Hook et al., 1985).

## **2.3 Open Channel Hydrology and Glacial Interaction**

The Slims River is a typical proglacial braided stream as it exists in a high energy fluvial environment with a relatively steep valley gradient, possesses large variable discharges, and has banks lacking stabilizing vegetation (Richards, 1982, p.211). There is some modification to the suspended sediment and discharge regimes between the source and any point downstream (Bogen, 1980). Once water leaves the glacier outlet it enters open-channel conditions, and in the case of the Slims River the stream becomes braided, and so is subject to common characteristics of alluvial channel hydrology. Smith (1985a) suggests that the similarities between glacial and non-glacial fluvial deposits are greater than their differences. Consequently, it is necessary to understand the hydraulic principles behind sediment transport and flow in open channels in order to fully explain the hydrological regime of the Slims River.

### 2.3 Open Channel Hydrology and Glacial Interaction

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Water moving in a stream exerts a tangential force that is resisted by a boundary shearing stress from the channel bed and banks. When no acceleration occurs the resistance equals the tangential force and the shear stress is transmitted from one layer of water to another through molecular interaction in laminar flow and turbulent exchange of momentum in turbulent flow (Leopold, Wolman and Miller, 1964, p.153). Because of boundary roughness, open channels never experience laminar flow throughout the entire water column. In turbulent flow, velocity decreases with the logarithm of depth and shear stress increases as the square of velocity (Leopold, Wolman and Miller, 1964, p.154, 172). Turbulent flow exerts larger shear stress than laminar flow for equal flow velocities (Petts and Foster, 1985, p.98). The ability of a channel to transport material depends on the balance of tangential forces of the moving water which change with velocity, viscosity, density and water surface slope and the resistance offered by the channel boundaries that vary with material type, area, plan form, particle size, and bedform development.

Stream power - the loss or expenditure of potential energy between two points - rather than force better describes the rate of energy supply at the channel bed for overcoming friction and initiating sediment transport (Petts and Foster, 1985, p.103-4; Richards, 1982, p.13). There is a critical level of stress necessary for movement of a given grain size (Leopold, Wolman and Miller, 1964, p. 172) - referred to as flow competence - which increases proportionally with stream power, discharge and velocity. Once inertial forces of the particle have been overcome it will move by rolling or saltating on the stream bed and, as vertical velocity components in turbulent eddies overcome the immersed weight, in suspension.

Suspension,  $s$ , occurs when the vertical velocity in terms of shear velocity  $v^*$  in the vertical, is greater than the fall velocity  $w_o$ , for a given particle assuming uniform density,

$$s = \frac{w_o}{v^*} \quad \{v^* > 0\} \quad (2.1)$$

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if  $s \leq 1$  the particle will remain in suspension (Middleton, 1975).

The vertical velocity of turbulent eddies is often approximated as 1/12 the mean forward speed of the current (Friedman and Sanders, 1978, p.100). Additionally, the fall velocity of particles varies as the root mean square of the vertical velocity component (Middleton, 1975).

The simple model of suspension is complicated by vertical velocity gradients which cause shear gradients and, hence, suspended sediment vertical concentration gradients where grain size and concentration increase with depth at a given flow competence. In addition, water viscosity increases with increasing concentration of fine particulate (<0.063 mm) and decreasing water temperature. Thus, high concentrations of fine particulate can reduce the effective grain size of individual sand grains by 30% to 80% (Richards, 1982, p.78). Consequently, actual fall velocity is less than  $w_o$  in turbulent and turbid water (Richards, 1982, p.102). Particles that fall below a given critical shear stress or fall velocity tend towards uniform concentration in the vertical. This is the case with washload sized material because  $w_o$  is so small that normally ineffective values of  $v^*$  will allow suspension,  $s$ , to be less than unity.

Easy mobilization and suspension of particles below 0.063 mm mean that they are almost always transported according to supply rather than competence (Petts and Foster, 1985, p.108, 115; Ward, 1984; Sidle and Campbell, 1985; Paustian and Beschta, 1979). Such factors make it difficult to determine total concentrations analytically. Additionally, large turbulent eddies can cause local shear stresses up to three times the average (Petts and Foster, 1985, p.108, 115).

Stream power expenditure and transport competence increase proportionally with stream gradient. This means hydraulic sorting takes place as particles are selected according to size for

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erosion and deposition as competence changes over a flood hydrograph (Petts and Foster, 1985, p.110) or in the long profile with gradient.

Cumulative frequency plots of grainsize distribution commonly show discontinuities from a single straight line relationship and these discontinuities can be broken into a series of straight line segments where each segment represents a distinct normal distribution (Visser, 1969; Middleton, 1975). The truncation or breaking point between two line segments identifies the size where one mechanism of transport changes to another, e.g., saltation to suspension (Middleton, 1975). With knowledge of the average bed material composition and the shear velocity of flow it is possible to predict the average size distributions and truncation points for the various types of suspended material classes in a stream based on the above equations (Middleton, 1975).

It has been suggested that if the grainsize distribution can be reproduced based on hydraulic theory, then this distribution is due to hydraulic sorting and, hence, is not necessarily limited by the grainsize distribution of the parent material, e.g., soil or bank material (Middleton, 1975). Despite capacity or competence limitations however, this argument does not apply to the washload component because it is supply dependant. Since one portion of the suspended sediment distribution is supply controlled while the other is hydraulically determined, it is not surprising that suspended sediment samples can be bimodally distributed, that is, a mix of two populations. Thus, the stochastic approach to prediction of sediment concentration often fails because of non-capacity loads.

In summary, once the Slims River enters the valley train it becomes braided and the hydrological regime becomes modified, such that entrainment and deposition occurs based on local hydraulic conditions. Depending on local turbulence there may be vertical suspended sediment concentration gradients. Under most flow conditions silt will be evenly distributed in the vertical

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and, if viscosity and shear velocity increase, so will larger grainsizes. In general, stream competence and capacity increase with increasing stream velocity and discharge per unit channel length. Consequent grainsize distributions of suspended material can be analytically determined but are complicated by the non-capacity sediment component, namely silt and clay, that are transported based on supply rather than hydraulic limitations.

### 2.3.1 Hydraulic Geometry

Sediment transport and discharge imposed from upstream affect local changes in channel morphology. Stream geometry reflects discharge/sediment interaction where channel adjustments take place in order to maintain sediment transport continuity. The basic continuity equation for steady flow is given by,

$$wdv = Q \quad (2.2)$$

where  $Q$  discharge,  $w$  is the average channel width,  $d$  is average depth,  $v$  average velocity. Concepts of hydraulic geometry suggest that six independent variables describe a given channel cross-sectional flow and are commonly given by power functions,

$$w = aQ^b; d = cQ^f; v = kQ^m; s_w = gQ^z; n = hQ^{y_1}; f_f = pQ^{y_2}; \tau_o = tQ^x \quad (2.3)$$

where  $s_w$  is water surface slope,  $n$  is Mannings roughness coefficient,  $f_f$  is a friction factor and  $\tau_o$  is bed shear stress (Richards, 1982, p.149). Since,  $wdv = Q$  the exponents  $b+f+m$  and coefficients  $a+c+k$  sum to unity and are therefore not independent of each other. Additionally, derived variables like  $n$ ,  $f_f$  and  $\tau_o$  make it unclear whether all should be given equal weight (Richards, 1982, p.151). Over a flood hydrograph many of the hydraulic geometry parameters experience hysteresis e.g., higher velocities, water surface slope, sediment concentration, and

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shallower depth on the rising limb for the same discharge on the falling limb (Leopold, Wolman and Miller, 1964, p. 230).

Studies of hydraulic geometry demonstrate that an at-a-station cross-section readily adjusts through negative feedback in such a way as to allow  $w$ ,  $d$ , and  $v$ , to change according to the upstream imposed sediment load and discharge (Leopold, Wolman and Miller, 1964, p.233) in order to maintain sediment transport continuity (Richards, 1982, p.172-3). By way of illustration, given two channels ( $C_1$  and  $C_2$ ) with equal change of width with discharge (exponents  $b_1=b_2$ ), if  $C_1$  has greater change of depth with discharge ( $f_1>f_2$ ) then  $C_2$  must have a smaller change of velocity with discharge. When compared to a single channel, in order to maintain sediment transport continuity, in the long profile, braided streams have greater distributary widths, smaller depths, and high channel boundary resistance to dissipate excess energy in order to compensate for steep channel gradients (Richards, 1982, p.211, 217).

Accordingly, long profile form and process exerts a fundamental independent control on the at-a-station hydraulic geometry. Changes in at-a-station hydraulic geometry occur in disequilibrium channels with excess load by fill, and channels with load deficits experience scour according to the transport regime (Richards, 1982, p.172). For a given sediment load, a narrow channel has high velocity rendering it load deficient so it will scour the bed and erode the banks until its form is appropriate for transport continuity (Richards, 1982, p.173). Similarly, a deep wide channel with a load in excess of what can be carried by a given velocity and discharge will aggrade the channel bed until transport continuity is attained (Richards, 1982, p. 173). Hydraulic geometric changes are also dependent on the particle size distribution of the imposed load where channels carrying a large proportion of silts and/or clays can be overcompetent at most flow velocities and cause bank erosion until channel form is in equilibrium to permit transport continuity.

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In summary, sediment transport and deposition partially depend on supply and local hydraulic geometry. At-a-station hydraulic geometry depends on the imposed sediment load, for which, the channel will modify itself appropriately in order to maintain sediment transport continuity. In effect, the proglacial zone can be thought of as a self-modifying system, which through negative feedback is able to transport the imposed sediment load from internal and external inputs as the season progresses.

### **2.3.2 Proglacial / Glacial Suspended Sediment Transport**

Large and variable discharges from glacier melt lead to high sediment loads and an active valley train (Smith, 1985a) and thus considerable changes in hydraulic geometry over a season. Active channels migrate and the stream constantly reworks the upper surface of the valley fill and carries relatively large sediment loads that reflect the perimeter sediment properties (Richards, 1982, p.170) as well as external inputs, namely, direct glacier sediment discharge (Bogen, 1980).

Although water velocity slows towards the bed and channel banks, turbulence transfers momentum laterally towards the banks and vertically towards the bed. Variation in velocity at the channel banks is significant in the context of bank erosion (Petts and Foster, 1985, p.114; Richards, 1982, p.72) and bank erosion is significant in terms of internal channel sediment supply. Bank failure depends on the forces of erosion, local flow geometry, bulk mechanical properties of the sediment (cohesiveness), bank geometry and structure and antecedent moisture conditions. The erodibility of cohesive silty-clay banks is a function of the chemical tractive force between the particles which is dependent on the mineralogy - mainly the presence of clay minerals and electrolyte (Richards, 1982, p.162; Friedman and Sanders, 1978, p.108).

Rapid draw-down of water after high stage furnishes positive pore-water pressures and effectively weakens bank material (Petts and Foster, 1985, p.113). This is amplified by removal of the basal

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material from saturated banks with large overburden weight (Richards, 1982, p.164). Rotational slips, non-circular slips or slab failure may then result because of overburden waterlogging, decreased frictional shear strength caused by increased pore water pressure, and increased relative height from basal scour (Petts and Foster, 1985, p.113; Richards, 1982, p.164). Depending on the position within the bank, retreat and failure is partly a function of the weakest component (Richards, 1982, p.162). Cantilever failure is common where river water thaws frozen sediments and enhances fluvial undercutting of the banks at the waterline (Petts and Foster, 1985, p.114). Bank erosion often supplies the majority of sediment load to rivers that run through alluvial valleys (Moore, 1984)

Perimeter sediments are important as internal storage supply mechanisms because in-channel storage can be ten times the annual export of total sediment (Bathurst, 1987). Sediment may be tapped as the rising stage breaks up bed or bank armour (Paustian and Beschta, 1979). Other storage may include accumulation of fines in bed gravels, bank sloughing, upper channels and bank supply, bank failure during low flows, pool storage, fluvial erosion and storage behind large organic debris (Sidle and Campbell 1985; Paustian and Beschta, 1979). These storage mechanisms have been shown to account for as much as 64 tons per lineal kilometre of stream channel in the Western Cascade Mountains (Paustian and Beschta, 1979). Internal storage is important to the sediment dynamics within a proglacial environment that produce hysteresis.

For estimation and prediction of suspended sediment load, accurate and precise specification of curvilinear relations between suspended sediment and discharge have often been complicated in glacier and non-glacier fed streams due to hysteresis. Hysteresis typifies the suspended sediment concentration discharge relation during stormflow events in fluvial streams. Clockwise hysteresis is most common in storm runoff events (Figure 2.1a). Most examinations of hysteresis have been done in order to better understand and predict storm sediment yield in small unglaciated

### *2.3 Open Channel Hydrology and Glacial Interaction*

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catchments, for example, Wood (1977), VanSickle and Beschta (1983) and Walling and Web (1982). Because of hysteresis, estimating stormflow sediment yield from sporadic sampling regimes is often erroneous.

It is commonly held that hysteresis takes place due to variation in the sediment supply and availability throughout the storm event. Clockwise hysteresis arises because the rising limb of the hydrograph taps and exhausts, before maximum discharge, supplies of fine grained sediment in the channel banks and bed accumulated since or deposited during the previous flood recession (Meade et al, 1990; Bathhurst, 1987; Petts and Foster 1985; VanSickle and Beschta, 1983). Thus, sediment concentration on the falling limb is lower, although total load may be greater because of hydrograph asymmetry.

Wood (1977) and Walling and Web (1982) found evidence of sediment depletion in compound storm events and conclude that the maximum sediment concentration in successive events is a function of the time elapsed between events when processes of weathering and erosion act to build-up supply. Moore (1984), and in a similar fashion VanSickle and Beschta (1983), modelled sediment storage by assuming that availability increases at an exponentially decreasing rate between storm events, which allows supply to be exhausted on temporally close discharge events. This is due to supply depletion, that is, the majority of the accumulated supply is depleted on the first event (Bathurst, 1985; Moore, 1984; VanSickle and Beschta, 1983; Wood, 1977; ). Therefore, the longer the elapsed time between storm events the more time is available for processes of weathering to accumulate products for erosion and, consequently, the greater the sediment concentration on those events.

Conversely, Moore (1984) also suggests that lower sediment concentrations on subsequent events may indicate a change in the sediments' resistance to erosion due to wetting or other factors and

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so, from a physical viewpoint, may not represent supply limitations. So exhaustion may be an equifinal product of various processes and not necessarily a supply function. Regardless of the physical assumptions, clockwise hysteresis (as an indicator of exhaustion between events) will not take place if the preceding peak flows are not of a sufficient magnitude to cause exhaustion of material (Wood, 1977).

Bogen (1980) considers the suspended sediment concentration of proglacial streams in terms of internal and external supply mechanisms (Figure 2.2). First, internal supply mechanisms, such as bank collapse, or re-entrainment contribute to the downstream suspended sediment concentration. As the stream passes various channel sections these mechanisms can be visualised as a discrete set of sediment inputs to the streams total sediment concentration, e.g.,  $I_1, I_2, I_3$ , etc. Hence the sum of these particular inputs at a given time,  $t$ , will yield the total internal contribution to the streams suspended sediment concentration,  $C_I(t)$ , such that,

$$C_I(t) = \sum_{i=1}^n I_i \quad (2.4)$$

Second, external inputs,  $E_1, E_2, E_3$ , etc., from tributaries or more importantly from direct glacier discharge also contribute to the downstream suspended sediment concentration. Thus, the external sediment concentration at some time,  $t$ ,  $C_E(t)$ , can be envisaged as,

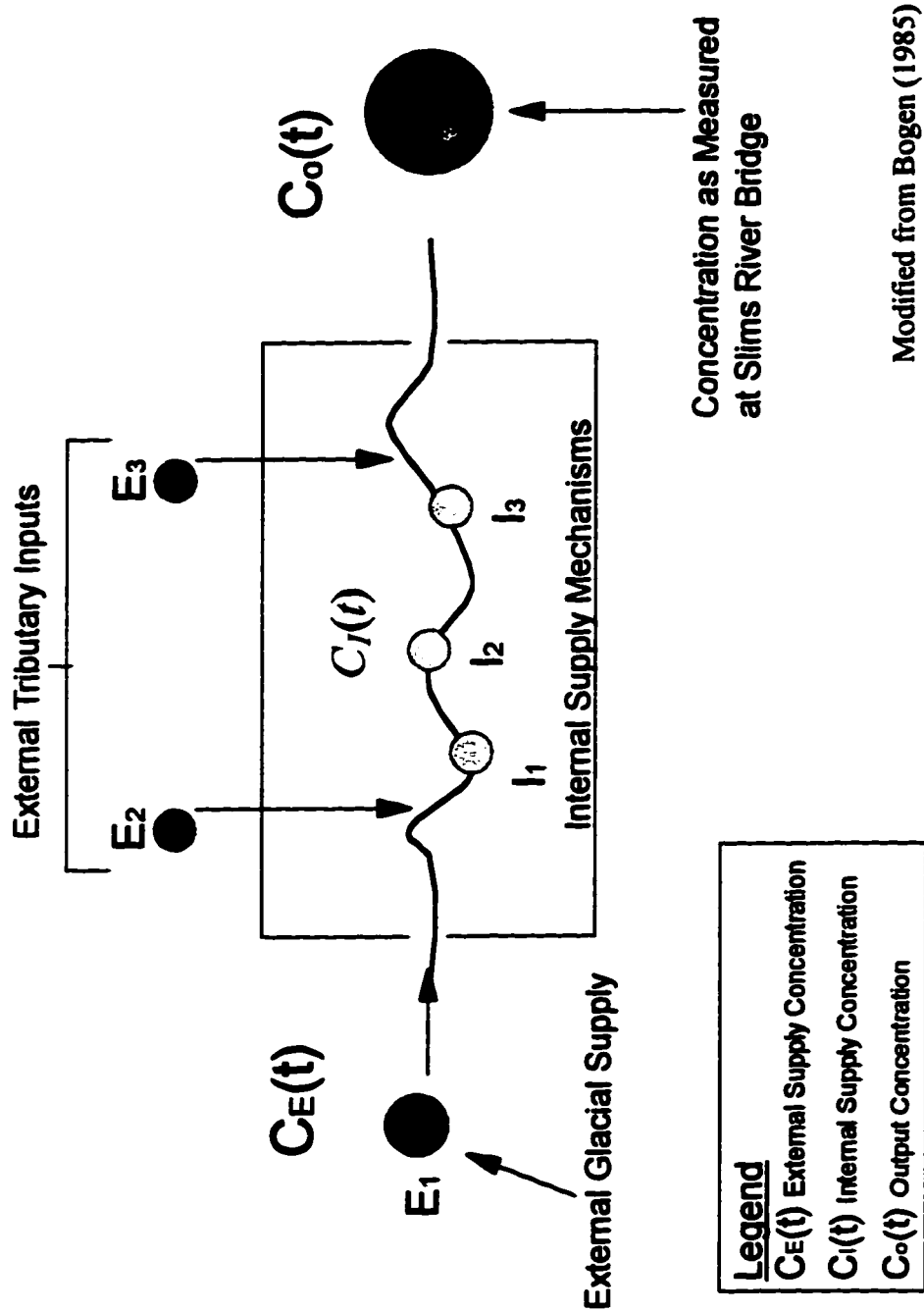
$$C_E(t) = \sum_{i=1}^n E_i \quad (2.5)$$

Consequently, the concentration at some point downstream from the glacier at time,  $t$ ,  $C_o(t)$ , is the sum of internal and external sources, in that,

$$C_o(t) = C_I(t) + C_E(t) \quad (2.6)$$

Bogen (1980) suggests, that in absence of external supply mechanisms other than the glacier, the time-varying difference between the input concentration and output concentration will reflect sedimentary processes within the proglacial reach. He found that the re-entrainment of

**Figure 2.2**  
Symbolic Proglacial Sediment Transport System



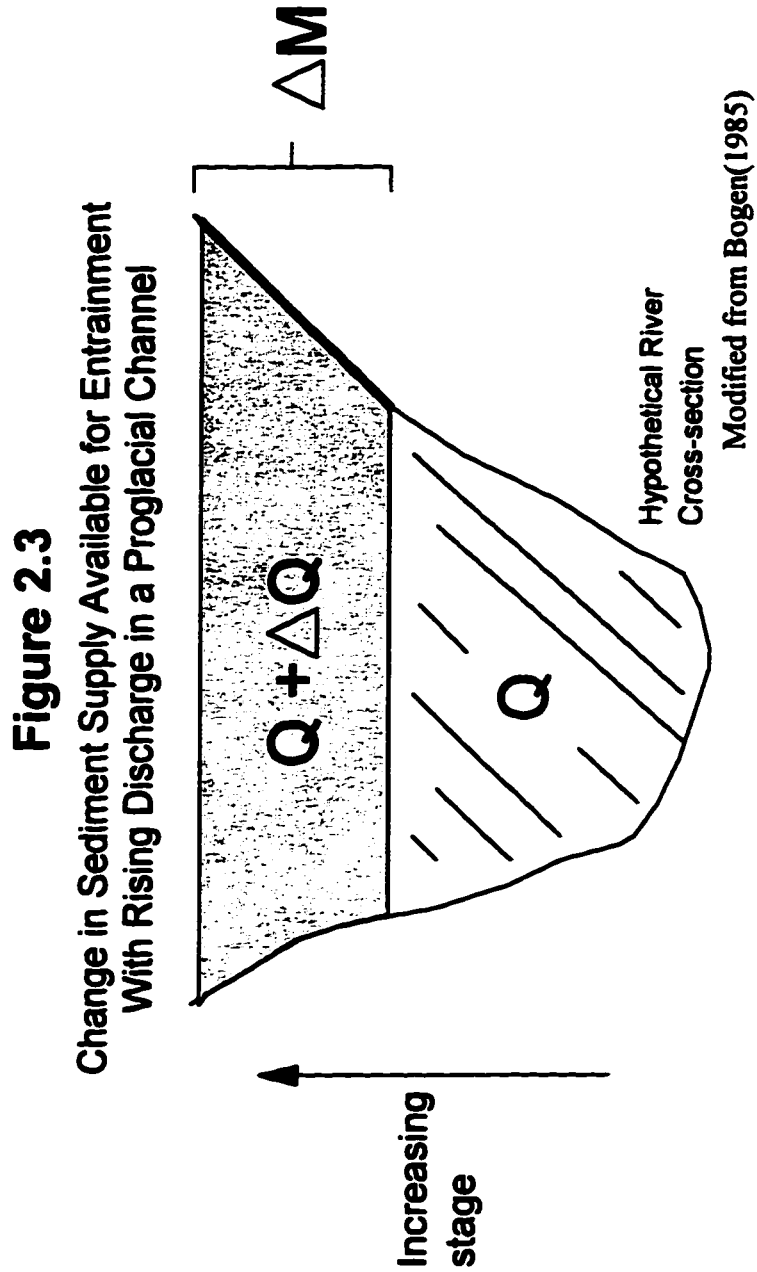
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temporarily stored external sediment produced higher suspended sediment concentrations on the rising limb of the hydrograph and, hence, hysteresis. Higher concentrations on the rising hydrograph limb occur because  $C_o(t)$  is the sum of  $C_E(t)$  plus the amount of sediments added to the water volume,  $(Q + \Delta Q)\Delta t$ , passing a given cross-section during a time interval  $\Delta t$  (Figure 2.3), that is to say,

$$C_o(t) = C_E(t) + \frac{\Delta M}{(Q + \Delta Q)\Delta t} \quad (2.7)$$

Equation 2.7 suggests that the suspended sediment concentration on the rising limb depends largely on the rate of change of discharge and hence flow volume on the rising limb of the hydrograph (Bogen, 1980). In the absence of  $C_E(t)$ , concentration is dependent on the length of time intervals necessary to fulfil the increase to peak discharge. During a rapid increase in discharge sediments will be added to a smaller volume and, consequently, concentration will be higher than during a slower increase to peak. That is to say, the rate of increase of discharge often outstrips the available sediment supply when concentrations above  $C_E(t)$  are largely provided by internal deposits (e.g., Collins, 1977). The most frequent internal supply mechanism that Bogen (1980) observed was remobilization on the rising limb of sediment deposited by loss of stream competence on the previous day's falling hydrograph limb. VanSickle and Beschta (1983) successfully used a similar storage function that partitioned supply,  $\Delta M$ , among several compartments,  $M_1, \dots, M_n$ , accessed at different flow levels in a non-glacierized catchment.

In summary, the suspended sediment concentration of a proglacial stream at any one time can be viewed as a function of internal and external sources of sediment. Externally, a glacier influences the suspended sediment regime as it melts and transports the products of glacial abrasion which may become deposited in a valley train or passed through without modification. Deposition or re-entrainment will depend on the particle size characteristics of the transported material as well as local hydraulic geometry. Internally, deposition, re-entrainment and erosion contribute to the



downstream suspended sediment concentration through mechanisms such as channel migration or bank collapse. Clockwise hysteresis is thought to indicate exhaustion of these internal and external sediment sources, however, hysteresis may also be indicative of differing susceptibility to erosion of internal supplies over a flood hydrograph. Clockwise hysteresis in proglacial streams can be thought of as a special case, in that, the effect may be due to a combination of supply factors internal and external to the system as well as hydraulic geometric variation in the at-a-station and long profiles over flood events.

### **2.3.3 Open Channel Bulk Hydrochemistry**

Aggregate dissolved solids fluctuations in fluvial streams often show an inverse relation with discharge during storm events that are similar to that of glacier fed streams experiencing diurnal discharge fluctuations (Hem, et al., 1990; Petts and Foster, 1985, p.83; Richards, 1982, p.95; Walling, 1977a). Individual ion concentrations will deviate from this simple relation as a function of basin characteristics and solute availability (Hem, et al., 1990; Petts and Foster, 1985, p.83; Walling, 1977a). Because specific ions deviate from bulk hydrochemical relations, Walling (1977a) suggests that conductivity trends are somewhat limited in revealing process mechanisms between storm period runoff and specific flow routing mechanisms within a catchment.

The dissolved solids concentration discharge relation is an example of the interaction of two adjacent process subsystems, reflecting both the surface and subsurface dynamics in a catchment (Gregory and Walling, 1973, p.219). Generally, the inverse relation is often a non-linear dilution process in which a near constant baseflow concentration exists at low flows and streamflow solute concentrations decline or rise when quickflow begins (Richards, 1982, p.95). This results in a chemograph trough that is not necessarily the mirror image of the hydrograph peak and consequently gives rise to hysteresis in the relationship (Richards, 1982, p.95). Clockwise hysteresis is due to readily soluble material such as evaporative deposits being flushed from the

## *2.4 Glacial Lake Sedimentation and Glacier Hydrology*

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ground surface or from chemically enriched soil moisture entering the stream on the rising limb of the hydrograph (Hem et al., 1990; Petts and Foster, 1985, p.83; Gregory and Walling, 1973, p.225). Anti-clockwise hysteresis may also occur because of a delayed throughflow effect whereby groundwater contribution and subsurface flow through the upper soil horizons has greater contributions to flow after the storm peak on the falling hydrograph limb (Petts and Foster, 1985, p.83; Walling, 1977a; Gregory and Walling, 1973, p.225).

In large river systems hysteretic effects may relate to kinematic waves, where the flood peak travels more rapidly than the water (Petts and Foster, 1985, p.83) and in small rivers relate to antecedent moisture conditions and the availability of water soluble material (Richards, 1982, p.85). Mixing models are often used to explain the variation of dissolved solids with discharge but even complex ones rarely explain more than 50% of the variation because of non-linear hysteresis (Richards, 1982, p.94).

## **2.4 Glacial Lake Sedimentation and Glacier Hydrology**

Kluane Lake is a 'distal glacier-fed lake' since it is not directly in contact with glacier ice but receives the majority of water from glacier melt. This distinguishes it from a 'glacial lake' which can be any lake of glacial origin in the past or present (Smith, 1985b). Distal glacier-fed lakes are efficient sediment sinks (Smith, 1985b) and so the nature of sediment deposition in Kluane Lake will be affected by the sediment input regime of the Slims River. The pertinent question for this research is: How does sedimentation take place in a glacier fed lake and, how could the rate of sedimentation be affected by short-term variations in glacial meltwater regimes, thus affecting paleoenvironmental interpretations.

It has been suggested that relatively large proglacial lakes trap a major proportion of the suspended sediment carried from tributary streams (Leemann and Niessen, 1994; Smith, 1985b).

## *2.4 Glacial Lake Sedimentation and Glacier Hydrology*

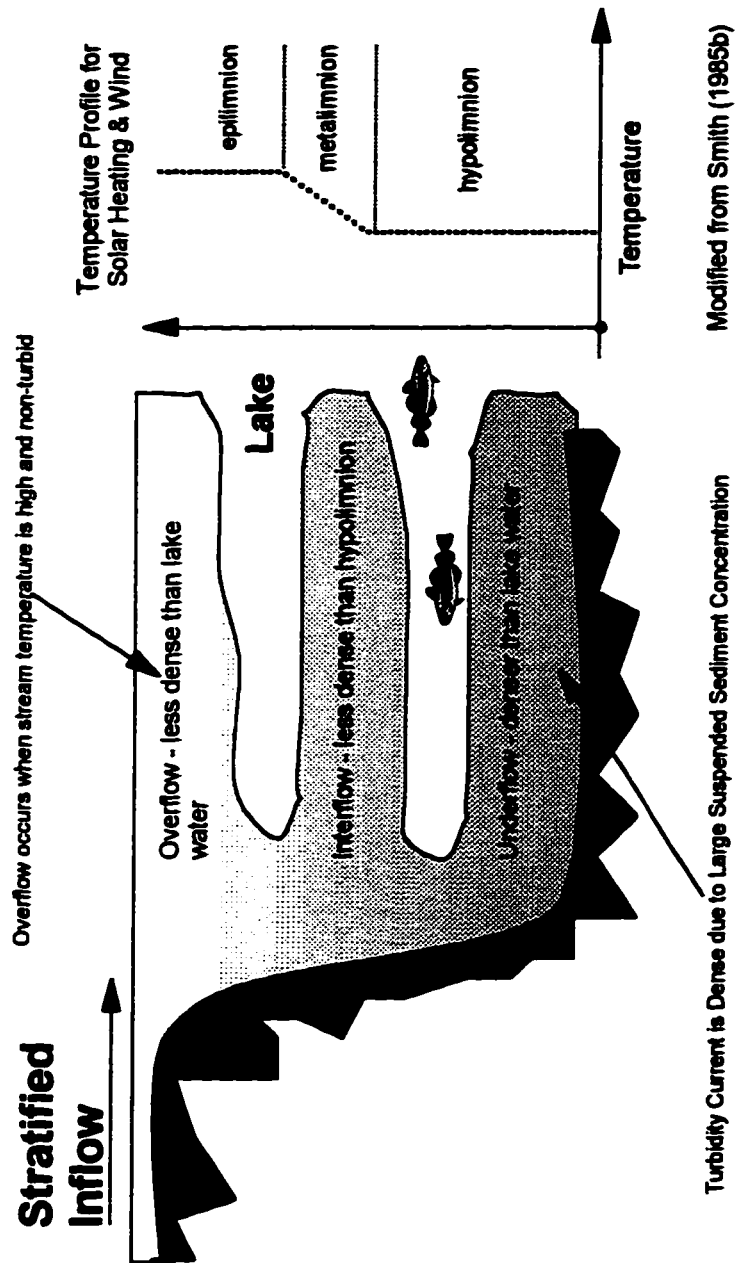
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In most cases, glacial erosion provides the quantities of silt and clay sized material necessary for the formation of varves (Leemann and Niessen, 1994). Winter accumulation in proglacial lakes is essentially negligible due to little to no flow from tributaries (Smith, 1985a) and lacustrine accumulation likely reflects the suspended sediment discharge rates from the catchment during summer, which in turn are controlled by meteorological and hydrological conditions (Leemann and Niessen, 1994). In a glacial lake, if the current is unidirectional away from the input point with no thermal or density stratification in the lake and sediment settling follows Stokes Law then lake bottom deposits beyond the limit of delta progradation will become increasingly fine towards the deepest part of the lake (Bryan 1974b). It is noticed, that this consideration is not pertinent here, however, because, in Kluane Lake there is no appreciable change in bottom sediment sizes as the distance from the input source (Slims River) increases (Bryan 1974b).

Leemann and Neissen (1994) found grainsize fluctuations in individual varves and attributed these to short-term variations in sediment input during summer caused by hot or cold weather and associated fluctuations in glacial runoff. Yearly variability in suspended sediment yield is partly a result of yearly variability in the glacier basal drainage network (Collins, 1991). Consequently, before hydrometeorological conditions are cited as causes of increased sediment accumulation, the nature of the sediment/discharge relationship should be established for a particular basin. For example, what may be construed as a warming period in a sediment record may be attributable to glacier rheological adjustments leading to high-magnitude short-term sediment discharge.

Many proglacial lakes are thermally stratified during the summer months (Leemann and Niessen, 1994; Smith, 1985b; Smith, 1981). Differences in the density between the inflowing water and that of the lake control the distribution of particles by overflows, interflows or underflows in the epilimnion and hypolimnion (Leemann and Niessen, 1994; Smith, 1981) (Figure 2.4). Underflows are controlled by bottom topography and seek the lowest parts of the lake floor (Smith, 1981).

**Figure 2.4**  
 Sketch Illustrating Principal types of Inflow to a Distal Glacier-fed Lake  
 During the Summer Months when Lake is Thermally Stratified



## *2.4 Glacial Lake Sedimentation and Glacier Hydrology*

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Turbidity underflows owe their density to their concentration of suspended sediment (Church and Gilbert, 1975) and are a major mechanism for transport and deposition of glaciolacustrine sediment in lakes (Smith, 1981; Smith et al, 1982). The most important variable in determining the type of density flow is the suspended sediment concentration of the inflow (Smith, 1981; Smith et al, 1982). Only when inflowing sediment concentrations are low do temperature and salinity make a difference in controlling the mechanism of clastic sediment dispersal within lakes (Smith, 1981; Smith et al, 1982). Rotational deflections from the Coriolis effect and wind stresses influence the distribution of sediment transported by interflows and overflows (Smith, 1981). However, Coriolis deflections are unlikely when considering the small size of Kluane Lake. Periods of overflow and interflow invariably occur at times of low sediment influx (Smith, 1981; Smith et al, 1982) and, consequently, it is important to understand the input regime and relationships between sediment concentration and temperature over time.

Contemporary reasoning proposes that sedimentation of clay size fractions takes place in calm ice-covered lakes, following Stokes Law, in winter (Leemann and Niessen, 1994) leading to the formation of varves. Contrary to this, Leemann and Niessen (1994) believe that decreasing grainsize and concentration of inflows over the summer melt season lead to varve formation rather than settling during winter. The relative contributions of either process require further study and, so, interannual variations of the varve thickness may be reflective of suspension-runoff rates from a glacierized catchment which in turn are controlled by meteorological boundary conditions (Leemann and Niessen, 1994). However, Leemann and Niessen's (1994) hypothesis could be important in glacially fed lakes that do not show distinct varving, whereby yearly sedimentation could be discernable by studying grainsize variations with depth.

Leemann and Niessen's (1994) hypothesis can, to some extent, be corroborated by current glacier hydrological evidence, namely, non-linearity and hysteresis seasonally between either dissolved

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load or suspended sediment with discharge in glacier fed streams indicate decreasing concentrations over melt seasons (Collins, 1991; Gurnell and Warburton, 1990; Collins, 1983; Ostrem, 1973). Furthermore, lower total suspended sediment concentrations (g/l) and fewer pulse events in the late melt season indicate that the cavity-conduit network has become established and exhaustion of the sediment supply therein has taken place (Collins, 1991). Decreasing grainsize of suspended material over a melt season must still be established.

Bryan (1974bc) proposes that the cold glacier waters entering Kluane Lake with their high sediment concentration should produce density currents, however, during warming periods of the Slims River this effect may be reversed. Differences in density flow must be considered in determining the progradation of sediment plumes and sedimentation in the basin as well as in choice of site for lake coring any paleoenvironmental reconstruction.

The water in the southern end of Kluane Lake is chemically similar to that of the Slims River (Bryan, 1974d). The concentration of dissolved solids in the Slims River is not high enough to affect the density of the inflow but could be used to identifying inflows once they have entered the lake (Bryan, 1974c). Bryan (1974d) suggests that since there is considerable variation in the Slims River dissolved load there should be periods when it is possible to identify parcels of the Slims River water inflowing into the lake.

## **2.5 Summary**

To summarize, in glacier dominated hydrological systems, the nature of the glacial hydrological and hydrometeorological regimes control the magnitude and timing of sediment transport and to an extent the deposition of sediment in glacial lacustrine environments. Consequently, there is a need to describe and explain at various temporal scales, the sediment, temperature and discharge

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variations in the Slims River as approximate baseline data for interpretation and paleoenvironmental reconstruction using Kluane Lake cores.

## Chapter 3

*On the rock  
waves can't reach,  
fresh snow*

Tantan (1674-1761)

### Methodology

#### 3.1 Context

This chapter describes the techniques used for investigation of the hydrological regime in the Slims River for 1993 and 1994. The choice of techniques are based on the objectives and hypotheses (Chapter 1) and review of the pertinent literature (Chapter 2). More importantly, this section attempts to give the rationale for the chosen procedures and techniques used, that is, the why's behind the how's (Mitroff and Turoff, 1973).

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## **3.2 Suspended Sediment**

In both 1993 and 1994 hourly samples of suspended sediment were collected from the Slims River at the Alaska Highway Bridge in the north channel (Figure 1.2) using an Alpha Sigma automatic pump-sampler. The pump sampler had a hose length of 4 m and was pre-purged and post-purged to ensure independence of samples. The hose nozzle was placed approximately 1.5 m from the shore at approximately 1/3 the depth. It is unlikely that the intake velocity equalled the stream velocity for all samples as is advised by Petts and Foster (1985, p.115). The stream bottom at this point is composed of rip-rap material (broken pieces of concrete and large boulders) which, as a rough bed, increases the turbulence in the reach (Richards, 1982, p.61). In spite of the fact that the sediment sampling site may not be fully in accordance with the conditions proposed by Petts and Foster (1985), processes akin to turbulent diffusion in the reach should negate some of cross-sectional suspended sediment stratification for sand-sized grains. And, as a further consideration, Østrem (1973) suggests that field experiments show that in very turbulent stream reaches suspended sediment is almost uniformly distributed in the vertical. Additionally the cross-section of the north channel did not significantly change over the monitoring period indicating continuity of transport through the reach. Consequently, the samples should be reasonably indicative of average concentration of the reach, although there is no physical evidence from the site to support this assumption.

In 1993 sediment concentrations were determined by filtering water samples through pre-weighed Whatman 44 filter papers using air pressure. These were then dried under three 100 watt lightbulbs for 24 hours. They were then re-weighed and the filter paper weight was subtracted from this to give sediment weight. In order to ascertain measurement error, samples were ashed and the resultant sediment weights compared to those measured in the field. Since measurement error will be more significant for small weights of suspended sediment the samples, and, such small weights correspond to those in the early melt season, the first 241 samples were used for

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comparison. The ashed samples had an average weight of 0.507 g and the field weighed samples 0.558 g and these weights are not statistically different (*alpha* 0.05 level with  $t = -1.535$ ). Consequently, the field weighed samples are sufficient to characterize the sediment regime in 1993.

In 1994, monitoring began later than 1993 and sediment concentrations were higher and so evaporation was used rather than filtration to determine suspended sediment concentration (Ward, 1984; Gregory and Walling, 1973). Water samples were allowed to settle for 24 hours - as this is sufficient for the settling of coarse suspensions from 2 to 1000  $\mu m$  (Ward, 1984) - and then the supernatant liquid was decanted. Following this, the samples were allowed to settle for another 24 hours and then a final decant was made with a syringe. The remaining sample was left with approximately 25 ml of supernatant liquid to avoid sample loss (Ward, 1984) and was poured into a pre-weighed, 100 ml petri dish. This dish was placed into an evaporation oven powered by two 100 watt lightbulbs that maintained a temperature between 52-65 °C. After 24 hours the samples were removed, weighed and then each sample was placed in an air-tight container for particle-size analysis. No correction for dissolved solids is needed because the suspended sediment weight is appreciably greater than the dissolved solids in the samples (Ward, 1984) with an average of 74 mg/l.

In 1994 the volumetric sample size was on average 420 ml and normally distributed. Actual sample sizes varied with a minimum of 215 ml to a maximum of 1060 ml. The variability of sample size is due to changes in hose length immersion as a result of rising and falling stage and to power fluctuations caused by changing an old battery to a new battery. There was a standard deviation of 64 ml which shows that the maximum sample size of 1060 ml was infrequent and that 99% of the samples were below 612 ml. This sample size is important as a representation of the horizontal integration of a given suspended sediment sample. Samples taken with a pump sampler

are not instantaneous and represent water withdrawal from the channel over a time of approximately 20 seconds which equates to horizontal sample integration of 10 m of water, that is, the amount of water passing the nozzle. This situation is similar to that of 1993.

Gurnell and Warburton (1990) suggest that hourly sampling may not be sufficient because it may not detect glacial sediment pulses that take place on intervals less than an hour. Turbidity data at 10 minute intervals does not exhibit any major sediment pulse events that are not also evident in the coincidental physical record. As an indicator of variation in suspended solids over time, a turbidity record taken at minute intervals for one 24 hour interval in 1994 indicated maximum turbidity variation between 03h00 and 04h00. For this time interval, the coefficient of variation suggests less than 10.8 % relative variation with 95% confidence that the 'true' mean variation lies between 11.1% and 10.5%. This variation can not be extrapolated with confidence to any given hour during the season but it does give some idea of the probable error in the physical samples, neglecting any outlying low frequency suspended concentration events.

### **3.3 Turbidity**

As a tool, turbidity can be used for many purposes including water quality monitoring, estimation of suspended solids, control of flocculation processes, and even particle size analysis (Gregory, 1984). In hydrological studies turbidity has been successfully used in open channels to monitor in-situ changes of suspended solids (Gurnell and Warburton, 1990; Humphrey, et al., 1986; Gurnell, 1982; Brabben, 1981; Grobler and Weaver, 1981). It has been recommended as the only way to accurately and reasonably depict temporal changes in suspended sediment concentration on small time scales for precise depiction of storm sediment yield (Olive and Rieger, 1988) and for detecting glacial sediment pulses or flux events (Gurnell and Warburton, 1990).

Turbidity is defined as the reduction in intensity of a beam of light passing through a suspension (Stone et al., 1993; Gregory, 1984; Melik and Fogler, 1983) and is given by,

$$I = I_0^{(-\tau L)} \quad (3.1)$$

where,  $I_0$  is the intensity of the unobstructed beam of light,  $I$  is the intensity of the light after passing through a suspension where the light path is of length  $L$  and  $\tau$  is the turbidity. Turbidity depends strongly upon the number concentration, cross-section area, spherical or non-spherical nature of the particulate, light scattering properties of the suspended particles and water colour (Gippel, 1989; Gregory, 1984). Turbidity measurements used for suspended sediment concentration can not be analytically related to particle size or volume fraction because assumptions about the suspended particles can not be maintained or for most purposes known - i.e., suspended particles in a stream are various geometrical shapes and have different light scattering properties.

Stone et al, (1993) suggest that the phototransistors have an approximate linear response for small changes over the collector-emitter voltage range. Hence, where the turbidity sensor is concerned, there is a direct relation between the light intensity  $I$  and the photo induced current  $I$ , in that the latter may be substituted for the former in (3.1). Furthermore, the photo-induced current may be rewritten using Ohm's law, in terms of voltage  $V$  and resistance  $R$ , and since the turbidity sensors operate with nearly identical components in close proximity the resistance  $R$  may be omitted, to obtain,

$$V = aI_0^{(-\tau L)} \quad (3.2)$$

where the constant  $a$  is a reference intensity correction value that is needed because of slight variations in light intensity between the light beam in clear water to the sample detector and to the internal detector. Equation 3.2 is the definition of turbidity in the current study.

One of the major problems with the use of turbidity is the expense of the sensors themselves which are subject to loss in flash events (Grobler and Weaver, 1981). Following plans by Stone, et al., (1993) a turbidity meter was constructed in little time with relatively inexpensive materials for testing in the Slims River. Ambient light in the ultraviolet can effect this particular sensor as its response wavelength is 940 *nm* (Stone, et al., 1993). Ideally, such a sensor should be housed in a length of PVC or similar material to thwart contamination. However, since water is an effective absorber of ultraviolet light, and, the Slims River is very turbid, these precautions were not considered necessary.

The turbidity sensor was placed approximately 3 m from the north channel bank and approximately 0.5 m from the bed - which in this area was rip-rap material. The sensor was approximately 2.5 m upstream of the mechanical sampler nozzle. Readings were taken at 10 minute intervals for various periods throughout the monitoring season. Additionally, readings were taken every 2 seconds for one 24 hour interval in order to give an idea for the actual variation in turbidity at small time intervals. This was done for the purpose of evaluating the possible error in taking one SSC sample each hour.

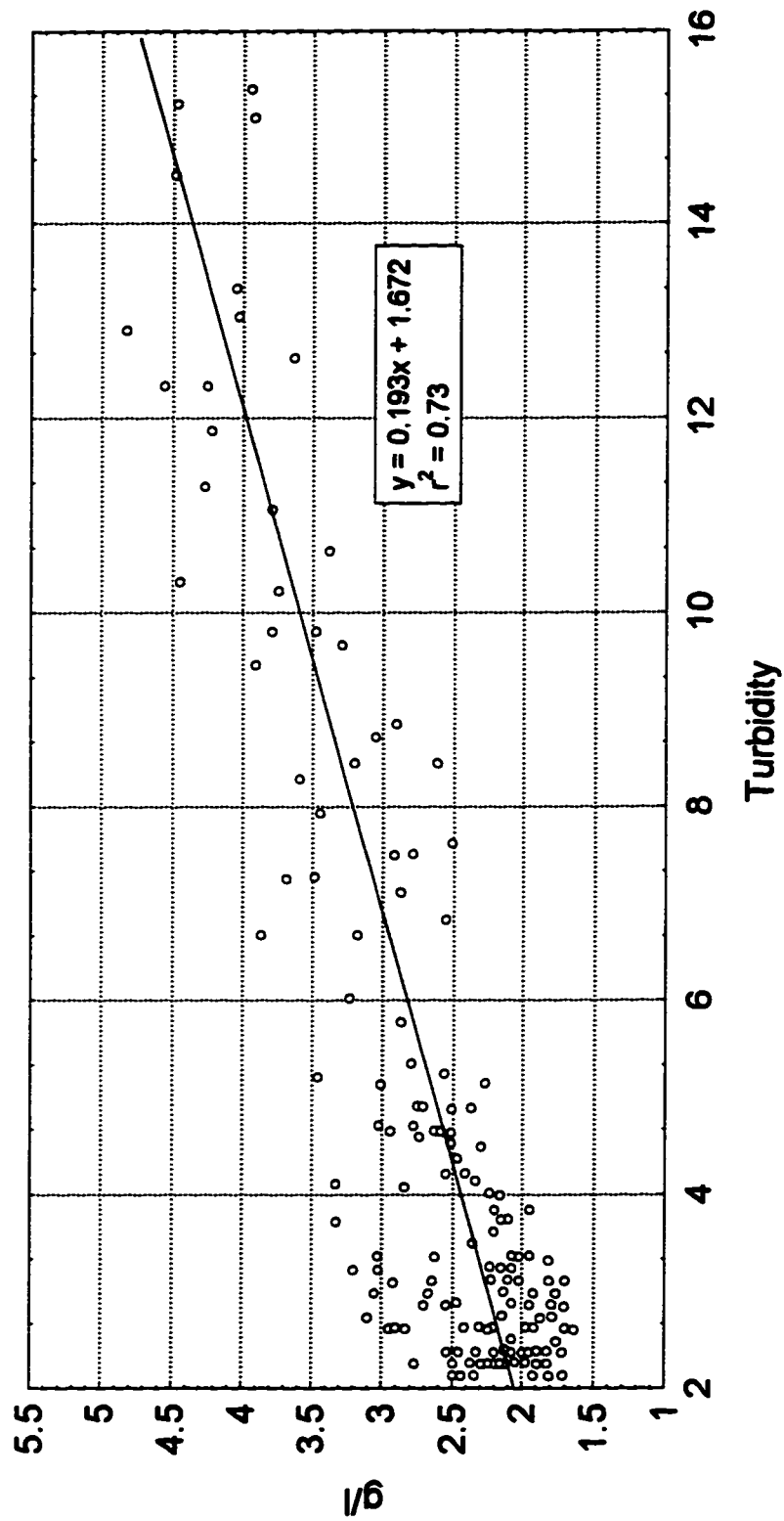
Frequent mechanical failure meant intermittent deployment of the sensor. The turbulent nature of the stream caused the sensor head to undergo constant swing (i.e., bending to and fro) and caused the single strand copper wire to break under its external sheath. Failure was so frequent that there was only a stub of wire on the sensor head by the end of the monitoring period. It would therefore be advisable to use a strong braided wire rather than single strand wire in construction. Re-soldering also changed the instrument response range presumably because of variable resistance associated with the solder connections.

Initial analysis of the relation between turbidity and suspended sediment concentration shows a residual variance of 28% from a linear relation (Figure 3.1) (Figure 3.10ab, Appendix D). Observed variance is due to a) the particle size distribution of the suspended material changing over time and space meaning constant differences in the light scattering properties of the turbid suspension (a significant portion of the suspended material of the Slims River is above 0.060 mm and thus falls above the limit where turbidity meters have been known to be stable (Brabben, 1981), which is especially true for visible light turbidity meters (Gippel, 1989) and, b) the water collected in the physical samples and that measured in the turbidity meter are not precisely simultaneous. The samplers were 2.5 m apart with the turbidity sensor deeper in the channel. As previously mentioned the pump sampler could take up to 20 seconds to extract a sample and integrated this sample over 10 m of water.

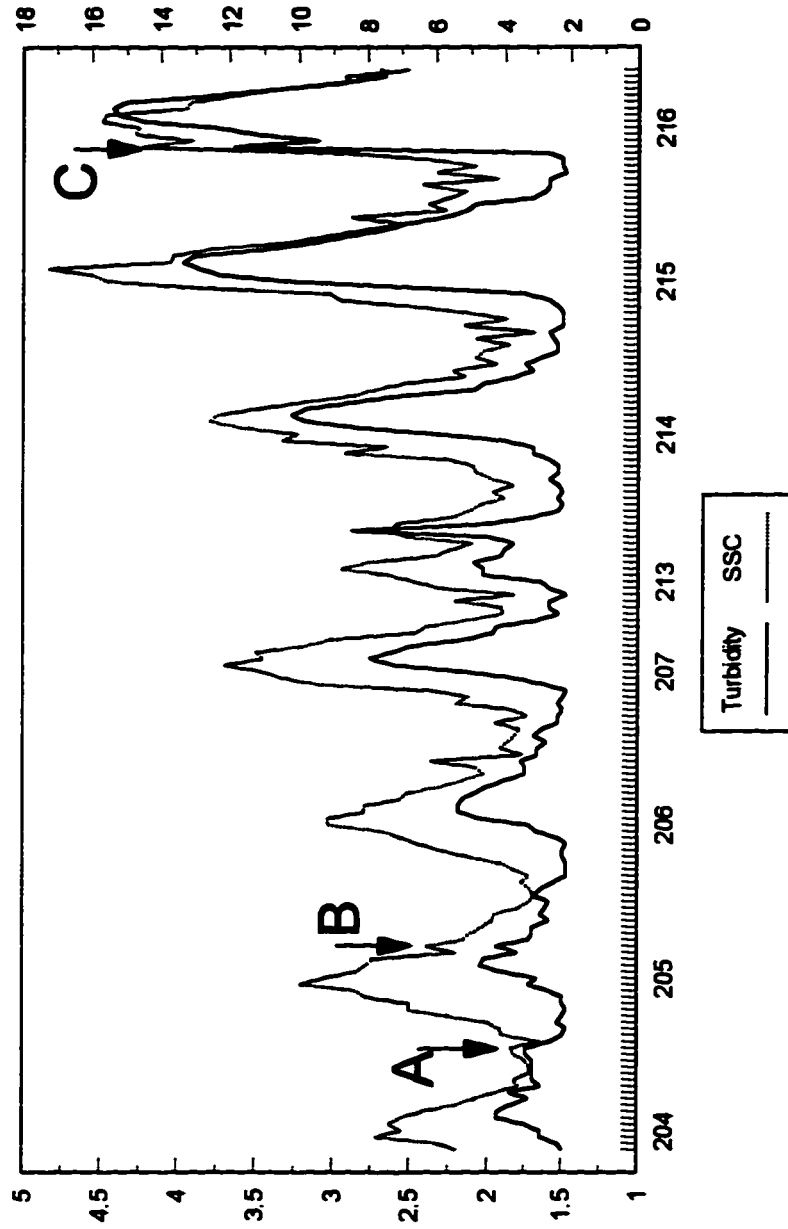
Despite the poor predictive capability of estimating concentration from turbidity, certain events clearly indicate that the turbidity meter is demonstrating the variation found in suspended sediment concentration as the time course of certain events are synchronous (Figure 3.2). Although at points the two variables are in synchrony, there is still a one or two hour time lag between peak turbidity and peak suspended sediment concentration (Figure 3.2).

Different sampling intervals as previously discussed and changing grainsize characteristics over the hydrograph may explain some of the lag, but it is beyond the scope of the present study to fully explain this phenomena. What can be suggested, however, is that if the turbidity timing is correct, then a slightly different picture would emerge regarding sediment transport over diurnal cycles. That being said, it is also noted that the lag is not large enough to negate the results that are presented with only the physical samples.

**Figure 3.1**  
 Suspended Sediment Concentration and Turbidity, Slims River, Summer 1994  
 Julian Day 203 2200h to 207 0900h and Day 212 2000h to 216 0900h



**Figure 3.2**  
Suspended Sediment Concentration and Turbidity: Visual Correlation



### **3.4 Grainsize**

Grain size analysis was performed on suspended sediment (chemically dispersed mineral fraction) and bank material to determine similarities and differences in distributions over successive diurnal cycles and the monitoring period. Analysis used a combination of sieving for coarse fraction (> 0.074 mm) and laser particle size analysis for grain-sizes below 0.074 mm. Analyzed samples were dispersed in a 10% solution of sodium metaphosphate. Grainsize was only considered in 1994 as this season has the longest continuous record of the two.

A Brinkmann PSA 2010 Particle Size Analyzer was used to analyze the volume percentage of grainsizes below .074 mm. The 2010 was fitted with a BCM-4 Liquid Flow Cell Module that facilitates the analysis of a turbid suspension prepared using a 50 ml sub-sample of a turbulent suspension taken by 10 ml splits and diluted in 1.8 l of water. The liquid borne particles were pumped through the cell and back to the sample suspension by a DIGI-STALTIC model 7518-10 pump house operating at 100 rpm passing approximately 9.8 ml/sec.

The laser particle size analyzer operates on a time-of-transition theory where the time it takes a particle moving at a fixed velocity to cross a laser beam depends on the particle diameter (Brinkmann Instruments Inc., 1989). Statistical significance for each sample is above 98% based on the sample size and variance (number of particles counted). The major drawback to the use of the laser particle size system is that it assumes a spherical particle while suspended sediments are more likely non-spherical which can introduce considerable error in the distribution (Harwood, 1977). The projected volume of the particles (assuming sphericity) is equivalent to mass or weight provided all particles are of uniform density (Brinkmann Instruments Inc., 1989). For all size class data less than .074 mm, uniform density and sphericity are assumed.

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Most fine particulate systems have particle size distributions that conform to a log-normal or Gaussian distribution (Knutson, 1977, p.17; Stockham, 1977, p.1-2). Particle size  $D$  in  $mm$  is converted to its logarithmic equivalent in  $\phi$  phi units by,

$$\phi = -1 \left( \frac{\ln D}{\ln 2} \right) \quad (3.3)$$

If logarithmically transformed particle size data plotted on probability paper as cumulative frequency yields a straight line then it is normally distributed and amenable to all statistical techniques designed for the Gaussian distribution. Most plots do not assume a continuous straight line and even a gentle curvature of the cumulative frequency plot may indicate that the sample is not a single population with a normal distribution but may be bimodal (Friedman and Sanders, 1978, p.71; Folk and Ward, 1957).

Inclusive graphic procedures are used for characterization of the samples rather than the method of moments for several reasons: first, they are easier to calculate; second, moments offers no distinct advantages over graphical techniques and may even have some drawbacks for natural sediments (Friedman and Sanders, 1978, p.74; Nickling, 1976, p.61; King, 1966, p.280); third, for comparative purposes and for building upon previous research and interpreting results there is a body of existing information based upon these techniques in general (eg., Folk and Ward, 1957), and specifically in the current study area (eg., Bryan, 1974c).

Hence, if the particle size distribution is close to normality then it can be characterized by the following graphic parameters,

$$M\phi = \phi_{50} \quad (3.4)$$

$$M_z\phi = \frac{\phi_{16} + \phi_{50} + \phi_{84}}{3} \quad (3.5)$$

$$\sigma_G\phi = \left[ \frac{\phi_{84} - \phi_{16}}{4} \right] + \left[ \frac{\phi_{95} - \phi_5}{6.6} \right] \quad (3.6)$$

$$SK_G = \left[ \frac{\phi_{84} + \phi_{16} - 2\phi_{50}}{2(\phi_{84} - \phi_{16})} \right] + \left[ \frac{\phi_{95} + \phi_5 + 2\phi_{50}}{2(\phi_{95} - \phi_5)} \right] \quad (3.7)$$

$$K_G = \frac{\phi_{95} - \phi_5}{2.44(\phi_{75} - \phi_{25})} \quad (3.8)$$

where the subscripts are percentile values read from the cumulative frequency curve in phi units as,  $M\phi$  is the graphic median (50th percentile),  $M_2\phi$  is the graphic mean,  $\sigma_G\phi$  is inclusive graphic standard deviation,  $SK_G$  is the inclusive graphic skewness, and  $K_G$  is the graphic kurtosis.

Equations 3.4 to 3.8 are given here because the measures they represent and notations used vary among investigators. These equations are given by Folk and Ward (1957) whereas the notation follows that of Bryan (1974c) and King (1966, p.280-5). These particular functions overcome difficulties with the often non-Gaussian distributions of natural sediments by considering the tails of the cumulative frequency distribution in calculations (King, 1966, p.280-1; Folk and Ward, 1957). McCammon (1962) proposes more efficient percentile measures for mean size and sorting, but for the previously mentioned reasons these are not adopted.

The mean and median are measures of central tendency, are equal in a symmetrical distribution and are sedimentologically useful as an indicator of the power of the transporting medium assuming no supply or size limitations of material. The standard deviation represents the sorting of the sample which depends on the method of transport of the sediment (Table 3.1a)(King, 1966, p.281). A well-sorted sample is represented by fewer grainsize classes than a poorly-sorted sample although bimodal samples can obscure sorting. Skewness represents the symmetry of the distribution, it is the position of the mean with respect to the median, and lies between -1 and 1 with a value of 0 indicating a normal distribution.

**Table 3.1**  
**A: Verbal Sorting Scale**

$\sigma_{G\phi}$	Sortedness
<0.35	Very Well Sorted
0.35-0.5	Well Sorted
0.5-1.0	Moderately Sorted
1.0-2.0	Poorly Sorted
2.0-4.0	Very Poorly Sorted
>4.0	Extremely Poorly Sorted

After Folk and Ward (1957)

**B: Verbal Skewness Scale**

$SK_G$	Skewness
-1.00 to -0.30	Very Negative-skewed
-0.30 to -0.10	Negative-skewed
-0.10 to +0.10	Nearly Symmetrical
+0.10 to +0.30	Positive-skewed
+0.30 to +1.00	Very Positive-skewed

After Folk and Ward (1957)

**C: Verbal Kurtosis Scale**

$K_G$	Kurtosis
< 0.67	Very Platykurtic
0.67-0.90	Platykurtic
0.90-1.11	Mesokurtic
1.11-1.50	Leptokurtic
1.50-3.00	Very Leptokurtic
> 3.00	Extremely Leptokurtic

After Folk and Ward (1957)

A positively skewed sample indicates an excess of fine particles and a negative skewed sample an excess of coarse particles (Table 3.1b). Most natural sediments do not have skewness values greater than  $\pm 0.8$  (Nickling, 1976, p.63; King, 1966, p.284; Folk and Ward, 1957). Kurtosis is a dimensionless measure of the peakedness of a curve (Table 3.1c) (Folk and Ward, 1957).

Kurtosis measures the ratio of the sorting in the extremes of the distribution compared with the sorting in the central part of the distribution (Folk and Ward, 1957). Normal curves have a value of 1, whereas an excessively peaked or leptokurtic distribution has a value  $> 1$  and is relatively better sorted in the central area than in the tails and a flatter or platykurtic distribution has values  $< 1$  and is better sorted in the tails than in the central area - most natural sediments never exceed a value of 8.0 (Folk and Ward, 1957).

### **3.5 Dissolved Load**

An aggregate measure of ionic activity or total dissolved solids is reflected in electrical conductivity measurements (Stone et al, 1993; Brown and Tranter, 1990; Collins, 1983; Steele, 1976). Rainwater and Thatcher (1960, p.275) propose that the concentration of major cations ( $\text{Ca}^+$ ,  $\text{Mg}^+$ ,  $\text{Na}^+$ ,  $\text{K}^+$ ), and anions ( $\text{HCO}_3$ ,  $\text{SO}_4$ ,  $\text{Cl}$ ) as well as other inorganic constituents can be assumed to vary linearly with conductivity if below  $10^4 \mu\text{siemen}$ . Lateral dispersion of solutes from a mid-channel source approximates a turbulent diffusion process, which means that for most river channels there is an approximate uniform concentration across a channel with less than 5% variation (Richards, 1982, p.92). This allows point samples to be indicative of mean channel concentration of solutes. Conductivity was measured hourly and continuously monitored in-situ with stream temperature at 10 *min* intervals. All conductivity values in 1994 are calibrated to a reference temperature of 25°C. This could not be done in 1993 because water temperatures were not taken with conductivity due to malfunctioning equipment.

The technique for temperature correction of conductivity values involves the measurement of conductivity at two temperatures,  $t_1$  and  $t_2$ , with the corresponding conductivity's  $L_1$  and  $L_2$  for a single sample. The sample is usually heated slightly for the second measurement. The conductivity at a reference temperature  $L_R$  is given by Gardiner and Dackombe (1983) as,

$$L_R = \frac{L_t}{1 + \alpha(T - R)} \quad (3.9)$$

where  $R$  is the reference temperature of 25 °C and  $\alpha$  is a temperature coefficient in %°C<sup>-1</sup> given by,

$$\alpha = \frac{(L_{t_2} - L_{t_1})}{[L_{t_1}(t_2 - R) - L_{t_2}(t_1 - R)]} \quad (3.10)$$

and lies between 1.5 and 2.4 %°C<sup>-1</sup>. This procedure was performed on twenty samples, an average value for  $\alpha$  was determined, and equation 3.9 was applied to the season's conductance data (Table 3.10 Appendix C).

Conductivity was measured on hourly intervals from hourly water samples at base camp using a Yellow Springs Instruments Model 33 Salinity, Conductivity and Temperature meter. In-situ monitoring used a 'home-made' conductivity sensor following plans by Stone, et al, (1993) with some minor modifications to materials. Empirical calibration of the sensor was performed at base camp using the following procedure:

1. Twenty solutions of distilled water and KCl were made at a constant temperature.
2. Each solution was measured with the YSI conductivity meter and recorded.
3. Each solution was then measured with the conductivity sensor in terms of bridge resistance and recorded.

The outcome of this procedure produced a non-linear relation between the two variables which was transformed to a linear relation by performing a natural log transform on both variables (Figure 3.3). A transfer function,

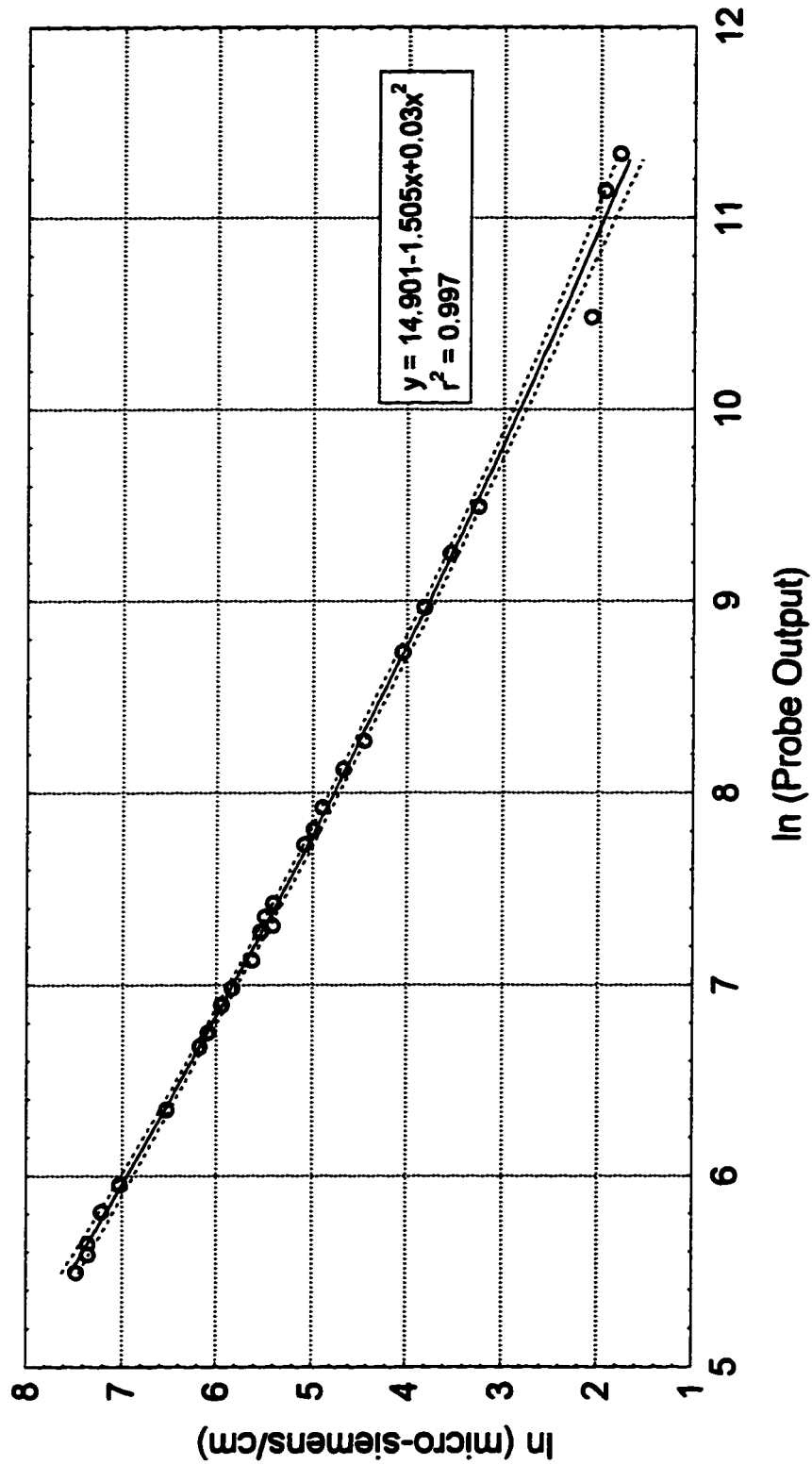
$$C_{\mu\text{siemens/cm}} = e^{14.901 - 1.505x + 0.03x^2} \quad (3.11)$$

where  $x$  is the log normal transform of the probe output, was determined by linear regression of the transformed data. Equation 3.11 was entered into the Campbell CR-10 data logger for conversion of bridge resistance to conductivity (Appendix A). The coefficient of determination between the two log transformed data sets is  $r^2 = 0.997$  and the r.m.s error is low, both of which attest to the accuracy and precision of this 'home made' instrument (Figure 3.3) (Figure 3.11ab, Appendix D).

The variation in the fit may be accounted for by the fact that the bridge resistance is measurable to 1/1000 whereas the conductivity is only measurable, from the YSI meter to at best 1/10 of a usiemen/cm. Additionally, *prima facia* there is a good fit but the response is clearly non-linear in the tails, as analysis of residuals indicates, and a quadratic fit would have been appropriate (Figure 3.11, Appendix D). However, during calibration, the observed range of conductance as read from the analogue YSI meter is greater than the r.m.s error of the calibration curve, and, consequently there is little practical error introduced by using Equation 3.11 as the transfer function, from a practical standpoint, and within the objectives of this study. Moreover, where fit error occurs is well outside of the range of observed conductance values in the Slims River.

Ideally, conductivity values would be transformed to total dissolved solids by evaporating Slims River water with known conductivity, measuring the weight of the residues and through regression determining a transfer function from the slope of this line (Gregory

**Figure 3.3**  
Conductivity Probe Calibration Function with 99% Confidence Bands



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and Walling, 1973, p.171). For most natural waters the residue upon evaporation is approximated by the conductivity of the water sample in  $\mu\text{siemen}$  multiplied by a factor of  $0.65 \pm 0.1$ . Gregory and Walling (1973, p.172) place this conversion value between 0.55 and 0.75 according to solute types. Generally, 0.65 is applicable to dilute solutions and increases as the TDS moves between 2000 and 3000 *ppm* (Rainwater and Thatcher, 1960, p.84). For this study the conversion factor chosen is 0.65, which reflects the dominance of  $\text{Ca}^+$  in the Slims River water, and, in general, has a greater rate of change of concentration with conductance (Gregory and Walling, 1973, p.170) (Table 4.1).

It is noted that the conversion factor does not yield an exact relation because the conductance is dependent on the type and total quantity of ions in solution (Gregory and Walling, 1973, p.172; Rainwater and Thatcher, 1960, p.84). Conductivity is meant as a surrogate measure of total dissolved solids (TDS) and, as a surrogate, involves measurement on a scale (conductance) that is different from the units of the target variable dissolved solids which is measured in mg/l. As a result of those 'flaws', problems associated with the use, such as,

1. variations in the surrogate will differ randomly from variations in the target variable (random error), and,
2. variations in the surrogate may differ systematically from variations in the target variable (systematic error) (Hammond and McCullugh, 1989).

Only with sufficient data on dissolved solids can estimation of the amount of random error in conductivity measurements be determined. Questions of this nature are beyond the current study with its interest in the temporal variation of dissolved solids with discharge, rather than the absolute changes in dissolved load over time. Consequently, a rough estimation of dissolved solids is sufficient for the purpose at hand.

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Dissolved ionic constituents (mainly  $\text{Ca}^+$ ,  $\text{Mg}^+$ ,  $\text{Na}^+$ ,  $\text{K}^+$ , as these are most abundant constituents in natural waters) from water samples taken each day at 1500 h were determined by Atomic Absorption Spectroscopy for the 1993 season. The 125 ml filtered samples were kept refrigerated in non-reactive ultra-violet shielded containers until they could be analysed in the university laboratory (<12 months). Similar analysis could not be performed in 1994 because of budget constraints.

### **3.6 Bedload**

The portion of total load transported as bedload has been estimated at 10% for alluvial channels (Richards, 1982, p.106) and may comprise as much as 40 or even 65 percent of total load (Bajewski, 1985; McPherson, 1971) in some channels. Bedload is the portion of total load supported by frequent contact with the unmoving bed or moving on or near the stream bed by rolling or saltation rather than in the bulk of the flowing water (Emmet, 1981). Problems with physically sampling bedload has largely meant its exclusion in most monitoring strategies in glacial streams (Bajewski, 1985; Smith, 1985), including this one. However, at the sampling site there is the possibility, if not evidence, that much of the bed material that passes through the monitoring reach may be thrown into suspension as the melt season progresses and baseflow increases. Such a condition would make measurements of suspended load more indicative of total load (Leopold, Wolman and Miller, 1964, p.185).

The velocity of water at the Slims River bridge, because of the constriction of flow, may carry in suspension what was previously carried as bedload. It is not uncommon, even in sand bed channels, for turbulence in constricted reaches to cause most of the bed material to go into suspension (Richards, 1982, p.106; Ward, 1984, p.39; Leopold, Wolman and Miller, 1964, p.185). The Slims River is shallow, broad and braided along its 22 km course from the Kaskawulsh terminus to the Slims River bridge. The hydraulic geometry at the Slims River bridge

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is drastically different from that of the upper stream. At the bridge, flow is constricted to two channels totalling 52 m in width. Water that was spread out in many channels over a sandur (Figure 1.2) with a width of 1 km is now discharged through two fixed channels of 20 m and 32 m in width. Consequently the hydraulic geometry of the stream must change to accommodate this water and imposed sediment load. Namely, there will be a change in width, depth stage and velocity to accommodate the upstream input. In this study, width and depth are relatively fixed at the monitoring reach after day 161 as the stream is scoured to a rip-rap bottom, and, although stage continues to increase, it is the third variable, velocity, which becomes the controlling factor in sediment transport. These observations are explained in detail in Chapter 4.

The increase in velocity equates to an increase in stream power and the ability of the water to do work in transporting sediment (Richards, 1982, p.13). The high turbulence observed in the reach will encourage momentum transfer of velocity between layers and bring rapid increases of velocity close to the bed (Richards, 1982, p.69). Because of this turbulence there is a rapid increase in shear stress (Petts and Foster, 1985, p.98). Furthermore, at increasing levels of shear stress the concentration of particles of a particular fall velocity are more likely to become uniformly distributed throughout the vertical (Leopold, Wolman and Miller, 1964, p.182). This means an increase in the competence and capacity of the stream to transport more sediment and larger grain sizes in suspension, which leads to the research question: Are the observed conditions able to produce the shear stresses required to suspend the clast sizes observed in bed and bank material? This question is pursued in Chapter 4 where the results of the investigation are presented.

Fahnestock (1969) found that in the Slims River the slope of a reach appeared to be inversely related to the efficiency of sediment transport through the reach (Fahnestock 1969). Materials transported in the form of large bars through a wide reach with high slope were transported with no bar formation by narrow reaches with a fraction of the slope that was required in the wider

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reaches with many more active channels (Fahnestock, 1969). This was particularly evident where alluvial fans constrict the Slims River and there is a decrease in the slope of the valley floor (Fahnestock 1969).

### **3.7 Discharge**

Discharge measurements were taken at the Slims River bridge 23 times in 1993 and 20 times over the sampling season of 1994. Velocity measurements were obtained every two meters along the channel beneath the Slims River bridge at vertical intervals of 0.6 of the depth.

At 0.6 of the channel depth the local velocity equals the mean velocity, assuming that the stream velocity is proportional to the logarithm of depth in any vertical. This assumption is necessary because the nature of the relation between velocity and depth in vertical is unknown, so, it follows that unknown error is admitted. Factors such as the extreme turbulence observed in the measurement reach likely allows considerable momentum transfer of velocity *via* turbulent eddies causing a steeper velocity gradient (Richards, 1982, p.68) and complicating the 0.6 depth rule.

Velocity measurements were made using an electrostatic velocity meter at two meters horizontal intervals and multiplied by the channel cross-sectional area given at the same two meters horizontal intervals obtained using a Raytheon sonar system in 1994 and weighted cable marked at 0.5 m intervals in 1993. Gregory and Walling (1973, p.171) suggest that in order to acquire adequate representation of bed elevation and horizontal velocity flux the interval between any two measurement verticals should not exceed 5% of the channel width. Because of inaccuracy in reading the analogue output of the sonar meter ( $\pm 0.25 m$ ) and for ease and consistency of measurement at equal intervals, ten and six percent of the width in the north and south channels, respectively, is accepted as an adequate representation of the cross-section in this study for both seasons. An *Ott type X* stage recorder was used to continuously trace water stage. Observations

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of water level were made daily from a fixed staff, and these were registered on the stage recording sheet in order to transfer stage to standard metric units.

Separate rating curves for each season are employed as transfer functions of stage to discharge for two reasons: first, stream geometric changes between the two seasons make a combined set less attractive, and, second, depth was not measured consistently by bathymmetry in 1993.

### 3.7.1 Season of 1993

In 1993 a third degree polynomial,

$$Q = 28.637 + 67.946s + 35.338s^2 - 30.552s^3 \quad (3.12)$$

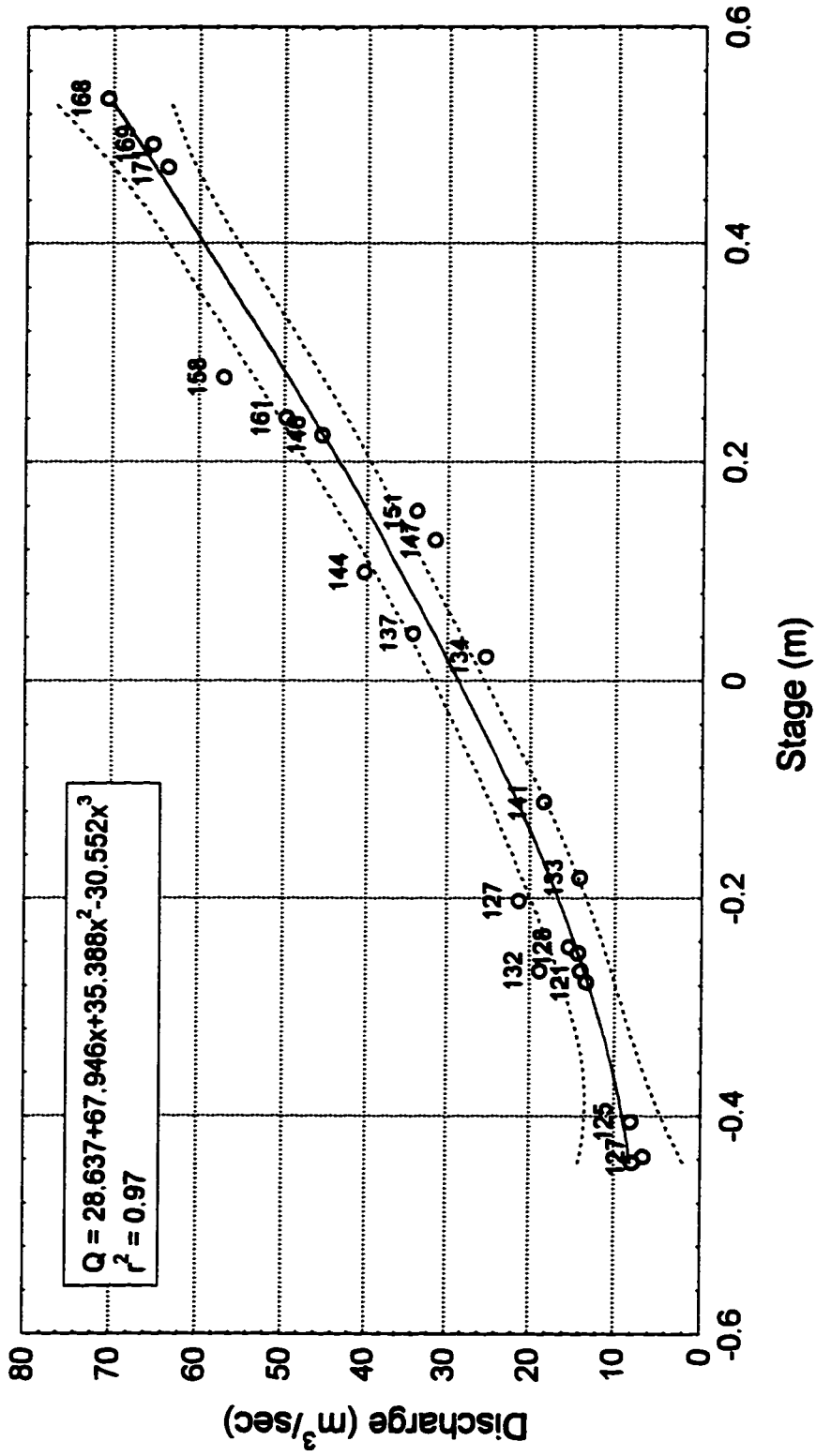
$$\{-0.4429m \leq s \leq 0.53361m\}$$

$$\{7.88m^3/sec \leq Q \leq 70.64m^3/sec\}$$

characterizes the stage-discharge relation where  $Q$  is discharge and  $s$  is stage in m (Figure 3.4a) (Figure 3.12ab, Appendix D). This degree is necessary to explain 97% of the variance and reasonable considering that measured stages are no more than  $\pm 0.03$  m outside the function domain. There was a decrease in adjusted  $r^2$  values with higher polynomials.

The major problem with the stage discharge relation in 1993 is the fact that at the beginning of the season the river was still ice covered and, during breakup, stage was initially high due to backwater effects far upstream because of local ice jamming and fast-ice conditions. Hydraulic conditions during ice break up are complex (Gerard, 1990; Belatos, 1983) and beyond the scope of this thesis. However, breakup and ice-cover conditions meant that discharges could not be measured with stage before day 121. Hence, because of high initial stages the discharge appears to be initially high but is actually low from day 109 to 121 (e.g., Figure 4.21a). Since discharge

**Figure 3.4a**  
Stage and Discharge Rating Best-Fit Curve, Slims River 1993



readings were not taken until day 121, none of the calculated discharges before that date are used in numerical analysis in this thesis. Although the values of discharge are not correct at this time, the regime in terms of periodic cycles is still valid for visual analysis.

### 3.7.2 Season of 1994

In 1994 a second degree polynomial relation,

$$Q = 105.2s^2 + 138.6s + 26.3 \quad (3.13)$$

$$\{-0.20m \leq s \leq 76.25m\}$$

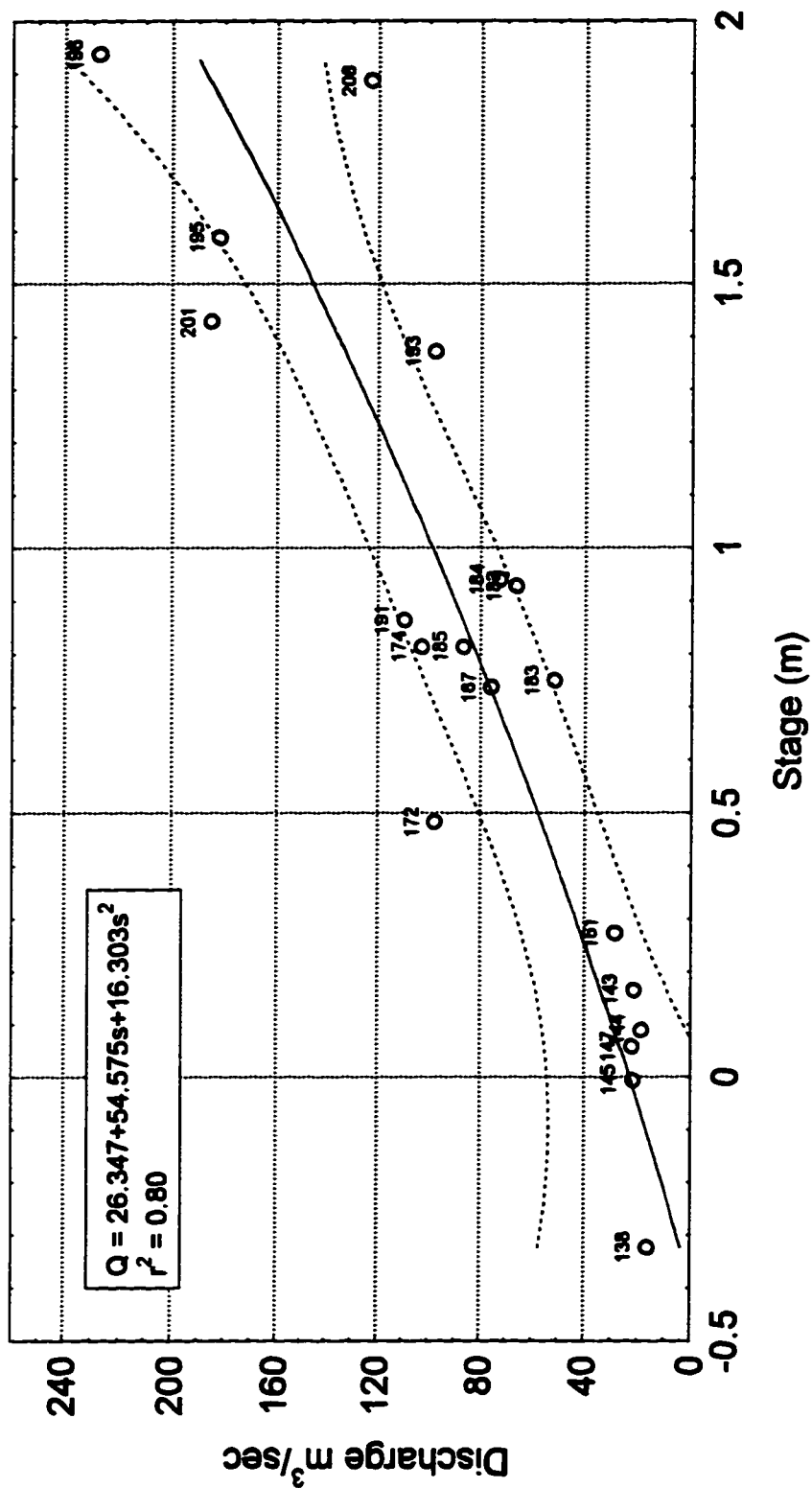
$$\{13.22m^3/sec \leq Q \leq 227.80m^3/sec\}$$

where,  $Q$  is discharge and  $s$  is stage, is used to describe the relation between stage and discharge (Figure 3.4b) (Figure 3.13ab, Appendix D). Equation 3.13 was used to transfer stage to discharge for the entire monitoring period and, in effect, is the definition of discharge for this year.

At best only 80% of the variation in discharge could be explained by stage as the only independent variable, and, because of the 20% residual variance the confidence bands are quite wide at the 99% level (Figure 3.4b).

A second degree polynomial was selected as the transfer function for two reasons: first, there was no bed accretion during the later season when stage and velocity were continuing to increase, and, therefore, extrapolation was necessary outside the function domain because it was safe to assume that discharge was increasing; and second, linear, cubic and quartic equations did not give reasonable values when their curves were extrapolated to values outside the function domain, being either too high or too low at the highest stages observed.

**Figure 3.4b**  
 Stage and Discharge Best-Fit Curve, Slims River 1994

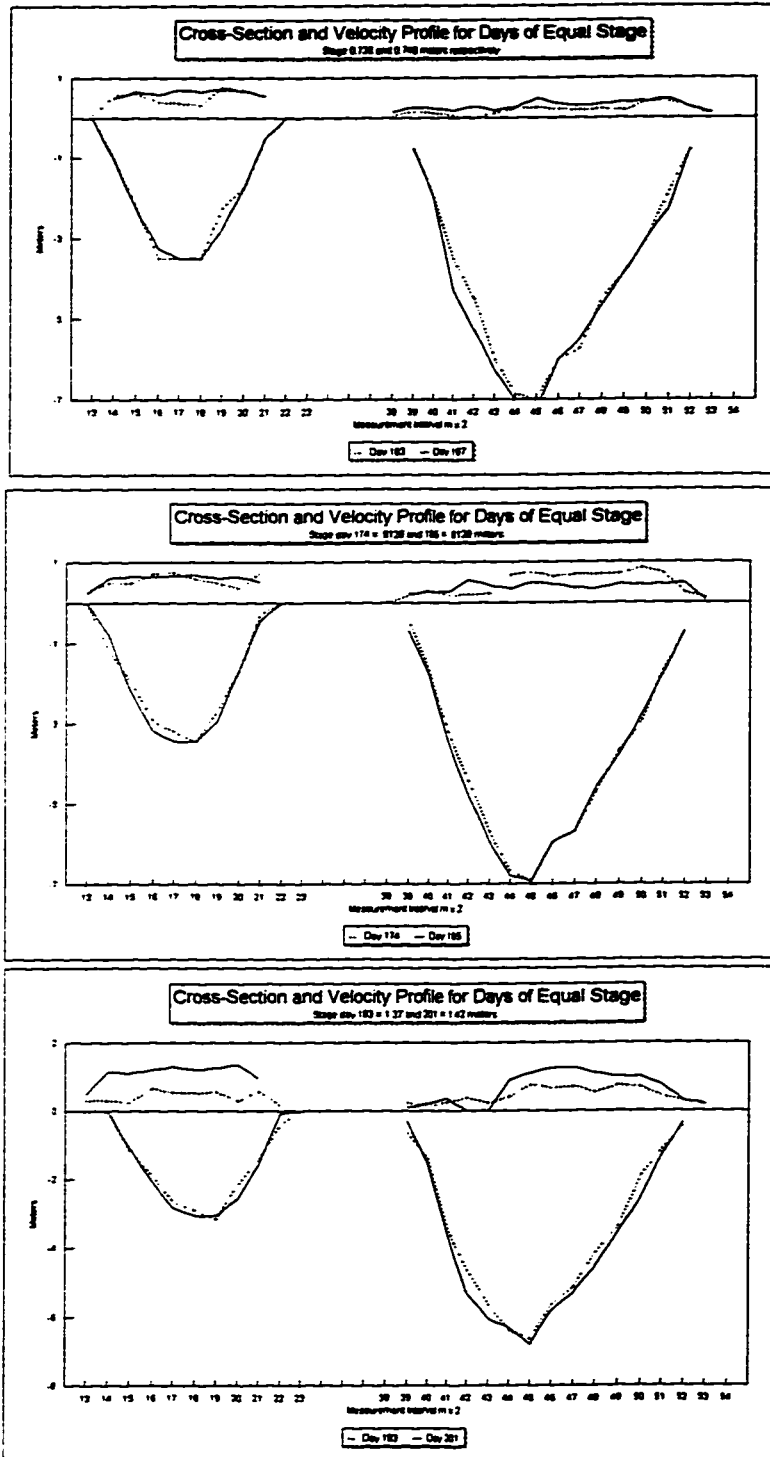


Discharge variability at similar stages increases with increasing stage. Although this may be misleading as there are no measurements of discharge at equal stages before day 161. The major error is that stages in the interval  $-0.43m \leq s \leq 1.14m$  are respectively 0.23 meters less than and 0.75 m greater than the lowest and highest stages (function domain) for which there are discharge readings. Consequently, extrapolation to observed stages is less rigorous when compared to that of 1993 and so some error is admitted.

The observed residual variance in the stage discharge relation has two main explanations. First, changes in the at-a-station hydraulic geometry were minor for width and depth (or at least at an undetectable level in the north channel) but more apparent for velocity. This is illustrated by the variation of velocity measurements at similar stages (Figure 3.5). Second, increasing baseflow generated extreme turbulence in the monitoring reach producing considerable variability in velocity measurements taken with the electrostatic current meter.

Turbulence beneath the bridge introduced considerable error in stage discharge relations in previous studies (Johnson, 1991b). Velocity readings often had to be visually averaged at certain points in the cross-section after day 161. For equivalent stages there is approximately a difference of 80 m<sup>3</sup>/sec in flow in the later season which is 41% greater than the average diurnal discharge amplitude range. This, seemingly, is caused solely by differences in velocity measurements. It would seem unlikely that velocity changes alone could account for changes in discharge greater than the average diurnal variation especially considering that velocity is not independent from discharge. In other words, for velocity to account for such changes in discharge there must be commensurate change in either width or depth. These factors can account for a large proportion of the variation in measured velocities and calculated discharges at similar stages.

**Figure 3.5**  
**Cross-Section and Velocity Profiles for Equivalent Stage**



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Although a stage - discharge rating curve is empirical in nature, the relation between the two variables is physically controlled. The residual error is indicative of a non-linear and possibly hysteretic component, but the data are lacking for the exploration of the source of the residual error. In absence of conclusive physical evidence to the contrary, and considering the error in velocity measurements as the season progressed, the error introduced by choosing a monotonic least-squares curvilinear function to represent the stage - discharge relation is equal or possibly greater than what would be introduced by eliminating points and using a step-wise function. Using a continuous function also allows continuity between the variation in stage and mean discharge variation over the entire season rather than only parts thereof, in that, we know there is a fixed geometric cross-section in the later 1994 season, and, with continuous increase in stage, it is assumed that discharge should follow likewise. Consequently, the subjective decision of when to use different rating curves, that is, a stepwise relation, for certain parts of the season can be avoided.

The better explanation of variance in 1993 may be due to the fact that channel cross-sections were measured by cable and weight which is not as accurate as sonar. Moreover, larger and more turbulent discharges were not encountered during the 1993 monitoring period because the season ended before July when greater discharges and more turbulence would have occurred.

## **3.8 Stream Temperature**

Stream temperature was continuously monitored at the same intervals as turbidity and conductivity using thermocouples. The thermocouples were connected to a Campbell CR-10 datalogger which provided a panel reference temperature designed for use with thermocouples - the program used can be seen in Appendix A.

### **3.9 Meteorological Data**

Site meteorological data were taken at the Slims River bridge and at the Arctic Institute of North America base camp for May, June, July and early August using Meteorological Research Incorporated (MRI) Mechanical Weather Stations. The data consist of precipitation, wind direction, windrun, short-wave solar radiation and temperature.

There is no *a priori* reason to assume that temperature monitored at base camp or the Slims River is indicative of high altitude regimes. However, it is postulated, as a best approximation, that lower altitude temperatures will follow the general tendencies of upper valley and higher altitude regimes over the monitoring period.

### **3.10 Other Measurements**

Ideally this field programme should have been implemented on all tributary streams to the Slims River. This was not the case for two reasons: first, past research has demonstrated that the high glacial contribution to discharge during the summer melt season transcends the effects that tributaries have on the discharge regime (Barnett, 1974; Bryan, 1974ab); and second, it was neither feasible nor an operationally reasonable undertaking with a field team of two persons.

### **3.11 Summary**

The techniques used in this study are aimed at the precise determination of the short-term suspended sediment and bulk hydrochemical regimes of the Slims River. To this end, field parameters measured included both suspended sediment concentration and conductivity - as a surrogate of total dissolved solids - as well as air and water temperatures. Turbidity was monitored in order to test its plausibility as a surrogate measure for suspended sediment concentration. The 'home made' turbidity sensors used are inadequate as a means of

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determining absolute quantities of sediment in suspension. However, these sensors show promise in monitoring relative changes in suspended sediment concentration with a minimum investment of time.

Filtration and evaporation were the two techniques used to physically measure suspended sediment concentration. Evaporation, was used in order to maximize the amount of processing that could be done in the field and shows no statistically significant difference in estimation of quantity when compared to filtration and ashing.

Extensive use of bathymetry for measuring channel cross-sections in 1994 allowed more precise determination of the hydraulic geometry while at the same time determining the dynamic nature of bed changes and improving estimation of discharge. With more precise channel cross-sectional areas, the major problem with the estimation of discharge in 1994 became water turbulence induced error on the electro-static current meter with increasing discharges accompanying glacier melt.

Finally, air temperatures were not measured synchronously with conductivity in 1993 and this, therefore, made calibration to a reference temperature impossible. This was not a problem in 1994 because water temperatures were measured at base camp with conductivity samples. Additionally, air and water temperature were measured *in situ* with a 'home-made' conductivity sensor. This sensor showed sufficient results for estimation of total dissolved solids and had the advantage of a minimal physical sampling effort.

## **Chapter 4**

*Never forget:  
we walk on hell,  
gazing at flowers.*

Issa (1763-1827)

### **Results**

The field season of 1994 was from May 11 to August 5, and that of 1993 was from April 19 to June 24. The data set for 1994 is more detailed and therefore is the main focus of this section. Where present, data for 1993 are discussed in comparison to those of 1994.

## 4.1 Discharge

### 4.1.1 Components

Over both monitoring periods, diurnal variation in discharge approximates a sinusoidal function with amplitude  $A$  and period  $B$  superimposed on mean discharge variation  $f(T)$ . Variation of discharge,  $b = f(T)$ , is defined as the mean daily discharge (e.g., discharge with the diurnal periodic component removed). It is evident that discharge variation,  $b$ , responds to long term temperature (temperature over a number of days),  $T$ , variations over the season, while discharge amplitude responds more directly to daily temperature variations. Therefore, discharge for any given time within the season is,

$$Q = f(t) + A \sin B(t - t_o) \quad (4.1)$$

where  $t$ , is time in hours,  $t = 0$  is day 126, and  $t_o$  represents time of maximum daily discharge on a given day,  $A$  is the diurnal discharge amplitude, and  $B$  is the diurnal discharge period. Since the discharge period,  $B$ , (length of time it takes for discharge to go from low to high and back to low) is almost constant at 24 hours over the season,

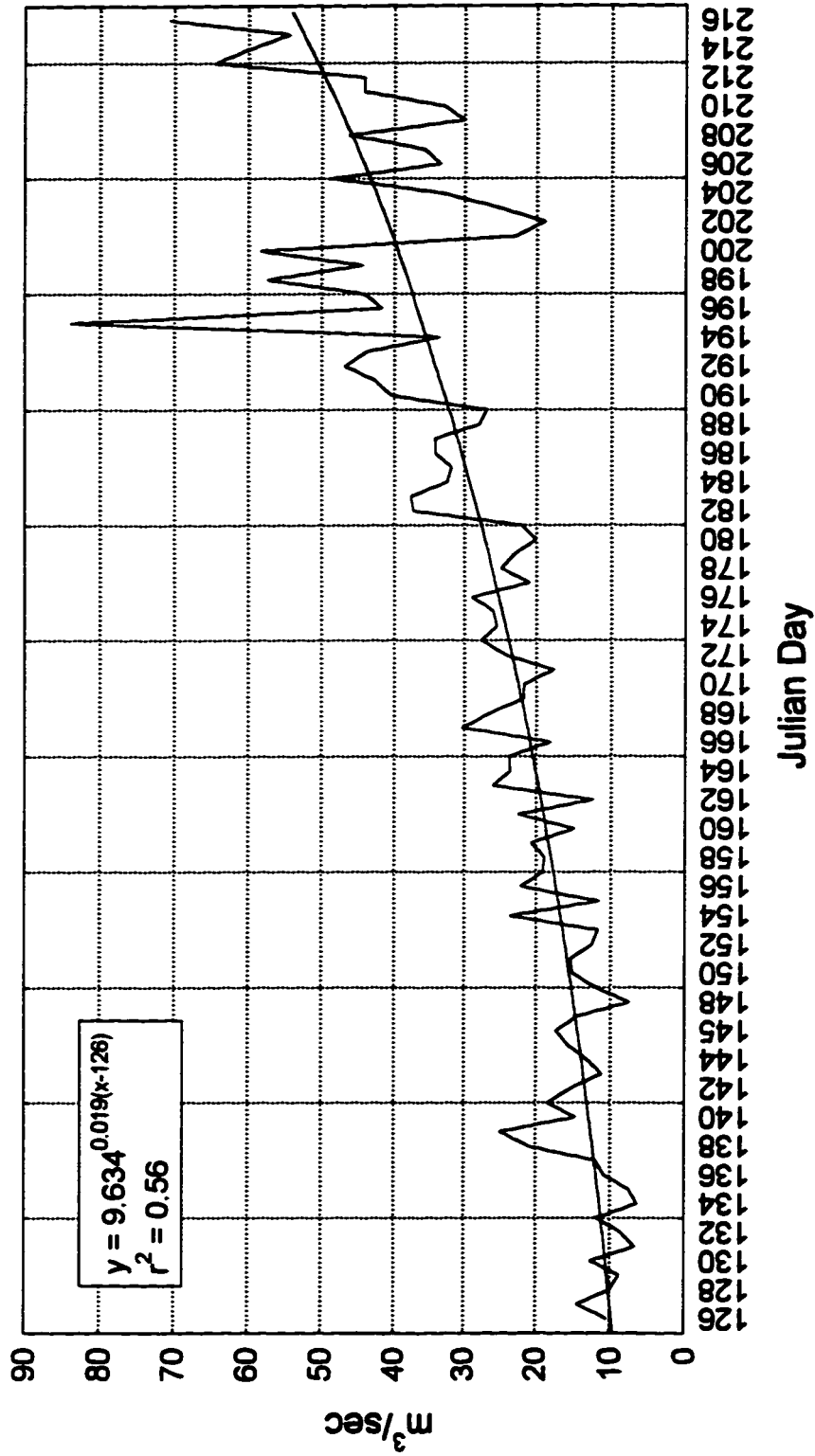
$$Q = f(t) + A \sin\left(\frac{\pi}{12}(t - t_o)\right) \quad (4.2)$$

The amplitude,  $A$ , increases over the monitoring period approximately exponentially<sup>1</sup> and can be given by  $A = (9.634)^{0.019t}$  (Figure 4.1), so discharge becomes,

$$Q = f(t) + \left(\frac{1}{48} 9.364^{0.013(t-126)}\right) \sin\left(\frac{\pi}{12}(t - t_o)\right) \quad (4.3)$$

<sup>1</sup> That is,  $A$  increases exponentially only because monitoring did not continue past peak discharge. Therefore,  $A$  increases exponentially only over the monitoring period.

**Figure 4.1**  
 Seasonal Diurnal Discharge Amplitude Changes, Sims River, 1994



which makes explicit the components of discharge which are allowed to vary over the season.

Each component of discharge is a composite but correlations fail to provide a strong foundation for prediction and modelling of discharge.

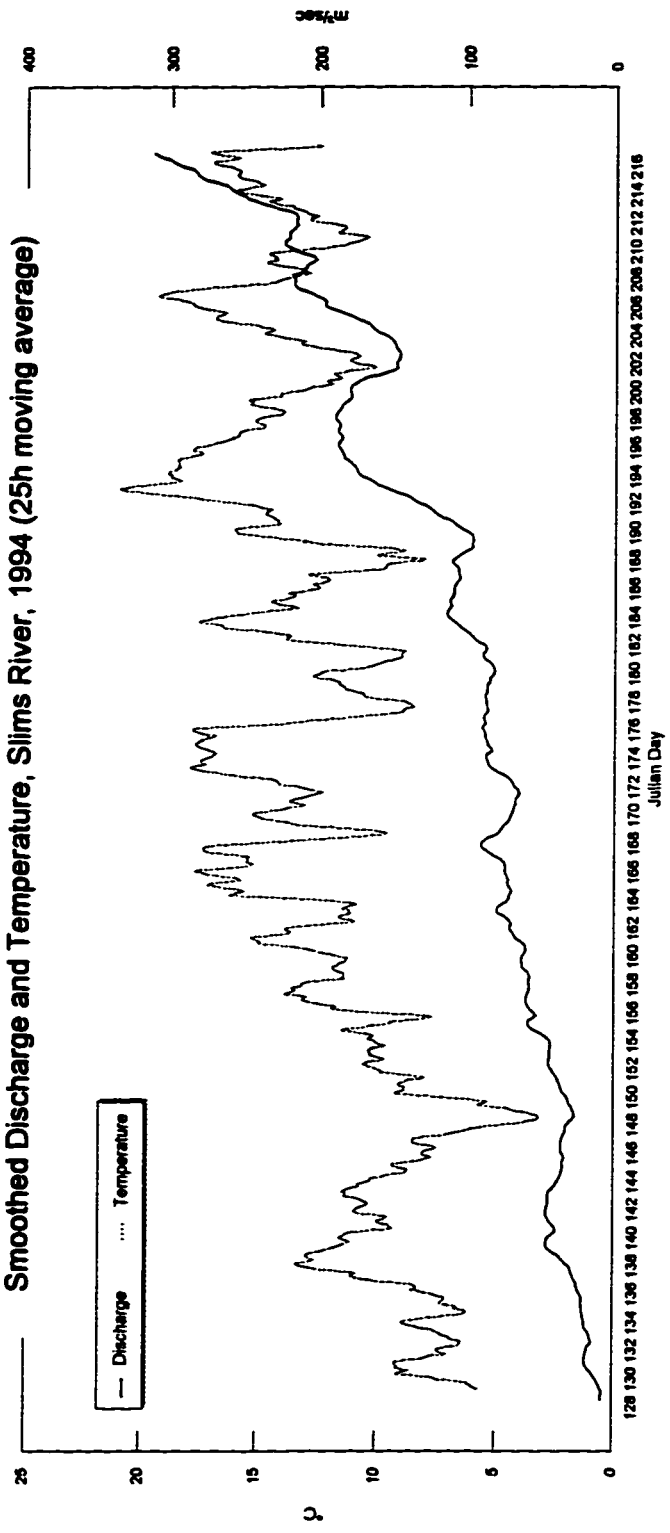
### **4.1.2 Discharge Amplitude and Temperature**

By way of illustration, decreased discharge amplitude during periods of sustained ablation was previously observed to take place when Slims River drainage shifts to the Kaskawulsh River (Figure 1.1) (Section 1.4). Barnett (1974) attributes a sudden drop in stage during a period of sustained ablation in 1970 to such a drainage shift, at which time daily hydrograph peaks and troughs were attenuated due to differential meltwater travel times from tributary outputs to the gauging station. *Prima facia*, a similar event takes place between Julian day 201 and 205 (Figure 4.1; Figure 4.30, Appendix B). However, this event is clearly preceded by a temperature trough (Figure 4.2a) and there is no attenuation of hydrograph peaks. The preceding temperature trough is not an outlying event and similar events in the early season do not produce similar discharge change. The explanation for this event is as follows: After snowmelt has ended, more clear glacier ice is exposed and, for similar temperatures, melt is amplified and therefore, discharge is more sensitive to temperature variations. However, considering that discharge was significantly rising previous to this event, increasing channel volume (by channel widening or piracy) could account for the decrease because of internal valley train storage-outflow adjustment. Similar events, though not of this magnitude, are clear in the 1993 record (Figure 4.2b).

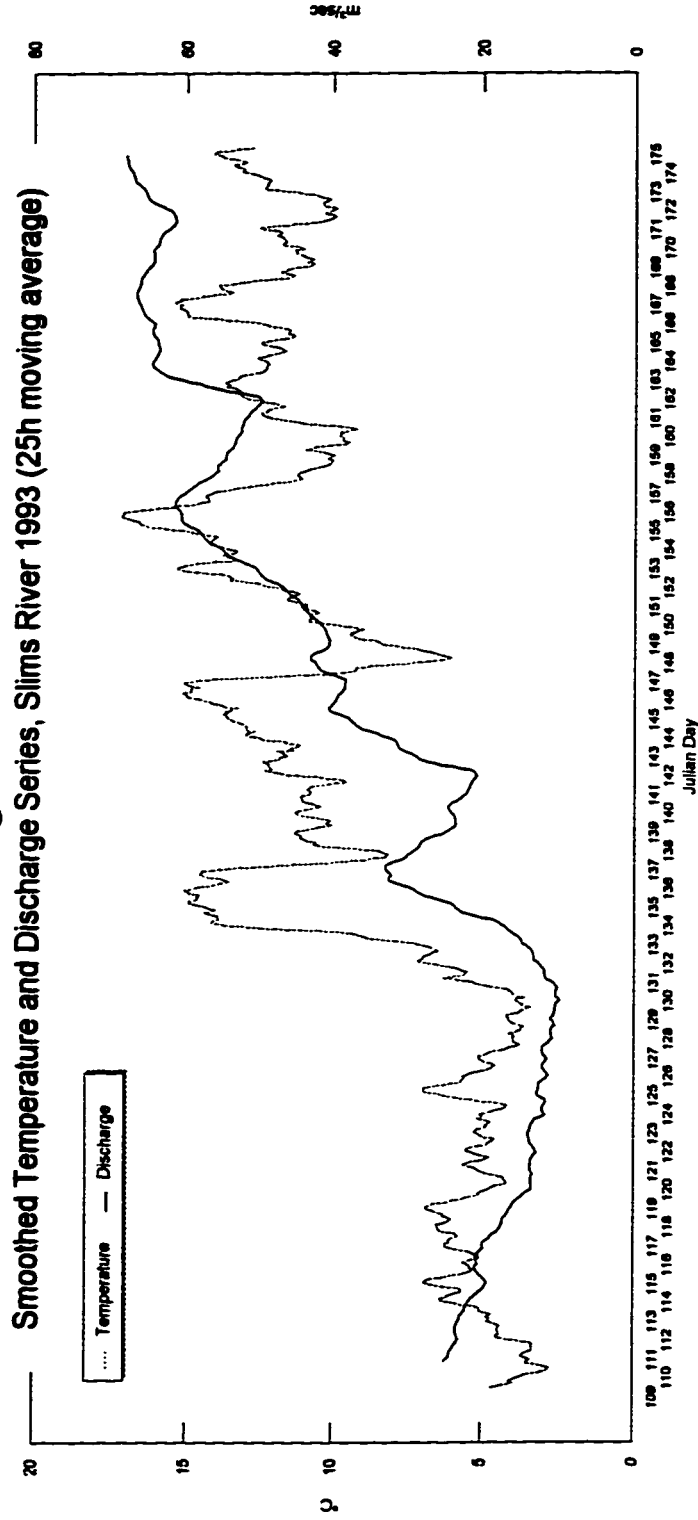
### **4.1.3 Temperature and Discharge**

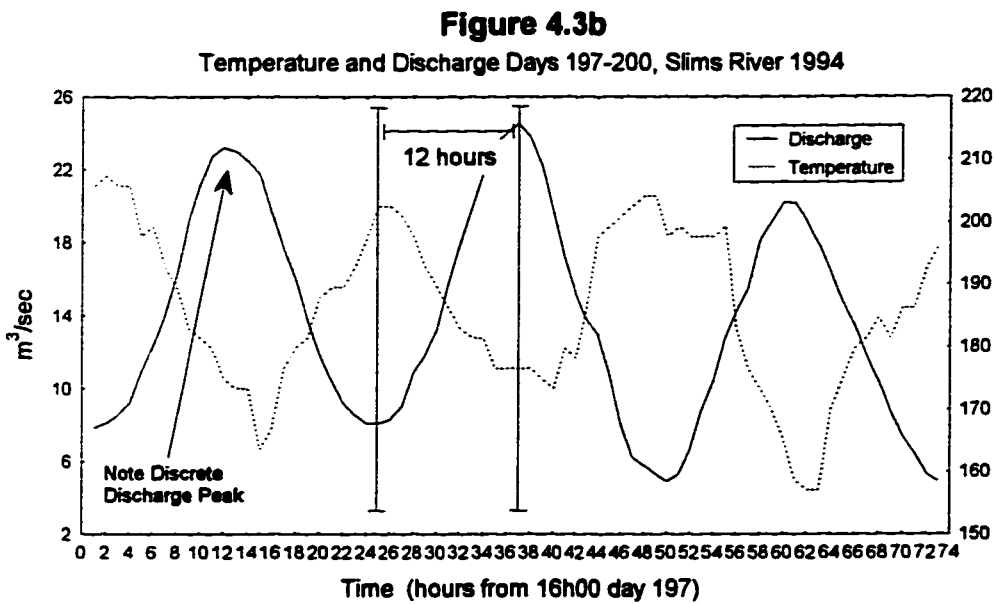
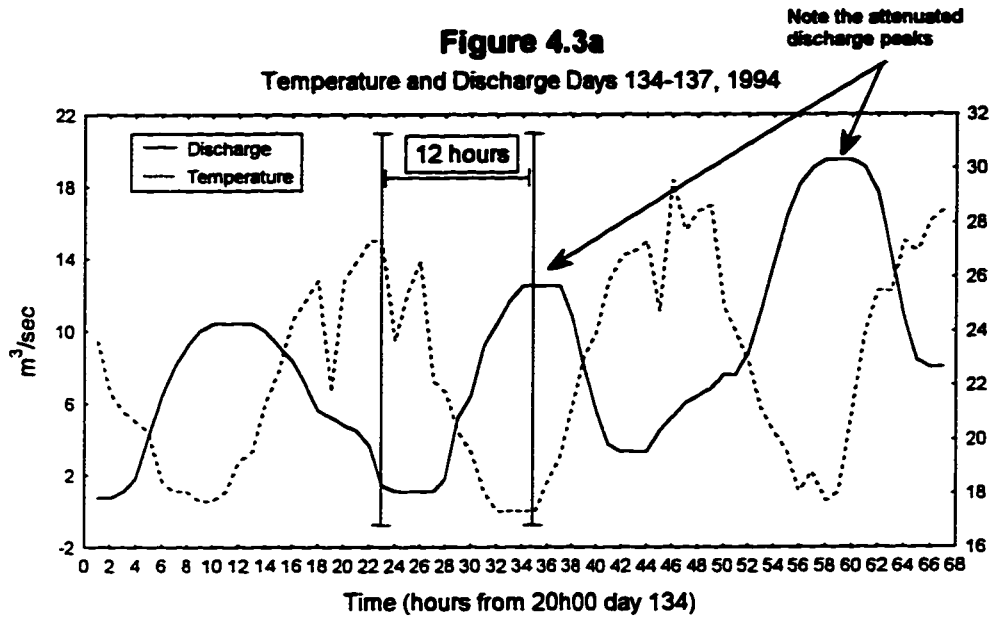
In both seasons there is a pronounced phase shift between temperature and discharge (Figure 4.3ab).

**Figure 4.2a**  
Smoothed Discharge and Temperature, Silms River, 1994 (25h moving average)



**Figure 4.2b**  
Smoothed Temperature and Discharge Series, Silms River 1993 (25h moving average)





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In 1993 there is a 10 hour time-lag between maximum discharge and daily temperature (Determined by lagged correlation) (Figure 4.4). In 1993, with a 10 hour time-lag, hourly temperatures explain at best ten percent of the variation in hourly discharge, however, mean daily temperature explains fifty-three percent of the variation in mean daily discharge.

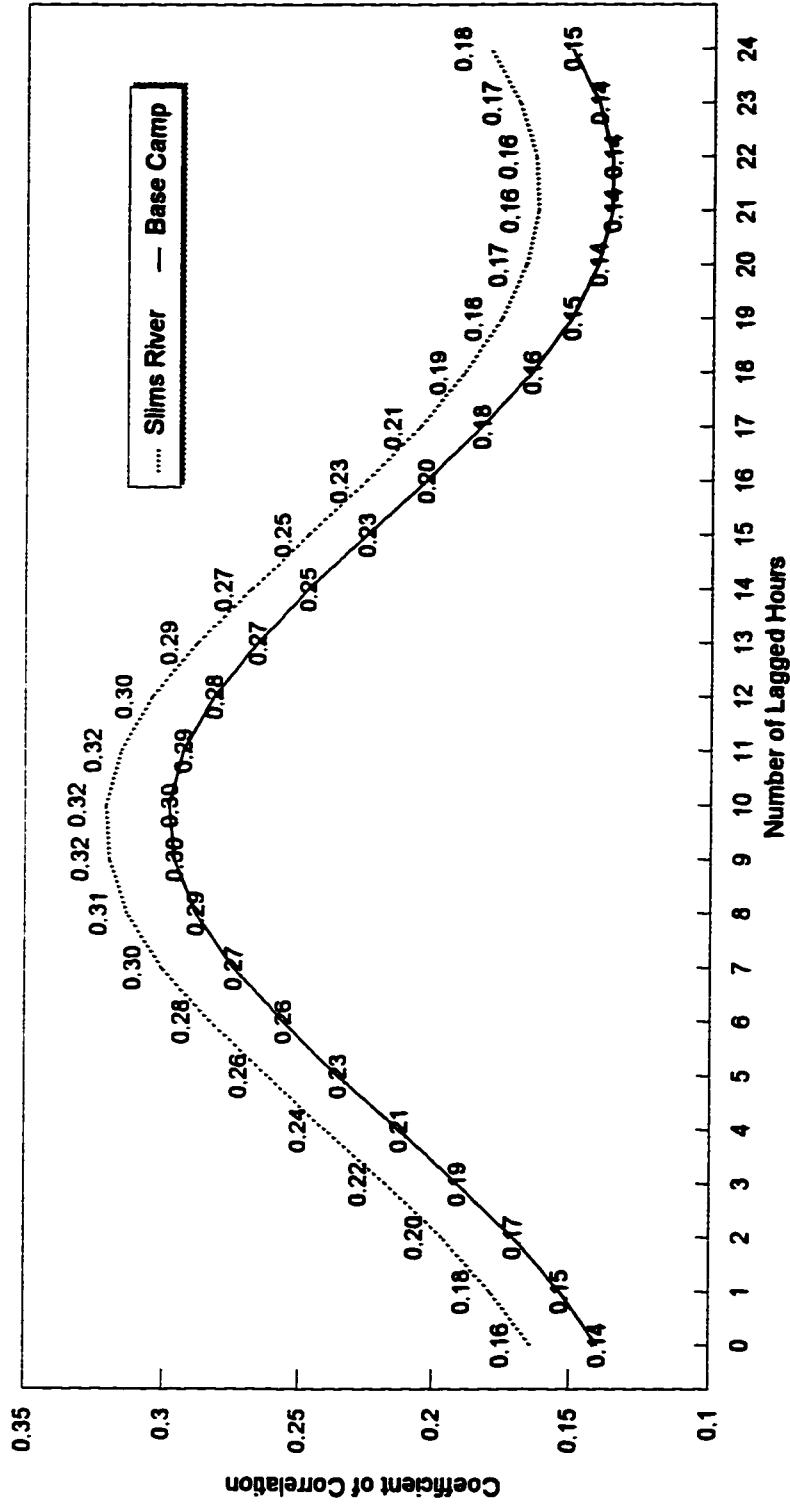
In 1994 there is an eleven hour time-lag between maximum daily temperature and discharge (Figure 4.5a). Hourly temperature data for 1994 with an 11 hour time lag explains 19% of the variation in discharge over the season. In 1994 mean daily temperature measured at base camp explains, at best, 37% of the variation in mean daily discharge ( $\text{m}^3/\text{sec}$ ) for the season and 49% of the variation in daily discharge amplitude.

When comparing the 1993 to 1994 data, a similar time interval in 1994 (Days 130-175), shows a 12 hour time-lag between temperature and discharge (Figure 4.5b). *Prima facie* this lag seems significant, however, observation indicates that  $r$  values vary by only  $\pm 0.005$  for  $\pm 1$  hour around the correlograms' global maximum. Consequently, these maximum correlations are quite flat and are unlikely physically significant. Mean daily temperature explains more variance in mean daily discharge in 1993 than in 1994, however, this relation is reversed for hourly values.

#### 4.1.3.1 Discussion

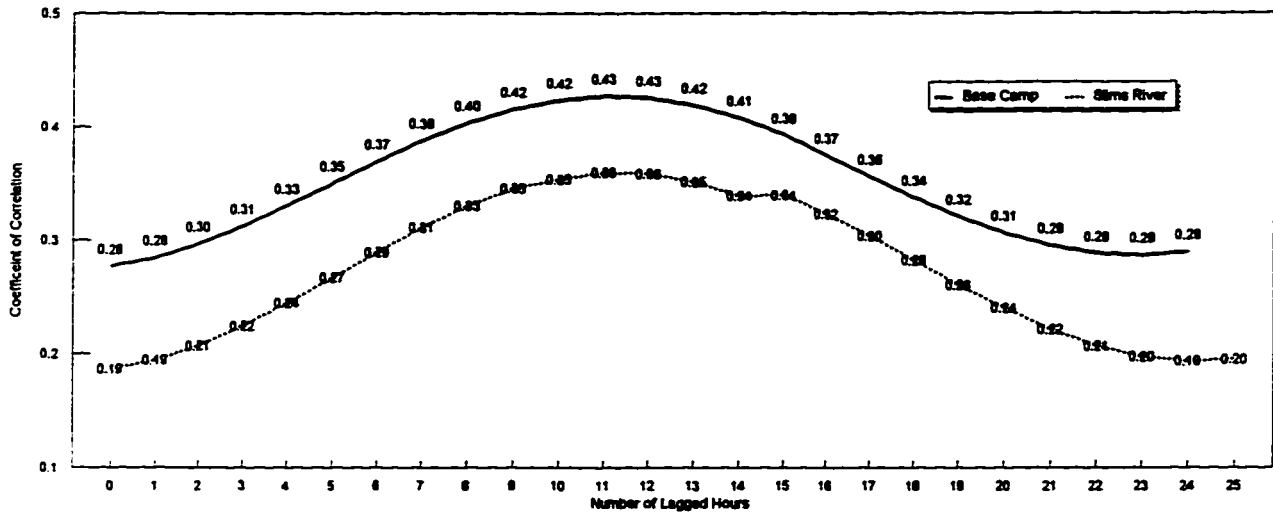
Periods of sustained high temperatures tend to increase meltwater production and discharge amplitude, while cooling periods with considerable cloud cover have the opposite effect (Gurnell et al., 1988; Bajewsky, 1985, p.41; Smith, 1985a; Binda, 1984, p.74). Diurnal discharge amplitude is low during break-up and nival melt and increases in amplitude as summer progresses and glacial melt appropriates a more substantial proportion of total runoff (Smith, 1985a). This is partially due to the enhanced melt rate of ice and firn on the glacier surface when compared to snow because of differences in albedo (Moore, 1993). So, with more glacier ice and firn exposed,

**Figure 4.4**  
 Lagged Correlation between Hourly Discharge and Air Temperature, Slims River, 1993



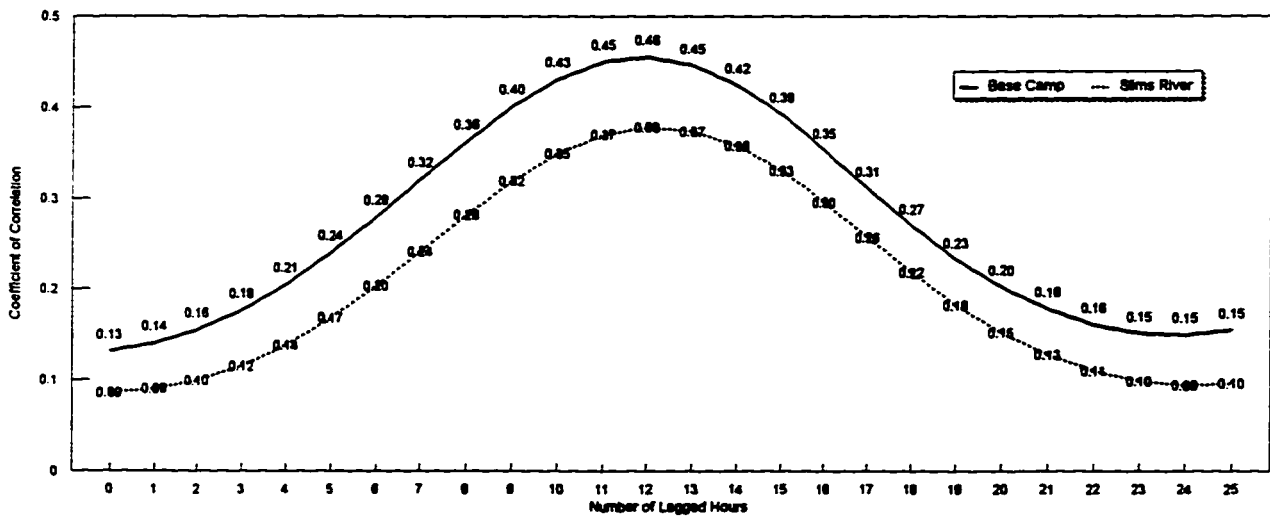
**Figure 4.5a**

Lagged Correlation between Hourly Discharge and Air Temperature, Slims River, 1994



**Figure 4.5b**

Lagged Correlation between of Hourly Discharge and Air Temperature for days 130 to 175, Slims River, 1994



meltwater production will be greater than if a similar area was covered with seasonal or periodic snow. It can be assumed that early season snowmelt on the glacier does not significantly contribute to runoff because of meltwater re-freezing within the snowpack (Binda, 1984, p.59). For these reasons, over the monitoring period the discharge amplitude increases in response to the dominance of the point source discharge from the Kaskawulsh Glacier due to the increasing contributing area of exposed snow-free ice.

The phase shifts between temperature and discharge are due to a 10-12 hour lag-time between maximum daily temperature and discharge for two reasons:

1. The delay-time of a large glacier ablation surface, and
2. Differential travel times of meltwaters from tributary outputs to the gauging site.

First, the time in hours for an aliquot of water  $t_Q$  from any source in the Slims River valley train to reach the gauging site at the bridge is approximately,

$$t_Q = \frac{d}{v} \quad v > \quad (4.4)$$

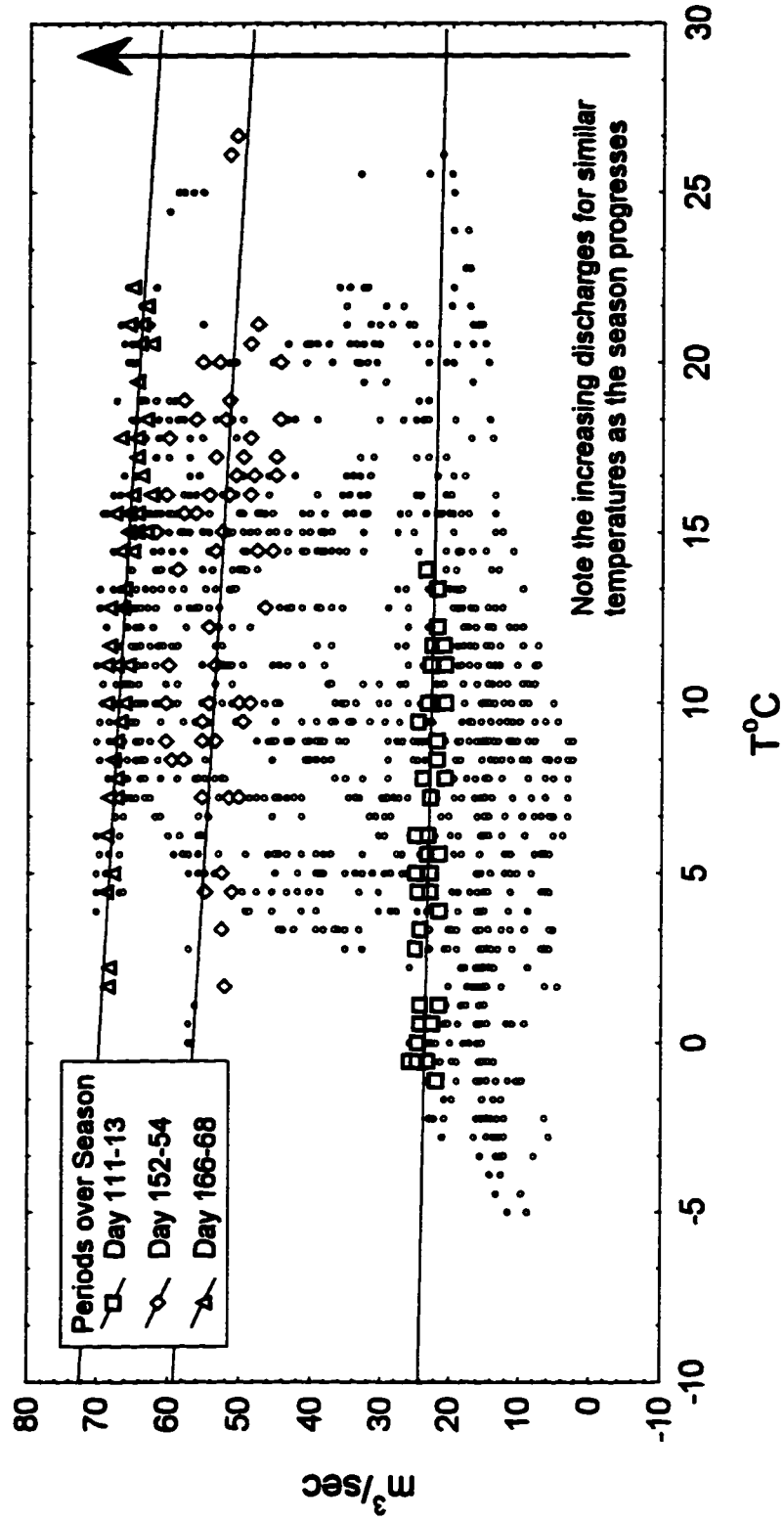
where  $d$  is distance from the gauging site in meters and  $v$  is the stream velocity in m/s. Equation 4.4 assumes that concentration times at tributary outlets to the Slims River are equivalent.

Equation 4.4 demonstrates that it will take a different amount of time for each peak discharge from a tributary to reach the Slims River bridge, thus accounting for the early season hydrograph attenuation. For example, in the early season of 1994, hydrograph peaks are attenuated by as much as 3 hours whereas in the later season, when glacial melt dominates the regime, there are discrete peaks to most events (Figure 4.3ab). As mean discharge increases because of glacier meltwater, the volumetric contribution of tributaries is not large enough to affect stage or attenuate hydrograph peaks. This is also evident in 1993.

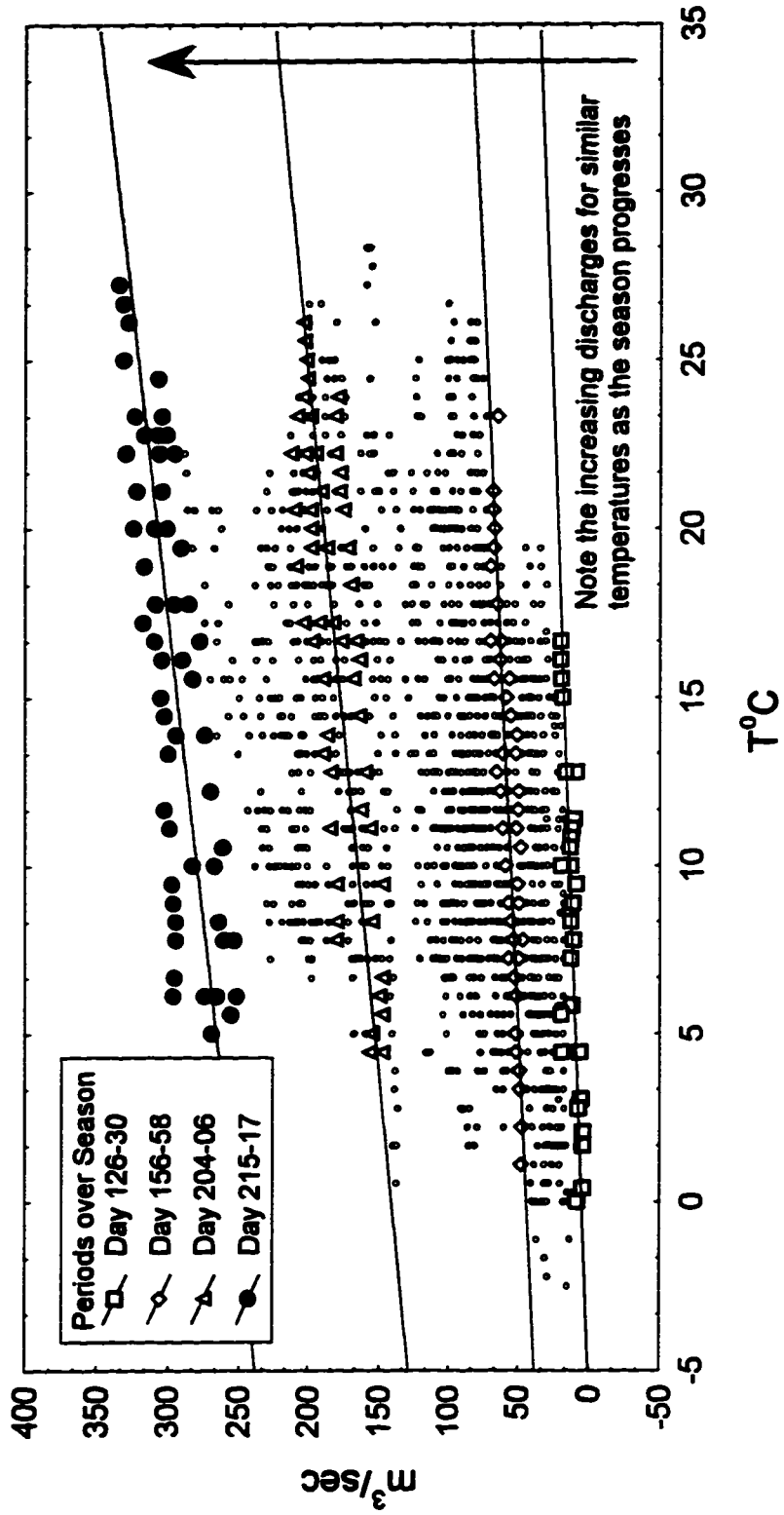
In the 1993 season solving Equation 4.4 for  $v$  with  $t = 10$  hours, requires a mean velocity 0.61 m/s for water to traverse the 22 km distance from the Kaskawulsh Glacier terminus to the monitoring site (Figure 1.1). However, if we assume that in 1993 tributary output is the major supplier of water to the Slims then the mean velocity required will be somewhat slower since less distance would have to be traversed for the same 10 hours. Likewise, in 1994, for  $t = 11$  and  $12$  hours requires mean velocities of 0.55 m/s and 0.51 m/s, respectively. The shift of delay time from 11 to 12 hours in 1994 values is because travel times decrease with increasing stream velocities that accompany discharge from glacier ice melt. However, these explanations are hypothetical because of the flat correlogram peaks, in that, phase shift periods do not significantly change over the season. Even the early/late season difference is more likely due to the inclusion of numerous attenuated peaks in the early season that are not encountered later. Barnett (1974) notes similar observations for the summer of 1970.

Weak explanation of variance of discharge by temperature in both seasons lies in a number of factors. First and foremost, scatter is produced by increasing discharge for similar temperature ranges over the season because of the amplified melt response of clear glacier ice (Figure 4.6ab). Therefore, the weak  $r^2$  between discharge and temperature over the 1994 season is primarily due to the progressive change in contributing area of glacier surface and exposed clear ice which causes an amplified melt response to non-outlying temperatures. Second, during the early season the contributing meltwater production area is less dynamic because meltwater production is primarily derived from snowmelt in the lower basin. Since, temperatures measured at the base camp are indicative of lower altitude temperatures during the snowmelt, at this time the contribution of low lying tributary streams like Vulcan, Canada, Bullion and Sheep Creeks to the

**Figure 4.6a**  
 Scatter-Plot Illustrating the Hourly Air Temperature and Discharge Relation  
 with 12 h Time Lag for Air Temperature and Discharge, Slims River 1993



**Figure 4.6b**  
Scatter-Plot Illustrating the Hourly Air Temperature and Discharge Relation  
with 12 h Time Lag for Air Temperature and Discharge, Slims River 1994

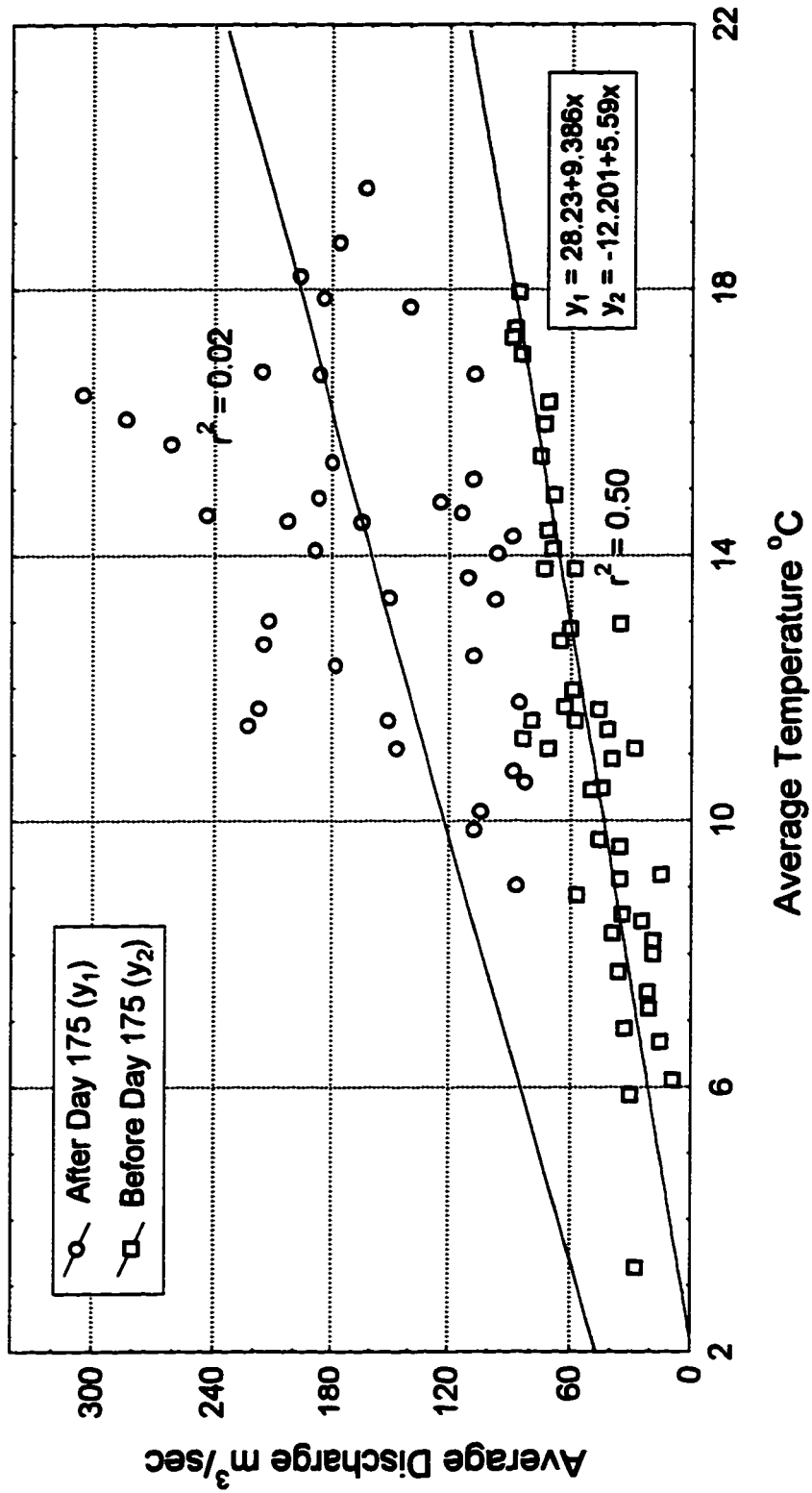


flow of the Slims River have a greater influence on total discharge. These contributions are volumetrically overshadowed by glacier melt in the later season. Additionally, explanation of variance is weak in part because of extrapolation of temperature data from the two low altitude monitoring sites to upper valley and mountain regimes.

For both seasons, mean daily temperature explains more of the variance in mean daily discharge than do similar hourly values. This is reflected in 25 hour moving averages of hourly discharge and temperature for 1993 and 1994. In both years there is a visible lagged response in the smoothed series, such that, mean daily temperatures first reach a maximum and are then followed by daily discharge peaks (Figure 4.2ab). In 1994 this lag is more pronounced in the later season and produces scatter in the mean daily discharge-temperature rating plot thus causing lower correlation than in 1993 (Figure 4.7). However, it is clear that during similar time periods in both seasons that correlation between mean daily temperature and discharge are also similar (compare  $r^2 = 0.53$  in 1993 to  $r^2 = 0.50$  for the same period in 1994) (Figure 4.7). Therefore, it is the later season of 1994 that produces the poor explanation of variance in the mean temperature-discharge relation for that year. This late 1994 season lag, may be due to the storage-outflow response to the general tendency of temperature over a number of days rather than the immediate diurnal temperature variations.

Generally, in 1994, daily discharge amplitude possesses the strongest correlation with mean daily temperature ( $r^2 = 0.49$ ). Amplitude should show a more immediate response to daily temperature variations than the mean discharge on short time scales because of possible time lags in the storage-outflow within the glacial system. Daily discharge amplitude increases exponentially over the season as glacier icemelt appropriates a larger proportion of the daily flow.

**Figure 4.7**  
 Mean Daily Discharge and Air Temperature Before & After day 175, Slims R. 1994



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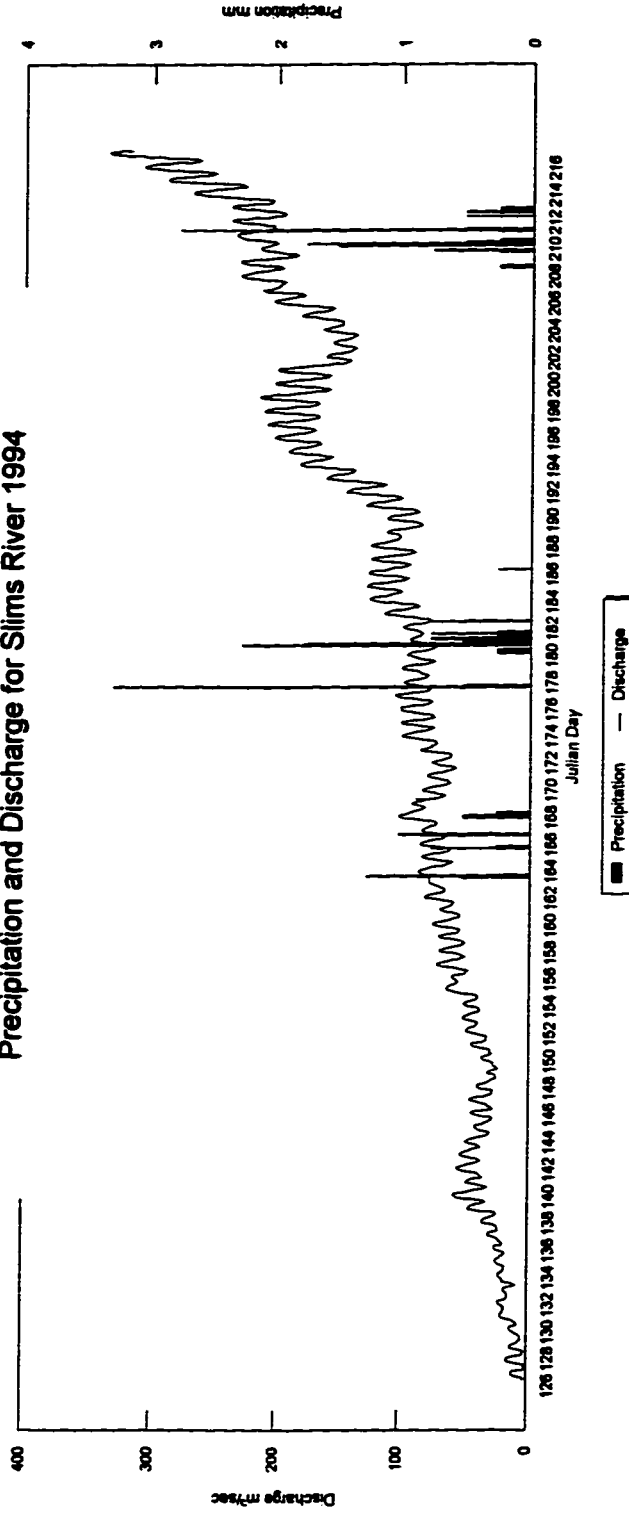
#### 4.1.4 Precipitation and Discharge

Precipitation was infrequent and of low magnitude over both monitoring seasons (Figure 4.8ab; Table 1.2). Smith (1985a) observes that the discharge of glacier fed streams is strongly periodic (diurnal) in the absence of rain-generated runoff. In 1994 when precipitation did take place it was associated with short-term changes in discharge amplitude. The precipitation events on day 163, 181-182, and 211 are associated with an effective elimination of the 'normal' diurnal amplitude such that the flood hydrograph does not completely recede to the previous day's or subsequent day's low flow. These precipitation events associated with discharge changes may be due to better established glacial and proglacial drainage paths combined with a bare glacier surface in the summer which would allow for more effective runoff (Smith, 1985a).

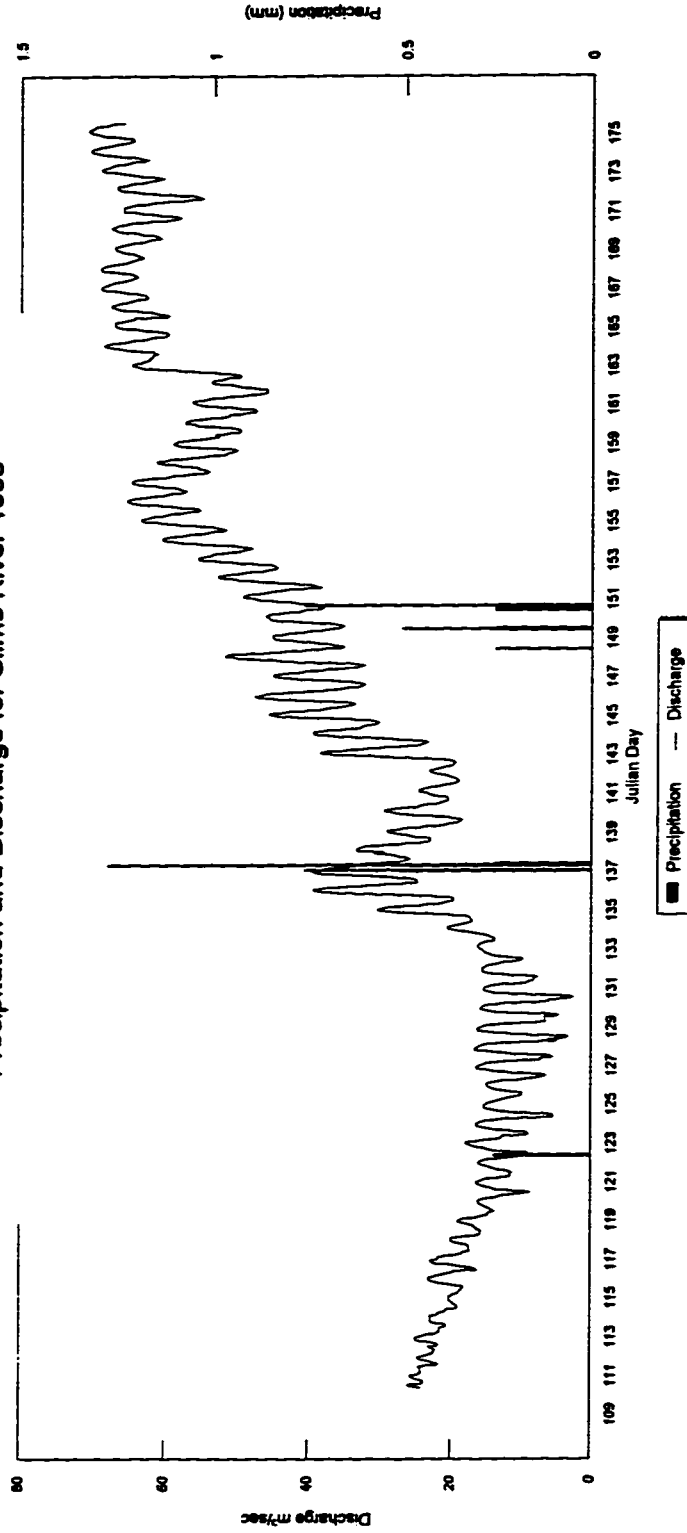
Similar discharge patterns on day 156 and 189 of 1994 were not associated with a measured precipitation event, although both were associated with periods of cloud cover as measured by a decrease in surface short wave insolation at base camp. One interpretation of this situation is that precipitation may have occurred in the upper basin that was not detected at base camp. In 1993 the measured precipitation events are not associated with changes in daily discharge amplitude. Precipitation events on days 137 and 151 are associated with some attenuation of their hydrographs falling limbs (Figure 4.8b). Smith (1985a) suggests that rain-induced discharge changes are less pronounced in the early melt season when snowpack is present to impede runoff.

Generally, precipitation was infrequent in both years and its occurrence decreased the regular diurnal discharge cycle in the summer of 1994 possibly because of the absence of retarding snowpack in the summer. In 1993 precipitation was associated with attenuation of hydrograph falling limbs but not to any great extent and so precipitation was either retarded by the existing snowpack or was of an ineffective magnitude. Due to the low frequency and ineffectiveness of precipitation in both years, temperature is the primary independent influence on discharge.

**Figure 4.8a**  
Precipitation and Discharge for Silms River 1994



**Figure 4.8b**  
Precipitation and Discharge for Slims River 1993



#### **4.1.4.1 Discussion**

Of the measured variables, temperature is the major independent influence on discharge and no significant relations could be found between variables such as solar radiation and windrun with discharge. However, the fact that solar radiation was measured at the base camp, some 22 km from any exposed glacier ice likely underestimates its' influence on meltwater production.

Unmeasured variables that may aid in the prediction of discharge, namely snowline observations, equilibrium line altitudes, graded temperature profiles, surface albedo and upper valley meteorological data could improve the explanation of variance for discharge over the season.

In addition, the physical aspects of storage-outflow within large valley glaciers may not necessarily follow the same regimes as smaller valley glaciers and so this remains an elusive independent influence on discharge. Consequently, although the discharge components can be adequately described, without reliable independent variables as predictors, such descriptions can not presently be transformed into useful predictive models.

## **4.2 Sediment Transport**

### **4.2.1 Channel Cross-sectional Changes**

Channel cross-sections were not taken by bathymetry in 1993, and, hence are not analyzed; however, in 1994 cross-sections were standardised to zero-stage for analysis of bed elevation changes. The south channel, the larger of the two (Figure 1.2), has a more dynamic cross-section and undergoes accretion and scouring while the north channel remains relatively constant for the 1994 monitoring season.<sup>2</sup>

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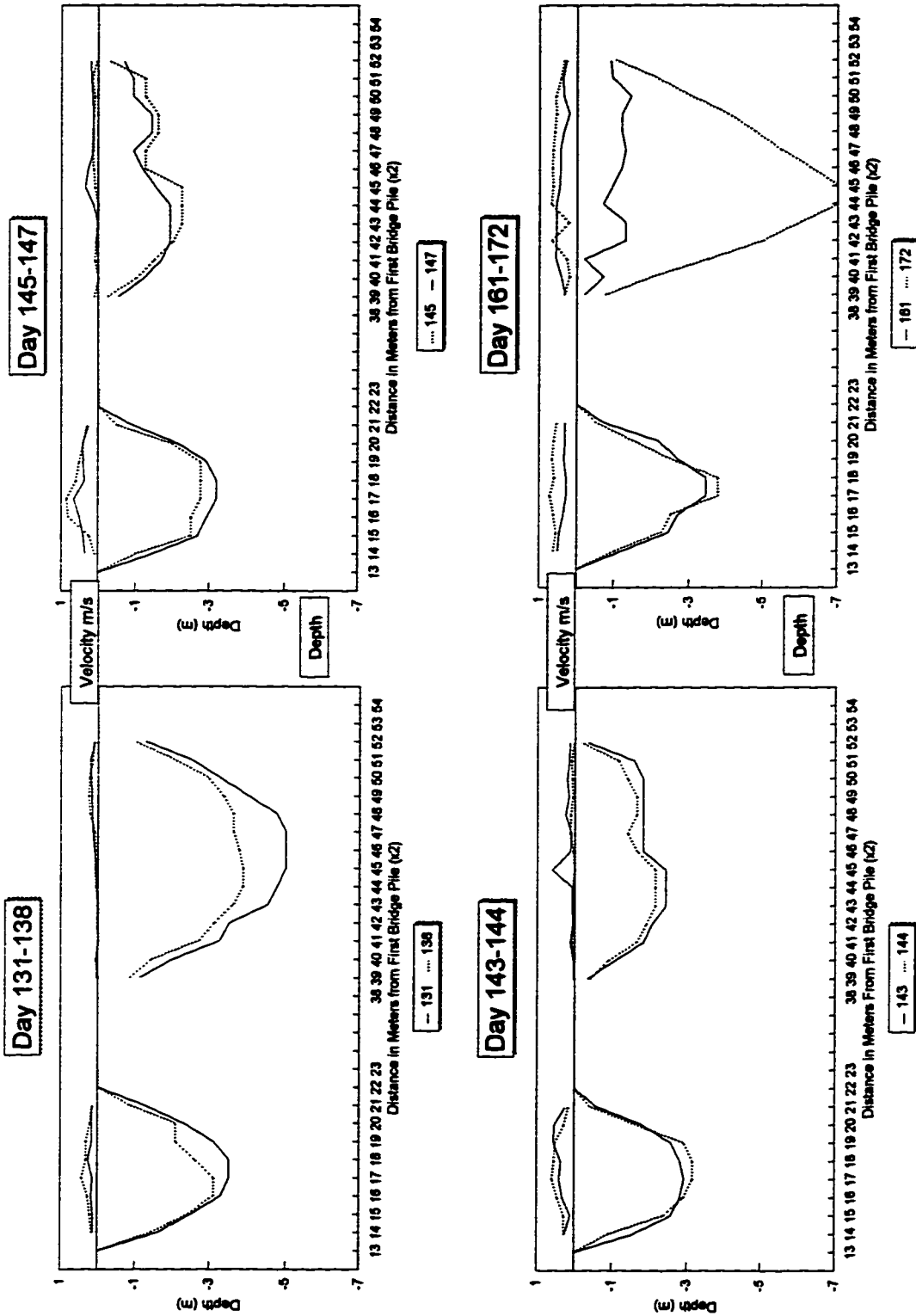
<sup>2</sup> It is not possible to extrapolate these cross-sectional changes to other points in the long profile since the effect of

In 1994 the south channel (Figure 1.2) underwent a period of accretion from Julian day 131 to 161 (Figure 4.9). The section changed from a depth of approximately 5 m to 1 m during this period. Between day 161 and 172 the cross section was scoured and remained comparatively constant for the rest of the monitoring season. During this time the north channel (the instrumented one) underwent relatively little accretion or scouring, or at least little was detected.

The question is: Why is fill taking place during rising discharge? In the south channel fill takes place between days 130 to 161 during increasing spring discharge (Figure 4.9). Although discharge is increasing, the north channel is sufficient to handle all of the water at this time but not all of the sediment. Channel cross-section must be decreased in order to maintain velocities capable of transporting the imposed sediment load. In order to achieve this condition the south channel fills, thus decreasing the cross-sectional area at the bridge until, at some point, discharge is high enough to maintain sediment transport through the measurement reach. Therefore, the fill in the south channel takes place, presumably, because it is not required to pass water and sediment.

Mean discharge continued to increase and between day 161 and 172 the channel was completely scoured and remains so for the rest of the season (Figure 4.9). Considering the imposed sediment load from upstream, a threshold is reached, after which the stream is more than capable of passing all of the sediment supplied to it through the measurement reach without further accretion. In other words, scouring is a response to a threshold discharge whereby the north channel can no longer contain the discharge level. In effect, the stream becomes overcompetent such that mean discharge increases as does the shear velocity and stream power available for sediment transport.

**Figure 4.9**  
Silms River Cross-section and Velocity Variation During Channel Infilling



#### *4.2.1.1 Changes in Hydraulic Geometry...*

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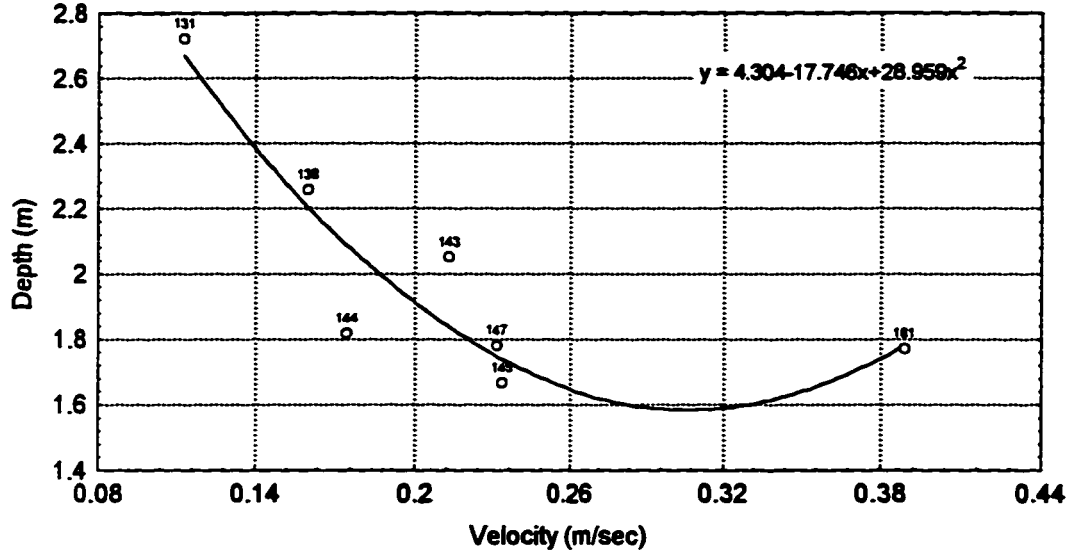
Whether or not accretion and scouring take place on shorter time intervals, for example, accretion and scouring within a day is unknown. Another question is whether the accretion represents bedform development or is simply fill. However, present evidence (time sequence of scour and fill) and that of Barnett (1970) would suggest that channel accretion takes place over a longer time interval, approximately 31 days at present, while the scouring takes place in a rather shorter time interval, presently a period of 11 days. The timing of this sequence suggests that it is controlled by the mean discharge changes rather than the diurnal flood magnitude (amplitude). However, in this regime peak discharge from day to day would be more effective in scouring because the shear stress at the bed is greater with the increase in discharge accompanying the passage of each day's flood wave.

#### **4.2.1.1 Changes in Hydraulic Geometry Accompanying Channel Cross-section Changes**

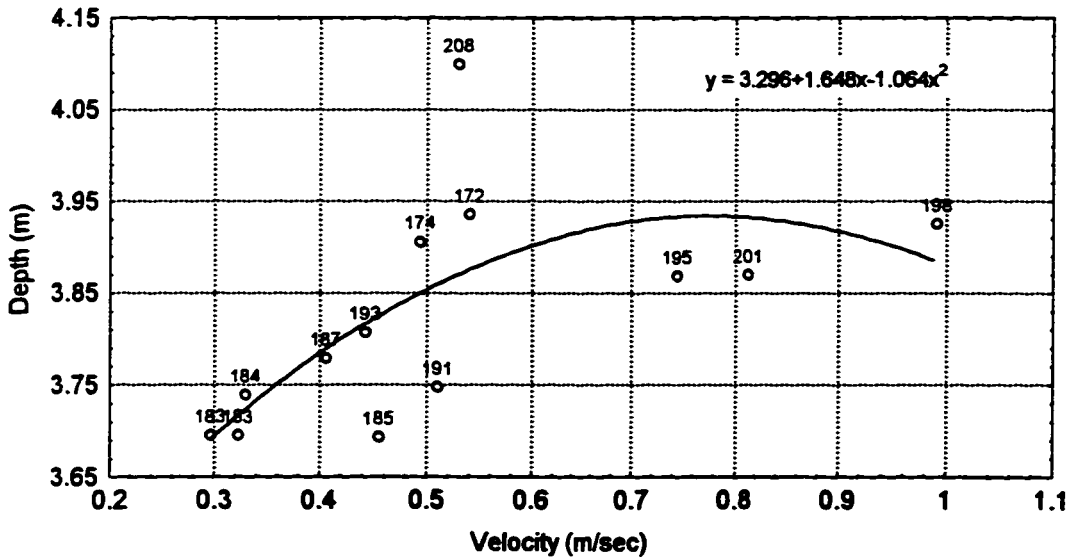
Data on width, velocity and depth are documented for 1994. Width is relatively constant at the monitoring reach because it is limited by the rip-rap banks making up the north and south channels. There are minor changes in width due to increasing water level, but width is not considered a primary independent influence on the cross-sectional geometry at the monitoring site.

Changes in hydraulic geometry are divided into an accretion and post-scour period. While the south channel fills there is an increase in velocity in the north channel until maximum filling on day 161, at which point, velocity also increases in the south channel and with it the shear velocity initiating scour which is completed by day 172 (Figure 4.9). Average velocity increases with decreasing depth during this time (Figure 4.10a). This relation is less clear once the channel is scoured at which time there is considerable variation in velocity (average  $\pm 0.8$  m/s) and minor variation in depth (average  $\pm 0.40$  m) (Figure 4.10b). This illustrates the considerable problems in

**Figure 4.10a**  
Average Channel Velocity and Depth During Channel Fill Sequence



**Figure 4.10b**  
Average Channel Velocity and Depth After Scour and Fill Sequence



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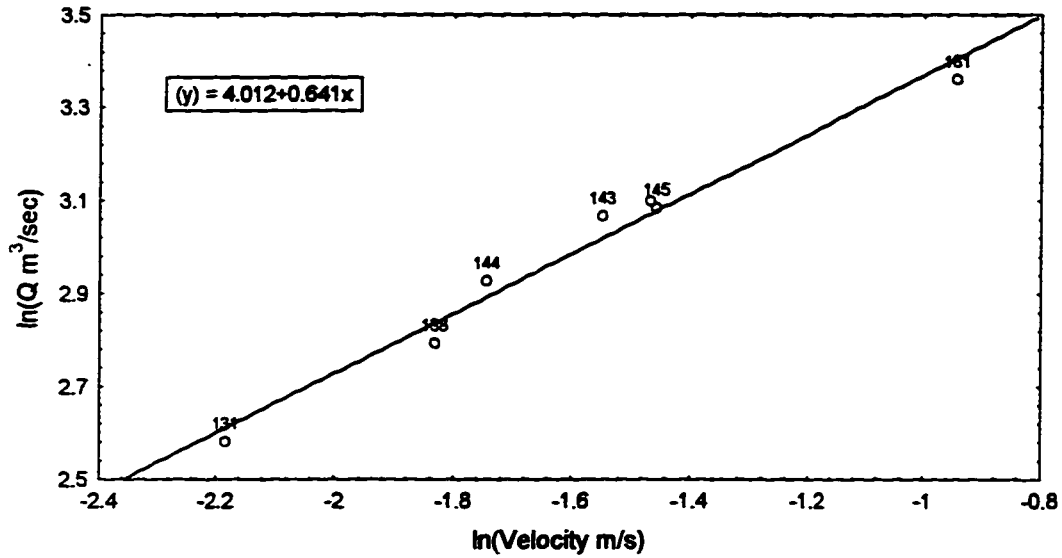
obtaining precise velocity values for similar stages because of turbulence as discharge increases in the later season as was illustrated in Chapter 3.

Scatter plots of both depth and velocity with discharge indicate that velocity is directly proportional to discharge both before and after the scour and fill sequence (Figure 4.11ab) (Figure 3.14ab, Appendix D). This illustrates the strong dependence of discharge on velocity in the measurement reach. Consequently, during the channel fill interval there is also an imperfect inverse relation between discharge and depth and a somewhat proportional relation after the channel is scoured. However, after the channel is scoured there is considerable residual variance between depth and discharge that is explained by the variable velocity measurements, due to turbulence, from which the discharge is calculated. Alternatively, since depth is limited by the channel rip-rap velocity becomes the fundamental independent control on discharge variation during this period and, consequently, has a strong positive relation regardless of measurement error.

The relation between discharge, velocity and depth is more complex in the monitoring reach than can be explained by the standard hydraulic geometry power functions. Richards (1982, p.152) proposes that many channels exhibit relations that depart from these simple curves, and that complex variations are best described by polynomials whose optimal degree is determined by improved explanation of variance. Cross-section changes in the early season demonstrate that accretion is likely a product of changes in the hydraulic geometry accommodating the upstream imposed sediment load in order to maintain sediment transport continuity. In other words, with the low velocities in the early season and high imposed sediment load, the channel 'silts up', decreasing the cross-sectional area and increasing velocity across the reach, thus allowing sediment transport. Generally, sediment transport continuity is maintained from day 138 - 161 by changes in hydraulic geometry, namely decrease in channel cross-section to increase velocity.

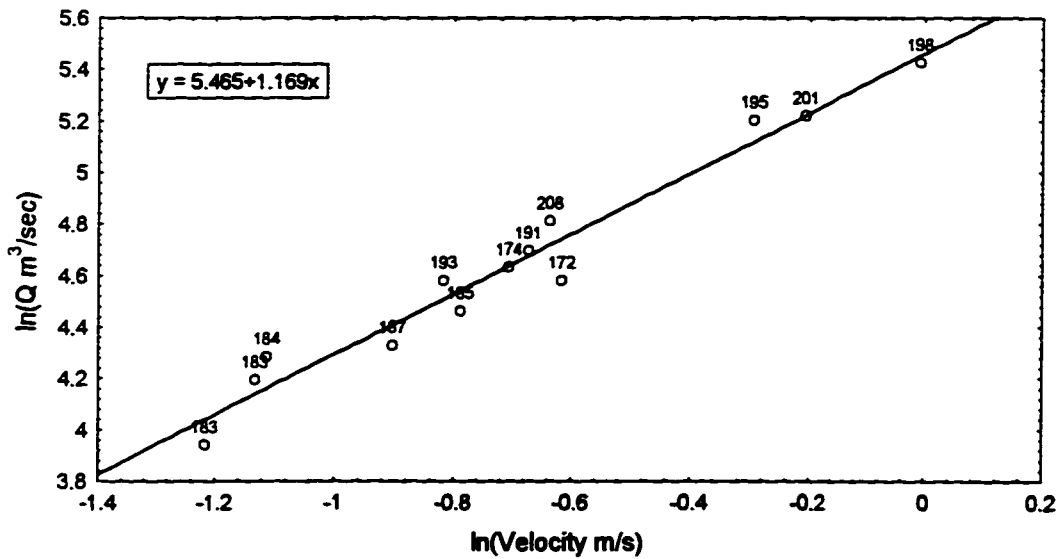
**Figure 4.11a**

Velocity and Discharge during Channel Fill, Slims River, 1994



**Figure 4.11b**

Velocity and Discharge after Scour, Slims River, 1994



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After day 161, continuity is maintained without geometric changes by increasing stream velocity and discharge in a fixed cross-section.

The scour and fill sequence suggests the question as to why the initial bed depth before the fill sequence is deep. Specifically, with decreasing discharges accompanying fall and winter, one would expect fill as the stream loses competence. The initial depth may be a relict of the previous season's scouring. With reasonable discharge, washload has little effect on stream competence and capacity and hence the scour and fill sequences. If this depth is a relict feature it may indicate overcompetence as the season progresses. In order for the stream to remain overcompetent with the low flows accompanying the fall and winter the proportion of fine sediment (<.063 microns) that is not deposited under normal flow conditions must increase as the season progresses. That this is the case is demonstrated by the particle size analysis of suspended samples.

Furthermore, a seasonal fining trend in suspended sediment is consistent with well established channels in the valley train. After peak discharge in summer, channels are left widened and, as stage decreases some channels are abandoned leaving the deeper ones as a well-established watercourse. During the winter, ice first forms in those deepened channels where water is still present. This presents an obstacle to early season meltwaters. Consequently, the channel infilling or accretion during the early melt season may be indicative of lateral channel migration of an undercompetent stream as the valley train responds to increasing flows and high imposed sediment load from the carving of new channels because old ones are blocked by ice.

#### 4.2.2 Grainsize Characteristics

Particle size is important in the determining whether the material carried by a stream will be deposited or carried through a low-energy environment like a reservoir or lake (Østrem, 1973).

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Smaller particles are more likely to pass through lakes and reservoirs. In addition, the particle size distribution reflects the nature of the source materials, processes of weathering, abrasion, corrosion, and sorting that act on particles during transport and deposition (Petts and Foster, 1985, p.179). Particle size influences cation exchange capacity between water and sediments (Lemmens and Roger, 1978) and is an important influence on heavy metal and contaminant loading of river and reservoir sediments (Walling and Moorehead, 1989).

The mean particle size provides a general indicator of the energy of the transporting stream, although this can be erroneously applied in situations when material is supply dependent (Petts and Foster, 1985, p.177; Friedman and Sanders, 1978, p.70). Walling and Moorehead (1989) suggest that a considerable portion of fine grained suspended sediment in rivers is transported as aggregates or flocs rather than discrete particles. In this case, the *in situ* or effective particle size distribution may be coarser than that measured from analysis of chemically dispersed mineral fractions. Particle aggradation is more common in rivers with high salinity and is usually of minimal importance in rivers with low dissolved solids concentrations (Walling and Moorehead, 1989). Sediment mineralogy also affects the potential for aggregation which increases with increasing proportions of clay minerals.

Because the 1994 season was longer, particle size was analyzed for this period only. There are two main controls on the particle size distribution of suspended sediment in the Slims River in 1994. First, the valley gradient decreases with distance from the terminus (Fahnestock, 1969) and with it the stream power expenditure, competence and capacity per unit length of channel. This results in selective deposition of coarser fractions with distance from the terminus (Richards, 1982, p.230). As noted by Fahnestock (1969) and Nickling (1976, p. 26) the Slims River experiences almost perfect proximal to distal sorting, and the small distance from the source to the delta does not encourage weathering as a major determination of particle size.

The second influence on particle size is the direct output from the Kaskawulsh Glacier.

Fahnestock (1969) notes that most of the material ejected, with the exception of fine sand to clay, is deposited near the sub-glacial resurgences. Consequently, much of the washload component could easily be transported in suspension directly from the source with little modification by weathering in the proglacial area. As a result, particle size characteristics measured at the input to Kluane Lake should reflect both processes of hydraulic sorting and changes in the source output.

#### **4.2.2.1 Properties - Sand, Silt and Clay**

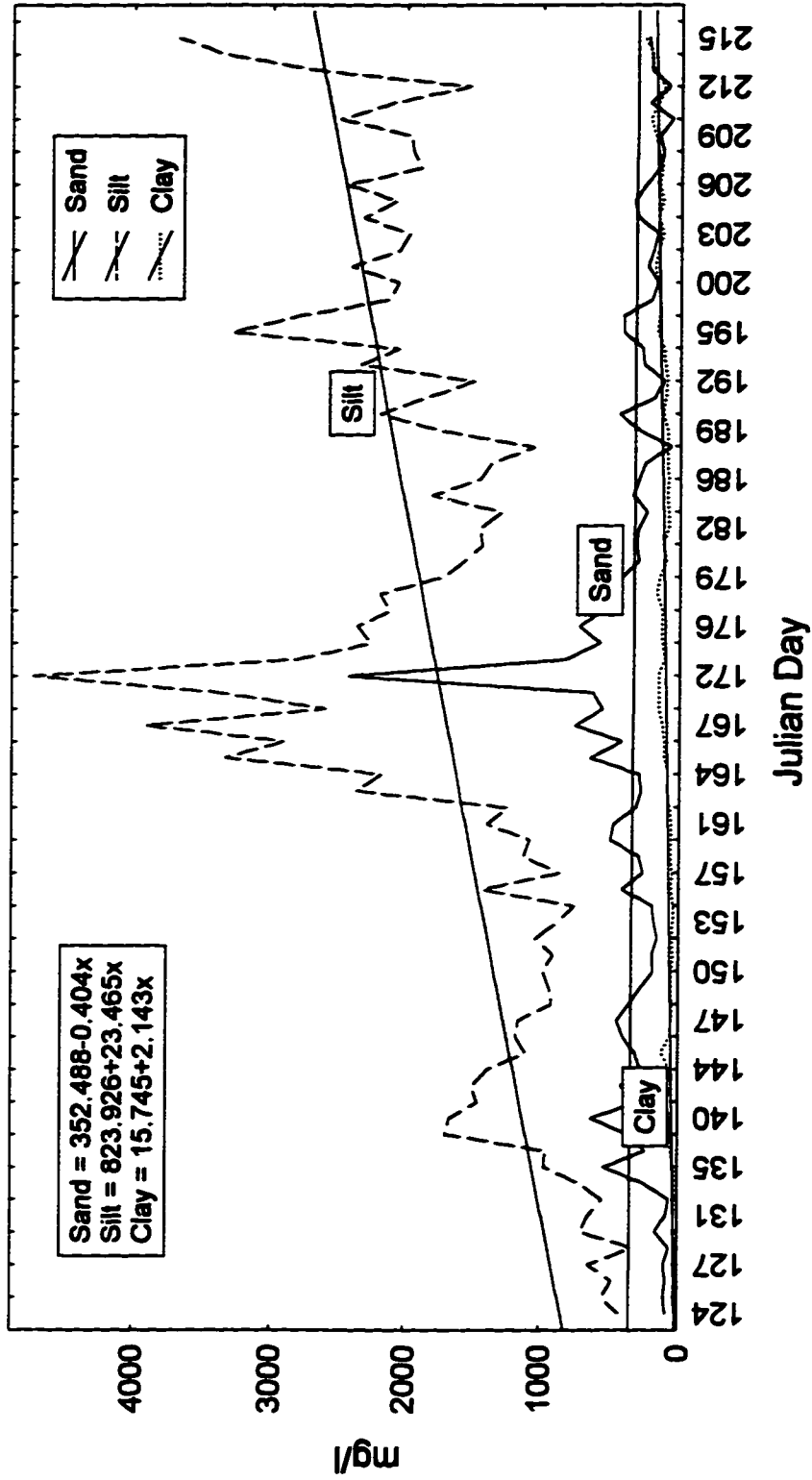
Silt is the largest component by weight of all suspended sediment samples followed by sand and then clay (Table 4.10 Appendix C). The concentrations of silt and clay both increase over the season while that of sand decreases, however, this latter trend is not significantly different from zero (Figure 4.12).

The importance of silt in the particle size distribution indicates that washload dominates the suspended sediment regime and consequently total load. Specifically, the average modal class is 8-4 microns accounting for 20 to 22% of the average suspended sediment concentration (Table 4.11 Appendix C). The decreasing trend in sand concentration and increase of silt over the season may reflect limited supplies of sand and no limited supply of silt. Hydraulically, the percentage of sand should increase as stream power increases if the supply is not limited.

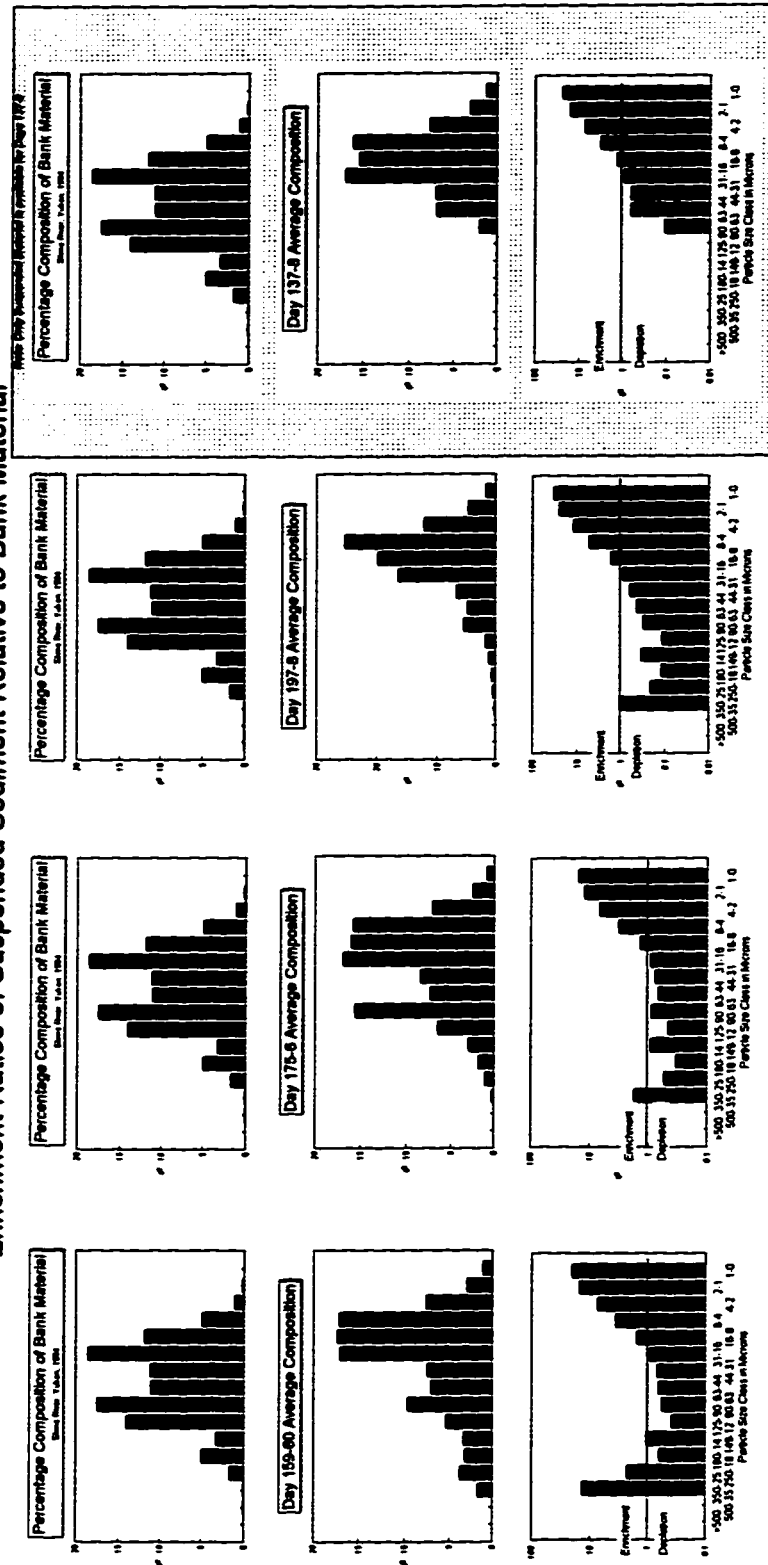
#### **4.2.2.2 Grainsize Distributions**

Both the bank and suspended material samples exhibit some evidence of bimodal distributions truncated at the 63 micron size class (Figure 4.13). This is more apparent in the early season samples of days 159-60 and 175-76 but disappears in the later season at days 197-8. Bimodality can be caused when two distinct sources of sediment contribute to the distribution. Visher (1969)

**Figure 4.12**  
 Concentration Sand, Silt and Clay over Season, Slims River 1994



**Figure 4.13**  
**Enrichment Ratios of Suspended Sediment Relative to Bank Material**



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found this truncation point to range between 88 to 148 microns in fluvial deposits and depending on the mode of deposition to generally range from 44 to 125 microns. Size ranges from 176 to 300 microns constitute the saltation population and those sizes greater than 300 microns constitute the traction or bed load portion of transport (Visher, 1969). By way of illustration, the glacier may provide a considerable proportion of the washload whereas local bank collapse and channel migration with deposition and erosion may provide the material in the coarse distribution. This is supported by the bulk seasonal distribution, in that, the proportion of silt and consequently washload increases over the season at the expense of the coarser sand fraction, a fraction more dominant in bank samples.

Furthermore, strong mixing between suspended material and saltating material is favoured with variable energy conditions and turbulence which results in the partial destruction of the boundary layer (Visher, 1969). Because of the previously mentioned hydraulic differences in the monitoring reach when compared to the upstream conditions, the bimodality observed in the present distributions (Figure 4.13) may be due to coarser bedload going into suspension and mixing with finer fractions in the monitoring reach and *vice versa* for bank material.

#### 4.2.2.3 Enrichment Ratios

Enrichment ratios were calculated for 4 individual diurnal cycles over the melt season (Figure 4.13). Enrichment is defined by Walling and Moorehead (1989) and Walling and Peart (1982) as the ratio of the proportion of a given size fraction in the suspended sediment to that in the source material. In the absence of sub-glacial and tributary sediment samples, bank material is the only alternative source material that can influence the grain size distribution of suspended sediment. The enrichment ratios have all been calculated relative to one bank sample (collected upstream of the monitoring site Figure 1.2) and when considering the spatial distribution of bank material the representativeness of this sample is questionable. The particle size data for the suspended sediment

samples represents the average distribution over an entire diurnal cycle for each of the days presented. Average values were necessary in order to provide a complete grainsize distribution because the amount of coarse material greater than 63 microns in individual suspended sediment samples was often less than 100 mg which, in itself, is not sufficient for sieving and too coarse for laser analysis.

Coarse fraction enrichment, 125 to 350 microns, is greatest in the early season, sample day 159-60, and disappears by the late season sample at days 197-8 (Figure 4.13). All three samples have marked enrichment in the fine silt and clay range (16 to 0.5 microns). The enrichment in fines begins at the 8-16 micron class consistently over the four samples. Thus, the magnitude of enrichment increases with decreasing particle size and is at a maximum for clay. Also, the modal class of 8-4 microns falls well within the common enrichment ranges of 16 to 0.5 microns. This reflects an almost complete absence of the very fine silt size class from the bank material.

The suspended enrichment of fines is consistent with both hydraulic sorting of bank and bed materials or, alternatively, glacier sediment discharge. The latter explanation is questionable for the early season on grounds that the modal class in the early season is the same as in the late season and, given that the early season is nival-dominated, largely precludes the effect of glacier sediment discharge on these classes. However, in the early season the suspended material samples more closely resemble the distribution for bank material, and this suggests that bank material may be the primary influence at this time. The question is whether or not the total load associated with the classes of < 16 microns could be derived from bank material alone in the early season. This question is pertinent considering that tributaries flow through colluvial and alluvial material and carry little suspended sediment to the Slims River.

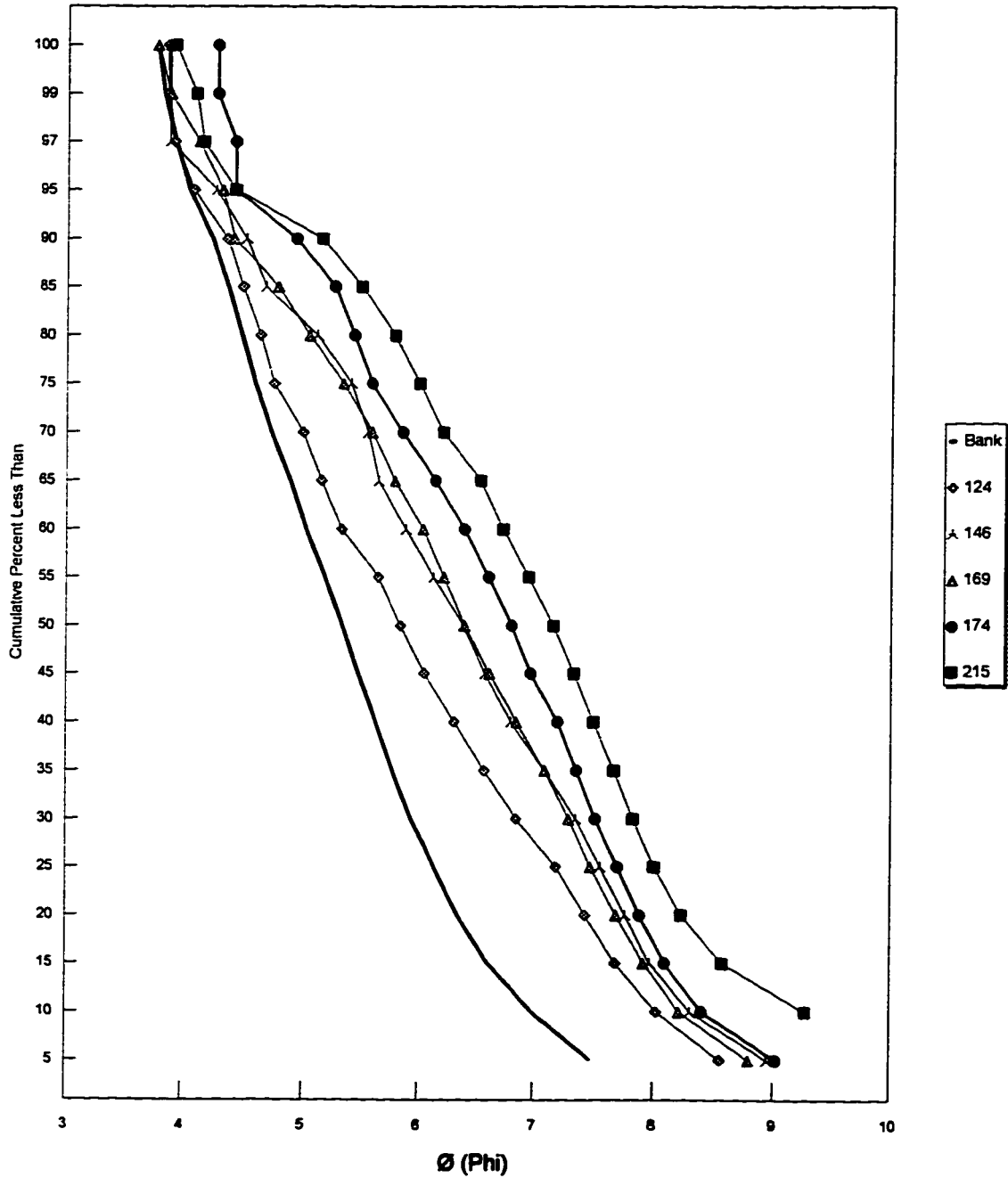
On average, the classes of 16 - 0.5 microns account for 51% of the total load over the season. The total load transported over the monitoring period is  $1.8 \times 10^6$  metric tons of suspended sediment. From Julian day 124 to 143, when the nival flow regime dominates,  $5.1 \times 10^4$  tons are transported by the Slims River, and 51% of this is  $2.6 \times 10^4$  tons. Hypothetically, if the grainsize composition of the Slims River valley for the 13 km where river silts are present was the same as that of bank material, to a depth of 1 m, there would be  $2.3 \times 10^7$  tons of material. Additionally, 18 % of which would yield  $4.2 \times 10^6$  tons of sediment less than 16 microns. This amount of material is two orders of magnitude greater than that transported by the Slims River up to Julian day 143. In the light of such figures, it is possible for the internal channel sources to supply the majority of fines for suspended sediment concentration in the early season.

#### 4.2.2.4 Washload

The washload component of suspended sediment concentration is supply-dependent rather than hydraulically-controlled, and a stream will transport all available under most flow conditions. Coarse sediment may be affected by erosion and re-deposition while washload may be transported to the monitoring site without substantial modification. Changes in washload may be indicative of changes in the dominance of the glacier hydrological system. Therefore, because of the importance of the silt-sized (< 63 microns) dispersed mineral fraction to the suspended sediment concentration, it is pertinent to look at this mineral fraction and its variations over the season.

Selected samples over the 1994 the season were analyzed to determine changes in the median diameter, mean diameter, inclusive graphic standard deviation, inclusive graphic skewness, and graphic kurtosis in accordance with the procedures laid out in the methodology (Table 4.12 Appendix C). Typical cumulative frequency plots covering the grainsize range encountered indicate a progression to a finer distribution over the monitoring period (Figure 4.14).

**Figure 4.14**  
 Typical Cumulative Frequency Curves for Suspended Silt and Clay



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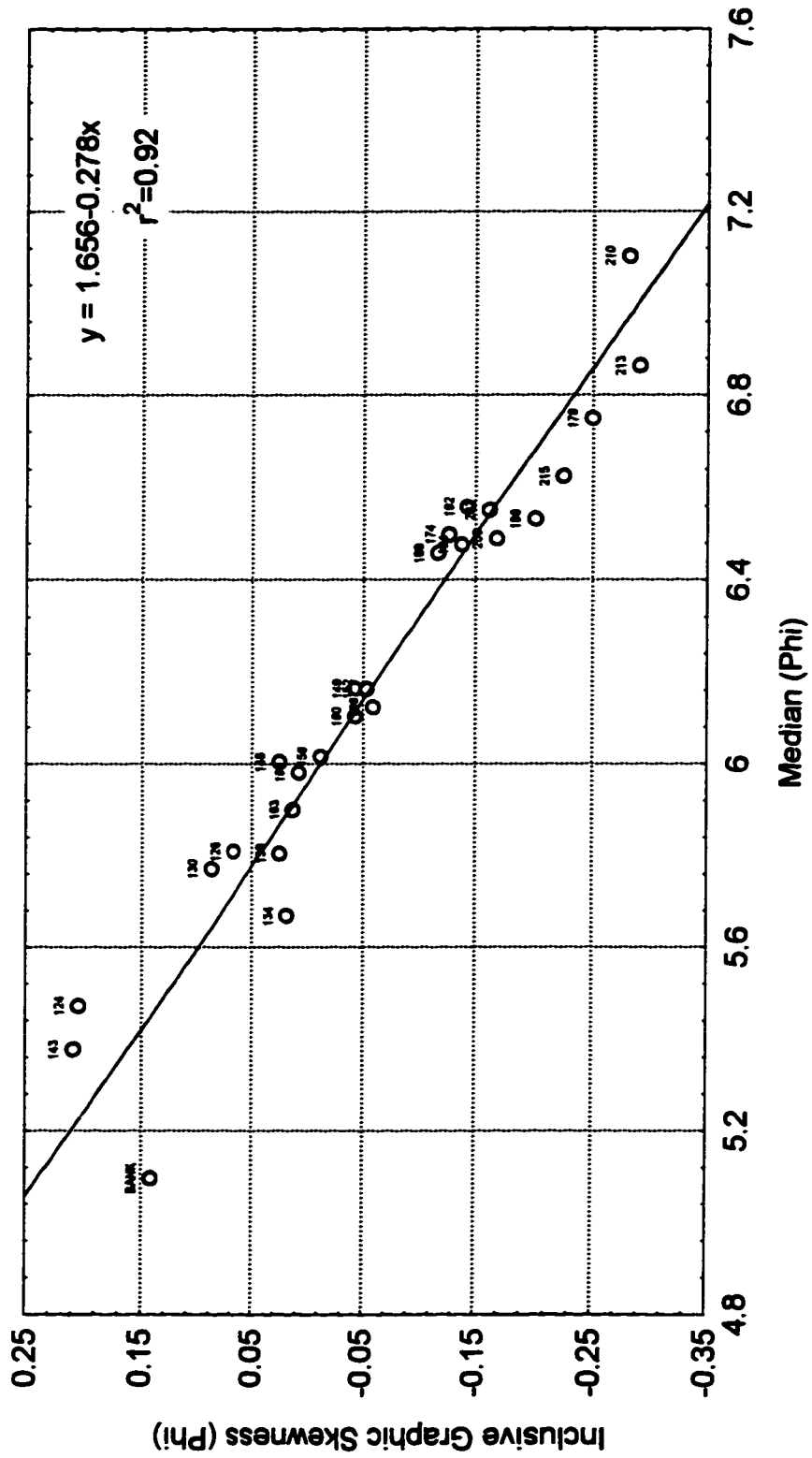
This fining trend is confirmed in the decreasing mean and median particle sizes (Figure 4.31, Appendix B). Conversely, skewness exhibits a trend towards the negative or coarse fraction as the melt season progresses (Figure 4.31, Appendix B). There is an inverse linear relation between median size and skewness which is not wholly temporally contiguous (Figure 4.15) (Figure 3.17ab, Appendix D). Generally, samples become finer over the season but are coarsely skewed in this trend. The suspended sediment samples are classified as nearly symmetrical in the early season to negative-skewed in the mid to late season. With limiting conditions of supply and increasing discharge, hydraulic sorting will favour coarser particulate for entrainment.

Sorting decreases over the season as the standard deviation of the samples increases (Figure 4.32 Appendix B). This trend is discernible in the cumulative frequency plots (Figure 4.14). Generally the steeper the curve the better sorted it is because it covers less size classes. There is a general trend of decreasing sorting (increasing standard deviation) with decreasing particle size that is non-linear (Figure 4.34 Appendix B). Folk and Ward (1957) suggest that the relationship between sorting and particle size is sinusoidal where local minima correspond to the modal classes in multimodal distributions.

The bank sample is classified as moderately sorted while all the suspended samples are poorly sorted. This is not surprising since the bank samples are a product of a depositional environment and should be deficient in fine fractions as these are easily transported in the Slims delta even by overbank flow conditions. Such an observation is consistent with the enrichment of fines in suspended samples.

Furthermore, the distribution of kurtosis values shows that most of the samples are platykurtic and better sorted in the tails of the distribution (Figure 4.35 Appendix B). This is likely due to the absence of extremely coarse or fine outliers in the distributions which may reflect fairly consistent flow rates but may also echo available sediment sizes.

**Figure 4.15**  
**Median and Skewness in Particle Size Distribution, Slims River 1994**



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The coarser distributions in the spring may reflect the dominance of channel sources of sediment while the fining as the season progresses may reflect dominance of glacier sources. Using a two tailed Student t-test for independent samples it is hypothesised that the sample means of the bank material and suspended sediment were the same and that any observed variation was due to chance. Rejection was set at  $\alpha = 0.05$  level. Analysis shows that the hypothesis can be rejected for all but days 124 through 143 and, as well, day 163. The same hypothesis was tested for the suspended sediment samples themselves and could not be rejected with the exception that day 124 and 143 were significantly different from day 210.

From these hypotheses and tests is inferred that the bank sample is significantly different from the suspended sediment samples after day 143 which is also evident in the enrichment ratios.

Furthermore, the suspended sediment samples themselves while showing fining trends over the monitoring period are not considered statistically different from each other. Whether this has physical meaning in support of early season samples reflecting perimeter sediments and later season samples reflecting direct glacial sources is questionable and suggests that another line of argument may be more effective.

Given that bank material is deficient in silt, and if it is assumed that the grainsize distribution of perimeter sediments influences the channel suspension to any significant degree as the season progresses then distributions should coarsen. Coarsening over the season is also encouraged since,

1. If vertical sediment concentration gradients exist, then with rising stage the zone of traction load transport increases in height above the bed and becomes closer to the sampling nozzle which in-turn increases the probability for coarser particles to be sampled.
2. In an internal supply situation sorting would increase as the season progresses because certain particle sizes would be exhausted. However, this is not the case in the Slims River.

Yet, it is observed that:

1. The percent by weight of sand in suspended sediment samples is lower in the later season.
2. The percent by weight of silt in suspension increases while at the same time sediment concentrations are similar to those in the early season.

Thus, with increasing discharge there is no hydraulic reason to expect finer distributions as the season progresses unless the distribution is largely supply dependent. Consequently, from the above observations, the suspended sediment concentration must be controlled by an external supply that is capable of maintaining an abundance of fine material. The only external mechanism that could increase and maintain the supply is the Kaskawulsh Glacier. Washload is a non-capacity load and so transport can occur without significant modification in the proglacial valley train. This is also consistent with well connected drainage streams in the valley train.

#### 4.2.2.5 Discussion

The largest proportion of sediment, fine silt, will be transported by most flow conditions for great distances without deposition, and so may be glacially-derived. Internally to the transport system, bank material exerts an influence on the particle size distribution of the suspension as it contains coarser material that can be transported based on local hydraulic conditions. It is not possible to determine how much deposition and re-mobilization takes place within the channel between the Kaskawulsh Glacier terminus and the Slims River bridge, but Bogen (1980) suggests that this can be significant for coarser particulate.

In the measurement reach as the discharge season progresses, it is evident that stream power is sufficient to suspend particle sizes larger than those found in suspension. This conclusion arises for two reasons: first, there is no channel accretion in the later season; second, discharge, stream power and turbulence increase as the season progresses but the concentration of coarse suspended sediment transported decreases.

### *4.2.3 Discharge and Suspended Sediment Seasonal Characteristics*

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Based on the amount of washload material transported and its increase in proportion over the season it is plausible to conclude that this material is supplied by glacial discharge since glacial erosion is capable of both maintaining and increasing supply as the season progresses. If supply was internal to the valley train, exhaustion of fines would take place first and sorting would increase over the season, whereas it does not.

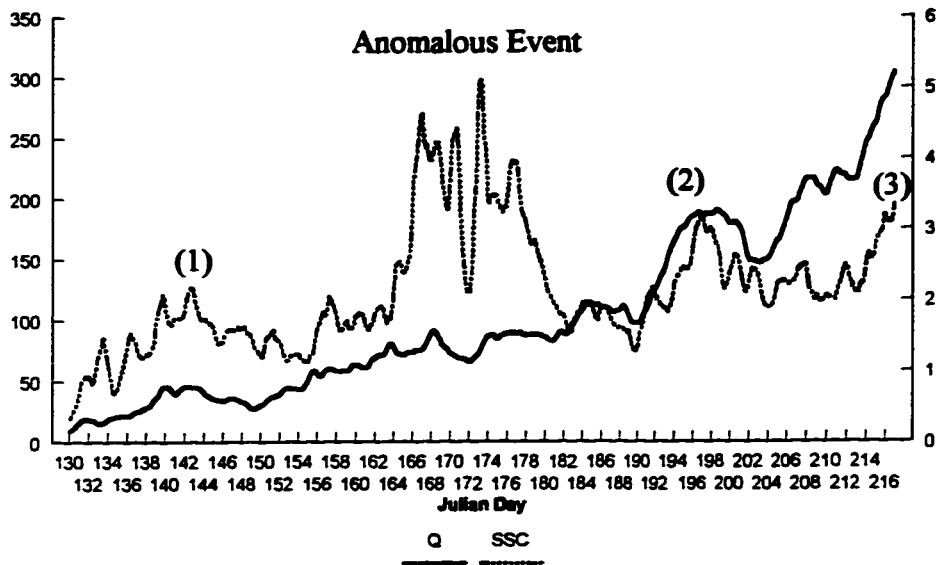
### **4.2.3 Discharge and Suspended Sediment Seasonal Characteristics**

Suspended sediment concentration rather than suspended load is discussed in the context of sediment dynamics because the latter is a function of discharge and therefore correlations are spurious (Gregory and Walling, 1973, p.215). For both seasons of 1993 and 1994, 25 hour moving averages of suspended sediment concentration follow the general tendencies of discharge changes over the melt seasons season (Figure 4.16ab). This trend is more evident in 1993 (Figure 4.16b).

For all of 1994 suspended sediment concentration events clearly follow changes in mean discharge, whereas the major seasonal peak does not and may indicate sediment delivery from subglacial sources (Figure 4.16a). There are three minor sediment concentration peaks labelled 1, 2 and 3 that clearly follow mean discharge variation in 1994 (Figure 4.16a). An anomalous sediment concentration event, in that, it is a seasonal outlier, between days 163-82 clearly deviates from this seasonal trend (Figure 4.17b). During this event there is no drastic increase in discharge accompanying these maximum sediment concentrations which are observed to range between 2.5 g/l to 11.9 g/l (Figure 4.17b; Figure 4.16a) (Figure 3.25, Appendix D). Discharge explains 11% of the variation in suspended sediment concentration over the 1994 season. When the anomaly is not included, the explanation improves to 45% (Figure 4.17b; Figure 3.24, Appendix D).

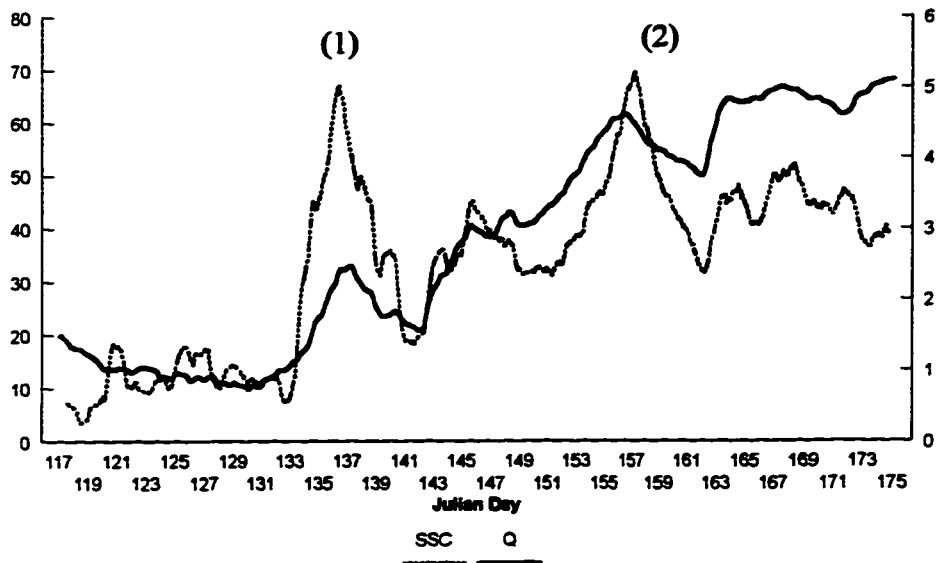
**Figure 4.16a**

Twenty-five Hour Moving Averages of Suspended Sediment Concentration and Discharge, Slims River, 1994

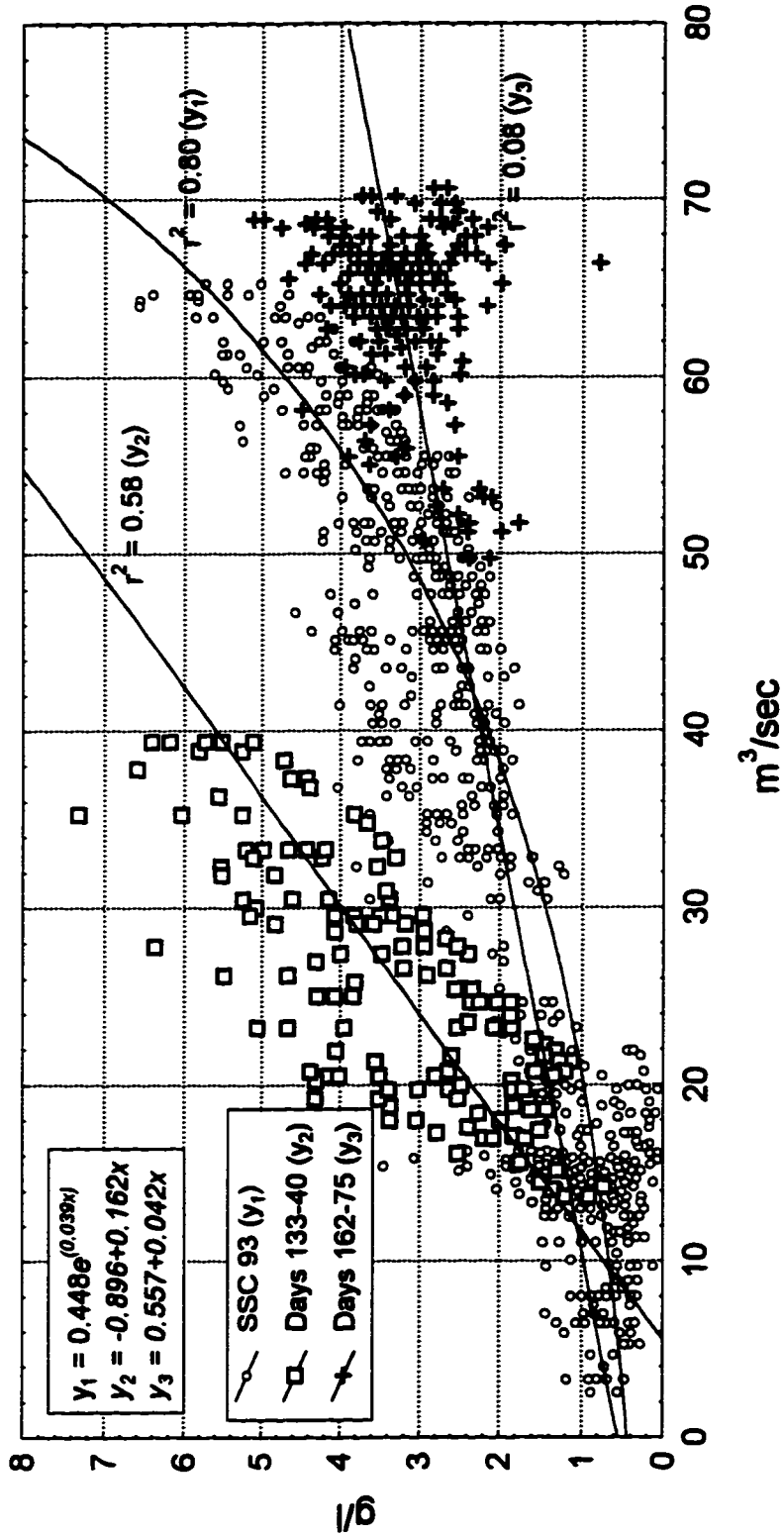


**Figure 4.16b**

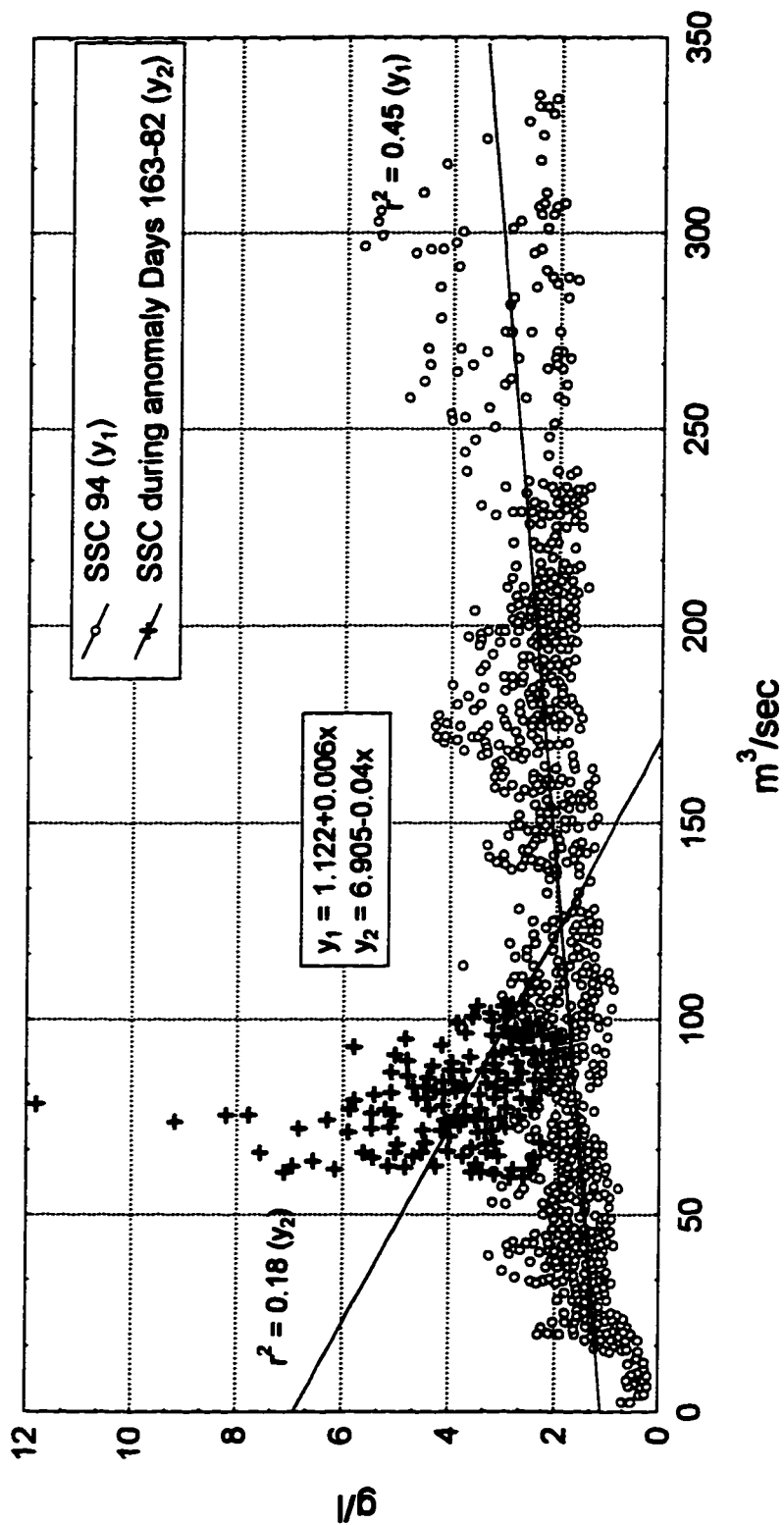
Twenty-five Hour Moving Averages of Suspended Sediment Concentration and Discharge, Slims River, 1993



**Figure 4.17a**  
 Categorized Scatter-Plot Illustrating Suspended Sediment Concentration (SSC) and Discharge over the 1993 Season, Simis River



**Figure 4.17b**  
 Categorized Scatter-Plot Illustrating Suspended Sediment Concentration (SSC) and Discharge over the 1994 Season, Sims River



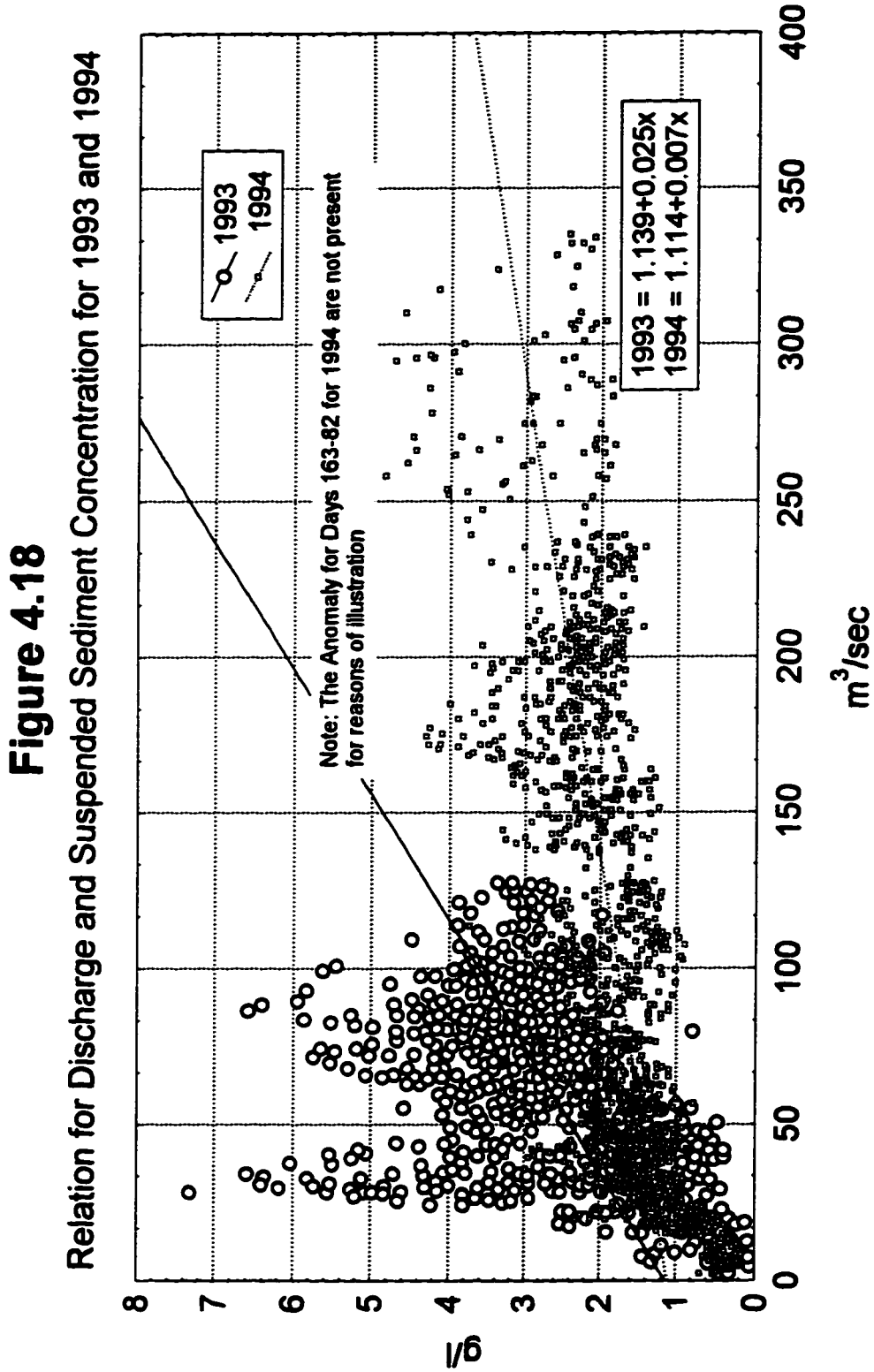
### 4.2.3 Discharge and Suspended Sediment Seasonal Characteristics

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Johnson (1991b) suggests that the opening of the glacier hydrological system coincides with greatest variability and highest concentrations of suspended sediment. The timing and magnitude of the present concentration peak is similar to that observed in the Slims River in 1983 by Johnson (1991b).

In 1993 suspended sediment concentration more clearly follows mean discharge variation. There are two major peaks in concentration between days 125-133 and 144-154 (Figure 4.16b). These peaks and troughs follow those of the mean discharge. In 1993, 50% of the variation in daily suspended sediment concentration can be explained by daily instantaneous discharge as the only independent variable. This weak  $r^2$  value is primarily because of two outlying events. The first of these takes place between days 133-140 where the highest sediment concentrations of the season are experienced - up to 7.4 g/l (Figure 4.17a; Figure 3.22ab, Appendix D). This outlying event has a strong positive relation with discharge as well as an  $r^2$  of 0.58. The second event is between days 162-175, where sediment concentrations are dropping off with increasing discharge and explanation of variance is low (Figure 4.17a) (Figure 3.23ab, Appendix D). Removal of these two events leaves discharge explaining 80% of the variation in suspended sediment concentration over the 1993 season, although, this is non-linear (Figure 4.17a; Figure 3.21ab, Appendix D).

Suspended sediment concentrations produced in 1993 for a given discharge are considerably greater than those produced in 1994. A scatter plot illustrating both years shows that for discharges between approximately 25 and 100 m<sup>3</sup>/sec, sediment concentrations in 1993 are 2.5 to 3 times greater than the similar discharges in 1994 (Figure 4.18). However, the anomaly in 1994 overlaps the 1993 concentrations and has greater concentration values. These facts show that the stream in 1994 is capable of carrying more sediment, and, consequently, 1994 discharges are operating under capacity. Furthermore, both the 1993 concentration peaks are comparable to the 1994 anomaly, but because of their short duration and clear variation with mean discharge, it



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would be premature to attribute them to the opening of the glacier hydrological system.

Consequently, the suspended sediment regime in 1993 is likely controlled by snowmelt and internal sediment supply rather than glacier discharge. The high concentrations in 1993 may be due to different supply mechanisms, more channel migration, or less braiding but these explanations remain speculative in the absence of data.

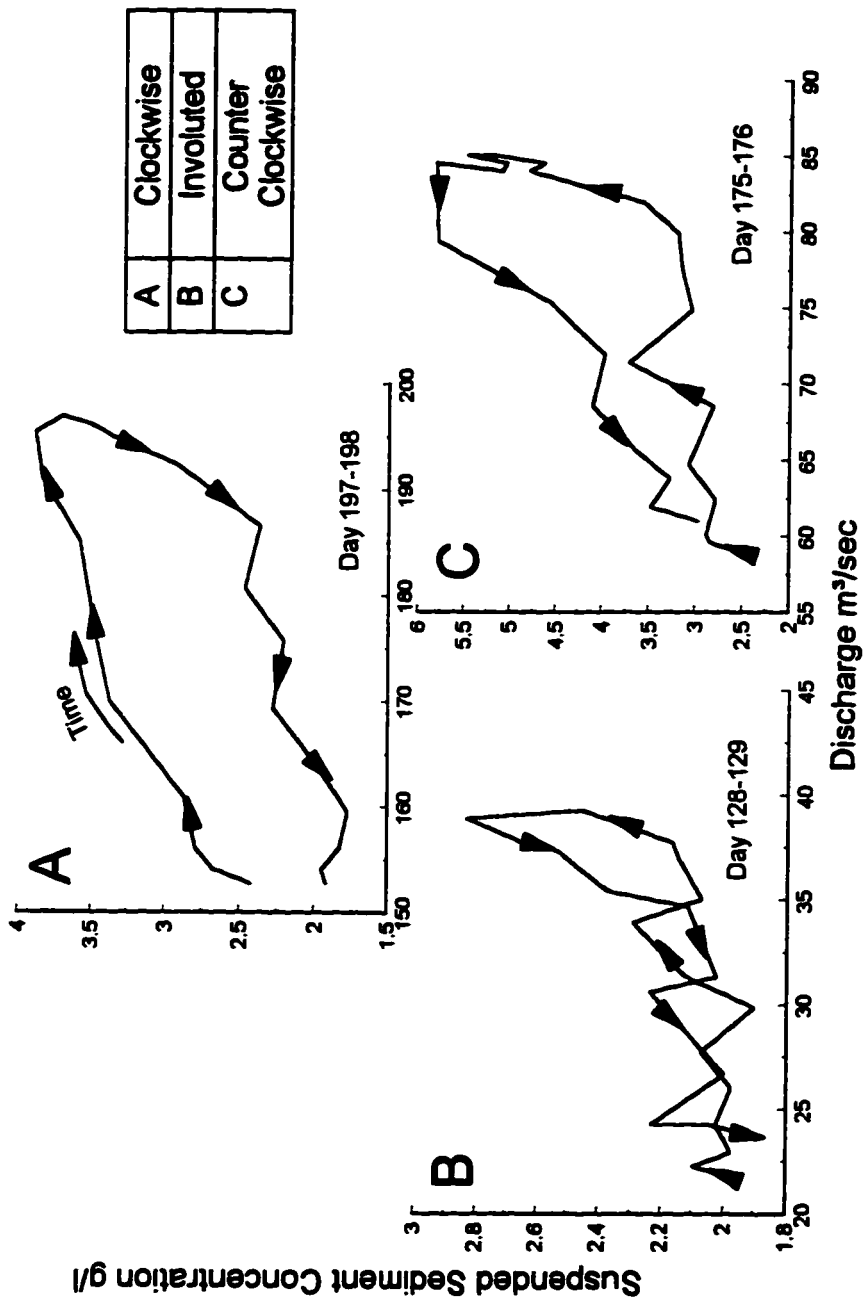
#### **4.2.4 Hysteresis in Sediment Discharge**

In both seasons hysteresis is present to some degree in every diurnal cycle, varying from well defined clockwise hysteresis to involuted relations with multiple clockwise and counter-clockwise segments to well defined counter clockwise hysteresis (Figure 4.19). Of the 88 days in 1994, from day 129 - 217, clockwise hysteresis is most frequent (Table 4.1). Hysteresis loops are best defined between days 178-217 with clockwise hysteresis occurring 91% of the time (Figure 4.20a; Table 4.1). The occurrence of counter-clockwise hysteresis coincides with the time when the glacier hydrological system is believed to open at days 163 and 164 as well as days 174 and 175.

In 1993, a picture similar to the early melt season of 1994 emerges. For the 58 days of data from day 117 to 175, clockwise hysteresis is frequent but the regime is dominated by the occurrence of involuted relations (Table 4.1; Figure 4.20b). This may be expected since monitoring began in 1993 12 days earlier than 1994 during the period when discharge is only beginning to establish channels in the valley train and flows are undercompetent for the imposed load.

Clockwise hysteresis is most frequent in 1994 with high flows accompanying glacial meltwaters. Clockwise hysteresis is less frequent in the early season of 1994 and the whole of 1993. Aside from the observation that involuted relations are more common for the whole of 1993, hysteresis does define the short-term sediment transport and discharge regime of the Slims River in both years, and so its causes warrant investigation, and, if preferably, explanation.

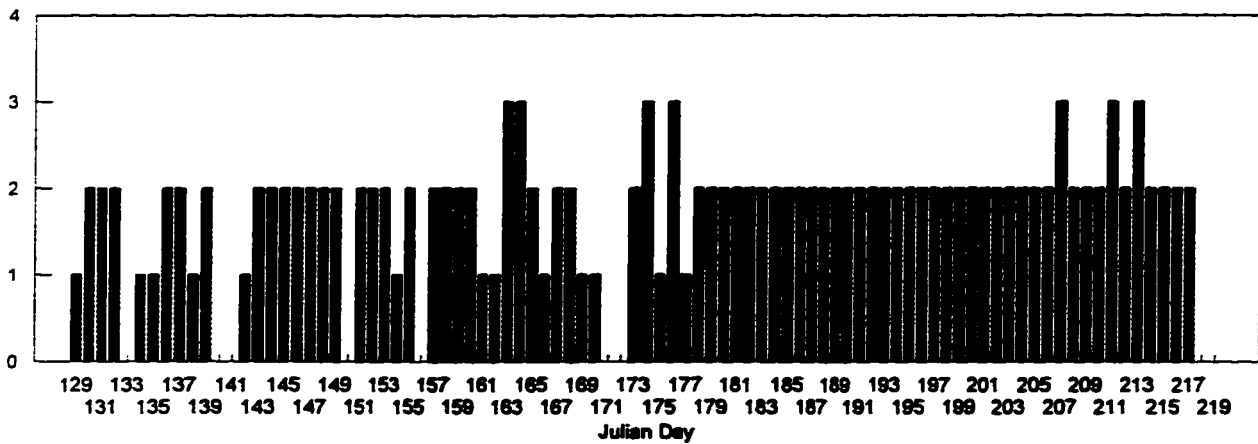
**Figure 4.19**  
Observed Types of Hysteresis



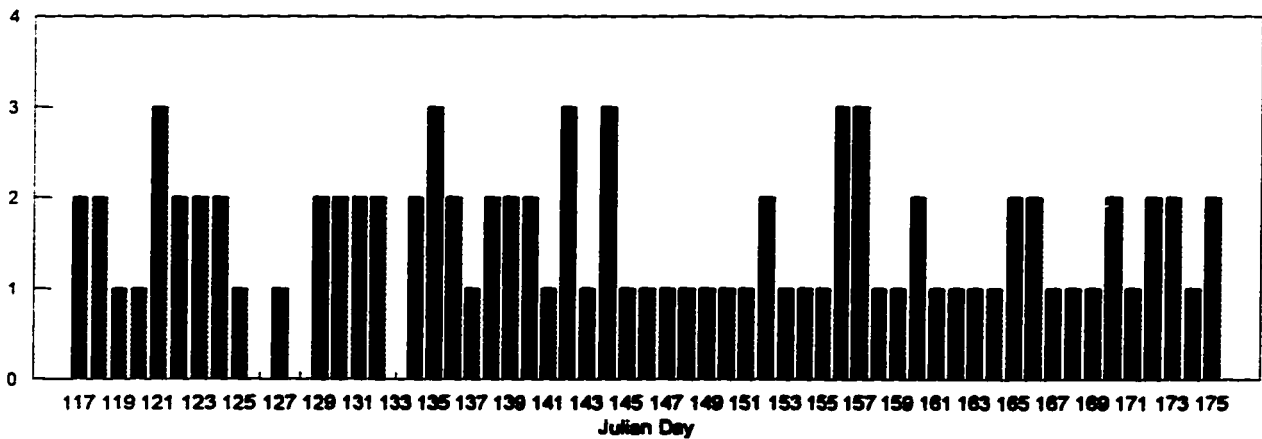
**Table 4.1**  
*Percentage Occurrence of Hysteresis over 1993 and 1994 Seasons*  
*Classified by Type of Hysteresis*

Type of Hysteresis	Percent of Days Hysteresis Occurs			
	1994	1993	Before Day 178 in 1994	After Day 178 in 1994
Involuted	15	48	27	0
Clockwise	69	37	59	91
Counter-clockwise	9	10	0	9
Missing Days	9	5	14	0

**Figure 4.20a**  
Occurrence of Hysteresis in Sediment Concentration & Discharge, Slims River, 1994



**Figure 4.20b**  
Occurrence of Hysteresis in Sediment Concentration & Discharge, Slims River 1993



1=INVOLUTED HYSTERESIS 2=CLOCKWISE HYSTERESIS 3=COUNTER-CLOCKWISE HYSTERESIS

## **4.2.5 Causes of Hysteresis**

Explanation entails determination of the physical reasons that account for the production of higher sediment concentrations on the rising hydrograph limb. To this end, hysteresis is thought to take place both because of differential dilution of sediment supply by varying flow volumes on the rising and falling limbs and supply or selective supply exhaustion in supply limited situations.

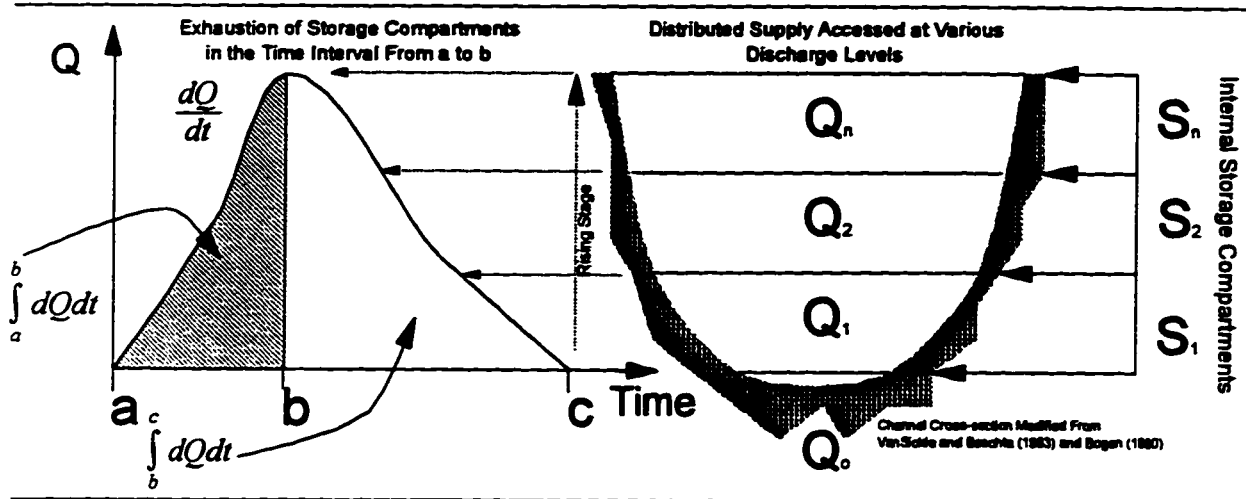
Dilution and supply exhaustion may act together to determine the occurrence of hysteresis and the magnitude of suspended sediment concentration on the rising hydrograph limb. The mechanisms for the production of hysteresis interact over individual and successive events.

### **4.2.5.1 Individual Events**

If there is a constant external sediment supply, e.g., a background suspended sediment concentration, then hysteresis may take place when there is a greater rate of change of discharge and therefore smaller total flow volume on the rising hydrograph. In other words, although diurnal discharge is periodic it is not strictly sinusoidal, but asymmetrical (Figure 4.21). As such, the total flow volume under the hydrograph to peak discharge is less than the total flow volume of recession or *vice versa* (Figure 4.21). It can be hypothesized that, if a smaller flow volume on the rising hydrograph limb exhausts internal channel supply compartments then sediment concentration on the rising limb should be greater than that of the falling limb and hysteresis will take place.

The ratio of the rate of change, approximated by linear regression, and flow volume on each rising and falling hydrograph limb were calculated for 1994. Double peaks are infrequent (Figure 4.30; Appendix B) and were treated as separate events in the analysis. Since ratios are used, numbers less than one indicate a smaller rate of change of discharge or greater total flow volume on the falling limb of the hydrograph when compared to the rising limb. Numbers greater than one indicate that the rising limb has a greater rate of change of discharge or greater total flow volume

**Figure 4.21**  
*Dilution and Supply Underlying Hysteresis Effect*



Given that,

$$Q'(t) \equiv \frac{dQ}{dt}$$

then,

$$Q(t) = \int_t^{t+\Delta t} \frac{dQ}{dt} dt$$

and from the graph of  $dQ/dt$ , it is clear that,

$$\int_a^b \frac{dQ}{dt} dt < \int_b^c \frac{dQ}{dt} dt$$

and since concentration,  $C_s(t)$ , is a function of total load,  $M_T(t)$ , divided by discharge,

$$C_s = \frac{M_T(t)}{\int_t^{t+\Delta t} \frac{dQ}{dt} dt}$$

then given a constant input of external sediment,  $M_E(t)$ , sediment in storage compartments,  $S_1 \dots S_n$ , is first made available to the rising discharge increment and therein partially or wholly exhausted. So, the concentration on the rising limb,  $C_r$ , will be greater than the falling limb,  $C_f$ , such that,

$$C_r = \frac{\sum_{i=1}^n S_i + M_E(t)}{\int_a^b \frac{dQ}{dt} dt} > C_f = \frac{M_E(t)}{\int_b^c \frac{dQ}{dt} dt}$$

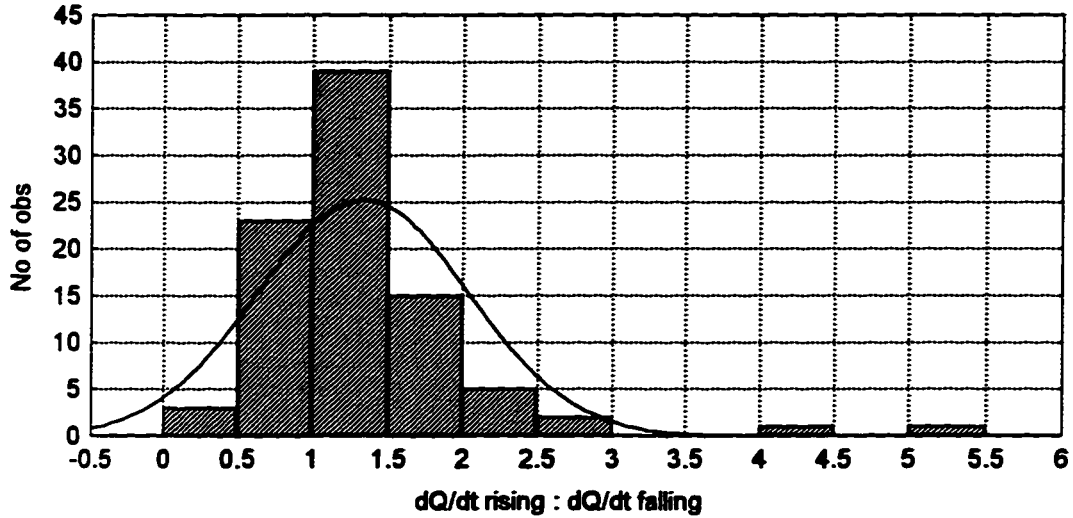
Higher concentrations on the rising limb are both a function of flow volume, represented by the time to hydrograph peak, and the number of internal storage compartments accessed at various stages. Greater rates of change on the rising limb should yield large concentration ratios because storage compartments will be exhausted by a smaller flow volume with the above symmetry.

with time, such that total flow volume and time to peak will be less than that of the falling limb flow volume and recession time. The number one itself, therefore, indicates a symmetrical mean rate of change and total flow volume. This analysis uses discharge as calculated from stage via Equation 3.13 and, therefore, assumes that stage accurately transfers to discharge. However, such an assumption is tenuous because velocity, channel roughness, water surface slope, and bedform development may be hysteretic with respect to stage, and if so, admits error when representing the rate of change of discharge or total flow volume from a stage record alone.

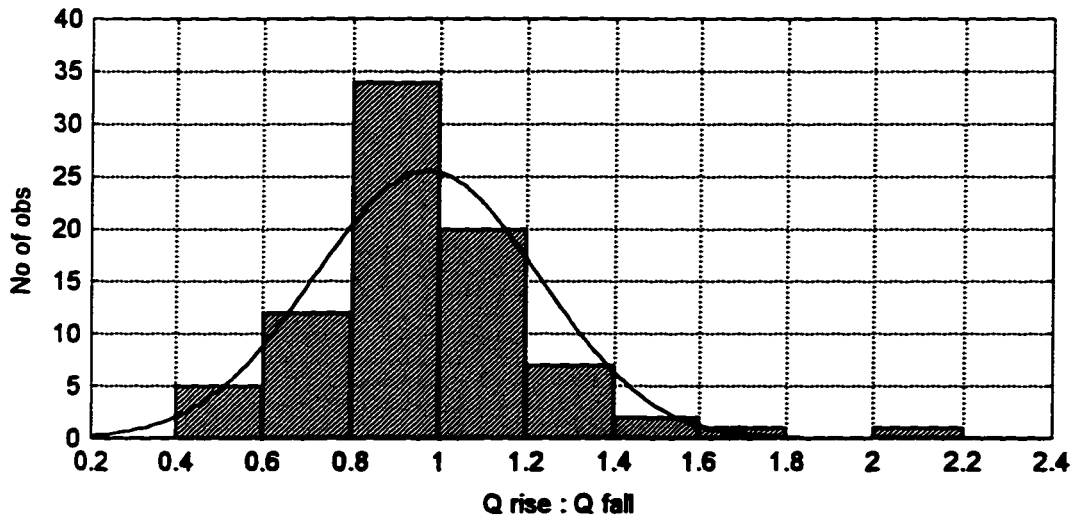
Analysis indicates that most diurnal hydrographs are asymmetrical with greater mean rates of change (Figure 4.22a) and smaller flow volumes on the rising limb (Figure 4.22b). This suggests that greater rates of change on the rising limb should be positively correlated with smaller flow volumes as is analytically true (Figure 4.21). However, Figure 4.21 is ideal, in that, discharges at  $t = a$  and  $t = c$  are equal, such that,  $Q(a) = Q(c)$ , but in the Slims River most often  $Q(a) \neq Q(c)$  because mean discharge varies over successive diurnal cycles and because of this, 84% of the time, greater rates of change of discharge on the rising limb do not correspond to smaller flow volumes. In other words, greater mean rates of change do not necessarily imply smaller flow volumes on the rising hydrograph limbs since they do not occur, for the most part, synchronously. Since the model in Figure 4.21 is concerned with flow volume it is this variable that is used in analysis with hysteresis.

For comparison with the occurrence of both clockwise and counter-clockwise hysteresis, the number one was assigned to flow volume ratios less than one and zero to those greater. If hysteresis occurred on a particular cycle it was given a number one or zero if it did not. These were then paired off with their respective flow volume ratios. The model presented in Figure 4.21 explains the occurrence of hysteresis 57% of the time in 1994. Consequently, *prima facie* there is no definitive relation between the total flow volume on the rising limb of a hydrograph and the

**Figure 4.22a**  
Distribution of Rising and Falling Limb Rate of Change of Discharge Ratio  
Sims River, 1994



**Figure 4.22b**  
Distribution of Rising and Falling Discharge Volume Ratios  
Sims River, 1994

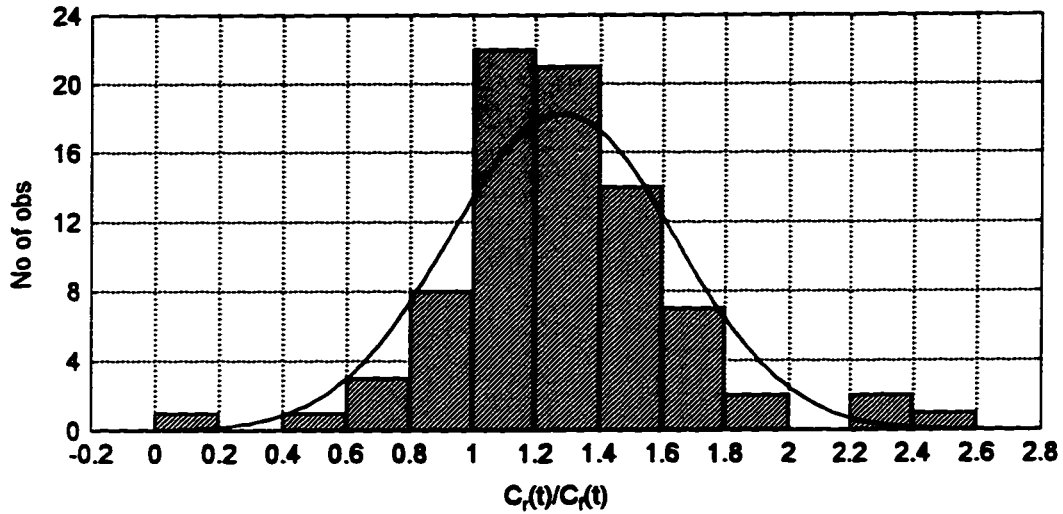


occurrence of hysteresis in the Slims River. In other words, hysteresis occurs regardless of flow volume on the rising limb. However, after day 171 in 1994, of the 31 days when flow volume on the rising limb was less than the flow volume on the falling limb, hysteresis occurred 81% of the time. It is noted that occurrence alone says nothing about the magnitude or timing of sediment concentration during a hysteretic cycle.

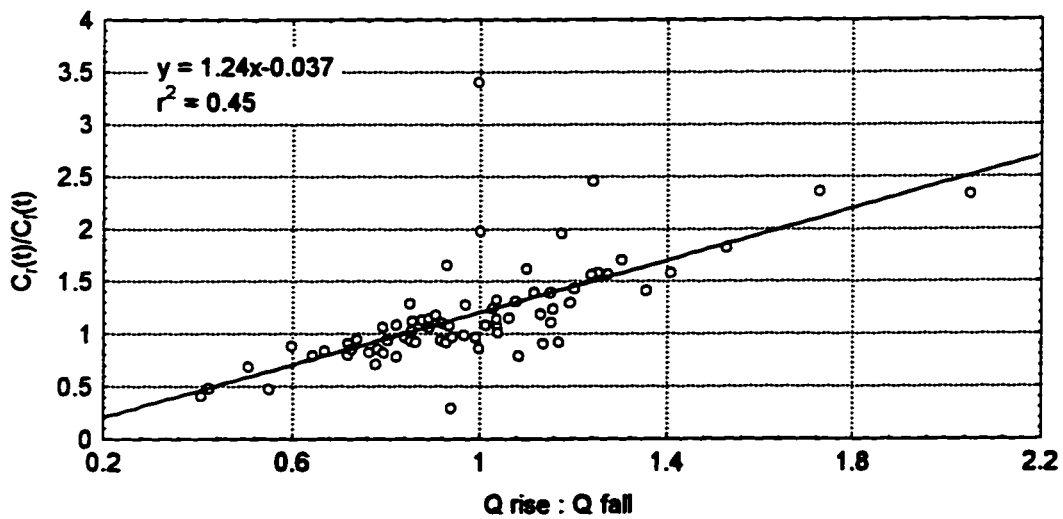
The flow volume ratio should be inversely proportional to the concentration ratio over successive events (Figure 4.21), assuming overcompetent flows. That is, concentration ratios should decrease as flow volume on the rising limb increases since the same amount of sediment is added to more and more water. In addition, the model in Figure 4.21 suggests that the concentration of sediment on successive hydrographs should be positively correlated with the rise in stage increment over the previous days stage, since, it is proposed, a greater rise in stage can tap more internal sediment supplies.

The distribution of suspended sediment concentration ratios for 1994, the ratio of concentration on the rising limb,  $C_r(Q_o)$ , to the concentration on the falling limb,  $C_f(Q_o)$ , for a given discharge,  $Q_o$ , indicate that, in general,  $C_r(Q_o) > C_f(Q_o)$  (Figure 4.23). There is no significant relation ( $r = 0.09$ ) between flow volume ratios and concentration ratios. This could mean internally that sediment availability is not constant over successive hydrographs. However, we should still expect sediment concentration ratios to increase if stage increases above the previous day's flow level since new storage compartments would be accessed. This is not the case, as concentration ratios are not significantly correlated with the stage increment above the previous day's flow ( $r = -0.01$ ). Rather, it is found that the total concentration ratio, that is, the total concentration on the rising or falling limb, is proportional to increasing flow volume ratios (Figure 4.24) (Figure 3.18ab, Appendix D). This means that more sediment is entrained by the rising limb of the

**Figure 4.23**  
Concentration Ratios for Suspended Sediment Concentration  
on the Rising and Falling Hydrograph Limbs, Slims River, 1994



**Figure 4.24**  
Relation Between Total Concentration Ratios and Flow Volume Ratios  
for Rising and Falling Limbs, Slims River, 1994



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hydrograph regardless of flow volume. Interpreted in terms of sediment availability: more sediment is available to the rising hydrograph limb. The question is why?

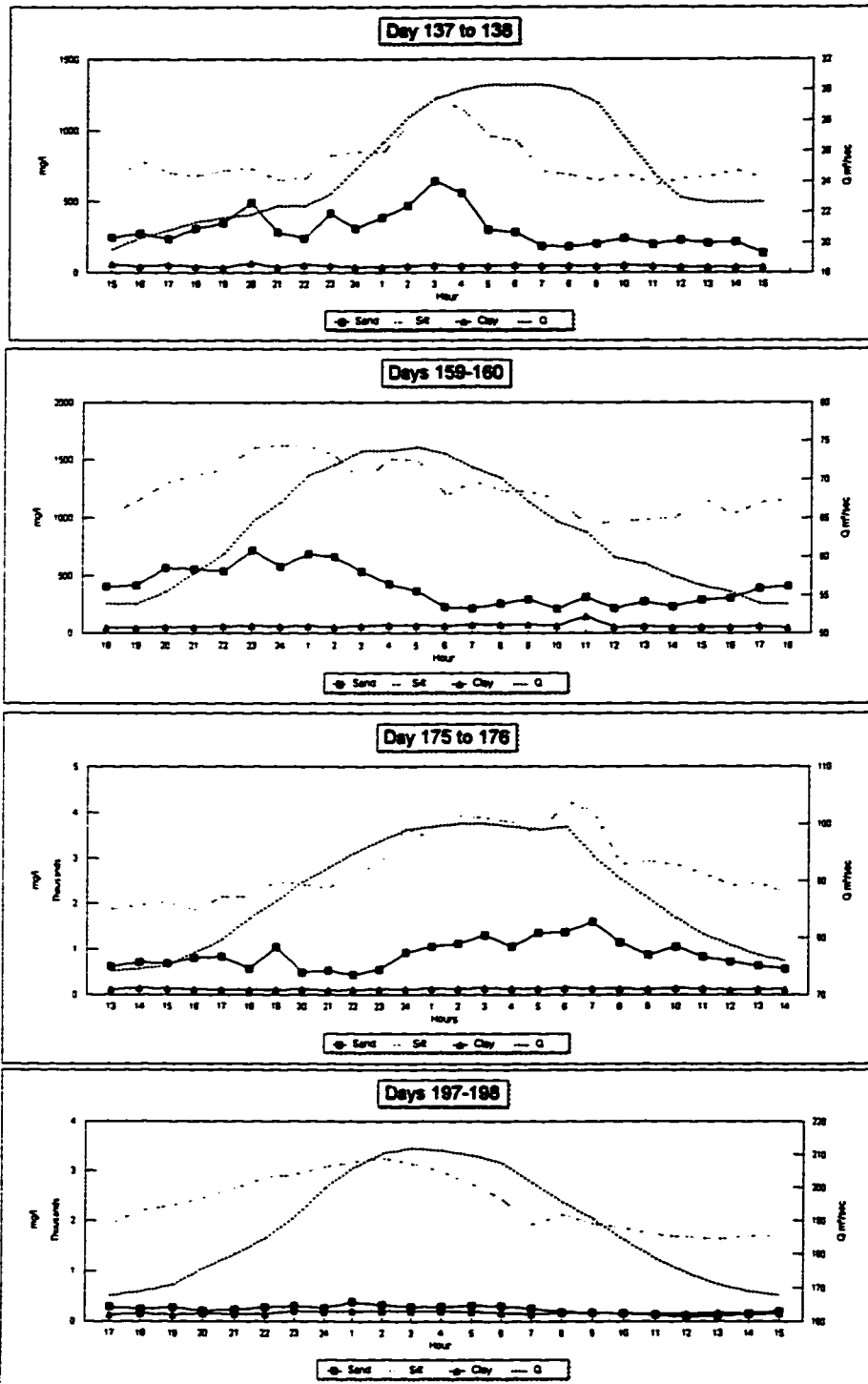
More sediment may be available to the rising limb of the hydrograph because there is initial liberation and selective exhaustion of readily transportable fine sediment deposited by the previous day's flood recession or accumulated since the last flood event (Sidle and Campbell 1985; Bogen, 1980; Paustian and Beschta, 1979; Collins, 1977). Thus, for well-developed individual hysteresis events there should be a greater concentration of fine sediment on the rising limb. Particle size analysis was completed for four diurnal cycles, three of which show clockwise hysteresis and one counter-clockwise hysteresis (Figure 4.25). For all events, silt and sand vary together and are both either higher on the rising limb if clockwise hysteresis takes place, days 137-8, 159-60 and 197-8, or on the falling limb if counter clockwise hysteresis takes place, days 175 to 176. In other words, there is no apparent 'flush' of fines on the rising limb. Consequently, there is no selective exhaustion over a flood hydrograph in the Slims River. However, the cycle of day 197-8 indicates that the proportion of sand has little influence on the suspended sediment regime of the late 1994 season.

#### 4.2.5.2 Successive Events

An unlimited supply of sediment or an insufficient discharge should yield a linear rather than hysteretic relation (Wood, 1977). Alternatively, the greater the exhaustion the more open the loop (VanSickle and Beschta, 1983) and the less likely hysteresis is to occur on temporally close subsequent events (Wood, 1977).

The concentration ratio's modal value indicates that suspended sediment concentration is between 1 and 1.2 times greater on the rising hydrograph limb for a given discharge (Figure 4.23). Consequently, it could be hypothesised that there should be less available sediment on a

**Figure 4.25**  
**Concentration of Sand Silt and Clay with Discharge**  
 Showing Variation Over Diurnal Cycles



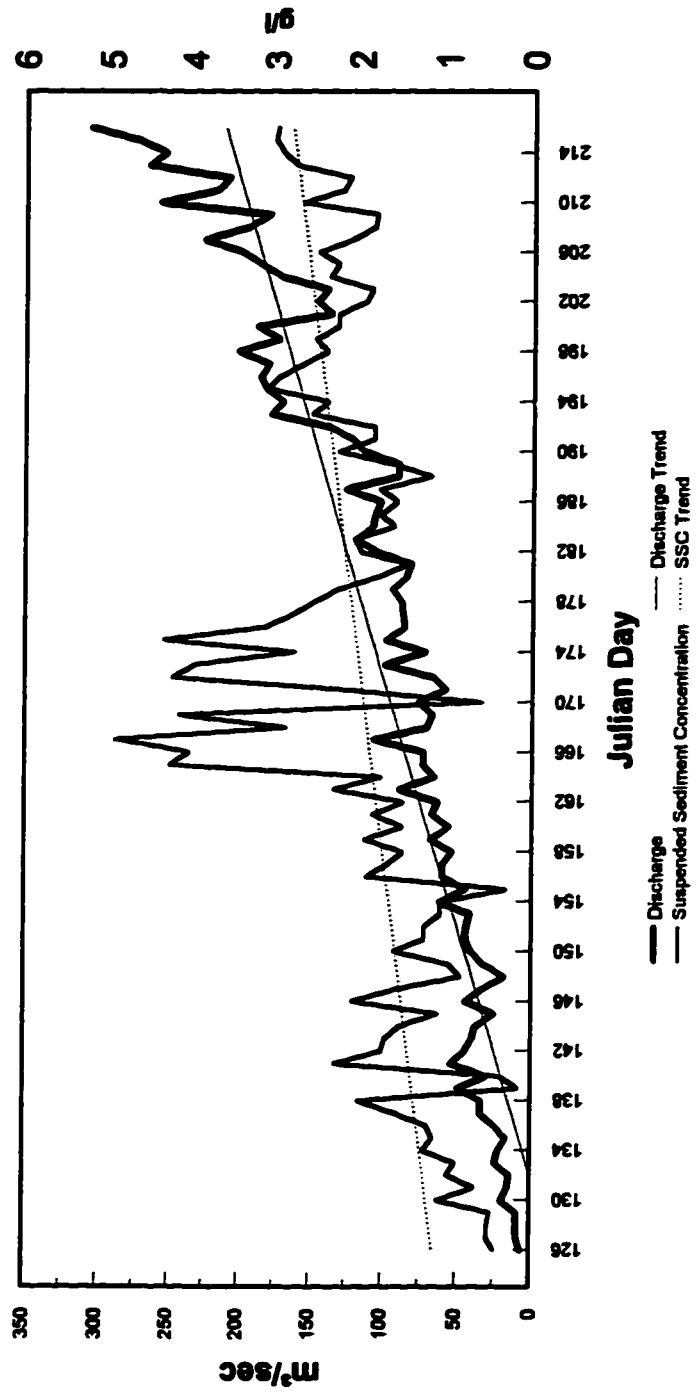
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successive event, if in fact, hysteresis entails sediment exhaustion. The problem is; hysteresis is quite frequent and, in spite of this, linear regression lines still show that discharge and suspended sediment concentration increase over the 1994 monitoring period (Figure 4.26).

Furthermore, Walling and Web (1982) propose that during stormflow hysteresis events, a horizontal relation between sediment concentration and discharge may reflect relatively constant background sediment concentrations in the stream from external inputs. In the later season of 1994, successive hysteresis loops associated with diurnal discharge fluctuations suggest a similar relation. However, the trend is not horizontal but rather positively sloped, in that, hysteresis is taking place over successive cycles but sediment concentration is also increasing (Figure 4.18). Consequently, this trend indicates limited vertical displacement of maximum and minimum sediment concentrations over successive hysteresis loops and therefore may be interpreted in terms of sediment availability. Suspended sediment may be higher on the rising limb because that is where internal channel supplies are first made available, for example, if channel erosion is the dominant supply mechanism then maximum suspended sediment concentration will be generated by the flow increment associated with the rise to peak discharge (Figure 4.21). Lower concentrations on the falling limb reflect exhaustion of supply on the rising limb as well as dilution by more flow volume given an asymmetrical hydrograph.

In the Slims River, because there is limited vertical displacement of successive hysteresis loops (e.g., Figure 4.18) and suspended sediment concentration increases over the season (Figure 4.26), there is apparently no significant exhaustion of supply between events. Consequently, no recovery period is necessary for replenishing sediment concentration between flood events in the Slims River as noted for other glacierized streams by Collins (1979a) and Kruzinsky (1993) and for non-glacierized streams by Walling and Web (1982), VanSickle and Beschta (1983) and Wood (1977). However, previous research is either proximal to the glacier or in absence of a glacier

**Figure 4.26**  
**Mean Daily Discharge and Suspended Sediment Concentration**  
**Slims River, 1994**



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and so may not necessarily apply to a fluvial-glacial environment. Alternatively, processes may act fast enough in the proglacial or glacial environment to replenish supply in the hours between events. On the other hand, hysteresis in the present situation is not a supply dependent phenomena but may be due to hydraulic conditions.

#### **4.2.6 Explaining Observed Hysteresis**

The aforementioned observations appear to be contrary to contemporary reasoning regarding the causes of hysteresis in fluvial streams and question whether conventional reasoning on the causes of hysteresis applies to a glacial/proglacial environment. Separately, none of the aforementioned causes of hysteresis can clearly explain the observed short-term suspended sediment regime in the Slims River. Therefore, there must exist a mechanism that can produce hysteresis every day, and apparently exhaust supply, but not decrease sediment concentrations on successive diurnal cycles. There are four possible explanations that fit the observations for 1993 and 1994, although no empirical evidence exists to aid in resolving the problem. The preferred explanations are as follows.

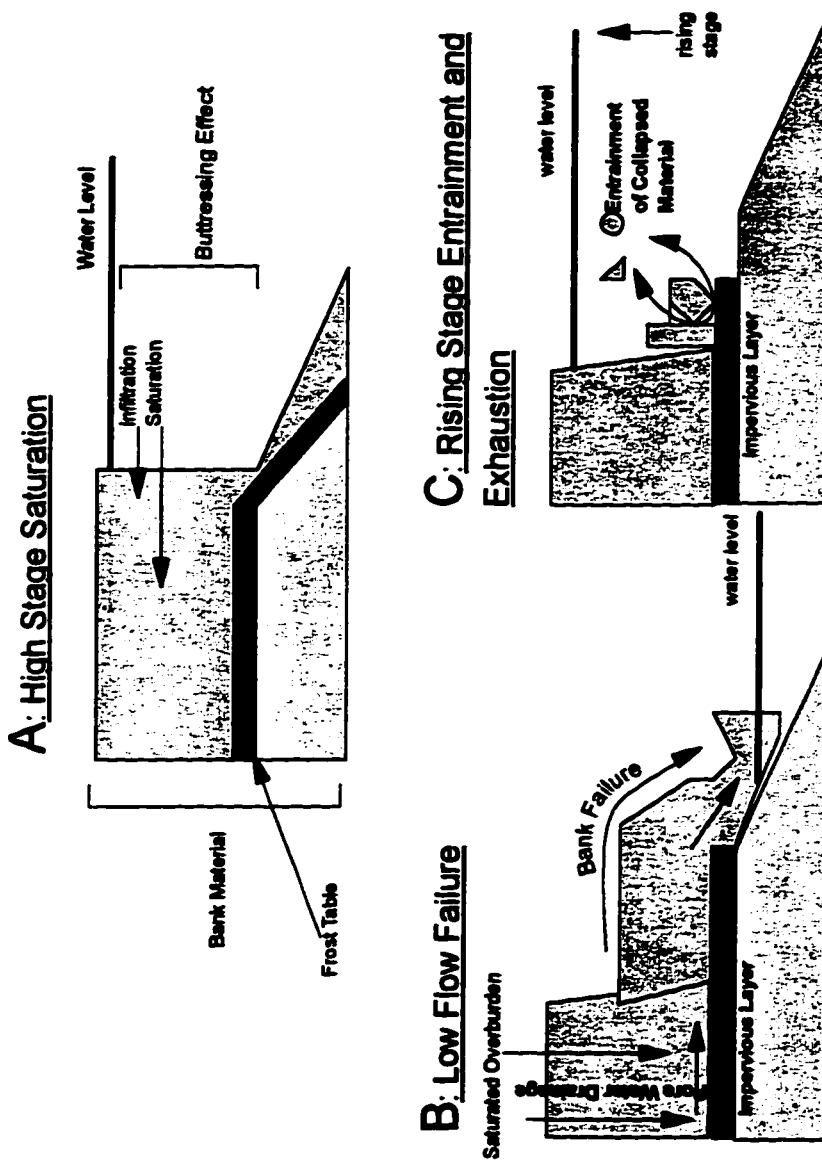
First, in the early season of 1994 and most of 1993, as evidenced by the more frequent broken loops, sediment supply is not limited and the rating curve of discharge and suspended sediment concentration better approximates a linear relation over successive events. This is supported by the rating plot for May of 1994 and 1993 where 68% and 50% respectively of the variation in suspended sediment concentration is explained by streamflow as the only independent variable. With spring snowmelt, in May and early to mid June, discharge increases over winter low flow conditions. The increasing discharge may cause channel avulsions as the Slims River establishes main channels in the valley train. Smith (1985a) suggests that this is common in valley trains as iced channels flush out and become integrated into efficient runoff systems in the spring. These valley train adjustments will cause variable sediment input to the stream and sediment pulses in the

form of multiple involutions in the daily rating plots. Such a scenario may also help explain the greater concentrations in the 1993 season.

The second explanation involves a relation between internal channel sediment sources supplied by bank collapse and their preferential removal by the rising hydrograph limb. As flows increase with the beginning of ice-melt, channels are well established in the sandur from the snowmelt runoff. At the same time increasing discharge provides the right conditions for bank undercutting and over-steepening during high flows. As fine sediment (< 63 microns) is difficult to move when wet due to cohesion, over-steepened banks would collapse during low flow and expose new sediment sources for entrainment during the next rise in stage. These sources would then locally be exhausted and the process repeated (Figure 4.27). This mechanism could allow higher concentrations on the rising limb of the hydrograph regardless of flow volume since it is essentially a supply limitation. This supply mechanism could supply sediment on short time scales leaving a necessary recovery period of only a few hours. So, assuming that the present hysteresis may be due to a supply limiting mechanism, namely bank collapse, the next task is to determine if the amount of bank collapse necessary to sustain hysteresis is reasonable.

Average concentration ratios for 1994 show that the mean percentage concentration difference between the rising and falling limbs is 14%, and range between 10% and 18% with 95% confidence. Since  $1.8 \times 10^6$  tons of sediment are transported over the monitoring period and sand and silt fractions vary together on the rising and falling limbs, 14% of the total load is  $2.5 \times 10^5$  tons. Assuming the background sediment concentration to be supplied by external sources (such as the Kaskawulsh glacier), then, to produce hysteresis, bank collapse must account for approximately 23.67 kg/s of material or 1.82 g/m/s for the 13 km where river silts are present. This means transport of 2045 tons/day above background concentrations, which assuming a

**Figure 4.27**  
**Mechanism of Bank Collapse for Maintenance of Hysteresis**



#### 4.2.6 Explaining Observed Hysteresis

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material density of  $2.045 \text{ g/cm}^3$  and banks 0.5 m in height, represents a daily bank retreat of approximately 15.4 cm/day.

These values are lower in the early season. For example, between day 124-143 bank collapse would have to supply 6.2 kg/s or 0.48 g/m/s in the 13 km where river silts are present in the early season to produce daily hysteresis. The above figures assume a single linear channel and so the actual amount of bank collapse per channel needed to sustain hysteresis and concentration would be much less than these figures since the Slims River is braided. In this case, the numbers are not unreasonable and bank collapse could explain the increased concentration on the rising limbs of the hydrographs during some events.

Thirdly, with the dramatic increase in hysteresis in July of 1994 (91% of the time) there must be a mechanism capable of both allowing sediment to be exhausted daily and replenished daily. At this time hysteresis may be produced in the subglacial channels and translated without modification through the proglacial zone. This too is not unreasonable considering that in the later season the percentage composition of suspended samples is predominantly washload sized material, a size fraction for which bank material is deficient. So, washload may be transported through the proglacial zone without modification. Consequently, bank collapse in the later season may not play a major role in the production of hysteresis at that time.

In 1994 counter clockwise hysteresis takes place at the beginning of icemelt and in late July at times of increasing discharge amplitude. In areas proximal to the terminus this type of concentration increase may be cited as a sediment pulse or flux event (Johnson, 1991a).

Although, counter-clockwise hysteresis is seen in 1993 when, presumably, maximum avulsions are taking place and there is little glacial influence. For counter-clockwise hysteresis to take place there must be an extraneous sediment supply over and above the supply on the rising limb. The

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Kaskawulsh Glacier is certainly capable of producing such a phenomenon. Alternatively, since discharges are orders of magnitude greater than the early season, it is possible that a channel avulsion at this time on the falling limb would transport enough sediment to cause counter clockwise hysteresis.

Fourth, hysteresis in a non-supply limited supply situation could be hydraulically controlled where greater stream velocities exist on the rising limb of the hydrograph and mobilize more bed or traction load. Higher velocities on the rising limb may be due to bedform and channel roughness differences although there is presently no evidence of these. This may not be unreasonable considering that discharge is dependant on velocity in the measurement reach, and so minor variations in velocity over a hydrograph may have considerable effects on sediment entrainment.

#### 4.2.6.1 Discussion

To summarize, based on conventional reasoning and the model presented in Figure 4.21, while, it is expected that a number of features should be true of the relation between suspended sediment and discharge however they were not found to be so at the monitoring site. These differences are:

1. Hysteresis should occur most often with smaller flow volumes on the rising hydrograph limb; it does not;
2. Flow volume ratios should be inversely related to concentration ratios; they are not;
3. Concentration ratios should be proportionally related to the stage height above previous days stage; they are not;
4. Selective exhaustion should be evident over successive loops; it is not;
5. Suspended sediment concentration should decrease over successive events because of supply exhaustion; it does not.

Observed hysteresis may an equifinal product of hydraulic, glaciologic, and sedimentological factors. Based on contemporary reasoning and observations of hysteresis found in the Slims River, a model of sediment transport is not yet attainable. Explanations presently remain impressions and limited findings from a few tested hypotheses, which indicates that further inquiry

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is needed to help decide between the various scenarios presented. More detailed data on hydraulic geometry over diurnal cycles and bank migration, in particular, is needed to attain a more complete picture of this environment and its production of hysteresis.

### **4.3 Dissolved Solids**

The in-channel solute concentration at any time is the sum of numerous specific concentrations derived from denudational processes like dissolution of bedrock, pollution, biological action and nutrient cycling processes (Richards, 1982, p.47,91). However, in basins comprised mainly of glacier ice as well as exposed rock, solutes are mainly derived from the atmosphere through precipitation and ice storage as well as from the lithosphere through liberation of ions from tills, moraine materials, and bare rock (Collins, 1979b; Zeman and Slaymaker, 1975). Atmospheric sources of solute have a benign effect on meltwater hydrochemistry as indicated by the very dilute nature of supraglacial streams (Lemmens and Roger, 1978). The ionic exchange capacity between the water and sediments, either suspended or in the form of basal till or moraine, influences meltwater hydrochemistry and is favoured with increased contact time which leads to solute enrichment and higher conductivity (Collins, 1983; Collins, 1979b; Lemmens and Roger, 1978). Furthermore, Zeman and Slaymaker (1975) found that runoff source areas vary spatially and dominate flow hydrochemistry at different times during a melt season. Consequently, any explanation of alpine hydrochemistry must take spatial and temporal variation of source runoff into account.

Non-glacial streams exhibit a non-linear dilution processes in which almost constant baseflow solute concentrations are diluted by quickflow or direct runoff during storms (Richards, 1982, p.95). In the case of glacierized basins, atmospherically-derived solute concentrations come from snow and ice-melt rather than liquid rainfall. Meltwater dilution results in a chemograph trough

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near peak discharge which because of hysteresis is rarely the mirror image of the hydrograph peak (Richards, 1982, p.95; Collins, 1979b). The dilution effect yields, on the long term, an inverse relation between solute and discharge and gives rise to empirical models of solute transport of the form  $c = aQ^b$  (where  $c$  is solute concentration,  $Q$  discharge, and  $a$  and  $-b$  (slope) empirically derived constants). This rating generally holds for most glacial and non-glacial streams but residual variability around most curves indicates that the conductivity - discharge relationship is rarely linear (Collins, 1983; Richards, 1982, p.91; Collins, 1979b).

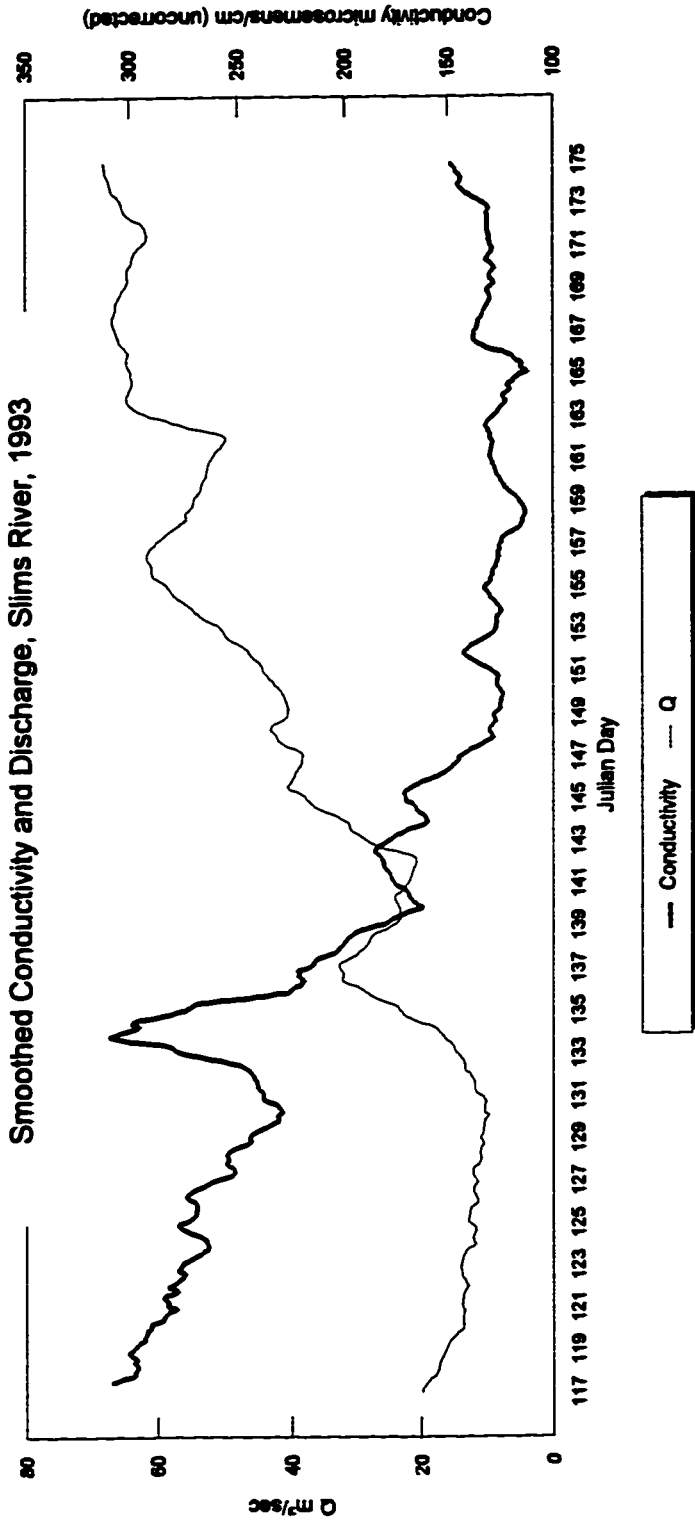
### 4.3.1 Seasonal Characteristics

For both 1993 and 1994 there is an overall decrease in the specific conductance of water samples extracted on an hourly basis.

In 1993 the trend is almost monotonic decreasing beginning at over 300 micro-siemens/cm and decreasing with a local maximum occurring between day 122 to 129 and then decreasing again to the end of the monitoring period to less than 150 micro-siemens/cm. There is evidence of another increase in conductivity near the end of the monitoring period (note that conductivity values for 1993 are not corrected to 25 °C and so appear lower than those of 1994) (Figure 4.28a)

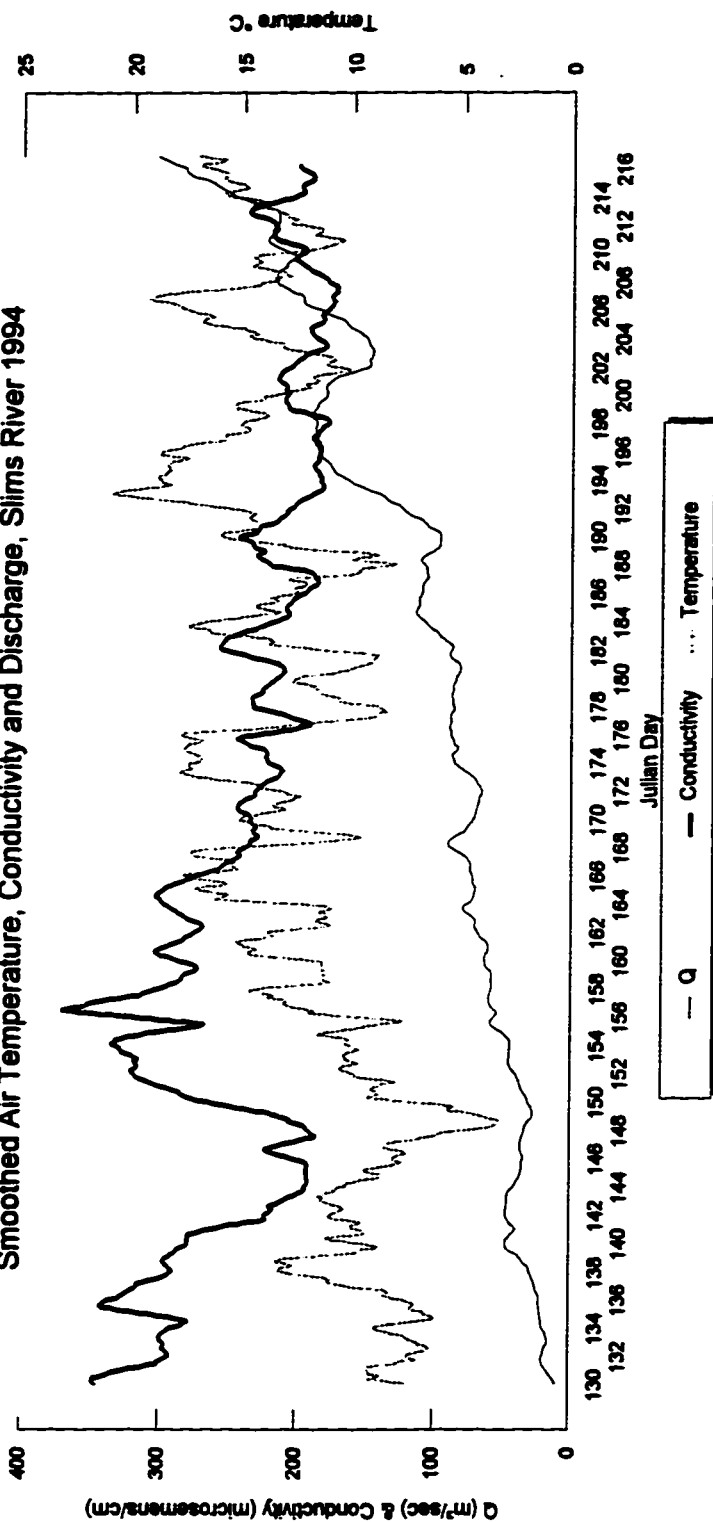
Within the decreasing trend of 1994 there are two major peaks in the concentration record (Figure 4.28b). First, conductivity is initially high, up to 390 micro-siemens/cm, and decreases to approximately 170 micro-siemens/cm by day 148. Second, from day 148 to 157 conductivity rises to a peak of over 400 micro-siemens/cm, and then decreases until the end of the monitoring period with minor local maxima and minima. The minimum and maximum conductance values for both years are greater than those observed in most glacier streams (Collins, 1979b).

**Figure 4.28a**



**Figure 4.28b**

Smoothed Air Temperature, Conductivity and Discharge, Slims River 1994



#### *4.3.1 Seasonal Characteristics*

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In the snowmelt period of runoff greater proportions of flow are derived from tributaries below the Kaskawulsh terminus. Water flowing through colluvial and alluvial materials will tend to become enriched in ions and supply greater amounts of dissolved solids than 'fresh' ice or snowmelt contributions alone (Lemmens and Roger, 1978). Proximal to the glacier, higher concentrations are expected when the sub-glacial hydrological system opens and enriched water that was in contact with basal materials is released (Brown and Tranter, 1990; Binda, 1984, p.85; Collins, 1983; Collins, 1979b). Once the glacier system is 'purged' conductivity decreases because of dilution that accompanies addition of 'fresh' ice meltwaters during sustained ablation.

From the data in 1993 and the same time period in 1994, it is suggested that the contribution of dissolved solids in the early season is derived from snowmelt through colluvial and alluvial materials in the lower basin tributaries. In 1994 there is a significant second peak in conductivity that corresponds to the opening of the subglacial hydrological system which is not correlated with precipitation, major increases in temperature or discharge, it must therefore reflect the expulsion of ion-enriched subglacial waters. There is no secondary peak in conductivity in 1993 that would suggest the opening of the subglacial system. Following the second 1994 peak conductivity decreases with increasing dilution by icemelt and greater diurnal fluctuations indicative of control of the bulk hydrochemical regime by sustained ablation. It is apparent in 1994, that as icemelt contributes greater proportions of water to the Slims it acts to dilute groundwater and subglacial contributions of dissolved solids to the flow, and causes proportional decreases in conductivity over the melt season.

In 1994, moving averages of air temperature and discharge and conductivity suggest that in general low temperatures occur first and then lower discharges and high conductivities (Figure 4.28b). Consequently, dilution of ion-rich meltwaters was maximal with high air temperatures and sustained ablation of the glacier surface which is common in streams proximal to glaciers

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(Bradley, 1990, p.65; Binda, 1984, p.78; Collins, 1979b). This relation is clearer after day 162, since before that date temperature and conductivity changes are more in phase, and support the idea that solute concentrations in the later season are glacially-controlled.

In 1994, precipitation events acted to decrease the diurnal amplitude of discharge fluctuations. During this season, following precipitation events the smoothed conductivity record shows local maxima (Figure 4.2a). This relation is poorly defined not evident in the 1993 record. Such observations suggest that some dissolved solids are acquired from non-glacierized areas in the lower basin during these events. However, since local maxima occur more often with no measurable precipitation, and are preceded by local minima in the temperature record, it is more likely that temperature is the primary independent variable in the relation. The global maximum and minimum conductivity values observed are greater than other proglacial streams. The secondary peak in conductivity that occurs in 1994 corresponds to the opening of the glacier system and therefore suggests that such concentrations can be produced and maintained in the subglacial system. In this particular case, this is not wholly unlikely considering that the source of the Slims is a number of subglacial resurgences.

Generally, the conductivity record suggests that because of the second peak in the 1994 season, the increase in discharge after day 148 can be explained by the release of subglacially stored waters which have become chemically enriched at the glacier bed. The conductivity record of 1993 in absence of a secondary peak in conductivity suggests that it is mainly controlled by proglacial snowmelt runoff.

### 4.3.2 Conductivity Discharge Relation

In both seasons discharge and conductivity exhibit an inverse relation with a correlation of  $r = -0.61$  in 1994 and  $r = -0.82$  in 1993 (uncorrected for temperature) and consequently discharge

#### 4.3.2 Conductivity Discharge Relation

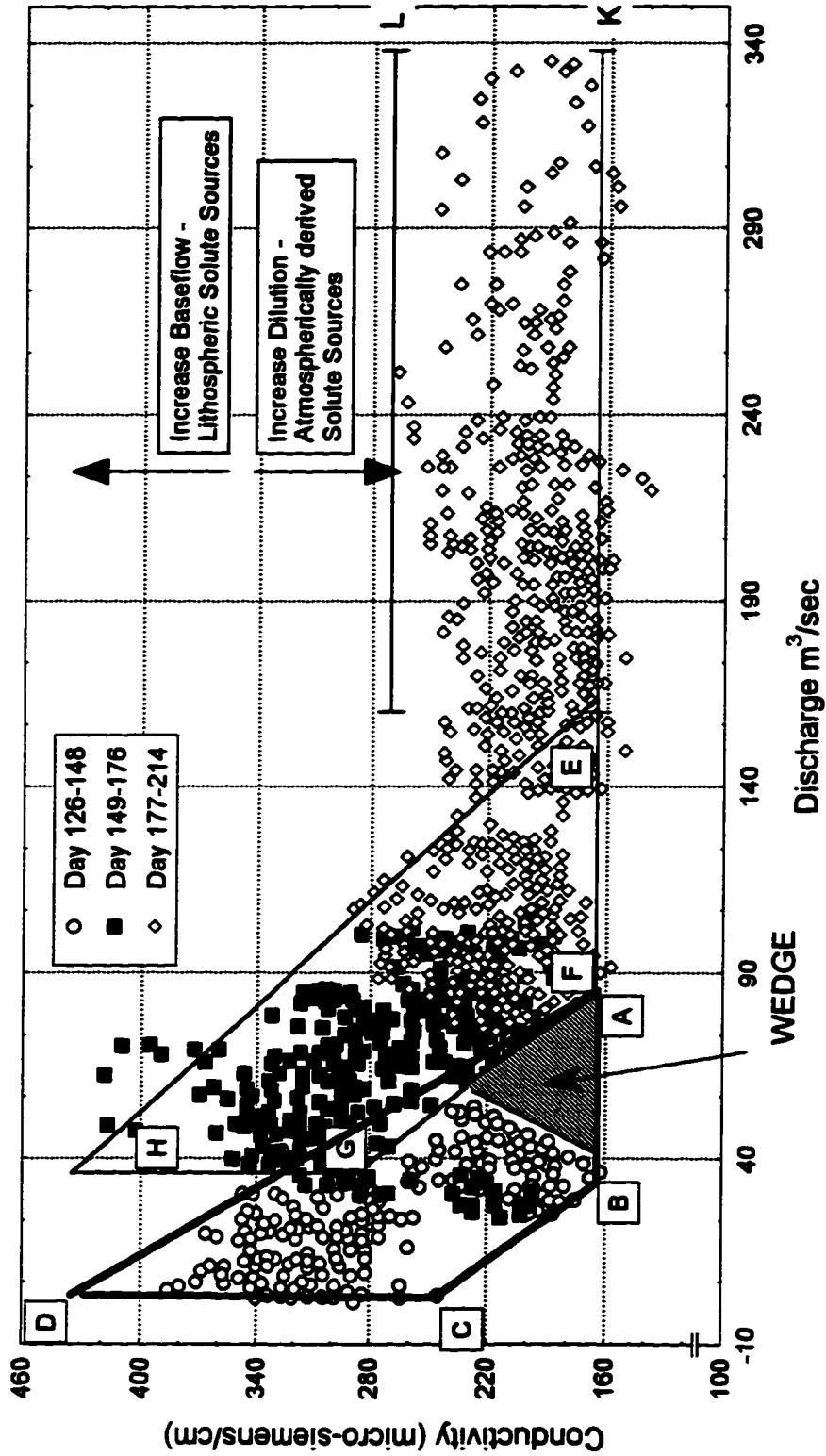
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explains less than 37% and 67% of the variation in conductivity in the two years respectively (Figure 4.29ab). In studies of sites proximal to glaciers, discharge has explained 65% (Bajewsky, 1985) and up to 89% and 97% of the variance in conductivity (Brown and Tranter, 1990). Conversely, Collins (1979b) found discharge to explain very little of the variation in conductivity for the Gornergletscher in the Swiss Alps.

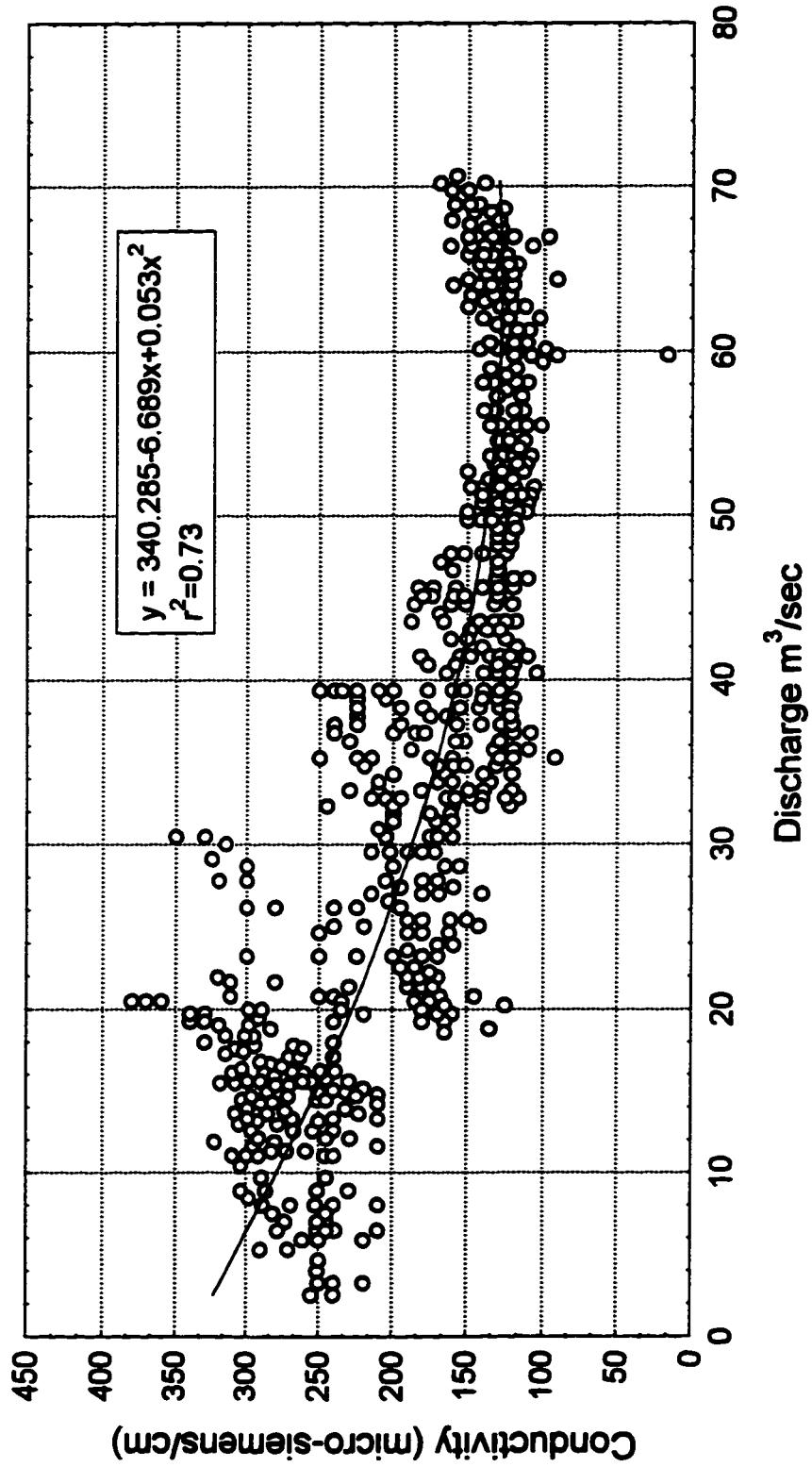
Bryan (1974c) found that as discharge increases in the Slims River there is a commensurate decrease in dissolved solids as measured by correlation of total alkalinity ( $r = -0.89$ ), total hardness ( $r = -0.93$ ) and  $\text{SO}_4$  ( $r = -0.89$ ) with discharge. This assessment may not be realistic considering that Bryan's sampling regime only involved monitoring of "several" 24 hour periods over the season. The present sampling resolution and number of observations raises some uncertainty regarding previous results. By way of illustration, in 1993, with only 58 days of monitoring, the relation between conductivity and discharge is much stronger than in 1994 when there were a total of 88 days of observation. *Prima facie* there is an apparent inverse relation between explanation of variance. However, this may also be due to year-to-year variance in runoff sources and routing.

Collins (1979b) suggests that because of a two-component englacial and subglacial routing system, the discharge conductivity relation in glacierized streams is best described by the margins of a trapezoid. Unfortunately conductivities of specific component contributions to flow, such as proglacial, subglacial, englacial and supraglacial were not monitored for logistical reasons. However, the variations in conductance with discharge at the gauging site attests to changes in dominant routing mechanisms both throughout the season and periodically. In the Slims River dissolved solids are related to three rather than two related subsystems, namely, englacial, subglacial and proglacial inputs which dominate Slims discharge at different times of the year.

**Figure 4.29a**  
Relation Between Conductivity and Discharge for 1994 with Trapezoidal Frame



**Figure 4.29b**  
Inverse Relation Between Conductivity and Discharge for 1993



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These routing mechanisms are made alluded to by the examination of the scatterplots in 1993 and 1994.

Two trapezoids characterize the distribution of conductivity and discharge over the monitoring period of 1994 (Figure 4.29a) (Figure 3.19ab, Appendix D). First, proglacial routing dominates from day 126 to 148 and is bounded by trapezoid ABCD, whereas glacier routing begins on day 149-177 and is bounded by trapezoid EFGH and dominates after day 177 and is bounded by two horizontal line segments K and L. In 1993 only proglacial routing is evident in the conductivity discharge record and its bounding trapezoid is similar to ABCD of the early 1994 season (Figure 4.29b) (Figure 3.20ab, Appendix D).

Points D (of ABCD) and H (of EFGH) are located at the maximum observed conductivities derived at low discharges from snowmelt runoff from the tributary streams and the opening of the subglacial system respectively.

Trapezoid ABCD is deficient in observations between AB, a segment that normally echoes the lowest conductivities from ice and snow meltwaters from the glacier surface during minimum subglacial output (Collins, 1979b). In this deficient area there is a well defined 'wedge' separating ABCD from EFGH. Observations are missing here because at this time the glacier drainage system begins to open - rises to the secondary peak in conductivity - and a new distribution is formed. This is not evident in the record of 1993 and supports the idea that ABCD in 1994 represents proglacial meltwater routing. Observations close to B represent the last of the tributary derived snowmelt. At this time conductivity decreases towards atmospherically derived levels because the accumulated evaporative deposits and long residence time water in the alluvial and colluvial materials has been depleted after maximum snowmelt discharge.

#### *4.3.4 Periodic and Hysteretic Relations*

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Segment EF (Figure 4.29a) represents meltwaters with characteristics derived primarily from atmospheric inputs (Collins, 1979b), namely, snow and icemelt on the glacier. These values would probably be less if the source of the river was not the subglacial resurgences. From day 177-214 an almost constant range of conductivity exists and is bounded by two line segments K and L. The values near K indicate maximum dilution by waters with concentrations approaching atmospherically-derived values and the upper line, L, represents the maximum conductivities of glacier night-time low flow from evening termination of ablation. Similar ranges of conductivity are evident in the early season at the end snowmelt, but considering the amount of flow derived from glacier snow and ice-melt in the later season, tributary contributions will effectively be diluted. Consequently, these conductivity values in the later season must reflect diurnal variations in glacier output.

In short, ABCD lacks observations in the wedge, that would be filled with observations representative of conductivities derived from atmospheric inputs as snowmelt in the basin receded and glacier melt begins. This is supported by the 1993 record which lacks a second conductivity peak. But, since the glacier system opened in 1994, a second distribution is formed that more fully represents a trapezoid. The maximum dilution is consistent over both monitoring periods whereas the maximum concentrations are only consistent in the later season of 1994.

#### **4.3.4 Periodic and Hysteretic Relations**

In 1994 (but not clearly evident in 1993) diurnal fluctuations of conductivity are better defined during the later season when meteorological conditions favour sustained ablation of the glacier. There is a marked increase in diurnal conductivity amplitude after day 173 in 1994 that corresponds to the timing of maximum diurnal discharge amplitude. Similar observations were found by Brown and Tranter (1990) with glacier discharge during uninterrupted diurnal cycles. At these times Collins (1979b) notes that restoration of surface ablation in the morning provides

#### 4.3.4 Periodic and Hysteretic Relations

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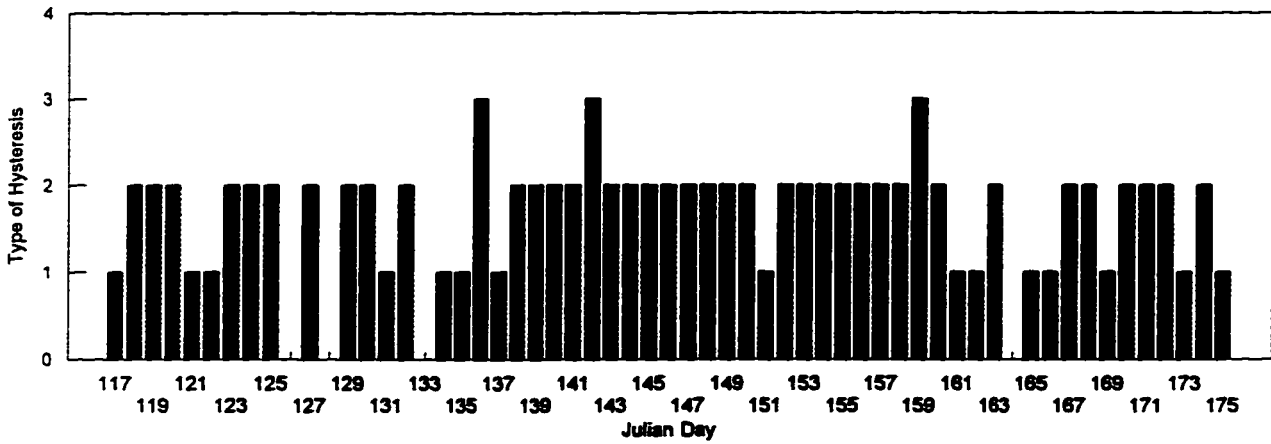
an increase in discharge at the glacier output that is paralleled by a drop in conductivity as night-time low flow is diluted. As discharge declines conductivity rises, as the proportion of flow derived from subglacial storage/routing increases. Similar mechanisms act in the early season as high conductivity low flow is partially diluted by daily snowmelt.

In 1993 and 1994 conductivity undergoes repeated periodic diurnal fluctuations which are out of phase with those of discharge. A representative period showing this phase shift is best defined from the continuous *in situ* monitoring of conductivity during 1994 (Figure 4.36 - Appendix B). The in-situ data was synchronously monitored with temperature and is not subject to the error of conductivity from physical samples (Section 3.5). In spite of this situation, periodic fluctuations in hourly data are clear in both years and phase shifts between the two variables gives rise to hysteresis in rating plots for individual events.

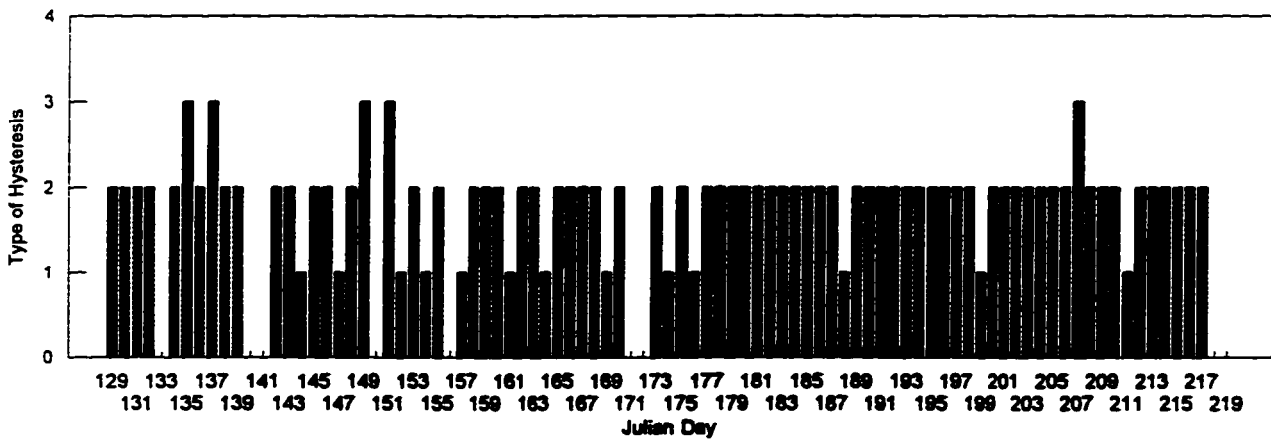
The scatter produces 67% and 37% residual variability, in 1994 and 1993 respectively, between conductivity and discharge and is largely explained by two factors: first, the presence of hysteresis in the diurnal relation of conductivity with discharge: second, different conductivity values for similar discharges on different diurnal cycles. Diurnal phase shifts between conductivity and discharge that give rise to hysteresis are easier to explain than fluctuating concentrations on successive diurnal cycles which depend on spatial and temporal changes of solute supply.

Hysteresis categories are defined identically to sediment concentration hysteresis (Figure 4.19). Clockwise hysteresis is most frequent in both 1993 and 1994 (Figure 4.30ab). Collins (1979b) notes that for the Gonerletscher with few exceptions each diurnal cycle showed clockwise hysteresis. Here, 63% and 72% of the time in 1993 and 1994, respectively, there is higher conductivity and total dissolved solids on the rising limb of the hydrograph when compared to similar discharges on the falling limb. Individual ions may deviate from this relation but the extent

**Figure 4.30a**  
Occurrence of Hysteresis in Conductivity Discharge Relation, Slims River 1993



**Figure 4.30b**  
Occurrence of Hysteresis in Conductivity Discharge Relation, Slims River, 1994



1=INVOLUTED HYSTERESIS 2=CLOCKWISE HYSTERESIS 3=COUNTER-CLOCKWISE HYSTERESIS

#### *4.3.4 Periodic and Hysteretic Relations*

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of similarity is unknown as the sampling strategy did not allow for the construction of diurnal relations for individual ions with discharge.

Collins (1979b) suggests that clockwise hysteresis takes place because there is a delay of 'fresh' ablation meltwaters reaching the glacier portal once daily ablation begins. Once ablation begins the stream is still fed from a proportionally greater amount of enriched subglacial discharge. It takes time for the 'fresh' ablation meltwaters to reach the conduits and cleanse them of the solute rich waters. Consequently, on the rising limb of the days hydrograph concentrations are higher as meltwater production increases and forces out solute rich basal waters. Once peak flow is reached the system has been partially 'cleansed' and the night-time low flow component of discharge proportionally decreases as ablation meltwaters reach the portal in increasing quantities (Collins, 1979b). Such an explanation, while sufficient for the glacier melt dominated part of the monitoring period, is not sufficient to explain hysteresis in the early season during the nival dominated regime.

In non-glacierized basins the higher concentration on the rising limb is often attributed to an initial flushing of ion-enriched water from the soil or banks by throughflow initiated from rainfall. Alternatively, overland flow can tap evaporative deposits which are flushed out before peak discharge is reached. Hysteresis in the early melt season can may explained by considering the travel times of water from the tributaries to the monitoring site in relation to the timing of snowmelt.

On a daily basis in a particular sub-basin, slow-moving solute-rich waters, from the ground or in channel, are both forced out and diluted by increasing snowmelt in response to rising temperatures. When concentration time is reached and the furthest areas of the sub-basins are contributing meltwater to Slims discharge, most of the solute rich water has been expelled.

#### *4.3.4 Periodic and Hysteretic Relations*

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Solute concentrations will, therefore, be higher on the rising limbs because all of the solute rich waters are exhausted for a given discharge on the falling limb. The travel time of water increases with distance from the monitoring site and so solute rich waters take progressively longer to reach the gauging site. Thus, given that solute concentration is greater on the rising limbs of the tributary hydrographs, they will also be greater on the rising limb of the Slims River hydrograph at the gauging site. Concentrations will begin to fall as the peak discharge waters from the furthest tributary reach the monitoring site and maximum dilution of the respective aliquots has taken place. Consequently, hysteresis in the early season is a product of the relation between two related subsystems of concentrated groundwater throughflow and dilute snow meltwaters. In this sense hysteresis in both the glacier and non-glacier portion of runoff is a function of the concentration time of snowmelt runoff of the individual meltwater streams and exhaustion of solute rich waters on successive diurnal cycles.

Finally, hysteresis may be explained throughout the season if concentration is always high in the tributary meltwater streams. Solute-rich discharge from tributaries streams reaches the gauging site first. Dilute water from Kaskawulsh glacier melt and snowmelt on the glacier will always take longer to reach the gauging site. This differential delay will always produce higher concentrations of solute on the rising limb of the hydrograph thus producing hysteresis.

Collins (1979b) notes that at peak discharge 'figure-of-eight' subloops may occur because of discharge irregularities caused by the release of pockets of diluted englacial waters. In the case of the Slims River, small subloops often occur at peak discharge that are counter-clockwise, and for a small period of time there are greater solute concentrations on the falling limb of the hydrograph.

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### 4.3.5 Major Cations

Major anions were not analyzed because of budget constraints. Daily information on major cations is only available for the monitoring period of April-June 1993, since only six samples - two per month - were analyzed for 1994. The time period of 1993, while not as extensive as that of 1994, will give an indication of the relative proportions of cations in Slims River waters as some long term research in glacierized basins has shown that the relative proportions of major cations remains similar from year to year (Collins, 1979b). Furthermore, the monitoring strategy involved only one sample of water for analysis of cations each day and limits the inferences that can be drawn from water chemistry to general trends and does not allow for the discrimination of individual solute sources.

In 1993 the dominant ion in terms of concentration is  $\text{Ca}^{2+}$ , followed by  $\text{Mg}^{2+}$ ,  $\text{K}^+$  and finally  $\text{Na}^+$  respectively (Table 4.2a). Samples analyzed over the 1994 period suggest that the same order is maintained (Table 4.2b). Over both monitoring periods cations follow the general trend of conductivity with decreasing concentrations in response to dilution from increasing discharge (Figure 4.31). Metcalf (1984) found that the amounts of free hydrogen as measured by pH are greater in the early melt season at the Gornergletscher Glacier in Switzerland. No such observations are presently available but the uptake of solutes is not only dependent on free hydrogen but also on sufficient contact time between suitable material either from the ground environment or glacial abrasion (Bradley, 1990, p.26). At present, the concentration of individual cations is not constant as the season progresses because enriched water, regardless of pH, is flushed out and contact time is decreased with increasing discharge.

In terms of variation, over the 1993 monitoring period  $\text{Ca}^{2+}$  shows the least variation, as measured by the coefficient of variation, and  $\text{Na}^+$  has maximum variation. The Slims River shows less proportional variation as measured by the percentage ranges for all cations when compared to the

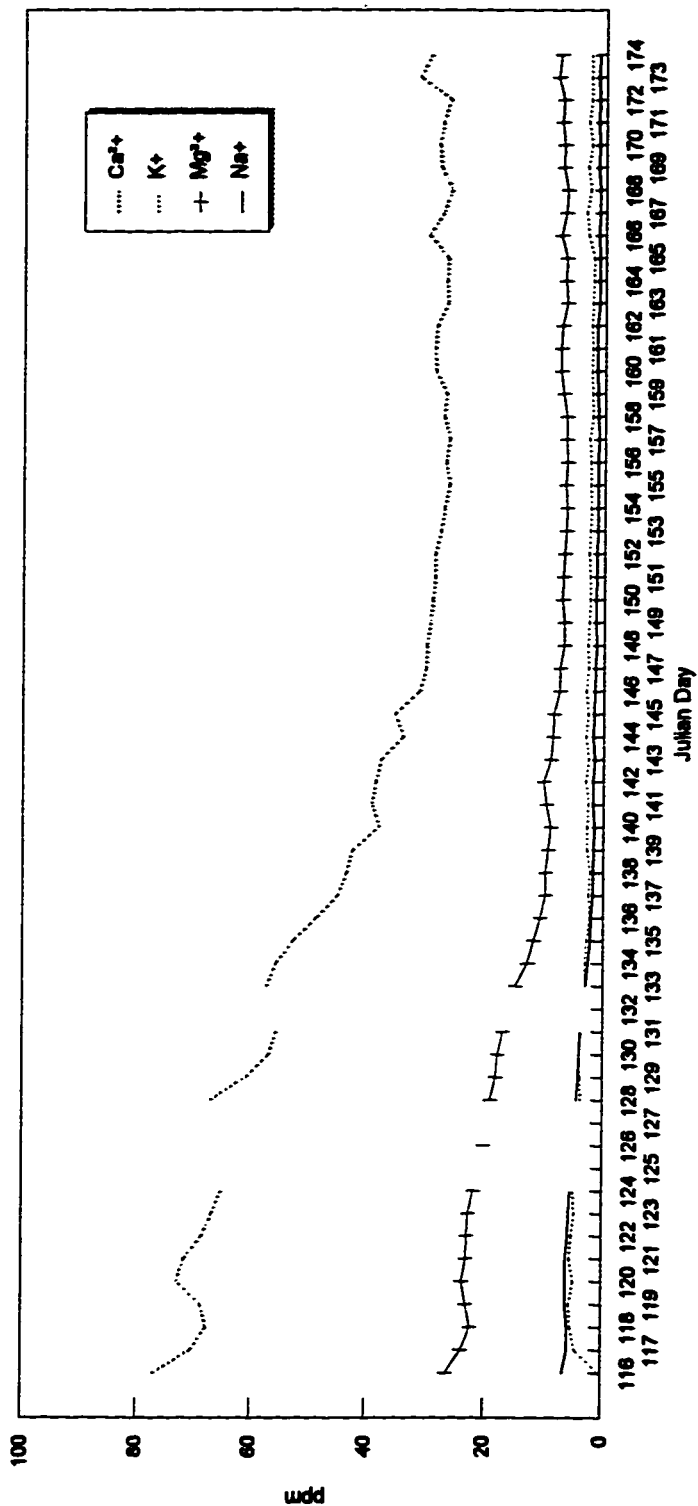
**Table 4.2a**  
*Ionic Concentrations and Proportions*  
*Determined from Meltwater from Slims River, April-June 1993*

Descriptor	Ionic Concentration (ppm)				% x of (Na <sup>+</sup> + K <sup>+</sup> + Ca <sup>2+</sup> + Mg <sup>2+</sup> ) per Sample			
	Ca <sup>2+</sup>	Mg <sup>2+</sup>	K <sup>+</sup>	Na <sup>+</sup>	Ca <sup>2+</sup>	Mg <sup>2+</sup>	K <sup>+</sup>	Na <sup>+</sup>
<i>Mean</i>	41.4	11.4	3.1	2.5	71.6	18.9	5.6	3.9
<i>Max</i>	76.9	26.3	5.5	6.6	76.3	23.8	8.7	6.3
<i>Min</i>	26.2	6.4	0.5	1.1	66.4	16.6	0.5	2.9
<i>S.D.</i>	17.3	6.4	1.1	1.8	2.5	1.9	1.3	1
<i>C.V.</i>	42.1	55.7	34.9	72.8	3.5	10.2	23.9	25.8

**Table 4.2b**  
*Ionic Concentrations Determined from Meltwater from the Slims River, 1994*

Julian Date	Concentration (ppm)			
	Ca <sup>2+</sup>	Mg <sup>2+</sup>	K <sup>+</sup>	Na <sup>+</sup>
128	50.1	13.7	6.2	2.74
142	32.6	6.9	5.2	1.41
166	34.7	6.5	4.8	1.48
178	31.9	6.3	4.5	1.54
196	31.9	5.5	5.6	1.09
209	27.3	5.4	3.9	1.08

**Figure 4.31**  
 Behavior of Major Cations, Slims River 1993



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other glacier streams (Table 4.3). The relative proportions of ions follow the same order as other glacier fed streams despite differences in basin size and percentage glacierization. The commonality between all basins, including the current one, is they are underlain by igneous and metamorphic rocks and have  $\text{Ca}^{2+}$  as the most readily mobilized ion (Collins, 1979b).

The small proportional variation is reflected in good correlations between the four major cations as the season progresses (Table 4.4). Lemmens and Roger (1978) suggest that good correlation probably indicates that the cation sources remain the same.

The ranges in percentage composition indicate that the proportions of ions do not remain constant as concentration of meltwaters varies, and precludes the use of conductivity as an index of individual ion concentration (Collins 1979b). Since the relative composition varies there are more factors affecting solute concentration than simple enrichment or dilution by meltwaters in the stream. The changing proportions indicate that along its course meltwater is routed to lithologies with different material composition (Collins, 1979b). In particular, the ranges of concentration of  $\text{Ca}^{2+}$  and  $\text{Mg}^{2+}$  indicate differential enrichment at different stages of flow (Collins, 1979b).

Although there is variance in the ranges of percentage composition, all measured cations decrease with increasing discharge but  $\text{Ca}^{2+}$  and  $\text{Mg}^{2+}$  show greater rates of change with time (Figure 4.31). Additionally, all major cations show strong positive correlation with conductivity in 1993 (Table 4.5a, Figure 4.32; Figure 3.26ab, 3.27ab, 3.28ab, 3.29ab Appendix D).<sup>3</sup> In 1993, Ca shows a linear relation with conductivity while the other major cations have 2nd degree polynomial relationships. The relation for conductivity and cations in 1994 is strong positive and linear for all cations but includes only 6 data points which obviously affects the results (Table 4.5b, Figure 4.33; Figure 3.30ab, 3.31ab, 3.32ab, 3.33ab Appendix D).

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<sup>3</sup> However, recall that these are uncorrected conductivity values and serial correlation with temperature must be taken into account in the interpretation of the correlations and relations.

**Table 4.3**

*Summary of Proportional Cationic Composition of Meltwaters from the Slims River and Comparison With Other Glacier Fed Streams*

Glacier or River	Sample Period	No. of Samples	Range of % as % ( $\text{Na}^+ + \text{K}^+ + \text{Ca}^{2+} + \text{Mg}^{2+}$ )				Source
			$\text{Ca}^{2+}$	$\text{Mg}^{2+}$	$\text{K}^+$	$\text{Na}^+$	
<i>Slims River Yukon</i>	April-June 1993	56	66.4-76.3	16.5-23.8	0.5-8.7	2.9-6.3	
<i>Castner Glacier</i>	July-August 1968	11	67.2	17.6	4.2	10.9	Collins (1979b)
<i>Gornera Switzerland</i>	July-August 1974	69	45.0-90.5	5.2-23.3	1.8-22.0	1.3-17.4	Collins (1979b)
<i>Moiry Glacier Switzerland</i>	Summer 1971	15	68.7-90.1	3.7-14.4	1.7-13.3	1.4-6.9	Lorraine and Souchez (1972)

**Table 4.4**

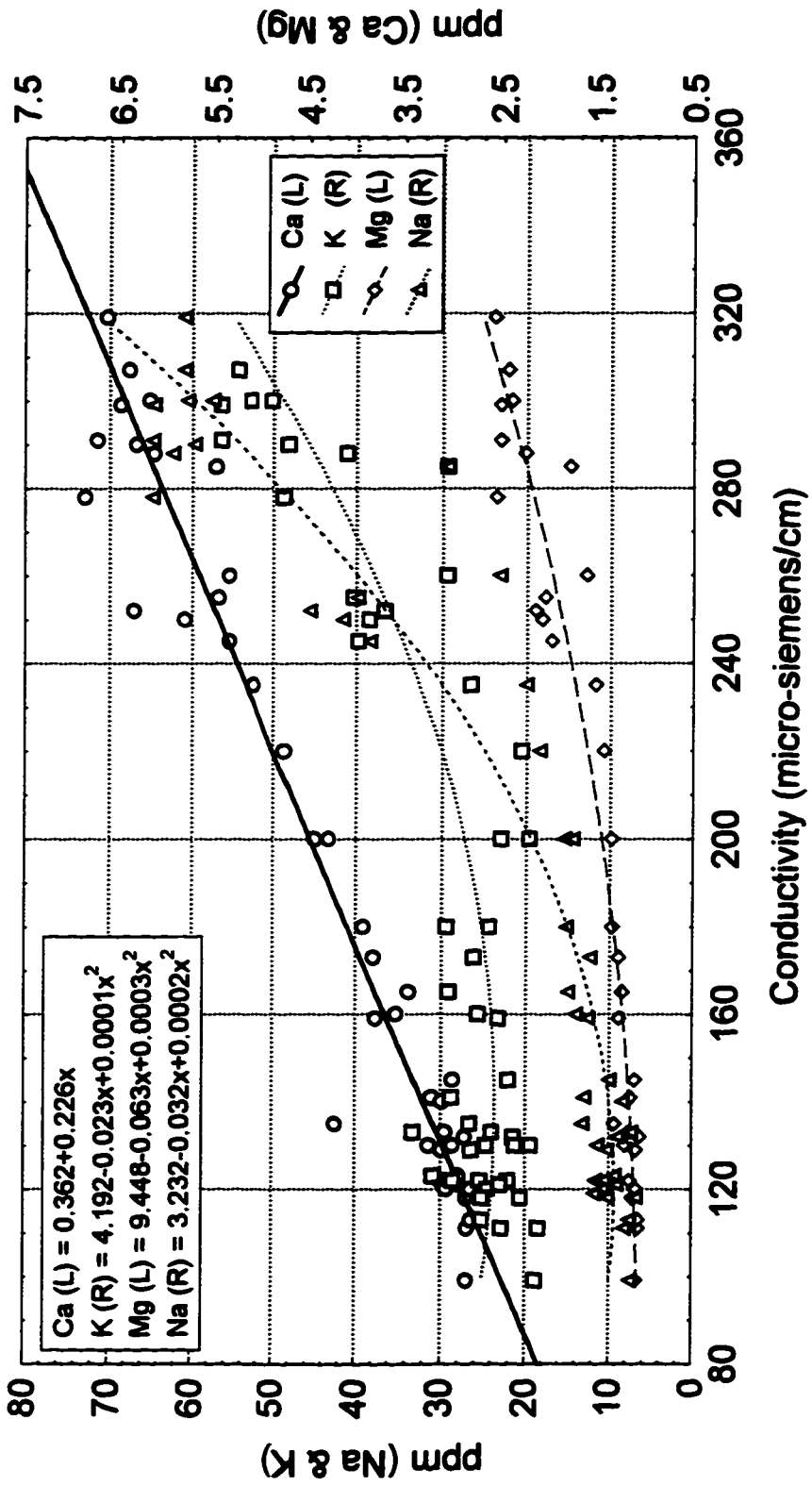
*Correlations Among Major Cations, Slims River, 1993*

Variable	$\text{Ca}^{2+}$	$\text{Mg}^{2+}$	$\text{K}^+$	$\text{Na}^+$
$\text{Ca}^{2+}$	1			
$\text{Mg}^{2+}$	0.98	1		
$\text{K}^+$	0.67	0.71	1	
$\text{Na}^+$	0.95	0.99	0.73	1

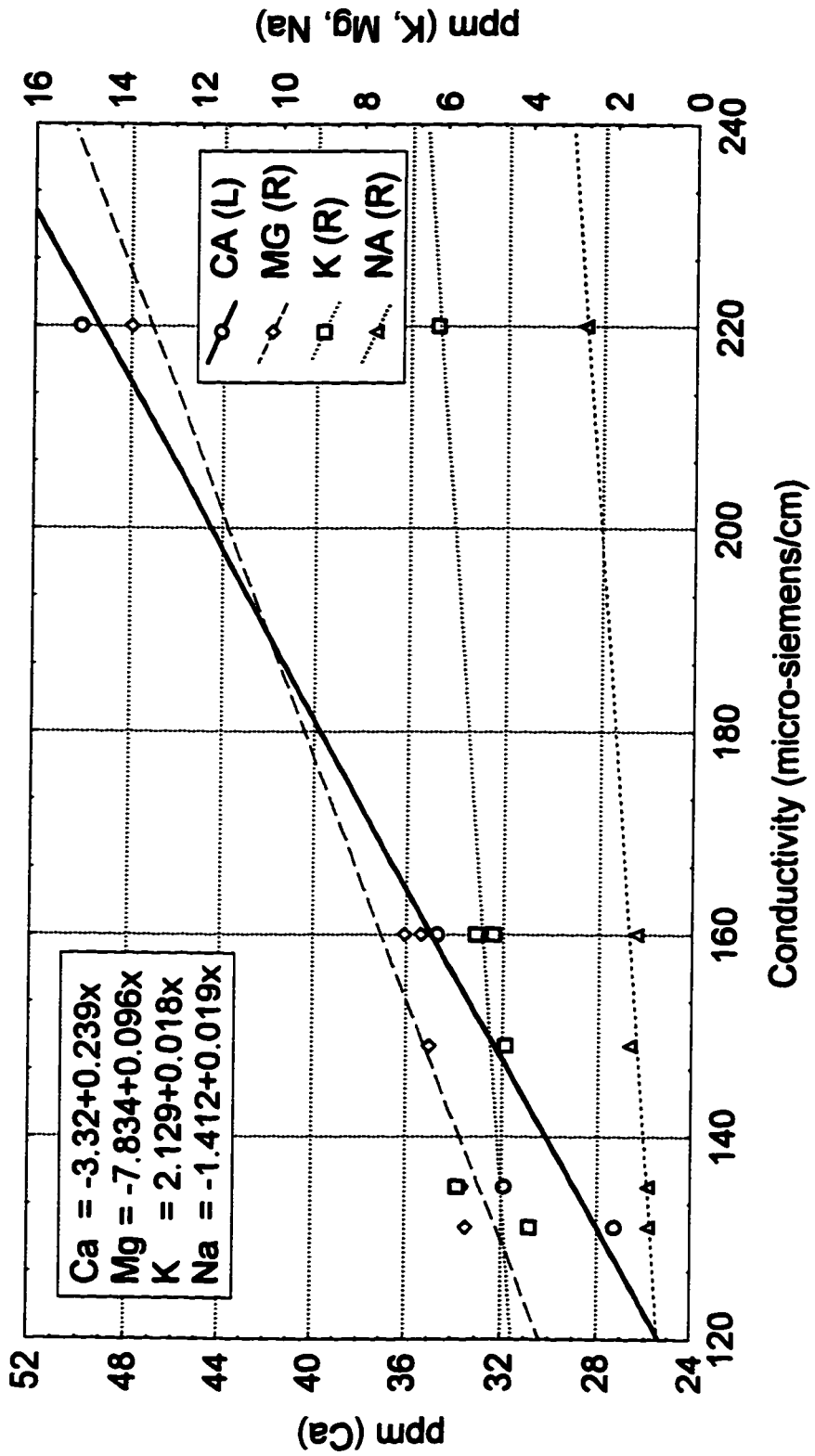
<b>Table 4.5a</b>	
<i>Correlation Matrix for Conductivity and Major Cations, Slims River 1993</i>	
<i>n = 53</i>	
	<b>Conductivity Uncorrected</b>
<b>Ca</b>	0.96
<b>Mg</b>	0.97
<b>K</b>	0.89
<b>Na</b>	0.96

<b>Table 4.5b</b>		
<i>Correlation Matrix for Conductivity and Major Cations, Slims River 1994</i>		
<i>n = 6</i>		
	<b>Conductivity Uncorrected</b>	<b>Conductivity Corrected</b>
<b>Ca</b>	0.95	0.92
<b>Mg</b>	0.96	0.94
<b>K</b>	0.52	0.4
<b>Na</b>	0.97	0.97

**Figure 4.32**  
Major Cations and Conductivity (Uncorrected), Slims River, 1993



**Figure 4.33**  
Major Cations and Conductivity (Uncorrected), Slims River 1994



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From the strong positive relations between the major cations and conductivity the following can be concluded of the dissolved solids regime:

1. This indicates dilution of chemically enriched proglacial sources by chemically dilute glacier meltwater or a flushing of ions from proglacial sources as the monitoring period progressed.
2. The small changes in relative proportions of  $\text{Ca}^{2+}$  and  $\text{Mg}^{2+}$  over the samples indicates that these likely reflect general lithological influences in the basin rather than particular changes in source areas.

However, the more drastic change in  $\text{K}^+$  and  $\text{Na}^+$  may indicate a change in source area.

Enrichment of  $\text{K}^+$  and  $\text{Na}^+$  has been shown to be a product of clay weathering (Lemmens and Roger, 1978). The small absolute amounts of these cations may be related to the small amounts of clay sized particles in suspension or an absence of clay minerals. Although, no mineralogical analysis was done, the delta deposits are thixotropic and therefore have little internal cohesion. Cohesion increases with large proportions of clay minerals (Glynn and Fergusson, 1986) and the presence of divalent electrolytes such as  $\text{Ca}^{2+}$  (Johnson and De Graff, 1988). Consequently, it is plausible, in spite of the lack of mineralogical evidence, that the small amounts of  $\text{K}^+$  and  $\text{Na}^+$  are partially due to lack of clay minerals in the sediments.

## **4.4 Summary**

In April, May and June of both years, discharge is primarily produced from proglacial snowmelt. Discharge fluctuates diurnally in response to diurnal temperature variations in both years. In 1994, peak seasonal discharge takes place after snowmelt has ended, at which time meltwater production, for non-outlying temperatures is amplified because the contributing area of exposed glacier ice is greater as the melt season progresses. Amplified glacier melt in response to non-outlying temperatures is reflected by exponentially increasing discharge amplitude in 1994. However, this exponential increase in amplitude is a consequence of the fact that monitoring does

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not extend past peak summer discharge. With more exposed glacier ice, discharge becomes extremely sensitive to temperature variations with the transition from snowmelt to icemelt. Precipitation was infrequent in both years. However, precipitation events were associated with periods of cloud cover and decreased or eliminated 'normal' diurnal discharge fluctuations.

In 1994 discharge was undercompetent, since, for equal discharges, higher sediment concentrations were produced in 1993. In 1993 smoothed suspended sediment and discharge series were more synchronous. However, in 1994 an outlying event containing the peak seasonal sediment concentration was observed to take place without any significant change in discharge. This indicates the opening of the glacier hydrological system and coincides with maximum sediment concentration variability and magnitude. No such event was observed in 1993 to suggest that the opening of the glacier system was encountered.

Diurnal clockwise hysteresis between suspended sediment concentration and discharge defines the short term suspended sediment regime in both 1993 and 1994. Various assumptions of contemporary sediment transport models were tested on the 1993 and 1994 data sets, however, these assumptions did not hold and thus contemporary reasoning fails to explain the observed frequency or magnitude of hysteresis in the Slims River. However, in the present situation it is proposed that hysteresis is either a supply-exhaustion, hydraulic or equifinal phenomena.

In both years, conductivity decreases in response to dilution by snowmelt and/or glacier icemelt producing a nonlinear inverse relation between conductance and discharge. The relation is best defined in 1993 because only a single nival melt regime was encountered. In 1994 a second peak in conductivity preceded and overlapped the proposed time of sub-glacial discharge establishment which decreased correlation between discharge and conductance. Contrary to other glacier streams, in both years, observed solute concentrations were high due to a combination of proglacial and subglacial routing at different parts of the season. High conductivity during maximum discharge suggests that the sub-glacial system is capable of producing these values.

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The diurnal amplitude of conductivity was greatest and most consistent with glacier melt in 1994.

Clockwise hysteresis dominates the diurnal regime in both years. Hysteresis between conductivity and discharge may be produced as a result of two scenarios:

1. As a result of two two-component meltwater routing systems. First, in the early season conductivity is higher on the rising limb due to the flushing of ion rich waters on the rising limb of tributary hydrographs by dilute daily snow meltwaters. Second, in the later season there is a delay of supraglacial meltwaters in reaching the glacier portal while at the same time flushing out subglacially enriched waters producing higher conductivity on the rising hydrograph limb.
2. Second, if solute concentration is always high in the tributary streams and if the tributary water always reaches the gauging site first, then this will always produce higher solute concentrations on the hydrograph rising limb. Concentrations will be reduced when more dilute and voluminous glacier meltwaters reach the gauging site and dominate the falling hydrograph limb.

The dominant cation in meltwaters is  $\text{Ca}^{2+}$ , followed by  $\text{Mg}^{2+}$ ,  $\text{K}^+$  and  $\text{Na}^+$ , respectively. This is not unusual for basins underlain by metamorphic and igneous rocks. All individual ion concentrations decrease over the season following conductivity trends but are not proportionately constant in this decrease and so preclude the use of conductivity as a surrogate measure of absolute individual ion activity. However, there are strong relations between conductivity and individual cation concentrations which suggest that in the present case, conductivity may be used as a relative estimate of individual ion activity.

## Chapter 5

*By the water god  
protected, a great river  
even in winter.*

Seishi Yamaguchi<sup>1</sup>

### Synthesis

#### **5.1 Context**

This chapter integrates the results of Chapter 4 within the hypotheses of Chapter 1. Hypotheses are not repeated, rather each section that deals with a particular hypotheses is headed by the pertinent reference. Following the synthesis are recommendations regarding the aspects of the flow regime of the Slims River that are most likely to affect Kluane Lake sedimentation on an annual basis. Finally, recommendations for future research are proposed.

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<sup>1</sup> Source: Takashi, K., and Marks, A.H. *The Essence of Modern Haiku: 300 Poems by Seishi Yamaguchi*. Mangajin, Inc: Atlanta USA. p.277.

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## **5.2 $H_1$ : Discharge**

There is a consistent 10-12 hour phase shift between temperature and discharge in both seasons. This is primarily due to tributary basin concentration times, and, in the later season, the combined delay time of a greater exposed glacier surface with increased travel time to the gauging site. There is no significant correlation between hourly discharge temperature because discharge increases over the season whereas the range of temperatures remains relatively stable. Therefore, the weak explanation of discharge variance is due to a progressive change in the contributing area of the glacier which, in turn, causes an amplified melt response for similar daily temperature variations.

In both melt seasons, mean discharge responds to temperature trends over a number of days. This is more evident in 1994 during the period when glacier melt is appropriating a greater percentage of runoff, at which point mean daily temperature explains less of the variance of mean daily discharge. The correlation between mean daily temperature and discharge is positive for 1993 and the same period in 1994. Positive correlation for the early melt season has two explanations: First, the position of the meteorological monitoring stations reflect lower altitude temperatures during the snowmelt portion of runoff. Second, snowmelt and discharge are more in phase because the amplified response of glacier melt to these temperature ranges has not yet begun. When the whole of the 1994 season is taken into account, the correlation between mean discharge and temperature decreases. This is because as the melt season progresses and glacier meltwaters influence flows, there are much higher discharges produced for similar temperatures when compared to the early 1994 melt season alone. Additionally, in the later 1994 season there is a somewhat delayed response of mean daily discharge to mean daily temperature variations. The differences in hydrometeorological and sedimentological factors before and after day 175 in 1994 suggest that before this date river flow is dominated by snowmelt and afterwards by glacier melt.

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Finally, in 1994, daily temperature explains more of the variance in daily discharge amplitude than variance in mean daily discharge. Consequently, discharge amplitude exhibits a stronger response to daily temperature variations. That circumstance is not surprising considering that there is more exposed glacier ice later in the season which because of its lower albedo responds faster to daily temperature than would similar snow-covered ice. The enhanced response of diurnal discharge amplitude is reflected in an event in 1994 between days 201 and 205 when there was a major decrease in discharge amplitude and mean discharge as a response to a temperature trough, which, in itself, was not an outlying event.

In summary, discharge is primarily produced from proglacial snowmelt in 1993 and for the same time interval in 1994. Discharge fluctuates diurnally over both seasons in response to diurnal temperature variations. In 1994, peak seasonal discharge takes place after snowmelt has ended at which time meltwater production, for non-outlying temperatures is amplified because of a larger contributing area composed mainly of exposed glacier ice. Amplified glacier melt in response to non-outlying daily temperatures is reflected by increasing discharge amplitude during late 1994, at which time discharge becomes extremely sensitive to temperature variations.

### **5.3 H<sub>2</sub>: Discharge and Sediment Transport**

The increasing sediment loads accompanying snowmelt runoff caused an at-a-station hydraulic response in the gauging reach. In order to maintain sediment transport continuity, channel accretion and scouring took place early in the season of 1994. Scour and fill were only encountered in the south channel whereas the north channel maintained a relatively constant cross-section throughout the 1994 season.

The considerable depth of the south channel before the fill sequence suggests that after the previous year's peak summer flow, the lower flows accompanying fall and winter are able to

maintain sediment transport continuity while at the same time requiring the use of both channels for water discharge. If this is the case then there must be either a seasonal decrease in sediment loads or commensurate decrease in grain size which would allow lower flows to proceed without accretion. In 1994, from snowmelt to peak summer discharge, there is decreasing sediment size while at the same time increasing sediment loads. This, combined with the proposal that perimeter sediments are deficient in fines, suggests that the glacial contribution of suspended sediment must increase over the season in order to maintain the increasing loads and finer material. It is clear that for the median grain sizes observed in the suspended sediment samples, the flows in 1994 after scour and fill are undercompetent. This conclusion arises because, for the same discharges in 1993, sediment concentrations were almost two times greater. However, there are no data on the at-a-station hydraulic geometry for the 1993 season.

The major influences on particle size in the Slims River are hydraulic sorting in the early season and changes in source output during glacier discharge. Fine silt is compositionally dominant in suspended sediment samples, and increases proportionally as the season progresses. Over the 1994 season suspended sediment becomes finer. However, bank material, a product of deposition, is low in fine silt and clay and thus the fines must be derived externally from the glacier. Furthermore, fining trends with increasing concentrations of fine material suggest that channels are well established after peak flows. This conclusion arises because, with increasing discharge there is no hydraulic reason to expect finer distributions. Sediment transport continuity in the late season may be due to the high proportion of washload which is transportable under modest flow conditions without channel modification. These observations are important to the frequency of occurrence of hysteresis and suggest that in the later season glacier sediment discharge may undergo little modification by deposition or re-mobilization in the valley train.

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Suspended sediment concentration follows mean discharge fluctuations in 1993 and the early season of 1994. Synchronous variations and similar correlation between sediment and discharge in the early season of 1994 and the whole period of 1993 suggest that concentration is supplied by sources internal to the valley train during the nivally dominated meltwater portion of the season. Generally, suspended sediment concentration follows discharge variations for the majority of 1994 with the exception of the peak between day 163 to 180. As an anomaly where peak seasonal sediment concentrations are encountered with no substantial change in discharge, it is typical of the glacier hydrological system opening. No event occurred in 1993 that would suggest that the glacier hydrological system had opened. However, greater sediment concentrations were produced in 1993 by flows that were well below the 1994 values discharge and, therefore, indicate that the stream is undercompetent in 1994. No definitive explanation can be proposed at present for these seasonal differences.

Hysteresis is evident in every diurnal cycle as either clockwise, involuted or counter-clockwise hysteresis and defines the short term sediment transport regime of the Slims River. Clockwise hysteresis is most frequent in the later season of 1994 when, after the inferred opening of the glacier hydrological system, it occurs 91% of the time. In the early season of 1994 and all of 1993 involuted hysteresis was most frequent. This has not previously been documented or explained for proglacial or glacial streams.

Contemporary reasoning cites exhaustion of supply on the rising limb of the hydrograph as the reason for the occurrence of hysteresis. In the Slims River, *prima facie*, there is no strong correlation between the occurrence of hysteresis and hydrograph asymmetry. The occurrence of hysteresis does not lower concentrations on successive diurnal cycles. Contrary to other research, there is no recovery period required, that is, the time interval between events when processes of weathering work to produce material for entrainment by the next flood.

Notwithstanding, there is also no definitive correlation between maximum sediment concentrations and the rise above/below previous day's stage. Furthermore, sand, silt, and clay fractions vary together during hysteretic cycles which suggests that there is no selective exhaustion of sediment over a given hydrograph. There are no definitive answers but rather the observations prompt questions such as:

If hysteresis indicates supply exhaustion then how can it take place daily?; and, Is there a sediment supply mechanism that can produce enough sediment to maintain hysteresis on diurnal time scales?

Either processes in the glacial and/or proglacial environment must act on short time scales to produce the sediment required to maintain diurnal hysteresis, or, hysteresis is not a supply exhaustion effect in this particular instance. Processes such as bank collapse during low flow will expose easily transportable material that is exhausted during rising stage. Because suspended sediment is mainly washload, and increases proportionally as the season progresses, then the later season hysteresis may be produced subglacially and translated to the gauging site without modification.

Alternatively, since all fractions vary together on rising or falling limbs, when hysteresis takes place, the effect may be hydraulically controlled by, for example, higher velocities on the rising limb because of immature bedform development. In this case hysteresis may not indicate supply exhaustion but another process altogether. Consequently, it is not unreasonable at this time to put forth the hypothesis that hysteresis is an equifinal product, that is, a product of various processes acting in conjunction or separately at different times of the melt season to produce the same effect. The causes of the phenomenon remain unresolved and further study is warranted,

### *5.4 H<sub>3</sub>: Bulk Hydrochemistry and Discharge*

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including investigating erosion in order to estimate bank collapse and timing, as well as a detailed study of changes in hydraulic geometry and stream power over individual diurnal cycles.

In summary, discharges are comparable for the same periods of 1993 and 1994 but this is not the case for suspended sediment concentrations. For equal discharges there are greater suspended sediment concentrations produced in 1993, and the suspended sediment regime closely follows that of discharge. In 1994 there is an anomalous event between days 168-182 when peak seasonal sediment concentrations are produced with no appreciable change in discharge. It is proposed that this indicates the opening of the glacier hydrological system. Since no similar event takes place in 1993 it is suggested that the opening of the glacier hydrological system was not captured in that year's data set. After the opening of the glacier hydrological system in 1994 clockwise hysteresis occurs diurnally. Hysteresis defines the short term suspended sediment regime in both 1993 and 1994 however current reasoning fails to explain its frequency or magnitude. Because of the frequency of hysteresis there is no definitive explanation for the phenomenon, which itself, suggests that hysteresis is either an hydraulic, supply-exhaustion or equifinal phenomena.

### **5.4 H<sub>3</sub>: Bulk Hydrochemistry and Discharge**

Over both melt seasons conductivity decreases in response to dilution by snowmelt and/or glacier icemelt. Therefore, there is a non-linear inverse relation between conductance and discharge in both years. Correlation between conductivity and discharge is strongest in 1993 because of a shorter monitoring season combined with a single nival melt regime. In 1994 there was a second peak in conductivity that corresponded to the opening of the sub-glacial system. This second conductivity peak in 1994 increased scatter and decreased correlation between discharge and conductance.

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Generally, decreasing air temperatures and discharge were followed by increases in conductivity in both years since at these times there was decreased surface melt and hence less dilution was taking place.

Minimum conductance values are consistent in both years but are higher than those recorded in other glacierized basins. High conductivities are expected in the early season when snowmelt percolates through alluvial and colluvial materials and liberates accumulated salts. Since the contribution of tributaries decreases substantially with the end of snowmelt and glacier melt dominates discharge, it is unexpected to find minimum values that are far in excess of atmospherically derived solute levels. High minimum conductance values may be due to the fact that the Kaskawulsh Glacier is relatively large when compared to glacier streams that are cited in contemporary literature. As well, the source of the Slims River is subglacial rather than supraglacial and, because of the glacier's size, this could mean longer contact times between waters in the subglacial system and the fine products of glacial abrasion produced there.

High solute concentrations are observed in both years, but in the 1994 record there is a second peak (equal in magnitude to the 1993 or early 1994 values) in conductivity that does not correspond to a decrease in the discharge rate. This peak precedes the previously mentioned anomaly in the sediment record and therefore may indicate a release of solute rich waters which would be consistent with the initial opening of the glacier system. Such an event was not observed in 1993, but the dilution regime in 1993 was similar to the early season of 1994 which suggests that ion enrichment is proglacially derived.

Diurnal variation in conductivity was better defined during the glacier melt portion of 1994, such that, maximum and minimum conductivity values are relatively constant due to diurnal dilution. In both years hysteresis is produced because conductivity is greater on the rising limb of the

hydrograph and is therefore out of phase with discharge. There are at least two plausible explanations for this:

1. As the result of two two-component meltwater routing systems acting at different parts of the melt season. First, in the early season conductivity is higher on the rising limb due to the flushing of ion rich waters on the rising limb of tributary hydrographs by dilute daily snow meltwaters. Second, in the later season the two components change to the delay of fresh supraglacial meltwaters in reaching the glacier portal while at the same time flushing out longer residence time subglacially enriched waters.
2. Tributary waters are solute rich throughout the season and reach the gauging site first due to faster travel times causing higher concentrations on the rising hydrograph limb. Second, dilute glacial meltwaters reach the tributary site last leaving lower concentrations on the falling hydrograph limb.

The dominant cation in meltwaters is  $\text{Ca}^{2+}$ , followed by  $\text{Mg}^{2+}$ ,  $\text{K}^{+}$  and  $\text{Na}^{+}$ , respectively. All decrease over the season following conductivity trends but are not proportionately constant in this decrease and so preclude the use of conductivity as a surrogate measure of individual ion activity. Major cations show less proportional variation than other glacierized basins which leads to good correlation among the cations which suggests that source areas are similar. However, the similarity likely reflects general basin lithology, rather than specific source areas since the percent cationic composition is not constant and suggests that meltwaters likely encounter different lithologies at different times of the season. This is reasonable, since meltwater production will reach higher altitudes as the season progresses and this increases the probability of routed waters encountering different lithologies.

## **5.5 Factors Affecting Kluane Lake Sedimentation**

The insignificant decrease in sediment size as distance from the delta increases found by Bryan (1974b) may in part be due to the high proportion of silt in suspension as well as the grain size distribution of this material, in that, it is almost well sorted. Moreover there is little clay entering the lake. Additionally, the high concentrations are capable of producing density flows that could

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remain stable (e.g., not undergo substantial mixing with lake waters) for some distance over the lake bottom. Stream temperature fluctuates substantially on a diurnal basis (Figure 5.1) but because of the high sediment concentrations this is unlikely, at any time, to produce overflow or interflow conditions. The braided streams running through the sandur are quite shallow and, therefore, respond quickly to air temperature changes in their 22 km journey to the gauging site.

## **5.6 Recommendations**

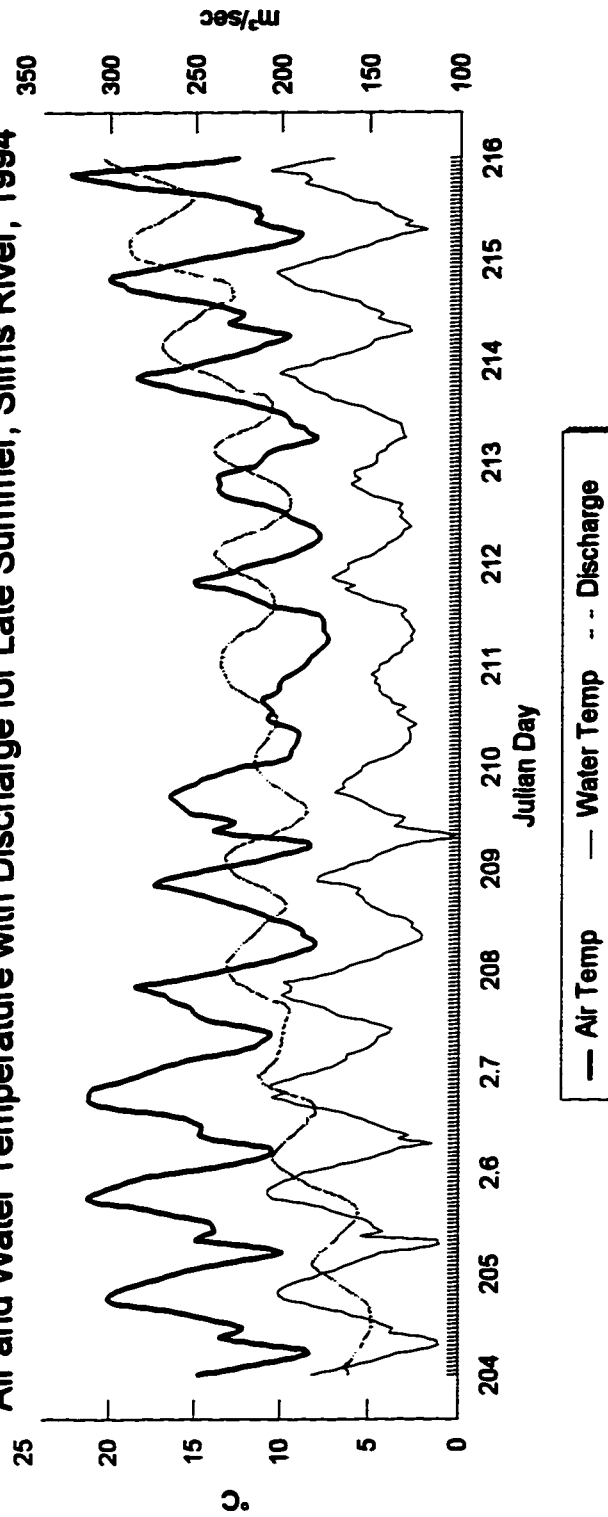
There is no end to learning as a process of inquiry (Wellar et al., 1994) and, as such, there is no end to methodologically sound inquiry as the process of 1) adding to knowledge and 2) adding to the ways and means of continuing to add to knowledge through better research methods, techniques and operations (Wellar and Harris, 1992). Following in this tradition, utility is gained in reflection by iteration of the lessons learned through a particular research process.

### **5.6.1 Discharge**

The major problem with measurement of discharge was the imprecision in the discharge and stage relation. If future work is to detail at-a-station hydraulic geometry, including bedform development and especially variation of velocity with stage over diurnal cycles, precision will be a primary objective. This would be difficult with electrostatic current meters considering that as discharge increased there were considerable problems in getting accurate velocity measurements for similar stages because of turbulence. The strong turbulence in the reach would be conducive to time integrated measurements using, for example, a tracer chemical. Detailed hydraulic geometry will be necessary if hydraulic causes of hysteresis are to be explained.

It would also be pertinent to re-survey the major tributaries to the Slims River throughout the season in order to assess the magnitude of their progressive change in contribution. Equilibrium

**Figure 5.1**  
**Air and Water Temperature with Discharge for Late Summer, Slims River, 1994**



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line altitudes from remotely sensed data would also help determine the transition from nival flow to glacially dominated meltwaters.

### 5.6.2 Sediment

Hourly suspended sediment data is ample for most purposes but tends to be rather labour intensive in processing which may prevent the acquisition of better resolution in other variables given finite human resources. Physical sampling would be extremely inefficient if a more detailed enquiry was to be undertaken. However, turbidity could be employed in order to get better resolution and more data from other parts of the river. Certainly, any in-depth study of the short-term suspended sediment transport regime should measure transport and discharge at numerous points along the stream from the glacier terminus to the gauging site. Automated turbidity stations would facilitate acquiring this data in a ready-to-use form. However, as shown in Chapter 3, the 'home made' turbidity meter performed rather poorly in the explanation of variance in simultaneous suspended sediment concentration. That being said, the exploration of the time varying flux of suspended sediment at various points along the stream is more important for explanation of the short term sediment transport, and, as indicated in Chapter 3, the turbidity meters are inexpensive, and in the respect of time varying flux, do perform adequately. Consequently a trade-off may be made in order to maximize utility of inquiry. In fact, a detailed future study may be impossible, from a practical standpoint, if turbidity is not employed.

One or two representative stream reaches should be chosen for a detailed bank stability study. Using erosion pins the amount of bank collapse could be estimated and its contribution as an internal supply mechanism ascertained. Furthermore, in this reach some investigation of diurnal deposition and remobilization would be useful.

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Finally, to determine the contribution of external sediment supplies, a survey of the time varying contribution of suspended sediment from tributaries and monitoring at the glacier terminus will facilitate understanding, explanation, and prediction.

### 5.6.3 Dissolved Solids

Considering that the conductivities observed in the Slims River are higher than those of other glacier fed streams there are obvious questions as to their source. As with suspended sediment, a survey of the time varying contributions of tributary streams and glacier output should be undertaken. Results could be effectively incorporated into a mixing model to aid in explanation and prediction of the downstream variations in bulk hydrochemistry. If turbidity is monitored at these stations then it is quite easy to incorporate stream temperature and conductivity, as since Chapter 3 shows, the 'home made' conductivity meters perform adequately for measurement of time-varying flux.

## 5.7 Summary

April, May and June of 1993 and for the same months plus July in 1994 a hydrometeorological study of the Slims River at its input to Kluane Lake was undertaken.

In 1994, peak seasonal discharge takes place after snowmelt has ended, at which time, meltwater production for non-outlying temperatures is amplified because of an increased contributing area composed mainly of exposed glacier ice. The increased response is also reflected by exponentially increasing diurnal discharge amplitude in the 1994 season. Peak discharge was not encountered in the 1993 season. However, the 1993 season had characteristics similar to the same period in 1994 and suggests that in both years April, May and June are predominantly dominated by snowmelt. Discharges in 1994 were under-competent, since, for equal discharges, 1993 produced higher sediment concentrations.

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Diurnal clockwise hysteresis between suspended sediment and discharge defines the short term suspended sediment regime in both years. However, current assumptions regarding the causes of hysteresis in glacial and fluvial environments fail to explain the magnitude and occurrence in the present situation. Therefore it is suggested that hysteresis in this glacial-fluvial stream is either a supply-exhaustion, hydraulic or equifinal phenomena.

In both years conductivity decreases in response to dilution by snowmelt and/or glacier icemelt producing a non-linear inverse seasonal relation between conductance and discharge. Diurnal conductivity amplitude was greatest and most consistent with the glacier melt portion of runoff in 1994. Clockwise hysteresis between discharge and conductivity is produced daily in both years due to either a two two-component meltwater routing system or, to consistently concentrated tributary discharges reaching the gauging site before dilute glacial meltwaters. The dominant cations in Slims River meltwaters are  $\text{Ca}^{2+}$ , followed by  $\text{Mg}^{2+}$ ,  $\text{K}^+$  and  $\text{Na}^+$ , respectively. All decrease over the season following conductivity trends but are not proportionally constant in this decrease. However, the strong positive relation between conductance and individual ion concentrations do not completely preclude the former as a proxy for the latter.

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## **Appendix A**

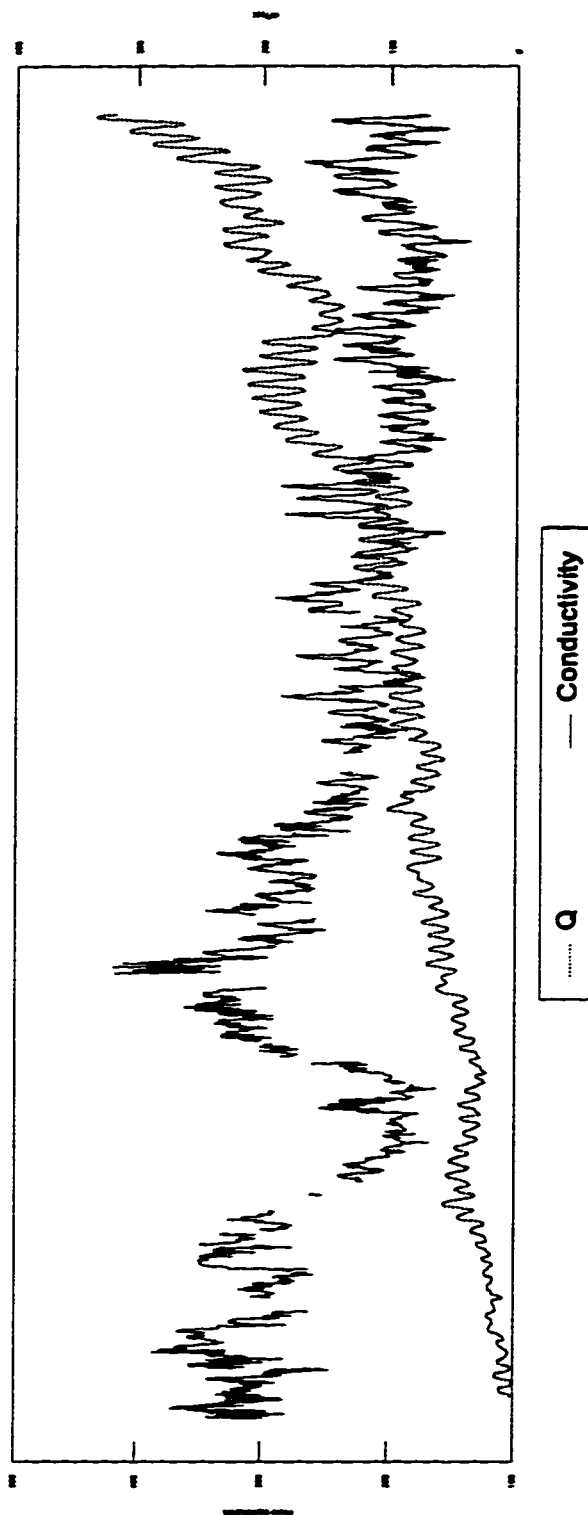
### **Campbell CR-10 Program**

# Campbell CR-10 Program For Continuous Monitoring Equipment

<p>* 1 Table 1 Programs 01: 600 Sec. Execution Interval</p> <p>01: P5 AC Half Bridge 01: 1 Rep 02: 15 2500 mV fast Range 03: 5 IN Chan 04: 3 Excite all reps w/EXchan</p> <p>05: 250 mV Excitation 06: 1 Loc : 07: 1 Mult 08: 0.0000 Offset</p> <p>02: P59 BR Transform Rf[X/(1-X)] 01: 1 Rep 02: 1 Loc : 03: 10000 Multiplier (Rf)</p> <p>03: P40 Z=LN(X) 01: 1 X Loc 02: 1 Z Loc :</p> <p>04: P37 Z=X*F 01: 1 X Loc 02: -.98017 F 03: 1 Z Loc :</p> <p>05: P34 Z=X+F 01: 1 X Loc 02: 12.753 F 03: 1 Z Loc :</p> <p>06: P41 Z=EXP(X) 01: 1 X Loc 02: 1 Z Loc :</p> <p>07: P2 Volt (DIFF) 01: 2 Reps 02: 25 2500 mV 60 Hz rejection Range 03: 1 IN Chan 04: 2 Loc : 05: .01 Mult 06: 0.0000 Offset</p>	<p>08: P37 Z=X*F 01: 3 X Loc 02: .92351 F 03: 3 Z Loc : Page 2 Table 1</p> <p>09: P38 Z=X/Y 01: 2 X Loc 02: 3 Y Loc 03: 6 Z Loc :</p> <p>10: P40 Z=LN(X) 01: 6 X Loc 02: 6 Z Loc :</p> <p>11: P37 Z=X*F 01: 6 X Loc 02: -66.667 F 03: 6 Z Loc :</p> <p>12: P17 Module Temperature 01: 4 Loc :</p> <p>13: P14 Thermocouple Temp (DIFF) 01: 1 Rep 02: 1 2.5 mV slow Range 03: 5 IN Chan 04: 1 Type T (Copper-Constantan) 05: 4 Ref Temp Loc 06: 5 Loc : 07: 1 Mult 08: 0 Offset</p> <p>14: P10 Battery Voltage 01: 7 Loc :</p> <p>15: P86 Do 01: 10 Set high Flag 0 (output)</p> <p>16: P77 Real Time 01: 0120 Day,Hour-Minute</p> <p>17: P78 Resolution 01: 1 High Resolution</p>	<p>18: P70 Sample 01: 7 Reps 02: 1 Loc</p> <p>19: P End Table 1</p> <p>* 2 Table 2 Programs 01: 0.0000 Sec. Execution Interval</p> <p>01: P End Table 2</p> <p>* 3 Table 3 Subroutines</p> <p>01: P End Table 3 Page 3 Mode A</p> <p>* A Mode 10 Memory Allocation 01: 28 Input Locations 02: 64 Intermediate Locations 03: 0.0000 Final Storage Area 2</p> <p>* C Mode 12 Security 01: 0 LOCK 1 02: 0 LOCK 2 03: 0000 LOCK 3 Page 4 Input Location Assignments (with comments):</p> <p>Key: T=Table Number E=Entry Number L=Location Number</p> <p>T: E: L: 1: 1: 1: Loc : 1: 2: 1: Loc : 1: 3: 1: Z Loc : 1: 4: 1: Z Loc : 1: 5: 1: Z Loc : 1: 6: 1: Z Loc : 1: 7: 2: Loc : 1: 8: 3: Z Loc : 1: 12: 4: Loc : 1: 13: 5: Loc : 1: 9: 6: Z Loc : 1: 10: 6: Z Loc : 1: 11: 6: Z Loc : 1: 14: 7: Loc :</p>
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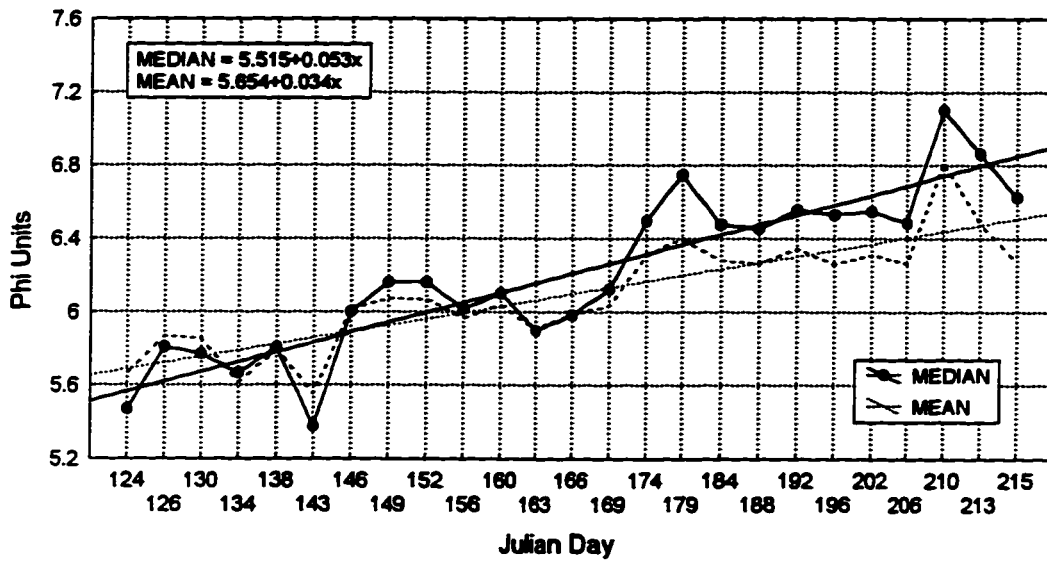
## **Appendix B**

**Figure 4.30**  
Diurnal Discharge and Conductivity, Slims River, 1994



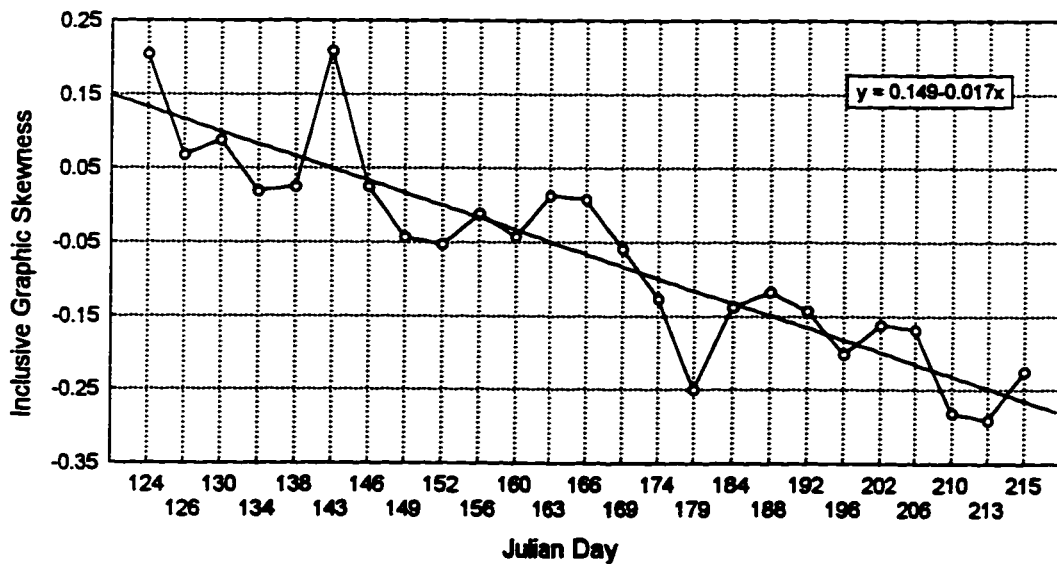
**Figure 4.31**

Mean and Median Particle Size Changes for <0.063 mm over Season



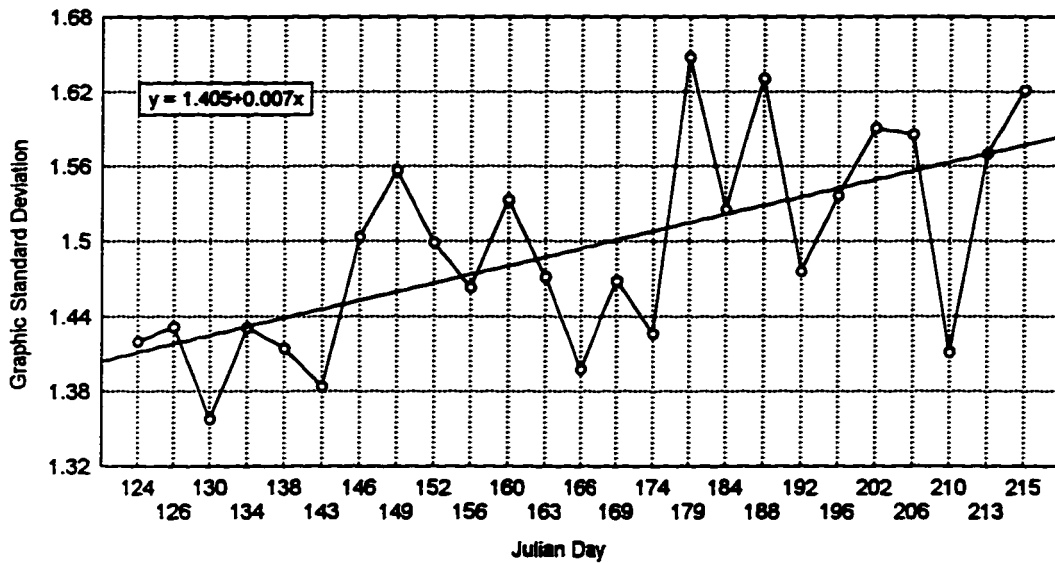
**Figure 4.32**

Skewness in Particle Size Distribution for Selected Samples, Slims River 1994



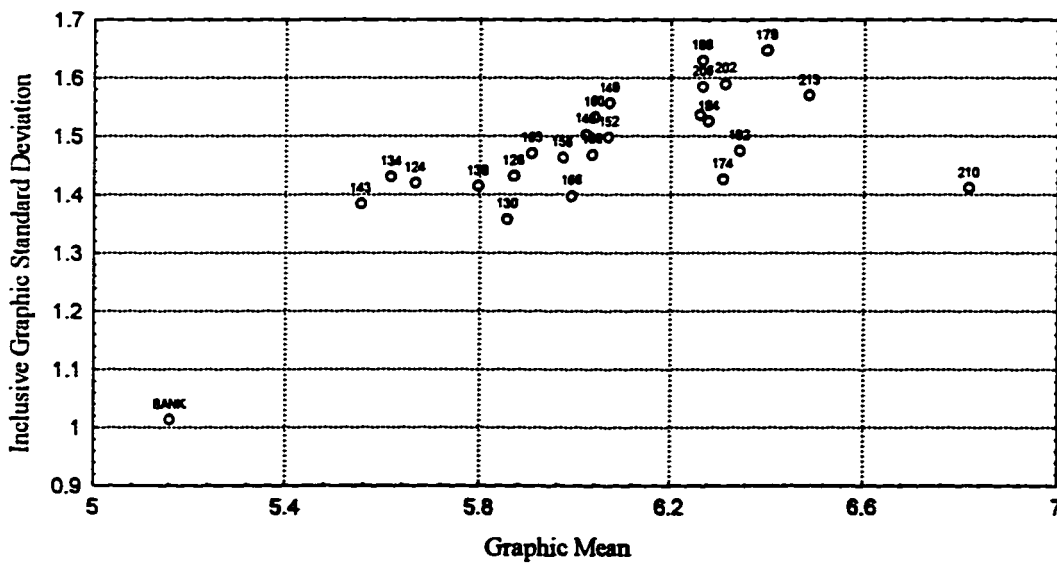
**Figure 4.33**

Sorting in Particle Size Distribution for Selected Samples, Slims River 1994

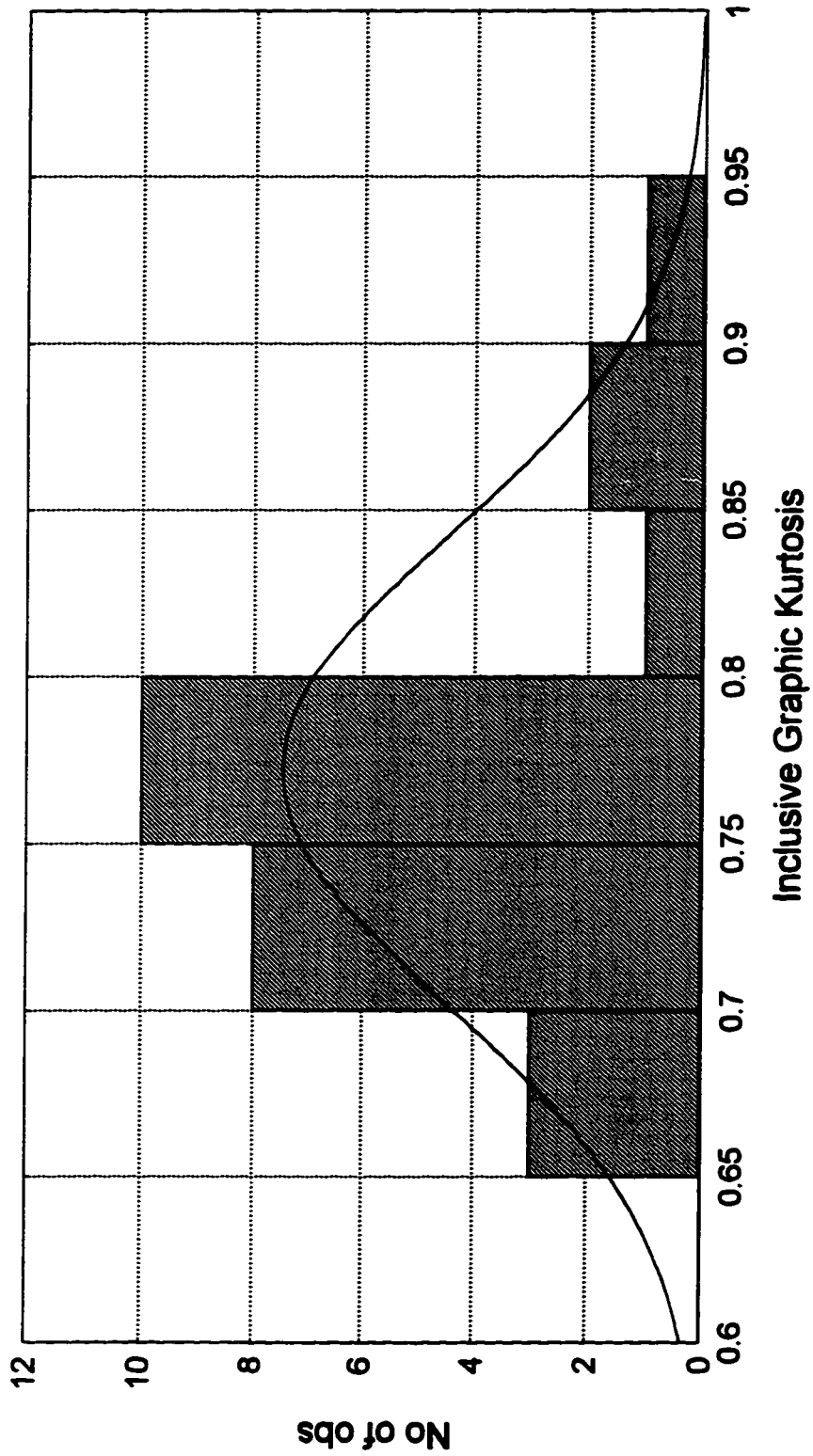


**Figure 4.34**

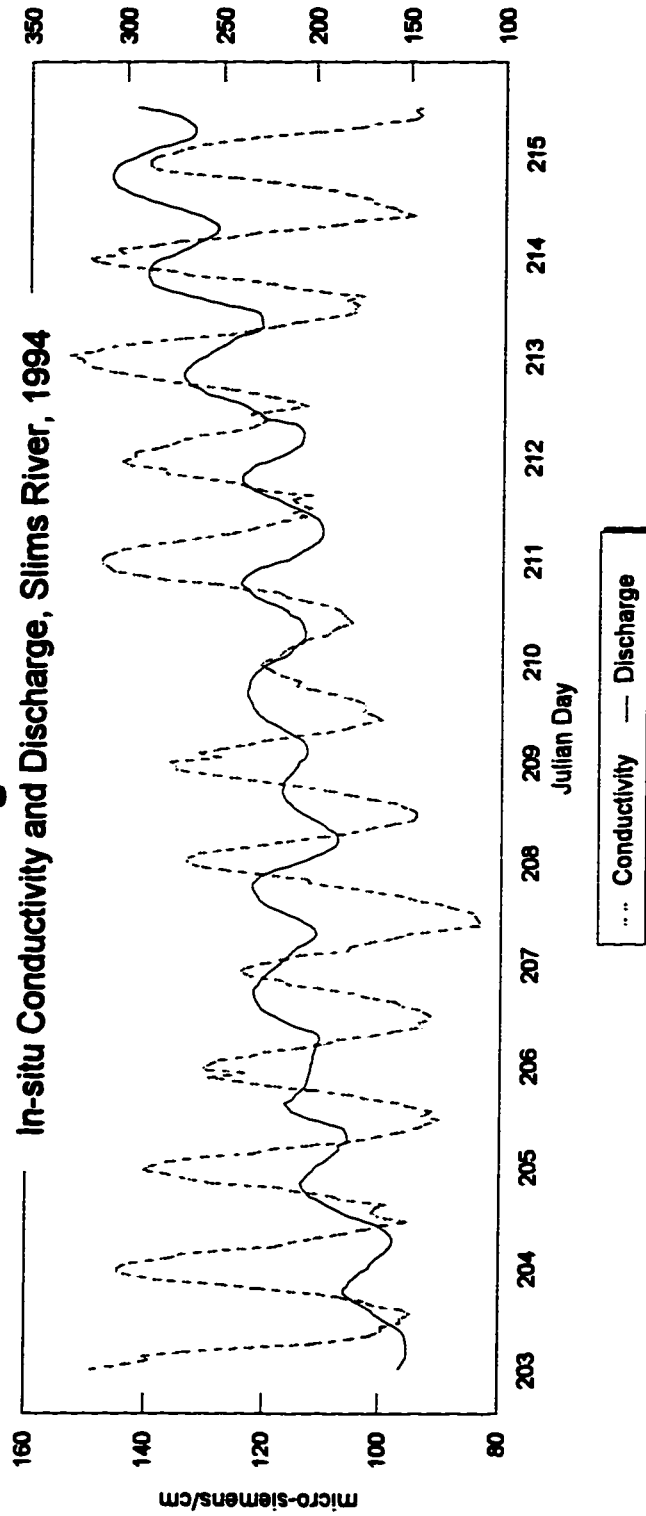
Graphic Mean and Inclusive Graphic Standard Deviation, Slims River 1994



**Figure 4.35**  
Distribution of Kurtosis Values for Selected Sediment Samples, Slims River 1994



**Figure 4.36**  
**In-situ Conductivity and Discharge, Slims River, 1994**



## **Appendix C**

**Table 3.10: Conductivity Reference Temperature Calibration Constant (a)**

t1	Con1 @ t1	t2	Con2 @ t2	L12-L11	L11(t2-R)	L12(t1-R)	(L12-L11)/((L11(t2-25)-(L12(t1-25)))	a in %/°C
8	142	21	198	56	-568	-3366	0.0200	2.00
10	160	23.1	235	75	-304	-3525	0.0233	2.33
14	220	23	265	45	-440	-2915	0.0182	1.82
14	212	23	263	51	-424	-2893	0.0207	2.07
14	212	22.8	266	54	-466.4	-2926	0.0220	2.20
14	201	22.2	251	50	-562.8	-2761	0.0227	2.27
14	201	22.1	250	49	-582.9	-2750	0.0226	2.26
14	199	22.1	248	49	-577.1	-2728	0.0228	2.28
14	195	22.1	240	45	-565.5	-2640	0.0217	2.17
14	190	22.1	236	46	-551	-2596	0.0225	2.25
14	185	22.1	228	43	-536.5	-2508	0.0218	2.18
14	180	22.2	219	39	-504	-2409	0.0205	2.05
14	184	22.2	227	43	-515.2	-2497	0.0217	2.17
8.7	140	22.5	192	52	-350	-3129.6	0.0187	1.87
16	208	30.5	297	89	1144	-2673	0.0233	2.33
14	206	31	301	95	1236	-3311	0.0209	2.09
14	202	29	287	85	808	-3157	0.0214	2.14
14.5	199	27.7	272	73	537.3	-2856	0.0215	2.15
14.8	192	27.5	264	72	480	-2692.8	0.0227	2.27
4	100	12.8	155	55	-1220	-3255	0.0270	2.70
							<b>Average value for a</b>	<b>2.18</b>

**Table 4.10**  
***Bulk Properties of Sand, Silt and Clay***  
***for Selected Samples***

<b>Variable</b>	<b>Mean</b>	<b>Standard Deviation</b>
<i>Percentages</i>		
Sand	15.5	6.8
Silt	80	5.9
Clay	4.6	4.6
<i>Concentration</i>		
Sand	336	296
Silt	1763	887
Clay	102	57

**Table 4.11**  
***Mean Grainsize Distribution for Season***

<b>Size Class (<math>\mu\text{m}</math>)</b>	<b>Mean %</b>	<b>Standard Deviation</b>
350-74	13.6	6.6
74-63	1.8	1.4
63-44	6.5	3
44-31	7.2	2.4
31-16	16.7	3.4
16-8	18.4	2.2
8-4	21.3	4.9
4-2	10.1	3.1
2-1	3.5	1.2
1-0	1.1	0.4

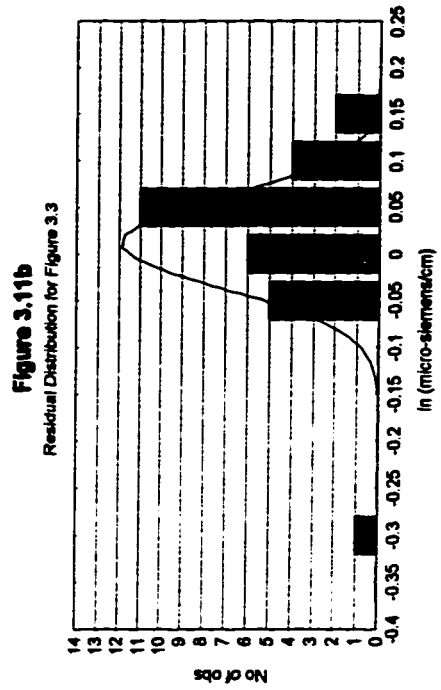
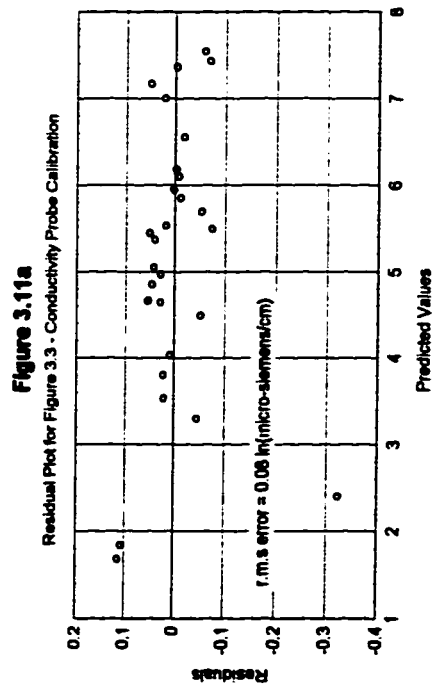
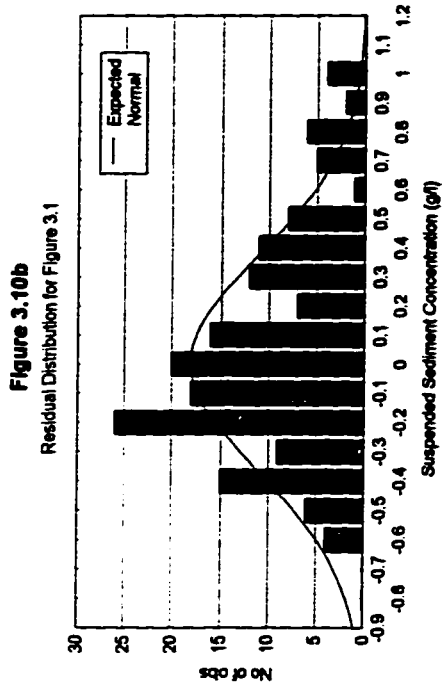
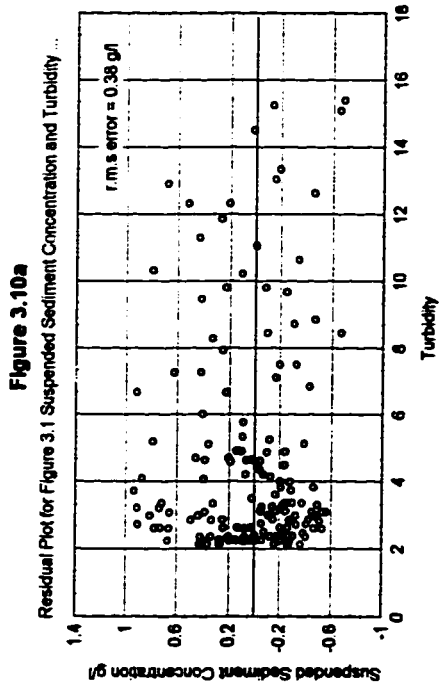
**Table 4.12**

***Size Analysis Data of Suspended Silt and Clay Over Season, Slims River, 1994***

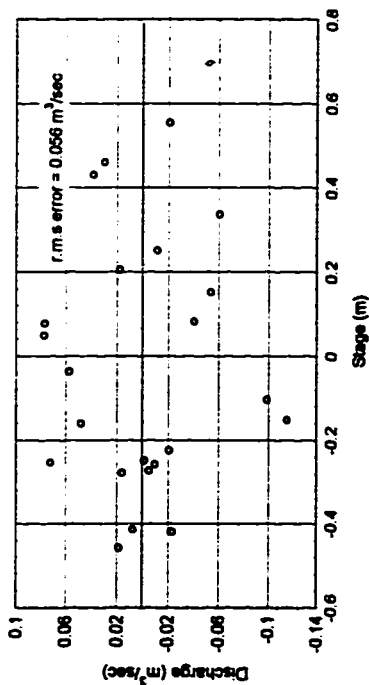
<b>Julian Day</b>	<b><math>\phi_{50}</math></b>	<b><math>M_z\phi</math></b>	<b><math>\sigma_G\phi</math></b>	<b><math>SK_G</math></b>	<b><math>K_G</math></b>
124	5.47	5.67	1.42	0.2	0.75
126	5.81	5.87	1.43	0.07	0.71
130	5.77	5.85	1.36	0.09	0.9
134	5.67	5.62	1.43	0.02	0.83
138	5.8	5.8	1.42	0.03	0.75
143	5.38	5.55	1.38	0.21	0.78
146	6	6.02	1.5	0.03	0.72
149	6.16	6.07	1.56	-0.04	0.71
152	6.16	6.07	1.5	-0.05	0.7
156	6.02	5.97	1.46	-0.01	0.7
160	6.1	6.04	1.53	-0.04	0.77
163	5.9	5.9	1.47	0.01	0.73
166	5.98	5.99	1.4	0.01	0.77
169	6.12	6.03	1.47	-0.06	0.74
174	6.5	6.31	1.43	-0.13	0.77
179	6.75	6.4	1.65	-0.25	0.79
184	6.48	6.28	1.53	-0.14	0.76
188	6.46	6.27	1.63	-0.12	0.75
192	6.56	6.34	1.48	-0.14	0.76
196	6.53	6.26	1.54	-0.2	0.7
202	6.55	6.31	1.59	-0.16	0.73
206	6.49	6.27	1.59	-0.17	0.73
210	7.1	6.81	1.41	-0.28	0.96
213	6.86	6.48	1.57	-0.29	0.92
215	6.62	6.27	1.62	-0.22	0.79
Bank Material	5.1	5.16	1.01	0.14	0.86

$\phi_{50}$  Median,  $M_z\phi$  Mean,  $\sigma_G\phi$  Standard Deviation,  $SK_G$  Skewness,  $K_G$  Kurtosis

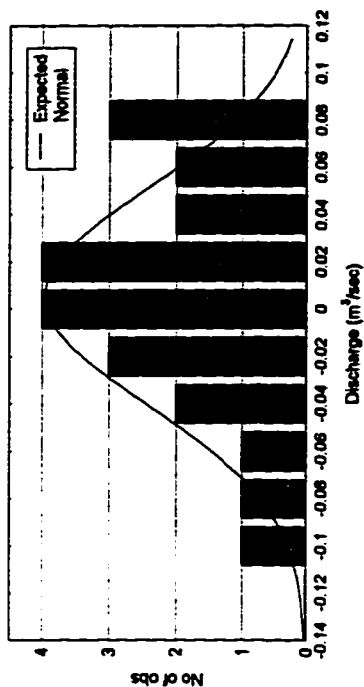
**Appendix D**  
**Analysis of Residuals**



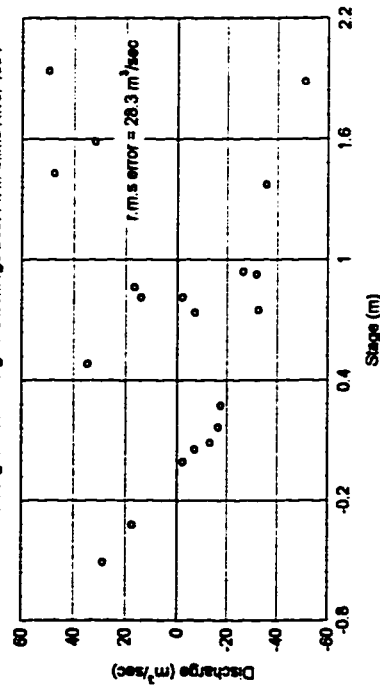
**Figure 3.12a**  
Residual Plot for Figure 3.4a - Stage & Discharge Best Fit ... Slims River 1993



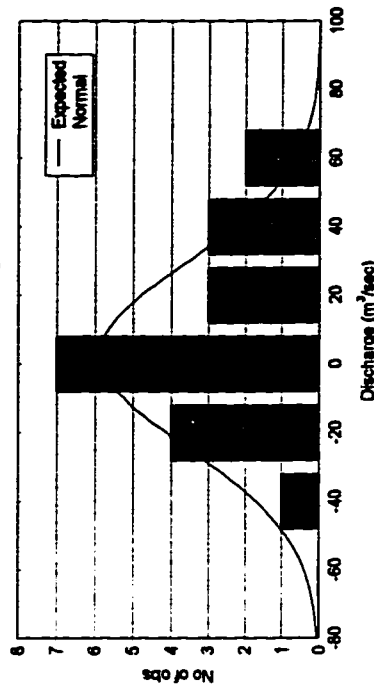
**Figure 3.12b**  
Residual Distribution for Figure 3.4a

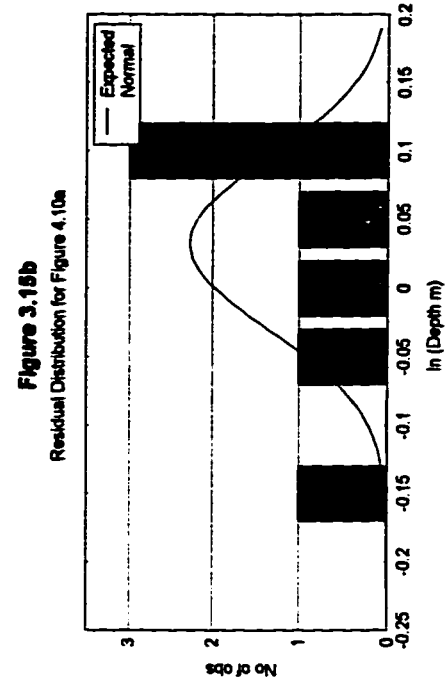
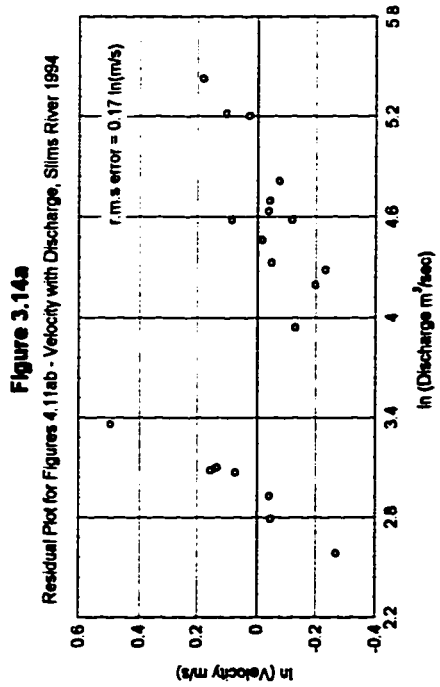
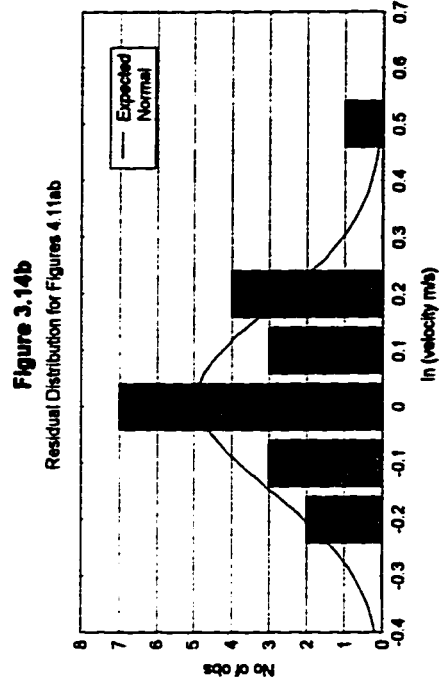


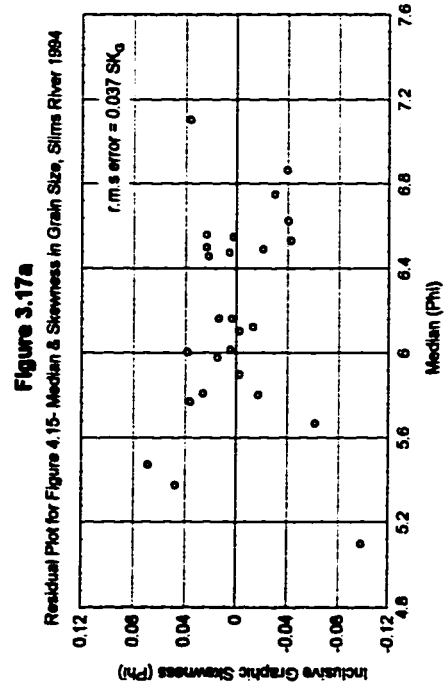
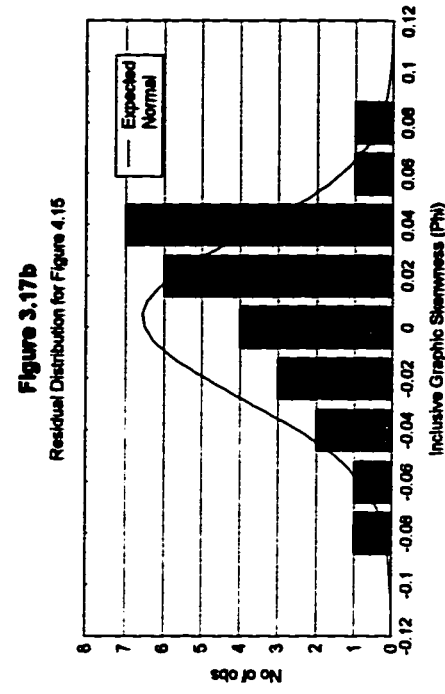
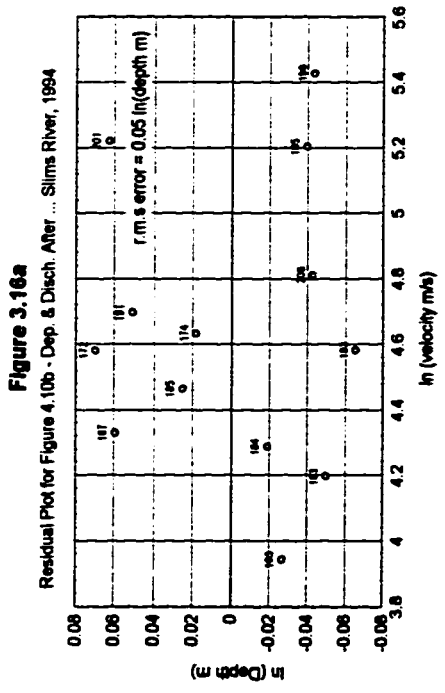
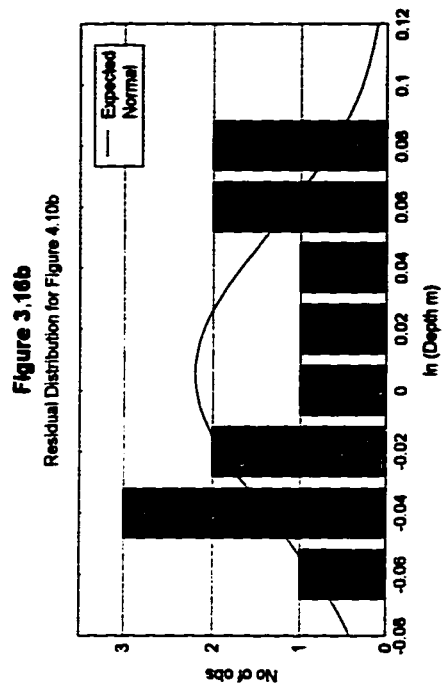
**Figure 3.13a**  
Residual Plot for Figure 3.4b - Stage & Discharge Best-Fit ... Slims River 1994

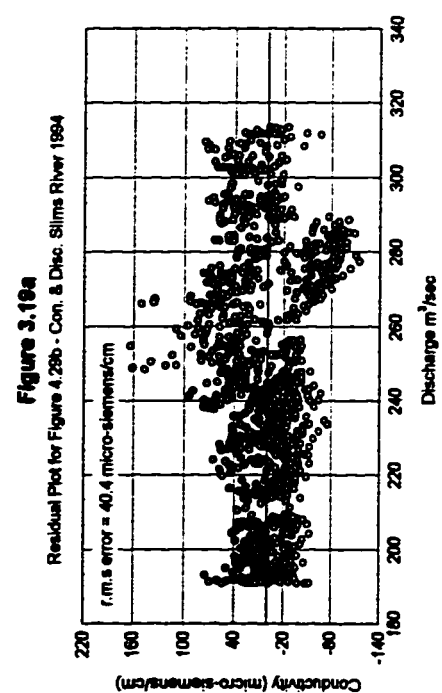
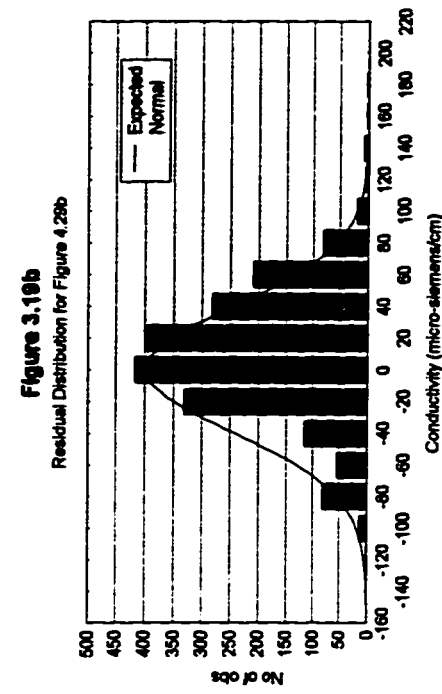
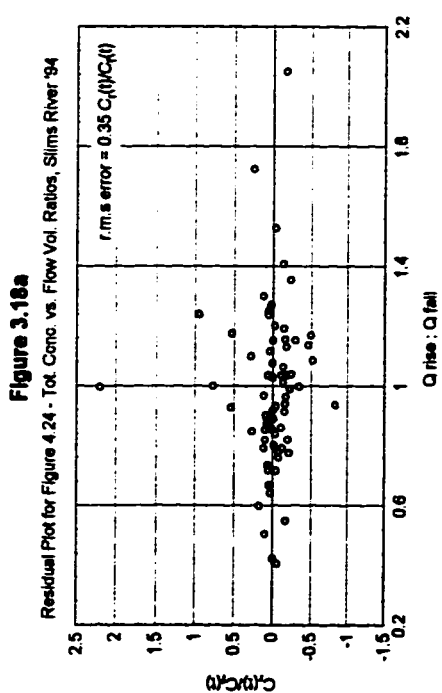
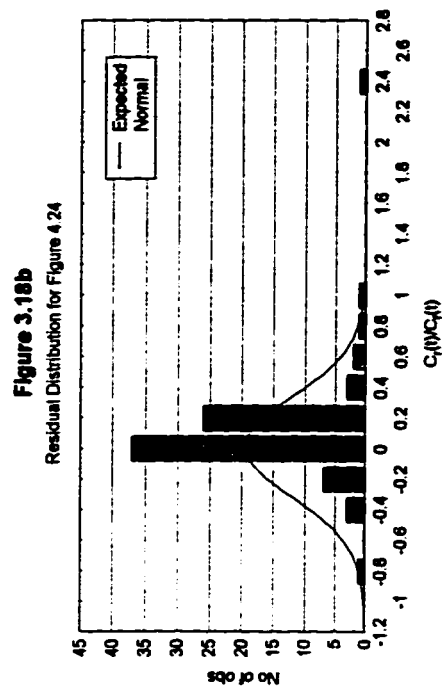


**Figure 3.13b**  
Residual Distribution for Figure 3.4b

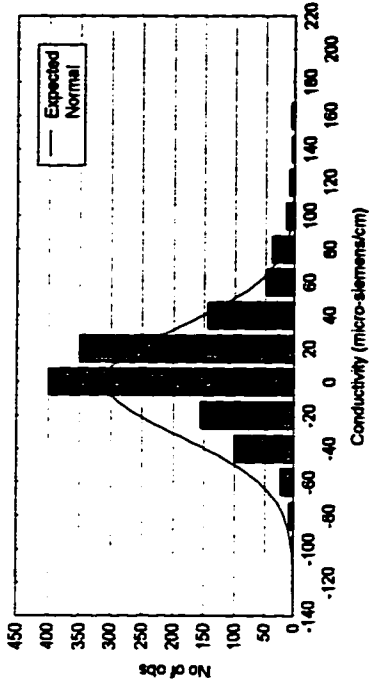




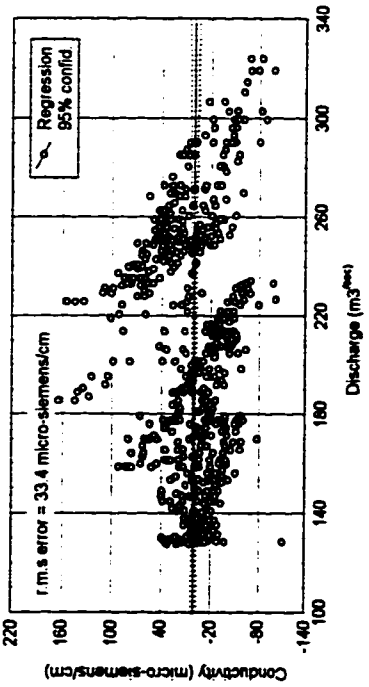




**Figure 3.20b**  
Residual Distribution for Figure 4.28b



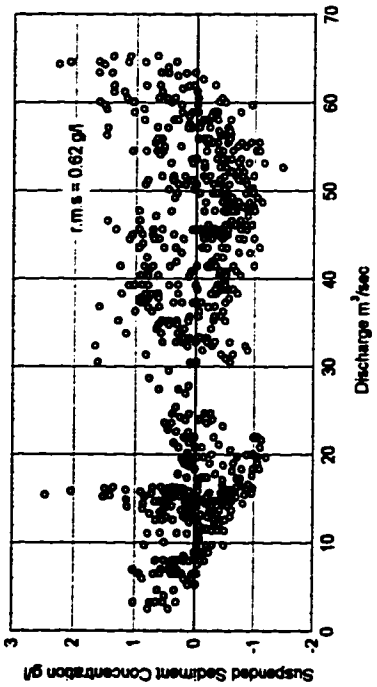
**Figure 3.20a**  
Residual Plot for Figure 4.28b - Con. & Disc. Slims River 1993



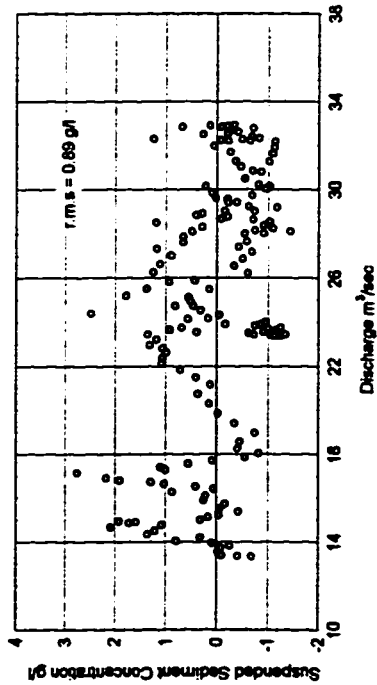
**Figure 3.21b**  
Residual Distribution for Figure 4.17a - SSC 1993



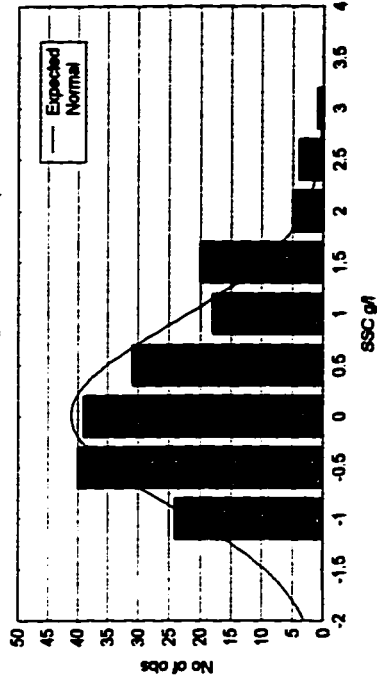
**Figure 3.21a**  
Residual Plot for Figure 4.17a - Q & SSC for 1993 without days 133-40 and 162-75



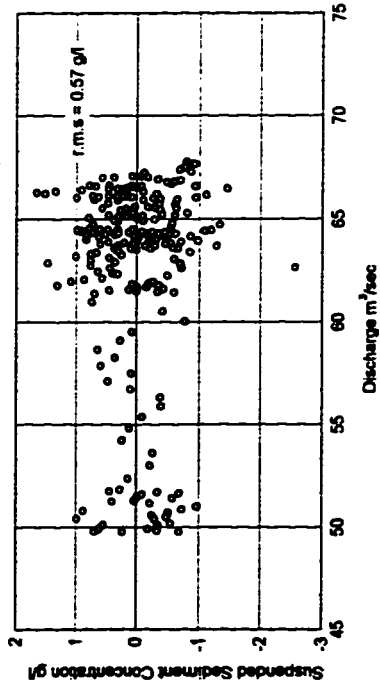
**Figure 3.22a**  
Residual Distribution for Figure 4.17a - Q & SSC for Days 133-140, 1993



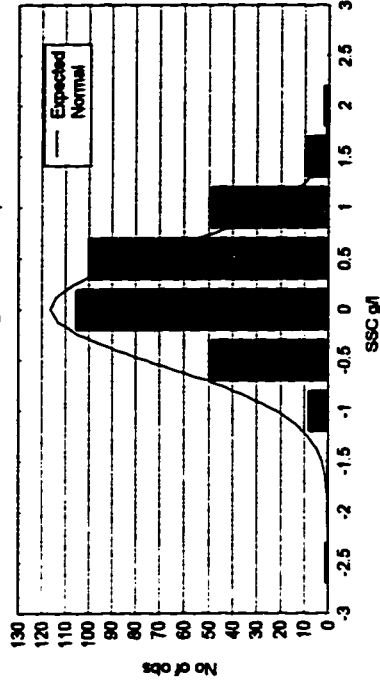
**Figure 3.22b**  
Residual Distribution for Figure 4.17a - For days 133-140

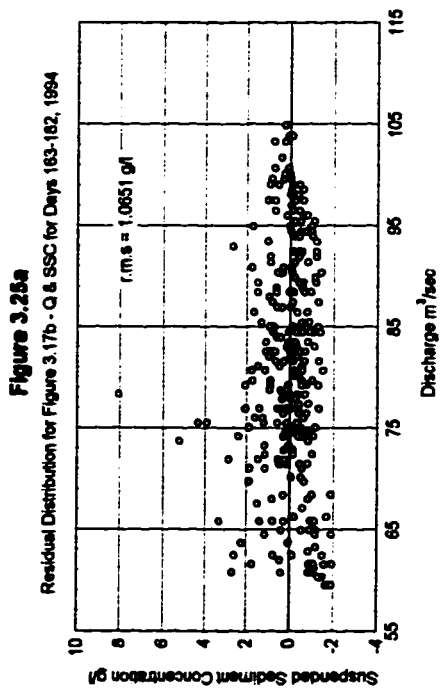
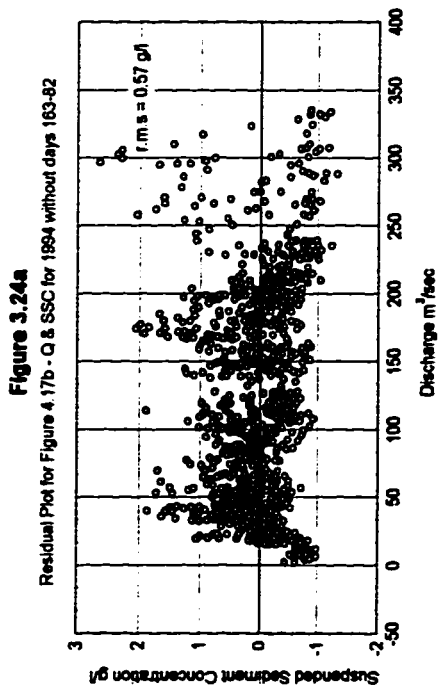
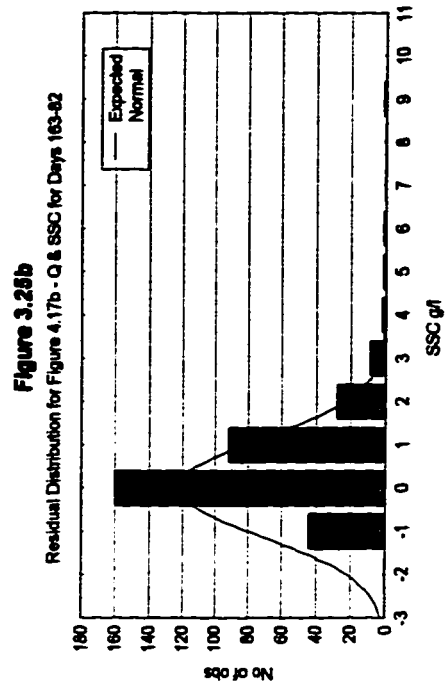
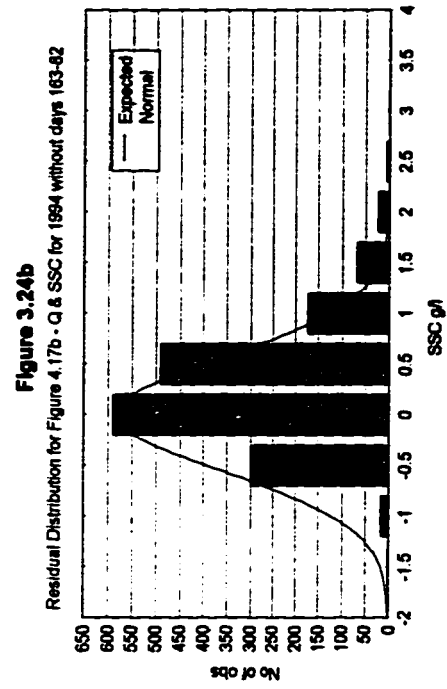


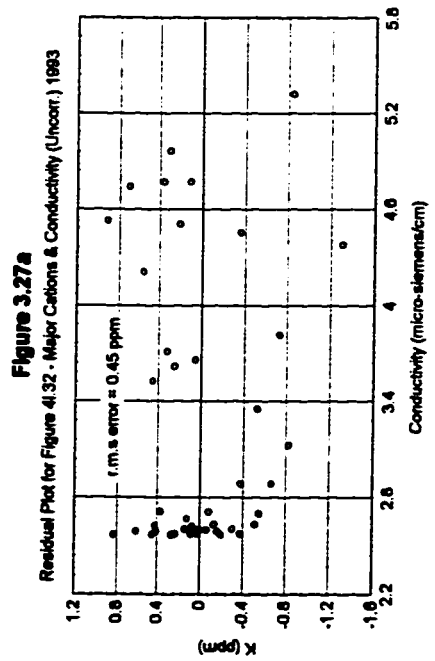
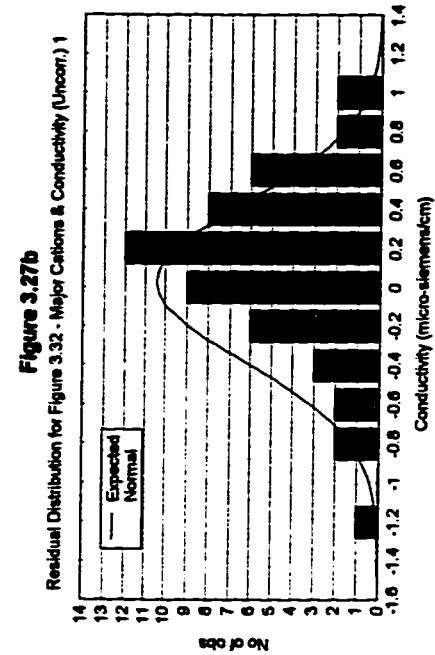
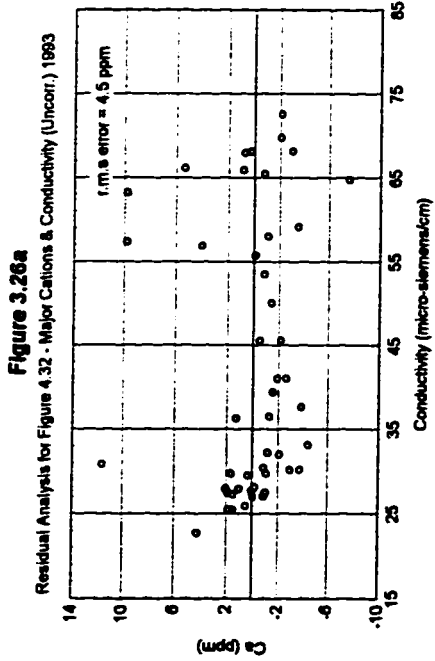
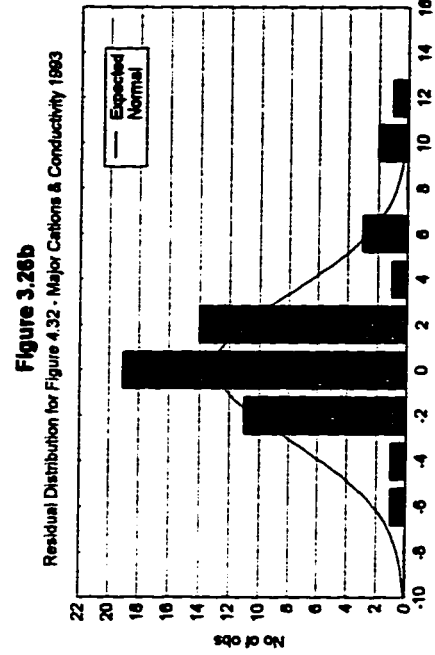
**Figure 3.23a**  
Residual Distribution for Figure 4.17a - Q & SSC for days 162-75

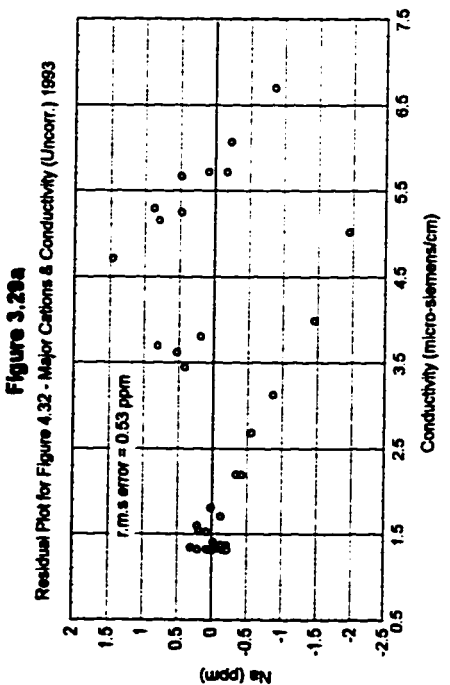
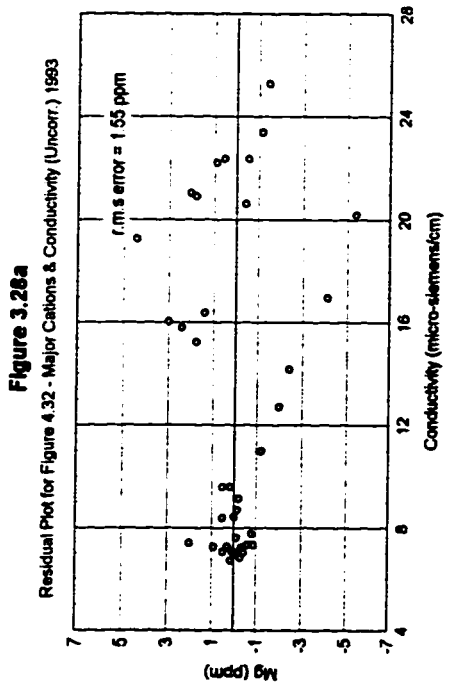
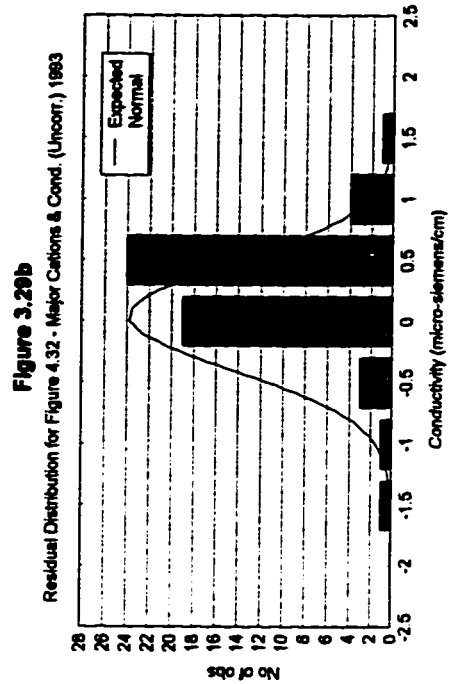
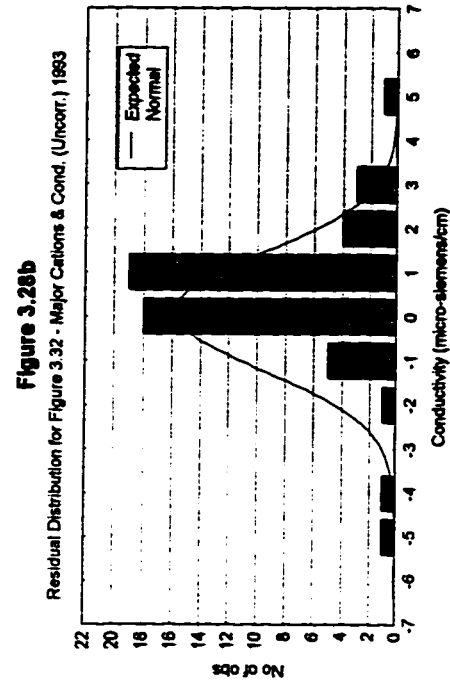


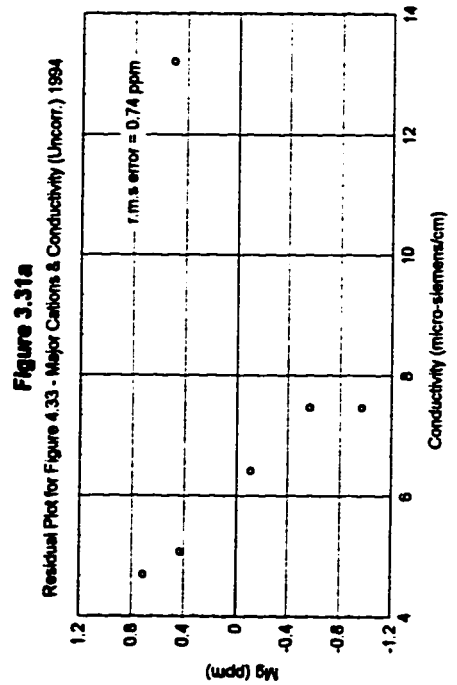
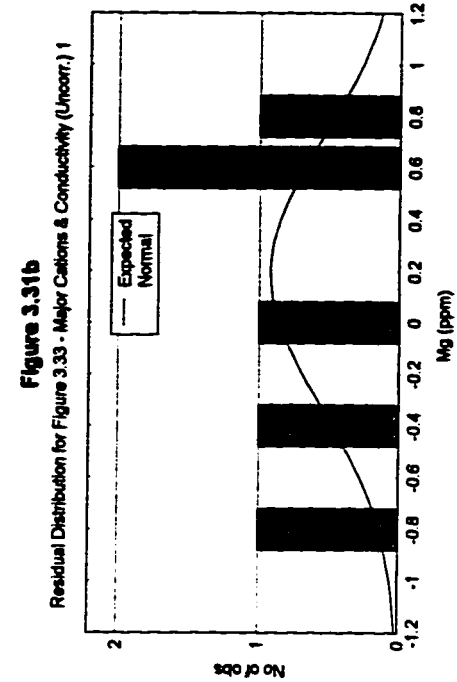
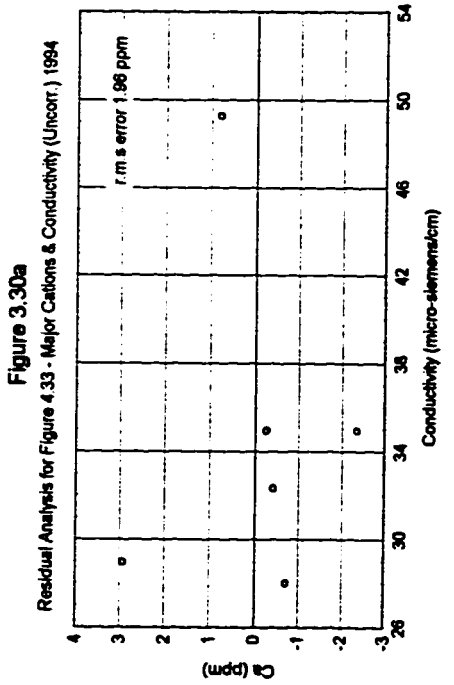
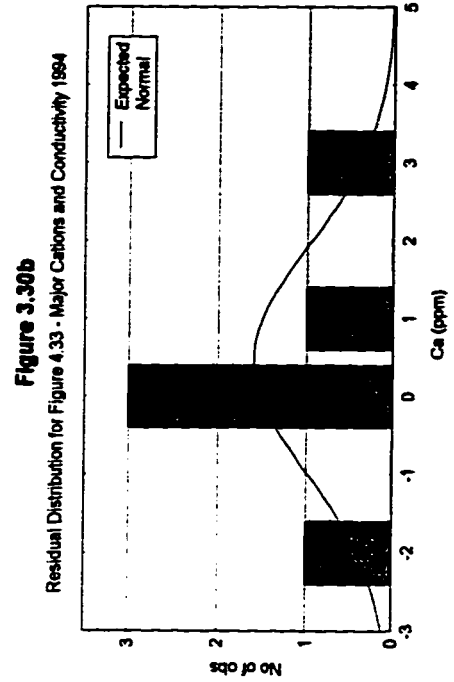
**Figure 3.23b**  
Residual Distribution for Figure 4.17a - Days 162-75



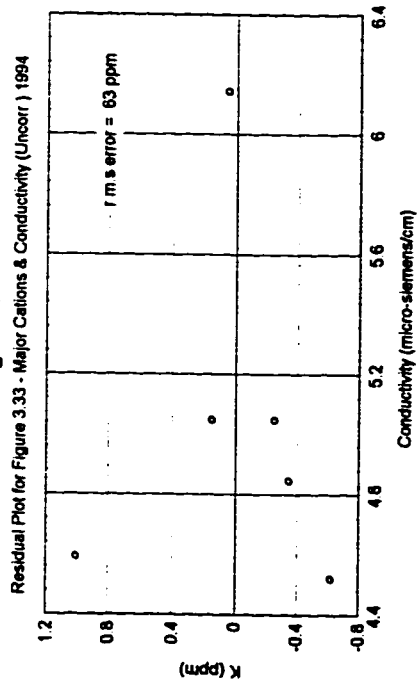




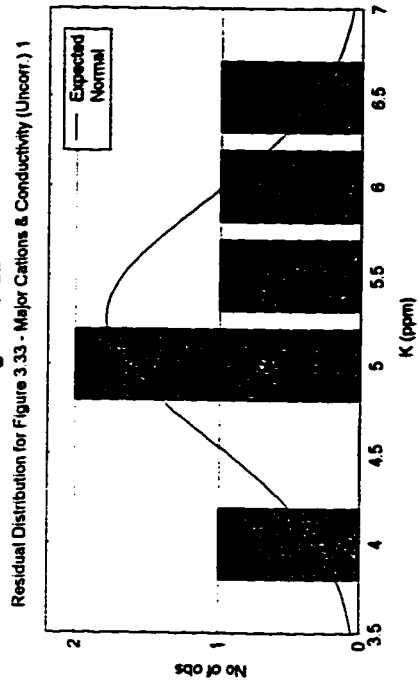




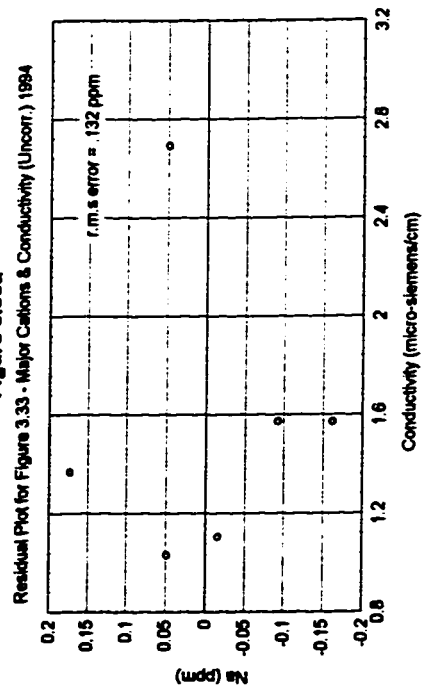
**Figure 3.32a**



**Figure 3.32b**



**Figure 3.33a**



**Figure 3.33b**

