GEOMORPHIC AND FLUVIAL RESPONSE TO RECENT PERMAFROST DISTURBANCES IN A HIGH ARCTIC RIVER, CAPE BOUNTY, NUNAVUT

by

Maryse Amélie Veillette

A thesis submitted to the Department of Geography

in conformity with the requirements for

the degree of Master of Science

Queen's University

Kingston, Ontario, Canada

(December, 2011)

Abstract

Using a sediment budget approach, suspended sediment transport dynamics were studied over the 2010 summer runoff season in the 8 km^2 West River catchment at the Cape Bounty Arctic Watershed Observatory (CBAWO), Melville Island, Nunavut. Research was carried out in an effort to determine the longer term impacts and response of recent (2007-8) active layer detachments (ALD) on the river system. In 2010, measured ALD inflows contributed 4.7% of the measured sediment yield, a decrease of 13.3% from 2007 when they initially formed. This indicates that while they continue to supply sediment to the main river, the impact they have on sediment fluxes, and hence the sediment budget has diminished, with time. Results from the sediment budget indicate that connectivity and the sediment delivery ratio within the system have also decreased with time. Sediment budget analysis shows that in response to this additional sediment, the West River progressively stores more sediment throughout the season, storing as much as 85% of sediment inflows during baseflow. Sediment was preferentially deposited within the channel, with coarser material deposited in the upper reaches, and finer material deposited in the lower reaches. Similarly, the transported and stored sediment became progressively finer with time, indicating the importance that river competence and wetted perimeter have on sediment transport as the larger sediment was entrained earlier in the season under higher flow conditions when the sediment was accessible, and finer sediment transported later in the season due to decreased competence and reduced accessibility of sediment. This sediment storage is expected to replenish sediment in the channel that is eroded during peak discharge in spring, and also dampens the effects that disturbances have on the sediment budget through storage. Sediment stored in the channel towards the end of the season does not contribute to sediment yield and may prove to be an important source of sediment in future years under late summer rainfall events.

Co-Authorship

The author wrote all chapters in this thesis with input from Dr. Scott Lamoureux and Dr. John Orwin, whose comments and suggestions helped improve the thesis.

Acknowledgements

First and foremost, I am especially grateful to my supervisor, Scott Lamoureux, for providing me with guidance, support and words of encouragement over the past two years. You have pushed me think about my project in the greater scheme of things and for that, I am extremely thankful. I especially enjoyed the Picasso's of my work.

Many thanks to everyone who has had to sit countless hours in the chem tent filtering water for SSC. Your patience and help was invaluable and greatly appreciated. Very special thanks to Kasey Kathan for your help, endless moral support, and for sharing your scientific knowledge and teaching me how to solder. To John Orwin, thank you for all your help and advice. I really enjoyed your story telling; it made the long hikes entertaining.

Thank you to my family and friends for always supporting and believing in me, and being there when I needed you. To all those fellow students, professors, and friends that I have not mentioned, you can rest assured that you are not forgotten, and I am grateful that you were there, for me, at all times. Thank you.

This research would not have been possible without the financial and logistical support of Natural Sciences and Engineering Research Council (grant to Scott Lamoureux), Arctic Net, Department of Indian Affairs and Northern Development (Northern Scientific Training Program), the Polar Continental Shelf Program, and Queen's University.

Table of Contents

List of Figures

List of Tables

List of Abbreviations

ALD	Active layer detachment
ALD05	ALD 05 subcatchment gauging station
a.s.l.	Above sea level
BS	Big Slide subcatchment gauging station
CBAWO	Cape Bounty Arctic Watershed Observatory
CR	Caribou subcatchment gauging station
d10, d50, d90	Grain size where 10, 50, and 90% of the distribution is smaller than it
G	Goose subcatchment gauging station
HI	Hysteresis index
PD	Pond water samples
PT	Ptarmigan subcatchment gauging station
RMS	Root mean square
RTS	Retrogressive thaw slumps
SSC	Suspended sediment concentration
SSY	Suspended sediment yield
SDR	Sediment delivery ratio
SWE	Snow water equivalence
WR1-7	West River gauging stations #1-7

Chapter 1

Introduction

In recent years there has been growing concern regarding the consequences of both observed and projected warming in the Arctic. Mean annual temperatures in the Arctic have already warmed at twice the average global rate of increase in the past decades, and are projected to continue increasing by another 4-7°C by 2100 if current trends continue (ACIA 2005). One of the potential consequences of this warming is an increase in permafrost degradation (Frey & McClelland 2009). Some models project that by the end of the century, there could be a 60-90% reduction in near surface permafrost across the Arctic (Lawrence & Slater 2005). One of the predicted effects of this degradation, especially in its early stages, is an increase in the occurrence of retrogressive thaw slumps (RTS) and active layer detachments (ALD). Both of these types of disturbance, if hydrologically connected to a river system, have the potential to increase the sediment loading to the river through increased exchanges between catchment and river, and higher rates of erosion (Woo & McCann 1994). This could have important effects on downstream ecosystems as the nutrients and potential contaminants transferred to rivers by thermokarst disturbances can be transported down the river system by means of the suspended sediment (Walling 1988, Walling et al. 2003). Moreover, increases in the concentrations of suspended sediment in rivers can reduce the amount of light transmitted through the water which can reduce the amount of photosynthesis occurring in river systems (Prosser et al. 2001).

Recent work in the High Arctic has demonstrated that unusually warm summer air temperatures, and/or high summer precipitation events can trigger the formation of ALD (Lewkowicz 1990, Lamoureux & Lafrenière 2009). Detachment slides of this nature can form over a period of days, as occurred at Cape Bounty, Nunavut, with detachment number 07-ALD-06 where it took four days for the slide to travel 650 m downslope (Lamoureux & Lafrenière 2009), or over a few hours. The latter occurred on the Fosheim Peninsula, where detachment slide BSC-2005-2, a much smaller ALD than 07-ALD-06 at Cape Bounty, took only a few hours to expand, after which little to no movement was observed (Lewkowicz 2007). The presence of ALD is not strictly limited to the Canadian Arctic Archipelago. They have additionally been seen in Alaska (Bowden et al. 2008), Russia (Leibman 1995), and in northern North America, such as the Mackenzie Valley (Lewkowicz & Harris 2005).

ALD are the primary type of landscape-scale disturbance occurring at the Cape Bounty Arctic Watershed Observatory (CBAWO) on Melville Island, Nunavut. They are slides that occur as a result of the rapid melting of ground ice located at the base of the active layer. This rapid melting of ground ice can be the result of intense rainfall, above average temperatures, or a combination of the two, as occurred during the summer of 2007 at the CBAWO (Lamoureux & Lafrenière 2009). The rapidity of the melt causes high porewater pressures, low effective strength and increased basal pore water pressures in the case of intense rainfall, which facilitate the downslope movement of the slide (Lewkowicz 2007). At the CBAWO, a combination of the aforementioned processes likely contributed to the formation of ALD. Many studies have documented the occurrence of such permafrost disturbances (e.g., Kokelj & Lewkowicz 1998, Lantuit & Pollard 2008, Bowden et al. 2008). Although the conditions under which ALD occur are well understood and there is growing literature on the erosion and runoff from ALD (Kokelj & Lewkowicz 1998, Lewkowicz & Kokelj 2002), there is still relatively little that is known regarding the potential effects that these localized disturbances may have on the larger channel systems that they drain into, and how the latter may be responding to such changes.

While numerous classifications of northern streamflow regimes exist, most northern and Arctic rivers are representative of nival regimes (Church 1974). Under such regimes, there is usually one major period of high discharge associated with spring snowmelt, followed by a period of recession and baseflow in summer (Woo 1986). It is during this period of peak discharge that the majority of the annual suspended sediment yield is transported (Lamoureux & Lafrenière 2009) as this typically represents the period when rivers are most competent (Hooke 2003). Other than the in-stream sediment available from the channel, it has been found that the transport of suspended sediment in streams is related to the amount of sediment supplied by the adjacent hillslopes (Gomi et al. 2004). Therefore, in a changing Arctic environment with an increased potential for landscape disturbances such as ALD and RTS, sediment transport dynamics could be potentially altered, resulting in greater sediment yields to the river. The disturbances, which can remain active for years, can represent significant means of sediment delivery if contiguous to the river system (Lewkowicz 1987, Kokelj & Lewkowicz 1998).

Previous short term observations of the impacts that ALD have on river systems have indicated that following their initial occurrence, downstream turbidity and sediment transport increased abruptly (Lamoureux & Lafrenière 2009). However, longer term observations of the fluvial impacts of ALD are unavailable, making it difficult to determine the ongoing effects to sediment yield and the recovery time of fluvial systems from these disturbances. Given the lack of available data to assess these longer term responses, the objectives of this research are twofold:

1. Determine how both sediment transport and fluxes in the West River (unofficial name) of the Cape Bounty Arctic Watershed Observatory (CBAWO), Melville Island, have responded to recent extensive ALD.

2. Determine the geomorphic changes to the West River channel system in response to these catchment permafrost disturbances.

These goals will address whether ALD constitute a significant source of sediment to river systems and provide results to assess the impact of this sediment to watershed-scale sediment dynamics. The second goal will additionally provide indications on whether the disturbances have accelerated dimensional and geomorphic changes to the river channel.

In relation to sediment storage and fluxes, it is hypothesized that the highly disturbed river reaches will contribute more sediment to the overall river system through increased tributary inflow and channel bank erosion. This will result in greater sediment availability compared to minimally disturbed reaches. Further, the largest contribution of sediment from ALD is predicted to be associated with peak discharge, and will diminish as the season progresses and the river flow competence decreases.

Determining the sediment budget of river basins is difficult if only measuring the sediment export at the river's mouth as this single measurement does not provide information regarding either the storage or release of sediment availability as it travels downstream (Walling et al. 1998). Understanding the transfer processes of suspended sediment along a river system from the headwaters to the mouth of the river through detailed knowledge of the spatial and temporal changes in suspended sediment yield can aid in the estimation of future recovery from landscape disturbances. Therefore, in an effort to answer these research questions, the West River of the CBAWO was divided into five reaches representing different degrees of hydrologically connected disturbance, ranging from highly disturbed to relatively undisturbed. At the outlet point of each reach, gauging stations were established allowing for velocity measurements and water samples for suspended sediment concentration (SSC). These measurements permit river discharge and sediment fluxes to be determined for each reach. An evaluation of the sediment flux entering and exiting each reach will determine whether the reaches are sources or sinks of sediment to the system.

It is important to relate field mapping of river morphology to the overall sediment budget for a river as the former is believed to reflect characteristics of the latter (Hooke 2003). Part of this thesis will be aimed towards linking the morphology of the channel to the measured sediment fluxes in an attempt to aid in the explanation of the observed flux differences in terms of channel morphology.

If warming in the Canadian High Arctic continues as projected, then it is likely that the rate of permafrost degradation will be accelerated and landscape disturbances such as ALD and RTS will become more widespread. To date, few studies have investigated the suspended sediment budgets in Arctic rivers and related them to known landscape disturbances. Therefore, with the risk of increased permafrost disturbance, it is becoming more crucial to understand how sediment fluxes, sediment storage and release dynamics, and river morphology are affected within the channel in the years following recent disturbances in order to better infer future response from such events. These results will contribute to a greater understanding of how fluvial systems are impacted by permafrost disturbances, and provide the initial sediment budget measurements for long term sediment budget monitoring at the CBAWO. This represents an important baseline for future long-term studies of sediment budgets and biogeochemical studies for the West River in analyzing how sediment dynamics compare on yearly time scales and how the fine sediment storage affects carbon and nitrogen cycles.

Chapter 2

Literature Review

2.1 Introduction

The Canadian High Arctic is a region that has, and will continue to undergo significant warming in comparison to other areas across the globe if current trends continue (ACIA 2005). Such warming can lead to an increase in the occurrence of permafrost degradation which could alter sediment transport and dynamics within alluvial systems. These changes can occur as the initiation of sediment slugs or the development of buffers, barriers, or blankets, which can interrupt the connectivity of sediment within the catchment. This chapter provides an overview on Arctic hydrology, sediment transport and dynamics, the sediment delivery ratio and how a lack of connectivity within the catchment can reduce this ratio. The chapter will finish with an overview of two types of thermokarst perturbations that are expected to occur more frequently under a warming climate in the Arctic.

2.2 Sediment transport and dynamics

Moving downstream from the headwaters, the sediment dynamics of fluvial systems can be divided into three general zones based on their location within a watershed (Church 1992). These are zones of sediment erosion, transport and deposition (Knighton 1998). The sediment delivery ratio (SDR) is a concept used to quantify the amount of sediment delivered from a watershed, by comparing the amount of sediment carried out of the stream at the watersheds outlet, to the total erosion that occurred within the watershed (Walling 1983). The delivery of sediment from the catchments headwaters to its outlet may not always be attained in one continuous motion. Instead, continuity can be interrupted by periods of sediment deposition and temporary storage as landscape features limit the delivery of sediment, thereby reducing the sediment delivery ratio (Fryirs et al. 2007, Walling 1983).

2.2.1 Erosion, transport, and deposition

There exist three stages of sediment motion in rivers: 1) initiation of movement, 2) transport, and 3) deposition. In order for each stage to occur, a critical or threshold value unique to each phase, and which is dependent on the physical properties of the sediment, notably grain size, must be attained (Miller et al. 1977). Each value is determined by hydrodynamic factors such as stream power, velocity and shear stress (Gordon et al. 1994). When stream competence or the maximum particle size that a water system can carry (Gordon et al. 1994, Hooke 2003) is reduced due to low discharge, the coarser particles will settle out of suspension first due to their higher settling velocities, and the smallest particles comprising the suspended sediment load, notably clay and silt, will remain in suspension the longest (She et al. 2005). These finer particles will additionally be the first type of sediment transported, initially through saltation (Knighton 1998). With increasing discharge, transport competence will increase, erosion of the bed and banks will begin, and particles will be entrained. If the critical threshold for sediment movement is reached, sediment transport will occur (Miller et al. 1977).

There are three general categories of erosion. These are weathering and weakening, mass failure, and hydraulic action (Couper & Maddock 2001). Weathering and weakening will generally be the first type of erosion acting on surfaces as it consists of the processes which weaken the soil so that the more direct erosion methods of mass failure and hydraulic action can occur. An example of this preliminary type of erosion would be the continual wetting and drying of the soil. However, in Arctic environments, cycles of freeze-thaw activity are a common example. This first step in the erosion process not only acts on channel banks, but everywhere within the catchment (Couper & Maddock 2001). Sediment eroded from the hillslopes can reach a river in two ways. The first is by overland flow, and the second, by direct deposit into the river (Arnell 2002). Hydraulic action and mass failure are more specific to fluvial systems, especially in the middle reaches of the catchment as this represents a region of enhanced stream power (Lawler 1992). Hydraulic action generally acts on non-cohesive surfaces, and represents the gradual removal of individual particles in response to the frictional drag force and shear stress caused by running water (Knighton 1998). Continual erosion by hydraulic action can scour and undercut channel banks creating overhangs, leaving them susceptible to failure (Couper & Maddock 2001). Contrarily, mass failure is the gradual weakening of the bank material from cycles of wetting and drying (general weathering and weakening) which promote expansion and contraction leading to the development of cracks in the soil. This decreases bank stability until the soils along the bank collapse into the channel (Couper & Maddock 2001). While hydraulic action is most effective on non-cohesive banks, the effectiveness of mass failure is dependent on soil properties and the structure and geometry of the bank. Once collapsed, the fallen blocks will either be transported by the river if the critical shear stresses are met which favour transport, or will remain in the channel and reinforce bank stability by decreasing the bank angle, and be subject to hydraulic action (Thorne 1982, Knighton 1998). In addition to the aforementioned erosion processes, a decrease in sediment input to the channel or an increase in its competence will lead to erosion. No matter which process occurs, all types of erosion will result in degradation of the bed over the length of affected river (Hey 1987).

2.2.1.1 Initiation of movement

In order for sediment transport to occur, a critical entrainment shear stress (τ_c) must be overcome in order to dislodge the particle from the river bed (Miller et al. 1977). The ability to overcome this shear stress is related to a particles stability which is dependent on the lift and drag forces of the flow, and the weight of the particle (Figure 2.1) (Carling 1992). Particles with larger pivot angles have reduced exposure to flow, making it harder for them to be dislodged and entrained (Carling 1992).

This figure has been removed due to copyright restrictions.

Figure 2.1 Necessary forces for entrainment acting on stationary particles. Larger particles have smaller pivot angles (ϕ) and are more exposed to the flow of water (Komar 1989 in Carling 1992).

The shear stress associated for any given particle can be determined with the following non-dimensional equation

$$\theta = \tau_c / (\rho_s - \rho)gD \tag{2.1}$$

where θ is Shields number, τ_c is the critical shear stress necessary for entrainment to begin, ρ_s and ρ are the density of the sediment and water respectively, g is the acceleration due to gravity, and D is the particle diameter (Carling 1992). A general relationship between Shields number and the grain Reynolds number (Re_*)

$$Re_* = u_* D / \nu \tag{2.2}$$

where u_* is the friction velocity, D is the particle diameter, and ν is the kinematic fluid viscosity can be used to graphically determine the threshold at which particles of various sizes will be entrained (Figure 2.2) (Miller et al. 1977). For high Reynolds numbers, when the fluid viscosity becomes unimportant, the Shields number remains relatively constant at a value near 0.045 (Miller et al. 1977).

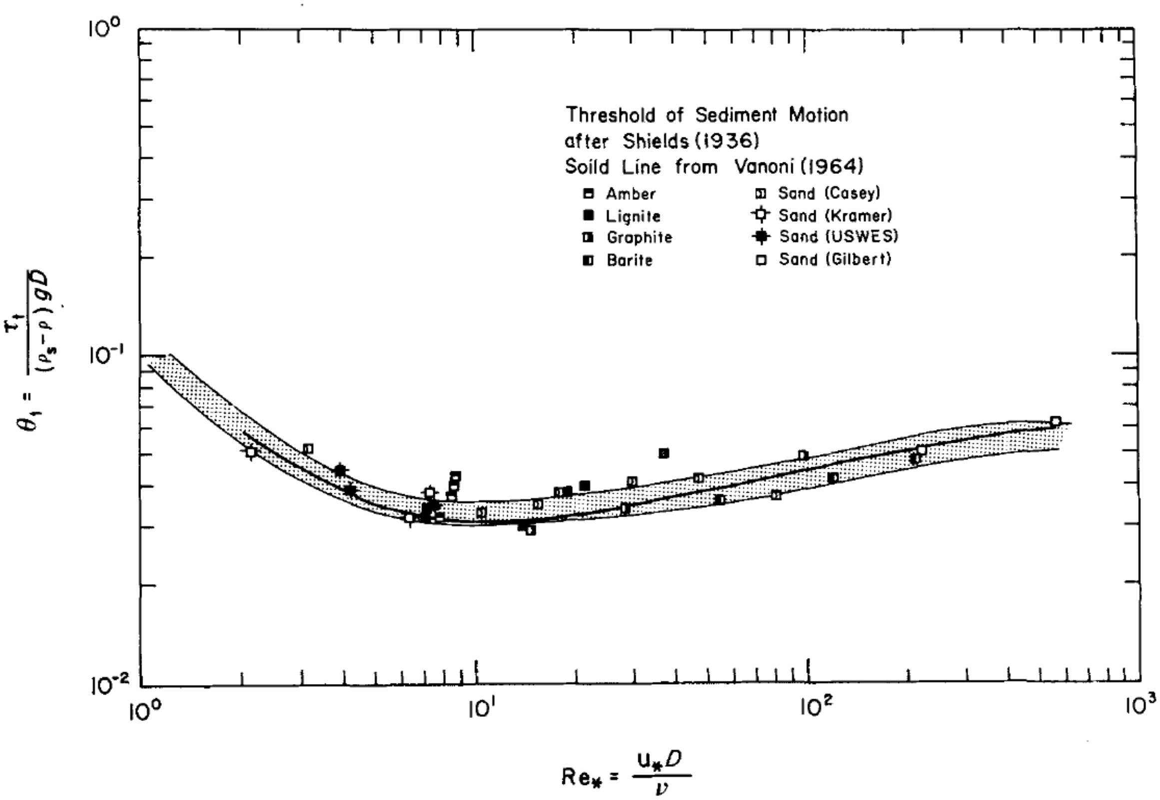


Figure 2.2 Shields curve depicting the relationship between Shields and Reynolds number (Miller et al. 1977).

2.2.1.2 Transport

Three types of load can be transported for natural streams: 1) dissolved load, 2) suspended load, and 3) bedload. Dissolved load represents any material (e.g., Ca^{2+}, Cl^{-}) that is transported in solution but does not affect channel morphology (Restrepo et al. 2006). Suspended sediment load is the sediment, usually smaller than 0.2 mm (sand, silt, clay) that is being transported in suspension by the upward eddies of water (Ewing 1996, Arnell 2002). Because clay particles have such small settling velocities due to their small grain size, they generally remain in suspension and are transported downstream at the same velocity of water.

Deposition will only occur when flow velocities are significantly reduced (Knighton 1998). However, the transport of coarser sediment (gravel and sand, and coarse silts) is strongly dependent on flow velocity and hence will not remain in continuous suspension (Figure 2.3).

This figure has been removed due to copyright restrictions.

Figure 2.3 Hjulström Diagram depicting the relationship between erosion, transport and deposition and flow velocity for material of various sizes (Hjulström 1935 in Knighton 1998).

Suspended sediment generally represents the largest fraction of a rivers sediment yield, and direct relationships have been established between suspended sediment concentration (SSC, mg/l) and discharge, where more sediment is carried in suspension during periods of high discharge. This relationship has been expressed empirically,

$$C = aQ^b \tag{2.3}$$

where C is the sediment concentration, a is a constant, Q is discharge, and b is a constant typically between 1 and 2 (Walling & Webb 1992). Contrary to dissolved load which does not affect channel morphology, suspended sediment can alter channel form (Knighton 1998). The third type of load is bedload, which represents material that is too large to be held in suspension by the waters turbulence and hence is transported downstream by saltation, which describes a rolling, bouncing and sliding movement along the channel bed as well as short distances travelled in the flow (Carling 1992, Arnell 2002). It generally represents material coarser than 0.2 mm and is typically found on the channel bed. Bedload will rarely remain in suspension more than few seconds, indicating that bedload transport is closely related to stream competence, discharge and turbulence (Beschta 1987, Knighton 1998). However, as long as the time-averaged entrainment forces exceed the gravitational force acting on the particles making up the bedload, they will remain in motion (Paiement-Paradis et al. 2011).

As previously mentioned, the lift and drag forces of water play an important role in the entrainment of material (Carling 1992). Bedload in particular can be quite dependent on flow turbulence for transport (Nelson et al. 1995, Paiment-Paradis et al. 2011). Various studies have found that turbulent accelerations in the water upstream of bed material results in an increase in the uplift forces that support bedload movement (Drake & Calantoni 2001, Paiement-Paradis et al. 2011). In a laboratory flume experiment, Nelson et al. (1995) determined that although fluctuations occurring at high frequencies are very important in bedload transport, there exists a strong relationship between streamwise turbulence of a high magnitude (greater than the mean) but low frequency, and bedload transport which cannot be ignored.

In an effort to determine the capacity of a given flow velocity to transport sediment, numerous sediment transport rate equations have been established whereby the sediment transport rate per unit width is related to excess discharge per unit width (Schoklitsch equation), excess shear stress (Du Boys equation), or excess stream power per unit width (Bagnold's equation). However, because the conditions under which each equation was determined is unique to a given flow type, an equation that can be generalized to all rivers has not yet been established (Knighton 1998). Additionally, because many of the equations are developed for a given grain size, it is not easily applied to a range of rivers unless used solely in those rivers with similar characteristics (Carling 1992).

2.2.1.3 Deposition

Deposition is the third stage of sediment movement. Suspended sediment is deposited when the shear velocity, or flow velocity, falls below that of the settling velocity. Settling velocities are dependent on grain size, with larger particles having higher settling velocities (She et al. 2005). Because of this, larger particles are deposited first, followed by increasingly smaller ones (Figure 2.3). The overall effect of this graded fallout is a horizontally and vertically sorted river bed. When sediment is deposited on the bed, it can alter channel dynamics through changes in bedload transport and flow resistance as is the case when channel bars form (Knighton 1998).

2.2.2 Fluvial channel structure

Although three types of stream channels exist; alluvial, semi-controlled and bedrock (Schumm 1985), the following section will concentrate on alluvial channel forms and their respective patterns as the bed of many high latitude channels are dominated by alluvial sediment (French 2007). River patterns are controlled by hydrologic parameters as discharge has a strong control on channel width and depth, and meander wavelength and amplitude (Schumm 1985).

2.2.2.1 Channel patterns

Because of the ease with which sediment can undergo erosion, transport and deposition, alluvial channels are highly susceptible to pattern changes. Such changes are influenced by the type and amount of sediment being moved through the system (Schumm 1985), as well as the runoff regime of the river because of the influence of flow strength on sediment transport (van den Berg 1995, Kleinhans 2010). Braided, straight, and meandering are the three broad categories of channel pattern (Schumm 1985). These patterns are influenced by discharge and the channel's bed and bank material.

The degree of sinuosity can be used to determine if a river is straight or meandering (van den Berg 1995). Sinuosity is calculated as the ratio of channel length to valley long-axis length, and varies from 1.0 for straight channels to approximately 3.0 for highly meandering rivers (van den Berg 1995, Schumm 1985). Braided channels for example consist of two main types; bar and island braided. Bar braided channels are less stable than island braided channels because the islands are more permanent features that contain vegetation which stabilizes the bars and hence the channel, while bar braided channels do not contain vegetation, rendering the channel more unstable (Schumm 1985). If a river contains islands whose width is greater than three times the total width of water during periods of average discharge, and whose banks are stable, then the river is considered to be anabranching (Nanson & Knighton 1996). This is different than anastomosing channels which are made up of multiple smaller channels that separate and rejoin the main channel over lengths that are much longer than the channel width (van den Berg 1995).

These smaller channels can be specifically classified as straight, braided, or meandering which is why this specific pattern does not fall into a classification of its own (Schumm 1985).

2.2.3 Landscape connectivity

Connectivity of sediment throughout a landscape refers to the ability of the sediment to be transferred from its source to end zones, or any two components, in one free motion (Harvey 2001). In fully connected river systems, the sediment introduced into the channel in the headwater zone would travel downstream and be discharged at the mouth without intermediate storage (Fryirs et al. 2007). However, this degree of connection rarely occurs without intermediate stages of either temporary or permanent sediment storage as certain landforms can limit the link between catchment and channel or upstream and downstream channels, thereby affecting the downstream sediment delivery (Fryirs et al. 2007). Catchment scale sediment budgets are a means of representing the level of connectivity within a stream through the identification of sediment sources, which are regions within the catchment where sediment is made available through erosion, and sinks, which are regions within the channel where storage of sediment occurs. Through the application of a sediment budget, regions acting as sinks become identifiable, and these indicate regions of sediment transport discontinuity. In the following section, three forms of decoupling agents, buffers, barriers, and blankets, will be distinguished based on how they interrupt this continuity.

2.2.3.1 Buffers, barriers, and blankets

Landforms within a catchment that prevent landscape connectivity and continuity can be classified as buffers, barriers and blankets (Fryirs et al. 2007). Buffers are those landforms that disrupt the longitudinal and lateral connectivity within a catchment. They prevent external sources of sediment within the catchment from entering the channel system (Fryirs et al. 2007).

Examples of buffers include fans that are found at the breaks in slope near valley margins and alluvial pockets of floodplains in low sloping areas (Harvey 2001). Once sediment has entered the channel, barriers can disrupt the longitudinal connectivity of the sediment by preventing it from readily being transported downstream (Fryirs et al. 2007). Step-pool river systems are an example of barriers because they reduce the slope of the channel resulting in trapped sediment within the pool, thus creating disconnect between upstream and downstream sediment movement. The natural widening of a channel can also be considered a barrier if stream competence is lowered sufficiently to induce sedimentation (Fryirs et al. 2007). As barriers get frequently reworked, they tend to be transient in nature (Fryirs et al. 2007). Finally, blankets are features that prevent the reworking or transport of bed sediment by shielding or armouring the bed (Church et al. 1998, Fryirs et al. 2007). Similar to buffers, blankets can be localized or extensive in space. Usually found along alluvial rivers where there is an abundance of sediment sinks, they include features such as the fine grained material that fills gravel bars or bed armour which prevents the reworking of the bed under the stress of the running water (Church et al. 1998).

2.2.4 Sediment delivery ratio

Before sediment can move through rivers, it must first be eroded from the river basin and transported to the channel. As previously mentioned in section 2.2.1, the erosion can occur as weathering and weakening, mass failure, or hydraulic action (Couper & Maddock 2001). However, the delivery of sediment from catchment to channel is not a straight forward process, and only a small portion of the eroded sediment will typically find its way to the outlet to be considered in the sediment yield (Walling 1983). Sediment delivery is further complicated once the eroded sediment reaches the channel because barriers can interrupt longitudinal connectivity (Fryirs et al. 2007), further decreasing the sediment available for yield. In an effort to better understand the conveyance of sediment within a catchment, and not only the sediment yield at the basin's outlet, Walling (1983) developed the concept of the sediment delivery ratio (SDR) which is the ratio between the sediment that reaches a basin's outlet to the total erosion that occurred within the basin.

$$SDR = \frac{ratio\ of\ sediment\ delivered\ at\ the\ catchment\ outlet\ (t/km^2/yr)}{gross\ erosion\ within\ the\ basin\ (t/km^2/yr)} \tag{2.4}$$

In addition to being influenced by the deposition and storage within the channel, there are many environmental and geomorphological processes that influence the magnitude of the SDR (Walling 1983). These are the types of sediment sources and their proximity to river networks, the topography of the catchment along with land use and vegetation cover. As buffers, barriers and blankets have the ability to limit the transfer of sediment through the catchment due to their impacts on sediment fluxes, they allow for the decoupling of catchment sediment sources from the channel network. This level of disconnect and decoupling is related to the location and persistence of these impeding landscape features (Fryirs et al. 2007). These features not only control the level of disconnect, but also the magnitude of the changes that the channel will undergo and the delivery of sediment. For example, if the flow of water is low, the energy input will be limited, and the likelihood of slope erosion will be reduced. Therefore, there will be a limited amount of sediment to circulate through the system, thus affecting the SDR (Fryirs et al. 2007). The delivery of sediment is highly scale dependent, whereby smaller catchments tend to have more sediment delivered to the outlet (Walling 1983). This is because larger catchments generally have gentler slope and channel gradients that reduce stream velocities. Resultant sedimentation in the channel system can act as a barrier and result in reduced yields (Fryirs et al. 2007). Additionally, larger catchments are generally better able to manage inflows of turbid water as they have greater sediment storage capacity in their lower reaches when stream deposition occurs (Walling 1983, Fryirs et al. 2007), resulting in a smaller SDR.

2.2.5 Sediment waves

2.2.5.1 Downstream migration of waves

When coupling between catchment and river is high, or when erosion rates are elevated, there is an increased opportunity for sediment delivery to the river channel. This can lead to the formation of sediment slugs, or waves, which are large pulses of sediment that move through the channel system by dispersion, translation, or a combination of the two (Figure 2.4) (Nicholas et al. 1995, Lisle et al. 2001). Sediment waves can be sourced intrinsically (endoslugs) by channel bed and bank erosion, or extrinsically (exoslugs) such as slope inputs from landscape disturbances (Nicholas et al. 1995). The formation of small sediment slugs is strongly influenced by the fluvial characteristics of the channel (e.g., water velocity, sediment supply/storage) and can occur regularly. However, larger waves are typically formed from external sources (Nicholas et al. 1995). Waves can occur at different time scales, and under different scenarios. For example, in response to major storm events or landscape disturbances located in the catchment that generate high erosion rates and sediment delivery to the channel (Potyondy & Hardy 1994, Nicholas et al. 1995, Benda & Dunne 1997a). Once formed, sediment waves can vary in length from a few meters to several kilometers (Benda & Dunne 1997b), and can last for long periods of time, up to 20 years in some documented cases (Madej & Ozaki 1996).

There are two ways by which sediment slugs can travel downstream, but often times, a combination of the two will be displayed (Nicholas et al. 1995). The first is dispersion. A dispersive wave is one in which the apex and back edge of the wave do not travel downstream; however the leading edge does, hence wave amplitude decreases with time (Figure 2.4). A small backwater effect on the stoss side of the dispersive wave can result in an accumulation of sediment on the trailing edge. As the wave continues to disperse and accumulate sediment, the back edge of the wave slowly migrates upstream creating a symmetric wave around the apex (Nicholas et al. 1995). Translation is the second way by which a wave can migrate downstream. Unlike a dispersive wave, the apex and leading and trailing edges of a translational wave all migrate downstream simultaneously (Figure 2.4). In order for this to occur, the trailing edge of the wave, which consists of a gentler slope than that of the channel bed, must undergo erosion. This condition is met when the wave is composed of finer material than that found in the channel, allowing it to be transported downstream easily (Lisle et al. 2001).

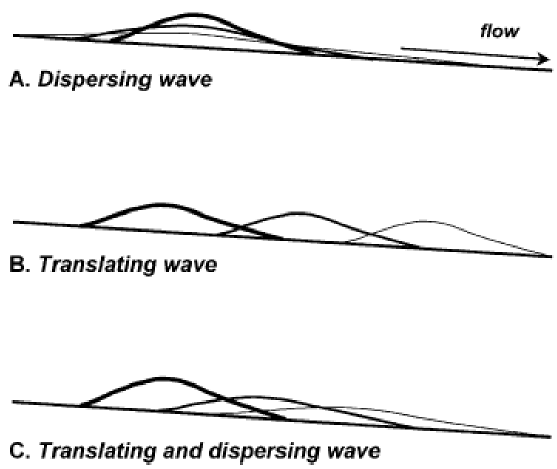


Figure 2.4 Evolution of a dispersive, translating, and translating and dispersive wave through time, with darker lines indicating earlier stages (Sutherland et al. 2002).

The rate at which a slug will move downstream is dependent on the method of migration (dispersion vs. translation) and the flood regime of the channel as this will determine if the river is capable of transporting sediment (Nicholas et al. 1995). However, it is also dependent on the grain size of both the sediment composing the wave and the channel bed (Cui et al. 2003). When the median particle size of the wave is coarser than that in the stream, the downstream sediment, along with the sediment underneath the wave will be armoured due to the coarser sediment input which acts as a protective layer. This renders the wave less mobile because a greater critical shear stress is needed to initiate sediment transport, thus reducing the rate of downstream migration (Cui et al. 2003).

As waves migrate downstream, their amplitude decreases, thus fewer sediment peaks are measureable. This is because greater amounts of sediment are stored in the lower sedimentation reaches of the channel. As the transport zones of rivers generally have narrower channels and higher flow velocities than depositional zones, transport rates will be higher in these reaches. As waves pass through constrictions, most of the sediment entering the zone will be transported through quickly. However, when the channel widens in the sedimentation zone, and transport rates decrease, a portion of the sediment composing the wave will become stored in the channel, resulting in a dampened wave (Nicholas et al. 1995).

2.2.5.2 Impact of sediment waves on river channels

Regardless of whether sediment waves are formed intrinsically or extrinsically, their passage through river systems will physically alter the channel (Nicholas et al. 1995). Generally controlled by the scale of the wave, larger waves typically have a greater impact and cause more changes to the channel than smaller ones (Kasai 2006). Such changes can range from modifications to the bedform at the scale of bar assemblages (Kasai 2006) to alterations of the valley floor configuration if the wave is large enough (Nicholas et al. 1995). However, this may not always be the case. In some instances, the magnitude of the impacts caused by sediment waves can be larger than expected solely based on the wave's size. This occurs when antecedent conditions are such that would favour enhanced changes (Nicholas et al. 1995). Additionally, the process of downstream movement of waves will impact how they alter the channel (Sutherland et al. 2002). The most common changes that a river will be subject to are aggradation and degradation of the channel bed. The former can be identified by an increase in the intensity of braid formations and channel infilling through sedimentation on the bed, while the latter is reflected by a decrease in channel braids, and an increase in bed scouring (Nicholas et al. 1995). If there is a high supply of sediment, and substantial aggradation, then the changes in the channel will be more pronounced, and in some instances, a single channel can be transformed into a braided channel due to the formation of bars (Schumm et al. 1987). If the sediment input is limited, aggradation will not occur to such a large degree and bar formation will be limited (Nicholas et al. 1995). This process of aggradation is important in terms of the possible impacts on the riverbed because if the level of sediment accumulation is maintained for long periods of time, the creation of barriers can occur (Cui et al. 2003), resulting in longitudinal disconnectivity of sediment. Under such circumstances, the magnitude of downstream geomorphic changes is limited until the barrier is breached and connectivity is restored (Fryirs et al. 2007).

If sediment wave movement is limited to dispersion, the downstream sedimentary impacts and channel recovery will be prolonged because only the leading edge migrates downwards. As such, downstream reaches are subject to the wave until it has completely passed through the system. However, this situation results in a weakened overall impact at any one location (Sutherland et al. 2002). If the leading edge of a dispersive wave is not moved through the system at a greater rate than upstream sediment transport, it could become a barrier to the flow of sediment by forcing sediment deposition behind the trailing edge, resulting in the accumulation of upstream sedimentation and the effective upward movement of the wave in the channel (Lisle et al. 2001). Since the leading and trailing edges of translational waves move as a unit, once they have passed, the channel and related bedforms will gradually recover to pre-wave conditions behind the wave (Lisle et al. 2001). However, the alteration caused to rivers from purely dispersive waves will be more localized in space, and will mainly affect those reaches, both upstream and downstream, that are nearest to the input of sediment responsible for the wave (Lisle et al. 2001).

Irrespective of whether waves are translational, dispersive, or composed of coarser or finer material than the average bedload in the channel, they will have an impact on the river. These impacts are generally an overall shallowing of pools, a rise in the channel bed, and banks that are low and poorly defined (Griffiths 1979, Madej & Ozaki 1996).

2.3 Arctic hydrology

2.3.1 Snow hydrology

In Arctic and other northern environments, the accumulation and redistribution of snow through drifting, sublimation, and accumulation are a primary element of the hydrological cycle (Church 1974, Woo 1983). During the winter months, the surface energy balance is strongly negative owing to continuous heat loss due to low air temperatures. This results in a frozen surface where temperatures are further decreased due to conductive heat loss (Woo 1983). It is during the period between the end of fall and spring that the majority of the annual precipitation falls in the form of snow. In Arctic environments where vegetation does not dominate the landscape, snow is unevenly distributed throughout the catchment, as the lack of vegetation allows for drifting of snow to occur (Woo 1983). Snow drifting results in snow-filled depressions and water tracks, representing zones of snow accumulation, while shallow snowpacks occur in exposed areas (Kane et al. 1991). This uneven distribution of snow is extremely important for the hydrological cycle because it results in a higher percentage of runoff since snow accumulates nearer to or in channels, compared to a more uniform snow distribution (Kane et al. 1991).

In late spring, early summer, net radiation at the surface of the snow remains negative. However, with the arrival of longer periods of sunlight energy, air temperatures slowly begin to rise, allowing for conductive heat gain into the snow from the warmer ambient air (Woo 1983). It is usually in late spring that snow ablation occurs (Kane et al. 1991). Not only does the incoming solar radiation help in warming the air, but the direction of the dominant wind plays a very important role in ablation. In a study of the Imnavait watershed in Northern Alaska, Kane et al. (1991) found that when the predominant winds were northerly, air temperatures tended to be lower, so the incoming shortwave radiation was used for the sensible heating of the atmosphere, thus slowing the snowmelt process. Conversely, when the dominant winds were southerly, air temperatures were warmer, so these warmer temperatures, in addition to the incoming solar radiation, could be directed towards enhanced snow melt. The heat that is gained or lost at the surface of the snowpack can be represented as an energy balance equation:

$$Q_M = Q^* + Q_H + Q_E + Q_R \qquad (2.5)$$

where

$$Q^* = (1\text{-}\alpha)\, K\!\downarrow + L^* \qquad (2.6)$$

and Q_M is the energy that either enters or leaves the snow surface (energy available for melt), Q^* is net radiation, Q_H and Q_E are the sensible and latent heat fluxes respectively, Q_R is the energy that is added to the snowpack by cause of precipitation, α is the surface albedo, $K\downarrow$ is the incoming shortwave radiation, and L^* is the net shortwave radiation (Woo 1983). By the end of

May, because temperatures remain above freezing for longer periods of time, the sensible heat flux is directed towards the snowpack and contributes energy for melt (Woo 1983), thus beginning the first of the three phases of snowmelt, “warming” (Dingman 1994). The snowpack progressively gains enough heat until it becomes isothermal at 0°C (Dingman 1994). Continued absorption of energy further warms the snowpack until it can no longer retain the water in its pores. At this point in the second phase of the snowmelt process, the snowpack is termed “ripe” (Dingman 1994). The last step in snowpack melt is “output”. During this phase, additional inputs of energy result in the percolation of water through the snowpack, and the release of water as runoff (Dingman 1994). Once snow has begun to melt, there are two ways by which melt water can percolate into the snowpack. The first is by following the advance of a wetting front, and the second by following the “vertical flow fingers” ahead of the slowly advancing wetting front (Marsh 1983). If the percolation of water through the snowpack is delayed due to changes in the properties of the snow, then the meltwater could refreeze in the snowpack, thereby releasing latent heat to the surrounding snow. This is an important and effective mechanism of warming the snowpack (Woo 1983). If the percolating meltwater reaches the base of the snowpack, more often than not it will refreeze because temperatures at the base of the snowpack have not had enough time to warm. This can result in the formation of basal ice and delays the release of runoff until sufficient energy accumulates for complete snowmelt to occur (Kane et al. 1991).

2.3.2 Channel development

When the snowpack within the river valley is saturated, channel development and flow can begin (Woo & Sauriol 1980). Meltwater initially moves by three methods; subsurface flow within the snow-filled valley, sheet flow on the surface of the snow, and slush flow. Subsurface flow in the snowpack will occur when there is a sharp rise of the hydrostatic water level in the snow (Figure 2.5). When this water level rises above the snowpack, sheet flow on the snow surface commences. As the water above the snowpack progresses downstream, it slowly and continually incises into the snowpack, thus shifting the newly channelized flow laterally and vertically until reaching the channel bed (Woo & Sauriol 1980). This channelized flow may not be continuous as saturation of large areas behind snowdrifts can result in ponding. The pond will expand until the snow dam blocking the pond disappears, often creating an outburst of water, further accelerating the downstream breakup of snow (Woo & Sauriol 1980). If there is deep snow accumulation in the valley, hydrostatic water levels may not reach the surface, in which case snow tunnels can form, allowing for the channelized sub-surface flow of water through them (Woo & Sauriol 1980). The initial location of the channelized flow rarely corresponds to its end-of-season location because of continual undercutting of snow and channel banks (Woo & Sauriol 1980).

As snowmelt progresses, there will be a gradual increase in the hydrograph associated with the spring freshet (Kane et al. 1991) which generates the annual peak flow. This brief period of peak flow is a defining characteristic of the runoff regime of Arctic and high latitude rivers (Woo et al. 2008). In some cases, this short spring freshet can account for up to 90% of the annual runoff (McCann et al. 1971). While annual peak flows for large Arctic rivers do tend to take place in the spring because of snowmelt, this may not always be the case. Large summer precipitation events, following peak runoff, can cause subsequent peaks in river discharge (French 2007, Dugan et al. 2009).

This figure has been removed due to copyright restrictions.

Figure 2.5 Response of the hydrostatic water level in snow over a two week period and the resultant development of the channel (Woo & Sauriol 1980).

2.3.3 Permafrost hydrology

Permafrost is any ground material that remains at temperatures below freezing for at least two consecutive years (Rowland et al. 2010). It covers approximately 80% of the Arctic Ocean drainage basin, and 25% of the Northern Hemisphere's land surface (Frey & McClelland 2009). The surface energy balance is critical in determining the geographical extent and presence of permafrost soils. Various classes of permafrost exist depending on its aerial extent and thickness, ranging from isolated, sporadic, discontinuous, and continuous patches with aerial extents and thicknesses of 1-10, 10-50, 50-90 and 90-100% and <10, 10-25, 25-100, and 100-800 m respectively (Frey & McClelland 2009). The role that permafrost, especially continuous permafrost, plays in the hydrological cycle of Arctic rivers and catchments is extremely important because it restricts the deep infiltration of water into the soil by acting as a semi-permanent aquitard (Carey & Pomeroy 2009). This defines the basin's contributing area because there will only be lateral flow of water in regions underlain by permafrost (Carey & Quinton 2004). By restricting deep infiltration of water into the soil, there is generally increased runoff and soil moisture in regions underlain by permafrost (Woo et al. 2008). Permafrost soils therefore have the opposite effect on regional hydrology than areas with seasonally frozen soils which tend to have well-drained soils, allowing for deeper percolation of water. This results in a water balance dominated by vertical inflows of water as opposed to lateral ones as is the case with permafrost dominated areas (Woo et al. 2008).

2.3.3.1 Continuous permafrost

The amount of water that can infiltrate into the ground is largely dependent on the type of permafrost present in the area, whereby an inverse relationship is recognized between permafrost classification and infiltration (Carey & Pomeroy 2009). In areas of continuous permafrost, hydrologic activity is limited to either the seasonally-thawed active layer or the ground surface (Woo 1983). The hydrographs of areas with continuous permafrost are characterised by rapid response times to snowmelt because of the distribution of highly saturated areas within the watershed near the lateral paths that the water follows, followed by a slow, long recession period because of the low water transmission associated with the deepened active layer (McNamara et al. 1997). As the summer season progresses, the thawing of the active layer allows for some transmission of sub-surface flow and water storage. However, the flow of deep water is limited by the presence of permafrost (Woo 1983, Carey & Pomeroy 2009) which increases the near surface water tables, allowing for the frequent rising of the water table above ground (Woo et al. 2008).

2.3.4 Runoff regimes

Hydrology in northern environments is strongly influenced by the presence of permafrost, seasonally frozen ground, and the thermal conditions associated with these higher latitudes. Specifically, in northern regions, runoff is controlled by the timing of snowmelt and the contribution of baseflow in the summer season (Woo 1983). Subarctic nival, arctic nival, proglacial, and muskeg regimes are four primary runoff regimes identified by Church (1974) for northern rivers. Each regime type is characterised by the timing and source of runoff, and has a distinct hydrograph (Figure 2.6).

The subarctic nival regime is characterised by high discharge in the spring caused by snowmelt, followed by a period of reduced summer flows which may be interrupted by rainfall events. Depending on latitude, flow can either completely stop in the winter given a frozen watershed, or if groundwater seepage is present, a limited baseflow can be maintained. The latter condition is mainly found in large northward draining rivers that have headwaters in areas of discontinuous permafrost (Church 1974).

Similar to the subarctic nival regime, rivers exhibiting arctic nival regimes generally experience one period of high runoff in the spring associated with the spring freshet. This period of high discharge is usually the largest of the season. However in small watersheds, late summer precipitation events have the ability to generate greater peak discharges than that observed from snowmelt. Rivers with an arctic nival regime drain watersheds with continuous permafrost and winter groundwater is not present to maintain baseflow (Church 1974).

Both subarctic and arctic nival regimes have a distinct spring freshet that can clearly be identified on a hydrograph. However, such a freshet is not typically observed in proglacial regimes. Under such regimes, glaciers or permanent snow continually melt through the summer when conditions favour it. This results in rivers with increasing discharges until late summer, at which point discharge decreases as glacial melt no longer occurs in catchment (Church 1974, French 2007).

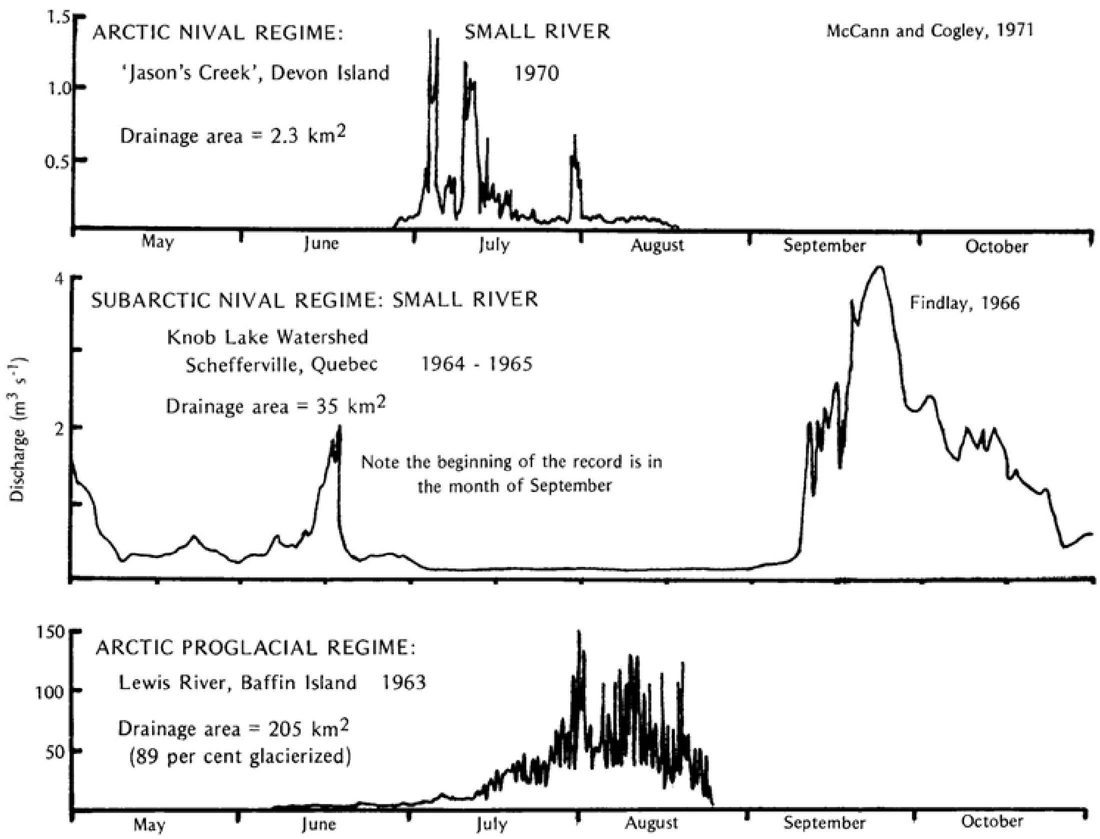


Figure 2.6 Typical hydrographs associated with regimes found in northern rivers (Modified from Church 1974).

2.4 Permafrost disturbances

Under a warming climate regime, permafrost slope disturbances such as active layer detachments (ALD) and retrogressive thaw slumps (RTS) can occur where ground ice is prevalent in the near-surface above the permafrost (Lewkowicz 1990). Impact studies in regions where ALD have been documented indicate that they can modify the landscape in important ways, and act as important geomorphic agents for sediment transfer. The following sections describe these disturbances in the context of catchment sediment dynamics.

2.4.1 Active layer detachments

2.4.1.1 Development

Thermokarst features can occur in many forms, one of which being active layer detachments. This type of shallow permafrost slope disturbance can develop on slopes ranging from very gentle to moderate (Harris & Lewkowicz 1993, Lewkowicz & Harris 2005) and are believed to occur as a result of high porewater pressures and low effective shear stresses which are triggered by the rapid melt of ground ice located at the base of the active layer. This in turn facilitates the downslope movement of the slide (Lewkowicz 2007). In order for a rapid melt of ground ice to occur, a period of above average temperatures is needed to support a positive subsurface energy balance (Lewkowicz 2007, Bowden et al. 2008). Other triggers to their development include forest fires (Harry & MacInnes 1988 in Lewkowicz 1990) and summer rainfall events (Lamoureux & Lafrenière 2009) as these can affect the strength and moisture conditions of the ground. Studies documenting ALD indicate that the timeframe associated with their formation can be quite variable. Some develop within minutes to hours, and form in one sliding movement, while others form over a period of days while advancing at sustained rates of a few meters per hour (e.g., 9 m/hr) and enlarging as it slides downslope (Lewkowicz 2007). There are two methods by which an ALD can slide; compression and translation (Figure 2.7), although many slides undergo a combination of the two (Lewkowicz 1990).

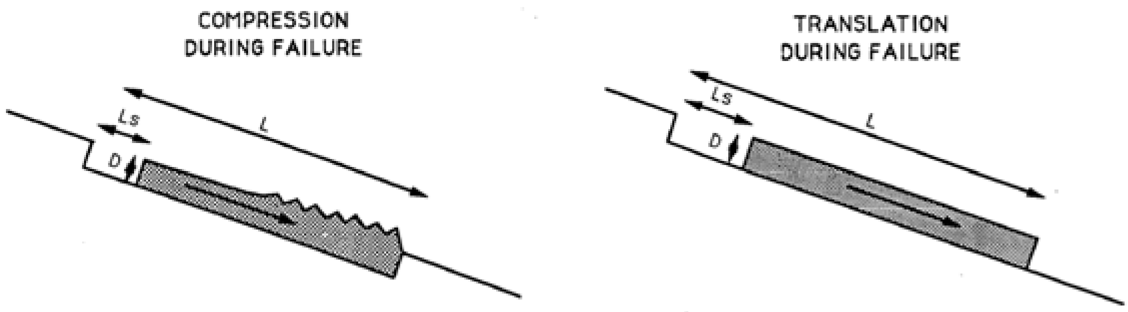


Figure 2.7 Methods of movement taken by active layer detachment slides (Lewkowicz 1990).

2.4.1.2 Morphology

ALD are composed of three zones. These are 1) the scar zone, 2) the extension or track zone, and 3) the compression zone or toe. However, it is also possible for ALD to have a fourth zone which is made up of disturbed material alongside and ahead of the main slide (Lewkowicz 1990).

The scar zone is the region delineating the undisturbed soil from the ALD. It represents the abrupt division line below which the disturbed ground will move downhill. Prior to, and early in the development of an ALD, the scar zone is smooth. However, as it expands, this zone can undergo erosion and surface ground material will be removed in blocks, resulting in further degradation of the slope (Lewkowicz 1990). The extension or track zone represents the path that the ALD travels down. The base of the track zone contains ridges created by the movement of individual blocks of disturbed material. Lastly the toe of the ALD is the farthest accumulation point of the slide. Because ALD can expand outwards as they slide, the toe is often wider than the extension and scar zone (Lewkowicz 1990).

2.4.2 Retrogressive thaw slumps: development and morphology

Retrogressive thaw slumps, a second major type of thermokarst feature, develop mainly in response to coastal or bank erosion (Lewkowicz 1987). Other factors can also trigger their initiation, and these include lateral stream erosion, wave erosion, ice push (French 2007), and active layer detachments (Lewkowicz 1990). These disturbances can expand inland rapidly, and perpendicular to the coastline or river bank (Lewkowicz 1987). They will form if the rate of ground ice melt exceeds the erosion rate at the coast or river bed (Lewkowicz 1987).

Retrogressive thaw slumps are composed of three main sections (Figure 2.8): 1) a vertical or near-vertical headwall containing the active layer and frozen ground material, 2) an ablating ice face which becomes smaller because of insolation and sensible heat fluxes, and is angled between 20 and 50°, and 3) a slump floor or mudflow which covers the underlying ice-rich ground (Lewkowicz 1987, Lantuit & Pollard 2008).

2.4.3 Active layer detachments

Active layer detachments have been recognized as agents in changing the physical characteristics of watersheds and the processes within them (Lewkowicz 2007). Because of this, when they are hydrologically connected to river systems they can have important impacts on the sediment dynamics of rivers. Those that form alongside rivers typically have compact forms and can be identified morphologically by their curved scar zone and straight edge at the bottom of the compression zone (Lewkowicz & Harris 2005).

Few studies surround the impact that ALD have on rivers and their sediment dynamics; however those that do indicate that they can have important consequences. The sliding of ALD into the main channel network can substantially restrict flow by narrowing channel width (Lamoureux & Lafrenière 2009), thus creating a temporary barrier to the flow of water and sediment. ALD can also increase sediment loading to rivers if connected to the main channel through erosion and by sediment-rich tributaries flowing through the disturbances and discharging into the main channel. A few days following the initial formation of ALD at Cape Bounty in the summer of 2007, Lamoureux and Lafrenière (2009) measured sharp increases in river turbidity downstream of the disturbances. The changes in turbidity lasted a few hours. The newly formed disturbances represented approximately 2% of the watershed, but were responsible for an 18% increase in seasonal sediment yield, thus indicating the important connection between suspended sediment transport and active layer detachment slides.

2.4.4 Retrogressive thaw slumps

Similar to ALD, RTS have also been shown to increase the sediment loading to fluvial environments in the short-term following their development (Bowden et al. 2008). Estimates of the total amount of eroded sediment from RTS on Herschel Island indicate that volumes can exceed 355000 m^3 (Lantuit & Pollard 2005). Because of the close proximity of most RTS to coastal ecosystems, the former can be significant agents in the supply of sediment (Lantuit & Pollard 2005). The total volume of sediment supplied from RTS can be highly variable depending on the size of the slumps, even if they are in close proximity to one another, indicating the variability associated with such disturbances (Lantuit & Pollard 2005).

2.5 Conclusion

If current trends of increasing temperatures and more frequent summer precipitation events in the Canadian High Arctic continue, there could be important consequences on the geographical extent and dynamics of permafrost. In particular, if permafrost degradation results in the formation of ALD or RTS and the disturbances are hydrologically connected to a river system, they can contribute substantial amounts of sediment to the channel. This localized sediment input holds the potential to generate sediment waves or slugs with lasting effects on the downstream fluvial system. ALD can additionally become barriers to the flow of sediment, thereby reducing the sediment delivery ratio of the catchment. It is therefore important to understand how Arctic rivers respond to such changes and how these changes in turn affect river patterns.

Chapter 3

Sediment storage dynamics and channel response to recent permafrost disturbances in a High Arctic river, Cape Bounty, Nunavut

3.1 Abstract

A sediment budget approach was used to investigate suspended sediment transport dynamics in the 8 km^2 West River catchment at the Cape Bounty Arctic Watershed Observatory (CBAWO), Melville Island, Nunavut. Research was carried out in an effort to determine the longer term impacts and response of recent (2007-8) active layer detachments (ALD) on river systems. In 2010, gauged ALD contributed 4.7% of the measured sediment yield, a decrease of 13.3% from 2007 when they initially formed. This decrease indicates that while ALD continue to supply sediment to the main river, the impact they have on sediment fluxes and hence the sediment budget diminishes with time. Sediment budget analysis shows that in response to the sediment originating from recent ALD, the West River progressively stored more sediment throughout the season. Results indicate that connectivity and the sediment delivery ratio within the system also decreased with time. Under baseflow conditions, up to 85% of sediment inflows went into channel storage in contrast to 26% under higher flow conditions early in the season. Spatially, the stored sediment was preferentially deposited within the channel, with coarser material deposited in the upper reaches and finer material in the lower reaches. Over time, the transported and stored sediment became progressively finer. This sediment fining indicates the importance of river competence and wetted perimeter in sediment transport as larger sediment was entrained earlier in the season under higher flow conditions when the sediment was

accessible, and finer sediment transported later in the season due to decreased competence and reducing access to sediment stores. The sediment stored in the channel mitigates the effects that disturbances have on the sediment budget, especially sediment stored towards the end of the season as it does not contribute to sediment yield and may prove to be an important source of sediment in future years under late summer rainfall events.

3.2 Introduction

Catchment-scale sediment budgets have been widely applied to rivers worldwide as a means of determining the degree of sediment connectivity within fluvial systems. They are also used as a method to quantify how fluvial landscapes evolve over time and to evaluate how external disturbances affect channel sediment dynamics (Hooke 2003, Fryirs et al. 2007). Sediment budgets allow for areas of sediment erosion, transport, and deposition to be identified, thereby enhancing our understanding of sediment delivery and connectivity within watersheds (Fryirs et al. 2007). Understanding fluxes of sediment within the channel (as opposed to a sediment yield for the basin as a whole) is also important as suspended sediment can act as a substrate on which contaminants and nutrients are transported through the river system. Sediment budgets, therefore, indirectly provide insight into the fate of nutrients and contaminants associated with suspended sediment (Walling 1988, Walling et al. 2003). Thus, sediment budgets can be used as a metric of catchment response to, for example, environmental disturbance resulting from a shifting climatic regime (Orwin et al. 2010).

During the last few decades, Arctic regions have been warming at nearly twice the average global rate, and are projected to continue increasing by another 4-7 °C by 2100 if current trends continue (ACIA 2005). This increase has raised concerns regarding the stability of permafrost which is predicted to degrade with continued warming (Bowden et al. 2008). In the early stages of permafrost thawing, there could be an increase in the occurrence of thermokarst processes such as active layer detachments (ALD) and retrogressive thaw slumps (RTS) (Woo & McCann 1994, Bowden et al. 2008). Many studies have already documented the occurrence of such thermokarst processes in response to increasing temperatures and summer precipitation events (e.g., Kokelj & Lewkowicz 1998, Bowden et al. 2008, Lamoureux & Lafrenière 2009). If water and sediment flow from these disturbances connects into larger channel systems, they have the potential to significantly alter the sediment loading to the larger channels through the introduction of sediment-rich runoff or by the direct contribution of soil (Lamoureux & Lafrenière 2009). This enhanced loading would in turn alter the sediment delivery ratio (SDR), the proportion of sediment discharged at the river mouth in proportion to that which is eroded in the catchment. However, this rapid increase in sediment delivery to the channel could act as a barrier to sediment transport through a series of potential storage mechanisms (Fryirs et al. 2007). These changes to the SDR can potentially be evaluated with the application of a sediment budget approach (Walling 1983).

An important component of any sediment budget is to determine where sediment originates and the type of sediment being transported downstream. Sediment sources can be difficult to monitor as they change seasonally. However, suspended sediment concentration – discharge hysteresis relationships can provide clarification into sediment availability and accessibility by indicating whether transported sediment originates from nearby, or upstream sources (Klein 1984, Lawler et al. 2006), and whether these sources are located within the channel system (Lawler et al. 2006). Due to their ability to indicate supply-limited and discharge-limited river conditions, hysteresis relationships are important tools towards the better understanding of sediment budgets. Additionally, sediment transport dynamics can be inferred from particle size analysis which reveals the type of sediment being transported throughout the season. Due to the correlation between the transport of a range of particle sizes and flow velocities, information pertaining to particle size can be used to understand the timing of sediment transport (Gordon et al. 1994, Knighton 1998).

Because there is a high potential of erosion and hence increased sediment delivery to rivers in catchments affected by ALD (Kokelj & Lewkowicz 1998, Lamoureux & Lafrenière 2009), this paper reports a study of the in-channel transport and catchment connectivity of sediment transport in a High Arctic river that has been subjected to recent permafrost disturbance. A key objective of the study was to identify whether localized ALD constitute a significant source of sediment to the overall fluvial channel system and how channel sediment dynamics have responded. The main objective of this study was, therefore, to determine how sediment transport and fluxes in the West River of the Cape Bounty Arctic Watershed Observatory (CBAWO) on Melville Island, Nunavut have responded to extensive initiation of ALD that occurred in 2007-8. Although sediment budgets should include measurements of bedload, dissolved and suspended load, this study will focus solely on the suspended sediment component. Hence, references to fine sediment will indicate sediment ranging in size from clay to sand, and coarser fractions >63 μm are excluded.

3.3 Study area

Research was carried out at the CBAWO located on the south-central coast of Melville Island (74°55’N, 109°35’W) in the western Canadian High Arctic (Figure 3.1). CBAWO was selected for this study because of the long-term, multi-disciplinary monitoring of the landscape and hydrological system that began in 2003. The specific focus of this study was the West River, located in an 8.0 km^2, unglacierized catchment underlain by continuous permafrost (Lamoureux et al. 2006). The West River has a moderate gradient profile with a mean slope of 0.015 that decreases slightly downstream (Figure 3.2).

In the catchment, gently rolling terrain rises to maximum elevations of 100 m above sea level (a.s.l.). Seasonal active layer depth varies with soil type, however, will on average reach depths between 50 to 100 cm (Lamoureux et al. 2006).

Typical of a polar climate, CBAWO experiences long, cold winters (September to May), and short, cool summers (June-August). Mean June and July air temperatures from 2008 to 2010 were 3.3 and 7.3 °C at the MainMet meteorological station, respectively (Figure 3.1). Mean winter air temperatures from December to February and summer temperatures from June to August for the period from 1971-2000, were -32.6 and 1.7°C, respectively at Mould Bay (Environment Canada 2011), a site located approximately 350 km west of CBAWO (Figure 3.1). The dominant form of precipitation is snow, which undergoes extensive redistribution by wind drifting and results in snow-filled depressions in low-lying areas and channels, and a shallow snowpack in exposed terrain (Woo 1983). Mean winter snowfall from 1971-2000 at Mould Bay is 84 cm (McDonald & Lamoureux 2009). Snowmelt usually begins in early to mid June. Streamflow begins a few days later and rises rapidly to peak flow in mid to late June.

Discharge then gradually recedes to baseflow in mid-July and continues until freeze up in late August or early September (Lamoureux & Lafrenière 2009). This region of the Arctic is characterised by infrequent heavy summer rainfall events (10+ mm/day).

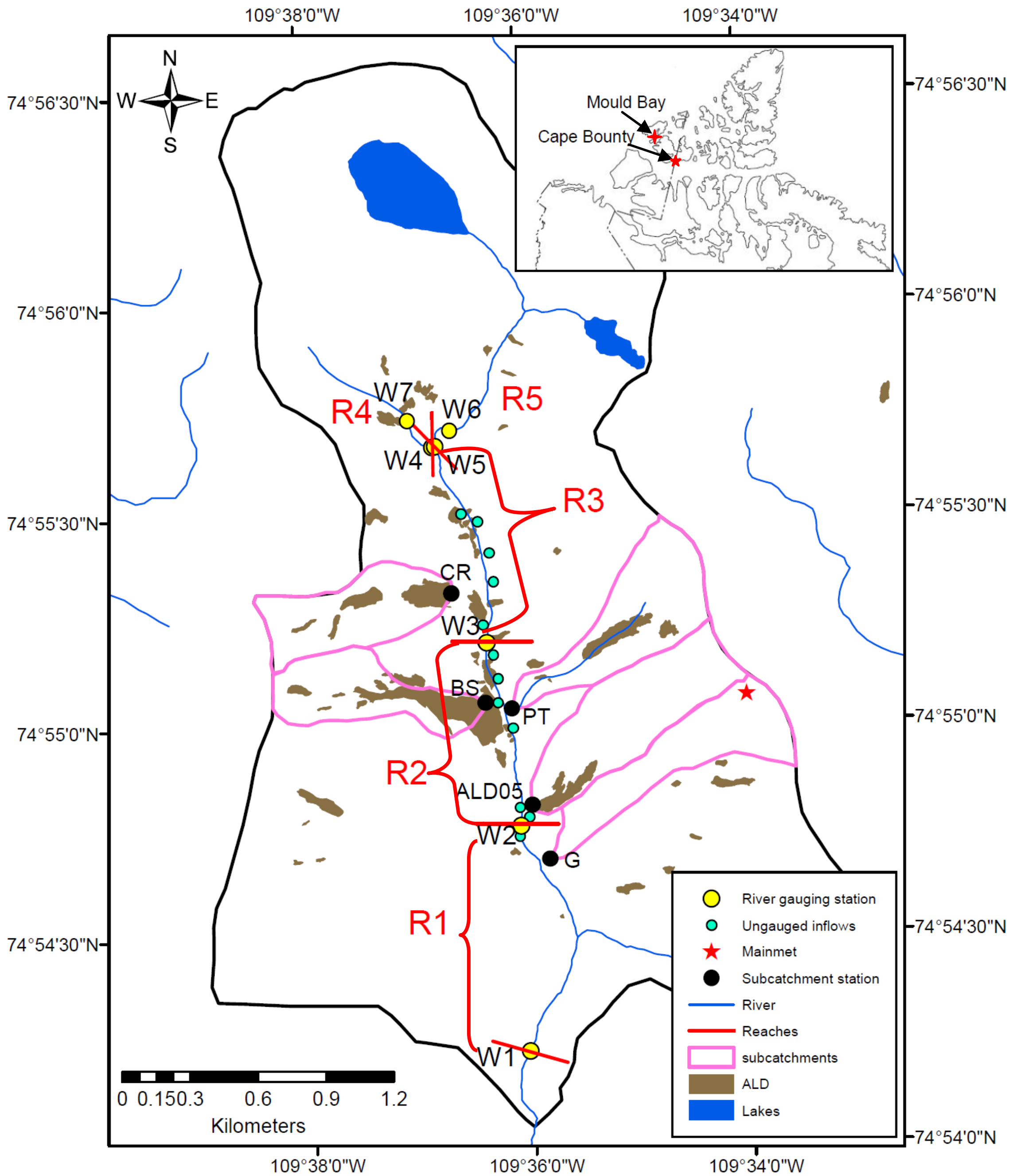


Figure 3.1 West watershed at the Cape Bounty Arctic Watershed Observatory (CBAWO), Melville Island, Nunavut Canada. Inset map indicates location of CBAWO in the Canadian Arctic Archipelago (inset basemap source: Natural Resources Canada 2001).

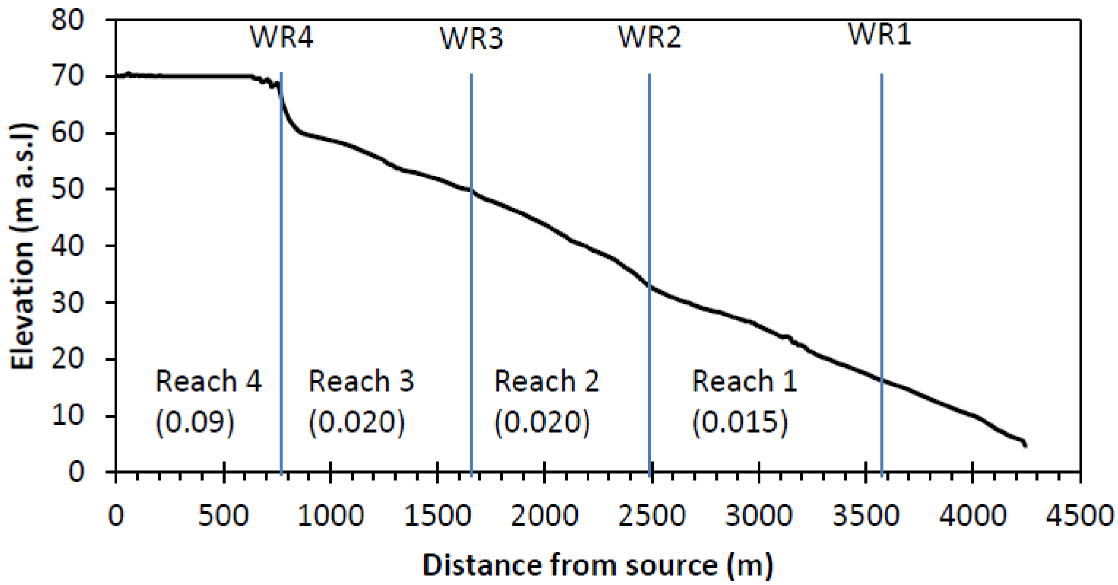


Figure 3.2 Longitudinal profile of the West River at the CBAWO. Numbers in parentheses indicate the slope for the given reach. Note that station WR1 is located approximately 700 m upstream from the outlet of the West River in a large lake.

An extended period of exceptionally warm air temperatures occurred during the summer of 2007 (Lamoureux & Lafrenière 2009) when maximum July air temperatures reached 21.3 °C, and total rainfall was 29.6 mm over the summer months (Dugan et al. 2009). This anomalous weather resulted in the formation of extensive ALD, which at the time covered approximately 1.9% of the watershed (Lamoureux & Lafrenière 2009). Continued disturbances in 2008 and 2009 have resulted in the disturbance of 2.9% of the West River watershed, primarily through new and expanded activity, of which several connect hydrologically through tributaries to the main river.

3.4 Methods

3.4.1 Meteorology

A meteorological station (MainMet) was established at the boundary of the West watershed in 2003. This location was chosen as it is central to the East River as well. It was equipped to measure air temperature, precipitation, relative humidity, net radiation, and wind speed and direction (Figure 3.1). However, only the results from air temperature and precipitation will be discussed in this study. Air temperature was measured with a shielded Humirel HTM2500 sensor (3% accuracy) located 1.5 m above the ground and precipitation was measured using a Davis Industrial Tipping Bucket (0.2 mm tip), both at 10 minute intervals and hourly means or totals were logged with a Unidata Prologger.

3.4.2 Hydrology

A sediment budget approach was applied to the West River catchment by installing five gauging stations along the West River, and an additional five in major subcatchments draining into the river (Figure 3.1). This sediment budget focused solely on suspended sediment, and was used to measure the influx and efflux of suspended sediment at each reach in order to isolate the effect of ALD, the tributaries running through them, and active erosion of the bed and banks in order to determine if the reaches were sources or sinks of sediment. Each station was equipped with an Onset HOBO U20 water level data logger (0.05% FS raw pressure) logged at ten minute intervals. At each station along the river (WR2-WR5), cross-sectional area and velocity was measured twice daily, at approximate high and low flow during the peak discharge season, and once daily afterward, using a Swoffer 2100 current velocity meter (1% accuracy). Measurements were taken at 0.50 m width increments at 60% stream depth from the surface. Additionally, cross-sectional velocity at WR1 was measured at 30 minute intervals over a 24 hour period coincident with the highest discharge of the season (June 17-18), and again for a six hour period during low flow conditions (July 25). Discharge (Q) was then calculated by applying the velocity-area method,

$$Q = \sum(\Delta x * \left(\frac{h_1+h_2}{2}\right) * \left(\frac{v_1+v_2}{2}\right)) \tag{3.1}$$

where Δx is the interval between two consecutive measuring points (m), h represents water depth (m), and v is the measured flow velocity (m/s). Rating curves (r^2 ranged between 0.75 and 0.99) were constructed for each station using a polynomial model to extrapolate hourly discharge using the recorded stage measurements (Table 3.1). The subcatchment stations were gauged with 8" cutthroat flumes and rated with standard rating tables (ASCE 1974) with stage measured using Onset HOBO U20 water level loggers.

Table 3.1 Rating curves and respective r^2 values at the five gauging stations along the West River of the CBAWO, where Q is discharge (m^3/s), and h is stage (m). WR5 has two rating curves because of a change to station WR5-b location on June 27.

Station	Rating Curve	r^2	n
WR5-b	$Q = 4.9534h^2 + 1.0877h - 0.0726$	0.87	26
WR5-a	$Q = 6.2579h^2 - 1.0224h + 0.022$	0.96	12
WR4	$Q = 11.042h^2 - 0.5484h + 0.0035$	0.75	37
WR3	$Q = 10.685h^2 - 0.9463h + 0.0197$	0.90	39
WR2	$Q = 16.748h^2 - 2.5781h + 0.1179$	0.95	35
WR1	$Q = 26.836h^2 - 3.237h + 0.1157$	0.99	40

3.4.3 Suspended sediment concentration and yield

At the main gauging station (WR1), water samples for suspended sediment concentration (SSC) were collected every three hours using an ISCO 3700 pump sampler with the water intake point located 10 cm above the channel bed. At gauging stations WR2-WR5 and the subcatchment stations, manual depth-integrated water samples were collected twice daily, once in the morning (approximate low flow) and again in the late afternoon (high flow) from the beginning of runoff to the end of June, and once daily afterwards during daily high flow for the remainder of the measurement period. During rainfall events, water samples were collected multiple times daily in order to identify any changes in SSC that may be caused by rainfall runoff.

Water samples of known volume were immediately vacuum filtered through 1.0 µm pre-tared glass fibre filters, and oven-dried at 45°C in the laboratory until weights stabilized between two consecutive measurements to calculate SSC (mg/l). A robust cubic spline function was applied to the measured SSC points to estimate hourly values when twice daily samples were available. When twice-daily sampling was not available, the SSC value of the daily measurement was assumed to represent the hourly average for the day. Suspended sediment yield was determined as the cumulative product of SSC and discharge.

In addition to the hydrometric stations, 13 additional inflows of water into the West River were identified (Figure 3.1). These occurred as overland flow which could not be constrained and as a result, discharge measurements are not available. However, water samples were collected for SSC analysis at various inflows and times throughout the season until the sources dried out. Although associated discharge measurements were not taken, SSC samples from stations WR6 and WR7, located upstream of stations WR5 and WR4 respectively (Figure 3.1, 3.3) were collected approximately daily in the evening over a two week period. The SSC values from WR4 and 5 were compared to those from WR7 and 6 respectively in order to measure the effect that a large section of unconsolidated sediment along both sides of the confluence had on the West River in terms of sediment concentration.

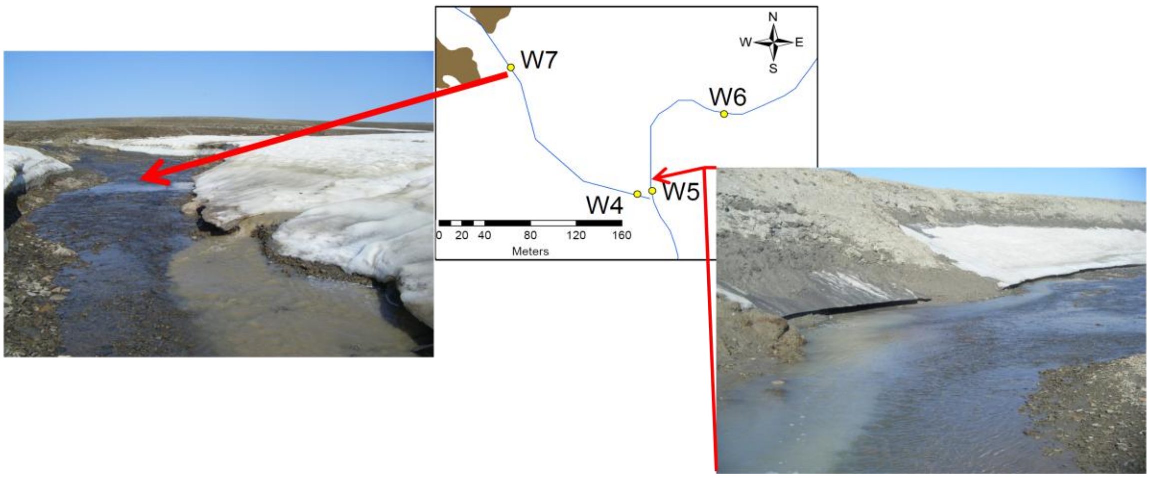


Figure 3.3 Location of sampling stations WR6 and WR7, both upstream from a large exposure of unconsolidated sediment above WR5 (right image) and a pond of turbid water (left image) respectively.

For sediment budget purposes, the net storage or release associated with each reach was determined by a comparison of the amount of suspended sediment entering and exiting the reach. Budgets could be calculated for WR1, 2 and 3 (Figure 3.1), but all other reaches and inflows were limited to sediment outflow determinations only. The budget was calculated for three time periods, low flow 1 and 2 (June 22 – June 30, July 1 – July 9 respectively), and baseflow (July 10 – July 28) which were determined based on the WR1 hydrograph. Budget work prior to June 22 is not possible as there was some delay in the installation of gauging stations due to excessive amounts of channel snow and difficulties in locating initial flow paths. Reaches acting as net sources of sediment to the main channel were found to have more sediment exiting the reach than entering it, whereas reaches that were net stores of sediment had a greater sediment inflow than outflow.

3.4.4 Suspended sediment yield-discharge hysteresis

Suspended sediment – discharge hysteresis relationships are used as a method of quantifying the response of sediment transport to variations in discharge during an isolated event, or season, by means of a hysteresis index (Lawler et al. 2006). Suspended sediment yield – discharge hysteresis relationships were used to determine the direction and magnitude of diurnal and period scale events. This measure was calculated using the method of Lawler et al. (2006), whereby the hysteresis index at the mid flow range (HI_{mid}) is calculated by determining the mid-point discharge (Q_{mid}) according to

$$Q_{mid} = k(Q_{max} – Q_{min}) + Q_{min}, \tag{3.2}$$

where k is a constant between 0 and 1 that represents the position at which the extent of the hysteresis loop is evaluated in relation to the flow range, and Q_{max} and Q_{min} are the maximum and minimum discharge values, respectively. The two suspended sediment yield (SSY) values associated with the value of Q_{mid} on the rising and falling limb (SSY_{RL} and SSY_{FL} respectively) are then interpolated from the SSY-Q graph, allowing for the determination of HI_{mid}, where

$$HI_{mid} = (SSY_{RL} / SSY_{FL}) – 1 \tag{3.3}$$

for clockwise hysteresis, and

$$HI_{mid} = (-1/(SSY_{RL}/SSY_{FL})) + 1 \tag{3.4}$$

for counter-clockwise hysteresis. Resultant HI values range from -2.9 to 1.91 and are positive (negative) for clockwise (counter-clockwise) hysteresis (Lawler et al. 2006). Clockwise hysteresis occurs when SSC values on the rising limb of the hydrograph are greater than those on the falling limb for the same discharge value. Such conditions indicate sediment limited conditions and occur when the sediment source is close to the measuring point, and when the sediment supply is rapidly depleted prior to the peak discharge (Klein 1984, McDonald & Lamoureux 2009). Counter-clockwise hysteresis occurs when SSC values on the rising limb are smaller than those on the falling limb for the same discharge value. Counter-clockwise hysteresis indicates discharge limited conditions, and sediment delivered from the upper portion of a catchment (Klein 1984, McDonald & Lamoureux 2009).

3.4.5 Grain size analysis

Particle size analysis was undertaken to determine how the size of suspended sediment changed during various discharge conditions, and to characterize the type of sediment transported under those conditions. Depth integrated 500 mL water samples were collected every half hour during three sampling campaigns at gauging station WR1 on June 17 (24 hours), July 07 (12 hours) and July 25 (6 hours). Samples were retained unfiltered in Whirlpak™ bags and subsequently analyzed for grain size in the laboratory without chemical pre-treatment using a Beckman Coulter LS 200 Particle Size Analyzer. Each sample was analyzed using seven runs for a total of 60 seconds per run; three runs without sonication, and four runs with. The latter were undertaken in order to separate the aggregated particles (McDonald & Lamoureux 2009). The last run of each analysis type was used unaveraged for comparative analysis after quality comparison with the other runs.

In order to determine the type of sediment being deposited in the channel during the season, grain size samples from either bars located within the channel, or sediment directly at the edge of the active channel were collected at 12 locations. Collection was carried out on June 23, July 1, and July 24 and samples were analysed for particle size distribution. Prior to particle size analysis, the bed samples were treated daily for ten days with successive applications of 1 mL of

5% sodium hypochlorite to remove organic matter. The samples were then analysed for particle size on the same instrument three times with sonication for 60 seconds. The third run was reported unaveraged for comparative analysis to determine particle size changes along the channel during the season.

3.4.6 Channel fine-grained sediment storage

In an effort to quantify the amount of fine sediment stored in the channel, an approximate 100 m^2 area was surveyed at the same 12 locations that the grain size samples were collected. Three transects, spaced five meters apart and extending across the channel were surveyed with a level every half meter to identify the presence of bars within the channel (Figure 3.4).   Bar area and percentage of fine sediment was visually estimated using a 0.25 m^2 quadrat. In situations where the bar or the fine sediment did not fill the quadrat in its entirety, the percentage filled was visually estimated by dividing the quadrat into quarters. Area was then determined as the product of quadrat area and percentage of quadrat filled.

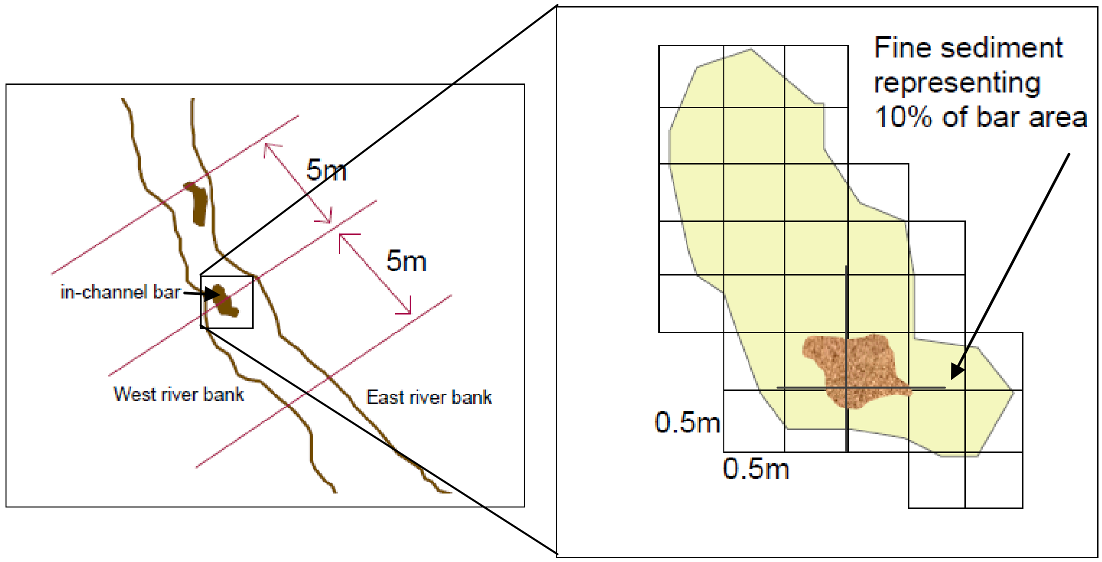


Figure 3.4 Example of how the 0.25 m^2 quadrat was applied to determine bar area. Presence and percentage of fine sediment (<2mm) was visually estimated according to the percentage in each quadrat.

3.4.7 Channel characterisation and change detection

In order to determine longer term changes in the West River channel that might reflect sediment sources or disturbance, aerial photographs and recent high resolution satellite images were compared. Aerial photographs from August 2, 1950 were orthorectified with the OrthoEngine tool in PCI Geomatics. The aerial photographs (1:20000, focal length: 137.16 mm) were scanned at 1200 dpi and orthorectified using a 2009 multispectral GeoEye satellite image (2 m resolution), and a 2009 digital elevation model (DEM) with a ten meter contour interval (Collingwood 2011, pers. communication). A false colour composite (bands 1, 2, and 3) of the GeoEye image was used for better distinction of the channel edges. Ten distinctive ground control points from the aerial photographs and the GeoEye image were used. The first photograph (A12858-164) had an overall root mean square (RMS) error of 1.35 m on the ground

(X RMS: 1.04 m, Y RMS: 0.86 m), while the second photograph had an overall RMS error of 1.35 m (X RMS: 0.96 m, Y MRS: 0.95 m).

In order to track how the channel edge, measured from bank to bank, had changed with time, the boundary outline was digitized in ESRI ArcGIS 10 software to create polylines that were overlaid for analysis. Additionally, a long profile of the West River, beginning at the source of reach 4 to the outlet of the river was created in ArcGIS by digitizing a 3D line from the ten meter DEM, then using the profile graph tool to create the long profile (Figure 3.2).

3.5 Results

3.5.1 Meteorology

The 2010 summer melt season at Cape Bounty represented a comparatively normal year. Beginning on June 9, temperatures remained above freezing, rising to a maximum of 14.3 °C on July 3 (Figure 3.5). While there were four days with rainfall exceeding 3.0 mm, these amounts are minor when compared to previous years (McDonald & Lamoureux 2009, Dugan et al. 2009) and had limited hydrological response. Mean temperature and precipitation for the months of June and July were 3.3 °C, 0.2 mm and 7.6 °C and 0.4 mm, respectively.

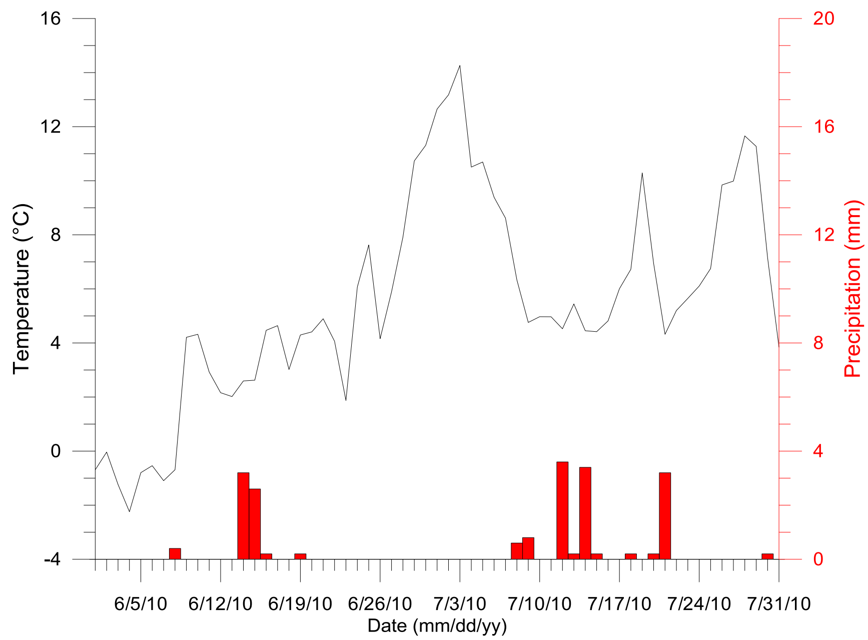


Figure 3.5 Mean daily temperature and precipitation for the West watershed of the CBAWO, recorded at MainMet (Figure 3.1).

3.5.2 Hydrology

Typical of an arctic nival regime, the hydrograph of the West River included three major hydrological phases. Flow began June 13 and continually increased until June 17 when the peak discharge of 1.81 m^3/s was reached at WR1 (Figure 3.6). This period was followed by a rapid decrease followed by a secondary increase on June 25 when discharge reached 0.70 m^3/s as a result of warm air temperatures (Figure 3.5). Flow receded to baseflow due to diminished snowpack after July 10 with mean discharge reduced to ~0.02 m^3/s. Total measured runoff for the West River catchment was 89.6 mm.

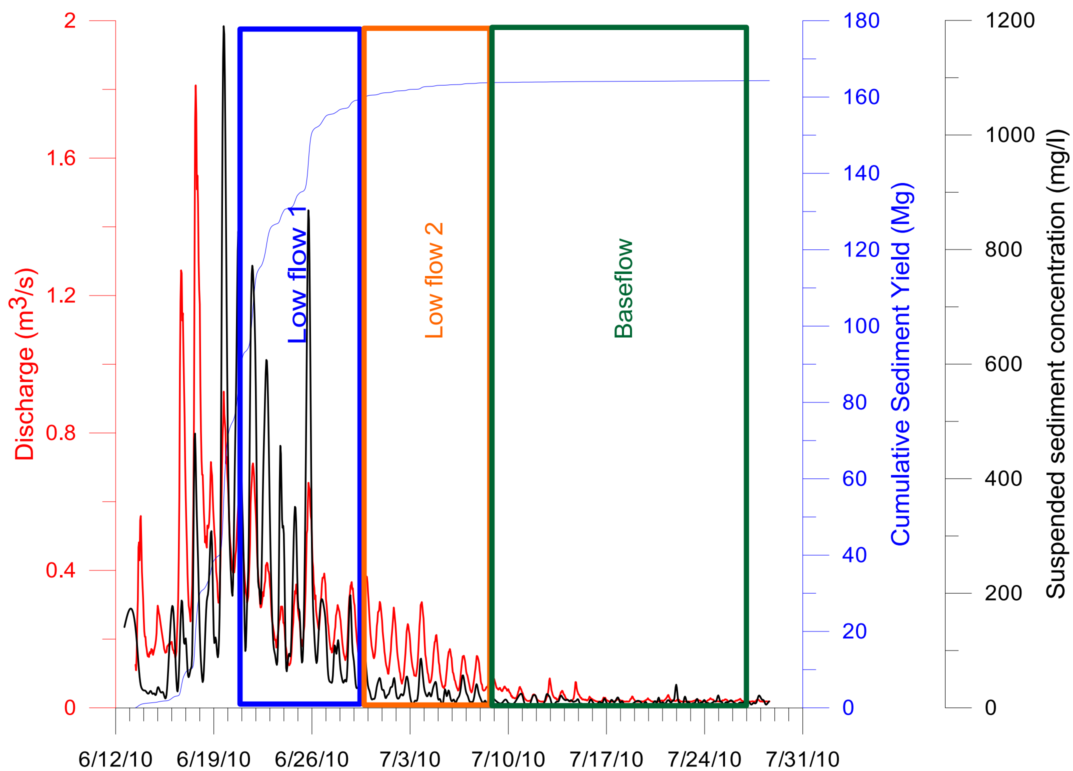


Figure 3.6 Discharge (m^3/s), SSC (mg/l) and cumulative sediment yield (Mg) for the West River, measured at WR1, during the 2010 hydrological season at the CBAWO.

Along the length of the main channel, measured discharge increased downstream during the three hydrological periods (Figure 3.7). Discharge measured at the WR4 gauging station was the lowest of all five stations along the main river. At all stations, discharge exhibited clear diurnal variations associated with daily changes in solar energy and snowmelt.

Of the five major gauged inflows, Ptarmigan (PT) had the highest discharge between June 11 and 22 and was the first to cease flow on June 29. By contrast, Goose (G) had the lowest maximum discharge (Figure 3.8). Big Slide (BS) was the only gauged tributary with continuous flow, due to a large perennial snow bank in the catchment.

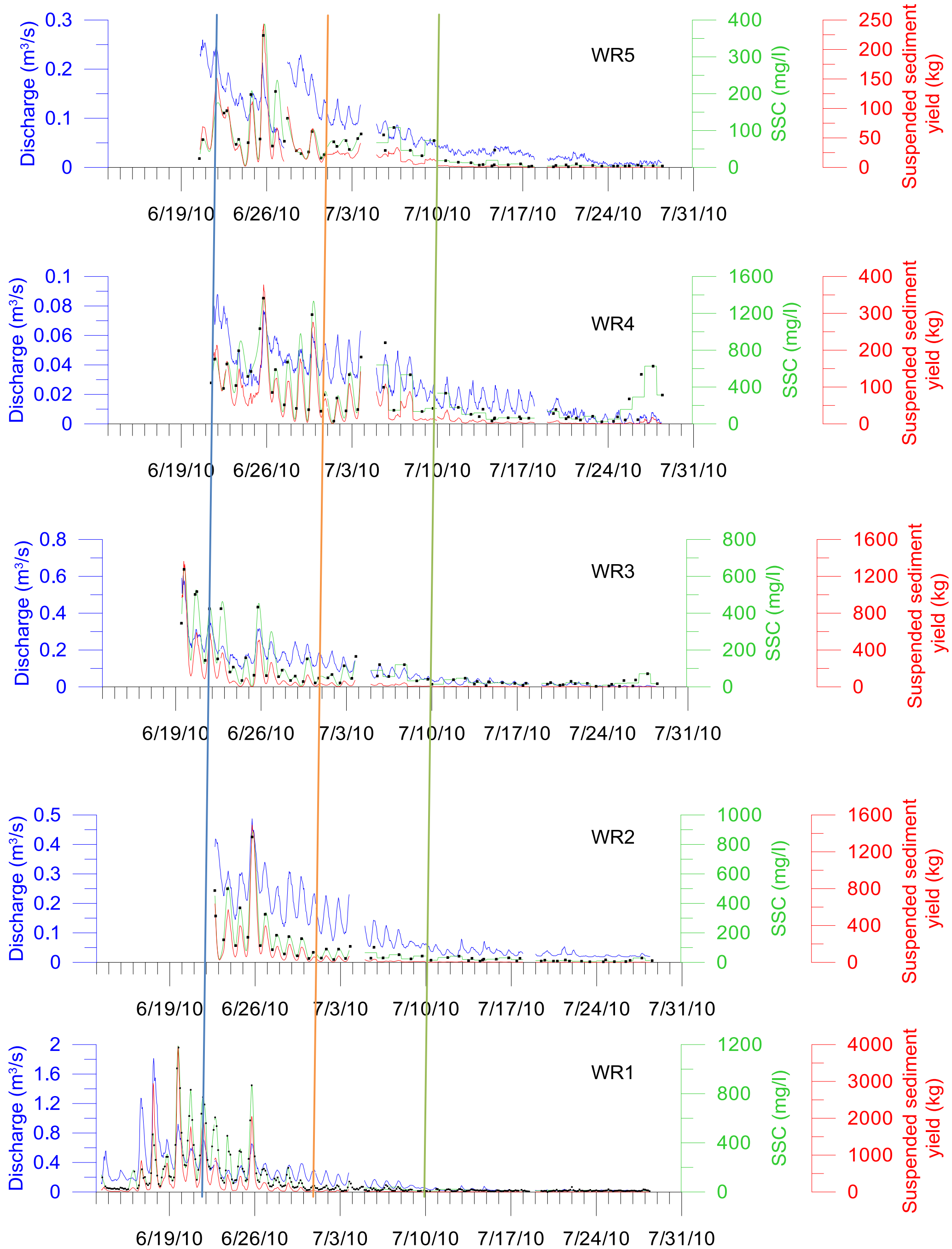


Figure 3.7 Discharge (m^3/s), SSC (mg/l), and suspended sediment yield (kg) for stations WR1-5 during the monitored 2010 season at the CBAWO. Black points indicate manual SSC measurements. Note the difference in discharge, SSC and suspended sediment yield scale at each station. Blue, orange and green lines indicate beginning of low flow 1, 2 and baseflow periods respectively.

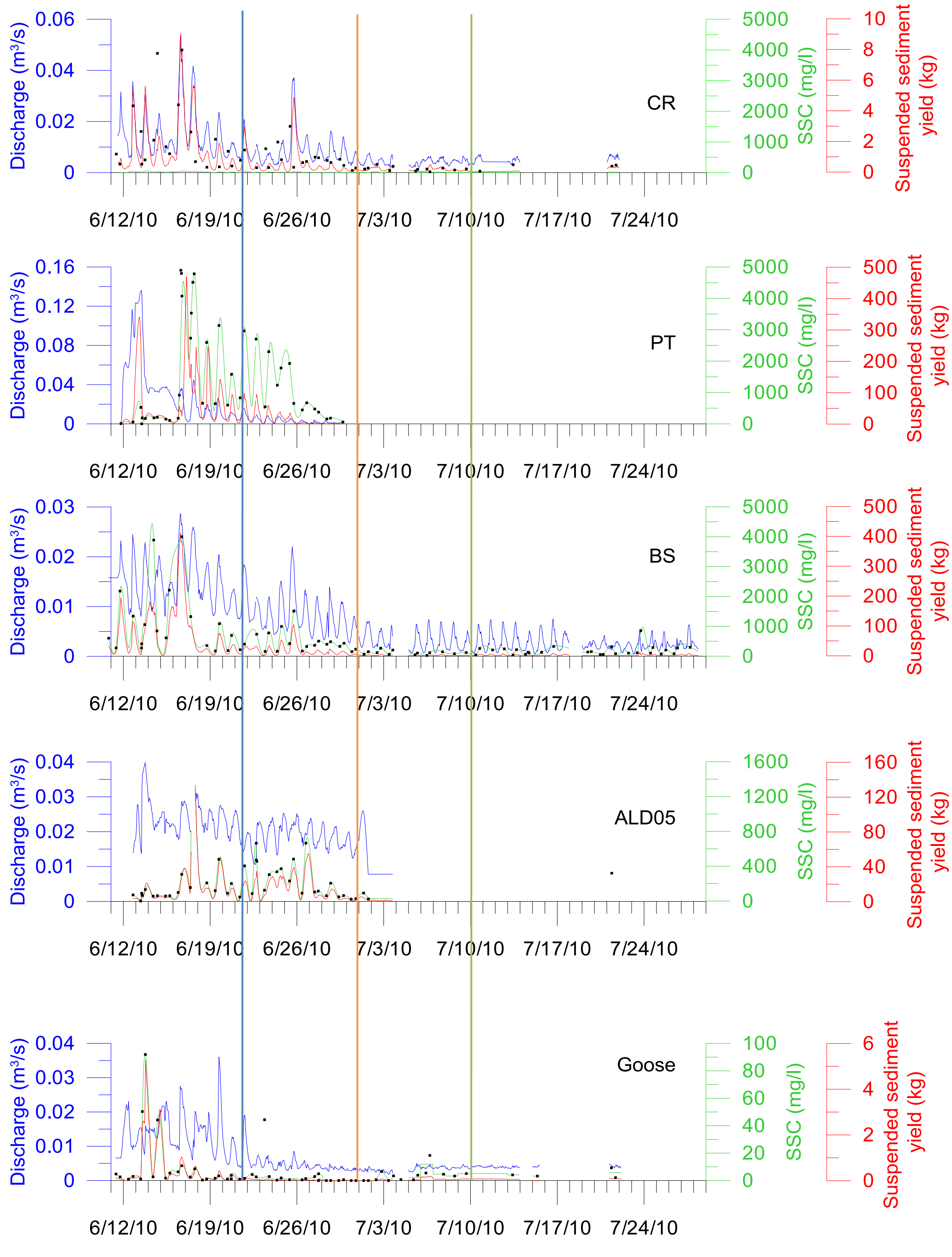


Figure 3.8 Discharge (m^3/s), SSC (mg/l), and suspended sediment yield (kg) for the subcatchments during the monitored 2010 season at the CBAWO. Black points indicate manual SSC measurements. Note the difference in discharge, SSC and suspended sediment yield scale at each station. Blue, orange and green lines indicate beginning of low flow 1, 2 and baseflow periods respectively.

3.5.3 Suspended sediment concentration

SSC along the West River varied considerably between stations. On average, the headwater station WR5 had the lowest SSC for the duration of the study, while the other headwater station, WR4, had the highest (Table 3.2). At the other stations, the relative concentrations varied. For instance, during the low flow 1 period, WR2 SSC were higher than those at WR3, while the opposite occurred during the following two periods. During the second low flow period, WR5, which typically had the lowest SSC values, had the third highest values.

Table 3.2 Mean SSC (mg/l ± S.D.) for gauging stations WR1-5 for the three hydrological phases and the entire study period. Note that WR4 and WR5 are located furthest upstream and WR1 is furthest downstream (Figure 3.1). Numbers in brackets represent sample size.

	Low flow 1	Low flow 2	Baseflow	Study period
WR5	102.6 ± 85.8 (17)	67.5 ± 22.3 (12)	7.4 ± 8.7 (28)	48.4 ± 63.6 (57)
WR4	569.3 ± 373.5 (16)	344.2 ± 270.1 (12)	128.8 ± 150.8 (28)	304.5 ± 312.0 (56)
WR3	115.7 ± 128.6 (16)	74.5 ± 44.6 (12)	21.0 ± 15.2 (28)	59.5 ± 82.1 (56)
WR2	218.4 ± 207.8 (17)	53.5 ± 34.0 (12)	15.6 ± 10.1 (27)	85.3 ± 144.6 (56)
WR1	156.5 ± 161.4 (65)	24.7 ± 15.4 (68)	10.9 ± 5.3 (135)	49.7 ± 100.1 (268)

During baseflow, SSC sampling was reduced to once daily in order to carry out mapping and channel sampling. Due to the potential bias of once daily sampling, a comparison between mean hourly SSC splined from 3-hour pump samples and a single daily sample at station WR1 was undertaken to quantify this bias, and results indicate a general agreement between the two daily estimates. However, on some days during baseflow, the once-daily SSC values differ from the splined mean (n=24) as much as 83% (Figure 3.9).

Still, when analyzing the baseflow period as a whole, this uncertainty is reduced to 16%. The uncertainty value was determined by comparing the mean hourly SSC at WR1 to the splined SSC values associated to the time nearest to when a sample at WR2 was collected as the latter is the station in closest proximity to WR1 (Figure 3.1). The mean hourly SSC at WR1 was determined by averaging the splined hourly SSC values for a given day. The uncertainty associated with the SSC measurements from the once-daily sampling will carry through to the suspended sediment yield measurements.

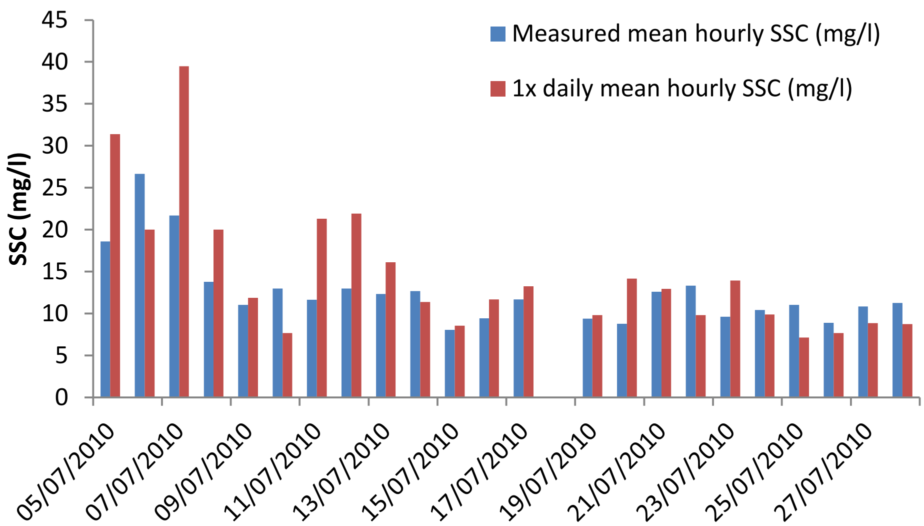


Figure 3.9 Comparison of measured and estimated mean hourly SSC values at WR1 during the period of once-daily SSC measurements. No data is available for July 18.

Of the five gauged tributaries, Ptarmigan generated the highest SSC of 4784 mg/l on June 17 and SSC measurements at Big Slide were the second highest, reaching a maximum value of 4435 mg/l on June 14. Throughout the study period, the undisturbed Goose tributary consistently had the lowest SSC values (Figure 3.8). Flow from these tributaries declined after the snowmelt period and ceased in early July, with the exception of Big Slide which flowed for the remainder of the season, and Goose and Caribou (CR) which ceased to flow in mid July until rejuvenated by July rainfall.

In addition to the gauged tributaries, 13 ungauged inflows of water were monitored throughout the season (Figure 3.1). Visual observations and SSC samples revealed these sources to be minimal contributors of water and sediment to the channel. SSC values were generally less than 10 mg/l (n=7) (Table 3.3) and most of these smaller inflows dried out early in the hydrological season.

Table 3.3 SSC values of the ungauged inflows of water for the West River watershed.

Sample date & time	Sample number	SSC (mg/L)
19/06/2010 18:25	WPT160	17.4
22/06/2010 11:00	WPT159-01	8.5
23/06/2010 13:10	WPT170-01	1.8
23/06/2010 16:30	WPT155-01	0.6
23/06/2010 16:45	WPT157-01	1.1
23/06/2010 17:00	WPT159-01	2.7
29/06/2010 19:00	WPT159-02	5.1

On June 29 there was a sharp SSC increase at station WR4 caused by the release of turbid water that had ponded behind deep snow upstream of the station. While no discharge measurements were taken from the location where the pond was draining, SSC samples from the draining water (PD), along with water directly upstream of the pond discharge (WR7) were collected. Water directly upstream from the turbid inflow had SSC values below 8 mg/l, while the water draining the pond was extremely turbid (Figure 3.10), and had SSC values in excess of 340 mg/l and approaching 1150 mg/l (Table 3.4).

A paired two-sample t-test of the comparative samples collected at WR5 and 6 indicated that the water at the upstream location (Figure 3.3) had significantly lower (t_{obs}=4.01, $p<0.001$, n=36) concentrations of suspended sediment than the water at WR5 (Table 3.5).

Figure 3.10 Release of turbid pond water upstream of WR4. Samples from the pond release are designated PD in Table 3.4.

Table 3.4 Comparison of SSC (mg/l) from WR4, the pond water (PD), and directly upstream of the pond (WR7).

Sample date & time	WR4	PD	WR7
05/07/2010 14:15	397.4	1149.3	5.4
06/07/2010 10:25	147.9	808.3	2.3
08/07/2010 18:00	134.3	340.5	7.8
09/07/2010 15:30	168.2	-	3.1
11/07/2010 17:10	179.1	-	1.7
12/07/2010 17:58	103.6	-	1.5
15/07/2010 19:11	65.8	-	0.2

Table 3.5 Comparison of SSC (mg/l) between WR5 (downstream) and WR6 (upstream).

Sample date & time	WR5	WR6
29/06/2010 17:25	97.3	23.7
30/06/2010 17:10	34.6	9.1
01/07/2010 18:20	57.0	6.2
02/07/2010 18:40	48.7	4.8
03/07/2010 17:55	90.8	7.8
05/07/2010 17:25	46.0	7.4
07/07/2010 18:45	46.2	6.6
08/07/2010 17:35	31.7	2.8
10/07/2010 16:15	18.3	6.9
11/07/2010 17:10	14.0	7.8
12/07/2010 18:06	12.7	3.5
15/07/2010 19:16	6.9	0.7
16/07/2010 19:43	9.5	1.2
17/07/2010 17:36	3.5	2.2
19/07/2010 17:07	5.3	0.8
20/07/2010 17:45	1.7	2.2
21/07/2010 17:33	2.5	1.5
22/07/2010 17:40	5.5	0.4

3.5.4 Suspended sediment yield

Total measured 2010 suspended sediment yield for the West River was 164 Mg, 91% of which was transported prior to June 26 (Figure 3.6). Sediment yields for the time period covering peak discharge are not available for all reaches as several gauging stations were installed late due to deep snow cover and rapidly shifting initial channel formation. The time when sediment fluxes are available is divided into three periods (low flow 1, low flow 2, and baseflow; Figure 3.6) corresponding to major changes in discharge in the West River.

The highest calculated daily sediment yields of the season, once all stations were established, occurred during the low flow 1 period. Notably, WR2 had the highest sediment yield during this period (Figure 3.11), while WR4, with the highest overall SSC had the third highest yields. However, during low flow 2 and baseflow, these quantities reversed and the reach with the greatest efflux of sediment was reach 4 (Figure 3.11). Additionally, during the low flow 2 period, WR1 changed from having the second highest sediment yield during the previous period to the lowest yield. The low yield in downstream WR1 continued during the baseflow period, and was comparable to the headwater WR5 reach.

On June 25 there was an increase in the sediment yield at all the main West River stations (Figure 3.7). However, this increase was not substantial in the gauged tributaries, with the exception of BS and CR, where sediment yields increased from 507 and 8 kg on June 24 to 1072 and 39 kg on June 25, respectively. The increase coincides with warmer temperatures and likely enhanced snowmelt, but the lack of response in some tributaries suggests that snow availability at the time was a significant controlling factor on sediment transfer.

During the release of the ponded water above station WR4 beginning on June 29, there was an increase in SSC which translated into an increase in the sediment yield at WR4 (Figure 3.7) that lasted several days, until the pond had fully drained. During this isolated event, the measured discharge at this station increased slightly by approximately 7% compared to the preceding days. The sediment yield prior to the pond influence, between June 28 1000h and June 29 0600h was 2022 kg (averaging 101 kg/hr). Once the pond started draining, the sediment yield increased by 1455 kg, for a total yield of 3477 kg (129 kg/hr) between June 29 0700h and June 30 1000h. In the 24 hours from June 30 1100h to July 01 1100h, sediment yield at this location decreased from 3477 to 769 kg (32 kg/hr). These values represent a decline of approximately 78% and suggest that the majority of the sediment had been flushed out of the pond. Although downstream transmittance of the pond release increased the sediment yield at station WR3, sediment yields at stations WR2 and WR1 decreased, suggesting a loss to storage in the intervening reaches.

Ptarmigan (PT) was the most turbid of all tributaries followed closely by BS. However, PT ceased to function as a source of flow and sediment on June 25 whilst BS continued to contribute flow and sediment throughout the measurement period. In contrast, discharge from G contributed negligible amounts of sediment to the WR. The tributaries CR, PT, ALD05, and G all ceased to function as sources of flow and sediment to the main channel at different times throughout the season (Figure 3.7) indicating that the contribution of ALD as sources of sediment to the West River was sporadic in time and gradually declined as snow cover diminished.

3.5.5 Suspended sediment yield-discharge hysteresis

During the annual peak flow on June 17, hourly sediment transport exhibited a strong clockwise hysteresis pattern (HI_{mid}=1.91, k=0.5). By contrast, during the peak discharge period (June 14-21), the West River (at WR1) was characterised by a strong counter-clockwise hysteresis relationship (Hi_{mid}=-2.9, k=0.5). The remainder of the sampling season (low flow and baseflow) was characterised by clockwise relationships (Hi_{mid}>0) between suspended sediment yield and discharge (Figure 3.12). These changes in hysteresis patterns suggest changing sediment sources and supply during the hydrologically active period.

Although most daily SSC-Q hysteresis patterns were dominated by clockwise relationships there were six days in July exhibiting counter-clockwise hysteresis, suggesting the occurrence of isolated events which would increase sediment availability following peak discharge on those given days. An analysis of the HI_{mid} revealed a trend towards increasingly stronger clockwise hysteresis as the season progressed (Figure 3.13).

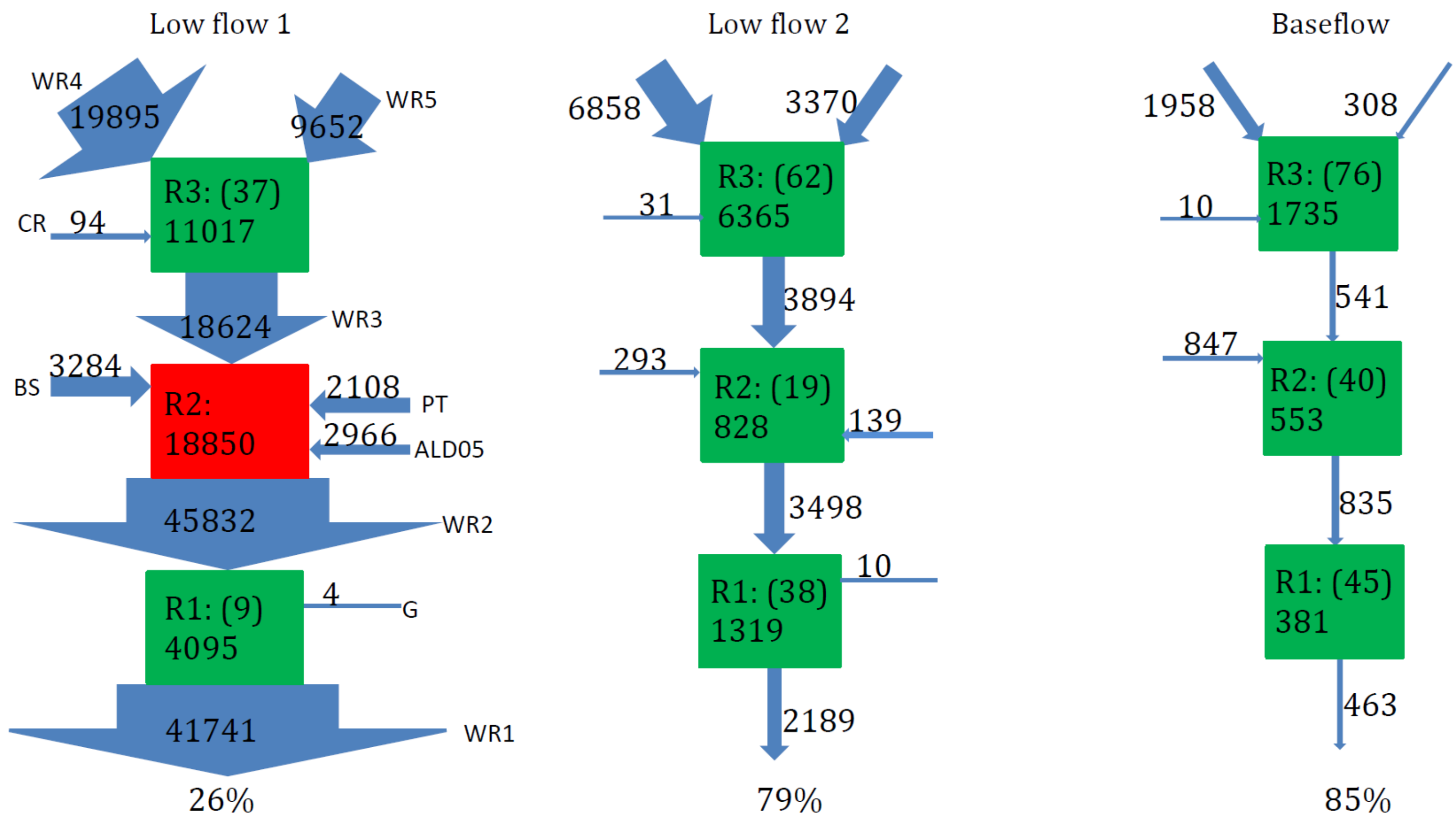


Figure 3.11 Sediment budget components (all values in kg) for West River reaches R1-3. Sediment inflows and transfers between reaches are indicated by blue arrows, with numbers representing flux values. Arrow width is scaled to represent sediment flux and the arrow length in the West River reaches (WR1-5) is scaled to channel length. Values in red boxes indicate net erosion and green boxes indicate net storage. Percentage values at the bottom represent the total percentage of sediment stored within the system, compared to measured inflows. Values in parentheses indicate the percentage of sediment stored in each reach relative to inflows.

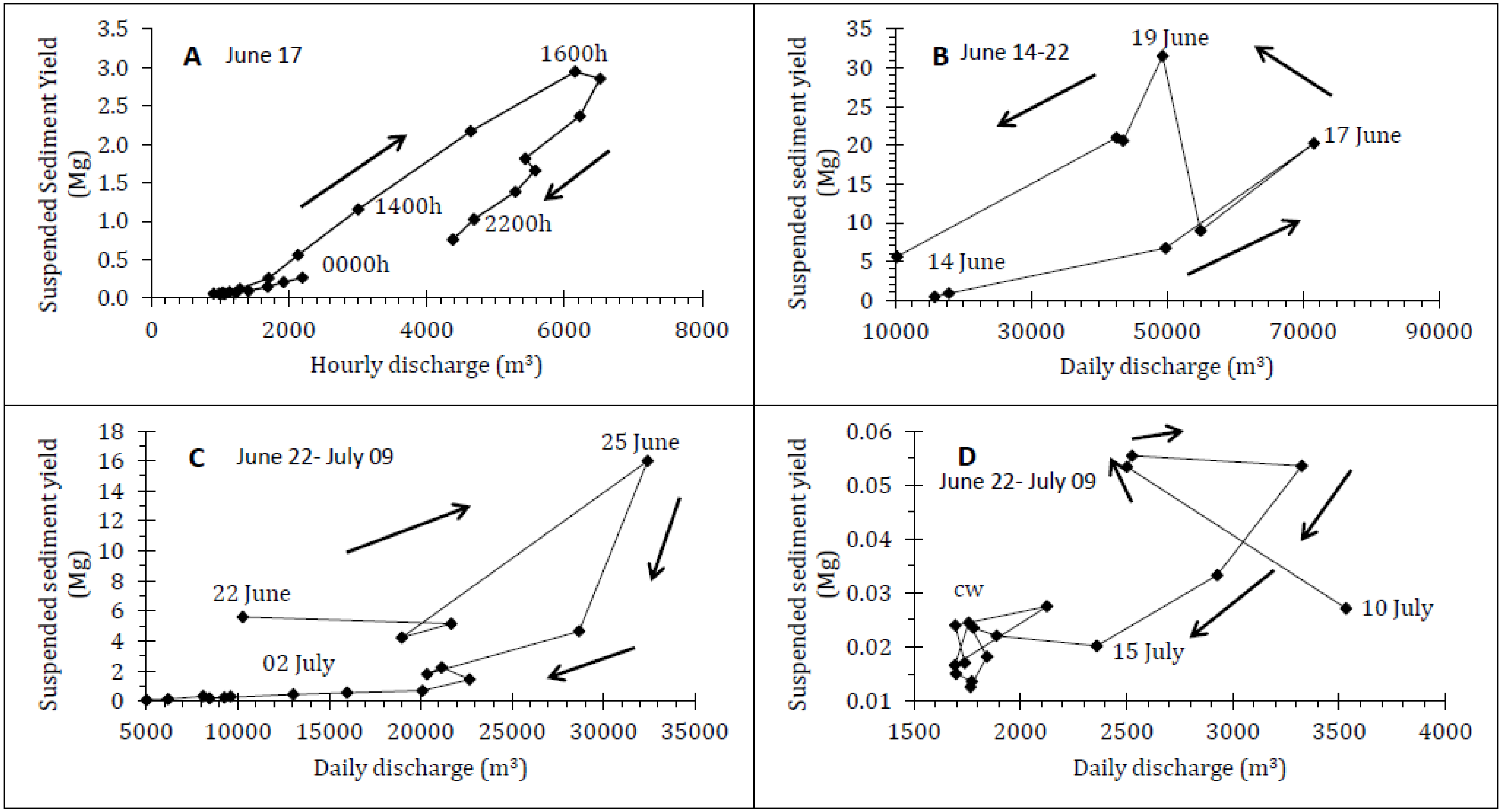


Figure 3.12 Suspended sediment yield-discharge hysteresis relationships at gauging station WR1 for A) peak discharge day (June 17); clockwise (HI_{mid}=1.91, k=0.5), B) peak discharge period (June 14-22): counter-clockwise (HI_{mid}= -2.9, k=0.5), C) recession period (June 22-July 09): clockwise (HI_{mid}=0.88, k=0.5), and D) baseflow (July 10-27): multiple weak clockwise relationships.

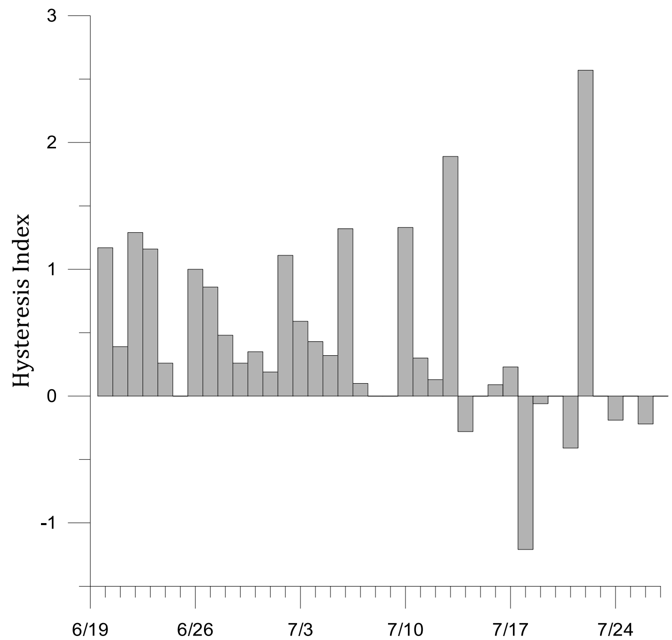


Figure 3.13 Daily hysteresis index at station WR1 where positive values indicate clockwise hysteresis, and negative values indicate counter-clockwise hysteresis. Larger HI values indicate a greater degree of hysteresis. Days with no bars indicate a HI index of 0.

3.5.6 Suspended sediment budget

Transfer of sediment from upstream reaches and tributaries changed substantially during the season both on a daily and period scale (Figure 3.14, 3.11). The rainfall events on July 12, 14, and 21 were not large enough to have a substantial impact on the sediment budget dynamics of the West River. As such, after all three events, each reach continued to store sediment (Figure 3.11). Reach 2 was the only segment which acted as a net source of sediment during low flow 1, at which time it released 18850 kg of sediment over a 9 day period (Figure 3.11). During the subsequent low flow 2 and baseflow period, reach 2 became a net sink for sediment, as was the case for reaches 1 and 3 throughout the study period. Reach 3 stored the largest amount of sediment in comparison to the two downstream reaches during the three periods, both in absolute and percentage terms (Figure 3.11). For instance, during the low flow 2 period, reach 3 stored 7.5 times more sediment than reach 2. As time progressed and sediment yields decreased, the relative proportion of sediment that each reach stored increased, and hence, the West River progressively stored a greater proportion of sediment as discharge decreased. For the three periods measured, the West River stored 26, 79, and 85% of the overall sediment supplied to it, respectively, thereby indicating a decreasing sediment delivery ratio of .74, .21, and .15 for the three time periods respectively, as the season progressed (Figure 3.11).

While the daily dynamics of sediment storage and release for the three reaches generally coincide with the period scale dynamics, there are exceptions. Most notably, on June 25, reach 1 became a source of sediment, releasing 1233 kg of sediment. This release may have been caused by thermal erosion as air temperature increased from 1.9 °C to 7.6 °C between June 23 and 25 (Figure 3.5, 3.14). Similarly on July 9, reach 2 which had in the previous days been a sediment sink, became a source of sediment, releasing 26.5 kg of sediment (Figure 3.14). This coincided with increased storage of sediment in downstream reach 1 (156.6 kg). The following days, storage in reach 1 decreased as reach 2 once again stored more sediment, therefore releasing less sediment downstream.

When the pond upstream of WR4 drained, suspended sediment storage dynamics changed throughout the West River (Figure 3.15). In the first 27 hours after the initial release of pond water, beginning June 29 0700h, the net amount of sediment stored in reaches 1 and 3 increased. This increase was especially evident in reach 1 which underwent a nearly five-fold increase compared to previous days. Even though the net storage increased in reach 3, the proportion of sediment stored decreased from 76 to 66%, indicating a net flushing of sediment to the downstream reaches. In contrast, the proportion of sediment stored in reach 1 increased from 7 to 37%. At the same time, while reach 2 was characterised by net erosion, the net release of sediment decreased and continued to decrease once the pond had fully drained. From June 30 – July 1, reach 3 was the only segment of river where net storage continued (storing 661 kg) while reach 1 underwent net erosion of 1622 kg.

Period	Date	R1	R2	R3
Low Flow 1	22/06/2010	-3411.1	1007.1	-682.6
Low Flow 1	23/06/2010	854.3	-2602.7	896.0
Low Flow 1	24/06/2010	1180.8	-2761.1	1721.2
Low Flow 1	25/06/2010	-1233.3	-8026.1	2254.5
Low Flow 1	26/06/2010	4103.9	-3868.7	257.7
Low Flow 1	27/06/2010	1455.8	-865.4	785.7
Low Flow 1	28/06/2010	45.6	-992.3	1951.3
Low Flow 1	29/06/2010	669.4	-544.9	2445.1
Low Flow 1	30/06/2010	429.5	-196.8	1387.8
Low Flow 2	01/07/2010	270.6	-3.2	512.4
Low Flow 2	02/07/2010	314.5	-39.0	978.4
Low Flow 2	03/07/2010	255.0	75.3	769.3
Low Flow 2	04/07/2010			
Low Flow 2	05/07/2010	246.2	105.0	1685.4
Low Flow 2	06/07/2010	49.6	181.5	682.7
Low Flow 2	07/07/2010	24.9	472.0	1032.7
Low Flow 2	08/07/2010	1.2	63.2	264.5
Low Flow 2	09/07/2010	156.6	-26.5	438.3
Baseflow	10/07/2010	36.9	32.2	537.9
Baseflow	11/07/2010	46.8	54.7	201.6
Baseflow	12/07/2010	71.9	26.8	77.4
Baseflow	13/07/2010	19.6	29.0	133.6
Baseflow	14/07/2010	18.9	12.7	72.2
Baseflow	15/07/2010	54.6	1.1	51.3
Baseflow	16/07/2010	64.1	21.1	69.8
Baseflow	17/07/2010	22.3	47.6	58.2
Baseflow	18/07/2010			
Baseflow	19/07/2010	11.1	19.4	119.7
Baseflow	20/07/2010	-4.9	9.4	43.5
Baseflow	21/07/2010	10.2	28.9	19.0
Baseflow	22/07/2010	0.7	21.8	16.2
Baseflow	23/07/2010	-17.3	75.2	6.4
Baseflow	24/07/2010	0.0	52.8	13.5
Baseflow	25/07/2010	-3.7	29.0	39.7
Baseflow	26/07/2010	2.1	27.6	53.1
Baseflow	27/07/2010	44.7	55.0	201.9
Baseflow	28/07/2010	3.1	8.3	20.0

Figure 3.14 Daily sediment budget components for West River reaches (R1-3). Values in red boxes indicate net erosion and green boxes indicate net storage (all values in kg). Data is not available for July 4 and 18th.

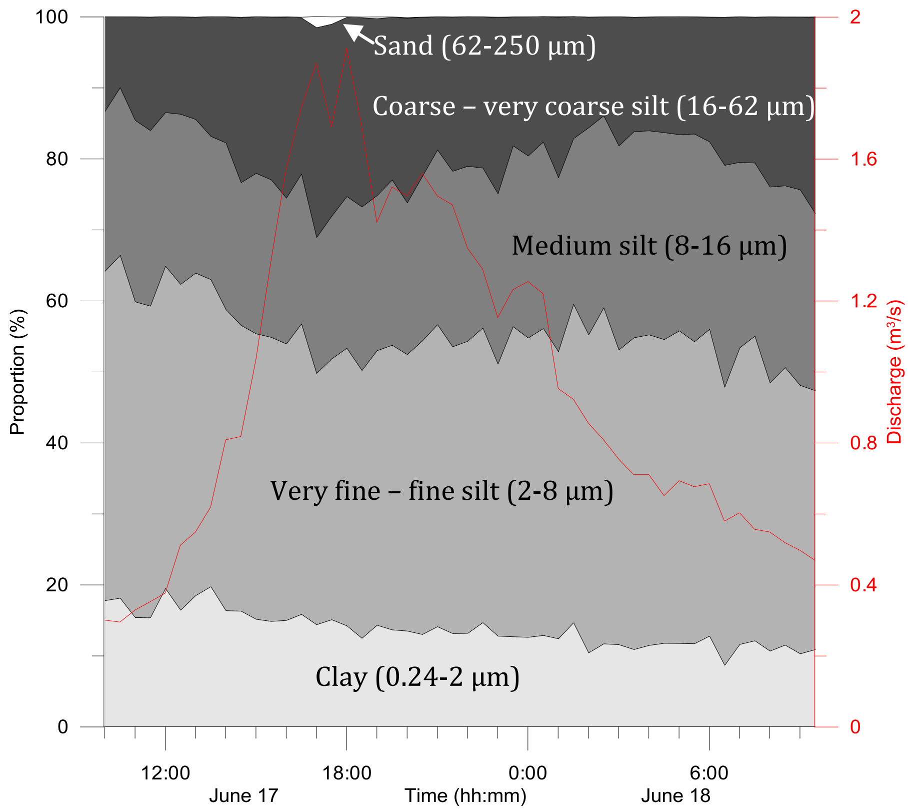
	28/06/2010 1000h - 29/06/2010 0600h	29/06/2010 0700h - 30/06/2010 1000h	30/06/2010 1100h - 01/07/2010 1100h
R3	2032.9 (76)	2952 (66)	661 (53)
R2	1252	247	4
R1	162 (7)	780 (37)	1622

Figure 3.15 Sediment budget components for West River R1-3 during the release of pond water upstream of gauging station WR4 (see Figure 3.10). Values in red boxes indicate net erosion and green boxes indicate net storage (both in kg). Values in parentheses indicate the percentage of sediment stored relative to the sediment entering the reach.

3.5.7 Grain size characteristics of suspended sediment

During the peak discharge period, there was a decoupling between grain size and discharge. Initially, as discharge increased, the river was able to transport coarser material (Figure 3.16). Small quantities of sand (~2-3%) were only transported when discharge peaked at 1.80 m^3/s. It was also at this time that more than 20% of the material transported was within the coarse to very coarse silt range. However, as discharge decreased, there was an increase in the volume of coarse to very coarse silt being transported after 0200 h on June 18.

During periods of high (June 17-18), low (July 7), and very low (July 25) flow, the d10 (the grain size where 10% of the grain size distribution is smaller than this value) and d50 grain sizes did not vary significantly. However, the change in d90 grain size decreased by approximately 10 µm between the high and very low flow periods, a statistically significant change (Table 3.6).



3.16 Discharge (m^3/s) and proportion (%) of particle size classes transported during peak discharge at station WR1 of the West River.

Table 3.6 d10, d50 (median), and d90 grain size (µm ± sd) of the West River at WR1 during peak discharge, low flow, and very low flow conditions. Numbers in brackets represent sample size.

Date & time range	d10	d50	d90
06/17/2010 - 06/18/2010 (10:00 - 10:00)	1.7 ± 0.3 (48)	7.9 ± 0.8 (48)	27.0 ± 5.8 (48)
07/07/2010 (10:00 - 22:00)	1.6 ± 0.2 (22)	6.5 ± 0.7 (22)	15.7 ± 1.9 (22)
07/25/2010 (9:00-12:00, 16:00-19:00)	2.3 ± 0.7 (4)	7.5 ± 0.2 (4)	17.3 ± 0.7 (4)

3.5.8 Extent and particle size characteristics of fine-grained channel deposits

The median grain size of the samples collected on July 1 indicates that coarser sediment was preferentially deposited in the upper reaches of the river, while finer material was generally deposited downstream, closer to the mouth (Figure 3.17A). Additionally, a comparison between the median values from July 1 and July 24 indicates that over time, the sediment stored became progressively finer, especially in the upper regions of the West River. The sediment median size on average became finer by 49 µm between the two dates (Figure 3.17B). However, downstream at sites 1, 2, 6 and 7, the sediment deposited became coarser over time.

The channel areas surveyed indicated diverse channel conditions and bar coverage. Although reach 3 had the greatest number of bars, especially at locations 8 and 11, reach 2 had on average, the largest extent of bars (Table 3.7). However, the bars in reach 2 were covered by a smaller proportion of fine material, resulting in an overall reduced accumulation of fine sediment in the bars. By contrast, upstream of WR4, at site 12, the channel was composed entirely of fine sediment (Figure 3.18).

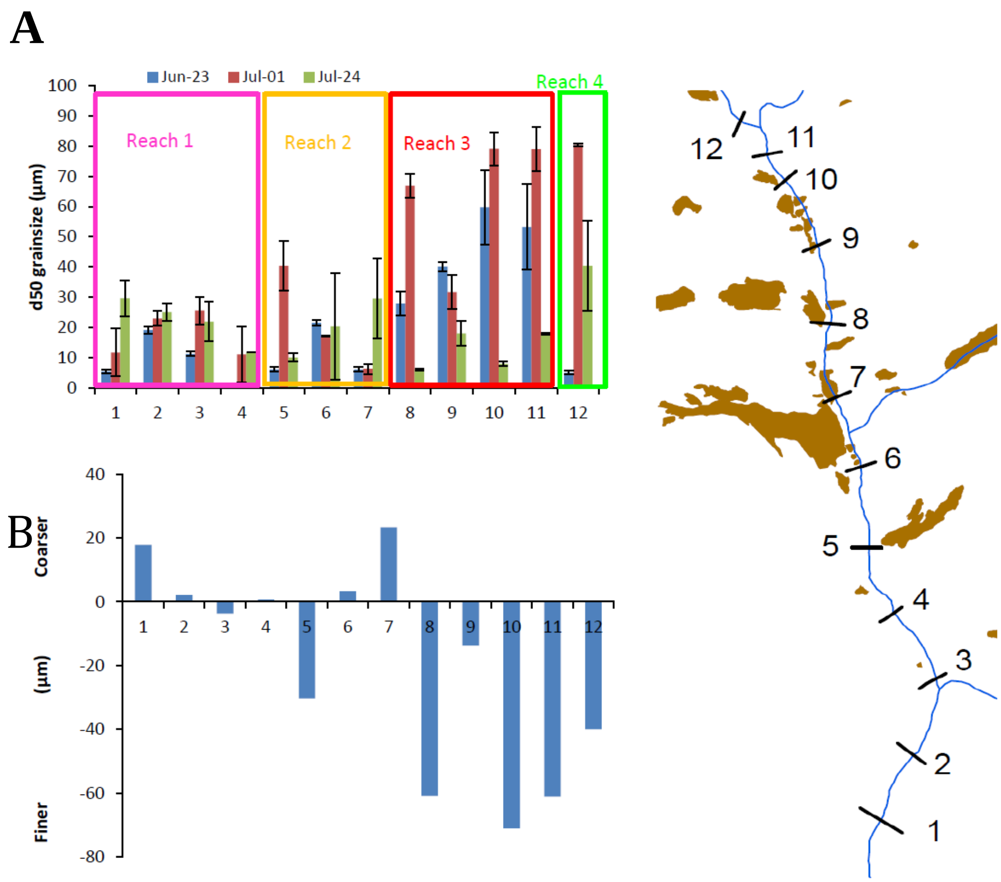


Figure 3.17 A. Median grain size of overbank grain size samples obtained on three dates during the season. B. Difference in median grain size between July 24 and July 1 repeat samples, where positive values indicate coarsening as the season progressed. Sample locations are indicated on the map. Error bars represent standard error.

Table 3.7 Description of bars at sample transects in the West River. Site numbers correspond to grain size sample locations and are mapped in Figure 3.17.

	Site	Bar area (m^2)	% fines in bar	area of fines (m^2)
Reach 1	2 bar 1	0.4	--	--
Reach 2	4 bar 1	10.3	25-30	2.6-3.1
	5 bar 1	5.5	45-50	2.5-2.7
	5 bar 2	0.2	5	0.0
	6 bar 1	15.6	0	0
Reach 3	8 bar 1	8.8	45-50	3.9-4.4
	8 bar 2	7.8	10-15	0.8-1.2
	8 bar 3	5.0	5-10	0.2-0.5
	8 bar 4	0.2	100	0.2
	9 bar 1	0.4	35	0.1
	9 bar 2	0.1	90	0.1
	10 bar 1	8.1	60	4.8
	10 bar 2	1.3	65	0.8
	11 bar 3	1.6	20	0.3
	11 bar 4	1.3	95	1.3
Reach 4	12 bar 1	4.3	100	4.3

Figure 3.18 Photograph of the fine sediment in the channel of reach 4 of the West River, taken July 24, 2010.

3.5.9 Channel characteristics and long term change

In the 59 year period between1950 and 2009, some areas of the West River have shifted laterally by tens of meters, while in other regions, the change has been minimal (Figure 3.19). For instance, at the location of the ALD associated with the Big Slide subcatchment, the West River channel narrowed by 3 m between 1950 and 2009. Observations in 2007 indicated that at this location, the channel was dammed by an ALD (Figure 3.20) (Lamoureux & Lafrenière 2009). This indicates that in the three years since 2007, the water has breached the dam and reconnected flow upstream and downstream of Big Slide. Overall, there has been a general narrowing of the channel width over the 59 year period (Figure 3.19). Based on the 2010 ground surveys, reach 2 had on average the widest floodplain, followed by reach 3 then 1 (Table 3.8).

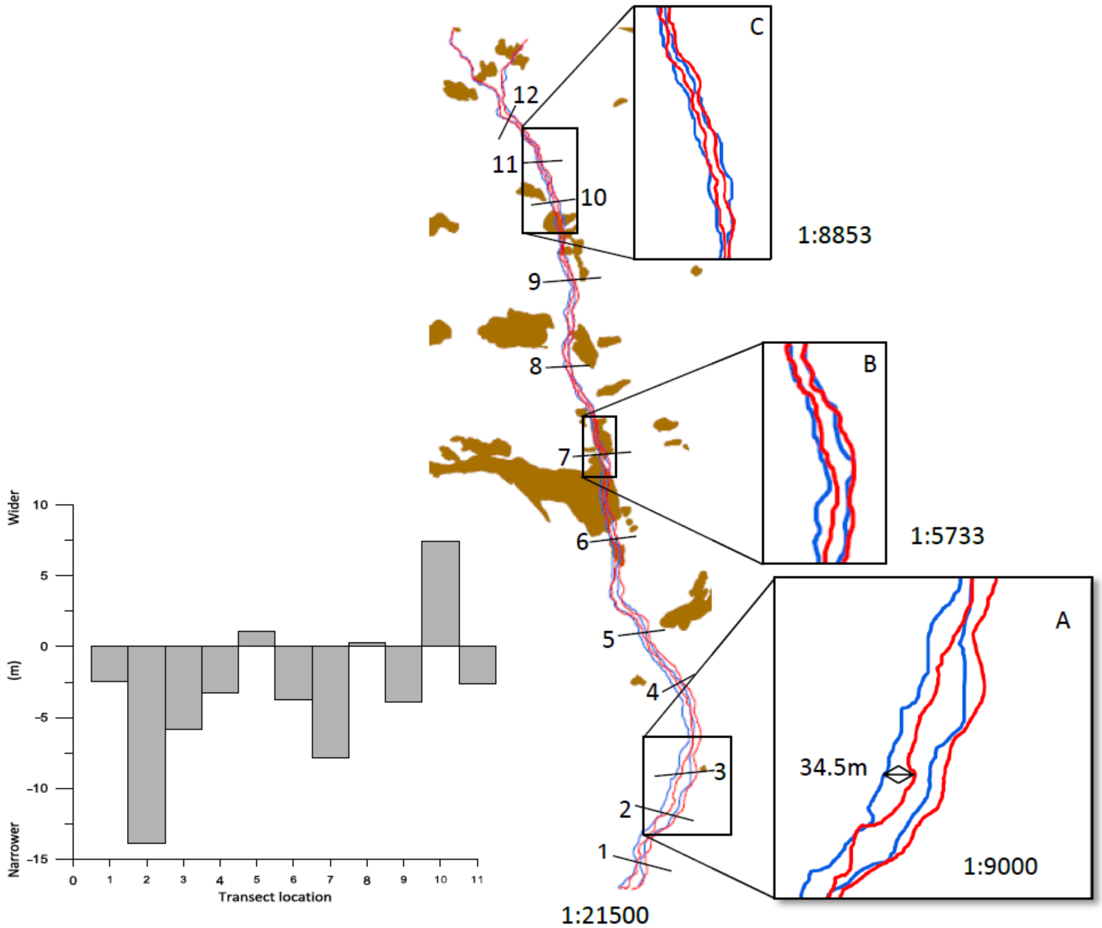


Figure 3.19 Changes in floodplain width from 1950 (blue channel) to 2009 (red channel) at transect locations. Regions along the West River where there has been large lateral changes (A), minimal changes (B), and a narrowing of the floodplain (C).

Figure 3.20 Photograph of the initial damming of the West River, following the development of ALD Big Slide during late July 2007. (Photo credit: S. Lamoureux).

Table 3.8 2010 floodplain width

	Transect location	floodplain width (m) (bank to bank)	Average (m)
Reach 3	11	15.5	14.0
	10	8	
	9	12	
	8	20.5	
Reach 2	7	16.5	17.7
	6	16	
	5	25	
	4	13.5	
Reach 1	3	13	7.8
	2	6	
	1	4.5	

3.6 Discussion

3.6.1 Impact of permafrost slope disturbances on sediment delivery

Active layer detachments along with other forms of slope disturbances have long been recognized as having the ability to change the physical characteristics of watersheds and the processes occurring within them, for example through enhanced erosion and sediment delivery (Woo & McCann 1974, Lewkowicz 2007, Bowden et al. 2008). In the summer of 2007, when an extended period of above average temperatures and heavy rainfall events resulted in favourable conditions for the formation of ALD at the CBAWO, these new disturbances covered approximately 2% of the West watershed. Even though their areal coverage was small and the disturbance occurred late in the season, they contributed to an 18% increase in the 2007 seasonal sediment yield of the West River (Lamoureux & Lafrenière 2009). It is not uncommon, in northern and Arctic environments, for small disruptions of the landscape to result in the delivery of large quantities of sediment to a river channel. Bowden et al. (2008) observed that the formation of ALD in a small watershed in the foothills of the North Slope, Alaska led to the delivery of eighteen times more sediment out of the catchment than would have normally been delivered by an adjacent and much larger (132 km^2) watershed over a 2-3 year period. Such landscape disturbances can yield elevated rates of surface wash erosion spanning multiple decades, with erosion rates averaging 1200 $g/m^2/a$ in recently disturbed regions, compared to erosion rates of 75 $g/m^2/a$ in undisturbed regions in the Fosheim Peninsula, Ellesmere Island (Lewkowicz & Kokelj 2002). However, results from this study indicate that three years following disturbance at Cape Bounty, ALD contributed only 4.7% of the seasonal sediment inflow into the West River. Hence, while they still actively contributed sediment to the main channel, the impact that they had on sediment fluxes, and hence their overall contribution to the sediment budget of the West River, decreased substantially.

The contribution of ALD to the sediment budget was most evident during the low flow 1 period in reach 2, where the major ALD (BS, PT, ALD05) are located (see Figure 3.1). At this time, the efflux of sediment from WR2 was much greater than the influx from WR3. While the ALD contributed 8358 kg of sediment, their contribution only represented 44% of the net sediment outflow from reach 2 during this period. Given that reach 2 was a net source of sediment, this implies that the additional sediment must have resulted from bed and bank erosion. Observations confirm this as blocks of soil and sediment collapsed from the bank into the river where three years prior, the river had been dammed (Figure 3.20). Active erosion and undercutting of the banks was visible where ALD made contact with the river (Figure 3.21, 3.22). It is possible that this active erosion may fully or partially account for the remaining 56% of net sediment released in reach 2 during the low flow 1 period. Such active bank erosion following landscape disturbances is not unique to northern environments, and has been documented in temperate regions as a result of various disturbances, such as volcanic eruptions and wildfires (Potyondy & Hardy 1994, Major 2004).

Active erosion of the banks and sediment release from large exposures of sediment and ground ice above reach 3 were important sources of sediment in the upper reaches of the river (Figure 3.3, 3.23). The melt from the retreating ground ice released large quantities of sediment into the channel and lined the river bed with sediment, making reach 4 the largest source of sediment compared to other inflows. This suggests that if current melting of ground ice continues, and the impact of ALD continues to decrease compared to three years prior, retreating ground ice at one location in the catchment will constitute a larger source of sediment than all of the ALD in the catchment. Further, these results suggest that under higher discharge conditions, such as those found immediately following peak discharge, river reaches subject to high levels of disturbance contribute more sediment to the river system than those with minimal disturbance.

Figure 3.21 River bank in reach 2 where active erosion is resulting in the collapse of the bank, providing a source of sediment to the West River.

Figure 3.22 River undercutting the bank in reach 2 of the West River.

Figure 3.23 East side above the river confluence, located between stations WR5 and WR6. Note the plume of sediment being released into the main channel.

3.6.2 Channel morphology controls on sediment storage

The importance of headwater tributaries as major sediment sources is increasingly being recognized (Gomi & Sidle 2003, Kasai 2006, Beel et al. 2011) as erosion from these regions can exert important impacts on downstream sediment dynamics through the passage of sediment slugs and related sediment storage (Benda & Dunne 1997b). The headwater reach 4 was consistently the largest source of sediment to the West River channel compared to all other inflows due to the retreat of ground ice located along the one bank. However, the sediment released from this reach was not entirely transported downstream, and storage of sediment in reach 3 was especially prominent (Figure 3.11, Table 3.7). Hence, storage and a wider floodplain in reach 3 created a barrier, thus decoupling sediment transport, and attenuating the impact that the erosion sources in reach 4 had on the sediment budget. This buffered the downstream reaches through sediment storage, and prevented the influence of reach 4 to fully be reflected in the sediment budget of a single year. Similarly, Madej & Ozaki (1996) found that following land development and disturbance, there was significant storage of sediment in the upper reaches of the Redwood Creek, California that buffered the lower reaches, thus preventing the introduced sediment from being immediately apparent in the basin's sediment yield. They noted that the stored sediment was gradually remobilized and passed through the system in the form of a sediment slug, thus becoming a source of sediment to lower reaches over time. Such decoupling between upper and lower reaches is not uncommon and prevents changes in the upper reaches from being reflected in the lower reaches (Shi et al. 2003, Phillips et al. 2004). Other types of disturbances can continue to impact sediment yields on much longer time scales. In a classic study, hydraulic-mining of the Yuba River, California in the mid to late 1800's resulted in the delivery of approximately 10^{12} kg of sediment to the river, half of which is believed to have been stored in downstream floodplains

(Gilbert 1917) and continues to supply sediment to the river as it is slowly remobilized (Curtis et al. 2005). Similarly, the West River has moderate channel gradients (Figure 3.2) which contribute to the high percentage of observed storage as low slopes indicate lowered stream power and hence greater opportunity for storage (Phillips & Slattery 2006). Reach 1 had the lowest slope (Figure 3.2), and stored the second largest percentage of sediment (Figure 3.11). However, differences in slope alone cannot account for the large difference in storage between reaches. Another important morphometric parameter that should be taken into consideration is the maximum valley width. Taylor and Kite (2006) compared three central Appalachian watersheds, and reported significantly greater volumes of sediment stored in the floodplain, fans, and channel bed of the river with the widest floodplain. This indicated that maximum valley width exerted a strong control over sediment transport, connectivity and hence storage. Similarly, reach 3 which had the second widest floodplain (Table 3.8), stored the largest percentage of sediment (Figure 3.11). As such, the reaches with the widest floodplain and lowest slope angles were the ones that stored the largest percentage of sediment, indicating that sediment transport and storage is dependent on a combination of morphometric parameters such as slope and valley width (Phillips & Slattery 2006, Taylor & Kite 2006), which in turn affect hydrometric parameters such as stream power and stage (Phillips & Slattery 2006).

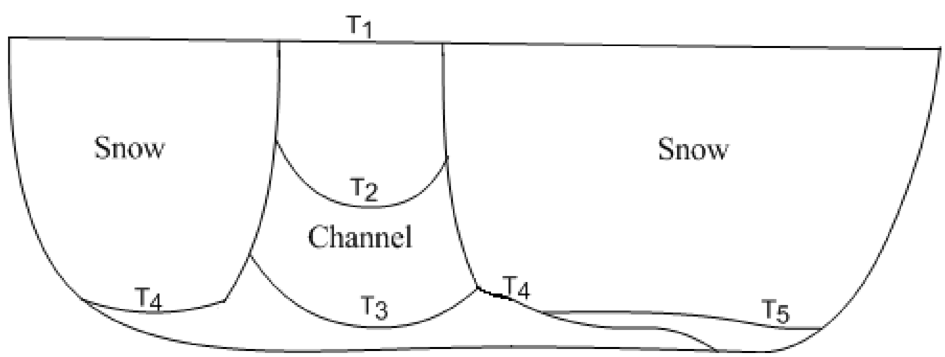
These results indicate that a substantial amount of the sediment originating from these slope disturbances is being stored in the channel. Three years following initial disturbance, the fluvial system is still actively adjusting to the changes in sediment supply caused by these permafrost disturbances. While a single year of budget work is insufficient to determine if storage and release dynamics are solely the result of the recent permafrost disturbance, the spatial patterns suggest that the disturbance has contributed to higher sediment inflows that are reflected in the budget quantities.

3.6.3 Controls of the flow regime on sediment accessibility, transport, and storage

Sediment transport and deposition in the West River began early with the onset of channelization of flow. The lateral shifting of the water on the snowpack as it incised into a channel resulted in the deposition of sediment on the surface of the snowpack where water once flowed (Figure 3.24).

Figure 3.24 Sediment deposit left behind as water above the snowpack shifted during channel formation on June 13, 2010. (Photo credit: S. Lamoureux).

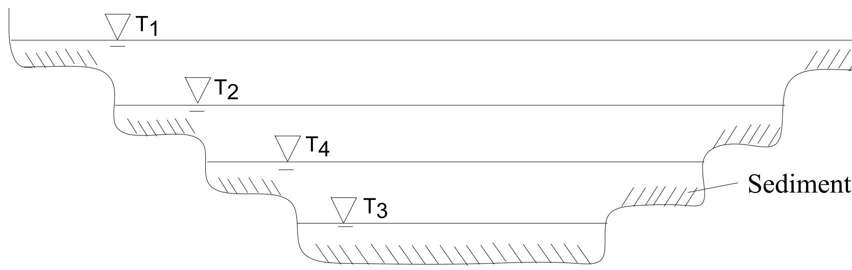
Sediment availability and accessibility was limited in the early phases of snowmelt because of frozen bed conditions, and minimal contact between channelized flow and river bed. However, as snowmelt progressed and the channel bed thawed and came into greater contact with the water, sediment availability and accessibility increased (Figure 3.25). These conditions account for the strong counter-clockwise hysteresis during the peak discharge period (Figure 3.12), which occurs when sediment is delayed as it comes from upstream sources in the catchment (Klein 1984, Lawler et al. 2006), and/or when the bed is frozen, thereby initially isolating sediment (McDonald & Lamoureux 2009). Given the frozen or thawing state of the bed early in the season, some of the available sediment likely came from external sources. The maximum suspended sediment yields on June 16 and 17 from the major tributaries through ALD, PT, BS, and CR (Figure 3.8) resulted in the delivery of large quantities of sediment to the channel when it had not completely thawed or was still isolated from flow by snow. Therefore, in addition to snowmelt, the tributaries at Cape Bounty played an important role in the sediment dynamics during the peak discharge period by providing sediment to the West River, thereby contributing to the strong counter-clockwise hysteresis (Figure 3.12).



3.25 Schematic stages of channel formation in a snow-filled channel at sequential times during snowmelt. T_1 indicates initial flow on the snow surface, followed by incision (T_2 and T_3) and lateral spreading on the channel bed (T_4 and T_5).

Although budget values are not available for the peak discharge period, this is when the majority of sediment was transported out of the system, and when flow velocities were highest (Figure 3.6). Previous studies of Arctic rivers have indicated that sediment yields are strongly influenced by snow availability (Lawson 1993, McDonald & Lamoureux 2009) and that sediment supply limits SSC more than streamflow (Braun et al. 2000). Given that most of the sediment was transported out of the system during the peak flow period (Figure 3.6), which is common for rivers in periglacial environments (Braun et al. 2000, Beylich & Gintz 2004), it is likely that the period of peak discharge flushed much of the available sediment out of the system, thereby exhausting the sediment supply once the snow had melted. This would result in the removal of the previous year's sediment accumulation and minimal storage of sediment during the peak discharge period. Similar to Arctic environments, the flushing of sediment during peak flow is common in more temperate climates, such as Western Europe, where Asselman (1999) determined that sediment in the Rhine river eroded away during the winter during high discharge conditions, and was re-deposited in the summer, thereby replenishing the channel bed for the following winter.

The counter-clockwise hysteresis during peak flow was followed by clockwise hysteresis during the recession period (Figure 3.12) as sediment supplies which had previously been deposited were exhausted following peak discharge (Lewkowicz & Wolfe 1994, Asselman 1999, Bogen 2004). Although in-channel storage increased with time (Figure 3.11), the waning stage in the river resulted in decreasing contact between flowing water and stored sediment. Therefore, previously deposited sediment was no longer accessible (Nistor & Church 2005) (Figure 3.26) and as such, the limited sediment which could be accessed and transported downstream was exhausted (Lawler et al. 2006).



3.26 Exaggerated schematic channel cross-sectional profile indicating sediment availability at various times throughout the hydrological season as stage diminishes, with T_1 representing peak discharge, T_2 low flow, T_3 baseflow, and T_4, a rainfall event that increases stage and allows access to stored sediment.

Daily SSC-Q hysteresis data support this interpretation, and similar results from the 2004, 2005, and 2006 pre-disturbance years at Cape Bounty (McDonald & Lamoureux 2009) demonstrate that daily hysteresis was dominated by clockwise relationships where SSC peaked on the rising limb (McDonald & Lamoureux 2009), indicating sediment-limited conditions and sediment that is transported short distances as it is sourced from nearby channel banks and slopes (Deel et al. 2011). In northern and semi-arid areas, this pattern occurs when previously eroded sediment is remobilized during the peak flood (Lewkowicz & Wolfe 1994, Asselman 1999), and/or when nearby sediment-rich tributaries drain into rivers (Lewkowicz & Wolfe 1994, Asselman 1999, Moliere et al. 2004). The resemblance in hysteresis dynamics across the pre- (2004, 2005, 2006) and post- (2010) disturbance years suggests that the West River channel has generally been undergoing the same dynamics of sediment accessibility for extended periods of time and that once snow is depleted, stage and wetted perimeter are important controls on sediment accessibility which drives the hysteresis relationship (Nistor & Church 2005). Such relationships are not unique to the Arctic, as daily clockwise hysteresis is also observed in rivers in temperate environments, and more urban settings (Asselman 1999, Wotling & Bouvier 2002). Decreased transport and increased storage of sediment in the later season under decreasing stage conditions indicates that a river's flow regime, particularly how it affects stage, exerts an important control over suspended sediment concentrations, sediment storage and release.

Although daily SSC-Q hysteresis was dominated by clockwise hysteresis, there were several days with counter-clockwise relationships. Such situations occur when sediment is readily available following maximum discharge, but limited beforehand (Lawler et al. 2006, McDonald & Lamoureux 2009). This could occur when sediment is delayed from upstream sources in the catchment (Klein 1984) as it is being eroded during high flows (Bogen 2004), and/or when the bed is frozen, thereby initially isolating sediment until it thaws (McDonald & Lamoureux 2009). Three of the days with counter-clockwise hysteresis coincide with the rainstorm events of July 14, 18 and 21. As there were no changes in the sediment yield from tributaries on these days, it is likely that the slight increase in stage following rainfall was sufficient to allow sediment previously isolated from flow to be remobilized with increased stage, thereby increasing sediment accessibility (Nistor & Church 2005) (Figure 3.26). Additionally, July 21, 24, and 26 exhibited counter-clockwise hysteresis and coincidently, the previous days all experienced net erosion in reach 1. It is possible that the erosion during the previous day limited the availability of sediment during the rising limb on July 21, 24 and 26, thereby resulting in counter-clockwise hysteresis, although this is not possible to demonstrate with the available data.

3.6.4 Long term sediment storage in Arctic rivers

Predicting how systems in the Arctic will respond to climate change is difficult because the changes are unlikely to be linear (IPCC 2007). Lewis and Lamoureux (2009) modeled climate trends for the 21st century at Cape Bounty and found that precipitation in the form of snow is expected to increase, findings which are consistent with predictions for other high latitude regions (Meehl et al. 2007). With greater end of winter snow cover and snow water equivalence (SWE), seasonal suspended sediment yields should increase as the former is a primary control on the latter (Forbes & Lamoureux 2005, Cockburn & Lamoureux 2008). Additionally, counter-clockwise hysteresis during the nival melt would be expected to continue as years with greater snow accumulation are predicted (McDonald & Lamoureux 2009). As end of season snowpack and total melt energy are expected to increase (Lewis & Lamoureux 2009), higher peak discharges will be reached, allowing for increased sediment transport (McDonald & Lamoureux 2009) under high flow conditions resulting from more active bank erosion when thawing begins, and increased stream power and competence.

While the large quantities of sediment stored in the West River channel is not unique to northern environments, it does demonstrate that Arctic river systems have the capacity to buffer the effects of increased sediment delivery under low flow and baseflow conditions, thereby both reducing the magnitude and also prolonging the impact that permafrost slope disturbances have on the downstream river system. The increasing percentage of sediment stored in the channel with time can be an important mechanism to offset the erosion of sediment that occurred under peak discharge conditions. More importantly, the sediment stored in the channel following peak discharge can become an important source of sediment for the seasonal sediment yield because of the predicted increase in rainfall-generated runoff (Beylich & Gintz 2004, Lewis & Lamoureux 2009). As no major late summer rainfall events occurred in 2010, the isolated sediment remained stored in the channel as it could not be remobilized and transported downstream to be included in the catchment's sediment yield. As a result, sediment storage exceeded release, and the sediment will remain in storage until it can be remobilized during a following year's peak discharge period (Smith 2008), or a major rainfall event (Dugan et al. 2009). However, if rainfall events in the Arctic were to occur more frequently as projected (IPCC 2007), there could be substantial increases in suspended sediment (Cogley & McCann 1976) which can create larger seasonal yields than those obtained from nival events (Dugan et al. 2009). Dugan et al. (2009) reported that two separate days of late summer precipitation totalling 20 mm, increased discharge and SSC in the West River and resulted in the transfer of 35% of the seasonal sediment yield, a total value greater than the yield during peak discharge. Similarly, 74 mm of rainfall in July and August of 2009 resulted in increased streamflow in the West River of the CBAWO, and generated increased sediment yields that accounted for 89% of the seasonal yield (Lewis et al. 2011). Increased sediment yields from rainfall events often result from increased hydrological connectivity between isolated sediment sources and rivers (Lewis et al. 2011), therefore highlighting the important role that stored sediment can have on sediment yields in years with large rainfall events.

In late summer, the wetted perimeter is much reduced in comparison to peak discharge conditions, therefore sediment accessibility is limited as it is stored a distance away from the active channel where water once flowed (Figure 3.26). As such, rainfall-generated runoff will become more important for sediment remobilization and transport in the later season. Given the large potential for storage in these rivers, if rainfall events become more frequent, the percentage of the seasonal sediment yield contributed from rainfall events could increase and even surpass the yield from the nival flood (Dugan et al. 2009, Lewis et al. 2011). This could offset the replenishment of the sediment lost during the initial nival flood, and cause sediment release to exceed storage, therefore tending towards a degrading river system.

In years like 2010 when no substantial rainfall events occurred, snowmelt-induced peak discharges will remain the process by which most sediment remobilization and transport occurs due to the active erosion of river banks during this time, indicating a large potential for sediment transport (Beylich & Gintz 2004, Smith 2008, Lana-Renault et al. 2011). For example, in a lowland river in Australia, sediment remained in storage for multiple years until the spring of 2005 when higher than average peak flows removed much of the sediment that had been stored in previous summers (Smith 2008). Hence, storage is an important buffering mechanism that can prolong the residence time of sediment within a channel system. The storage in the later season and potential remobilization of stored sediment during peak flows indicate a seasonal variability to the sediment dynamics of rivers (Smith 2008).

Sediment transport in rivers can be a complex process (Benda & Dunne 1997b), and due to the slow downstream migration of slugs, it can be many years (decades to centuries) before the impact of catchment disturbances fully disappear (Madej & Ozaki 1996, Benda & Dunne 1997b, Sutherland et al. 2002). There are no cases in the Arctic to compare with our results. However, after analysing annually-laminated sediment cores from Nicolay Lake, Cornwall Island, Lamoureux (2002) showed that large sediment transport events (inferred to be rainfall events) sustain high sediment yields in the following 10-30 years. Although these sustained yields were inferred to be the result of rainfall events, a similar pattern can be potentially expected following ALD formation given that they immediately increase sediment yield (Lamoureux & Lafrenière 2009) and sediment storage is a major process in the years that follow (this study).

3.7 Conclusion

Sediment budget approaches are an effective method for determining the level of connectivity within fluvial systems through the identification of river reaches undergoing net aggradation, degradation, or are graded. The West River's post-peak discharge sediment budget revealed a persistent pattern of sediment storage from various types of landscape disturbances

(ALD and bank collapse), thereby mitigating the effect of disturbance on the seasonal sediment yield.

Although the ALD at Cape Bounty are contributing less sediment to the seasonal sediment yield of the West River three years following their initial formation, they continue to act as sediment sources. A large portion of the sediment transported and introduced into the West River went into storage prior to reaching the outlet. Although this storage may act to reduce the seasonal sediment yield for the 2010 season, it may prove to be an important source of sediment in subsequent years, especially for late summer precipitation events. The stored sediment could therefore contribute to sustained periods of elevated sediment yields and prolong the impact that these disturbances have on fluvial systems.

Connectivity and the sediment delivery ratio within the system as a whole decreased with time as more sediment was stored in the channel resulting from decreased flow velocity, stage, competence and wetted perimeter. Flow velocities and floodplain width were important controls over storage as the slopes between river reaches varied minimally. Over the last few weeks of the study, as much as 85% of the transported sediment was stored in the channel (SDR= 0.15). Sediment was preferentially deposited within the channel, with the coarser material deposited in the upper reaches and the finer material in the lower reaches. With time, the sediment stored became progressively finer, highlighting the importance of river competence on sediment transport.

While slope disturbances continue to contribute sediment to the main channel, the relative amount that they are supplying has decreased since their initial formation in 2007. Given the large storage component of this fluvial system, it is evident that the West River is still actively storing and adjusting to the additional input of sediment from ALD.

Chapter 4

Conclusion

The likelihood of permafrost disturbances such as RTS and ALD are expected to increase in response to the predicted warming of the Arctic (Frey & McClelland 2009). Because these disturbances can release large quantities of sediment into rivers if hydrologically connected (Woo & McCann 1994, Bowden et al. 2008, Lamoureux & Lafrenière 2009), it is critical to understand the impact that such disturbances can have on the river systems that they are draining into and how the rivers are in turn responding to these changes. Due to the limited data on the longer term response of river channels to such disturbances, a sediment budget study was undertaken at the CBAWO in order to determine how a river subject to ALD is responding to such disturbances three years following their initiation.

Five major conclusions have been revealed from this sedimentary budget research of the West River at Cape Bounty.

1. Active layer detachments at Cape Bounty continue to play an important role in the sediment budget of the West River by contributing sediment from sources external to the river and therefore increasing the seasonal sediment yield. However, three years following their initiation, ALD are only contributing 4.7% of the measured sediment yield, compared to 18% in 2007. This indicates that while they are still contributing sediment to the West River, the impact that they have on sediment fluxes and hence their overall contribution to the sediment budget is decreasing.
2. While ALD are contributing additional sediment to the river and enhanced bank erosion is occurring in near proximity to them, the West River is actively adjusting to these

changes by storing a large percentage of this additional sediment in the post peak discharge period, resulting in the aggradation of river segments. The results from this research demonstrate that as river competence decreases later in the season due to lower flow velocities, the amount of sediment stored considerably increases, especially during baseflow when as much as 85% of all sediment introduced into the main channel is stored. This indicates that the sediment delivery ratio of the West River decreases substantially in the later season.

3. The sediment dynamics of the West River can change from one hydrological period to the next. These dynamics were dependent on the presence of ALD, especially in reach 2 which was the only reach where net erosion occurred, during the low flow 1 period. It was during this period that the three largest ALD (Big Slide, Ptarmigan and ALD05) had their peak sediment fluxes and sediment yields being discharged into the main channel, which contributed to a larger suspended sediment efflux from this reach. Additionally, because the river is still adjusting to the damming of the river caused by these disturbances three years earlier, there is increased erosion as flow velocities are elevated in this narrow section of the river.
4. Mobilization of stored sediment occurs primarily during the peak discharge period, where 91% of sediment was flushed through the system, and during late season rainfall events (Dugan et al. 2009). Because large quantities of sediment are stored in the channel following peak discharge, within perennial snowpacks, and in the floodplains isolated from the flow of late season water, many sources of sediment will only become accessible and be remobilized during the peak discharge of future years. This indicates that storage is occurring on an inter-annual basis, therefore the impact that ALD have on

the sediment budget and hence the seasonal sediment yield of Arctic rivers can be prolonged until ALD are no longer actively contributing sediment to the river and the sediment that they have contributed have been completely flushed out of the system.

5. In response to the active layer detachments that dammed the river in 2007, the West River is actively carving a new channel through the dam, resulting in increased erosion and sediment delivery to downstream reaches. In the 60 years since the aerial photographs of the West River were taken, the West River has shifted laterally in the lower reach 1, and has become narrower in width. This region of the channel is a primary site for bank erosion as the channel is carving through the narrow reach, resulting in additional sediment to the channel.

In addition to adding another year of hydrological results to the West River database, the longest running study of an Arctic river, the results from this research have contributed to a better understanding of how permafrost disturbances are impacting Arctic fluvial systems. In order to better understand sediment dynamics for the entire runoff season, future work should concentrate on the sediment budget during the peak discharge period. This work would determine whether storage or erosion is predominant during this period, thereby indicating if for the duration of the entire runoff season, the West River is subject to net aggradation, degradation, or is graded. Furthermore, research will be needed to determine the timescale associated with the temporary storage of in-stream sediment and determining the type of sediment being transported by these disturbances in order to understand if they are mainly flushed out of the system because they are fine-grained, or if they contribute to the stored sediment by being coarser-grained. By providing the initial measurements of a sediment budget for the West River, this research represents an important baseline that can be used for long-term monitoring of sediment budgets to determine the timescale associated with the impacts of ALD at Cape Bounty.

References

ACIA (2005), Arctic Climate Impact Assessment, Scientific Report, Cambridge University Press, Cambridge.

Arnell, N. (2002), Hydrology and global environmental change., 346 pp., Pearson Education Limited, Essex.

ASCE (1974), Generalized discharge relations for cutthroat flumes, Journal of the Irrigation and Drainage Division, 98, 569-583.

Asselman, N. E. M. (1999), Suspended sediment dynamics in a large drainage basin: the River Rhine, Hydrological Processes, 13(10), 1437-1450.

Beel, C. R., J. F. Orwin, and P. G. Holland (2011), Controls on slope-to-channel fine sediment connectivity in a largely ice-free valley, Hoophorn Stream, Southern Alps, New Zealand, Earth Surface Processes and Landforms, 36(7), 981-994.

Benda, L., and T. Dunne (1997), Stochastic forcing of sediment supply to channel networks from landsliding and debris flow, Water Resources Research, 33(12), 2849-2863.

Benda, L., and T. Dunne (1997), Stochastic forcing of sediment routing and storage in channel networks, Water Resources Research, 33(12), 2865-2880.

Beschta, R. L. (1987), Conceptual models of sediment transport in streams., in Sediment transport in gravel-bed rivers, edited by C. R. Thorne, et al., pp. 387-420, John Wiley & Sons, Canada.

Beylich, A. A., and D. Gintz (2004), Effects of high-magnitude/low-frequency fluvial events generated by intense snowmelt or heavy rainfall in Arctic periglacial environments in northern Swedish Lapland and northern Siberia, Geografiska Annaler Series a-Physical Geography, 86A(1), 11-29.

Bogen, J. (2004), Erosion and sediment yield in the Atna River basin, Hydrobiologia, 521(1-3), 35-47.

Bowden, W. B., M. N. Gooseff, A. Malser, A. Green, B. J. Peterson, and J. Bradford (2008), Sediment and nutrient delivery from thermokarst features in the foothills of the North Slope, Alaska: Potential impacts on headwater stream ecosystems, Journal of Geophysical Research-Biogeosciences, 113(G2), DOI:10.1029/2007JG000470.

Braun, C., D. R. Hardy, R. S. Bradley, and M. J. Retelle (2000), Streamflow and suspended sediment transfer to Lake Sophia, Cornwallis Island, Nunavut, Canada, Arctic, Antarctic, and Alpine Research, 32(4), 456-465.

Carey, S. K., and W. L. Quinton (2004), Evaluating snowmelt runoff generation in a discontinuous permafrost catchment using stable isotope, hydrochemical and hydrometric data, Nordic Hydrology, 35(4-5), 309-324.

Carey, S. K., and J. W. Pomeroy (2009), Progress in Canadian snow and frozen ground hydrology, 2003-2007, Canadian Water Resources Journal, 34(2), 127-138.

Carling, P. A. (1992), In-stream hydraulics and sediment transport, in The rivers handbook: hydrological and ecological principles, edited by P. Calow and G. E. Petts, pp. 101-125, Blackwell Scientific Publications, Oxford.

Church, M. (1974), Hydrology and permafrost with reference to northern North America., paper presented at Proceedings of Workshop Seminar on Permafrost Hydrology, Canadian National Committee, International Hydrological Decade., Ottawa, 7-20.

Church, M. (1992), Channel morphology and typology, in The river's handbook: hydrological and ecological principles, edited by P. Calow and G. E. Petts, pp. 126-143, Blackwell Scientific Publications, Oxford.

Church, M., M. A. Hassan, and J. F. Wolcott (1998), Stabilizing self-organized structures in gravel-bed stream channels: Field and experimental observations, Water Resources Research, 34(11), 3169-3179.

Cockburn, J. M. H., and S. F. Lamoureux (2008), Hydroclimate controls over seasonal sediment yield in two adjacent High Arctic watersheds, Hydrological Processes Hydrological Processes, 22(12), 2013-2027.

Cogley, J. G., and S. B. McCann (1976), An exceptional storm and its effects in the Canadian High Arctic, Arctic and Alpine Research, 8(1), 105-110.

Couper, P. R., and I. P. Maddock (2001), Subaerial river bank erosion processes and their interaction with other bank erosion mechanisms on the river arrow, Warwickshire, UK, Earth Surface Processes and Landforms, 26(6), 631-646.

Cui, Y. T., G. Parker, T. E. Lisle, J. Gott, M. E. Hansler-Ball, J. E. Pizzuto, N. E. Allmendinger, and J. M. Reed (2003), Sediment pulses in mountain rivers: 1. Experiments, Water Resources Research, 39(9), DOI: 10.1029/2002WR001803.

Curtis, J. A., L. E. Flint, C. N. Alpers, and S. M. Yarnell (2005), Conceptual model of sediment processes in the upper Yuba River watershed, Sierra Nevada, CA, Geomorphology, 68(3-4), 149-166.

Dingman, S. L. (1994), Physical hydrology, 575 pp., Macmillan Publishing Company, New York.

Drake, T. G., and J. Calantoni (2001), Discrete particle model for sheet flow sediment transport in the nearshore, Journal of Geophysical Research-Oceans, 106(C9), 19859-19868.

Dugan, H. A., S. F. Lamoureux, M. J. Lafrenière, and T. Lewis (2009), Hydrological and sediment yield response to summer rainfall in a small high Arctic watershed, Hydrological Processes, 23(10), 1514-1526.

Environment Canada, 2011, National climate data & information archice, Fredericton, New Brunswick, available from: http://www.climate.weatheroffice.gc.ca/Welcome_e.html

Ewing, R. (1996), Postfire suspended sediment from Yellowstone National Park, Wyoming, Water Resources Bulletin, 32(3), 605-627.

Forbes, A. C., and S. F. Lamoureux (2005), Climatic controls on streamflow and suspended sediment transport in three large middle Arctic catchments, Boothia Peninsula, Nunavut, Canada, Arctic, Antarctic, and Alpine Research, 37(3), 304-315.

French, H. M. (2007), The periglacial environment, 3 ed., 478 pp., Wiley, Chichester.

Frey, K. E., and J. W. McClelland (2009), Impacts of permafrost degradation on Arctic river biogeochemistry, Hydrological Processes, 23(1), 169-182.

Fryirs, K. A., G. J. Brierley, N. J. Preston, and M. Kasai (2007), Buffers, barriers and blankets: The (dis)connectivity of catchment-scale sediment cascades, Catena, 70(1), 49-67.

Gilbert, G. K. (1917), Hydraulic-mining debris in the Sierra Nevada, United Sates Geological Survey Professional Paper 105, 216 pp. Washington.

Gomi, T., and R. C. Sidle (2003), Bed load transport in managed steep-gradient headwater streams of southeastern Alaska, Water Resources Research, 39(12).

Gomi, T., R. C. Sidle and D. N. Swanton (2004), Hydrogeomorphic linkages of sediment transport in headwater streams, Maybeso Experimental Forest, southeast Alaska, Hydrological Processes, 18(4), 667-683.

Gordon, N. D., T. A. McMahon, B. L. Finlayson, C. J. Gippel, and R. J. Nathan (1994), Stream hydrology: an introduction for ecologists, 444 pp., John Wiley, Chichester.

Griffiths, G. A. (1979), Recent sedimentation history of the Waimakariri River, New Zealand, Journal of Hydrology, 18, 6-28.

Harris, C., and A. G. Lewkowicz (1993), Form and internal structure of active-layer detachment slides, Fosheim Peninsula, Ellesmere Island, Northwest-Territories, Canada, Canadian Journal of Earth Sciences, 30(8), 1708-1714.

Harry, D. G., and K. L. MacInnes (1988), The effect of forest fires on permafrost terrain stability, Little Chicago-Travaillant Lake area, Mackenzie Valley N. W. T., edited by G. S. O. Canada.

Harvey, A. M. (2001), Coupling between hillslopes and channels in upland fluvial systems: implications for landscape sensitivity, illustrated from the Howgill Fells, northwest England, Catena, 42(2-4), 225-250.

Hey, R. D. (1987), River dynamics, flow regime and sediment transport., in Sediment transport in gravel-bed rivers., edited by C. R. Thorne, J. C. Bathurst and R. D. Hey, pp. 17-42, John Wiley & Sons.

Hooke, J. (2003), Coarse sediment connectivity in river channel systems: a conceptual framework and methodology, Geomorphology, 56(1-2), 79-94.

IPCC (2007), Intergovernmental Panel on Climate Change, Cambridge University Press, Cambridge.

Kane, D. L., L. D. Hinzman, C. S. Benson, and G. E. Liston (1991), Snow hydrology of a headwater Arctic basin 1. Physical measurements and process studies, Water Resources Research, 27(6), 1099-1109.

Kasai, M. (2006), Channel processes following land use changes in a degrading steep headwater stream in North Island, New Zealand, Geomorphology, 81(3-4), 421-439.

Klein, M. (1984), Anti clockwise hysteresis in suspended sediment concentration during individual storms: Holbeck catchment; Yorkshire, England, Catena, 11(2-3), 251-257.

Kleinhans, M. G. (2010), Sorting out river channel patterns, Progress in Physical Geography, 34(3), 287-326.

Knighton, D. (1998), Fluvial forms & processes: a new perspective, Oxford University Press, New York.

Kokelj, S. V., and A. G. Lewkowicz (1998), Long-term influence of active-layer detachment sliding on permafrost slope hydrology, Hot Weather Creek, Ellesmere Island, Canada., Permafrost - Seventh International Conference (Proceedings), Yellowknife (Canada), Collection Nordicana, 55, 583-589.

Lamoureux, S. F., D. M. McDonald, J. M. H. Cockburn, M. J. Lafrenière, D. Atkinson, and P. Treitz (2006), An incidence of multi-year sediment storage on channel snowpack in the Canadian High Arctic, Arctic, 59(4), 381-390.

Lamoureux, S. F., and M. J. Lafreniere (2009), Fluvial impact of extensive active layer detachments, Cape Bounty, Melville Island, Canada, Arctic, Antarctic, and Alpine Research, 41(1), 59-68.

Lana-Renault, N., B. Alvera, and J. M. García-Ruiz (2011), Runoff and sediment transport during the snowmelt period in a Mediterranean High-Mountain Catchment, Arctic, Antarctic, and Alpine Research, 43(2), 213-222.

Lantuit, H., and W. H. Pollard (2008), Fifty years of coastal erosion and retrogressive thaw slump activity on Herschel Island, southern Beaufort Sea, Yukon Territory, Canada, Geomorphology, 95(1-2), 84-102.

Lawler, D. M. (1992), Process dominance in bank erosion systems, in Lowland floodplain rivers: Geomorphological perspectives, edited by P. A. Carling and G. E. Petts, pp. 117-143, Wiley, Chichester.

Lawler, D. M., G. E. Petts, I. D. L. Foster, and S. Harper (2006), Turbidity dynamics during spring storm events in an urban headwater river system: The Upper Tame, West Midlands, UK, Science of the Total Environment, 360(1-3), 109-126.

Lawrence, D. M., and A. G. Slater (2005), A projection of severe near-surface permafrost degradation during the 21st century, Geophysical Research Letters, 32(24), 5.

Lawson, D. E. (1993), Glaciohydrologic and glaciohydraulic effects on runoff and sediment yield in glacierized basins, U.S. Army Corps of Engineers, Cold Regions Research and Engineering Laboratory, Monograph.

Leibman, M. O. (1995), Cryogenic landslides on the Yamal Peninsula, Russia: Preliminary observations, Permafrost and Periglacial Processes, 6, 259-264.

Lewis, T., and S. F. Lamoureux (2009), Twenty-first century discharge and sediment yield predictions in a small high Arctic watershed, Global and Planetary Change, 71(1-2), 27-41.

Lewis, T., M. J. Lafrenière, and S. F. Lamoureux (2011), Hydrochemical and sedimentary responses of paired High Arctic watersheds to unusual climate and permafrost disturbance, Cape Bounty, Melville Island, Canada, Hydrological Processes.

Lewkowicz, A. G. (1987), Headwall retreat of ground-ice slumps, Banks-Island, Northwest-Territories, Canadian Journal of Earth Sciences, 24(6), 1077-1085.

Lewkowicz, A. G. (1990), Morphology, frequency and magnitude of active-layer detachment slides, Fosheim Peninsula, Ellesmere Island, N.W.T., paper presented at Permafrost-Canada. Proceedings of the Fifth Canadian Permafrost Conference, Québec City.

Lewkowicz, A. G., and P. M. Wolfe (1994), Sediment transport in Hot Weather Creek, Ellesmere Island, N.W.T., Canada, 1990-1991, Arctic and Alpine Research, 26(3), 213-226.

Lewkowicz, A. G., and S. V. Kokelj (2002), Slope sediment yield in arid lowland continuous permafrost environments, Canadian Arctic Archipelago, Catena-Giessen Then Amsterdam, 46(4), 261-283.

Lewkowicz, A. G., and C. Harris (2005), Morphology and geotechnique of active-layer detachment failures in discontinuous and continuous permafrost, northern Canada, Geomorphology, 69(1-4), 275-297.

Lewkowicz, A. G. (2007), Dynamics of active-layer detachment failures, Fosheim Peninsula, Ellesmere Island, Nunavut, Canada, Permafrost and Periglacial Processes, 18(1), 89-103.

Lisle, T. E., Y. Cui, G. Parker, J. E. Pizzuto, and A. M. Dodd (2001), The dominance of dispersion in the evolution of bed material waves in gravel-bed rivers, Earth Surface Processes and Landforms, 26(13), 1409-1420.

Madej, M. A., and V. Ozaki (1996), Channel response to sediment wave propagation and movement, Redwood Creek, California, USA, Earth Surface Processes and Landforms, 21(10), 911-927.

Major, J. J. (2004), Post eruption suspended sediment transport at Mount St. Helens: Decadal-scale relationships with landscape adjustments and river discharges, Journal of Geophysical Research-Earth Surface, 109(F1), DOI:10.1029/2002JF000010.

McCann, S. B., P. J. Howarth, and J. G. Cogley (1972), Fluvial processes in a periglacial environment, Transactions of the Institute of British Geographers, 55, 69-82.

McDonald, D. M., and S. F. Lamoureux (2009), Hydroclimatic and channel snowpack controls over suspended sediment and grain size transport in a High Arctic catchment, Earth Surface Processes and Landforms, 34(3), 424-436.

McNamara, J. P., D. L. Kane, and L. D. Hinzman (1997), Hydrograph separations in an Arctic watershed using mixing model and graphical techniques, Water Resources Research, 33(7), 1707-1719.

Meehl, G. A., et al. (2007), Global climate projections. In: Qin, D., et al. (Ed.). Climate change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change., Cambridge University Press, Cambridge.

Miller, M. C., I. N. McCave, and P. D. Komar (1977), Threshold of sediment motion under unidirectional currents, Sedimentology, 24(4), 507-527.

Moliere, D. R., K. G. Evans, M. J. Saynor, and W. D. Erskine (2004), Estimation of suspended sediment loads in a seasonal stream in the wet-dry tropics, Northern Territory, Australia, Hydrological Processes, 18(3), 531-544.

Nanson, G. C., and A. D. Knighton (1996), Anabranching rivers: Their cause, character and classification, Earth Surface Processes and Landforms, 21(3), 217-239.

Nelson, J. M., R. L. Shreve, S. R. McLean, and T. G. Drake (1995), Role of near-bed turbulence structure in bed-load transport and bed form mechanics, Water Resources Research, 31(8), 2071-2086.

Nicholas, A. P., P. J. Ashworth, M. J. Kirkby, M. G. Macklin, and T. Murray (1995), Sediment slugs: Large-scale fluctuations in fluvial sediment transport rates and storage volumes, Progress in Physical Geography, 19(4), 500-519.

Nistor, C. J., and M. Church (2005), Suspended sediment transport regime in a debris-flow gully on Vancouver Island, British Columbia, Hydrological Processes, 19(4), 861-885.

Orwin, J. F., S. F. Lamoureux, J. Warburton, and A. Beylich (2010), A Framework for characterizing fluvial sediment fluxes from source to sink in cold environments, Geografiska Annaler Series a-Physical Geography, 92A(2), 155-176.

Paiement-Paradis, G., G. Marquis, and A. Roy (2011), Effects of turbulence on the transport of individual particles as bedload in a gravel-bed river, Earth Surface Processes and Landforms, 36(1), 107-116.

Phillips, J. D., M. C. Slattery, and Z. A. Musselman (2004), Dam-to-delta sediment inputs and storage in the lower Trinity River, Texas, Geomorphology, 62(1-2), 17-34.

Phillips, J. D., and M. C. Slattery (2006), Sediment storage, sea level, and sediment delivery to the ocean by coastal plain rivers, Progress in Physical Geography, 30(4), 513-530.

Potyondy, J. P., and T. Hardy (1994), Use of pebble counts to evaluate fine sediment increase in stream channels, Water Resources Bulletin, 30(3), 509-520.

Prosser, I. P., I. D. Rutherfurd, J. M. Olley, W. J. Young, P. J. Wallbrink, and C. J. Moran (2001), Large-scale patterns of erosion and sediment transport in river networks, with examples from Australia, Marine and Freshwater Research, 52(1), 81-99.

Restrepo, J. D., P. Zapata, J. M. Díaz, J. Garzón-Ferreira, and C. B. García (2006), Fluvial fluxes into the Caribbean Sea and their impact on coastal ecosystems: The Magdalena River, Colombia, Global and Planetary Change, 50(1-2), 33-49.

Rowland, J. C., C. E. Jones, G. Altmann, R. Bryan, B. T. Crosby, G. L. Geernaert, L. D. Hinzman, D. L. Kane et al. (2010), Arctic landscapes in transition: Responses to thawing permafrost, Eos, 91(26), 229-230.

Schumm, S. A. (1985), Patterns of alluvial rivers, Annual Review of Earth and Planetary Sciences, 13, 5-27.

Schumm, S. A., et al. (1987), Experimental fluvial geomorphology, Wiley, Chichester.

She, K., L. Trim, and D. Pope (2005), Fall velocities of natural sediment particles: a simple mathematical presentation of the fall velocity law, Journal of Hydraulic Research, 43(2), 189-195.

Shi, C. X., D. D. Zhang, and L. You (2003), Sediment budget of the Yellow River delta, China: the importance of dry bulk density and implications to understanding of sediment dispersal, Marine Geology, 199(1-2), 13-25.

Smith, H. G. (2008), Estimation of suspended sediment loads and delivery in an incised upland headwater catchment, south-eastern Australia, Hydrological Processes, 22(16), 3135-3148.

Sutherland, D. G., M. H. Ball, S. J. Hilton, and T. E. Lisle (2002), Evolution of a landslide-induced sediment wave in the Navarro River, California, Geological Society of America Bulletin, 114(8), 1036-1048.

Taylor, S. B., and J. S. Kite (2006), Comparative geomorphic analysis of surficial deposits at three central Appalachian watersheds: Implications for controls on sediment-transport efficiency, Geomorphology, 78(1-2), 22-43.

Thorne, C. R. (1982), Processes and mechanisms of river bank erosion, in Gravel-bed rivers, edited by R. D. Hey, et al., pp. 227-259, Wiley, Chichester.

van den Berg, J. H. (1995), Prediction of an alluvial channel pattern of perennial rivers, Geomorphology, 12(4), 259-279.

Walling, D. E. (1983), The sediment delivery problem, Journal of Hydrology, 65(1-3), 209-237.

Walling, D. E. (1988), Erosion and sediment yield research - Some recent perspectives, Journal of Hydrology, 100(1-3), 113-141.

Walling, D. E., and B. W. Webb (1992), Water quality. I. Physical characteristics, in The rivers handbook: hydrological and ecological principles, edited by P. Calow and G. E. Petts, pp. 48-72, Blackwell Scientific Publications, Oxford.

Walling, D. E., P. N. Owens, and G. J. L. Leeks (1998), The role of channel and floodplain storage in the suspended sediment budget of the River Ouse, Yorkshire, UK, Geomorphology, 22(3-4), 225-242.

Walling, D. E., P. N. Owens, J. Carter, G. J. L. Leeks, S. Lewis, A. A. Meharg, and J. Wright (2003), Storage of sediment-associated nutrients and contaminants in river channel and floodplain systems, Applied Geochemistry, 18(2), 195-220.

Woo, M. K., and S. B. McCann (1974), Climatic variability, climatic change, runoff, and suspended sediment regimes in Northern Canada, Physical Geography, 15(4), 201-226.

Woo, M. K., and J. Sauriol (1980), Channel development in snow-filled valleys, Resolute, N. W. T., Canada, Geografiska Annaler. Series a- Physical Geography, 62(2), 37-56.

Woo, M. K. (1983), Hydrology of a drainage basin in the Canadian High Arctic, Annals of the Association of American Geographers, 73(4), 577-596.

Woo, M. K. (1986), Permafrost Hydrology in North America, Atmosphere-Ocean, 24(3), 201-234.

Woo, M. K., and S. B. McCann (1994), Climatic variability, climatic change, runoff, and suspended sediment regimes in northern Canada, Physical Geography, 15(3), 201-226.

Woo, M. K., D. L. Kane, S. K. Carey, and D. Yang (2008), Progress in permafrost hydrology in the new millennium, Permafrost and Periglacial Processes, 19(2), 237-254.

Wotling, G., and C. Bouvier (2002), Impact of urbanization on suspended sediment and organic matter fluxes from small catchments in Tahiti, Hydrological Processes, 16(9), 1745-1756.

Appendix A

A. 1 Daily sediment yields at gauging stations WR1-5.

	Sediment yield (kg)					
	date	W1	W2	W3	W4	W5
Recession flow 1	22/06/2010	5597.4	2186.0	2408.7	1053.5	669.0
	23/06/2010	5144.7	5997.8	2206.1	2150.9	944.6
	24/06/2010	4218.0	5398.3	1014.3	1655.2	1072.4
	25/06/2010	16001.2	14767.4	4697.6	4387.6	2525.6
	26/06/2010	4641.8	8745.3	3884.7	2749.9	1375.8
	27/06/2010	1789.7	3244.3	1664.6	1555.5	888.7
	28/06/2010	2231.2	2276.6	792.7	1895.3	841.0
	29/06/2010	1435.9	2105.3	1189.5	2776.8	853.9
	30/06/2010	681.5	1111.0	765.5	1669.8	480.7
Recession flow 2	01/07/2010	550.2	820.8	686.8	626.2	569.9
	02/07/2010	432.3	746.2	619.1	1061.6	529.7
	03/07/2010	232.4	487.1	513.4	839.4	439.7
	04/07/2010	Data not available				
	05/07/2010	301.5	547.0	626.5	1835.2	474.2
	06/07/2010	169.0	215.1	368.4	400.7	646.4
	07/07/2010	320.0	343.4	757.8	1524.9	259.3
	08/07/2010	123.7	123.8	159.6	278.3	142.9
	09/07/2010	59.6	214.4	162.6	291.4	307.5
Baseflow	10/07/2010	27.1	62.3	46.5	518.3	65.3
	11/07/2010	53.3	98.4	86.1	247.8	38.5
	12/07/2010	55.4	125.6	88.0	133.0	32.4
	13/07/2010	53.5	71.7	61.0	175.1	19.5
	14/07/2010	33.3	52.2	45.1	64.5	52.9
	15/07/2010	20.2	74.1	49.3	78.1	22.4
	16/07/2010	22.1	86.2	30.0	74.7	25.1
	17/07/2010	23.6	45.9	18.8	71.0	5.9
	18/07/2010	Data not available				
	19/07/2010	17.0	28.1	20.1	134.4	5.4
	20/07/2010	24.0	19.1	8.9	45.5	6.9
	21/07/2010	27.5	36.1	32.6	31.6	12.2
	22/07/2010	16.6	17.0	17.3	26.3	7.2
	23/07/2010	24.5	7.2	0.7	4.7	2.3
	24/07/2010	18.2	18.2	1.9	13.9	1.5
	25/07/2010	12.6	8.9	3.8	41.2	2.3
	26/07/2010	13.6	15.7	3.4	53.4	3.1
	27/07/2010	15.0	59.7	25.4	223.8	3.5
	28/07/2010	5.8	8.9	1.8	20.3	1.5

A. 2 Daily sediment yields at the subcatchment gauging stations.

		Sediment yield (kg)				
	date	ALD05	Goose	PT	BS	CR
Recession flow 1	22/06/2010	138.1	0.4	473.2	173.2	3.6
	23/06/2010	293.5	1.1	580.0	315.6	6.5
	24/06/2010	594.4	0.5	521.6	506.9	7.9
	25/06/2010	594.1	0.5	377.7	1071.8	38.9
	26/06/2010	598.2	0.4	71.9	321.8	16.8
	27/06/2010	387.3	1.2	55.7	271.3	6.0
	28/06/2010	190.6	0.2	20.4	280.5	7.7
	29/06/2010	120.9	0.1	7.3	242.7	3.9
	30/06/2010	48.8	0.0	0.0	99.9	2.8
Recession flow 2	01/07/2010	81.6	0.0	0.0	49.1	3.3
	02/07/2010	33.5	0.7	0.0	54.5	6.2
	03/07/2010	23.4	0.4	0.0	25.6	3.6
	04/07/2010	Data not available				
	05/07/2010	0.0	0.6	0.0	25.6	2.5
	06/07/2010	0.0	3.4	0.0	28.3	4.1
	07/07/2010	0.0	1.6	0.0	57.5	6.4
	08/07/2010	0.0	1.0	0.0	27.4	2.9
	09/07/2010	0.0	1.7	0.0	25.3	2.0
Baseflow	10/07/2010	0.0	1.8	0.0	48.0	0.7
	11/07/2010	0.0	1.8	0.0	66.9	1.4
	12/07/2010	0.0	1.7	0.0	64.4	0.0
	13/07/2010	0.0	1.4	0.0	39.7	0.0
	14/07/2010	0.0	0.0	0.0	19.7	0.0
	15/07/2010	0.0	0.7	0.0	25.8	0.0
	16/07/2010	0.0	0.0	0.0	77.3	0.0
	17/07/2010	0.0	0.0	0.0	74.7	0.0
	18/07/2010	Data not available				
	19/07/2010	0.0	0.0	0.0	27.4	0.0
	20/07/2010	0.0	0.0	0.0	19.6	0.0
	21/07/2010	0.0	1.6	0.0	32.4	7.8
	22/07/2010	0.0	0.4	0.0	21.5	0.0
	23/07/2010	0.0	0.0	0.0	81.7	0.0
	24/07/2010	0.0	0.0	0.0	69.1	0.0
	25/07/2010	0.0	0.0	0.0	34.0	0.0
	26/07/2010	0.0	0.0	0.0	39.9	0.0
	27/07/2010	0.0	0.0	0.0	89.3	0.0
	28/07/2010	0.0	0.0	0.0	15.4	0.0

A. 3 Sediment yield at all gauging stations for the different hydrological periods.

	Sediment yield (kg)		
Station	Low flow 1 (June 22-June 30)	Low flow 2 (July 1-9)	Baseflow (July 10-28)
5	9652	3370	308
4	19895	6858	1958
CR	94	31	10
3	18624	3894	541
BS	3284	293	847
PT	2108	0	0
ALD05	2966	139	0
2	45832	3498	835
G	4.42	10	9
1	41741	2189	463