POST-FIRE DYNAMICS OF A GRAVEL BED STREAM: FISHTRAP CREEK, BRITISH COLUMBIA

by

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ABSTRACT

In August 2003, the McLure Fire burned through several watersheds north of Kamloops, including Fishtrap Creek. The burn was extensive, killing almost all the trees in the riparian area. Given our current knowledge, it is not possible to accurately predict channel response to wildfire, the rate of change, or the time scale at which response and recovery may be expected to occur. The purpose of this study is to document the timing and magnitude of initial changes in channel morphology, peak flows, sediment supply, and sediment mobility in Fishtrap Creek in the aftermath of the McLure Fire. We take an empirical approach that combines stream channel monitoring with physically based data analysis. The results of this study have contributed to our understanding of the response of gravel bed rivers to severe vegetation disturbance and have produced an estimate of the timescale at which significant channel transformations may begin to occur.

During the summer of 2006 (3 years after the fire), we conducted an intensive field study at Fishtrap Creek. Results indicate that Fishtrap Creek is in a state of transition, triggered by the loss of bank strength and subsequent sediment input to the stream channel. Cross-sectional bed elevation surveys conducted annually from 2004 to 2006 have documented changes in channel form. In localized areas, up to 42 cm of aggradation has been observed; aggradation has typically occurred on the channel bars and behind large woody debris. Avulsions and overbank flooding have also occurred as aggradation forced peak flow stage height higher. Analysis of the relationship between stage height and discharge at a variety of locations indicate that the vast majority of aggradation occurred during the first peak in the 2006 hydrograph. Magnetic tracer stone analysis indicates that sediment mobility was high during the 2006 freshet. The burial depth distribution of tracers suggests substantial scour and a thorough mixing of the active layer. The current trend of channel adjustment is expected to continue until the root systems of emerging vegetation begin to stabilize the banks.
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LIST OF SYMBOLS

\( A \) \hspace{1cm} \text{channel cross-sectional area at bankfull discharge (m)}
\( d \) \hspace{1cm} \text{water depth at bankfull discharge (m)}
\( D_n \) \hspace{1cm} \text{particle size associated with the nth percentile of surface bed material (mm)}
\( D_{n(sub)} \) \hspace{1cm} \text{particle size associated with the nth percentile of the subsurface bed material (mm)}
\( g \) \hspace{1cm} \text{acceleration due to gravity (m sec}^{-2}\text{)}
\( h \) \hspace{1cm} \text{stage height}
\( n \) \hspace{1cm} \text{Manning's roughness coefficient}
\( P \) \hspace{1cm} \text{channel wetted perimeter at bankfull discharge (m)}
\( Q \) \hspace{1cm} \text{discharge (m}^3\text{ s}^{-1}\text{)}
\( Q_{bf} \) \hspace{1cm} \text{bankfull discharge (m}^3\text{ s}^{-1}\text{)}
\( q_s \) \hspace{1cm} \text{sediment transport rate (kg s}^{-1}\text{)}
\( Q_s \) \hspace{1cm} \text{bulk sediment transport rate (kg event}^{-1}\text{)}
\( R \) \hspace{1cm} \text{hydraulic radius - A/P at bankfull discharge}
\( R_b \) \hspace{1cm} \text{radius of the curvature of the channel centerline}
\( S \) \hspace{1cm} \text{energy slope, estimated from water surface slope (m m}^{-1}\text{)}
\( v \) \hspace{1cm} \text{mean water velocity (m s}^{-1}\text{)}
\( V \) \hspace{1cm} \text{volume of sediment (m}^3\text{)}
\( W \) \hspace{1cm} \text{width of the water surface at bankfull discharge (m)}
\( \alpha \) \hspace{1cm} \text{constant used in the Meyer-Peter and Müller equation}
\( \gamma \) \hspace{1cm} \text{the unit weight of water (kN m}^3\text{)}
\( \gamma_s \) \hspace{1cm} \text{the unit weight of sediment (kN m}^3\text{)}
\( \rho \) \hspace{1cm} \text{density of water (kg m}^3\text{)}
\( \rho_s \) \hspace{1cm} \text{density of sediment (kg m}^3\text{)}
\( \rho_b \) \hspace{1cm} \text{density of bulk bed material (kg m}^3\text{)}
\( \tau \) \hspace{1cm} \text{reach average boundary shear stress (N m}^2\text{)}
\( \tau_{crit} \) \hspace{1cm} \text{critical shear stress for a given grain size (N m}^2\text{)}
\( \tau^* \) \hspace{1cm} \text{critical dimensionless shear stress for the median surface grain size (Shield's parameter)}
\( \Delta \) \hspace{1cm} \text{change in a given variable}
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Chapter 1

1. WILDFIRE AND THE STABILITY OF GRAVEL BED STREAMS

1.1. Introduction

The summer of 2003 in southern British Columbia was an unusually hot and dry summer that followed one of the driest winters on record. Unprecedented low summer flows were recorded for several streams in the Kamloops region (Doyle, 2004). As a result, the 2003 fire season was marked by several large, destructive forest fires. Over 104,000 ha were burned in the Kamloops Fire Region alone (Scott and Pike, 2003).

On July 30, 2003, the McLure fire ignited. High fuel loads, dry initial conditions, and strong winds created a high intensity fire that was exceptionally difficult to contain. The fire’s perimeter quickly grew to hundreds of kilometres over the following days, ultimately destroying the town of Louis Creek and parts of Barriere, B.C. The Tolko sawmill in Louis Creek was completely destroyed. The regional map of southwestern British Columbia in Figure 1 shows the location of the McLure Fire limits.

Figure 1. Regional map of Southwestern British Columbia indicating the location of the McLure Fire. Baseline mapping data from ESRI Data and Maps (2002).
The McLure Fire was a high-intensity stand-replacing crown fire. Several watersheds north of Kamloops were severely burned, including Fishtrap Creek. The fire burned an estimated 75% of the watershed with the burn of the riparian area being extensive, killing virtually all the vegetation. The map in Figure 2 identifies the location of Fishtrap Creek Watershed within the McLure Fire limits.

![Figure 2. Area map illustrating the McLure Fire limits and the burnt area of Fishtrap Creek Watershed. Baseline mapping data obtained from British Columbia Ministry of Sustainable Resource Management (MSRM) Terrain Resource Information Mapping (TRIM) data.](image)

The effects of wildfires on forest ecosystems are of primary concern for researchers and resource managers in North America. There have been numerous studies conducted to analyze the effects of fire on stream channels and aquatic ecosystems. Recently, the research community has reported an increasing number of post-fire analyses, likely because: i) high-intensity crown
fires are increasingly common as fire suppression practices have lead to unusually high fuel loads in many forests (Daigle 1996), ii) forest mortality as a result of insect infestation, such as the Mountain Pine Beetle, is contributing to increased fuel loads in many forests (Keane et al., 2002); and iii) climatic change is expected to influence fire frequency (Johnson and Larsen, 1991).

Much of the leading research on landscape response to wildfire has been conducted in southern California (e.g. Rice, 1982; Wells 1987; Florsheim et al., 1991), in the Colorado Front Range (e.g. Moody and Martin, 2001a; Kunze and Stednick, 2005), and in Yellowstone National Park (e.g. Meyer et al., 1995; Legleiter et al., 2003). A large proportion of post-fire studies have been principally concerned with changes in hydrology, surface erosion, and the effects of mass movements from hillslopes. In fact, surprisingly little research has been conducted to analyze the role of bank erosion on the overall stability of channel morphology following fire. This is likely due to the highly variable nature of channel morphology, the complexity of governing conditions, and the lack of pre-fire data. In addition, there is a bias towards researching systems where fires are followed by large rain events that increase the sediment supply through mass wasting (Wondzell and King, 2003). Large, exogenous sediment inputs from mass wasting events may overwhelm the signal from bank erosion. Currently, there are no methods established for quantifying the magnitude of change or determining the time scale of recovery following fire.

Stream response to wildfire may lead to a number of adverse environmental and socio-economic effects. Research on post-fire ecology indicates that water temperature, water quality, and instream habitat features may be affected by wildfire (Agee, 1993). Increased sedimentation of fine material in riffles may reduce density and diversity of aquatic organisms (Roy et al, 2003). Loss of canopy cover due to fire may also lead to increased summer maximum water temperatures in small streams (Helvey, 1972). Furthermore, there may be significant hazards such as debris flows, floods, and landslides associated with watershed response, which are especially relevant where wildfires occur adjacent to populated areas. The probability of hazardous, high magnitude events has been shown to increase following intense wildfires (Benda et al., 2003a; Wondzell and King, 2003). Jordan et al. (2004) observed an unusually large number of debris flows following the 2003 wildfires in southern British Columbia, triggered by high runoff from fire-induced hydrophobic soils.
The response of a given watershed to wildfire is highly variable, both spatially and temporally, depending on several factors including the initial conditions in the watershed as well as the intensity, severity, and extent of fire. Recovery time to pre-fire conditions varies dramatically and is estimated to be 25 to 300 years based on the stability of the system (Debano et al., 1998). Specifically, changes in post-fire channel morphology vary considerably between regions. Wondzell and King (2003) noted that post-fire channel morphology is highly dependent on the regional variations in climate that influence sediment supply mechanisms. Where debris flows and landslides are common such as in the highly active climate of the Pacific Northwest the addition of large quantities of Large Woody Debris (LWD) in combination with coarse sediments can create stable physical structures that dramatically alter channel shape. In contrast, the relatively dry climate of the Rocky Mountain Region is characterized by post-fire input of fine sediments from surface erosion. Fine sediments have less of an impact on channel morphology as finer material is more easily moved through the system (Wondzell and King, 2003). Collins and Ketcham (2001) also documented significant regional variation in California between the dry Chaparral landscape in the south and the moist north coast. They hypothesized that the differences are due to the apparent cohesion provided from vegetation root networks.

1.1.1. Post-Fire hydrology

Water yield may increase following fire as a result of decreased interception of precipitation and evapotranspiration from the loss of vegetation and litter layer. A literature review by Moody and Martin (2001b) documented significant increases in post-fire peak discharge and annual water yield with a recovery time of 3-10 years. They found that the change in peak discharge is greater than the change in annual water yield and is therefore a more sensitive measure of hydrologic response. In British Columbia, Cheng (1980) conducted a paired watershed analysis to examine the hydrologic effects of the Eden Forest Fire near Salmon Arm. The study reported higher and earlier peak flows and an increase in total water yield, which is generally corroborated by other studies examining the influence of forest harvesting in British Columbia (Winkler et al., 2005). In addition, intense ground fires may lead to the formation of hydrophobic soils and other changes in the upper soil horizons. Where clays and organic material are present, the upper soil surface horizons may be altered by high fire temperatures, which may result in decreased total porosity and pore size, thus reducing infiltration rates (Debano et al., 1998).
Changes in snow accumulation and melt in burned and logged watersheds are well documented in the literature. Typically, the loss of forest canopy results in greater snow accumulation and earlier melt (Winkler, 2000). Greater accumulations are largely due to lack of interception and redistribution from altered wind patterns (Scott and Pike, 2003). Snowmelt typically occurs up to two weeks earlier in clearcut areas compared to forested ones depending on forest type, aspect, and elevation (Winkler, 2000). The same is true following fire. In his study following the wildfire near Salmon Arm, BC, Cheng (1980) documented earlier snowmelt. These findings are also in agreement with the literature on snow accumulation and melt following other vegetation disturbances such as disease (Scott and Pike, 2003).

1.1.2. Role of Vegetation

Where fires cause the mortality of riparian vegetation, stream banks may become vulnerable to erosion. Numerous researchers have demonstrated the influence of riparian vegetation on alluvial channel form (Andrews, 1984; Hey and Thorne, 1986; Millar and Quick, 1993; Millar, 2000; Eaton et al., 2004; Eaton, 2006). Empirical research shows that alluvial channels that support healthy riparian vegetation are deeper, narrower, and migrate more slowly than channels without riparian vegetation (Abernethy and Rutherfurd, 2000). Vegetation controls channel process and form mainly by altering channel flow resistance and bank strength. Hickin (1984) demonstrated that variation in bank vegetation in small streams may result in significantly altered channel roughness and width.

In forested riparian areas following fire, bank erosion may be responsible for increasing the amount of instream LWD as trees on the banks are recruited into the channel. Early recruitment of LWD occurs as a result of initial tree mortality, while later recruitment of LWD largely depends on bank stability and rates of lateral erosion (Debano et al., 1998). Recruited trees may likewise introduce sediments from their exposed root wads, and often promote further erosion by redirecting flow into the bank or forcing channel avulsions (Hogan, 1986; Bisson et al., 1987; Montgomery et al., 2003; Hassan et al., 2005). Indeed, recruitment of LWD following disturbance has been found to cause significant changes in channel morphology (Collins and Ketcham, 2001; Hogan, 1986; Benda et al., 2003b). Increased woody debris loading may lead to channel instability by: i) forcing areas of local scour below LWD obstructions; ii) increasing
bank erosion due to flow diversion; iii) promoting deposition of sediments that may form bars, steps, and plunge pools (Montgomery et al., 2003). LWD jams form at random along the length of the channel and provide structure for sediment storage upstream. The formation and decay of LWD jams over time may influence the sequence of bars and force multiple channels through avulsion (Hogan et al., 1998).

The degree of stabilization provided by bank vegetation depends on the rooting depth of vegetation relative to the maximum channel depth. Bank stabilization is greatest in small and intermediate streams where the rooting depth is equal to or greater than the maximum channel depth (Eaton and Church, 2007). In small- or intermediate-sized watersheds, bank erosion may increase following fire due to the absence of cohesive strength provided from dense root systems of live riparian vegetation (e.g. Collins and Ketcham, 2001). Healthy riparian vegetation has both a mechanical and hydrologic influence on bank stability. The hydrologic influence comes as a result of interception of precipitation that would otherwise have been infiltrated, or by loss of soil moisture through transpiration. Both processes result in increased shear strength due to the creation of negative pore-water pressure, called matric suction (Rinaldi and Casagli, 1999).

The mechanical influence of vegetation arises from greater soil shear strength provided directly from the root structure (Simon and Collison, 2002). In root-permeated soils the shear strength of the soil-root composite increases as a function of tensile strength and spatial distribution of roots (Abernethy and Rutherfurd, 2001). The tensile strength of roots has been measured by many researchers and has been shown to vary with: i) root diameter (Abernethy and Rutherfurd, 2001; Gray and Barker, 2004; Simon and Collison, 2002); ii) species (Gray and Barker, 2004); and iii) growing conditions (Gray and Barker, 2004). While many studies have quantified variation of tensile strength with root size and species, few have documented differences due to growing conditions.

1.1.3. Governing Conditions and Channel Stability

The shape of an alluvial channel has long been recognized to qualitatively represent a response to several factors, or governing conditions (Lane, 1957). The most important governing conditions controlling alluvial channel morphology are: i) the volume and distribution of peak water discharge; ii) the volume and calibre of sediment supplied to the channel; iii) the nature of
channel banks (including vegetation); and iv) the landscape history of the drainage in which the river flows (Church, 1992). The landscape history is important as it generally determines the type and quantity of sediments that are available and controls the valley gradient. The peak water discharge is mainly responsible for channel scale while the nature of sediment supplied to the channel largely controls channel morphology. The balance among the governing conditions determines the level of channel stability (Church, 2006).

An equilibrium channel may be defined as one in which the cross-sectional shape, roughness, and gradient allow transport of the sediment load contributed from upstream (Leopold and Bull, 1979). Sediment can be delivered to the stream channel by landslides, debris flows, surface erosion, in-channel storage sites, bank erosion, tree throw (Dietrich and Dunne, 1978). The quantity and sediment size distribution differs with each sediment delivery process. The channel form at any point in a stream network is the result of a balance between the supplied sediment load and the transport capacity of the channel. A channel that is in a state of quasi-equilibrium (while it may have stage-dependent adjustments to channel form) will return to the same form and surface texture after every flood hydrograph (Lisle et al., 2000). Alluvial channels may adapt to changing water and sediment discharge largely through adjustments in flow resistance. The hypothesis put forward by Eaton et al. (2004) is that equilibrium is achieved when the frictional resistance to flow for the fluvial system is maximized by maximizing flow resistance, and thereby, the system’s ability to deform the channel is minimized (see also Davies and Sutherland, 1983).

An alluvial channel response to sediment supply changes may occur at three scales: grain scale (surface texture); bedform scale (bars, pools); or reach scale (thalweg sinuosity). At the grain scale, the relationship between sediment supply and surface texture is well documented (Dietrich et al., 1989; Church et al., 1998). Experiments by Dietrich et al. (1989) and Lisle et al. (1993) indicate that a decrease in sediment supply entering the channel (of the same size distribution as the bed material) will cause the bed surface to become coarser and sediment mobility to decrease by way of an increase in the grain-scale flow resistance. Other researchers have documented an increase in sediment supply leading to bed surface fining (Lisle, 1982; Madej, 2001). At the bedform scale, variations in the spatial distribution of shear stress due to vertical adjustments of bars and pools cause zones of increased sediment transport and an overall increase in transport efficiency (Lisle et al., 2000; Eaton et al, 2006). Montgomery et al. (1999)
documented exceptionally high sediment transport rates in high supply conditions. They attributed increased transport rates to changes in grain size and bed roughness that increase transport capacity. And finally at the reach scale, experiments conducted by Eaton and Church (2004) indicate that a system with erodible banks preferentially adjusts flow resistance by varying thalweg sinuosity to account for increased sediment supply.

Field observations and experiments conducted by several researchers indicate that event-scale adjustments in response to increased sediment supply are triggered at threshold conditions (Montgomery et al., 1999; Lisle and Madej, 1992). Although researchers can clearly identify the point at which the threshold is reached in the field, the exact nature of these thresholds is not well understood due to the complexity of the system and measurement difficulties. The author’s field observations indicate that once threshold conditions are reached, the response is characterized by rapid adjustment with intense sediment transport rates. Over a longer time scale (on the order of decades), we speculate that changes in channel morphology may occur as a response-and-recovery sequence characterized by a period of rapid adjustment followed by a long recovery to pre-disturbance state. Or, the result may be a dynamic transition to a disparate state depending on the magnitude of disturbance and subsequent effects on the governing conditions.

1.1.4. Sediment Transport

It is important to clarify the terminology used to describe sediment transported in rivers. Transport is characterized by bursts of movement followed by stationary periods. The term 'path length' was first used by H.A. Einstein (1950) to describe the total distance moved by particles from initial entrainment to final deposition (Pyrcz and Ashmore, 2003b). This differs from 'step length' which describes the intermittent movements between 'rest periods'. In this paper, we will be examining path lengths.

Sediments may be categorized depending on the mechanism by which they are moved. Bed load is the material that moves along the bed due to tractive forces (rolling, sliding, or bouncing). Suspended load refers to sediments that are lifted up into the water column and are carried in suspension. Saltation of sediments is sometimes considered a third category of movement in rivers; however, it is not a dominant form of transport. Sediments may also be
categorized depending on their morphological role in the river system, which is the more appropriate categorization when examining channel morphology (Church, 2006). Bed material forms the bed and lower banks of alluvial streams. Wash material is characterized by fine sediments that are found mainly in the upper banks and once entrained are transported long distances. Bed material and bed load differ in that bed material may move in suspension during high flows where bed load always moves by tractive forces (Church, 2006). For the remainder of this thesis we will be concerned with bed material since it is most closely related to channel morphology.

Sediment is transported through the channel network in a complex sequence of storage and transport phases (Lisle and Church, 2002). Sediment transport is believed to be proportionally related to the shear stress acting on the bed. Reach average boundary shear stress may be approximated using the following equation:

\[ \tau = \rho g R S \]

where \( \tau \) is the shear stress acting on the boundary (N m\(^{-2}\)), \( \rho \) is the fluid density (kg m\(^{-3}\)), \( g \) is the acceleration due to gravity (m sec\(^{-2}\)), \( R \) is the hydraulic radius (area/wetted perimeter) and \( S \) is the energy slope (in practice, the water surface slope).

In many gravel bed rivers the mean shear stress only slightly exceeds the entrainment threshold of the median surface grain size at bankfull flows (Andrews, 1983). These channels are commonly referred to as “threshold channels”. The threshold channel concept was first presented by Lane (1955) for the design of stable canals. Natural threshold channels are generally gravel to cobble alluvial channels with bed load-dominated transport regimes. Total sediment transport in these channels is generally low intensity and therefore changes in channel morphology are expected to be slow (Church, 2006).
The critical shear stress for the entrainment of a characteristic grain size may also be estimated with a variation of the Shields equation. The equation to estimate the critical shear stress for the $D_{50}$ has the form:

$$\tau_{\text{crit}} = 0.06g(\rho_s - \rho)D_{50}$$

where $\tau_{\text{crit}}$ is the critical shear stress value for the $D_{50}$ and the value of 0.06 is Shield’s parameter. The estimation of Shield’s parameter is discussed in Section 3.6.2.

Many threshold channels have a characteristic bed surface coarsening - called armouring (Parker and Klingeman, 1982). Armouring is generally represented as the ratio of the surface median grain size to the median grain size of the subsurface:

$$D_{50} / D_{50,\text{sub}}$$

A high degree of surface armouring is often associated with low supply, low mobility systems. One theory states that surface armouring occurs at flows that are sufficient to gradually winnow away the fines but not move the larger material. It is hypothesized that selective transport of fines occurs until the surface is coarse enough to prevent further sediment transport (Hassan et al., 2006). This type of armouring is called static armour.

Another armouring mechanism was presented by Parker et al. (1982) which describes the formation of a mobile armour layer. The hypothesis may be described using the concept of equal mobility. Equal mobility occurs when all size fractions are transported in proportion to their presence in the sediment supply. During equal mobility transport the probability for entrainment is related to a particle’s size relative to neighbouring particles (Hassan et al., 2006). This is due to relative exposure to tractive forces and movement obstructions (Andrews, 1983). This results in smaller particles becoming less mobile as they are sheltered and larger particles becoming more mobile as they are exposed to entrainment forces. Under equal mobility conditions, a high degree of armouring does not necessarily indicate low sediment supply or low sediment mobility.
Surface structures, such as grain imbrication, grain lines, grain clusters and grain nets, may occur in conjunction with surface armouring, but are contingent on a maintained period of low sediment supply (Church, 2006). There have been numerous field observations and flume experiments that have documented the presence of surface structures in low supply, low mobility systems (Brayshaw, 1983; Hassan and Church, 2000; Church and Hassan, 2002). Highly developed surface structures may reduce sediment transport rates by orders of magnitude (Church and Hassan, 2002).

The flow regime is apparently also important as a secondary influence on the development of an armour layer. Hassan et al. (2006) documented that in arid climates with flashy flow regimes there is a characteristic lack of surface armouring, while in snowmelt-dominated flow regimes there is a distinctly greater degree of surface armouring. This can be theoretically explained through the concept of equal mobility. In flashy flow regimes the duration of competent flows able to produce conditions of equal mobility may not be long enough to form a coarse surface layer. Flume experiments also indicate that hydrographs of long duration, like those of snowmelt-dominated systems, increase the degree of surface armouring (Hassan et al., 2006).

There are numerous equations available for estimating sediment transport in rivers based on flow intensity. They are all fundamentally based on the assumption that a functional relationship exists between the movement of individual grains and key hydraulic variables. Each equation has inherent inaccuracies; in fact, several studies have documented significant discrepancies among standard sediment transport equations (Gomez and Church, 1989; Habersack and Laronne, 2002). Ashmore and Church (1998) referred to sediment transport equations based on hydraulic variables as the “forward approach”. In contrast, the “inverse approach” or “morphologic method” infers information about sediment transport based on observed changes in channel form. Beginning with Popov (1962), there have been many precedents for this technique in the literature (e.g. Neill, 1971, 1987; Carson and Griffiths, 1989; Ashmore and Church, 1998; Martin and Church 1995; Ham and Church, 2000; Eaton and Lapointe, 2001). The technique requires that detailed measurements of channel topography be made both measured before and after a given event. Sediment transport rates are then back-calculated, under certain assumptions, based on a sediment budget or travel distance approach (Ashmore and Church, 1998).
1.2. Objectives and Research Questions

The seemingly basic question of how Fishtrap Creek will respond to the loss of riparian vegetation is not easily answered given our current knowledge. The manner in which fire affects sediment transport processes and channel morphology is highly variable and poorly understood for most geographic areas. At this time, it is not possible to accurately predict channel response to wildfire, the rate of change, or the time scale at which response and recovery may be expected to occur. Documenting the response of Fishtrap Creek following wildfire will be important for expanding our knowledge of the hydrogeomorphic response to fire as well as to other disturbances that result in the loss of vegetation, such as disease, logging, blow down, or insect infestation. In the longer term, ongoing monitoring of channel changes and sediment dynamics will provide a basis for testing theories and developing models of longer-term channel response, particularly in relation to bank strength and riparian vegetation.

The overarching goal of this research is to better understand the mechanisms of hydrogeomorphic response of alluvial channels to vegetation disturbance. Specifically, the objectives of this thesis are: i) to document the short-term adjustments in reach-scale channel morphology, hydrology, sediment sources, and sediment mobility in Fishtrap Creek in the aftermath of the McLure Fire; ii) to attempt to explain empirical observations through the description of process-form interactions; and iii) to identify the temporal and spatial scales, and magnitude at which response is currently occurring and may be expected in the future. These objectives prompted the following research questions:

1) Has the reach-scale channel morphology of Fishtrap Creek changed in the response to the McLure Fire? If so, what is the magnitude, rate, and timing of channel change?
2) Has the magnitude or timing of peak flows changed as a result of the McLure Fire?
3) What is the predicted future hydrogeomorphic response of Fishtrap Creek and what implications might this have?
4) How does the post-fire response of Fishtrap Creek differ from other studies in the literature?
With these objectives and research questions in mind, the study was designed to quantitatively and qualitatively document the change that has occurred in Fishtrap Creek in the years following the fire through detailed empirical observation; and, to relate these observations to changes in key factors controlling alluvial channel form—called governing conditions—to determine if channel change may be directly linked to the fire.

1.3. Thesis Organization

Chapter 2 begins with an introduction to the general principles of the study design and follows with a detailed description of Fishtrap Creek. The description begins at the watershed scale, including a description of the climate, soils, geology, vegetation, and watershed history. At the reach scale, field data collected in 2006 are used to describe the channel morphology, LWD, bankfull channel dimensions, grain size distributions, and longitudinal profile. The chapter concludes with a description of the previous post-fire studies that have occurred in the watershed and details the data available for comparison in this study.

Chapter 3 contains a description and analysis of the hydrogeomorphic response that occurred during the 2006 freshet. The chapter begins with an introduction to the historic and current hydrologic conditions in the watershed with an emphasis on the timing and magnitude of annual maximum daily peak discharge. Then, analysis of multiple data sources including cross-sectional bed elevation surveys, stage height recorders, and magnetic sediment tracers, are used to document the magnitude and timing of observed channel change and sediment mobility. And finally, sediment transport rates are calculated using conventional flow-intensity-based equations and the morphological method.

Chapter 4 offers conclusions of the research. The chapter begins with a discussion that summarizes the response of the 2006 freshet and synthesizes the results from all data sources. Predictions are made about the expected future state of Fishtrap Creek as well as a theoretical discussion concerning the role of fire on the equilibrium channel morphology over a long time scale. The thesis concludes by discussing the implications of the research, and suggestions for possible directions for further work.
Chapter 2

2. STUDY DESIGN AND STUDY AREA DESCRIPTION

2.1. General Principles of Study Design

An empirical approach was taken to document a descriptive field study of the hydrogeomorphic impacts of fire. Descriptive field studies involve in-depth collection of both qualitative and quantitative data from a specific site. These data provide multiple lines of evidence to explain the physical processes driving change. A well-planned and carefully designed field study is informative for designing further research or generating and testing new hypotheses, particularly where natural disturbances occur. If the geomorphic response of interest has a significant lag time, then data collected soon after the disturbance may be assumed to represent pre-disturbance conditions, and in these cases it is possible to conduct reasonable pre-disturbance vs. post-disturbance comparisons. The study methods for this project combine stream channel monitoring with physically based data analysis to examine process-form interactions in a dynamic channel.

This approach differs from other research methods, such as experiments or meta-analysis, in that it considers a single case in detail rather than analyzing a large sample with a limited number of variables. The paired catchment experiment is perhaps more robust, statistically, but requires that one or more control watersheds be monitored, in addition to the disturbed watershed, for a number of years before and after the disturbance. This is almost always impossible to achieve. In the longer term, meta-analysis of numerous field studies may provide more statistical power and will be beneficial for expanding our knowledge of regional variation and for model testing. However, this requires that detailed, careful field studies be conducted in the first place.

2.2. Watershed Description

The Fishtrap Creek Watershed is approximately 170 km², located in the Interior Plateau physiographic region of British Columbia. The elevation of the watershed ranges from 370 m at the confluence with the North Thompson River to 1600 m on the high plateau. The region is
characterized by gently rolling uplands incised by steep-sided channels that are tightly coupled with the hillslopes.

The modern day landscape in this region is predominantly the result of rapid deglaciation of the Interior Plateau in the early Holocene around 12 000 BP (Ryder et al., 1991). While the Cordilleran Ice Sheet was retreating, rapid and intense erosion occurred as meltwater channels cut through valley fill leaving glaciofluvial terraces high on valley walls. The formation of the current drainage pattern occurred from the post-glacial deformation within the first few hundred years of deglaciation (Fulton, 1967). Mass wasting events constructed large colluvial fans at many of the valley bottoms as channels incised and valley walls were over-steepened.

Figure 3. Surficial geology within the Fishtrap Creek Watershed. The watershed consists of approximately 80% volcanic sediments. Data Source: British Columbia Ministry of Forests and Range.

Today, thick deposits of glacial drift cover virtually the entire surface of the plateau except for bedrock outcrops and rocky slopes in the entrenched channels. The hillslopes flanking
the channel are dotted with bedrock outcrops and boulders, with soils that are generally thin and poorly developed. The surficial geology of the watershed consists mainly of volcanic sediments, although Syenite and Diorite are found on some of the hillslopes in the upper reaches (Figure 3). A major north-south fault bisects the eastern edge of the watershed. The fault line is visible as the linear feature on the hillshade in Figure 4 and Figure 3 that intersects Fishtrap Creek downstream from the study site. Field observations of glaciofluvial terraces at the site indicate that a glacial meltwater channel previously occupied the fault line.

The stream channel through the upper reaches of Fishtrap Creek watershed is incised into the plateau and tightly coupled (in the sense proposed by Rice and Church, 1996) with the hillslopes. Near the study area, the floodplain widens as the creek intersects a glacial meltwater channel that follows a major fault line. Below the study site, the creek flows along the meltwater channel for approximately a kilometre before downcutting through a large glaciofluvial terrace associated with the North Thompson River (Figure 4).

The longitudinal profile contained in Figure 5 illustrates the complex variation in channel elevation throughout the watershed. The profile exhibits two distinct concave sections, the second of which ends approximately a kilometre below the study area. The shape of the profile is largely controlled by tectonic events and other geologic history. Breaks in slope appear to
correspond with large glaciofluvial deposits. However, downstream changes in discharge and sediment calibre are also thought to result in profile concavity (Knighton, 1984).

![Longitudinal profile of channel elevation](image)

**Figure 5.** Longitudinal profile of channel elevation throughout the watershed. Data source: Tim Giles, British Columbia Ministry of Forests and Range. Data Source: British Columbia Ministry of Forests and Range.

The climate of the area is semi-arid due to its location in the rain-shadow of the Coast Mountains. Winters in the region are relatively mild and short, while summers are hot and dry. Precipitation is produced by frontal systems during the winter months and convective thunderstorms in the summer. The hydrologic regime is snowmelt-dominated with a spring freshet driven by accumulations of snow at the mid and upper elevations (see Figure 13 in Section 3.2 for characteristic hydrographs). The Mayson Lake snow research area is located approximately 15 km from the upper elevations of the watershed at an elevation of 1250 m. Data from the site indicate that the area receives about 700 mm of precipitation annually, 60% of which falls in the form of snow (Winkler, 2005).

The vegetation in the watershed is best described by the Biogeoclimatic Ecosystem Classification (BEC) system. The BEC system, developed in the 1970s, incorporates vegetation type, climate, geology, and soils data to classify ecosystems within British Columbia. The majority of the watershed (60%) is in the Montane Spruce Biogeoclimatic (BEC) Zone (Figure 6). This zone is characterized by frequent stand-replacing fires. The vegetation in this zone is predominantly Subalpine Fir and White Spruce underneath a canopy of Lodgepole Pine. Lodgepole tends to dominate because of superior post-fire regeneration due to its serotinous cones that open in the heat of fire. Douglas-fir, Western larch, Western redcedar, Trembling
aspen, and Black cottonwood may also be found in this zone (McDowell and Lloyd, 1999). The Interior Douglas-fir BEC zone is found in the lower elevations of the watershed. The study site is located within this zone. Douglas-fir, Lodgepole pine, Ponderosa pine, and Trembling aspen are all commonly found here.

Figure 6. Biogeoclimatic (BEC) zones located within the Fishtrap Creek Watershed. The study site is in the Interior Douglas-Fir BEC zone while the majority of the middle elevations of the watershed are in the Montane Spruce BEC zone. Data source: British Columbia Ministry of Environment.

In recent history, fire, insect outbreak, and windthrow have been the dominant disturbances in the dry interior region. Large floods, landslides, and debris flows also occur; however, they are not as frequent as in the wet coastal region. Fires have likely been the principal disturbance mechanism since glaciation due to their large spatial extent and magnitude of influence. Fire cycles are not easily determined since they are influenced by climate change, and since modern-day forest practices remove most of the evidence from which they could be reconstructed. However, stand-replacing fires are thought to occur on average every 150-200 years in the Interior Douglas-fir and Montane spruce BEC zones (Egan, 1996).
The first people to occupy the North Thompson River basin were the Simpcw First Nation, a division of the Shuswap Nation. The Simpcw First Nation spent summers at hunting camps high on the plateau and winters in the valleys close to the rivers. The first European settlers arrived to the area in the early 1800s. The gold rush of the 1860s and the construction of the Canadian Pacific Railway spurred further growth. By the early 1900s, commercial timber harvesting began in the area. Timber harvesting and processing became the leading industry. There has been significant timber harvesting within the Fishtrap Creek watershed. Several roads have been built in the watershed to access harvesting areas and a microwave tower. Currently, grazing occurs within the basin and cattle have been known to occupy the riparian areas of Fishtrap Creek.

2.3. Study Area Description

The study site begins 20 m upstream from the Water Survey of Canada (WSC) weir and gauging station and continues for 400 m upstream. The channel has an average width of approximately 10 m and a maximum depth of less than 1.5 m. The bed material is predominantly gravel and cobble, making it an intermediate-sized stream channel (after Church, 1992). The morphology of the study reach may be categorized as ‘plane-bed’ with ‘forced pool-riffle’ sequences in the classification system of Montgomery and Buffington (1997). The channel lacks the regular spacing of bedforms characteristic of a ‘pool-riffle’ or ‘step-pool’ channel and the gradient is less than generally found in a ‘cascade’ morphology. Long sections of the channel are straight with a relatively featureless bed. Flow obstructions in the form of LWD and LWD jams result in localized and irregularly spaced bar, pool, and riffle morphologic units.

Currently, the riparian area within the study reach is densely populated with burned, standing-dead vegetation including alders, cottonwoods, and a variety of conifers. Salvage harvesting has not occurred on the floodplain, although it has in other areas within the watershed. The floodplain is marked by remnants of previously occupied channels that are overgrown with vegetation. The floodplain is wider in the study area than in the uplands and is not as tightly coupled with the hillslopes as it is further upstream. Hillslopes flanking the floodplain consist of thin, poorly developed soils over bedrock or till.
Channel banks are steep and undercut in many places. Fine sediments and dense root systems form the upper bank. The lower bank consists of unconsolidated gravel to cobble-sized material comparable to that in the channel bed, while the toe of the bank contains predominantly boulders ($>D_{90}$). Reinforcement from the root systems of both trees and grasses are identifiable in the bank profile. Figure 7 illustrates a characteristic bank profile within the study site where trees are found near the channel banks. The influence of grass root systems can be seen at a depth up to 20 to 30 cm, while the influence of riparian trees reaches approximately 1 to 1.5 m.

![Bank profile sketch](image)

Figure 7. Bank profile sketch located at cross-sectional survey transect number 8 is typical of the banks in the study reach. Banks are steep and undercut in many places throughout the study site. The influence of the root system of trees and grasses is identifiable in the bank profile.

2.4. Study Area Mapping

During the 2006 field season, a detailed survey was conducted to document the current conditions of Fishtrap Creek within the study reach. The study area mapping program was designed to provide a detailed longitudinal profile and a generalized planimetric map that would
serve as a way to organize spatial data as well as provide a baseline for the detection of significant longitudinal and planimetric channel change in the future. Several important measures of channel morphology were obtained from the mapping program including sinuosity, pool spacing, LWD characteristics, and channel gradient.

The longitudinal profile and planimetric form of the study reach were mapped using a laser range finder equipped with an electronic compass attachment. The survey data were collected in an arbitrary coordinate system and later transformed into UTM using the known coordinates of cross-sectional survey pins 1-11 and permanent benchmarks located within the study site. The precise coordinates of the survey pins and benchmarks were obtained from a total station survey conducted in 2004 by Fluvial Systems Research Inc.

The longitudinal profile of the thalweg was surveyed by recording the position of the thalweg, elevation, and water depth at less than 5 m intervals in the longitudinal direction and at sites that best represented the changing bed surface. Survey locations were specifically placed at changes in unit morphology (pool, riffle, glide) and at the point of maximum pool depth. Where the thalweg bifurcated, both forks were surveyed.

The planimetric mapping documented the location of bars, banks, LWD, and LWD jams. Bars were surveyed as polygons; the water’s edge at the time of survey formed the inside of the polygon and either the bank or the edge of vegetation formed the outside. Elevation of the bar top was also surveyed as a measure of bar volume. Bank position was recorded at less than 5 m intervals in the longitudinal direction at sites that best represented the planimetric form of the banks. Individual pieces of LWD greater than 0.1 m in diameter and 1.0 m in length, within the bankfull channel, were surveyed by collecting representative endpoints. The diameter of each LWD piece was estimated and grouped into four size classes and the orientation of each piece relative to flow direction was estimated. Polygons were surveyed around the perimeter of LWD jams and estimates of the number of individual pieces and average diameter were recorded.

There were several sources of error in the study area mapping program associated with the laser range finder and electronic compass attachment. First, the laser range finder reports an error of ± 0.01 m in the horizontal distance measurement and an error of ± 0.1° in the inclination measurement. The second source of error is related to the sighting device. The precision with which the target can be
sighted decays with increasing distance, thereby increasing error to an estimated ± 0.05 m at a horizontal distance of 30 m from the target. The third source of error was found in the electronic compass attachment. The compass is able to resolve the bearing to ≤ 0.5°. The absolute error in x and y coordinates resulting from the bearing resolution increases as a function of distance from the target. Error analysis indicated that error in the vertical was minimal (≤ 0.05 m). Horizontal error was larger, increasing radially from the instrument location to a maximum error of 0.30 m at a horizontal distance of 30 m from the target.

Total error in the planimetric map was reduced by spacing the instrument location to minimize distance from the target and adjusting the coordinates of each instrument location relative to the known coordinates of cross-sectional survey pins and permanent benchmarks. In this way, instrument error was not propagated in the longitudinal direction as the instrument location was moved. The error in the study area map necessitates that it principally be used for visualization and only for change detection greater than 1 m.

Figure 8 (A) contains the planimetric map of the study area. The two dark lines on the planimetric map separate the study area into the lower, middle and upper reach. It is in the middle reach where the original cross-sectional surveys and low-level reach photography were conducted after the fire, as described in Section 2.7. Only a small section of channel in the upper reach is confined by bedrock on one side; the rest of the bed and banks are alluvial material. The channel is bifurcated in the upper reach, apparently in response to a LWD jam that is currently buried. The channel throughout the study area exhibits a relatively low sinuosity except for the bend in the channel located at the downstream end of the middle reach (Figure 8 (A)).

Channel sinuosity was calculated as the ratio of the length of the channel centerline over the length of the valley ($L_{(channel)} / L_{(valley)}$). The sinuosity ratio over the entire study site was calculated to be 1.19. The expected range of sinuosity values is from 1.0 to 3.0 (Schumm, 1985). The distinction between straight and meandering is somewhat arbitrary; however, channels with sinuosity values in excess of 1.5 are often referred to as meandering (Knighton, 1984). The value of 1.19 within the study area indicates a relatively low sinuosity.
Figure 8. (A) Planimetric map and (B) longitudinal thalweg profile within the study area. Created by surveying the location of bars, banks, LWD, and the thalweg with a laser range finder equipped with an electronic compass attachment.
The factors controlling the degree of channel sinuosity are complex and result from the interaction of channel slope, bank strength, discharge, and sediment load (Knighton, 1984). Low sinuosity channels are generally associated with small width-to-depth ratios due to high bank strength relative to available stream power. The study site exhibits a relatively small width-to-depth ratio (see Section 2.4) and low sinuosity ratio (1.19) suggesting that the available stream power is not sufficient to freely erode the banks under normal conditions. Therefore, channel stability is expected to be high in the study area provided discharge and bank strength are unchanged. In such channels, if instability and lateral adjustments do occur, they may be related to either an increase in discharge or a reduction in bank strength.

The location of individual pieces of LWD and LWD jams are documented on the map in Figure 8 (A). Individual pieces of LWD were relatively frequent, averaging approximately 2 pieces per bankfull width. Jams were channel-spanning, moderately sized (10-20 individual pieces), and spaced on average every 8-10 bankfull widths. In total, 79 individual pieces of LWD and 8 jams containing an estimated 100 pieces of LWD were surveyed within the entire study reach. Of that, 50% were located in the right channel in the upper reach, which was abandoned following the 2006 freshet (see Figure 8 (A)). The reason for the greater quantity of LWD in the right channel is uncertain. It is possible that the reduction in gradient, discharge, and width associated with channel bifurcation has restricted throughput of LWD.

The number of pieces of LWD in the study site averaged 4.5 pieces per bankfull width; the average was 2.25 pieces excluding the right channel in the upper reach. These quantities are near-average when compared with other similarly sized channels in the literature. Examining data from a variety of regions, Fox (2001) found a weak but significant relationship between the number of pieces of LWD and channel width. Linear regression conducted by Fox (2001) predicted 4 pieces per channel width for a 10 m wide channel. For channels that were 8-12 m wide, the documented number of LWD ranged from 1 to 12 pieces per channel width.

Figure 8 (B) contains the longitudinal profile of thalweg elevation. The lower dark line represents the bed, while the upper light line represents the water surface. Note that the x-axis is the horizontal distance along the thalweg, not the centreline of the channel. Pools are easy to distinguish by the dips in bed elevation; glides and riffles appear similar as the shallow areas...
connecting the pools. The profile is distinguished by poorly defined and irregularly spaced morphologic units.

The pool-to-pool unit spacing is a common descriptor of channel morphology. The pool spacing for the entire longitudinal profile was calculated as the distance along the thalweg between points of maximum pool depth. Spacing varied widely, ranging throughout the reach from approximately 16 to 80 m. The mean pool spacing within the study area was calculated to be 38 m – or approximately 4 bankfull widths. Pool location and maximum depth were irregular and appear to be heavily influenced by the presence of LWD. The deepest and most tightly spaced pools were located at LWD jams (Figure 8.

The irregular pool spacing is in agreement with the observation that the study site exhibits a plane-bed rather than a pool-riffle morphology. In free-formed pool-riffle morphologies, pool spacing is expected to average 5 to 7 channel widths (Leopold and Wolman, 1957). However, Montgomery et al. (1995) found that in forested channels with LWD loading, pool spacing decreased with increasing LWD in both pool-riffle and plane-bed morphologies. They found that pool spacing decreased from greater than 13 channel widths per pool to less than 1 with increasing LWD loading. The pool spacing of 4 channel widths within the study area supports the conclusion that there is a moderate amount of woody debris loading compared with other forested channels.

Channel gradient was calculated from the thalweg profile. The gradient was calculated as $\Delta y / \Delta x$, where $\Delta y$ is the rise calculated from riffle to riffle at the point of minimum water depth and $\Delta x$ is estimated as the run calculated as the downstream distance along the thalweg. The average gradient for the entire study site was 0.022 m m$^{-1}$ (1.3°); this value falls within the observed range for pool-riffle, plane-bed, and forced pool-riffle morphologies (Montgomery and Buffington, 1997).

Slope declines in the downstream direction throughout the study area (upper reach, 0.031 m m$^{-1}$; middle reach, 0.022 m m$^{-1}$; lower reach, 0.014 m m$^{-1}$). The cause of the slope reduction is not entirely clear. The decreased slope in the lower reach may be in response to the local base level established by the weir located 20 m downstream of the study site. However, the amount of elevation change due to the weir is not known and the upstream effect would be expected to
decay rapidly. The local slope reduction trend may also be a result of the transition from the more confined uplands above the study to the glaciofluvial meltwater deposit located below the gauging station. This is in agreement with the profile of channel elevation, which exhibits concavity in this area (Figure 5).

2.5. Bankfull Channel Dimensions and Flow Parameters

Wolman and Miller (1960) defined the dominant or channel-forming discharge as that which performed the most work in terms of sediment transport. The dominant discharge is often associated with bankfull flow, based on an observed consistency between the two measures (Knighton, 1984). Bankfull channel dimensions such as average width, depth and velocity are useful as a way of generally describing channel behaviour and for estimating sediment transport. Bankfull channel dimensions and flow parameters are used to estimate sediment transport in Section 3.6.

The expected recurrence interval for bankfull discharge is approximately 1-2 years (Knighton, 1984). However, Pickup and Warner (1976) found that bankfull discharge is not necessarily of constant frequency. In this case, bankfull flow was observed at the study site on April 30, 2006 during the freshet. Discharge data from WSC indicate that the observed flow ($Q_{bf}$) was $7.5 \text{ m}^3\text{s}^{-1}$, which, based on historical records, corresponds to a recurrence interval of 1.5 years.

Surveys at cross-sections 1 to 11 (described in Section 2.7) were used to estimate the channel dimensions corresponding to bankfull flow. The reach average bankfull width was measured directly from channel cross-section surveys. Bankfull depth was estimated by averaging the depths ($d$) from the bed to an estimated water surface that filled the channel up to the height of the lowest bank. Figure 9 contains a diagram that shows an example of how the bankfull channel dimensions were calculated from the cross-sections.
Assuming a constant water surface elevation across the channel introduces error in the depth estimation because it ignores super-elevation that occurs on the outside of bends due to centrifugal forces. The amount of super-elevation may be estimated using the following equation:

\[ \Delta d = C \frac{v^2 W}{g R_b} \]

where \( \Delta d \) is the change in water surface elevation from the channel centerline to outside bend (m), \( C \) is a constant based on channel curve type, and \( R_b \) is the radius of the curvature of the channel centerline (m) (USACE, 1994). Cross-sections 1-5 are located at bends where super-elevation would be expected whereas cross-sections 6-11 are relatively straight. At the tightest bend in the study reach, near cross-sections 3-5, the range of super-elevation values was estimated at 0.07 to 0.14 depending on the constant chosen. This corresponds to a total elevation change across the channel ranging from 0.14 to 0.28 m. Because the water depth was estimated from the height lowest bank, the super-elevation error likely resulted in an under-estimated water surface elevation and thereby bankfull flow depth.
The reach average bankfull width \( (W) \) and depth \( (d) \) were estimated at 10.36 m and 0.58 m respectively. Using the recorded bankfull discharge \( (Q_{bf}) \) and the estimated width \( (W) \) and depth \( (d) \), bankfull velocity \( (v) \) was estimated with the continuity equation:

\[
Q = Wdv
\]

Equation 5

Accordingly, the bankfull velocity \( (v) \) was estimated at 1.34 m s\(^{-1}\) for the 2006 peak flows.

Table 1 displays the reach average estimated bankfull channel dimensions and flow parameters for the 2006 freshet.

| Bankfull Channel Parameters (2006) |
|-------------------------------|-----------------|-----------------|
| \( Q_{bf} \) (m\(^3\) s\(^{-1}\)) | 7.5             | \( v \) (m s\(^{-1}\)) | 1.34            |
| \( W \) (m)                    | 10.36           | \( S \) (m m\(^{-1}\)) | 0.022           |
| \( d \) (m)                    | 0.58            | \( W/d \)         | 17.9            |

The width to depth ratio \( (W/d) \) is often used as a measure for channel classification. The \( W/d \) ratio for the study site is 17.9. This is a relatively low value which is characteristic of plane-bed channels (Montgomery and Buffington, 1997).

2.6. Grain Size Distributions

Sediment sampling and grain size analysis are common techniques for monitoring gravel bed streams. Streambed texture is recognized as an important indicator of sediment supply and channel stability (Lisle and Madej, 1992). During the 2006 field season, sediment samples were collected to characterize surface and subsurface particle size distributions. Reference grain sizes were then used to calculate relative roughness and surface armouring. Reference grain sizes were also used as parameters for estimating sediment transport in Section 3.6.

Surface particles were sampled using a frequency-by-number methodology called a Wolman (1954) pebble count. It was conducted by measuring the \( b \)-axis (intermediate axis) of at least 100 surface particles collected in a grid pattern. The particles were measured in half-phi size classes; any particles smaller than 8 mm were classed as "fines". It was important that grid spacing was
significantly larger than the largest stone in the sampling area to ensure the independence of each sample (Wolman, 1954). Actual grain size distributions vary significantly both across and along the channel. Surface sampling techniques were designed to represent reach average conditions. Cumulative frequency distributions were developed for each sample. Reference grain sizes ($D_{50}, D_{95}$, etc.) were then calculated from the distribution using the following equation:

$$\phi_x = (x_2 - x_1) \left( \frac{y_x - y_1}{y_2 - y_1} \right) + x_1$$

where $\phi_x$ is the particle size of $x^{th}$ percentile; $y_2$ and $y_1$ are cumulative frequency percent values above and below the desired frequency; $x_1$ and $x_2$ are the particle sizes in phi units associated with the cumulative frequencies $y_2$ and $y_1$ (Bunte and Abt, 2001).

The subsurface sediment was sampled using a frequency-by-weight bulk sampling method. In this method, a volume of material was excavated from the subsurface and sieved into half-phi size classes. Each size class was then weighed and cumulative frequency distributions were developed. The frequency-by-weight sampling method was far more labour intensive than the frequency-by-number sampling method due to the large volume of material needed. The volume of material to sample is based on the largest particle on the surface. The sampling. volume must be large enough that the largest particle is less than 1% of the sample by weight to obtain significant results (Church et al., 1987). In gravel bed rivers, the largest stone on the surface is often 0.5 – 1.0 kg, resulting in sample masses of 500 – 1,000 kg.

Five Wolman pebble counts were conducted throughout the study reach during the 2006 field season. The samples were taken late in the season when flows were low and the bed had stabilized. Sampling grids were located in areas of homogenous substrate at the head of bars to best represent the median surface grain size of the main channel. Four out of five of the sample sites were located on bars that had recent accumulations deposited during the 2006 freshet. The last sample was taken at a site that had not had recent accumulation because the bar had shifted laterally. Table 2 contains the reference grain sizes of the old and new deposits.
Figure 10 contains the resulting cumulative frequency distribution and reference grain sizes for both the new 2006 and the old surface sediments. The 2006 surface distribution in Figure 10 represents the average distribution of the four samples from new deposits. Comparing the old and the new deposits, the samples differ at both ends of the distribution. The old deposit has a higher frequency of both the smallest and largest particles in the distribution. The increased frequency in the small end of the distribution is all in the “fines” size class (< 8 mm). This is not surprising since the sample site was located near the channel margin. However, the sample site location does not explain the increase in frequency of the largest particles. The largest stone from all of the samples was documented at this site (> 256 mm). It is difficult to draw any conclusions from one sample; however, the distributions suggest that the surface texture is fining, particularly in the coarse end of the distribution ($D_{90}$, $D_{95}$). Other researchers have documented surface fining under high sediment supply conditions (Lisle, 1982; Madej, 2001).

One bulk sample was conducted on October 1, 2006; the cumulative frequency distribution, $D_{30sub}$, and $D_{95sub}$ are displayed in Figure 10. The sample was taken on the bar near the weir pool. This location was chosen because the weir pond was excavated right after the fire and therefore the sample provides an estimate of the distribution of the material in transport since the fire. Because of the hydraulics at this location, it is expected that the bulk sample approximates the lower range of subsurface sediment size found throughout the study site.

<table>
<thead>
<tr>
<th>Size Percentile</th>
<th>Size (mm)</th>
<th>Size Percentile</th>
<th>Size (mm)</th>
</tr>
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<tbody>
<tr>
<td>$D_{50}$</td>
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<td>$D_{50}$</td>
<td>45</td>
</tr>
<tr>
<td>$D_{90}$</td>
<td>114</td>
<td>$D_{90}$</td>
<td>119</td>
</tr>
<tr>
<td>$D_{95}$</td>
<td>134</td>
<td>$D_{95}$</td>
<td>181</td>
</tr>
</tbody>
</table>
The ratio of the largest grains to the flow depth ($D_{90} / d$), termed relative roughness, is an indicator of the magnitude of influence of grain-scale flow resistance, and is often used as a measure for channel classification. Plane-bed morphologies generally have a higher relative roughness than that found in pool-riffle channels but less than that of cascade. The relative roughness for the study site was calculated to be 0.20, which is within the overlapping range of both plane-bed and pool-riffle morphologies (Montgomery and Buffington, 1997).

Likelihood for bar development can be evaluated with a combination of three indicators: gradient, width-to-depth ratio, and relative roughness. In natural channels, bar formation appears to occur at gradients $\leq 0.02$ (Ikeda, 1977; Florsheim, 1985 – cited in Montgomery and Buffington, 1997), which is a condition not met in the upper reaches of the study site (0.031). In addition, Church and Jones (1982) state that alluvial bar development requires a sufficiently large width-to-depth ratio and low relative roughness. Although the boundaries for these categories are not absolute, the values within the study site are within the range where bar development may or may not occur. This suggests that bar development might be expected under high supply conditions and not under low supply conditions.
The degree of vertical sorting, or surface armouring (discussed in section 1.1.4), was found to be high in the study site, with a value of 3.3. The armour ratio is calculated as the median grain size of the surface ($D_{50}$) divided by the median grain size of the subsurface ($D_{50sub}$); an armour ratio of 1 represents no surface coarsening and an armour layer $> 2$ is thought to be well developed (Hassan et al., 2006). Surface armouring is often associated with low sediment supply and low sediment mobility. However, as discussed in Chapter 1, a high degree of armouring does not necessarily indicate low sediment supply or low sediment mobility. Sediment mobility is also influenced by bed surface structures and flow regime. Lisle et al. (2000) observed highly armoured beds in high-supply, high-mobility systems compared to only moderately armoured beds in low-supply systems. Hassan et al. (2006) documented an apparent relation between flow regime and surface armouring by compiled data from humid, arid, and snowmelt-dominated streams. They documented that the armouring ratio was highest in snowmelt-dominated systems, ranging from 2 to 7, with a mean of 3.4.

2.7. Existing Data and Previous Analysis

Continuous discharge data has been collected at Fishtrap Creek since 1971 by WSC. Sporadic and seasonal measurements are also available for the time period 1914-1929. The weir and gauging station are located near the downstream end of the study site, at which point the drainage area is approximately 135 km$^2$. In this study, analysis was conducted using hourly, daily, and maximum daily peak discharge data provided by WSC. A description of the pre- and post-fire hydrology of Fishtrap Creek is located in Section 3.2.

During the first year following the fire, Tim Giles, Research Geomorphologist for the British Columbia Ministry of Forests and Range, established a monitoring program at Fishtrap Creek that included annual cross-sectional surveys of bed elevation and low-level reach photography. Mr. Giles surveyed 11 transects, monumented with rebar posts, over a 115 m reach located upstream from the WSC gauging station. The location of the 11 cross-sections is shown in Figure 11. The cross-sectional bed elevations were surveyed in 2004, 2005, and 2006 using an automatic level, tape, and stadia rod. In 2006, the research area was expanded from the original 115 m section of creek containing 11 cross-sections to a nearly 400 m section containing 27 cross-sections. For the remainder of this document, the portion of the study site containing the
original 11 cross-sections will be referred to as the *middle* reach. The portions of the study site upstream and downstream of the middle reach will be referred to as the *upper* and *lower* reach respectively.

CROSS-SECTIONAL SURVEY LOCATIONS

Projection: UTM zone 10N, Datum: NAD 83

* location of individual pieces of LWD surveyed at points within bank-full width

** The bounding polygon of LWD clusters were surveyed rather than individual pieces

Figure 11. Location of cross-sectional survey transects 1-11. Bed elevations have been surveyed annually at each location 2004-2006.

In 2004, Mr. Giles collaborated with Steve Bird of Fluvial Systems Research Inc., an expert in river photography and photogrammetry, to capture the low-level reach photography. They photographed the creek by suspending a camera on a 10 m telescoping pole with a gimbal mount. Then, using image interpretation software, they were able to generate a photographic mosaic of 2004 channel conditions. We repeated the photographic mosaic of the middle reach in 2006 for comparison.

Suspended sediment concentrations have been measured continually since March 29, 2004. An ISCO automatic water sampler is in operation near the WSC gauging station. Currently, the suspended sediment sampling program is maintained by Tim Giles and Dr. R.D.
Moore. Using these data, Petticrew et al. (2006) examined the short-term influence of the McLure fire on fine sediment delivery and retention (see Section 3.1 for discussion).

Bob Grace and Dennis Einarson from the BC Ministry of Environment have been examining water quality since the fire. They operate an autostation at Fishtrap Creek that continuously measures water chemistry, conductance, temperature, and turbidity. Other researchers have begun post-fire analyses pertaining to stream temperature, aquatic invertebrates and organic matter, turbidity, and snow processes.

The analyses contained in the following chapters rely on several of these data sources. The long record of discharge data from WSC provides detailed information on the pre- and post-fire hydrology. In addition, cross-sectional bed elevation surveys, and low-level reach photography data previously collected are used for comparison against data we collected during the 2006 field season. Water quality, water chemistry, water temperature, suspended sediment, and aquatic invertebrates were not considered in any analysis contained in this thesis. Sources for these data are mentioned above to provide general description of the data available and as a reference for future researchers.
3. ANALYSIS OF HYDROGEO MORPHIC RESPONSE - 2006

3.1. Introduction

The lack of pre-disturbance data for comparison is one of the key challenges when studying natural disturbances. Fortunately, in the case of Fishtrap Creek, there is a long history of pre-disturbance discharge data from WCS (Table 3). This provided an exceptional opportunity to study the hydrologic and geomorphic effect of forest fires. To take advantage of the research opportunity, several researchers began collecting baseline monitoring data following the McLure Fire (see Section 2.7).

Several of these data sources provide insight into the initial hydrogeomorphic response of Fishtrap Creek – specifically, suspended sediment concentrations and cross-sectional surveys. Suspended sediment concentration data for the first 15 months following the fire were examined by Petticrew et al. (2006). They documented minimal increases of suspended sediment from surface erosion and noted that the sediment yields were surprisingly low for a system so severely burned. They suggested that the hydrologic drivers (precipitation and snowmelt) were not large enough to move significant amounts of material: From the data, they further suggested that channel and bank erosion were also minimal in the first 15 months following the fire (Petticrew et al., 2006). Channel cross-sectional surveys also indicate minimal channel change between the 2004 and 2005 surveys (see Section 3.3.3). Therefore, surveyed cross-sections and low-level photography collected in 2004 (described in section 2.7) are assumed to be representative pre-fire channel conditions and are used for comparison.

It is believed that Fishtrap Creek watershed did not exhibit the soil hydrophobicity that is commonly observed following other fires (Tim Giles, pers. comm., 2005). The reasons for this are not apparent, since the fire intensity and ground temperature reported were sufficient for the formation of hydrophobic soils. It appears as though the lack of soil hydrophobicity following the fire has averted event-based surface erosion (e.g. Rice, 1982; Florsheim et al., 1991), mass movements (e.g. Meyer and Wells, 1995; Cannon, 2001; Benda et al., 2003a; Jordan et al., 2004), and increased water yield (e.g. Helvey, 1972; Cheng, 1980; Moody and Martin, 2001) that
have been observed in the first few years following other wildfires. Instead, the response of Fishtrap Creek has exhibited a lag related to the decay of riparian vegetation root systems and subsequent bank instability.

To the author's knowledge, no other study has specifically examined the effect of post-fire bank erosion on sedimentation and channel stability. A good example illustrating this point is that in Shakesby and Doerr (2005), which is a frequently cited literature review of wildfire as a hydrogeomorphic agent, the only mention of bank erosion is in reference to over-steepening banks, which in turn reactivates existing landslides. Where bank erosion is mentioned in the literature it is generally associated with increased peak flow (e.g. Wondzell and King, 2003). It is possible that the type of response observed at Fishtrap Creek is more common than it appears in the literature, for two reasons. First, funding for stream channel monitoring is often short-term and therefore stream channel response that exhibits a significant lag time may often be missed. Second, in systems where dramatic post-fire events occur (e.g. debris flows, landslides and surface erosion), the sediment input from bank erosion may be eclipsed by the larger-scale input. The circumstance at Fishtrap Creek provides an excellent opportunity to isolate the influence of post-fire bank erosion on channel stability.

Even though it is relatively intuitive that bank erosion would follow post-fire vegetation decay, there is little mention of it in the literature. The author found one study by Collins and Ketcham (2001) that documented the onset of bank erosion in the second year after the Vision Fire in northern California. In contrast, there have been many studies conducted to examine the effects of decreasing root cohesion on slope stability following forest harvesting. Review of a collection of studies indicate a 2- to 10-fold increase in mass wasting events in the period from 3 to 15 years following harvesting (Sakals and Sidle, 2004).

As with slope stability, there is a similar window of vulnerability for streambank stability following mortality of riparian vegetation. Live riparian vegetation increases soil strength through both mechanical and hydrologic processes (Simon and Collison, 2002). The degree to which mechanical processes increase bank stability is highly dependent on the tensile strength of individual roots and root density (see Section 1.1.2). The time scale of vulnerability is related to the decay rates of vegetation root systems and the regeneration of subsequent vegetation root systems. Figure 12 illustrates a generalized time sequence of the decay and recovery of
maximum root cohesion following vegetation disturbance. There is a sharp decrease in total root cohesion as a result of decaying root systems, but as regrowth colonizes the banks, total root cohesion returns to a maximum value. Rates of decay are known to vary by species and by growing conditions (Abernethy and Rutherfurd, 2001; Watson et al., 2001).

The remainder of this chapter presents the data and analysis that describe the hydrogeomorphic response of Fishtrap Creek to the McLure Fire. The data are organized in the following manner: i) discharge data are used to describe the historic and current hydrologic conditions, ii) cross-sectional bed elevation surveys are used to describe the patterns and magnitude of changes in channel form, iii) morphologic analysis of rating curves document the timing of channel adjustment, and iv) tracers reveal sediment mobility and the depth of the active layer. The chapter concludes with a section estimating sediment transport rates during the 2006 freshet using both hydraulic-based equations and the morphological method.

3.2. 2006 Hydrologic Conditions

Post-fire hydrology has been analyzed in terms of the magnitude and timing of peak flows using discharge data from 1971-2006. The maximum daily discharge is used for analysis
rather than maximum instantaneous discharge, as it is assumed that significant geomorphic work requires sustained flows over at least the course of a full day.

Both the maximum daily and maximum instantaneous discharge data for all years are contained in Table 3. The maximum daily peak discharge in 2006 was 7.5 m$^3$ s$^{-1}$, recorded on April 30. This value is near the long-term mean of maximum daily discharge for the period of record, which is 7.35 m$^3$ s$^{-1}$. Maximum daily peak discharge ranges over the years from a low of 2.97 m$^3$ s$^{-1}$ recorded on May 14, 2001, to a high of 14.9 m$^3$ s$^{-1}$ recorded on May 15, 1997. The highest maximum instantaneous discharge recorded was 15.3 m$^3$ s$^{-1}$ also on May 15, 1997.

The data indicate that the magnitude of peak flows during the freshet of 2006 were near average for the period of record. In fact, maximum daily discharge for all post-fire years (2004-2006) is within the historic range of variability (Table 3). The data imply that if there has been an increase in water yield following the fire it is likely minimal or confounded by climatic variability. Also, the maximum daily discharge was higher in 2005 (8.93 m$^3$ s$^{-1}$) than in 2006 (7.5 m$^3$ s$^{-1}$), but yet did not induce significant bank erosion or aggradation (see Section 3.3.3). This suggests that the observed bank instability in 2006 is more likely due to a loss of bank strength rather than an increase in hydraulic forces.

The timing of peak flows was also analyzed. Figure 13 contains the post-fire hydrographs and a representative subset of the hydrographs from the pre-fire years. The dark lines represent the post-fire hydrographs and the light lines represent the pre-fire hydrographs. Pre-fire years before 2000 were selected at 5-year intervals for image clarity. Pre-fire peak flows at Fishtrap Creek generally occurred during the month of May with a mean peak flow date of May 15th. In contrast, the mean peak flow date for the post-fire years (2004-2006) is April 27th.
Table 3. Maximum instantaneous and maximum daily discharge 1971-2006*

<table>
<thead>
<tr>
<th>Year</th>
<th>Instantaneous Q (m³ s⁻¹)</th>
<th>Time and Date</th>
<th>Maximum Daily Q (m³ s⁻¹)</th>
<th>Date</th>
</tr>
</thead>
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<td>---</td>
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Mean 7.50 7.35

* Data provided by Water Survey of Canada.
Figure 13. Annual hydrographs indicating that post-fire peak flows are of average magnitude and have been occurring approximately two weeks earlier than average pre-fire peak flows.

The hydrographs in Figure 13 illustrate that since the fire, peak flows have tended to be similar in magnitude but occur approximately two weeks earlier than before the fire. These findings are in agreement with other researchers documenting earlier peak flows in snowmelt-dominated catchments following vegetation disturbance (Cheng, 1980; Scott and Pike, 2003; Winkler et al., 2005). See Section 1.1.1 on post-fire hydrology.

3.3. Cross-Sectional Channel Adjustments

3.3.1. Introduction

The cross-sectional form of gravel bed rivers is sensitive to changes in sediment supply. Shifts in cross-sectional channel form may indicate changes in sediment supply and transport capacity. Rates of bank erosion and aggradation may be measured directly. In the following section, three years of cross-sectional bed elevation surveys (2004-06) are examined to determine the type and magnitude of post-fire changes in channel morphology. In addition, the discussion contains a comparison of the two reach photographic mosaics from 2004 and 2006 (described in Section 2.7) and other photographs that illustrate channel change. The discussion concludes with a theoretical model that provides a physically based explanation for the data presented.

Sediment supply in steep watersheds is often associated with exogenous variables such as hillslope surface erosion and mass wasting events. However, bank erosion is one of the primary mechanisms of sediment supply to streams (Knighton, 1984). Sediments input to the channel
from bank erosion are characterized by discrete, episodic events that are concurrent with moderate-sized flood events. The grain size distributions of sediments supplied from this type of input typically represent the entire spectrum of those found in the channel (Sullivan et al., 1987).

Aggradation from increased sediment supply due to bank erosion is well documented in the literature (Knighton, 1975; Griffiths, 1979; Lisle, 1982). Once transport capacity is exceeded, bed material is preferentially stored in bars and behind flow obstructions. This leads to asymmetry in cross-sectional form, where deposited material creates greater relief from the thalweg to the bar top. This asymmetry in channel form leads to variance in the across-channel distribution of shear stress, because shear stress is proportional to flow depth. The conceptual model presented by Eaton et al. (2006) indicates that increasing the variance in the shear stress distribution results in an overall increase in sediment transport capacity. Other researchers have also documented zones of increased sediment transport (Wathen et al., 1995) and an overall increase in transport efficiency (Lisle et al., 2000; Ferguson, 2003) with increasing variance in the shear stress distribution. Lisle (1982) reported that aggradation increased the effectiveness of the channel to transport sediment, particularly at low flows.

If channel banks are sufficiently weak, bar aggradation may cause further bank erosion. This occurs as a result of flows being diverted around the bars and concentrated near the banks. Continued bank erosion then increases the sediment supply to downstream reaches, thus creating a positive feedback loop. If the conditions are such that this process is allowed to continue for a sufficient amount of time, the channel will both widen and shallow to the point at which braiding may occur, provided gradient and sediment supply are suitable (Church, 2006).

3.3.2. Methods

In the three years post-fire, bed elevations were surveyed annually at 11 transects, over a 115 m research site (see Section 2.7 for description). In 2006, the research area was expanded from the original 115 m section of creek to a nearly 400 m reach containing 27 cross-sections. The map in Figure 14 shows the location of all 27 cross-sections across the entire study site. As discussed earlier, the section containing the original 11 cross-sections, XS 1 - XS 11, is referred to as the middle reach, while the upper reach contains cross-sections XS 12 - XS 19 and the lower reach contains cross-sections XS A - XS H. In the middle and lower reaches, cross-
sections are spaced in the longitudinal direction approximately every 10 m and in the upper reach every 20 m. All surveys in the middle and lower reach were conducted by Tim Giles, a Research Geomorphologist for the British Columbia Ministry of Forests and Range. With the help of a field assistant, I conducted the 2006 survey of the upper reach cross-sections. The analysis to follow uses only the 11 original cross-sections, XS 1 - XS 11, because there are three consecutive years of data with which to make comparisons. The surveyed bed elevations from the remaining 16 cross-sections (XS 12 - XS 19 and XS A - XS H) are presented in appendices A-F to illustrate the morphology of the upper and lower reaches.

The surveyed transects in the middle reach are oriented perpendicular to flow direction and spaced evenly in the longitudinal direction – roughly every bankfull width. The location of cross-sections 1-11 are shown on the map in Figure 15 (A), while Figure 15 (B) represents the thalweg longitudinal profile in this reach. Surveys were conducted using an automatic level, tape, and stadia rod. Within the active channel, the bed elevation was surveyed at ≤ 0.5 m intervals at sites that best described bed topography. On the floodplain, elevation was recorded at ≤ 1.0 m intervals. The edge of water, location of LWD, and the bankfull channel were also documented where possible. When points were within the wetted perimeter, water depth was recorded. In addition, morphology and characteristic sediment at each interval were recorded.

Measurement error in the survey of the cross-sections may be divided into vertical and across-channel error. The vertical error has two sources: first, the instrument error of the automatic level, which is estimated to be relatively small (≤ 0.001 m); and second, human error as data are misread from the stadia rod, which is likely the greater source of error. However, the magnitude of this error is unknown and therefore cannot be quantified. Across-channel error depends on the placement of the stadia rod and the accuracy with which the reading is taken from the tape. Measurement error in this direction is expected to be greater - on the order of several centimetres. Measurement error in either the vertical or across-channel direction is not expected to have exceed ± 0.05 m, which is approximately the same scale as the variability in the channel boundary itself. The measurement interval across the cross-section dictates the precision to which the channel bed is represented. The interval of ≤ 0.5 m is appropriate for examining bedform-scale channel topography.
Figure 14. Study area map indicating the location of the 27 surveyed cross-sections. Only cross-sections 1-11 are used in the analysis below.
Figure 15. (A) Location of cross-sectional bed elevation surveys 1-11. (B) Cross-sectional survey locations are labelled on the thalweg longitudinal profile.

Figure 15 (B) contains the longitudinal profile of the thalweg bed and water surface. The location of transects XS 1 - XS 11 are labelled on Figure 15 (B) in order to illustrate the influence of channel shape and the presence of LWD on the thalweg longitudinal profile (see Section 3.3.4 for discussion). Note that the horizontal distance on the x-axis of the thalweg
profile is the distance along the thalweg and is not the same as the straight-line distance along the valley.

3.3.3. Results

The 2005 and 2006 bed and water surface elevations from the eleven cross-sections located in the middle reach of the study site are plotted in Figures 16-18. The survey data from 2004 are not included on the figures for image clarity but are included in the analysis to follow. The figures are viewed as the reader is looking downstream; the distance across the channel (x-axis) increases from the left bank to right bank.

Cross-sections 1-7 all exhibited aggradation from 2005 to 2006, with as much as 0.42 m of aggradation occurring at cross-section 3 (Figure 16 and Figure 17). The pattern at cross-section 4 is different than at the other cross-sections, since deposition occurred not only on the bars but in the thalweg as well. This suggests that channel form and roughness adjustments at cross-section 4 were not sufficient to compensate for the increase in sediment supply. Therefore, the transport capacity was exceeded and aggradation occurred across the entire channel width. In addition, bank erosion was documented at cross-sections 3, 4, and 6 (Figure 16 and Figure 17), measuring as much as 1.5 m at cross-section 4. Cross-sections 8-10 experienced localized scour of the bed likely related to LWD movement within the jam located in this section of the reach (Figure 17 and Figure 18). And lastly, cross-section 11 upstream from the jam was relatively stable (Figure 18) exhibiting only a few centimetres of bed scour.

Changes in bankfull width and bar relief were used as indicators of the magnitude of channel adjustment. Table 4 displays the estimated bankfull width, and bar relief for 2004, 2005, and 2006. In addition, the table contains the calculated percent change since 2004 of both measures for each cross-section. Bankfull width was estimated from topographic breaks in the channel cross-section (see Section 2.5). The mean bankfull width in 2005 was 10.09 m compared to 10.36 m in 2006. This represents a 2.5 % increase in bankfull width from bank erosion during the 2006 freshet. Bar relief was calculated as a measure of aggradation or degradation at each cross-section. Bar relief was measured as the vertical distance from the thalweg to the bar top. There was an average increase of 63 % in bar height between 2004 and 2006. Isolating the cross-sections that had aggraded resulted in a 109 % increase in bar relief.
Figure 16. Surveyed bed elevations and water level for 2005 and 2006 at cross-sections 1-4. As much as 0.42 m aggradation is recorded in this section of the study reach. Cross-sections 4 and 5 exhibit the greatest bank erosion measuring 0.50 to 1.5 m. Data source: British Columbia Ministry of Forests and Range.
Figure 17. Surveyed bed elevations and water level for 2005 and 2006 at cross-sections 5-8. Cross-sections 5-7 document bar aggradation while cross-section 8 exhibited scour in the thalweg and no aggradation. Data source: British Columbia Ministry of Forests and Range.
Figure 18. Surveyed bed elevations and water level for 2005 and 2006 at cross-sections 9-11. Cross-section 9 exhibited measurable amount of erosion of the bed while cross-sections 10 and 11 were relatively stable. Data source: British Columbia Ministry of Forests and Range.
Table 4. Comparison of bankfull width and relative bar height from 2004 and 2006 cross-sections.

<table>
<thead>
<tr>
<th>XS</th>
<th>Bankfull Width (m)</th>
<th>Bar Relief (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>04-05</td>
<td>04-06</td>
</tr>
<tr>
<td>XS 1</td>
<td>8.50</td>
<td>8.50</td>
</tr>
<tr>
<td>XS 2</td>
<td>10.25</td>
<td>10.25</td>
</tr>
<tr>
<td>XS 3</td>
<td>9.00</td>
<td>9.00</td>
</tr>
<tr>
<td>XS 4</td>
<td>8.25</td>
<td>8.25</td>
</tr>
<tr>
<td>XS 5</td>
<td>9.50</td>
<td>9.50</td>
</tr>
<tr>
<td>XS 6</td>
<td>7.75</td>
<td>7.75</td>
</tr>
<tr>
<td>XS 7</td>
<td>13.75</td>
<td>13.75</td>
</tr>
<tr>
<td>XS 8</td>
<td>11.00</td>
<td>11.00</td>
</tr>
<tr>
<td>XS 9</td>
<td>15.25</td>
<td>15.25</td>
</tr>
<tr>
<td>XS 10</td>
<td>8.25</td>
<td>9.00</td>
</tr>
<tr>
<td>XS 11</td>
<td>8.75</td>
<td>8.75</td>
</tr>
</tbody>
</table>

3.3.4. Discussion

Cross-sectional surveys indicate that there was a small change in channel morphology from 2004 to 2005 followed by a much greater magnitude change from 2005 to 2006. During the 2006 freshet, field observations and cross-sections documented significant bank erosion, aggradation, and the formation of a channel avulsion downstream from cross-section 4 (shown in Figure 19).

Three cross-sections (XS3, XS4, and XS6) showed bank erosion, and 8 of the 11 cross-sections exhibited significant bar aggradation. In contrast, only minimal bank erosion and aggradation occurred following the 2005 freshet, even though the peak discharge was larger (8.93 vs. 7.5 m³ s⁻¹). From 2004 to 2005, only one cross-section (XS 10) showed bank erosion and several exhibited only minimal bar aggradation. It appears that the limited bank erosion observed in 2005, and the contrasting extensive bank erosion observed in 2006, may indicate a lag time in channel response. This lag time is similar to that found by Collins and Ketcham (2001) following the Vision Fire in northern California. Furthermore, this lag time is in agreement with the decay rates of vegetation root systems (Abernethy and Rutherfurd, 2001; Gray and Barker, 2004) and the window of vulnerability for landslide potential following the loss of vegetation (Sakals and Sidle, 2004).
Figure 19. Channel avulsion of the left bank near cross-section 4. Photo taken on April 30, 2006. $Q = 7.5 \text{ m}^3 \text{ s}^{-1}$.

Figure 20 (A and B) are photos of the bank erosion that occurred in the middle and upper reaches of the study site. Bank erosion was localized and varied in magnitude; the erosion was most dramatic in the upper reach of the study site. Unfortunately, there were no existing cross-sections in this reach to quantify the erosion; however, the author's field observations estimate hundreds of cubic m of material were excavated from a length of channel approximately 20 m long (Figure 20 (A)). The rate of erosion here was apparently amplified by the breakdown of an old LWD jam buried in the bank. In these systems, the formation and decay of LWD jams influences channel evolution. It is probable that the creation of the buried jam forced the channel bifurcation in the first place.

In addition to bank erosion, Figure 17 and Figure 18 indicate that bed scour is also occurring within the middle reach of the study site. Bed scour is likely contributing a portion of the material that is deposited on the bars downstream. Interestingly, Table 4 documents that more cross-sections exhibited bar aggradation than bank erosion during the 2006 freshet. This suggests that, in this case, only localized areas of lateral adjustment (bank erosion) were required to force extensive vertical adjustment. This may be in part due to the overall instability of the system resulting from increased sediment supply and LWD loading.
Figure 20. (A) Bank erosion in the upper reach of the study site. Bank erosion here was apparently amplified by the breakdown of a buried LWD jam. (B) Bank erosion in the middle reach of the study site, near cross-section 4. Dense root systems are exposed where banks have retreated approximately 1 m. The tree in the upper right hand corner was standing vertical before the 2006 freshet.

The growth of bars and movement of LWD is clearly identifiable on the low-level reach photographic mosaics collected in 2004 and 2006. A section of the mosaics that spans from cross-sections 3 to 6 is presented in Figure 21 (A and B). The 2006 mosaic shows the growth of point bars and the formation of a mid-channel bar near cross-section 4 that occurred during the 2006 freshet (Figure 21 (B)). The formation of a mid-channel bar is significant because it is an indicator of the propensity for channel braiding. In contrast, the 2004 photographic mosaic in Figure 21 (A) exhibits a uniform, relatively featureless bed. The methods for the pole photography, and photographic mosaic techniques, as well as an estimation of the error, are located in Appendix G. In addition, Appendices H and I contain the photographic mosaic of the middle reach for 2004 and 2006, respectively.

Figure 21 (A and B) also documents the movement of individual pieces of LWD within the study reach. Although woody debris loading was observed as a result of bank failure during the 2006 freshet, the quantity was limited to a few trees. The quantity of instream LWD per channel width has not dramatically increased following the fire. A great deal of the burned riparian vegetation remains standing. Barring a major blow-down event it may take decades for the standing dead to fall. The principal mechanism for woody debris loading over the next several years is expected to be as a result of streambank failures.
The response of alluvial channels to increasing sediment supply is complex and highly variable; therefore, interpretations are inherently over-simplified and should not be expected to apply in all cases. However, it is useful to speculate on the mechanisms behind the observed response as it aids in the creation of new hypotheses and clarifies questions for further study. The interpretations to follow are based on the author’s field observations and understanding of the relationship between bank erosion, sediment supply, and channel form.

The response of channel form to increasing sediment supply in Fishtrap Creek may be described in stages. Figure 22 contains a conceptual diagram illustrating the stages of cross-sectional adjustment in response to increasing sediment supply in a channel with stable banks. In the first stage of low sediment supply, the channel cross-section is relatively uniform and featureless (plane-bed morphology). During the second stage, increasing supply leads to material being preferentially stored in bars. This process increases the transport capacity of the channel by increasing the variance in the shear stress distribution (see Section 3.3.1). The channel cross-section achieves the most efficient form when sediment transport is maximized at all flows. In
the study reach, preferential deposition of material on the bars of cross-sections 1, 3, 5, and 6 appears to be increasing overall transport efficiency through cross-sectional form adjustments to (Figure 16 and Figure 17). In the third phase, sediment supply exceeds transport capacity across the entire channel width. In this phase, aggradation is evenly distributed across the channel. This aggradation may force overbank flows and channel avulsions to occur as stage height is forced higher. In the study reach, this appears to be occurring in the study site at cross-sections 2 and 4 (Figure 16).

![Diagram](image)

Figure 22. Theoretical diagram representing the relationship between sediment supply, cross-sectional channel form, and transport capacity.

3.4. Timing of Adjustment – Stage-Discharge Regressions

3.4.1. Introduction

Rating curves are commonly power-law functions that relate river stage and discharge. However, not all rating curves are power-law functions; the type of relationship depends on the existing cross-sectional channel shape and the relationship of mean velocity with water depth (Ferguson, 1986). Stage-discharge relationships are not static; alteration of channel form or flow velocity at any point in the hydrograph leads to a shift in the relation. Intuitively, aggradation or degradation of the channel bed or a change in water velocity will result in divergent stage height values for a given discharge. Therefore, if detailed discharge and stage height data are available, breaks in the relationship may be used to document the timing of channel adjustments. Leopold and Maddock (1953) documented the timing of within-event scour and fill in several Western United States streams using stage-discharge curves that were generated from gauging station data.
In the following section, the traditional rating curve is inverted to examine the timing of changes in channel form within the middle reach of the study site during the 2006 freshet. To our knowledge this is the first study that has used multiple stage height recorders to document the spatial and temporal variation in channel form adjustments within a single reach.

Traditionally, the rating curve is constructed as a power-law relation between discharge and stage at a specific location. The rating curve is fitted to the data using a power-law formula of the following form:

\[ Q = a(h - h_0)^b \]

where \( Q \) is the water discharge \( (\text{m}^3\ \text{s}^{-1}) \), \( a \) is the coefficient and \( b \) is the exponent of a power-law regression, \( h \) is the stage height \( \text{(m)} \) and \( h_0 \) is the stage height at which discharge is zero \( \text{(m)} \).

Changing channel gradient and roughness also may also influence stage height by way of flow velocity adjustments. As discussed in Section 1.1.3, alluvial channel form is the result of a complex balance between cross-sectional shape, roughness, and gradient, that allow for the transport of the sediment load contributed from upstream. The association of slope, roughness and velocity is given by Manning’s equation:

\[ n = \left( \frac{R^{2/3} S^{1/2}}{v} \right) \]

where \( n \) is Manning’s roughness coefficient, \( R \) is the hydraulic radius \( \text{(area/wetted perimeter)} \), \( S \) is the energy slope, and \( v \) is the flow velocity. The relation between mean velocity and cross-sectional area is described by the continuity equation (Equation 5), which states that discharge \( (Q) \) is the product of the cross-sectional area \( (W \times d) \) and flow velocity \( (v) \). Therefore, a decrease in flow velocity for a given discharge results in a proportional increase in cross-sectional area \( (W \times d) \) and vice versa.
3.4.2. Methods

On April 1, 2006, nine submersible water level recorders – called divers – and a barometer were installed in a variety of morphologically distinct sites within the middle reach of the study site. The divers recorded pressure and water temperature at ten-minute intervals from April 2 to July 2, 2006. They were housed in a protective PVC casing attached to lengths of rebar driven into the channel bed. The map in Figure 23 displays the location of each of the divers relative to the cross-sectional survey locations. The divers are labelled FD 1-9 incrementing upstream; diver 4 is not represented on the map because it was swept away early in the season and never recovered. The elevation of each diver was surveyed before and after the freshet to identify if they had been displaced during peak flows.

Water level above the diver was calculated based on the relationship between water pressure and water depth at a given temperature. The pressure transducer does not measure water pressure directly; it measures the absolute pressure, which is the sum of the water pressure and the atmospheric pressure. The water pressure (due to the overlying column of water, if we assume that the pressure distribution is hydrostatic) is calculated by subtracting the atmospheric pressure (due to the ambient atmospheric conditions) from the recorded values of absolute pressure. In order to document the changes in atmospheric pressure, an additional barometer data
logger must be installed near the pressure transducers in the stream channel in order to compensate for fluctuations in atmospheric pressure.

The water level data were reduced from 10 minute resolution to daily measurements taken at 4 a.m. and 4 p.m. to lessen the amount of data while still retaining diurnal cycles. Next, the corresponding discharge for each time step was obtained from the WSC gauging station. And finally, the data were fitted using ordinary least squares linear regression or a power-law regression depending on the best fit to the data.

3.4.3. Results

Early in the season, two of the divers were displaced as a result of instream LWD movement. As mentioned above, diver 4 was swept away on April 29\textsuperscript{th} and never recovered. Diver 6 was dislodged on May 4\textsuperscript{th}; however, it remained attached to a safety cable and was recovered. The following results were obtained from the seven remaining divers (FD 1-3, FD 5, FD 7-9).

The time series in Figure 24 illustrate the variation in stage height at the divers over time. The stage height was set to an initial value of zero for each diver so stage changes could be compared directly. Notice that on the first rising limb of the time series stage height values are closely aligned. They begin to scatter on the falling limb of the first peak and through the second peak. Following the third peak in the time series, stage height values diverge distinctly following different trajectories. Diurnal variations in stage are also evident, particularly later in the time series.

The discharge upon installation of the divers (1.9 m\textsuperscript{3} s\textsuperscript{-1}) is marked on both the rising and falling limb of the time series in Figure 24. This discharge is labelled to illustrate the point that the stage height recorded from several of the divers does not return to the same value on the falling limb of the hydrograph as on the rising limb at the same discharge. This indicates that channel cross-sectional shape, slope, or roughness were not constant throughout the time series. The stage at divers 3, 5, and 9 is 5 to 12 cm higher than at the same discharge on the rising limb; at diver 2 the stage is 21 cm higher.
Figure 24. Time series of changing stage height. Data from all divers were set to an initial value of zero for comparison. Channel form adjustments result in divergent relative stage height over time.

The logarithmic plots of $h$ against $Q$ in Figure 25 (B) illustrate an abrupt change in the stage-discharge regression at many, but not all, of the divers following the first peak in the hydrograph (Figure 25 (A)). In particular, divers 2, 3, and 9 exhibit a clear change in the stage-discharge curve following the first peak. Divers 1 and 5 also exhibit a change but of smaller magnitude and divers 7 and 8 are relatively uniform. Even though the shape of the relation differs among divers, all appear to have a reasonably well defined relationship between stage height and discharge in recession limb in the hydrograph. There is also a subtle break in the WSC relation which may be a result of sediment accumulations at the weir pond or a result of not including the stage when discharge is zero ($h_a$) in the model (Figure 25 (B)).
Figure 25. (A) 2006 hydrograph – the data from May 2 to July 2 is highlighted for comparison with logarithmic plots in B. The mean maximum daily peak for the period of record is displayed as a horizontal line. Data source: Water Survey of Canada. (B) Logarithmic plots of $h$ over $Q$ indicate an abrupt break in many of the relations following the second peak of the hydrograph.

Interpretation of the stage-discharge curve relies on the assumption that the WSC discharge data are accurate. The discharge data are expected to contain error as they are themselves derived from a rating curve. However, the error is expected to be minimal and would affect all the stage-discharge relationships similarly and therefore could not be responsible for the observed breaks in the stage-discharge curves. This is evident in the plots of divers 7 and 8 in Figure 25 (b) which have a single-valued relation.

To identify the timing of channel change, best fit stage-discharge equations were generated using diver data from May 2 to July 2. These data were chosen because the stage-discharge curve, at nearly all of the divers appears to be single-valued during this time period.
(Figure 24 and Figure 25). As discussed earlier, not all rating curves are power-law functions. The type of stage-discharge relationship depends on channel shape and the relationship water velocity and depth. To find the best fit equation, the stage height data from each of the divers were fitted using a power-law regression, least squares linear regression, and a 2nd order polynomial equation. Table 5 contains the equations and associated $R^2$ values for each diver. Power-law regressions reported the highest $R^2$ values for four of the divers (FD 2-5 and 9) while the 2nd order polynomial equation recorded the highest $R^2$ values for the remaining three divers (FD 1, FD 7, and FD 8). The $R^2$ values for each of the best fit equations were 0.94 or higher indicating good fits to the data (Table 5).

<table>
<thead>
<tr>
<th>Diver</th>
<th>Power-law equation</th>
<th>$R^2$</th>
<th>Linear equation</th>
<th>$R^2$</th>
<th>Polynomial equation</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>FD 1</td>
<td>$h=7.99Q^{1.09}$</td>
<td>0.95</td>
<td>$h = 8.6Q + 0.51$</td>
<td>0.97</td>
<td>$h = -0.62Q^2 + 11.98Q - 2.8$</td>
<td>0.98</td>
</tr>
<tr>
<td>FD 2</td>
<td>$h=35.73Q^{0.94}$</td>
<td>0.94</td>
<td>$h = 4.26Q + 32.01$</td>
<td>0.89</td>
<td>$h = -0.55Q^2 + 7.26Q - 29.1$</td>
<td>0.91</td>
</tr>
<tr>
<td>FD 3</td>
<td>$h=23.0Q^{0.50}$</td>
<td>0.98</td>
<td>$h = 7.21Q + 16.67$</td>
<td>0.95</td>
<td>$h = -1.04Q^2 + 12.88Q - 11.17$</td>
<td>0.98</td>
</tr>
<tr>
<td>FD 5</td>
<td>$h=22.7Q^{0.40}$</td>
<td>0.97</td>
<td>$h = 5.44Q + 17.65$</td>
<td>0.94</td>
<td>$h = -0.51Q^2 + 8.22Q - 14.96$</td>
<td>0.95</td>
</tr>
<tr>
<td>FD 7</td>
<td>$h=8.13Q^{1.21}$</td>
<td>0.84</td>
<td>$h = 9.67Q + 1.2$</td>
<td>0.97</td>
<td>$h = -0.78Q^2 + 14.01Q - 3.06$</td>
<td>0.98</td>
</tr>
<tr>
<td>FD 8</td>
<td>$h=6.97Q^{1.2}$</td>
<td>0.91</td>
<td>$h = 8.66Q - 0.2$</td>
<td>0.97</td>
<td>$h = -0.67Q^2 + 12.32Q - 3.79$</td>
<td>0.97</td>
</tr>
<tr>
<td>FD 9</td>
<td>$h=26.32Q^{0.50}$</td>
<td>0.98</td>
<td>$h = 8.51Q + 18.52$</td>
<td>0.97</td>
<td>$h = -0.77Q^2 + 12.71Q + 14.43$</td>
<td>0.98</td>
</tr>
</tbody>
</table>

The best fit model for divers 2, 3, 5 and 9 was a power-law regression while the best fit model for divers 1, 7 and 8 was a 2nd order polynomial equation (Table 5). Residual values (observed stage height – predicted stage height) from the best fit equations were examined to determine if the data exhibited heteroscedasticity (differing variance of residual values with the independent variable) or non-linearity, both of which violate the assumptions of least squares regression. Figure 26 (A-G) contains the residual values for each of the divers. The data do appear to exhibit weak heteroscedasticity, which does not cause the estimates of the regression to be biased, but may result in an underestimation of the standard error. The residuals do exhibit non-linearity in residual values, particularly at divers 2 and 7 (Figure 26 (B, and E)). Non-linearity simply indicates that the model does not completely accurately describe the data, which may be a result of explanatory variables that are not included in the model.
Figure 26. Residuals from May 2 to July 2 stage-discharge regressions for all divers. Divers 2, 3, 5 and 9 (B, C, D and G) were fit using a power-law regression and divers 1, 7 and 8 (A, E and F) were fit using a 2nd order polynomial model.

The residual values were calculated for all of the data from the best fit equations and are reported as a time series for each diver in Figure 27 (A-G). The horizontal lines on the graph designate ± 3 standard errors as calculated from the distribution of residuals for the May 2 to
July 2 data. It is assumed that residuals greater than ± 3 standard errors contained in the remaining data (April 2 to May 2) indicate the timing, direction, and magnitude of the change in the stage-discharge relationship.

Figure 27 (A-G). Residuals calculated for all data using the regression equations generated from the May 2 to July 2 stage and discharge data. Horizontal lines indicate ± 3 standard errors.
The timing, direction, and magnitude of change in residual values vary between divers. First, residual values from divers 1, 3, and 9 all indicate a relatively abrupt change in the stage-discharge relation on about April 25 (Figure 27 (A, C, and G)). The direction and magnitude of change are in agreement with previous analysis indicating that the stage height at these divers is 10 to 20 cm lower from April 2 to April 25 than at the same discharge on the falling limb of the hydrograph. Second, diver 2 and 5 exhibit a different pattern that suggests a sequence of scour and fill events occurred during the freshet (Figure 27 (B and D)). The stage-discharge relationship at diver 2 is particularly interesting; stage changes are steep during the rising limb of the first peak in the hydrograph. This helps to explain the unique appearance of the diver 2 logarithmic plots of $h$ over $Q$ in Figure 25 (B). Lastly, divers 7 and 8 also appear to have a subtle break in the stage-discharge relationship but not a large enough one to be statistically significant. These results are in agreement with the cross-sectional changes documented in Section 3.3.3. Divers that exhibited higher stage height on the falling limb of the time series were located close to cross-sections that had significant aggradation (Figure 23).

3.4.4. Discussion

Analysis of the stage-discharge regressions from the water level recorders and the WSC discharge data suggests that the majority of channel form adjustments during the 2006 freshet occurred between April 2 and May 2. In fact, at several of the divers, it appears that most of the channel adjustment occurred between the first peak and second peak in the hydrograph (April 9 to April 30). This is an interesting observation since little is known about the timing of channel form adjustments in snowmelt-dominated systems. There are particularly little data available because it is difficult to measure channel form directly at flood stages.

It was not possible to separate the influence of aggradation, slope, and roughness on stage height in this analysis given the complexity of the system and the lack of appropriate data. However, it is likely that stage height changes due to slope and roughness, although significant, are relatively small compared to the direct effects of bed aggradation.

Our field observations and cross-sectional surveys support the site specific changes in channel morphology documented by the divers. Areas of the channel where significant bed aggradation was documented with the cross-sections also demonstrate an abrupt break in the
stage-discharge regression, while areas that remained stable or experienced only modest bed aggradation maintained a relatively uniform relationship (Figure 25 (B)). Also, the plot of water surface in cross-section 2 (near diver 2) agrees with the stage-discharge analysis in that the stage height has indeed increased at that location (Section 3.3.3., Figure 16). The stage height in cross-section 4 also increased substantially suggesting that if diver 4 had not been lost it would have also documented a dramatic change in the stage-discharge relationship.

3.5. Sediment Mobility – Magnetic Tracers

3.5.1. Introduction

Magnetic tracer stones have widely been used to investigate sediment transport patterns in gravel bed rivers. Sediment tracers have been used for gathering information about the average distance of movement, sediment sources, depth of active layer, and volume of mobile sediment (Hassan and Ergenzinger, 2003). In the following section, magnetic sediment tracers are used to document the distribution of transport distances and identify morphologic constraints on travel distance. Tracer burial depth distributions are used to document the amount of scour and fill of the bed and the degree of vertical mixing that occurred during the 2006 freshet. In addition, in Section 3.6, the mean tracer travel distance is used as a parameter for estimating sediment transport rates.

Tracer travel distance has been examined by many researchers. Tracer studies conducted by Church and Hassan (1992) determined that, for unconstrained particles (free stones on the surface), there was a nonlinear decay relationship between grain size and mean travel distance. Other researchers have also documented that the mean travel distance decreases only slightly as a function of increasing grain size up to roughly the surface $D_{50}$ – where it drops off rapidly at the coarse end of the grain size distribution (Wilcock, 1997; Ferguson and Wathen, 1998).

The distribution of tracer travel distance has been shown to be well represented by a gamma function in many cases. The theoretical basis for this distribution was presented by H. A. Einstein (Einstein, 1937 cited in Hassan et al., 1991). He argued that particle movement occurs as a series of steps of random length that are followed by rest periods of random duration. If the step length and rest period are assumed to be serially independent, negative exponential variates,
then it follows that the travel distances over a given number of steps follow a gamma distribution (Hassan et al., 1991). However, others indicate that the travel distance distributions are influenced by morphologic constraints and do not necessarily follow a gamma distribution; a review of available tracer studies found that path length distributions take a variety of forms based on particle mobility and morphological constraints (Pyrcz and Ashmore, 2003a). Pyrcz and Ashmore concluded that path length distributions are heavily influenced by the length scale of the morphology (pool-to-bar spacing). Flume experiments by Pyrcz and Ashmore (2003b) indicate that, if the conditions for sediment transport are sufficient to mobilize bed material, then, the mean transport distance should be equivalent to the pool-to-bar spacing.

The rate of tracer movement – termed virtual velocity – is the distance of tracer movement divided by the duration of competent flow (Hassan et al., 1992). Most tracer studies have examined virtual velocity on event timescales; however, Ferguson et al. (2002) documented the rate of tracer movement over longer timescales. They mapped the dispersion of tracers after 2 and 8 years to identify the importance of burial depth and exchange on sediment mobility. The results were virtual velocities that were approximately 50% lower after 8 years than after 2 years. They concluded that the reduced rates of travel were a result of tracers becoming buried more deeply in the active layer or becoming trapped in long-term storage sites.

The burial depths of tracers have been used as an indicator of scour and fill depth, the volume of available sediment, and the degree of vertical mixing (Hassan and Church, 1994). Hassan (1990) documented that tracers tended to be more deeply buried in bars than in pools and riffles and that the probability for movement decreased with increasing burial depth. Hassan and Church (1994) reported that, under certain conditions, tracer burial depth distributions were well represented by an exponential decay function. They concluded from available data that short and single-peaked events led to a negative exponential burial depth distribution while snowmelt and events with multiple peaks did not. In addition, the shape of the frequency distribution of burial depths provides information about the depth of scour and the degree of vertical mixing (Hassan and Church, 1994)

This study is fundamentally different than most other tracer studies in that we are specifically examining the effectiveness of given flows to mobilize unconstrained particles near the $D_{50}$ of the surface – subject to morphologic constraints. Most other tracer studies have
attempted to represent the entire grain size distribution and have replaced recovered particles in their buried positions in an attempt to replicate natural conditions. Unconstrained particles near the $D_{50}$ represent the portion of the bed load that has the highest probability for movement.

3.5.2. Methods

Before the 2006 freshet, 400 magnetically tagged tracers were placed in Fishtrap Creek to examine sediment mobility and the depth of the active layer. The size classes of the tracers were selected as those bracketing the estimated median grain size of the bed surface. Tracer sizes ranged from 22-91 mm as illustrated in Figure 28. A total of 100 tracers were constructed for each size class: 22-32 mm; 32-45 mm; 45-64 mm; 64-91 mm. Tracers from the largest three size classes were constructed from natural alluvial material. These were constructed by drilling a small hole where magnets and a label could be inserted. They were painted a bright blue color using fish-friendly aquarium paint for easy visual identification. The magnets and label were bonded to the stone using epoxy. The smallest size class (22-32 mm) was too small for drilling and therefore they were artificially constructed from epoxy and lead shot. Magnets and labels were also inset into these tracers.

On April 2, 2006, the tracers were divided evenly into four heterogeneous groups of 100 particles each (labelled A, B, C, and D) and placed on the launch lines labelled in Figure 29. Launch lines were placed in the upper, middle, and lower reach of the study site to document sediment mobility through the entire reach. Tracers were randomly placed with respect to size class across the channel transect. The tracers were individually tamped into the bed in an attempt to avoid excessive exposure to the flow. In areas of higher velocity, some tracers were immediately moved up to 0.5 m by the flow before finding stable positions in the bed. The discharge reported by WSC at the time of installation was 2.2 m$^3$ s$^{-1}$, which is near the estimated threshold of motion for the surface $D_{50}$ (see Section 3.6).

In August, during low flows, the tracers were recovered using a magnetic locator. The tracers were not replaced in the location they were found; instead they were reclaimed to be launched again in subsequent years. The distance travelled, burial depth, and depositional morphology were recorded for each stone. The distance travelled was measured as the centerline
distance from the launch line to the point of deposition. Where applicable, the depositional morphology was recorded as pool, riffle, glide, bar, bar edge, thalweg or LWD storage.

Burial depth was measured from the base of the tracer to the bed surface. Burial depth is difficult to measure precisely. Particles may easily become displaced during the recovery process. Also, it is difficult to determine the exact height of the bed surface as it often changes around the hole dug to recover the stone. Efforts were made to minimize this error by measuring to a flat plane spanning the hole in the direction of flow. The burial depth error is estimated as ± 4 cm for deeply buried particles.
3.5.3. Results

In total, 83% of the tracers were recovered. However, labels that identified the launch line, and thereby distance travelled, were unidentifiable in a small percentage of the tracers due to damage from abrasion. The remaining 72.5% of the tracers that were undamaged were used to conduct the following analysis. It is possible that tracer Group D particles that travelled to the middle reach may have taken the right or left channel through the upper reach (Figure 29). However, it is highly unlikely that they were able to be transported through the right channel.
Erosion early on during the freshet led to the left channel being favoured early in the freshet. The particles that were recovered in the right channel were found in the upstream half. The downstream half consists predominantly of low velocity pools filled with fine sediments. Therefore, travel distances are calculated based on movement through the left channel.

Nearly all of the tracers recovered were moved from their starting location (96%). Tracers of all sizes traveled long distances, up to 530 m – approximately 52 bankfull widths ($W_b$). Twenty percent of the tracers recovered passed over the WSC weir. The mean travel distance for all tracers combined was 103 m (~10 $W_b$). This distance is more than twice the pool-to-bar spacing (~4 $W_b$) that is the expected travel distance suggested by other researchers (e.g. Pyrce and Ashmore, 2003b). Many of the tracers were also quite deeply buried; several were recovered over 50 cm below the surface. This depth is estimated to be near the detection limit for the magnetic locator.

The Fishtrap Creek tracer travel distance and burial depth were compared with other streams from a wide variety of fluvial environments. Table 6 contains the mean distance of travel for all particles ($\bar{L}$), for the surface $D_{50_s}$ (\(\bar{L}_{D_{50_{surf}}}\)) and the burial depth for six rivers from a range of environments. Detailed site descriptions may be found in the papers under the source column. In brief, the Allt Dubhaig is located in the Scottish Highlands. The channel width varies from 10 to 20 m. These are the only data that represent tracer movement over two years. Carnation Creek is located on the west coast of Vancouver Island, British Columbia. It is a highly active system that is dominated by intense rain events; the channel ranges from 5 to 20 m wide and the bed surface is weakly armoured (armour ratio, 1.6). Harris Creek is located in the interior of British Columbia; the channel is 10 to 20 m wide and is heavily armoured (armour ratio, 3). Nahal Hebron, (Israel) and Nahal Og (West Bank) are both ephemeral desert streams. They are moderately armoured (armour ratios of 2.0 and 2.3 respectively) and range in size from 3-12 m wide. Seale’s Brook, located in Quebec, is 4 to 8 m wide with reported bed packing and grain clustering. The tracers used at Seale’s Brook were painted clasts without magnets and therefore recovery was limited to stones exposed on the surface. Harris Creek and Seale’s Brook are the only two snowmelt-dominated systems.
Table 6. Comparison of Fishtrap Creek tracer data to other published studies.

<table>
<thead>
<tr>
<th>River</th>
<th>Event</th>
<th>Bankfull Q (m³ s⁻¹)</th>
<th>Range(mm)</th>
<th>Mean Travel Dist.</th>
<th>Mean LD₅₀ Burial Depth (m)</th>
<th>Mean Number of Particles</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Carnation</td>
<td>Aug. 29, 1991</td>
<td>24.5</td>
<td>16-180</td>
<td>129</td>
<td>--</td>
<td>0.06</td>
<td>50 (Haschenburger and Church, 1998)</td>
</tr>
<tr>
<td>subreach 3</td>
<td>Nov. 19, 1991</td>
<td>30.4</td>
<td>99</td>
<td>58</td>
<td>--</td>
<td>0.08</td>
<td>149</td>
</tr>
<tr>
<td></td>
<td>Jan. 29, 1992</td>
<td>22.6</td>
<td>58</td>
<td>27</td>
<td>--</td>
<td>0.04</td>
<td>43</td>
</tr>
<tr>
<td></td>
<td>Oct. 20, 1992</td>
<td>17.7</td>
<td>--</td>
<td>69</td>
<td>--</td>
<td>0.11</td>
<td>49</td>
</tr>
<tr>
<td></td>
<td>Jan. 24, 1993</td>
<td>36.3</td>
<td>26</td>
<td>--</td>
<td>--</td>
<td>0.08</td>
<td>11</td>
</tr>
<tr>
<td>Fishtrap</td>
<td>2006</td>
<td>7.5</td>
<td>22-90</td>
<td>103</td>
<td>107</td>
<td>0.14</td>
<td>290 this study</td>
</tr>
<tr>
<td>Harris</td>
<td>1989</td>
<td>15.2</td>
<td>6-512</td>
<td>57</td>
<td>24</td>
<td>--</td>
<td>564 (Church and Hassan, 1992)</td>
</tr>
<tr>
<td></td>
<td>Jan. 23, 1983</td>
<td>33.0</td>
<td>63</td>
<td>70</td>
<td>0.20</td>
<td>125</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Oct. 17, 1984</td>
<td>18.0</td>
<td>11</td>
<td>14</td>
<td>0.17</td>
<td>45</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Nov. 8, 1986</td>
<td>49.8</td>
<td>66</td>
<td>86</td>
<td>0.15</td>
<td>73</td>
<td></td>
</tr>
<tr>
<td>Og</td>
<td>Nov. 8, 1986</td>
<td>36.7</td>
<td>45-180</td>
<td>46</td>
<td>--</td>
<td>0.22</td>
<td>137 (Hassan et al. 1991)</td>
</tr>
<tr>
<td></td>
<td>Jan. 6, 1987</td>
<td>5.7</td>
<td>14</td>
<td>14</td>
<td>--</td>
<td>0.15</td>
<td>135</td>
</tr>
<tr>
<td>Seale’s Brook</td>
<td>1972</td>
<td>5.75</td>
<td>6-250</td>
<td>87</td>
<td>33</td>
<td>--</td>
<td>238 (Laronne, 1973; from Church and Hassan, 1992)</td>
</tr>
</tbody>
</table>

* bankfull discharge – this discharge was exceeded 13 times during the study period.

It is difficult to directly compare tracer travel distances between studies from different stream channels for several reasons. First, the flow intensity and the duration of competent sediment transporting flows varies between studies. Second, the flow regime varies between studies from snowmelt to rain-dominated. And lastly, the relative range of tracer sizes and recovery rates also vary between studies. However, it is useful to compare tracer data with other studies to provide insight about the relative mobility of bed material. The mean travel distance \( \bar{L} \) for Fishtrap Creek was relatively large when compared with other systems (Table 6). Only Carnation Creek (August 29, 1991) and Alt Dubhaig (1991 to 1993) recorded mean travel distances larger than Fishtrap Creek. The mean travel distance of the \( D_{50} \) at Fishtrap Creek was several times greater than at Harris Creek and Seale’s Brook, which are the only two other snowmelt-dominated flow regimes.
Frequency and cumulative frequency distributions of tracer travel distance were generated for each group and all groups combined. Frequency distributions were created by placing tracer data into bins based on the distance of travel of individual tracers divided by the mean distance of travel for that group \( \frac{L}{L_{\text{mean}}} \) following the methods of Hassan et al. (1991). The frequency distribution of tracer travel distances were similar to others reported in the literature (e.g. Hassan et al., 1991; Pyrce and Ashmore, 2003a). Figure 30 (A-E) contains the distributions for each tracer group and all groups combined. The frequency distribution of tracer groups A and B (Figure 30 (A and B)) has a steep decay with peak frequencies in the first one or two bins indicating that the majority of tracers from those groups moved less than half the mean travel distance for the group. The distribution of tracer groups C and D (Figure 30 (C and D)) exhibits a more gently sloping decay with peak frequencies that are not in the first bin. The distribution of all tracer groups combined (Figure 30 (E)) has a moderately sloping decay with frequencies generally well distributed among groups – i.e. there are few gaps in the distribution.

Travel distance distributions were fitted using a gamma function. Distribution statistics including the mean distance of travel for each group \( L_{\text{mean}} \) and standard error (S.E.), the shape parameter for the function \( k \) and standard error (S.E.) are located in Table 7. The gamma distribution is represented by the dark line in Figure 30 (A-E). Frequency distributions for all groups individually and all data combined appear to be relatively well represented by the gamma function; however, visual inspection shows systematic departure from the gamma distribution particularly in groups tracer groups A and B (Figure 30 (A and B)). Goodness-of-fit for the gamma function was assessed using the chi-square \( \chi^2 \) test. Table 7 contains the \( \chi^2 \) values as well as the critical \( \chi^2 \) values corresponding to the 0.01 significance level. The results were that individual tracer groups (A, B, C, and D) did not meet the goodness-of-fit test at the 0.01 significance level. However, the data combining all tracer groups did pass the \( \chi^2 \) test with a \( \chi^2 \) value of 16.2 and a critical \( \chi^2 \) value of 36.19.
Figure 30. Tracer travel distance frequency and cumulative frequency distributions for each tracer group and all groups combined. Gamma model fits to the distributions are shown as dark lines.
Table 7. Gamma function and chi-square statistics for distance distributions

<table>
<thead>
<tr>
<th>Group</th>
<th>L_{MEAN}(m)</th>
<th>S.E.</th>
<th>k</th>
<th>S.E.</th>
<th>$\chi^2$</th>
<th>$\alpha = 0.01$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Group A</td>
<td>88.09</td>
<td>14.39</td>
<td>0.51</td>
<td>0.14</td>
<td>57.5</td>
<td>29.1</td>
</tr>
<tr>
<td>Group B</td>
<td>75.79</td>
<td>12.38</td>
<td>0.51</td>
<td>0.15</td>
<td>71.2</td>
<td>26.2</td>
</tr>
<tr>
<td>Group C</td>
<td>137.33</td>
<td>28.4</td>
<td>1.36</td>
<td>0.29</td>
<td>54.0</td>
<td>21.7</td>
</tr>
<tr>
<td>Group D</td>
<td>112.30</td>
<td>15.08</td>
<td>0.87</td>
<td>0.22</td>
<td>59.7</td>
<td>26.2</td>
</tr>
<tr>
<td>All Groups</td>
<td>103.40</td>
<td>7.02</td>
<td>0.76</td>
<td>0.10</td>
<td>16.2</td>
<td>36.19</td>
</tr>
</tbody>
</table>

Fishtrap Creek tracer travel distance appeared to be related to the proximity of the nearest depositional area. The propensity for stones to be preferentially trapped in morphologic features was examined by calculating the residuals from the predicted gamma distribution of tracer travel distance. The observed density distribution was calculated as the number of tracers per channel width divided by the number of tracers available for deposition at that distance along the channel and the predicted tracer density was calculated from the gamma functions discussed above. The residuals were then calculated as the observed tracer density minus the predicted tracer density. Figure 31 contains the residuals of the density distribution for all groups along the channel centerline distance.

Figure 31. Tracer density residuals along the channel centerline (observed density minus density predicted from gamma model). The distribution highlights the influence of morphological constraints.

The mid-channel bar near cross-section 4, the WSC weir, and the tracer launch lines are labelled on Figure 31 to link the density distribution to the mapped channel morphology in Figure 15 and Figure 29. The distribution is distinctly non-uniform and exhibits several localized peaks in residual values. The highest residual value is located in the middle reach of the study site at the mid-channel bar near cross-section 4. Other high residual values in the study site are also located at areas where bars aggraded during the 2006 freshet. The last peak in the residuals is approximately 150 m downstream from the weir at a LWD jam where the gradient decreases.
and the channel becomes wider. The lowest residual value is located at tracer launch line B. Two other areas of low residual values are downstream from both tracer launch line D and the WSC weir. In both areas, the channel is steep, straight, and lacking major depositional features.

Neither travel distance nor burial depth was highly correlated with tracer group or grain size. The plot of burial depth over travel distance in Figure 32 illustrates a weakly decreasing trend. This relationship is somewhat intuitive as particles that are transported longer distances are likely to be deposited later in the freshet when flows are lower and the depth of the active layer is expected to be less. There is a distinct break in burial depths falling from a maximum of 0.55 m to a maximum of 0.25 m at roughly 115 m. This likely represents the range of travel distances of tracers that were deposited during major channel restructuring (Figure 32). Also notice clustering, by group, of transport distances on the x-axis indicating preferential deposition.

![Graph](image_url)

**Figure 32.** Tracer burial depth plotted against travel distance for all tracer groups. The light grey line represents the linear decay trend of maximum burial depth with travel distance.

Tracer burial depth distributions from Fishtrap Creek were compared to studies at four other sites: Nahal Hebron; Nahal Og; Carnation Creek; and Harris Creek. See Hassan and
Church (1994) for full river details and event descriptions. In order to compare burial depths, the
data were grouped into fixed intervals that equalled the size of the surface $D_{50}$—called layers.
The relative proportion was then calculated for each layer and they were labelled 0 to $x$
increasing with burial depth. For example, if the $D_{50}$ of the surface was 0.10, then layer 0 burial
depths ranged from 0.0 to 0.10 and layer 1 burial depths ranged from 0.11 to 0.20 and so on.

Figure 33 (A-I) contains the burial depth distributions for nine events from five sites
including Fishtrap Creek. Distributions vary widely both between sites and among individual
events at a site. The Fishtrap Creek distribution (Figure 33 (I)) is most similar to that of
Carnation Creek on Feb. 2, 1990 (Figure 33 (C)), which was a single peaked event and that of
Nahal Hebron on Jan. 23, 1983 (Figure 33 (F)), which was a long event with multiple peaks.
This was the main event after the particles were seeded on the Nahal Hebron. The Harris Creek
burial depth distribution (Figure 33 (H)), which is the only other snowmelt-dominated event,
document the majority of particles in the layer number 0 with a steep decay in the frequency of
subsequent layers. Others events document similar distributions (Figure 33 (A, B, D, and G)). In
contrast, the distribution from the Nov. 8, 1986 event of the Nahal Og (Figure 33 (E)) exhibits a
gently sloping decay with the frequency distributed quite evenly among layers. This was also an
event with multiple peaks and the first after the particles were seeded.

The burial depth distributions were fitted using an exponential decay function. The
exponential distribution is represented by the dark lines in Figure 33 (A-I). Goodness-of-fit for
the exponential function was assessed using the chi-square ($\chi^2$) test. Hassan and Church (1994)
conducted $\chi^2$ tests for the eight events (A-H) and the author conducted the test for Fishtrap
Creek (I). Table 8 displays the resulting $\chi^2$ value, the critical $\chi^2$ value for the 0.01 significance
level, and the number of tracers ($N$). The results of Hassan and Church (1994) indicated that the
distributions of the Nahal Hebron January 23, 1983 (Figure 33 (F)) and November 8, 1986
(Figure 33 (G)) events and the 1990 snowmelt event of Harris Creek (Figure 33 (H)) were not
satisfactorily described by the exponential decay function while all other events were. Hassan
and Church (1994) state that they know of no physical reason for the observed exponential
distribution of the burial depths. However, they concluded from these data that short and single-
peaked events led to a negative exponential burial depth distribution while snowmelt and events
with multiple peaks did not. Furthermore, they suggested that armouring of the bed surface at
Harris Creek and Nahal Hebron prevented substantial vertical mixing and inhibited the likelihood of an exponential decay distribution of burial depths.

The Fishtrap Creek data tend to contradict those conclusions. The 2006 hydrograph for Fishtrap Creek was long and complex with multiple peaks and the bed surface exhibits a high armour ratio of 3; yet, the burial depth data fit the exponential decay function remarkably well. The Fishtrap data passed the $\chi^2$ test yielding a value of 8.3 and a critical $\chi^2$ value of 23.21 (Table 8). The slope of the exponential decay was gentle, which Hassan and Church (1994)
suggest is an indication that the active layer was deep and experienced substantial vertical mixing.

<table>
<thead>
<tr>
<th>Event</th>
<th>$\chi^2$</th>
<th>$\alpha = 0.01$</th>
<th>$N$</th>
</tr>
</thead>
<tbody>
<tr>
<td>A. Nahal Hebron, Jan 23, 1983</td>
<td>42.63</td>
<td>16.81</td>
<td>244</td>
</tr>
<tr>
<td>B. Nahal Hebron, Oct. 17, 1984</td>
<td>12.23</td>
<td>15.09</td>
<td>141</td>
</tr>
<tr>
<td>C. Nahal Hebron, Nov. 8, 1986</td>
<td>35.36</td>
<td>13.28</td>
<td>100</td>
</tr>
<tr>
<td>D. Nahal Og, Nov. 8, 1986</td>
<td>18.37</td>
<td>21.67</td>
<td>142</td>
</tr>
<tr>
<td>E. Harris Creek, 1990</td>
<td>19.38</td>
<td>13.28</td>
<td>151</td>
</tr>
<tr>
<td>F. Carnation Creek, Dec. 3, 1989</td>
<td>14.35</td>
<td>16.81</td>
<td>171</td>
</tr>
<tr>
<td>G. Carnation Creek, Feb. 2, 1990 – yellow group</td>
<td>15.95</td>
<td>20.09</td>
<td>96</td>
</tr>
<tr>
<td>H. Carnation Creek, Feb. 2, 1990 – orange group</td>
<td>15.9</td>
<td>18.48</td>
<td>104</td>
</tr>
<tr>
<td>I. Fishtrap Creek, 2006</td>
<td>8.3</td>
<td>23.21</td>
<td>279</td>
</tr>
</tbody>
</table>

3.5.4. Discussion

Comparisons indicate that travel distances do indeed seem to be high relative to other systems even though flows were near average (Table 6). The mean travel distance of the $D_{50}$ was several times greater than in both Harris Creek and Seale’s Brook, which are the two other snowmelt-dominated systems. However, it is important to note that these comparisons do not account for the duration of competent flows or the relative size of the system. Larger systems may more freely erode their banks and adjust bars, which may reduce transport distances.

Tracer transport distances appeared to be heavily influenced by morphologic features. The residuals of tracer density distribution in Figure 31 illustrate the influence of channel morphology on travel distance. The largest peak in the distribution is in the middle reach of the study site located around the mid-channel bar near cross-section 4. This area also recorded the greatest amount of aggradation in the cross-sections (Section 3.3.3, Figure 16 and Figure 17). The author’s field observations indicate that peaks in residuals from the gamma model of tracer travel in the upper and lower reaches are also located at bars that formed during the 2006 freshet.
The distribution of transport distances for groups A, B, C, and D were not sufficiently described by the gamma function while the distribution containing all of the data passed the goodness-of-fit test. There are at least two possible explanations for this observation: the number of tracers in the four individual groups may not have been sufficient to represent the gamma distribution due to low density, or combining the four groups from different launch lines serves to remove the influence of tracers clustering in depositional features. Either way, the findings do support the reach-average tendency for transport distance to be well represented by a gamma function.

Fishtrap Creek transport distances do not conform with the expected pool-to-bar spacing suggested by Pyrce and Ashmore (2003b). The mean transport distance of 103 m is more than twice than the average pool spacing (38 m). The longest recorded path lengths in Fishtrap Creek reached 52 times the channel width. In a literature review of published tracer studies, Pyrce and Ashmore (2003a) define cases where the longest path lengths are much greater than 15 times the channel width as “extreme cases” (Pyrce and Ashmore, 2003a, pp.183). They suggest several possible reasons for “extreme cases”: the absence of bars; flow magnitude so great that the bar morphology is restructured; the flow event is long enough that particles are deposited and re-entrained through several pool-to-bar paths; events may be sufficient for bars to be removed and/or deposited. Whether or not any of these explanations applies to Fishtrap Creek is not entirely clear; however, the author speculates that long transport distances are related to the long duration of competent flows, high sediment supply, and the lack of well-developed bars in Fishtrap Creek before the 2006 freshet.

The mean burial depth was also relatively high compared to other systems. The burial depth distributions at Fishtrap Creek are well represented by an exponential decay function despite having a long and complex hydrograph. This is in contrast with the conclusions of Hassan and Church (1994) that short and single-peaked events lead to an exponential distribution while snowmelt and events with multiple peaks do not. The shape of the burial depth distribution at Fishtrap Creek does suggest substantial scour and thorough mixing of the active layer despite having an armoured bed surface. The physical explanation for the observed exponential burial depth distribution at Fishtrap Creek is not entirely clear. Hassan and Church (1994) identify three factors influencing the shape of the burial depth distribution: i) the magnitude and duration of the
flow event; ii) the number of events since the particles were placed; iii) and the surface structure and texture. We speculate that the difference between Fishtrap Creek and the other sites is due to the lack of surface structures rather than bed surface armouring or hydrologic conditions as they concluded. It is likely that the layer of mobile sediment at Fishtrap Creek consisted of loose, unconsolidated material recently eroded from the banks rather than a heavily structured bed surface. These sediments were likely easily mixed and provided little resistance to vertical exchange. However, it is difficult to draw any conclusions from this limited analysis of the data; further investigation is needed.

The relationship between depth and travel distance exhibits an abrupt break at approximately 115 m (Figure 32). The break represents the likely maximum distance of travel for tracers captured during major channel restructuring. Analysis of the timing of channel adjustments in Section 3.4 indicated that channel restructuring occurred by April 30, 2006. It is assumed that tracers buried at a depth of 50 cm would have been deposited during this time. Therefore, a rough estimate of the rate of tracer travel during peak flows may be estimated at 4.1 m day\(^{-1}\). Rather than an estimation of the transport rate this should be considered to be a speculation of the capacity of the system to transport unconstrained particles near the \(D_{50}\) during peak flows. The values are, however, reasonable when compared to other calculated virtual velocities in the literature (e.g. Hassan and Church, 1992; Eaton and Lapointe, 2001).

3.6. Sediment Transport Estimation

3.6.1. Introduction

Bed load transport rate is a fundamental factor in understanding the development and determining the stability of alluvial channels. Numerous equations are available for estimating sediment transport rates based on an assumed relationship between flow intensity and the transport of sediment. An alternative technique – called the morphologic method – utilizes observed changes in channel form to estimate sediment transport rates. In the following section, sediment transport rates are estimated for the 2006 season using both traditional flow intensity equations and the morphologic method. In addition, the reach-averaged duration of sediment transporting flows is estimated using a simple model constructed from cross-sectional bed elevations, stage height from divers, and discharge data provided by WSC.
Researchers have documented significant discrepancies among standard sediment transport equations. Both Gomez and Church (1989) and Habersack and Laronne (2002) provided a thorough description and quantitative assessment of traditional sediment transport equations. There are at least two important limitations to flow-intensity-based transport equations. First, even if a transport equation properly represents the physics of sediment movement at one point on the bed, it does not reflect the across-channel and downstream discontinuities in transport. Second, the measurement of key hydraulic variables that are used as parameters in transport equations is exceedingly difficult at channel-forming flows. The latter explains why many sediment transport equations are based on flume experiments rather than field data. A good example of this is the Meyer-Peter and Müller (1948) equation, which is a widely used empirical equation based on flume data.

Church (1985) argued that all the hydraulic variates are highly correlated and therefore all flow-intensity-based transport equations will work reasonably well (no matter how well, or poorly, they represent the physics of sediment transport) because sediment transport is roughly a scale factor of flow. This suggests that reasonable estimates of reach average sediment transport may be obtained from equations that directly relate sediment transport to discharge – for example, the Schoklitsch 1943 formula (Schoklitsch, 1950).

In contrast to flow-intensity-based transport estimation, the morphologic method uses net changes in channel storage to estimate sediment transport rates. One way to apply the morphological method is by assuming a typical path length for the total volume of erosion measured within a reach. In small to intermediate sized rivers, the total volume of erosion is most often measured by cross-section surveys (Ashmore and Church, 1998). The volume of erosion between cross-sections is usually estimated by prismatic approximation between cross-sections (e.g. Goff and Ashmore, 1994; Martin and Church, 1995). Since the typical path length of eroded sediment is rarely known, it is often assumed to be related to bar spacing (e.g. Carson and Griffiths, 1989), may be estimated by pairing discrete zones of erosion and deposition (Goff and Ashmore, 1994), or, may be estimated from excess stream power (e.g. Eaton and Lapointe, 2001). The path length morphological method may lead to an underestimation of actual sediment transport rates because it does not account for scour and fill of the channel bed between surveys (Ashmore and Church, 1998).
This study has the distinct advantage of recording the path length of eroded sediments (at least those near the median grain size of the surface) with magnetic tracers. However, as mentioned earlier, the fact that the tracers are unconstrained particles near the $D_{50}$ means that they represent the portion of the bed load that has the highest probability for movement. Therefore, estimated transport rates using the mean tracer path length is equivalent to estimating the sediment transport potential of the total measured volume of erosion, rather than the actual sediment transport rate.

3.6.2. Methods

The reach-averaged duration of sediment transporting flows was investigated using a simple model constructed from cross-sectional bed elevations, stage height from divers, and discharge data from WSC. The duration of sediment transporting flows was defined in the model as the duration in which the reach-average boundary shear stress ($\tau$) exceeded the critical shear stress for the $D_{50}$ ($\tau_{crit}$) — in other words, when the excess shear stress ($\tau - \tau_{crit}$) value was positive. The model was fitted to the recession limb of the hydrograph when stage-discharge regressions indicated the channel was relatively stable (discussed in Section 3.4.3).

To construct the model, a daily time-series of stage height was generated for five cross-sections located near stage height recorders. Then, the channel cross-sectional area ($A$) and the hydraulic radius ($R$), were calculated for each cross-section at every time step. This allowed shear stress ($\tau$) to be estimated at each time step using Equation 1, assuming a constant slope ($S$) of 0.019 m m$^{-1}$. The mean shear stress value for all cross sections at each time step was taken as the reach average value. The critical shear stress for the $D_{50}$ ($\tau_{critD50}$) was then calculated using Equation 2. And finally, the excess shear stress was calculated as ($\tau - \tau_{crit}$).

The variation in channel roughness was evaluated over the time-series using Equation 8 (Manning’s Equation, $n = (R^{2/3}S^{1/2})/v$). Velocity ($v$) was calculated for every time step by way of Equation 5 (continuity of flow $Q = v \times A$). Again a constant slope ($S$) of 0.019 m m$^{-1}$ was assumed. It is not expected that the slope (actually the energy slope) remains constant over time. However, the covariation of slope and roughness over time is complex and we were not
able to discern the exact relationship in this study due to measurement difficulties. Another model was constructed holding Manning's $n$ constant which yielded qualitatively similar results in terms of the duration of sediment transporting flows.

Sediment transport rates during peak flow were estimated using two flow-intensity-based transport equations: the Meyer-Peter and Müller (1948) equation and the Schoklitsch 1943 equation (Schoklitsch, 1950). The Meyer-Peter and Müller equation is in a class of shear stress based transport equations. It is an empirical equation and is the most widely used sediment transport equation for alpine rivers. A simplified version of the equation takes the form:

Equation 9: $q_s = W\alpha \left[ \tau - \tau_c^* (\gamma_s - \gamma) D_{50} \right]^{3/2}$

where $q_s$ is the bed load transport rate (kg s$^{-1}$), $W$ is the width of the channel (m), $\alpha$ is a constant ($\alpha = 0.04$), $\tau_c^*$ is the critical dimensionless shear stress for a chosen grain size (estimated at 0.06 – see section 3.6.2 for discussion), $\gamma_s$ is the unit weight of sediment (kN m$^{-3}$), and $\gamma$ is the unit weight of water (kN m$^{-3}$).

In contrast to the Meyer-Peter and Müller equation, the Schoklitsch equation is a discharge based equation which is written:

Equation 10: $q_s = 2500 S^{3/2} \left( Q - 0.6 \left( D^{3/2} / S^{7/6} \right) \right)$

where $Q$ is the water discharge (m$^3$ s$^{-1}$), and $D$ is the characteristic grain size (m). For graded sediments, Schoklitsch recommended that the $D_{40}$ should be used (Gomez and Church, 1989).

The path-length-based morphologic method was also used to estimate sediment transport rates. In order to use this method two important sources of data are required: first, detailed measurements of the total volume of erosion over the time step of interest; and second, the typical path length of those eroded sediments.
The total volume of erosion was estimated using the cross-sectional area of erosion measured at each of the 11 cross-sections in the middle reach of the study site between the 2005 and 2006 surveys. The total volume of erosion was interpolated for the zones between adjacent cross-sections. Interpolation was conducted using simple geometric assumptions following the methods of Martin and Church (1995). The equation for the change in volume of erosion was written:

\[
\Delta V_e = \frac{\Delta A_j + \Delta A_{(j+1)}}{2} L_{(j,j+1)}
\]

where \( \Delta V_e \) is the volume of erosion (m\(^3\)) between adjacent cross-sections \( j \) and \( j + 1 \), \( \Delta A_j \) is the area of erosion (m\(^2\)) as measured at cross-section \( j \), \( \Delta A_{(j+1)} \) is the area of erosion (m\(^2\)) measured at cross-section \( j + 1 \), and \( L_{(j,j+1)} \) is the distance along the channel centerline (m) between the two cross-sections (Martin and Church, 1995). The volume of deposition was also calculated using the same equation to identify if the erosion and deposition were balanced within the reach.

Path length was assumed to be the mean travel distance for all of the tracers. The error in the estimation of the mean was taken to be the standard error of the distribution shown in Table 7. Once the total volume of erosion and typical path length were known, the bulk sediment transport rate was calculated using the equation presented in Eaton and Lapointe (2001):

\[
Q_s = \rho_b \frac{v_e (L_{\text{MEAN}} / L_r)}{t}
\]

where \( Q_s \) is the bulk sediment transport rate (kg event\(^{-1}\)) over time \( t \) (event), \( \rho_b \) is the bulk sediment density (kg m\(^{-3}\)), \( v_e \) is the total volume of erosion (m\(^3\)), and \( L_r \) is the length of the reach (m) in which \( v_e \) was determined. The bulk sediment density was estimated by assuming a porosity of 0.29 (Eaton and Lapointe, 2001) Assuming a particle density of 2650 kg m\(^{-3}\), the corresponding bulk density for the load is 1890 kg m\(^{-3}\).

The largest source of error in the path-length-based morphologic method is likely the inability to account for scour and fill of the channel bed between surveys. Unfortunately, there is
no way of quantifying the magnitude of underestimation due to this error. The accuracy of the estimation also depends on the precision with which net erosion is measured and the uncertainty in the bulk sediment density due to variations in porosity. Eaton and Lapointe (2001) estimated the uncertainty from these sources to be on the order of 18%.

3.6.3. Results and Discussion

The model described in the methods section was used to estimate the reach average duration of competent flows during the 2006 freshet. Figure 34 contains the model output in the form of the shear stress, excess shear stress, and Manning’s $n$ values as a function of discharge. Manning’s $n$ values are on the left $y$-axis and shear stress is on the right $y$-axis. Manning’s $n$ values exhibit a gently decreasing trend with increasing discharge. Above about 2 m$^3$ s$^{-1}$, $n$ values range from 0.05 to 0.08. Values of $n$ are only greater than 0.10 at the lowest flows. Most likely, this effect is somewhat subdued in the field by changes in the energy slope at lower flows.

The model output also indicates shear stress ($\tau$) increases as a power function of discharge (Figure 34). This is because as discharge increases, shear stress ($\tau$) increases proportionally with the hydraulic radius ($R = A / P$). In this model shear stress ($\tau$) is calculated as $\tau = \rho g R S$; therefore, values are directly proportional to the increase in the hydraulic radius ($R$) as the remaining variables are held constant. The hydraulic radius ($R$) was calculated from cross-section surveys with the stage height determined from the recorded stage at nearby divers.

The critical shear stress for entrainment of the $D_{50}$ ($\tau_{crit}$) of the 2006 surface sediments was estimated to be 49 N m$^{-2}$. This shear stress value likely corresponds to a hydraulic mean flow depth of approximately 30 cm at threshold conditions during the 2006 freshet. In the model, excess shear stress ($\tau - \tau_{crit}$), increases approximately logarithmically as a function of discharge. Excess shear stress values indicate that threshold conditions for transport of the surface $D_{50}$ occurred at a discharge of approximately 1.8 m$^3$ s$^{-1}$ (Figure 34). This corresponds to roughly 65 days of potential sediment transporting flows during the 2006 season.
The estimated critical discharge of 1.8 m$^3$ s$^{-1}$ was supported by the observations of initial tracer movement. While we were not in the field to observe the onset of tracer movement during the 2006 freshet, we were there during the 2007 freshet to observe the initial movement of the same tracers reseeded in the same locations. The first tracer movement during the 2007 freshet was observed on April 8 at an estimated discharge of approximately 2 m$^3$ s$^{-1}$. This discharge is in reasonable agreement with the model output threshold conditions in Figure 34.

Sediment transport rates for the 2006 freshet were estimated using both the Meyer-Peter and Müller and Schoklitsch transport equations. The Meyer-Peter and Müller equation is sensitive to the value chosen for initiation of motion ($\tau^*$). Meyer-Peter and Müller (1948) suggest a critical dimensionless shear stress value of 0.047 be used. Other researchers have

![Figure 34. Model output reach average shear stress ($\tau$), excess shear stress ($\tau - \tau_{cri}$), and Manning's $n$ values for the recession limb of the 2006 freshet. The distribution of Mannings $n$ and shear stress are fitted with power functions while excess shear stress is fitted with a logarithmic function.](image)
proposed a range of $\tau^*$ values from 0.03 to 0.07 depending on channel characteristics. For this analysis, a $\tau^*$ value of 0.06 was estimated based on the observed initiation of motion of the 2007 tracers (Figure 35).

![Figure 35. (A) Initial seeding of 2007 tracers on launch line B, March 31, 2007. Photo taken from right bank (B) Observation of initial tracer movement on launch line B, April 8, 2007. Q ~ 2 m$^3$s$^{-1}$. Photo taken from left bank.]

The sediment transport rates ($q_s$) are calculated from each equation and are reported in mass transport per second (kg s$^{-1}$) and per day (kg day$^{-1}$). Values represent the sediment transport rate per unit length of channel at bankfull discharge. The results are located in Table 9 as well as the calculated values of bankfull shear stress ($\tau$), and the critical shear stress for the $D_{50}$ ($\tau_{crit}$). The bankfull shear stress (88 N m$^{-2}$) was nearly twice the critical shear stress for entrainment of the $D_{50}$ (49 N m$^{-2}$) assuming a critical dimensionless shear stress ($\tau^*$) of 0.06.

<table>
<thead>
<tr>
<th>$\tau$ (N m$^{-2}$)</th>
<th>$\tau_{crit}$ (N m$^{-2}$)</th>
<th>$q_s$ (kg s$^{-1}$) (Meyer-Peter and Müller)</th>
<th>$q_s$ (kg day$^{-1}$) (Meyer-Peter and Müller)</th>
<th>$q_s$ (kg s$^{-1}$) (Schoklitsch)</th>
<th>$q_s$ (kg day$^{-1}$) (Schoklitsch)</th>
</tr>
</thead>
<tbody>
<tr>
<td>88</td>
<td>49</td>
<td>101</td>
<td>8,730,000</td>
<td>45</td>
<td>3,850,000</td>
</tr>
</tbody>
</table>

There are significant discrepancies between the two transport equations (101 kg s$^{-1}$ vs. 45 kg s$^{-1}$). It is unclear which value may be more accurate; however, it is likely that together they represent a reasonable estimation of the range of sediment transport rates that may have occurred.
during bankfull discharge. The value of 101 kg s\(^{-1}\) estimated using the Meyer-Peter and Müller is high, but not unreasonable, when compared with other published transport rates (Gomez and Church, 1989). However, most transport formula tend to over-predict actual transport rates and have been reported to have errors of an order of magnitude. This is possibly due to the failure to account for the affect of surface coarsening or variations in the rate of sediment supply (Gomez and Church, 1989). For these reasons, the more conservative estimate of 45 kg s\(^{-1}\) calculated with the Schoklitsch formula is likely the more accurate estimation. Daily estimates of transport rates (kg day\(^{-1}\)) from both equations are exceedingly high (Table 9). This suggests that peak transport rates are not sustained for long periods of time. The duration of peak transport rates may depend more on the availability of unconstrained particles than flow duration.

An estimate of the sediment transport rate was also calculated using the path-length-based morphologic method. The total volume of erosion was calculated from the measured cross-sectional area of erosion and deposition surveyed from 2005 to 2006. The cross-sectional areas of erosion and deposition were calculated for each cross-section (XS 1 - XS 11). The results are located in Table 10. Ten zones were created representing the area between adjacent cross-sections. The distance between adjacent cross-sections \(L_r\) was measured as the distance along the channel centerline (Table 10). The total volume of erosion and deposition for each zone between adjacent cross-sections was calculated using Equation 11. The total volume of erosion throughout the reach \(V_e\) and the reach length \(L_r\) were then calculated as the sum of all zones.

The total volumes of erosion and deposition are well balanced for the reach. If we assume ± 4% uncertainty in the estimates as suggested by Eaton and Lapointe (2001), the difference in volume is nearly within the uncertainty. This is somewhat unexpected as the author’s field observations documented an area of erosion that was roughly hundreds of cubic meters in the upper reach of the study site during the 2006 freshet. It was assumed that much of the material from that erosion was deposited in the middle reach and therefore the volume of deposition would be much greater than the volume of erosion. The reason for the balanced erosion and deposition may be in the estimation the volume of erosion in the upper three zones (8, 9, and 10) where scour of the bed was documented. The bed scour at these cross-sections may have been related to the movement of LWD, and, therefore, may not have been distributed evenly throughout the entire zone. This could cause an over-estimation of the volume of erosion in these
zones. In any case, the balance of the two measures suggests that it is likely a reasonable estimate.

<table>
<thead>
<tr>
<th>Cross Section</th>
<th>XS Area (m²)</th>
<th>XS Area (m²)</th>
<th>Erosion Zone</th>
<th>Distance (Lₑ) (m)</th>
<th>Erosion (Vₑ) (m³)</th>
<th>Deposition (Vᵈ) (m³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>XS1</td>
<td>0</td>
<td>1.142</td>
<td>--</td>
<td>--</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>XS2</td>
<td>0.212</td>
<td>0.318</td>
<td>1</td>
<td>11.14</td>
<td>1.18</td>
<td>8.14</td>
</tr>
<tr>
<td>XS3</td>
<td>1.29</td>
<td>1.539</td>
<td>2</td>
<td>12.10</td>
<td>9.11</td>
<td>11.24</td>
</tr>
<tr>
<td>XS4</td>
<td>0.774</td>
<td>1.461</td>
<td>3</td>
<td>11.20</td>
<td>11.58</td>
<td>16.80</td>
</tr>
<tr>
<td>XS5</td>
<td>0.153</td>
<td>0.764</td>
<td>4</td>
<td>10.75</td>
<td>4.98</td>
<td>11.96</td>
</tr>
<tr>
<td>XS6</td>
<td>0.660</td>
<td>0.886</td>
<td>5</td>
<td>11.40</td>
<td>4.63</td>
<td>9.40</td>
</tr>
<tr>
<td>XS7</td>
<td>0.303</td>
<td>1.023</td>
<td>6</td>
<td>11.15</td>
<td>5.37</td>
<td>10.64</td>
</tr>
<tr>
<td>XS8</td>
<td>0.808</td>
<td>0</td>
<td>7</td>
<td>11.85</td>
<td>6.59</td>
<td>6.06</td>
</tr>
<tr>
<td>XS9</td>
<td>1.928</td>
<td>0</td>
<td>8</td>
<td>11.55</td>
<td>15.81</td>
<td>0</td>
</tr>
<tr>
<td>XS10</td>
<td>0.689</td>
<td>0</td>
<td>9</td>
<td>14.50</td>
<td>18.97</td>
<td>0</td>
</tr>
<tr>
<td>XS11</td>
<td>0</td>
<td>0</td>
<td>10</td>
<td>11.75</td>
<td>4.05</td>
<td>0</td>
</tr>
<tr>
<td>Total</td>
<td>6.821</td>
<td>7.133</td>
<td>--</td>
<td>117.39</td>
<td>82.27</td>
<td>74.24</td>
</tr>
</tbody>
</table>

155,000 kg 140,000 kg

Tracer travel distance was used to estimate the path length of eroded sediments. The mean path length for all tracers (Lₑ) was 103 m ± 7 m (see Section 3.5.3). Using the standard error of the mean (7 m), the minimum (96 m) and maximum (110 m) path length (Lₑ) were used to calculate a range estimate of the bulk transport rate (Qₛ) using Equation 12. The results are located in Table 11. Sediment transport rates are presented in kg per event (kg event⁻¹) for the reach and the same units per unit length of channel (kg m⁻¹ event⁻¹).

<table>
<thead>
<tr>
<th>Min. Lₑ (m)</th>
<th>Max. Lₑ (m)</th>
<th>Min. Qₛ (kg event⁻¹)</th>
<th>Min. unit Qₛ (kg m⁻¹ event⁻¹)</th>
<th>Max. Qₛ (kg event⁻¹)</th>
<th>Max. unit Qₛ (kg m⁻¹ event⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>96</td>
<td>110</td>
<td>127,150</td>
<td>1,083</td>
<td>145,693</td>
<td>1,241</td>
</tr>
</tbody>
</table>

Estimates of the sediment transport rate using the morphologic method appear to be far lower than those from the two flow-intensity-based transport equations. The maximum estimate
of 1,241 kg m$^{-1}$ event$^{-1}$ derived from the morphologic method would only require a short duration event at the transport rate of 101 kg s$^{-1}$ estimated for bankfull flow using the Meyer-Peter and Müller equation. There are at least two possible reasons for this disparity. First, as discussed earlier, scour and fill within events are not accounted for with the morphologic method. Given the long, complex, and multiple peak hydrograph that occurred during the 2006 season, it is possible that several scour and fill events may have gone undetected by the 2005 and 2006 surveys. Second, it is possible that the higher rates of sediment transport estimated by the flow-intensity-based equations roughly represent local maximum sediment transport rates during peak flow. However, those transport rates may be highly spatially and temporally variable and only occur for short durations if at all. In any case, the transport estimates indicate that the actual sediment transport rates are uncertain and the results are inconclusive.

If we assume that flow-intensity-based equations represent an idealized maximum transport rate (i.e. transport capacity), and estimates of transport rates based on the morphologic method represent a minimum seasonal transport rate estimation, then it follows that high intensity transport rates likely occur for short periods followed by long periods of moderate to low transport rates. In the case of Fishtrap Creek, this condition appears to have occurred despite the fact that shear stress conditions were above the threshold for entrainment of the $D_{50}$ for a long duration (~65 days). These conclusions are in agreement with the stage-discharge regressions presented in Section 3.4.3 that documented the majority of the channel adjustment (and possibly sediment transport) occurred over a relatively short period of time. Other researchers have also documented high temporal variability in sediment transport rates (e.g. Tunnicliffe et al., 2000).
Chapter 4

4. SUMMARY AND CONCLUSIONS

4.1. Summary

The McLure Fire of 2003 was a significant disturbance to Fishtrap Creek watershed. During the summer of 2006 (3 years after the fire) an intensive monitoring program was established at Fishtrap Creek. The primary purpose of this study was to document the timing and magnitude of initial changes in channel morphology, hydrology, sediment sources, and sediment mobility in Fishtrap Creek in the aftermath of the McLure Fire, and to link those observations to changing governing conditions by way of physical processes. The monitoring program was designed to compare and contrast current conditions with existing data that was collected in the years following the fire as well as establish a baseline for future researchers to build upon. Monitoring techniques included study area mapping, grain size analysis, cross-sectional surveys, low-level reach photography, multiple water level recorders, and magnetic tracer stones.

Results from the monitoring indicate that there was little change in the morphology of Fishtrap Creek in the first two years following the fire. Then, during the 2006 freshet, widespread bank erosion began occurring, apparently in response to decaying riparian root systems. As a result, dead vegetation on the banks is beginning to be recruited into the channel and the sediment input from bank erosion has considerably increased the supply. Currently, the channel is in an active state of transition resulting from the increased sediment inputs from bank erosion and bed aggradation.

Cross-sectional surveys indicate that channel form adjustments were much greater in 2006 than in 2005 despite the fact that the discharge was larger in 2005. Localized bed aggradation is occurring in areas where the sediment supply has exceeded transport capacity. A channel avulsion and overbank flows were observed as the stage height was forced higher due to aggradation. In addition, analysis of stage-discharge relationships indicate that channel form adjustments observed during the 2006 freshet occurred over a relatively short time period between the first and second peak in the hydrograph.
Travel distance recorded from magnetic tracer stones suggest that sediment transport distances were relatively high during the 2006 freshet when compared with other studies. The longest recorded travel distance reached 52 times the channel width. The mean tracer transport distance was more than twice the average pool spacing which is the expected mean transport distance suggested by several researchers. In addition, tracer burial depth distributions support the field observation that Fishtrap Creek does not have well-developed surface structures. In addition, the burial depth distributions at Fishtrap Creek are characteristic of systems that experience substantial scour and thorough mixing of the active layer.

The model for estimating the duration of competent sediment transporting flows indicates that the reach average boundary shear stress was above the threshold for entrainment of the $D_{50}$ for an estimated 65 days during the 2006 season. Actual sediment transport rates are uncertain because there was a large discrepancy between the flow-intensity-based sediment transport estimations and the morphologic method of sediment transport estimation. It is possible that the flow-intensity-based sediment transport estimations are representative of the high-intensity transport that likely occurs at Fishtrap Creek for short periods whereas the morphologic method incorporates the entire duration of sediment transport which may include long periods with low to moderate sediment transport; however, the results are inconclusive.

4.2. Predictions of Future Conditions

Given our current knowledge, it is not possible to accurately predict the exact future conditions of an alluvial channel. However, there have been studies that have documented the long-term response of stream channels to fire. Robinson et al. (2005) examined the long-term response of a number of streams, in terms of channel morphology, and found that the majority of channel change occurred during the first 10 years following fire with little change in the next 10 years. This time scale roughly corresponds with the theoretical model of maximum root cohesion due to the decay and regeneration of riparian root systems (Section 3.3.1 Figure 12). Therefore, if our assumptions are correct, the current trend of channel transformation would be expected to continue in Fishtrap Creek until the root systems of emerging vegetation begin to stabilize the banks. In the next several years, it is expected that the morphology of Fishtrap Creek will transition from a relatively stable, predominantly single-thread channel to an unstable system that episodically transitions among multiple channels.
This theory was supported by the author’s field observations during the 2007 freshet. Observed changes in channel morphology in 2007 were much greater than in 2006; and again, they occurred in response to average flow conditions. The volume of erosion documented during the 2007 freshet was an order of magnitude greater than that observed during the 2006 freshet. The greatest amount of bank erosion in 2007 occurred in the upper reach of the study site where 14 m of the bank retreat occurred over the period of two days. The erosion resulted in a large fan deposit at the downstream end of the upper reach. Aggradation was widespread in the middle and lower reaches. In many places, the elevation of the top of the bar is equivalent to the top of the bank. This aggradation forced stage height higher and caused overbank flows – even at relatively low discharges. Figure 36 illustrates the change in the left channel in the upper reach of the study site. At this point, the channel widened by roughly 12 m during the 2007 freshet.

The question remains: will Fishtrap Creek eventually recover to a channel morphology similar to that before the fire, or will it transition to a new equilibrium state. We speculate that, in many watersheds, particularly in the interior of British Columbia, this type of response is quite common although it is under represented in the literature. Furthermore, we expect that in the dry interior streams of British Columbia, forest fire may be the single largest source of sediment and LWD to streams. In fact, it is likely that in many of these systems, channel morphology remains relatively stable for decades between fire events; and may only change significantly following fire. This suggests that the channel morphology of Fishtrap Creek is likely an equilibrium morphology that, over a sufficiently long time scale, is the product of the response to repeated stand-replacing fires that have a recurrence interval of 150-200 years.
4.3. Improvements and Suggestions for Further Research

Modifications could be made to improve the stream channel monitoring presented in this study. Most notably, results could be improved by increasing the spatial resolution and frequency of the measurement of channel form. The spatial resolution issue could be addressed by a more detailed characterization of the bed topography possibly using photogrammetry or a LIDAR or total station survey of the bed. The frequency of measurement issue is more difficult to address for two reasons. First, surveys are time-consuming and expensive. Second, changes in channel form are occurring at near bankfull flows when traditional channel survey methods are not possible. There may be other methods of surveying the channel bed at high flows, such as bathymetric-type surveys; however, this may prove difficult in shallow flow and when the bed is mobile. I believe that increased frequency bed surveys, especially during peak flows, are the key to obtaining accurate sediment transport estimations using the morphologic method.

Further research is needed to develop models for identifying the likely response of a given stream channel to fire based on the post-fire conditions in the watershed (e.g. soil hydrophobicity, slope stability, burn severity etc.). Continued monitoring of changes in channel morphology and sediment dynamics in Fishtrap Creek and at other sites is needed for testing and developing models of channel response and recovery. I believe that watershed scale models of sediment and water input coupled with numerical modeling of stream channel dynamics (e.g. Eaton, 2006) are the necessary tools to be able to reasonably predict stream channel response to fire.

4.4. Conclusion

The research in this thesis indicates that channel morphology may remain quite stable for a period of years following fire. This delayed response may be relatively common in post-fire systems where significant soil hydrophobicity does not occur. The lag time between disturbance and channel response is directly related to the decay of riparian vegetation root systems. Once the root systems have decayed sufficiently for bank erosion to begin, the response is dramatic. There appears to be an abrupt threshold that, once exceeded, results in high-magnitude, rapid adjustment with intense sediment transport rates. The result is bar growth, channel aggradation, and channel avulsions once transport capacity is exceeded.
It is important for resource managers to be aware of the lag time associated with the loss of bank strength following the burn of riparian areas. Having the ability to identify these systems early may mitigate damage or destruction of fish habitat, and damage to roads, property, and other infrastructure, due to flooding. In addition, understanding the process-form interaction that occurs during the post-fire response and recovery sequence is key to improve our understanding of landscape evolution in the interior of British Columbia.

The results of this study have improved our understanding of the response of gravel bed streams to severe wildfire and have provided an estimate of the time lag following fire when significant channel transformations may be expected to occur in alluvial channels where riparian root systems provide bank strength. The results are also relevant to the impacts of other forest fires, and potentially to the impact of mountain pine beetle and other vegetation disturbances on stream channel stability.
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APPENDICES


**XS A**

BED JULY 12, 2006

WATER SURFACE, Q = 0.53 m$^3$s$^{-1}$

**XS B**

**XS C**

**XS D**

Data source: British Columbia Ministry of Forests and Range.

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Appendix B. Surveys XS E - XS H – 2006.

XE

BED JULY 12, 2006
WATER SURFACE, Q = 0.53 m$^3$s$^{-1}$

XF

XS G*

XS H*

DISTANCE ACROSS CHANNEL (m)

* Note that the horizontal scale is 0 to 20 m rather than 0 to 16 m.

Data source: British Columbia Ministry of Forests and Range.
Appendix C. Surveys XS 12, XS 13, and XS 14 – 2006 (right and left channel).

* Note that scale of the horizontal axis changes and XS 14 is divided into the right and left channels.

Data source: British Columbia Ministry of Forests and Range.
Appendix D. Surveys XS 15 and XS 16 – 2006 (left and right channels).

* Note that 4 m of the floodplain between the right and left channel in XS 15 and XS 16 is not displayed.

Data source: British Columbia Ministry of Forests and Range.
Appendix E. Surveys XS 17 and XS 18 – 2006 (left and right channels).

* Note that 2 m of the floodplain between the right and left channel of XS 17 is not displayed. XS 18 is divided into the right and left channel with the entire cross-section displayed.

Data source: British Columbia Ministry of Forests and Range.

Data source: British Columbia Ministry of Forests and Range.

Graph showing ARBITRARY ELEVATION (m) vs. DISTANCE ACROSS CHANNEL (m).

Legend:
- - - BED JULY 10, 2006
- - - WATER SURFACE, Q = 0.53 m³ s⁻¹
Appendix G. Pole photography and photographic mosaic methods.

As discussed in Section 2.7, low-level reach photography of the middle reach of the study site was captured in 2004 by Tim Giles of British Columbia Ministry of Forests and Range, and Steve Bird of Fluvial Systems Research Inc. They photographed the creek by suspending a camera on a flexible 10 m telescoping pole with a gimbal mount. Then, using image interpretation software (ERDAS), they were able to generate an orthorectified photographic mosaic of the 2004 channel.

The 2004 and 2006 mosaics are presented as Appendix H and Appendix I respectively. Image resolution has been dramatically reduced in both images for display. The resolution of the original images is easily sufficient at the scale of individual grains. There are several observations to be made in these images. In the 2004 mosaic, the two people that repeat throughout the image are actually the photographers holding the pole with the camera on it. Notice that in the 2004 mosaic there are no exposed bars except where sediment is accumulating behind LWD and the bed appears relatively uniform. On the contrary, in the 2006 mosaic there are exposed bars in many places and the bedforms appear to be highly developed. This is not exactly a fair comparison though as the flow is higher in the 2004 image. However, estimating the change in for individual cross-sections indicates that most of the bars would still be exposed even at the higher discharge. And finally, the movement of several pieces of LWD can also be found by comparing the two mosaics.

The 2006 mosaic was captured using a digital SLR camera. The camera was mounted to a rigid 9 m telescoping pole. The angle of the camera was adjusted 20° from vertical so that the base of the pole and the photographer would not be in the image. A grid of camera locations was established and marked with wooded stakes. The grid was strategically placed to capture the entire channel with 60% overlap in the downstream direction and 20% overlap in the across channel direction; this was done to have enough overlap to use the images for photogrammetry.

Hundreds of Ground Control Points (GCPs) were placed throughout the reach. The average spacing was roughly 3-7 GCPs per image. The $x$, $y$, and $z$ coordinates of the GCPs were surveyed using a total station. The photos were then registered to geographic coordinates using the known location of the GCPs. This was done using a technique called georegistration in a GIS.
(ArcGIS). Error in this process increased radially out from the principle point of the image. The absolute error from the GCP in the image and the surveyed GCP ranged from 0 to 0.15 m. Notice that the LWD that is elevated off the bed does not line up from image to image. This is because of the camera angle and is not representative of the amount of error at the bed surface. Finally, the individual images were mosaiced into a single image in ArcGIS.
Appendix H. Middle reach photographic mosaic – 2004.

Photography taken September 22, 2005. $Q = 1.0 \text{ m}^3 \text{ s}^{-1}$. Data source: British Columbia Ministry of Forests and Range.
Appendix I. Middle reach photographic mosaic – 2006.

Photography taken August 9 -12, 2006. $Q = 0.18 \text{ m}^3 \text{ s}^{-1}$. 