LINKING TEMPORAL AND SPATIAL VARIABILITY OF MILLENNIAL AND
DECADAL-SCALE SEDIMENT YIELD TO AQUATIC HABITAT IN THE
COLUMBIA RIVER BASIN

by

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A thesis submitted in partial fulfillment
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ABSTRACT

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by

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Utah State University, 2014

Major Professor: Dr. Patrick Belmont
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Eco-geomorphic interactions occur across a range of spatial and temporal scales from the level of the entire watershed to an individual geomorphic unit within a stream channel. Predicting the mechanisms, rates and timing of sediment production and storage in the landscape are fundamental problems in the watershed sciences. This is of particular concern given that excess sedimentation is considered a major pollutant to aquatic ecosystems. Rates of sediment delivery to stream networks are characteristically unsteady and non-uniform. Because of this, conventional approaches for predicting sediment yield provide incomplete and often inaccurate information. Terrestrial cosmogenic nuclides (TCNs) provide an estimate of spatially averaged rates of sediment yield from $10^1$ to $10^4$ km$^2$ and temporally integrated from $10^3$ to $10^5$ years. Here, I used TCNs to constrain unsteadiness and non-uniformity of sediment yield within specific catchments of the Columbia River Basin (CRB). This is in combination with GIS analysis optically stimulated luminescence (OSL), Carbon-14 (C14) dating of fluvial deposits, and rapid geomorphic assessments.
Results showed an order of magnitude spatial variability in the rates of millennial-scale sediment yield at the scale of the entire CRB. At the broadest scale long-term rates of sediment yield generally are poorly predicted from topographic and environmental parameters. A notable exception is the observed positive correlation between mean annual precipitation and sediment yield. Where functional relationships exist, the nature of those relationships are scale and situation-dependent. In addition to the broadest scale, each smaller watershed (e.g., ~ 10 – 2,000 km²) has a distinct geologic, geomorphic, and disturbance history that sets the template for the modern sediment dynamics and the physical aspects of aquatic habitat. Chapter 2 presents results of broad-scale trends while Chapter 3 is comprised of case studies from smaller watersheds. Finally, Chapter 4 explores the relationship between long-term sediment yield and modern channel form.

(236 pages)
PUBLIC ABSTRACT

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Historically the Columbia River was amongst the most productive salmonid-bearing watersheds in North America. Currently, salmonid populations have collapsed and are estimated at ten percent of historic levels. Because of this, there are thirteen, Endangered Species Act (ESA) listed distinct population segments within the Columbia River Basin (CRB). The Bonneville Power Administration (BPA) operates a substantial number of hydroelectric dams within the CRB. As the dams effectively block anadromy and are thought to be extremely detrimental to the long-term persistence of salmonid populations within the CRB, the BPA funds a series of programs to mitigate the impacts of dam operation. One such program is the National Oceanographic and Atmospheric Administration’s (NOAA) Integrated Status and Effectiveness Monitoring Program (ISEMP). ISEMP’s goal is to develop and test strategies for determining the status and trend of salmonid populations and their habitat in the CRB. In 2010 ISEMP started the Columbia Habitat Monitoring Program (CHaMP) to initiate and implement a standardized approach to salmonid monitoring within the CRB. ISEMP and CHaMP are
partial funders of this thesis study and desired watershed-level geomorphic context for their restoration and monitoring efforts within the CRB. Specifically this project consisted of a first-order examination of the following broad research questions: 1) How do long-term rates of sediment supply vary spatially and temporally throughout the Columbia River Basin (CRB)? 2) How have human activities influenced (amplified or dampened) processes of erosion and sediment transport? 3) At what scales do long-term and near-term rates of sediment yield influence aquatic habitat metrics?

I addressed these questions at multiple spatial and temporal scales within the CRB primarily using cosmogenically derived catchment-averaged millennial-scale erosion rates. The principal findings were:

- There is an order of magnitude variability in long-term sediment yield at the scale of the entire CRB.
- This spatial variability is generally poorly predicted from simple topographic and environmental parameters. A notable exception is a strong positive correlation between long-term sediment yield and mean annual precipitation.
- Where functional relationships exist between topographic and environmental parameters and sediment yield the nature of those relationships is scale and situation dependent.
- Within Bridge Creek, John Day Watershed, Oregon I showed how CAER can be coupled with C14 or OSL dates and stratigraphic analysis to understand timing and duration of channel cut and fill cycles.
• Within the Cascade Range in Washington I illustrated how the geomorphic history of a watershed (e.g., glaciation) can have a profound influence on modern sediment dynamics.

• CAER can be used to effectively quantify changes in sediment supply at the watershed level if those changes are drastic (e.g., transition from glacial dominated to fluvial dominated erosion) but appears to be too coarse in the absence of near-term rates (i.e., yearly to decadal sediment yield) to effectively quantify the temporary pulse of sediment associated with wild-fire (e.g., Tower fire, EBNFC, JDW) if the scale of disturbance isn’t sufficiently large.

• I identified a positive correlation in the Cascades, WA between long-term sediment yield and potentially alluvial reaches.
ACKNOWLEDGMENTS

Many different organizations and people made this work possible. I would like to thank my funders: ISEMP, CHaMP, Ecological Research Inc., Terraqua Environment Consulting Inc., Utah State Universities’ Agricultural Experiment Station and the Geological Society of America Research Grant. I also want to thank my advisor Patrick Belmont. His scientific curiosity and keen ability to read the landscape inspired and guided me throughout this effort. Patrick has greatly influenced my development as a geomorphologist in a wide variety of ways but specifically he strengthened my technical writing, spatial analysis and field-based observational skills. Patrick also showed me that wide-eyed wonder and respect for the complexity and beauty of natural systems can guide and inspire rigorous scientific inquiry. I also want to thank my committee members Nick Bouwes and Tammy Rittenour. Nick provided an important perspective in attempting to make the link from sediment yield to aquatic habitat. Tammy provided important geomorphic interpretations and allowed me to use her Luminescence Laboratory to process my optically stimulated luminescence samples.

I also want to thank Caty Clifton, a hydrologist with the Malheur National Forest. Caty provided logistical information about sampling in the John Day Watershed and sediment yield data. Greg Nagle also contributed geomorphic information about the John Day Watershed. Marwan Hassan provided model predicted sediment yield information from the CRB. I also want to send a big thank you to Justin Stout, a past graduate student of Patrick Belmont. Justin helped me in a plethora of ways with every aspect of my project. His enthusiasm and competency were inspiring and I couldn’t have done this without him. He is truly a good friend. Scott Shahverdian was also a big help in the field
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Most importantly I want to thank my family and friends that have provided emotional support throughout this difficult process. My wife, Sara Bangen, has supported me in every way imaginable. Lastly, I would like to dedicate this effort to my grandfather William Shepard, who passed away during my master’s degree. My grandfather Bill was my teacher and friend throughout my life and I will always remember and be inspired by his compassion, wit, and enthusiasm for living life to the fullest.

Elijah Portugal
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CHAPTER 1
LINKING TEMPORAL AND SPATIAL VARIABILITY OF MILLENNIAL AND
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1.1 Introduction

Understanding and predicting the mechanisms, rates and timing of sediment
production and storage in the landscape are fundamental problems in geomorphology and
Earth surface sciences (Trimble, 1977; Wolman, 1977; Walling, 1983; Schaller et al.,
2001; National Research Council, 2010). Sediment production and storage within a
watershed are difficult to quantify and predict because of high levels of ‘natural’ and
human-induced variability in both time and space (Kirchner et al., 2001; Ferrier et al.,
2005; Smith et al., 2011). This variability is driven by a host of interrelated hydrologic,
geomorphic and environmental processes that can be influenced by human land-use
(Walling, 1983). Sediment entrainment, transport and deposition mechanisms are highly
non-linear in regards to hydrologic flux and sediment availability. Further, sediment can
be stored and remobilized many times during transport from primary source to ultimate
sink. The percentage of sediment that experiences temporary storage can range from 0 to
100%, with storage lasting from a few days to thousands of years (Allen, 2008; Skalak
and Pizzuto, 2010; Smith et al., 2011). For these reasons, many widely applied
conceptual and empirical models that attempt to predict long-term erosion and/or
sediment yield from simple parameters like mean watershed slope and mean annual
precipitation have been shown to be incomplete and inadequate (Riebe and Kirchner,
2001a; Riebe and Kirchner, 2001b; Von Blanckenburg, 2006).
Trimble (1999) provides an excellent case study illustrating the complexity of sediment yield dynamics over time in Coon Creek, an agricultural watershed in Wisconsin. Trimble constructed a sediment budget accounting for 140 years (1853-1993) since Euro-American settlement and found that total sediment efflux leaving the watershed remained relatively steady over time 39,000 T yr$^{-1}$, despite the fact that terrestrial erosion decreased from 441,000 to 117,000 T yr$^{-1}$, due to changes in agricultural practices. Large shifts in sediment sources and sinks within the watershed were not reflected in the total efflux leaving the watershed. This is compelling evidence of the complexity of sediment source, sink and flux dynamics within a watershed.

Geochemical techniques based on the concentration of terrestrial cosmogenic nuclides (TCNs) in river sediment have developed rapidly over the last 20 years and are now widely used to quantify rates of sediment yield at the watershed scale (Gosse and Phillips, 2001; Von Blanckenburg, 2006; Belmont et al., 2007). Coupled with well-established geochemical dating techniques (e.g., radiocarbon, C14) and other emerging methods, (e.g., optically stimulated luminescence, OSL) these approaches are revolutionizing our ability to understand and predict watershed-wide sediment dynamics and improving our ability to link estimates of landscape-scale sediment supply with aquatic habitat.

In a seminal global review, Von Blanckenburg (2006) compared millennial-scale catchment-averaged erosion rates (CAER) derived from TCNs in river-borne quartz, to the simple environmental parameters of mean watershed slope and mean annual precipitation and found no correlation. In another recent global review, Portenga and Bierman (2011) compared 1599 CAER and rock outcrop erosion rates derived from TCN
concentrations to a host of simple environmental parameters (e.g., latitude, elevation, relief, mean annual precipitation and temperature, seismicity, watershed slope, watershed area, and vegetation cover within a basin). They found that individually none of the parameters accurately explained the variability in millennial-scale erosion rates but when combined in multiple regression analysis the variability was better explained (e.g., final $R^2 = 0.60$). In the tectonically active Sierra Nevada Range, California, Riebe and Kirchner (2001a; 2001b) found that millennial-scale CAER correlate to mean watershed slope only when watersheds were located in close proximity to active faults. These examples are further evidence that simple bivariate relationships between long-term erosion and simple environmental parameters often do not sufficiently account for the complexity of the erosion and sediment routing system at the watershed, or landscape, scale. With the use of TCNs we are beginning to unravel the complex and spatially variable processes of long-term watershed-wide erosion and sediment yield in ways that were previously not possible. This is particularly relevant within the Columbia River Basin (CRB) as it contains a wide range of physiographic settings.

Another example from Kirchner et al. (2001), working in 32 mountainous catchments in central Idaho, analyzed sediment yield over decadal, millennial and million year timescales and found that long-term yield (millennial and million year time scales) was roughly spatially uniform, and on average 17 times higher than decadal yields. This data suggests that modern sediment delivery from mountainous terrain can be relatively low, with long-term averages significantly higher due to low-frequency episodic delivery via high magnitude mass-movement events following forest fires. Because of this temporal variability, conventional approaches for predicting sediment yield (e.g., suspended and bed-load sediment yield gages and sediment accumulation rates in reservoirs) provide incomplete, and potentially misleading information
(Schaller et al., 2001; Ferrier et al., 2005). Additionally, the time and cost involved in establishing and maintaining an effective watershed-wide gaging network (decades of work and $10^5 - $10^6) is substantially greater than the time and cost required for a small number of TCN samples to derive a catchment-averaged erosion rate (CAER; one year and ~$10^3 per sample).

Though difficult to constrain, developing a better understanding of long-term watershed-wide sediment dynamics is critical in the context of both basic and applied research particularly as it pertains to analyzing the link between sediment supply and channel form, which forms the physical template of aquatic habitat. Lane (1954) established the now famous qualitative proportionality commonly referred to as Lane’s Balance (Figure 1.1) that directly relates the four primary controls on channel form, including water discharge and channel gradient, which are balanced by the quantity and caliber of sediment supplied to the channel. Fundamentally, this relationship illustrates that changes in any one of the variables (e.g., amount and/or caliber of sediment flux or timing and magnitude of water flux) directly affects the others such that they adjust to maintain continuity in the water and sediment routing system (i.e., the channel network).

Within basic research, unraveling the processes and rates of sediment yield is crucial for understanding landscape evolution in mountainous environments (Molnar and England, 1990; Pazzaglia, 2003), and evaluating erosional impacts of natural perturbations (e.g., forest fires, landslides, etc.) (Robichaud and Waldrop, 1994; Moody and Martin, 2009). From an applied perspective, analyzing sediment yield can identify the erosional impacts of anthropogenic land use (Montgomery et al., 2000; Hewawasam et al., 2003; Belmont et al., 2011), predict rates of sediment loading in stream ecosystems (Anderson, 1996; Gregory and Bisson, 1997), and inform in-stream anadromous salmonid restoration
efforts that attempt to rehabilitate or restore the physical component of aquatic habitat (Montgomery, 2004; Roni et al., 2008)

Figure 1.1: Lane’s Balance. A conceptual diagram linking water discharge and channel gradient (i.e., transport capacity) with the quantity and caliber of sediment supplied to the channel. Fundamentally, this relationship is responsible for the form of the channel which is the physical template of instream aquatic habitat. From Rosgen (1996) adapted from Lane (1954).

Developing a better understanding of erosion and sediment supply to small (2 – 6th order) channels throughout the Columbia River Basin (CRB) is the overarching theme of the research conducted here. Specifically, we use CAER to estimate the long-term, natural rates of sediment yield and evaluate the spatial and temporal variability of those rates for the purposes of basic research, monitoring, and restoration related to in-stream anadromous salmonid habitat.
1.2 Introduction to Fluvial Terraces

When a floodplain is abandoned by incision of its associated river the resulting landform is called a terrace (Anderson and Anderson, 2010; Pazzaglia, 2013). Terraces are typically formed when the river undergoes a change in dynamic equilibrium from an existing regime of sediment and water flux to another (Dury et al., 1972; Veldkamp and Vermeulen, 1989). There are a variety of external variables that can cause a shift in sediment and water flux including changes in climate (Warner, 1992; Hancock and Anderson, 2002; Fuller et al., 2009), tectonics (Riebe and Kirchner, 2001a), base level (typically either through uplift or sea level change) (Pazzaglia et al., 1998; Belmont, 2011) or anthropogenic effects (DeLong et al., 2011). The interpretation of terrace sequences, stratigraphy and the dating of deposition and incision events can be used to understand past river dynamics.

Two main types of alluvial terraces exist, fill and strath. A fill terrace is primarily formed by depositional processes, such as when sediment accumulates behind a temporary obstruction (e.g., beaver dam, mill dam, woody debris jam) in the channel and the channel subsequently incises through the deposit when the obstacle is removed (Merritts et al., 1994). The sequences of erosion and deposition of these terraces can be difficult to interpret because of the episodic nature of their development and the complexity of the stratigraphic sequences (Leopold et al., 1964; Bridge, 2009). With the advent of modern dating techniques (e.g., optically stimulated luminescence (OSL), cosmogenic nuclide dating and Carbon-14 (C14)), understanding the process of fill terrace formation has improved but complexity and challenges remain.
Strath terraces differ from fill in that they are primarily formed by channel incision. Typically strath terraces are composed of thinly mantled alluvium (i.e., approximately the thickness of the floodplain) capping laterally planed bedrock (Hancock and Anderson, 2002). The conceptual formation of strath terraces was first described by Gilbert and Dutton (1877) in the Henry Mountains of Utah where he assumed that during periods of increased aggradation relative to stream power the vertical incision component of river migration was lessened, leading to increased potential for lateral incision. Mackin (1937) expanded on this idea and proposed that increased lateral incision into valley walls would contribute to valley expansion during strath planation. Vertical incision of the channel subsequent to lateral planation leads to floodplain abandonment and thus formation of a terrace, a flat alluvial feature that is no longer inundated by floods on a regular basis (Hancock and Anderson, 2002). Lithology of the bedrock and the level of valley confinement affect a river’s ability to laterally incise and form a strath terrace (Finnegan and Dietrich, 2011).

Terraces provide information regarding past geomorphic and climactic regimes. Thus many techniques have been developed to interpret their stratigraphy and date their formation. Stratigraphic interpretation of old floodplain deposits has been and continues to be a major component of geomorphic inquiry (Bridge, 2009; Anderson and Anderson, 2010). The relative thickness and arrangement of depositional layers as well as the lithology and coloring of those layers can be mapped and correlated to other terraces to aid in understanding a river’s history. Sediment size and grading are also indicative of past processes. For example, Baker (1974) and Pierce and Scott (1983) found that many terrace deposits show evidence of sediment transport of larger grains that would not be
possible to transport with the stream power of the modern river, indicating a depositional period with increased peak discharge or slope. Other researchers (Ritter, 1967; Sinnock, 1981; Jackson et al., 1982) have found evidence of changes in river planform over time (i.e., braided instead of meandering).

1.3 Research Context

The CRB was historically one of the most productive anadromous salmonid fisheries in the world (Williams et al., 1999) providing critical food, cultural and economic resources for the original native inhabitants of the Pacific Northwest and subsequent settlers. Prior to the construction of the Dalles dam in 1957, Celilo Falls, a tribal salmon fishing area on the lower Columbia, was the longest continuously inhabited native community in North America dating back over 15,000 years (Dietrich, 1995). Over the last century human activities have led to an approximate 90% reduction in salmonid population levels (Williams et al., 1999). This severe reduction is thought to be the result of a combination of four main factors: commercial overfishing, modified hydrology through large and small scale dams and irrigation diversions, genetic pollution of wild stocks due to unintended effects of hatchery operations, and large-scale degradation of in-stream physical habitat (Dietrich, 1995). Anadromous salmonids life history strategy requires a return to the natal stream for spawning and thus salmonids are particularly vulnerable to in-stream barriers to anadromy and the disturbance history and degradation of physical habitat conditions.

The collapse of anadromous salmonid populations throughout the CRB led Congress in 1980 to pass the Northwest Power Act mandating watershed-wide mitigation for salmonids affected by dam operations. The Endangered Species Act (ESA) listing of
salmon and steelhead populations in the CRB and subsequent legal actions by tribal entities and state agencies as well as biological opinions by the National Oceanic and Atmospheric Administration (NOAA) mandated the Bonneville Power Administration (BPA), who operates the Columbia River dams, to invest hundreds of millions of dollars on anadromous salmonid monitoring, restoration and conservation projects. In 2011 alone BPA’s Fish and Wildlife Program’s (FWP) mitigation budget was estimated at $236 million with a large portion of that allocated to monitoring and restoring in-stream physical habitat conditions (BPA, 2010). The research proposed here will seek to provide geomorphic context by establishing rates of long-term sediment yield and identifying sediment erosion or deposition hotspots to inform salmonid restoration and habitat monitoring efforts for specific sub-basins throughout the CRB. Specifically, research objectives are to understand spatial and temporal variability of sediment supply to small tributaries (2-6th order) distributed throughout the diverse landscape that is the CRB. Secondly, I seek to understand if/how different parts of the CRB are more or less sensitive to anthropogenic and natural erosion. Third, I investigate how differential rates of sediment supply influence the physical component of aquatic habitat. Specifically, I compare the type and distribution of channel units (e.g., pool/riffle, plane-bed, cascade, etc.) delineated by Dr. Tim Beechie (NOAA’s Northwest Fisheries Science Center) to CAER in specific sub-watersheds within the CRB.

1.4 Long-Term Sediment Supply and Aquatic Habitat

Many researchers recognize the linkage between physical and hydrological processes and aquatic habitat at multiple temporal and spatial scales (Resh et al., 1988; Naiman, 1992; Bayley and Li, 1994; Montgomery and Buffington, 1997). Yet, our
ability to make quantitative predictions based on understanding of these linkages remains underwhelming (Imhof et al., 1996; Naiman et al., 1999; Petts, 2000). At the broadest scale, long-term erosion rates (i.e., $10^3$ – $10^5$ yrs) provide an estimate of sediment contributed to the channel network averaged over timescales much longer than most sources of variability. Long-term sediment yield coupled with complex feedbacks involving climate and tectonic setting control watershed topography including valley width and stream gradient (Montgomery, 2004; Anderson and Anderson, 2010). Valley width and stream gradient are, in turn, two primary controls determining the sediment transport regime, which structures reach-scale channel morphology and local sediment routing behavior (Montgomery and Buffington, 1997; Church, 2002; Brierley and Fryirs, 2005) (Figure 1.2).

Channel morphology and ultimately aquatic habitat diversity are linked to long-term physical processes (Yarnell et al., 2006). Specifically, the balance between the rate of long-term sediment supply and the transport capacity of the channel (which is fundamentally established by the channel gradient and amount of flow) functions as the primary control in shaping the physical template of in-stream aquatic habitat in alluvial (i.e., self-formed) channels (Imhof et al., 1996; Montgomery and Buffington, 1997). For example, channel incision is common throughout the arid and semi-arid regions of the world, including the CRB (Pollock et al., 2007), and is thought to be linked to changes in sediment supply or flow regime and often results in degraded aquatic habitat conditions (Williams et al., 1991; Elmore et al., 1994). Channel incision can simplify the channel planform (e.g., channel straightening) and restrict access of high flows onto the floodplain resulting in a paucity of instream flow diversity for salmonids, among other
negative attributes. The primary driver(s) of wide-spread channel incision is/are debatable but in many locations is thought to be related to anthropogenic changes in land-use (e.g., soil compaction) (Cooke and Reeves, 1976). This change can result in an increase in run-off, which increases the channel boundary shear stress, leading to vertical incision and aquatic habitat degradation.

Salmonids and other stream biota have specific substrate requirements that differ at various life stages (Gregory and Bisson, 1997; Montgomery, 2004). These requirements force salmonids to use different portions of the channel network at different times for spawning and rearing habitat, which are directly structured by channel morphology. For example, the type and distribution of channel units (pools, riffles, and bars) also determine population-level salmonid habitat availability and capacity. Thus, the detrimental effects of excessive fine sediment or coarse sediment starvation on salmonids (Bisson et al., 1992; Waters, 1995; Suttle et al., 2004) and other stream biota (Death and Winterbourn, 1995; Rabenf et al., 2005) is ultimately controlled by the sediment yield and transport regime within a watershed (Yarnell et al., 2006).

Considering the importance of physical and hydrological processes in structuring aquatic habitat, as well as the complexity of erosion, transport, and deposition processes, and the diverse and pervasive human alterations to sediment routing systems, landscape to watershed-scale geomorphic context is essential for developing effective salmon recovery practices. My research involves a novel approach at providing that watershed-scale geomorphic context. The concentration of TCNs in quartz, within river sediment, can be used to estimate catchment-average erosion rates that integrate over millennial time-scales, providing insight into long-term (e.g., $10^3$-$10^5$) rates of sediment supply as
well as variability of sediment yield over time (Hewawasam et al., 2003; Gellis, 2004; Ferrier et al., 2005; Fuller et al., 2009). Additionally, contrasting these millennial-scale erosion rates with paleo-erosion rates (estimated via TCN concentrations in dated terrace deposits) and/or modern sediment yield (i.e., decadal-scale, past 150 years, measured from sediment gaging or modeling) provides useful insights into the degree of perturbation (both natural and anthropogenic) within particular watersheds (Kirchner et al., 2001). Ultimately, quantifying the spatial and temporal variability of sediment yield over a range of scales will inform the aquatic ecosystem monitoring and restoration efforts throughout the CRB (Figure 1.2).

1.5 Broad Research Questions

This project will begin to inform our understanding of three broad research questions:

1. How do long-term rates of sediment supply vary spatially and temporally throughout the Columbia River Basin (Figure 1.3)?
2. How have human activities influenced (amplified or dampened) processes of erosion and sediment transport?
3. At what scales do long-term and near-term erosion rates influence aquatic habitat metrics
1.6 Specific Objectives

Specifically, this project consists of seven related objectives that form the following thesis chapters:

Chapter 2: SPATIAL AND TEMPORAL VARIABILITY OF SEDIMENT YIELD IN THE CRB: WHAT THE LANDSCAPE CAN TELL US ABOUT LONG-TERM EROSION RATES
1. Compile a database of CAER from TCNs and other sources (sediment yield gage data and modeling outputs) throughout the Columbia River Basin (Kirchner et al., 2001; Densmore et al., 2009; Harris et al., 2010; Moon et al., 2011; Portenga and Bierman, 2011).

2. Use OSL and C14 dating in combination with TCN concentrations in fluvial terraces and modern channel alluvium to observe how CAERs vary spatially and temporally within four sub-watersheds in the CRB (John Day watershed (JDW), Oregon, Pahsimeroi watershed (PRW), Idaho, Entiat (ERW) and Wenatchee watersheds (WRW), Washington) and one watershed directly adjacent to the CRB (Skykomish watershed (SRW)) to document spatial and temporal variability in long-term erosion rates and sediment supply throughout the CRB.

3. Use hypsometry, longitudinal profile analysis, channel steepness, slope, relief, drainage area, bedrock geology and flow duration curve analysis to attempt to explain the variability observed in the CAER and sediment yield from the same small catchments from point two.

**Chapter 3: SPATIAL AND TEMPORAL VARIABILITY OF SEDIMENT YIELD IN THE CRB: CASE STUDIES FROM SPECIFIC STUDY SUB-WATERSHEDS**

4. Use OSL and C14 dating in combination with TCN concentrations in fluvial terraces and modern channel alluvium to observe how CAERs vary spatially and temporally (Schaller et al., 2002; Schaller et al., 2004; Fuller et al., 2009) within four sub-basins in the John Day watershed and two sub-basins within the Wenatchee watershed and the Skykomish watershed. This will illustrate if and how rates have changed and may be used to determine the degree of recent perturbation or disturbance within these basins.
5. Use a combination of rapid geomorphic assessments, geomorphic mapping, grain-size analysis and longitudinal analysis coupled with CAER to unravel the processes of watershed-wide sediment dynamics within the same watersheds outlined above.

Chapter 4: LINKING LONG-TERM SEDIMENT YIELD TO AQUATIC HABITAT IN THE COLUMBIA RIVER BASIN

6. Conduct rapid geomorphic assessments based on the Fluvial Audit (Newson and Sear, 1997; Newson et al., 1998) and River Styles frameworks (Brierley and Fryirs, 2005; Brierley et al., 2010) within six sub-basins of the John Day Watershed. These assessments informed the TCN, OSL and C14 sampling strategy as well as our understanding of the linkages between long-term sediment supply and modern channel morphology.

7. Relate spatial variability in millennial-scale, catchment-averaged erosion rates with the type and distribution of channel units used by NOAA’s Columbia Habitat Monitoring Program (CHaMP).

1.7 Physiographic Setting

General Overview of the Columbia River Watershed

The Columbia River watershed is large and diverse; topographically, geologically and climatically (Figure 1.3). It covers over 673,000 km² and includes parts of seven states and British Columbia. The physiography of the watershed is extremely diverse, ranging from the semi-arid crest of the Rocky Mountains to the moist maritime climate of the western side of the Cascade Range. Elevations range from sea level to 4392 m at Mt. Rainier. Steep mountains and low valleys comprise a large portion of the Basin with mountain ranges such as the Bitterroot, Selkirk, Steens, Cabinet, Salmon River, Lemhi,
and Purcell ranges, averaging over 1,500 m (Mckee, 1972). Many of these ranges have been sculpted by both continental and mountain glaciations. In addition to these mountains and valleys, a large portion of the watershed, called the Columbia River Plateau is composed of high elevation desert (e.g., Channeled Scablands of eastern Washington, Snake River Plain of southern Idaho, Owyhee Desert in southwestern Idaho and southeastern Oregon and the Oregon High Desert of eastern Oregon) with average elevations over 1200 m (Figure 1.3).

1.8 Geology

The bedrock geology of the CRB is diverse; composed of metamorphosed schists and gneiss, marine sedimentary rocks, granitic batholiths, bedded sandstone, basalt, and belt series metasediments (Baldwin, 1959; Mckee, 1972) (Figure 1.4). Much of the watershed’s landscape has been sculpted by volcanism. Pyroclastic and lahar flows, flood basalts and ash deposits cover over 163,700 km², forming the vast Columbia River Plateau of eastern Washington, southwestern Idaho and northern Oregon. The plateau is composed of the Columbia River Basalt Group which is actually four separate flood basalt formations, the Imnaha Basalt (17.5 Ma), followed by the Grande Ronde Basalt (16.5 to 15.6 Ma), the Wanapum Basalt (15.6 to 14.5 Ma) and finally the Saddle Mountains Basalt (14.5 to 6 Ma). In the Pasco Basin (Figure 1.3) of eastern Washington, on the basis of geophysical evidence (e.g., gravity and magnetic surveys, seismic refraction profiles, thermal gradient measurements and oil test wells), basalt deposits are thought to reach a maximum thickness of 4870 m and are primarily responsible for the landscape of rolling hills, plains, prairies and deserts (Quigley, 2000).
Figure 1.3: A portion of the CRB upstream of the Cascade Mountains. Study sub-watersheds are shown in black and prominent landforms are labeled.

Additionally these flood basalts are responsible for forcing the ancient Columbia River into its present course. The Snake River Plain is a unique geologic feature within the CRB composing most of southern Idaho. It stretches 640 km westward from the Wyoming border to the Idaho-Oregon border. The eastern Snake River Plain traces the path of the North American plate as it migrated across the Yellowstone hotspot. This
region is underlain almost exclusively by basalt generated from large shield volcanoes. The rest of the Snake River Plain sits in a large rift valley filled with interbedded layers of basalt and fluvial/lacustrine sediments between two to three kilometers thick (Wood and Clemens, 2002). In more recent times, (i.e., last 4,000 years) volcanoes of the Cascade Range erupted on average twice per century (Quigley, 2000) adding to the dynamic history of volcanism that has shaped the region.

Figure 1.4: A simplified representation of the bedrock geology of the CRB. Geologic layers were collected from state-wide digital GIS portals and categorized into the categories shown above. For more information on the sources of data and the classification scheme see Appendix C.
As the TCN-derived CAER as well as the OSL dating technique both utilize quartz as the target mineral, the distribution of quartz-bearing rock types within the CRB is of interest. Quartz distribution is variable throughout the CRB (Figure 1.5) but generally is abundant within the Rocky Mountains and the Cascade Range as well as areas with plutonic intrusions (i.e., sections of the John Day watershed). The most common quartz bearing rock types in these mountainous regions are: granite, granodiorite, gneiss, quartzite, rhyolite and dacite. The portions of the watershed that experienced the greatest amount of volcanism (i.e., the Snake River Plain and the Columbia River Plateau) are primarily absent of quartz-bearing rock types (Figure 1.5) except where felsic plutons have intruded.

1.9 Surficial Geology

In addition to volcanism, Pleistocene glaciations have left their imprint on the CRB landscape. During the Pleistocene, the Cordilleran ice sheet advanced south into Idaho forming a large ice dam on the Clark Fork River at the border of present-day Montana. The massive lake that formed, Glacial Lake Missoula (Figure 1.3), contained roughly the same volume as present-day Lake Michigan and was over 600 m deep at the dam. Repeated breaching of Glacial Lake Missoula from 12,700 to 15,300 years ago caused massive amounts of scour, forming the landscape known as the Channelized Scablands in eastern Washington (Dietrich, 1995; Jackson, 2005). The Scablands are composed of erosional features such as: coulees, buttes, mesas, dry water falls, hanging valleys, and enormous ripples (decimeters to tens of meters of relief) (Bretz, 1969). The magnitude of the Missoula floods, which were thought to have occurred at least 40 times throughout the Pleistocene, had discharges that were ten times the combined flow of all
Figure 1.5: Quartz bearing rock types occurring in the CRB are shown in blue. Also shown (in black) are the locations of sample sub-watersheds. Quartz content was inferred from mapped rock types. For more information on the sources of data and the classification scheme see Appendix C.

The present day rivers of the world combined (i.e., 10 – 20 X 10^6 cms) (Denlinger and O’Connell, 2010). Glacial outwash and deposits from glacial lake flood events including Lake Missoula were eroded by the wind and sequentially redeposited as thick layers of loess (>60 m) over what is now the Columbia Valley, Columbia Plateau and the Snake River Plain. These deposits create highly productive, moisture retaining soils that are ideal for agriculture, making the Columbia River Plateau one of the major grain producing regions of the world. Also during the Pleistocene, a number of large lakes
developed throughout the Great Basin region (Grayson, 1993), these include late Pleistocene pluvial Lake Bonneville and Lake Lahontan (Beck and Jones, 1997).

1.10 Climate

Climate also varies considerably throughout the large and diverse Columbia River Watershed, but is generally influenced by three distinct air masses; moist marine air from the Pacific Ocean, dry continental air from the east and south, and dry arctic air from the north. Elevation and the spatial relation to the rain shadows caused by the dominant North-South running mountain ranges greatly affects the climate in localized areas (Pfister et al., 1977; Finklin et al., 1987; Franklin and Dyrness, 1988; Jain et al., 2008). For example, the Cascade Range experiences a pronounced orographic effect, which is especially prominent in the summer. Generally, higher elevation areas experience cold winters (mean January temperatures of -10 to -5 °C) and cool, short summers (mean July temperatures of 8 to 13 °C) with intermountain regions experiencing extreme temperature variability between summer and winter (-4 to 16 °C) and intermittent droughts. Some areas experience maximum precipitation in the winter (e.g., west of the Cascade Mountains) with Pacific storm-driven climactic patterns, while other eastern regions like the shrub-steppe ecoregion tend to have continental climate patterns with maximum precipitation occurring in early summer. Precipitation patterns are varied across the watershed with over 400 cm per year on the western side of the Cascade Range and under 20 cm in portions of the interior.
Figure 1.6: Spatial trends in precipitation throughout the CRB. Note the pronounced orographic effect around mountain ranges. Data was collected from the PRISM Climate Group and represents the past 30 year annual averages. Cell resolution is 800m. See Appendix B for more details.

1.11 Columbia River Channel Network

Originating in Lake Columbia in the Rocky Mountains of Canada, the Columbia River flows circuitously through the CRB starting northwest then south before turning west again to form most of the border between Oregon and Washington on its 2,000 km route to the Pacific Ocean. The Columbia River experiences an elevation drop of over 750 m giving it four times the relief of the Mississippi River in roughly half the distance. Additionally, the Columbia has the fourth largest amount of discharge of any North American river. For these reasons the Columbia has been the biggest hydroelectric
Figure 1.7: Mean maximum summer temperatures (recorded in August) throughout the CRB. Data was collected from the PRISM Climate Group and represents the past 30 year annual maximum averages. Cell resolution is 800m. See Appendix B for more details.

energy producer of any North American river. Dams and the resulting upstream reservoirs exert a strong influence on the longitudinal profile of the Columbia River as evidenced by the stair-step profile of the main-stem Columbia and the Snake River (Figure 1.9). Water diversions and dam construction began as early as 1900, lasting until 1984 with the completion of the Revelstoke Dam in British Columbia (Committee on Water Resources Management et al., 2004). There are currently 14 major hydroelectric dams on the main-stem and 47 total within the CRB including the main-stem dams, not
Figure 1.8: Mean minimum winter temperatures (recorded in February) throughout the CRB. Data was collected from the PRISM Climate Group and represents the past 30 year annual maximum averages. Cell resolution is 800m. See Appendix B for more details.

including many other run-of-the-river diversions and small dams within sub-basins to the Columbia. In addition to dams and reservoirs, bedrock canyons exert a strong influence on the gradient in the CRB (Figure 1.9). In many places, the location of dams and bedrock canyons coincide as dams were built in the narrowest sections of the channel network.
Figure 1.9: Longitudinal profile of the Columbia River (black), major tributaries (Snake: dark blue, Salmon: light blue) and study sub-basins (Pahsimeroi: cyan, Wenatchee: red, Entiat: pink, NF John Day: dark green, John Day: light green). Major canyons and dams are also shown with dams alphabetically labeled (A: Mica Dam, B: Revelstock Dam, C: Grand Coulee Dam, D: Milner Dam, E: Brownlee Dam, F: John Day Dam). DEM cell resolution: 100 m, smoothing window: 1000 m. Dams exert a strong control on gradient as evidenced by the stair-step profile of the Columbia and Snake rivers as well as changes in rock type resulting in canyons and gorges. Also note the relatively greater steepness of the Entiat and Wenatchee profiles compared to the John Day and Pahsimeroi study sub-watersheds.

1.12 Human Perturbations

Historic land use within the portion of the CRB that falls within the political boundaries of the United States varies regionally depending on the timing and intensity of Euro-American settlement, but general trends can be observed post-settlement. Native Americans have inhabited the CRB for at least 15,000 years (National Research Council, 2004) while the era of large-scale Euro-American settlement occurred from 1810 to the 1930’s. Mining, livestock and agriculture followed from the 1850s to 1910 with large-scale and intensive timber harvesting occurring from 1920, peaking in the 1990’s, and continuing to this day (National Research Council, 2004). The history of timber extraction has profoundly influenced the sediment dynamics within the CRB, impacting
run-off and groundwater fluxes and temporarily increasing fine sediment supply to parts of the channel network (National Research Council, 2004). Coupled with timber extraction, the legacy of in-stream woody debris removal for navigation and flood control over the last ~150 years has had profound impacts on sediment routing and channel morphology. Currently, commercial timberland (public and private) is the dominant land use in the region (35.6 % of the US land base), followed by land for grazing (35.2 %), croplands (15.9 %), non-commercial timberlands (10.2 %), developed lands (urban and transportation, 1.8 %), and National Parks (1.3 %) (Jackson and Kimerling, 1993). The combined totals for commercial timberland and grazing (70.8 %) make these the eminent land uses in the region by area (Jackson and Kimerling, 1993) (Figure 1.10).
Figure 1.10: Land cover and land use in the CRB. Data was collected from the United States Geologic Survey and is based on information from the 1970-1980’s.
CHAPTER 2

SPATIAL AND TEMPORAL VARIABILITY OF SEDIMENT YIELD IN THE COLUMBIA RIVER BASIN: WHAT THE LANDSCAPE CAN TELL US ABOUT LONG-TERM EROSION RATES

2.1 Introduction

Variability of erosion, transport, and storage within a watershed can occur across a wide range of spatial and temporal scales (Walling and Webb, 1996). This variability can be driven by a complex combination of ‘natural’ conditions (e.g., tectonic setting, rock type, climate, vegetation) and disturbances (e.g., wildfire, floods) that may be exacerbated or diminished by anthropogenic processes (e.g., agriculture, forest clearing, road building). The relative importance of natural and anthropogenic processes that give rise to the variability of any given system is often difficult to determine (Walling, 2006). Nevertheless, quantifying the rates and variability of sediment yield at a variety of scales can provide important geomorphic context for watershed and aquatic ecosystem management (Montgomery et al., 2000; Belmont et al., 2011).

Quantifying long-term (i.e., millennial-scale) sediment yield by measuring the concentration of Berillium-10 ($^{10}$Be) in river-borne sand can help establish the ‘natural’ or reference conditions, in terms of sediment yield, that aquatic ecosystems have experienced. It should be noted that such $^{10}$Be derived catchment-averaged-erosion-rates (CAER) do not allow for the quantification of individual processes of erosion or sediment routing (e.g., rill erosion, individual debris flows or landslides, etc.) but instead identify broad trends of variability (Portenga and Bierman, 2011). CAER are also sensitive to
recent, very large changes in catchment erosion rates (e.g., removal of 30 cm of material). Long-term erosion rates can identify erosional hotspots within a landscape, which can help guide reach-scale salmonid restoration and habitat monitoring efforts. Further, these rates can be interpreted relative to modern sediment gaging techniques (e.g., dam infilling rates, in-stream sediment transport measurements, etc.) and/or the CAER measured in dated terrace deposits to potentially quantify the degree of recent perturbation or disturbance within a watershed, which could provide useful context for restoration actions (Brown et al., 1998; Hewawasam et al., 2003).

Many conceptual and empirical models have been developed to predict long-term erosion and/or sediment yield from simple topographic and environmental parameters like mean watershed slope and mean annual precipitation. Land managers and restoration practitioners would greatly benefit from the development of such simple predictive relationships that could provide critical information for developing sediment budgets, watershed conservation and restoration plans based on readily accessible information (e.g., DEMs, precipitation data, etc.). But these predictions are only valuable if accurate. Such predictions are challenging because of the spatially variable and stochastic nature of sediment yield coupled with extremely limited and incomplete empirical data for model development. Furthermore, estimates of long-term sediment yield derived from the concentration of terrestrial cosmogenic nuclides (TCNs) in river sediment has shown that simple relationships are often incomplete and inadequate (Riebe and Kirchner, 2001a; Riebe and Kirchner, 2001b; Von Blanckenburg, 2006). Instead, previous work with TCNs has shown a stronger relationship between rates of tectonic uplift and long-term erosion (Binnie et al., 2007; Siame et al., 2011), but that relationship
can be confounded by a number of other factors. For example, Chapter three provides evidence that long-term rates of sediment yield in the Cascade Range of northern Washington are still influenced by the recent (11-17 ka) history of glaciation and may also be influenced by the strong precipitation gradient created by the Cascade Range (Moon et al., 2011). Additionally, changes in base level from eustatic processes can influence long–term sediment yield (Merritts et al., 1994).

In this chapter I compare rock type, slope, relief, hypsometry and mean annual precipitation to long-term rates of sediment yield measured with $^{10}$Be in river sediment within four sub-watersheds in the CRB (John Day watershed (JDW), Oregon, Pahsimeroi watershed (PRW), Idaho, Entiat (ERW) and Wenatchee watersheds (WRW), Washington) and one watershed directly adjacent to the CRB (Skykomish watershed (SRW)). We use these data to assess if and where functional relationships between sediment yield and readily available topographic, climatic and geologic metrics exist. Additionally, I provide examples from select study sub-watersheds where I evaluate the legacy effects of natural processes on long-term sediment yield.

### 2.2 Background on Catchment Averaged Erosion Rates (CAERs)

For decades geologists and geomorphologists have used a variety of tools to measure the rates of sediment generation and transport through a drainage network at a variety of spatial and temporal scales (Burt and Allison, 2010). Conventional methods have relied on measurements from suspended and bedload sediment gages and sediment accumulation rates in reservoirs or are based on empirical sediment yield models (e.g., Soil and Water Assessment Tool (SWAT) and Water Erosion Prediction Project (WEPP)) that relate simple topographic and climatic parameters to long-term sediment yield via the
Universal Soil Loss Equation (USLE). Recent Terrestrial Cosmogenic Nuclide (TCN) studies (Kirchner et al., 2001; Hewawasam et al., 2003; Ferrier et al., 2005) have shown that direct short-term (i.e., decadal-scale) measurements of sediment flux via stream gages or sediment in-filling rates in reservoirs is often not representative of long-term (i.e., millennial-scale) averages and further, that basin morphological characteristics (e.g., watershed slope) are not good predictors of long-term erosion rates in many places (Von Blanckenburg, 2006).

Cosmogenic nuclide theory and practice has been evolving rapidly over the last 25 years and is now widely applied in the Earth Surface Sciences (Gosse and Phillips, 2001; Von Blanckenburg, 2006; Granger et al., 2013). Terrestrial cosmogenic nuclides are produced within mineral grains at and near Earth’s surface by the interaction of high-energy sub-atomic particles (i.e., cosmic rays) with target atoms within minerals (Lal and Peters, 1967). Specifically, our research focuses on the TCN $^{10}\text{Be}$, which is produced within quartz. Quartz is widely used in TCN studies because it is ubiquitous across much of Earth’s surface, is resistant to weathering, and has a simple internal geometry (Gosse and Phillips, 2001).

Rates of $^{10}\text{Be}$ production are well constrained throughout the world (Granger et al., 2013). The cosmic ray flux attenuates rapidly with depth into the Earth’s surface, thus TCN production is highest within the first meter and drops to nearly zero after approximately 1.5-2 meters depth (Bierman and Nichols, 2004). The accumulated TCN concentration in a mineral grain (e.g., quartz) records the rate at which that grain has been unearthed, which coincides with the time it took for the grain to pass through the approximately two meter window of the soil profile wherein the grain was exposed to
incoming cosmic rays (Granger and Kirchner, 1996). Because most erosional processes take place within the first few meters of Earth’s surface, TCN studies are well suited to constrain the rates of these processes.

Rivers are the great integrators of the landscape, transporting sediment eroded from hillslopes throughout the channel network ultimately to the river’s mouth. A sample of sand collected from any given location in a river is composed of hundreds of thousands of grains eroded at different rates from all different parts of the upstream contributing area. Thorough mixing during hillslope and fluvial transport homogenizes TCN concentrations in the bulk sample collected at the watershed outlet (Von Blanckenburg, 2006). Assuming that the upstream areas contribute sediment in proportion to their long-term erosion rate, spatially-averaged erosion rates can be calculated from the concentration of TCNs in the sample using equation 2.1 (Brown et al., 1988; Granger and Kirchner, 1996; Kirchner et al., 2001; Schaller et al., 2001; Von Blanckenburg, 2006)

\[
N = \frac{P_0 \Lambda}{E}
\]  

(2.1)

where \(N\) is the concentration of \(^{10}\text{Be}\), \(P_0\) is the production rate of \(^{10}\text{Be}\) (corrected for watershed-wide variability in latitude/elevation and adjusted for shielding), \(\Lambda\) is the attenuation length of cosmic ray penetration and \(E\) is erosion rate.

The time period over which CAERs integrate depends on the time it takes to erode the ‘exposure window’ within which most of the TCN inventory is accumulated
(~60 cm). Thus in rapidly eroding landscapes such as the Himalayas the integration time may be less than a thousand years (Vance et al., 2003) while in slowly eroding landscapes such as portions of Antarctica or Namibia the integration time may be as long as $10^5$ to $10^6$ years (Bierman and Caffee, 2001). TCNs have been used for a wide range of applications, including: understanding sediment transport pathways (Brown et al., 1998), establishing the rates and styles of local and large-scale erosion, soil development, and landscape evolution (Schaller et al., 2001; Bierman and Nichols, 2004; Guralnik et al., 2011), comparing the relative importance of climate and tectonics in chemical weathering processes (Riebe and Kirchner, 2001a; Riebe and Kirchner, 2001b).

The following several studies have compared erosion rates over decadal, millennial, and in one case, million year time-scales using a combination of TCN derived CAER, sediment yield gage data and/or apatite (U-Th)/He analysis.

- Kirchner et al. (2001) compared erosion rates over decadal, millennial and million year timescales within 32 mountainous catchments of central Idaho and found the millennial and million year rates were roughly equal and on average 17 times higher than the rates collected with decadal gage data. This data suggests that sediment delivery from mountainous terrain is typically low, punctuated by episodic delivery of high magnitude debris flow type events following intense forest fires. Over long time periods aquatic ecosystems are occasionally subjected to intense sediment disturbance.

- Ferrier et al. (2005), in two catchments in the Coast Range of Northern California, compared decadal and millennial-scale erosion rates and found the millennial
rates roughly in agreement with or slightly higher than decadal rates inferred from >30 years of sediment-yield gage data. They did not have a definitive interpretation of these findings but suggested that intensive logging in the late 19th and early 20th century may have stripped a significant portion of soil from the study catchments artificially elevating the millennial-scale erosion rates. Their results also suggest greater variability in decadal-scale erosion rates over millennial timescales, lending further support to the finding of Kirchner et al. (2001) regarding the episodic nature of sediment delivery in small mountainous catchments.

- In contrast, Hewawasam et al. (2003), in a deforested tropical highland catchment of Sri Lanka, found modern sediment yield from suspended sediment data was 10-100 times greater than those inferred from millennial-scale TCN-derived erosion rates. Recently elevated rates are attributed to large-scale deforestation and intensive agriculture.

- Gellis et al., (2004), in the arid Arroyo Chavez catchment of New Mexico, derived sediment yield data from colluvial slopes, gently sloping hillslopes, the mesa top and the alluvial valley using sediment traps and dams and compared that with millennial-scale erosion rates derived from TCNs (10Be). They found modern rates of sediment production measured from modern gaging techniques were similar to millennial-scale rates except on the alluvial valley floor where modern rates were >10 times higher than the background rates. The 10x increase in
sediment production was attributed to the significantly higher level of human influence (e.g., grazing and gas-pipeline activity) within this region of the catchment.

- Densmore et al. (2009) used TCN-derived catchment averaged erosion rates in ten small watersheds in the Sweetwater Range in south-western Montana to test the hypothesis that millennial- scale erosion rates would spatially vary with rock uplift rates along an active normal fault, with maximum erosion associated with the fastest rates of uplift. He found that erosion rates did vary along strike, depending on the fault length and displacement but that pre-faulting topography had decoupled millennial-scale erosion rates from fault displacement in some locations along the fault.

- Moon et al. (2011) used a combination of millennial-scale erosion rates derived from TCNs and million year scale erosion rates from apatite (U-Th)/He analysis to assess the role of climate and past glaciations on erosion in the Washington Cascades. They found rates averaging over millennial time-scales were four-fold higher than rates integrated over million-year timescales. Additionally, the millennial-scale erosion rates increase linearly with the modern precipitation gradient, an observation that stands in contrast to the more general trend that long-term erosion rates are not correlated to precipitation because of negative feedbacks provided by vegetation. Thus, the difference in erosion rates integrated over different timescales, combined with the correlation between precipitation and
millennial-scale erosion is attributed to a mis-match between landscape form and modern erosional process. Specifically, recent glaciations sculpted a landscape characterized by over-steepened topography relative to the modern mechanisms of erosion, which are primarily fluvial processes.

2.3 Methods: Geochemical Techniques: Cosmogenic Nuclides, Optically Stimulated Luminescence (OSL), and Carbon-14 (C14) Dating

Collection and processing of stream alluvium and terrace deposits for $^{10}$Be concentration

Samples of river-borne sand (size fraction 250-600um) were collected near the outlets, and in some locations, along the longitudinal gradient of the main channel of 19 study sub-watersheds. Modern alluvium was collected either from small point bars or other deposits within the active channel. Terrace sand deposits were sampled primarily from terraces immediately adjacent to the modern stream that were identified as active sediment sources. Sampled terrace deposits were dated in the JDW with either OSL or C14. Sampled terraces in the Wenatchee and Entiat watersheds were determined to be late Pleistocene/early Holocene aged glacial outwash terrace deposits based on stratigraphic interpretation and previous research (Lorang and Aggett, 2005; Godaire et al., 2010)

Samples were processed to isolate quartz by a combination of physical and chemical methods. The target grain size was extracted by either wet-sieving in the field or dry sieving in the lab. Depending on the amount of quartz present, approximately 10-15 kg of sand was collected per sample. Approximately 75 g quartz was purified by a combination of physical separation (e.g., magnetic and with heavy liquids), aqua regia (HCl/HNO$_3$), dilute (5%) hydrofluoric (HF), and nitric (HNO$_3$) acid treatments on heated
agitators (Kohl and Nishiizumi, 1992). Each acid treatment typically lasted twelve hours. The HF treatment removes any meteoric $^{10}\text{Be}$ that adheres to the outside of quartz grains while also removing many unwanted minerals (non-quartz). After physical separation, aqua regia, and the first series of one to two HF treatments, samples were sent to the Frankel Cosmogenic Geochronology Laboratory (Georgia Institute of Technology) for final processing. Samples from relatively quartz-rich lithologies (e.g., Wenatchee and Entiat) required one to two additional HF/HNO$_3$ treatments while samples from the JDW, which is a volcanic, generally quartz-poor landscape, required a more involved purification process to obtain the minimum amount of quartz for Accelerator Mass Spectrometry (AMS) analysis. This consisted of three to four additional HF/HNO$_3$ treatments and physical separation using a heavy liquid solution with a density of (2.85 g cm$^{-3}$).

Following quartz purification, a known quantity of $^9\text{Be}$ carrier was added to each sample and process blank to act as a benchmark during AMS analysis. The quartz and carrier were then dissolved using concentrated HF, dried for 24 hours, and subjected to a series of perchloric acid fuming treatments to remove fluorides. Ion exchange chromatography was then used to remove Fe, Al, Ti, and other metals. The Be is subsequently dried and dissolved in a 50% HCl solution, precipitated out using ammonium hydroxide (NH$_4$OH), and heated to form Beryllium-oxide (BeO). BeO was then mixed with Nb before being packed into targets, which were sent to Purdue University Rare Isotope Measurement Laboratory (PRIME) for AMS analysis.
TCN production rate and denudation rate calculations

Globally, cosmic ray attenuation increases as atmospheric pressure increases (Stone, 2000). As a result, TCN production rates (e.g., $^{10}\text{Be}$ production rates in quartz) vary based on latitude and altitude. Scaling factors were developed (Lal, 1991; Stone, 2000; Dunai, 2001; Dunai, 2010) to transfer a site-specific production rate derived from calibration sites to any location where TCNs are applied. Here, we scaled catchment-averaged production rates from our study watersheds using a sea-level, high latitude production rate of 4.5 atoms g$^{-1}$ yr$^{-1}$, recommended by PRIME Laboratory’s revised $^{10}\text{Be}/^{9}\text{Be}$ standard and Stone (2000).

In addition to atmospheric shielding, cosmic ray attenuation is also affected by shielding due to topography and snow cover (Gosse and Phillips, 2001). To account for this affect, topographic shielding was calculated for each 90 X 90 m DEM cell following methods developed by Balco and Stone (2008) and Nishiizumi (1989) and using equation 2.2:

$$S_t = 1 - \int_0^{2\pi} \frac{\sin(h(\theta))^{3.5}}{2\pi} d\theta$$  \hspace{1cm} (2.2)$$

where $S_t$ is the topographic shielding factor, $h(\theta)$ is the ‘horizon’ in the azimuthal direction, $\theta$. A mean topographic shielding factor was generated within each study sub-watershed from all pixels upstream of the sample locations. For a detailed description of the computational methods used see Appendix A. We conducted a sensitivity analysis comparing the catchment-averaged topographic shielding value based on DEM cell size.
resolutions of 30 m, 60 m, and 90 m. We found cell resolution was negligible in terms of shielding but substantial in terms of computational time.

**Snow Shielding, John Day Watershed, Oregon**

Gosse and Philips (2001) found that, depending on snow density (0.1 – 0.3 g/cm³), snow depths less than 50 cm (i.e., 5-15 cm Snow Water Equivalent, SWE) for four continuous months did not produce a noticeable attenuation of cosmic rays and did not warrant the inclusion of a snow shielding factor. It should be noted, that millennial-scale erosion rates average over a much longer time period than any directly measured precipitation gage. The conditions of the snow pack measured over the last ~ 30 years are not necessarily representative of conditions for the last 10 kyr. Despite this, the following methods described below have been used by others in similar mountainous regions (Gosse and Phillips, 2001; Moon et al., 2011) to account for the cosmic ray attenuation due to snow cover. A reliable proxy-based paleoclimatic record was unavailable in almost all of the study sub-watershed.

To determine if a snow shielding factor was required in our analyses of quartz samples from the JDW, I consulted 14 snow telemetry (SNOTEL) sites located within or immediately adjacent to the basin. SNOTEL data was solely available as monthly Snow Water Equivalency (SWE) depths averaged over the winter months of a 30 year period from 1981 to 2010. SWE values were converted to snow depths using equation 2.3 developed by the Natural Resources Conservation Service (NRCS):

\[
\frac{SWE}{p_{\text{Snow}}} = \text{Snow Depth}
\]  

(2.3)
where $p_{Snow}$ is the maximum snow density. Here, I assumed a $p_{Snow}$ of 0.3 g/cm$^3$ (Moon et al., 2011). We found that 13 of the 14 sites did not meet the criterion of a SWE $\geq 15$ cm for a minimum of four continuous months. The one site that met the criteria was not located within any of my study watersheds.

**Snow Shielding, Wenatchee and Skykomish Watersheds, Washington**

The effect of snow shielding was considered in the Wenatchee and Skykomish watersheds as SWE $\geq 15$ cm for four continuous months within the higher elevation portions of these watersheds. Values were obtained from Moon et al. (2011) as our sample locations were located in close proximity. For a complete description of the method used to calculate the snow shielding factors see Moon et al. (2011) supplementary information (DOI: 10.1038/NGEO1159) but generally it was the same method I employed within the Pahsimeroi watershed.

**Snow Shielding, Pahsimeroi Watersheds, Idaho**

A catchment-averaged snow shielding factor was calculated for the three study sub-watersheds within the Pahsimeroi watershed, as snow depths are considerable. First, we developed a linear regression predicting SWE as a function of elevation based on four USGS precipitation gages located within or immediately adjacent to the Pahsimeroi watershed with a record extending from 1981 to 2012. Second, we implemented two software tools developed by Portenga and Bierman (2011) that use hypsometry to effectively reduce the distribution of elevations within a watershed to a single point. Within the study sub-watersheds drainage area ranges from 45 – 170 km$^2$ and elevations range from 1900 -3000 m. The hypsometrically derived elevation point, for each sub-
watershed, was used to predict SWE for every month of the year. SWE depth was converted to snow cover depth as explained above with an assumed snow density of 0.3 g/cm$^3$. The catchment-averaged snow shielding factor was then calculated in CosmoCalc (Vermeesch, 2007) using equation 2.4:

$$S_c = \frac{1}{n} \sum_{i=1}^{n} e^{\frac{\rho(i)z(i)}{\Lambda_0}}$$  \hspace{1cm} (2.4)

where $n$ is an individual month, $z$ is snow thickness, $\rho$ is snow density and $\Lambda$ is equal to the attenuation coefficient for spallation reactions.

**Calculating Catchment-Averaged Erosion Rates**

Ultimately the catchment-averaged scaling and shielding factors were combined with the concentration of $^{10}$Be from the samples as inputs into the Cosmic-Ray Produced Nuclide Systematics on Earth Project (CRONUS-Earth Project version 2.2) $^{10}$Be – $^{26}$Al calculator to derive catchment-averaged erosion rates for each study watershed. CRONUS uses the time-dependent corrections of Lal (1991) and Stone (2000) following the method of Balco et al. (2008).

**Calculating Sediment Yield from CAER**

Catchment-averaged erosion rates can be converted to sediment yield using equation 2.5.

$$\text{Sed. Yield} = \left(\frac{1}{1000}\right) \varepsilon \ast DA \ast \rho \left(1 - DL\right)$$  \hspace{1cm} (2.5)
where \( \text{Sed. Yield} \) is expressed in Mg/yr, \( e \) is equal to the millennial-scale erosion rate (mm/yr), \( DA \) is drainage area (m\(^2\)), \( \rho \) is equal to rock density (2.5 g/cm\(^3\)) and \( DL \) represents the dissolved load. Anderson and Anderson (2010) report a figure of ~40% of the total sediment yield from the CRB is contributed as dissolved load. This percentage is a compilation from two other sources (Maybeck, 1976; Milliman and Meade, 1983) that are in agreement with this approximate value. Clearly the CRB is a large and diverse watershed in terms of geology and climate and as such the amount of dissolved load should spatially vary. I chose to account for this variability by using a model developed by Olson and Hawkins (2012) that predicts the natural base-flow stream water chemistry within the western United States based on geologic and environmental (e.g., climate, atmospheric deposition, soils, vegetation, topography) parameters. Additionally, they generated continuous maps of percentages of CaO, MgO, S, compressive strength, and hydraulic conductivity based on the chemical and physical properties of bedrock that are incorporated into the model. For a complete list of model parameters see Olson and Hawkins (2012). The model output is expressed as electrical conductivity (EC) which we used as a proxy for the total dissolved load. We developed a simple classification system that linked the range of model-predicted EC values generated from my study sub-watersheds to a reasonable proportion of dissolved load derived from literature values (Anderson and Anderson, 2010) and expert opinion (personal communication with Dr John Olson). Table 2.1 shows this classification system with the model generated values for EC and dissolved load used for each study sub-
watershed. Sensitivity analysis indicated that this value should be applied at the scale of the entire study sub-watershed (e.g., JDW, WRW, SRW, etc).

Table 2.1 The top three columns show typical values for electrical conductivity (EC) and dissolved load (DL). Bottom three columns show model predicted EC and DL values for each study sub-watershed

<table>
<thead>
<tr>
<th>Categories</th>
<th>Very Low</th>
<th>Low</th>
<th>Average</th>
<th>High</th>
<th>Very High</th>
</tr>
</thead>
<tbody>
<tr>
<td>Typical EC Values</td>
<td>&lt; 50</td>
<td>&lt; 100</td>
<td>250</td>
<td>300</td>
<td>&gt; 400</td>
</tr>
<tr>
<td>Typical Dissolved Load</td>
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<td>5 %</td>
<td>10 %</td>
<td>20 %</td>
<td>40 %</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Watershed</th>
<th>John Day</th>
<th>Pahsimeroi</th>
<th>Entiat</th>
<th>Wenatchee</th>
<th>Skykomish</th>
</tr>
</thead>
<tbody>
<tr>
<td>Predicted EC</td>
<td>166</td>
<td>67</td>
<td>45</td>
<td>46</td>
<td>45</td>
</tr>
<tr>
<td>Dissolved Load</td>
<td>7 %</td>
<td>5 %</td>
<td>4 %</td>
<td>4 %</td>
<td>4 %</td>
</tr>
</tbody>
</table>

**Optically Stimulated Luminescence Dating**

OSL dating provides an age estimate for the last time sediment was exposed to light, which resets the luminescence signal (Huntley et al., 1985). During burial the luminescence signal is reacquired due to exposure to ionizing radiation from the surrounding sediments. The intensity of luminescence measured in the lab is directly related to the length of burial and amount of radioactivity in the surrounding sediments. In its simplest form the OSL age of a sample of quartz (T) can be expressed as the equivalent dose of total ionizing radiation required to produce the natural luminescence signal (Dn) divided by the rate of irradiation from radionuclides in the surrounding sediments (Dr) and contributed from cosmic rays (equation 2.6).
\[ T = \frac{D_e}{D_r} \]  

I collected 14 OSL samples from terrace deposits, nine from the JDW in Oregon and five from the WRW and ERW in Washington. Standard sample collection methods were followed per Utah State University Luminescence Laboratory (USU-LL) instructions. Due to financial and logistic constraints, a small sub-set of these samples \((n=3)\) were chosen for analysis. Quartz was purified and the samples were analyzed in the USU-LL using the latest generation single-aliquot regenerative dose technique (Murray and Wintle, 2000).

**Carbon-14 Dating**

Carbon-14 is the most common numeric dating method and is based on the decay rate of C14 to N14 atoms. C14 atoms are cosmogenic nuclides produced in the atmosphere by the collision of cosmic rays with gases. The C14 is then incorporated into CO\(_2\) gas which in turn is incorporated into plants via photosynthesis and distributed throughout the food chain. The CO\(_2\) taken up by the plant generally contains a similar proportion of C14 as that found in the atmosphere during the life of the plant. Although there is some fractionation due to vital effects, this can be accounted for by measuring the C13/C12 ratio in the organic material. Upon death, the organisms are no longer in equilibrium with atmospheric C14 production and C14 levels decay at a known rate (i.e., half-life = 5730 yrs.). The remaining levels of C14 in the sampled material is directly related to the age of the sample, after correction for variations in C14 production rates and calibration to calendar years (Trumbore, 2000).
I collected four C14 samples of charcoal and bone from fluvial sand deposits within terraces from three sub-watersheds in the JDW. Samples were sent to Beta-Analytics for accelerator mass spectrometry (AMS) analysis. Ages were calibrated to calendar years by Beta-Analytics following the approach of Talma and Vogel (2006). Ages were used to constrain the timing of deposition so that CAER measured in the same terrace deposits can be put into context for comparison with modern alluvium samples.

2.3.2 Methods: Longitudinal profile analysis

Analysis of river longitudinal profiles is predicated on theory and empirical observation from fluvial systems worldwide that show the power-law relationship of channel slope decreasing with increasing upstream contributing drainage area (Hack, 1973; Flint, 1974; Wobus et al., 2006). This relationship can be simply expressed by equation 2.7:

\[ S = k_s A^\theta \]  

(2.7)

where \( S \) is equal to local channel slope (m/m), \( A \) is the upstream contributing drainage area (m\(^2\)), \( k_s \) represents the steepness of the channel profile and \( \theta \) is the profile concavity.

The underlying basis for the relationship is the fundamental tradeoff between the two critical drivers of sediment transport, slope and flow (for which \( A \) is a proxy). Concavity indicates the rate of decrease in channel slope in the downstream direction. Typically, \( \theta \) falls within the relatively narrow range of 0.3 to 0.7 for channels across a wide range of geologic, climatic, and tectonic environments. Given sufficient adjustment
time this relationship can be used to identify portions of the stream channel that exhibit anomalously steep or flat profiles, which likely represent discontinuities in the sediment transport system. These regions can be indicative of different geomorphic or tectonic processes that influence the rate of erosion or sediment yield within portions of the watershed. For example, a steep section could represent relatively more focused erosion or rapid sediment transport as opposed to a flat area that may function as a depositional environment. The distinct transition from a ‘normal’ slope to an anomalously steep or flat slope is called a knickpoint (Wobus et al., 2006) or knickzone depending on the spatial extent of the anomaly. Often knickpoints or knickzones identify a geomorphic transition that is reflected in an observable shift in channel morphology and represent discontinuities in sediment transport.

2.4 Broad Trends in Spatial Variability of Sediment Yield from the CRB

The CRB is an incredibly diverse landscape, exhibiting a wide range of rock types, climate regimes, and vegetation communities that have been subject to highly varied geologic and land use histories, as discussed in Chapter 1. It naturally follows that throughout the CRB exists a wide range of landscape erosion rates. Using data compiled from numerous previous studies as well as new data discussed in Chapter 3, Figure 2.1 illustrates large-scale variability of long-term erosion rates throughout the CRB. Figure 2.2 illustrates the spatial variability of long-term sediment yield (Equation 2.5) at the scale of the sub-watersheds sampled as part of this project. Figure 2.2 also includes the work of Moon et al., (2011) from the Cascade Range (e.g., Wenatchee, Entiat and Skykomish watersheds). Together, these datasets provide a wide range of environments
in which to examine whether or not this spatial variability is directly related to any simple GIS based metrics (e.g., rock type, topography, hypsometry, and climate).

2.4.1 How Do Rates of Sediment Yield Compare between the End Member Watersheds: JDW and the WRW/SRW?

Rates of sediment generation and delivery to the stream network as measured from modern alluvium and terraces (though the terraces do not strictly adhere to the steady-state erosion assumption) within WRW and SRW are significantly higher and considerably less steady in time (i.e., comparing terraces versus alluvium) than rates of sediment yield from the JDW. At the entire sub-watershed scale (i.e., JDW vs WRW and SRW) Figure 2.2 provides a clear example of spatial variability of sediment yield over the millennial-scale. The substantial difference between rates of sediment yield measured in the JDW versus the Cascades is probably due to the profound, recent glaciation within the Cascades that have left the watersheds in a state of disequilibrium in terms of sediment supply (Moon et al., 2011). In the Cascades, fluvial processes of erosion are still being influenced by the extremely high rates of sediment supply associated with past glaciations. The spatial variability of erosion observed in the WRW and SRW is probably driven by a combination of glacial and climatic drivers. The process of glaciation itself has been shown to be spatially variable because of differences in the timing, intensity and magnitude of glacial advance and retreat (Fabel et al., 2004). Additionally, the strong precipitation gradient in the Cascades could be an important driver (Moon et al., 2011). In contrast, the JDW did not experience substantial glaciation during the late Pleistocene and appears to be dominated by fluvial processes of erosion alone.
Figure 2.1 CAER derived from the concentration of $^{10}$Be in active channel alluvium and terrace alluvium (shown with black crosses in inset boxes) within the entire CRB and two adjacent watersheds (Skykomish Watershed, WA, Sweetwater Range, MT). Sub-watersheds are identified within black boxes. Kirchner (2001) samples included from central Idaho (Center Image), Densmore et al., (2009) samples included from the Sweetwater Range, MT (Inset Top Right) and Moon et al., (2011) samples are also shown from the Skykomish, Entiat and Wenatchee Watersheds (Inset Top Left).
Figure 2.2 Variability of long-term sediment yield at the scale of entire study sub-watersheds. Rates of sediment yield are derived from the catchment-averaged erosion rates based on the concentration of $^{10}$Be in active channel alluvium (black dots) and terrace alluvium (red dots). Black boxes denote the individual study sub-watershed. Error bars represent the analytical uncertainty in AMS measurement.

In addition to the different geomorphic histories there is possibly a difference in the tectonic environments between the regions (e.g., JDW versus WRW and SRW), as evidenced by fault maps, longitudinal profiles and measurements of relief.

The landscape of the JDW is dominated by basalt flows that have a low density of faults compared to the WRW and SRW. Additionally longitudinal profile analysis from the JDW revealed well-graded watersheds with no evidence of recent tectonic activity. In the WRW and SRW, longitudinal analysis primarily revealed watersheds with abundant knick points that are not in a well-graded condition. We attributed many of these knick points to the effects of glaciation but cannot rule out the possibility that there is an older uplift driven imprint in the longitudinal profiles. In the absence of recent glaciations, the
WRW and SRWs would likely still yield relatively higher rates than in the JDW.
Unfortunately this relationship cannot be isolated because of the profound influence of glaciation on the rates measured in the Cascades.

2.5 Rock Type

Mechanical and chemical weathering of bedrock into saprolite and mobile regolith (i.e., soil) provides the material that can then be physically eroded from hillslopes into the channel network (Anderson and Anderson, 2010). There is a dynamic exchange between the rate of weathering and physical erosion where physical erosion may be limited by the rate at which rock is weakened and made available for transport. Conversely, physical erosion is necessary to expose fresh mineral surfaces where chemical weathering can occur (Riebe et al., 2004). The concentration of $^{10}$Be in river sediment measures the catchment-averaged denudation rate which includes mechanical and chemical weathering as well as physical erosion. TCN analysis quantifies the rate that material moves through the top ~ 1.5 meters of Earth’s surface which includes bedrock, saprolite and mobile regolith. Because bedrock is highly variable in chemical and mineral composition, past work has shown that rock type influences the rate of soil formation (Ahnert, 1987). This could influence denudation rates and ultimately sediment yield into the channel network. Because of this I chose to examine the relationship between dominant rock type within study sub-watersheds and the long-term rate of sediment yield.

Details of the generation of broad categories of rock type (e.g. extrusive, intrusive, sedimentary and metamorphic) are outlined in Appendix C. With these categories established I calculated the percent area of each rock category within the study
sub-watersheds using the Calculate Area tool within ESRI’s Arc GIS 10.0. The dominant rock type was identified as comprising at least 30% of the total watershed area but most study sub-watersheds were dominated by one rock type that comprised more than 40% of the watershed area. Figure 2.3 illustrates the results of this analysis.

![Figure 2.3](image.png)

**Figure 2.3** Dominant rock types within study sub-watersheds and associated rates of watershed averaged long-term sediment yield. Thick black bars represent the mean and black boxes represent the standard error. Dashed bars are maximum and minimum values. Metamorphics on average have a higher sediment yield than all other rock types though sample size was low for all rock types.

Metamorphics have the highest average sediment yield with standard error taken into consideration. This is probably reflective of the tectonic setting that generated metamorphism within these rocks. Despite this, rock type appears to be a poor sole-predictor of long-term rates of sediment generation across the wide range of environments sampled. This is consistent with recent cosmogenic work in the Rwenzori Mountains of East Africa where Roller et al. (2012) found no direct correlation between
bedrock type and long-term erosion rates. Though they suggest that rock strength may be partially responsible for generally slow erosion in this tropical mountain range. This does not imply that variability in rock type does not influence the rate of chemical weathering and soil formation within the study sub-watersheds. Despite this, the role of rock type alone on driving rates of soil formation and ultimately sediment yield may be dwarfed by tectonic forcing.

2.6 Hypsometry

Watershed hypsometry shows the proportional representation of normalized elevations within a watershed. It is a means to characterize landscape morphology and was traditionally used to classify different developmental stages in landscape evolution (Strahler, 1957). Strahler identified three general phases of landscape evolution (e.g., disequilibrium, equilibrium, and monadnock stage) that correspond to the sinuosity of the hypsometric curve and the value of the hypsometric integral (HI), which is simply the area under the hypsometric curve. He suggested that the distribution of topographic forms captured by hypsometric analysis must be quantitatively related to the rates of erosional processes.

Currently hypsometry is being applied in a variety of ways. For instance, Montgomery et al. (2001) characterized different geomorphic process zones along the latitudinal gradient of the Andes based, in part, on the shape of the hypsometric curves. Brocklehurst and Whipple (2004) applied hypsometry to classify the degree of glacial erosion within watersheds that experienced different levels of glaciation. Additionally, Marstellar (2012) found a strong positive correlation between the HI and catchment-averaged erosion rates within watersheds of the Appalachian Mountains. For these
reasons, I chose to conduct a simple analysis comparing the HI from all study sub-watersheds to long-term rates of sediment yield derived from erosion rates. I calculated hypsometric curves and integrals for all study sub-watersheds using an open-source python script developed by Jerry Davis (Institute for Geographic Information Science) on 10 m DEMs. The script executes the following simplified equation (Equation 2.8) for calculating the hypsometric integral (Brocklehurst and Whipple, 2004).

\[
HI = \frac{\text{mean elevation} - \text{minimum elevation}}{\text{maximum elevation} - \text{minimum elevation}}
\]  

(2.8)

Results of this simple analysis showed no discernible relationship between the HI and long-term sediment yield when comparing across all study sub-watersheds (Figure 2.4, Top), but when we consider the level of recent glaciation within study sub-watersheds an interesting relationship emerges (Figure 2.4, Bottom). Watersheds dominated by fluvial erosion show a weak positive correlation ($R^2 = 0.34$) while glaciated watershed show a strong negative-correlation ($R^2 = 0.59$). Within fluvial erosion dominated sub-watersheds a high HI indicates that relatively more of the total watershed area occurs at high elevations. Conversely a low HI would have relatively more area at low elevations. This translates to relatively lower relief watershed area. The high HI watersheds, within fluvial erosion dominated watersheds, are primarily found in the Pahsimeroi watershed where we found a relationship between tectonic uplift and rates of sediment yield. This is discussed in more detail in section 2.6. This indicates that tectonic
forcing could account for relatively more of the watershed area occurring at higher elevations.

A different process may account for the strong negative correlation ($R^2 = 0.59$) within the glacial erosion dominated sub-watersheds. In this case, the HI may be recording the level of glaciation that occurred within these watersheds. Montgomery et al., (2001) found that generally watersheds that experienced a high level of glacial erosion had a distinct ‘shoulder’ in the hypsometric curve because of selective removal of the highest elevation portions of the watershed by glacial erosion. My data are consistent with this explanation (Figure 2.5). Despite a small amount of scatter, Figure 2.5 shows a clear difference in the relative amount of high elevation area between non-glaciated and glaciated watersheds. Watersheds that experienced a high amount of Pleistocene and Holocene glacial erosion also have higher long-term rates of sediment yield. The glaciated watersheds shown in Figures 2.4 and 2.5 are all located in the Cascade Range. In addition to the legacy of glaciation itself, in Chapter 3 section 3.5 I discuss other potential reasons behind the high rates of sediment yield from this area which include: potential bias in measured rates due to dilution with $^{10}$Be poor sediment from actively sourcing glacial outwash terrace deposits and the strong precipitation gradient of the Cascade Range.

2.7 Watershed Averaged Slope and Total Relief

G.K. Gilbert (1877) famously stated while working in the Henry Mountains of southern Utah that, “Erosion is most rapid where slope is the steepest.” This appealing
Figure 2.4 Hypsometric integral plotted against long-term rates of sediment yield. The top plot is all samples combined while the bottom plot is organized by the amount of glaciation within the watershed. Error bars represent analytical uncertainty from AMS measurement.
Figure 2.5 Normalized hypsometric curves of non-glaciated watersheds dominated by fluvial erosion processes (in red) and glaciated watersheds (in black). Note the relative lack of high elevation watershed area in the glaciated watersheds compared to the non-glaciated watersheds.

observation was fundamental to the early development of conceptual models that link mean watershed slope and relief to long-term erosion rates. Another widely cited study (Ahnert, 1970) reported a linear relationship between long-term erosion and relief. In contrast to this view, Hack (1960) hypothesized that over time landscapes attain a state of dynamic equilibrium where slope is adjusted to attain uniform erosion. For any given erosion rate, harder rock types adjust to steeper slopes and softer rock types adjust to gentler slopes. Generally the former researchers posited that steep slopes provide more potential for downslope transport of materials and hence faster erosion. Despite the fact that recent work (Montgomery and Brandon, 2002; Von Blanckenburg, 2006) has questioned this view, predictive models of sediment yield based on simple topographic parameters are still widely used for watershed management purposes. Cosmogenic
analysis (Riebe and Kirchner, 2001a) and other recent work (Willett, 1999; Pazzaglia and Brandon, 2001) has provided compelling evidence of strong coupling between tectonic activity and erosion rates. Further, Montgomery and Brandon (2002) demonstrated that relief can become decoupled from erosion rates when rates of tectonic uplift exceed certain thresholds, above which hillslopes primarily erode by landsliding in order to keep pace with fluvial incision. Because of the variability of findings from different physiographic regions regarding the relationship between watershed topography and long-term erosion rates, I investigated this relationship within study sub-watersheds in different regions of the CRB.

Mean watershed slope was calculated in degrees on 30 m DEMs using the Surface Slope Spatial Analyst tool within ArcGIS 10.0. Total watershed relief was calculated from the same DEMs by simply subtracting the maximum watershed elevation from the minimum. In cases where TCN samples were not collected exactly at the outlet of a study sub-watershed, surface slope and relief were measured only for the watershed upstream from the sample location. These values were then compared to long-term rates of sediment yield derived from TCN analysis. The results of this analysis are shown below in Figure 2.6.

When viewed as a whole across all broad study sub-watersheds there appears to be a positive correlation between mean watershed slope ($R^2 = 0.40$), total watershed relief ($R^2 = 0.28$) and long-term sediment yield. This finding is consistent with Portenga and Bierman’s (2011) review that found a weak, positive correlation between basin-averaged erosion rate and slope ($R^2 = 0.35$) or relief ($R^2 = 0.102$) from a global dataset including 1149 basins representing a wide variety of physiographic settings.
Figure 2.6 Sediment yield plotted against mean watershed slope (Top) and against total watershed relief (Bottom). All study sub-watersheds active channel $^{10}$Be samples were included (including Moon et al., 2011) (Black Dots) and a sub-set of the terrace alluvium $^{10}$Be samples (Red Dots). A weak positive correlation is evident for both slope and relief.
This relationship is interesting from a broad perspective but does not support a mechanistic relationship between topography and long-term sediment yield and fails to account for watershed (e.g., JDW, PRW, ERW, etc) specific variability. The smaller scale variability in the relationship between topography and long-term sediment yield becomes clear when we analyze individual sub-watersheds (Table 2.2) and potentially allows for a greater understanding of the specific drivers of long-term sediment yield in particular watersheds.

Table 2.2 Summary of $R^2$ values for linear regressions between slope, relief and long-term sediment yield. The general positive correlation is not consistent within each study sub-watershed.

<table>
<thead>
<tr>
<th>Watersheds</th>
<th>n</th>
<th>Sed Yield vs. Slope: $R^2$</th>
<th>Slope Signif.</th>
<th>Sed Yield vs. Relief: $R^2$</th>
<th>Relief Signif.</th>
</tr>
</thead>
<tbody>
<tr>
<td>All Samples</td>
<td>35</td>
<td>(+) 0.394</td>
<td>Yes</td>
<td>(+) 0.277</td>
<td>Yes</td>
</tr>
<tr>
<td>JDW</td>
<td>9</td>
<td>(-) 0.057</td>
<td>No</td>
<td>(+) 0.562</td>
<td>Yes</td>
</tr>
<tr>
<td>PRW</td>
<td>6</td>
<td>(+) 0.743</td>
<td>No</td>
<td>(-) 0.006</td>
<td>No</td>
</tr>
<tr>
<td>Cascades</td>
<td>20</td>
<td>(+) 0.273</td>
<td>Yes</td>
<td>(-) 0.003</td>
<td>No</td>
</tr>
</tbody>
</table>

Within the JDW mean watershed slope appears to be decoupled from long-term sediment yield, which may be reflective of a tectonically inactive landscape (Riebe and Kirchner, 2001a). The narrow range of relatively low rates of sediment yield (58 - 169 Mg/yr/km$^2$) within the JDW is consistent with an inactive tectonic setting. Decoupling of slope to long-term sediment yield is consistent with Hack’s (1960) hypothesis which
asserts that over time landscapes attain a state of dynamic equilibrium where slope is adjusted to attain uniform erosion.

Sub-watersheds sampled from the Cascade Range (e.g., ERW, WRW and SRW) exhibit a weak positive correlation ($R^2 = 0.27$) between slope and long-term sediment yield but no discernible relationship exists between relief and long-term sediment yield. Figure 2.7 shows the general positive trend of slope and long-term sediment yield but with considerable variability, especially in watersheds with slopes ranging from 24 to 28 degrees. From the distribution of the data it is clear that within specific sub-watersheds the simple bivariate relationship between slope and long-term sediment yield does not hold, except in the Pahsimeroi watershed which will be discussed in more detail in section 2.10.

2.8 Mean Annual Precipitation

The relationship between climate and long-term sediment yield is an active area of geomorphic inquiry. Over geologic time-scales (e.g., Ma) hemisphere-scale climate shifts (e.g., Pleistocene glaciations) play a large role in setting the pace of erosion and sediment yield (Montgomery et al., 2001; Fuller et al., 2009). While at the smallest time-scales high magnitude storm events can mobilize large amounts of sediment within a matter of hours to days. The role of climate over the intermediate, millennial-scale is not well understood (Riebe and Kirchner, 2001a; Riebe and Kirchner, 2001b; Von Blanckenburg, 2006).

Climate, which includes both temperature and precipitation, is thought to control the chemical and mechanical weathering of bedrock.
Figure 2.7 Sediment yield plotted against mean watershed slope from Cascade Range study sub-watersheds (ERW, WRW, SRW). All study sub-watersheds active channel $^{10}$Be samples were included (including Moon et al., 2011) (Black Dots) and a sub-set of the terrace alluvium $^{10}$Be samples (Red Dots). A weak positive correlation is evident for slope.

Additionally, climate driven vegetative feedbacks can stabilize regolith and increase infiltration rates which dampens the effects of surface erosion from runoff. While in arid climates with little vegetative cover precipitation can mobilize and transport large amounts of unconsolidated material through sheetwash (Horton, 1945). Despite this, recent work (Von Blanckenburg, 2006) assessing the role of climate on millennial-scale denudation rates suggests that it is a relatively minor driver (Riebe and Kirchner, 2001a; Riebe and Kirchner, 2001b; Portenga and Bierman, 2011) and that tectonic activity dwarves the role of climate. The CRB represents a natural laboratory to assess the role of
climate on long-term rates of sediment yield because of the abundance of diverse climatic regimes within the large watershed.

I generated sub-watershed specific values of mean annual precipitation (MAP) using data from the approximately the last 30 years (i.e., 1981 – 2010). Data was downloaded as ASCII data grids from the Oregon State University Climate Group’s Parameter-elevation Regressions on Independent Slopes Model (PRISM). PRISM data with a cell resolution of 800 m was then clipped to each study sub-watershed using the Extract by Mask tool in ArcGIS 10.0. MAP was then plotted against rates of long-term sediment yield derived from each study sub-watershed, as shown in Figure 2.8.

Figure 2.8 Mean annual precipitation plotted against long-term sediment yield for all alluvium samples including a sub-set of Moon et al., (2011) (1, Top Left), the Cascades (2, Top Right), the John Day Watershed (3, Bottom Left) and the Pahsimeroi watershed (4, Bottom Right).
When including all active channel alluvium samples from all study sub-watersheds there is a strong ($R^2 = 0.90$) positive correlation between sediment yield and MAP. The trend revealed in Figure 2.8 is surprising given the relatively minor role that MAP appears to play in driving millennial-scale erosion on a global scale (Von Blanckenburg, 2006; Portenga and Bierman, 2011). This relationship becomes more complex at the scale of the JDW and PRW where there appears to be an negative correlation though this may be an artifact of the small sample size within these sub-watersheds. An alternative explanation is that in the semi-arid JDW and PRW there may have been a shift in climate over the same millennial-scale that the rates of sediment yield average over. If this were the case, an arid landscape (like the JDW and PRW) with vegetative communities adjusted to low MAP would be highly sensitive to changes in precipitation as vegetation would be unable to provide a buffer to a changing climates effect on sediment yield. Unfortunately our data cannot address this hypothesis.

Within the JDW and PRW both precipitation and sediment yield are low which is consistent with the general trend of increasing sediment yield with increasing precipitation. Additionally the range of precipitation and sediment yield within these watersheds is small. In contrast, within the Cascade Range where the range of precipitation and sediment yield is large there is a strong positive correlation ($R^2 = 0.85$). From this data set it appears that mean annual precipitation and long-term sediment yield are strongly correlated across the wide range of environments sampled. In order to determine the strength of this relationship throughout the entire CRB it would be advantageous to have a larger sample size particularly from the arid portions of the CRB.
2.9 A Case Study from the Entiat River Watershed

Comparing long-term rates of sediment yield derived from the concentration of $^{10}$Be in active channel alluvium from two study sub-watersheds within the ERW provides a compelling example of how variability of topographic metrics, longitudinal profile analysis, and MAP, might lead one to reasonably expect a large amount of variability in the long-term rates of sediment yield between the two study sub-watersheds. Despite this variability, cosmogenically derived rates of sediment yield derived from active stream alluvium from the two watersheds are nearly identical.

2.9.1 Study Setting

The bedrock of the ERW is composed of the Mad River, Swakane, and Chelan Mountains geologic terrains. These are resistant, metamorphic and intrusive rocks. The rock types are primarily quartz diorite, tonalite, schist, granodiorite, gneiss, and migmatite. The Duncan Hill granodiorite pluton intruded into the Chelan Mountains terrain and underlies a large portion of the Entiat River (Tabor, 1987) including the upper elevation study sub-watershed, Silver Creek (Figure 2.9). The lower elevation study sub-watershed, Roaring Creek is underlain almost exclusively by Precambrian Swakane biotite gneiss.

A large portion of the upper valley of the ERW, including Silver Creek, was glacially sculpted during the Pleistocene Epoch by a 56 km-long valley glacier (Long, 1969) and abundant tributary glaciers in the upper elevation portion of the watershed. This glacier sculpted the pre-existing river valley into the U-shaped profile characteristic of glaciated valleys and left a prominent terminal moraine across the valley (Figure 2.9). Below this point the valley morphology exhibits the typical V-shape driven by fluvial
incision (Long, 1951). Roaring Creek is located exclusively in this low elevation, fluvial erosion dominated region.

Longitudinal profile analysis conducted on the mainstem Entiat River and major tributaries reveals distinct topographic features associated with the glacial history of the landscape (Figure 2.9). Upper elevation tributaries (upstream of the prominent terminal moraine of the main valley glacier, noted in Figure 2.9) experienced a high level of glaciation and exhibit either entirely convex profiles or have substantial portions of the longitudinal profile that are convex. Additionally, all upper elevation tributaries and the upper mainstem river itself have pronounced knickpoints that are likely related to past glaciations. In contrast, the low elevation tributaries that are located downstream of the terminal moraine show no topographic signs of glaciation and exhibit well-graded concave up profiles with no major knickpoints (Figure 2.9).

2.9.2 Climate/Hydrology of the Entiat Watershed

Temperatures in the Entiat watershed vary from 32 to 38 °C in the summer to below 0 °C in the winter (based on three climatological stations within the area (Plain, Chelan, Lake Wenatchee), period of record: 1949-2012). The majority of precipitation falls as snow during the winter months and spring rain with a mean annual precipitation of 96 cm. Topography influences this general trend and there is a marked difference in the total amount of precipitation in the higher versus lower elevation portions of the ERW. Higher elevations experience approximately 250 cm of precipitation annually while the lower reaches nearby the confluence with the Columbia only have 25 cm annually (Kirk et al., 1995). Silver Creek receives 112 cm of precipitation annually while Roaring Creek has 90 cm of precipitation annually.
Figure 2.9 Longitudinal profiles of the Entiat River and major tributaries (Top). Location of tributaries (Bottom Left). Silver Creek’s longitudinal profile is shown in red and the watershed boundary is highlighted in orange (Bottom Right), cosmogenic sample locations are shown in red. Roaring Creek’s longitudinal profile is highlighted in dark blue and the watershed boundary is illustrated in orange. Bottom right shows the bedrock geology of the ERW.
Like much of the CRB, the ERW is primarily a snowmelt driven system with peak flows typically corresponding to peak timing of snowmelt in the late spring and early summer. Lowest flows are typically in the late summer and early fall.

The annual to two year peak discharges are typically on the order of 28-96 cms (USGS gage: 12452800, Entiat R. near Ardenvoir, WA period of record: 1957-2013) (Godaire et al., 2010). Deviations from this pattern can be from rain on snow events or intense short duration summer thunderstorms that temporarily cause a spike in the hydrograph. Flow duration curve analysis of USGS gage 12452800 shows a large amount of decadal variability with a general increase in the mid-range flows (20 – 55% exceedance) and a decrease in the base flows (90 – 100 % exceedance) from 1970 – 2000’s, except for the most recent years (2010 – 2013) (Figure 2.10, A). These data suggest that high flows have not increased or decrease appreciably due to climate or human-induced changes.

The observed decrease in base flows from the 1970’s through to the 2000’s may be caused by a major increase in water claims for agricultural irrigation filed with the Washington department of Ecology starting in the early 1970’s (Kirk et al., 1995). Additionally, the increase in mid-level flows observed from the 1970’s through to the 1990’s could have been influenced by the change in run-off characteristics following the massive fire (486 km²) that burned much of the watershed in 1970. The wildfire of 1970 was among the most extensive to have occurred in the Pacific Northwest during the period of 1846 through 1971 (Helvey, 1980). Helvey conducted flow duration curve analysis pre- and post-fire in three experimental watersheds located in the Entiat watershed and found that the entire range of flows doubled post-fire (1973-1977).
Consistent with that analysis, Woodsmith et al., (2007) found that peak flows increased dramatically in the ten years following the fire as shown in Figure 2.10 B. Forest regrowth during subsequent decades reduced peak and mid-range flows to near or below their 1960’s levels. Of course, natural decadal variability also complicates these trends.

2.9.3 Methods

River borne sand was collected from the active stream alluvium near the watershed outlet in Silver Creek and Roaring Creek (Figure 2.9, Bottom Right). Samples were analyzed for $^{10}$Be concentration at the Purdue University Rare Isotope Measurement Laboratory (PRIME) using accelerator mass spectrometry. The concentration of $^{10}$Be was then converted to catchment-averaged erosion rates and ultimately sediment yield using equation. Topographic attributes (relief and slope) were generated using the Spatial Analyst Tools within ArcGIS 10.0 on 30 m DEMs. Channel steepness (Ksn) was generated using the Stream Profiler Tool (Whipple et al., 2007) on 30 m DEMs with a reference concavity of 0.45 and a smoothing window of 250 m. Mean annual precipitation represents data from the approximately the last 30 years (i.e., 1981 – 2010). Data was downloaded as ASCII data grids from the Oregon State University Climate Group’s Parameter-elevation Regressions on Independent Slopes Model (PRISM). PRISM data with a cell resolution of 800 m was then clipped to the two study sub-watershed using the Extract by Mask tool in ArcGIS 10.0.

2.9.4 Summary of Topographic and Environmental Parameters

Pleistocene glaciation within Silver Creek left a distinct topographic signature on this watershed compared to the unglaciated Roaring Creek watershed.
Figure 2.10 Flow duration curves organized by decade for the Entiat River, WA. Note the temporary decrease in base flows (90-100%) (A), possibly related to agricultural diversions, and the temporary increase in mid (20-50%) and high level (<1%) flows associated with the changes to run-off characteristics associated with the large wildfire of 1970 (USGS gage 12452800, Entiat R. near Ardenvoir, WA period of record: 1957-current).
Upper Silver creek has a broad, relatively low gradient valley and channel (Ksn = 77.5) morphology with steep hillslopes. Downstream this transitions at a substantial knickpoint (Figure 2.9) into a steep channel (Ksn = 269.4) until the confluence with the Entiat River. The knickpoint appears to be an artifact of the downstream limit of the alpine glacier. Despite the steep channel section, mean watershed slope is low compared to Roaring Creek (Table 2.3). In contrast to the glacially carved morphology of Silver Creek, Roaring Creek reflects the dominance of fluvial erosion processes. For example, the valley morphology is confined with direct coupling of steep hillslopes to the mostly straight channel. Additionally, the concave-up profile has no knickpoints and maintains a fairly constant channel steepness (steeper than upper Silver but less steep than lower Silver Creek: Table 2.3) throughout the length of the longitudinal profile. Mean slope in Roaring Creek is considerably higher than Silver because of the lack of a broad, glacially carved upper valley. Total relief is much higher in Roaring Creek compared to Silver (Table 2.3) but MAP in Silver Creek is higher as it is closer to the range crest. The difference in rock types between the two watersheds appears negligible as the high-grade metamorphic biotite gneiss that underlies Roaring Creek has similar erosive properties to the intrusive, quartz diorite pluton that composes the bedrock of Silver Creek.

2.9.5 Discussion

Based on traditional geomorphic relations it is difficult to predict which watershed one would expect to yield more sediment over the long-term. Watershed slope and relief are considerably higher in Roaring Creek suggesting that these topographic parameters would encourage a relatively higher rate of sediment yield.
Table 2.3 Summary of topographic and environmental parameters from Silver Creek and Roaring Creek, Entiat River Watershed, WA. Also shown are the long-term rates of sediment yield which are nearly identical.

<table>
<thead>
<tr>
<th>Watershed Parameters</th>
<th>Silver Creek</th>
<th>Roaring Creek</th>
</tr>
</thead>
<tbody>
<tr>
<td>Glaciation</td>
<td>Yes</td>
<td>No</td>
</tr>
<tr>
<td>Presence of Knick Points</td>
<td>Yes</td>
<td>No</td>
</tr>
<tr>
<td>Channel Steepness (Ksn)</td>
<td>180.7</td>
<td>122.8</td>
</tr>
<tr>
<td>Slope (degrees)</td>
<td>20.7</td>
<td>27.0</td>
</tr>
<tr>
<td>Relief (m)</td>
<td>1072</td>
<td>1443</td>
</tr>
<tr>
<td>MAP (cm)</td>
<td>112</td>
<td>90</td>
</tr>
<tr>
<td>Rock Type</td>
<td>Intrusive</td>
<td>Metamorphic</td>
</tr>
<tr>
<td>Long-term Sediment Yield (Mg yr$^{-1}$km$^{-2}$)</td>
<td>255 +/- 21</td>
<td>257 +/- 25</td>
</tr>
</tbody>
</table>

But MAP is higher in Silver Creek and Figure 2.8 (2) suggests that there is a strong positive correlation between MAP and long-term sediment yield within the Cascade Range. Average channel steepness (Ksn) is considerably higher in Silver Creek which could facilitate more rapid transport of eroded material once it reached the channel network. This could lead to a higher total rate of sediment yield over time if the watershed had ample sediment sources. Glaciation within upper Silver Creek could mean more abundant near-channel sediment sources as the alluvial valley is undoubtedly underlain by glacial deposits. The relative importance of each of these parameters in regards to the driver(s) of long term sediment yield are impossible to untangle from my data alone but based on the variability of parameters it is surprising that the rates are nearly identical (Table 2.3). This case illustrates the inadequacy of bivariate relationships between long term sediment yield and simple topographic and climatic parameters.
2.10 A Case Study from the Pahsimeroi Watershed, Idaho

2.10.1 Introduction and Study Setting

Unlike the previous example where long-term sediment yield appears to be decoupled from topography the following case study provides an example where a presumed variability of tectonic uplift rates within the Lemhi Mountain Range has generated unexpected relationships between topography and long term sediment yield. The Pahsimeroi watershed is located in the northeastern portion of the Basin and Range geologic province in eastern Idaho with the Snake River Plain directly to the south. The broad U-shaped alluvial valley is bounded by the Lemhi Range to the northeast and the Lost River Range to the southwest (Figure 2.11).

The southwest dipping, active extensional Lemhi fault array establishes the eastern boundary of the alluvial valley (Janecke, 1993) and is responsible for the uplift of the Lemhi range. This 150 km long fault array consists of six or seven segments (Crone and Haller, 1991; Janecke, 1993) (Figure 2.11). Faulting activity is thought to have begun during the late Miocene and continues to this day, with 5-6 km of total fault throw (Densmore et al., 2004). There is also evidence of an older, Paleogene normal fault within the northern region of the Lemhi Range footwall. Dramatic evidence of modern fault activity within this region comes from the neighboring Lost River normal fault, to the south. In 1983 the Mt Borah earthquake was caused by tectonic activity within the Lost River fault array. The earthquake raised the Lost River range by ~ 30 cm with an accompanying drop of the headwall (i.e., alluvial valley) by ~ 1.2 m (Shumar et al., 2001).
Bedrock within the three study sub-watersheds in the Lemhi Range is fairly uniform (Figure 2.11). Precambrian metamorphics (quartzite and siltstone) underlie over 90% of the study sub-watersheds with Pleistocene glacial deposits overlying the bedrock in small areas within the upper elevations of each study sub-watershed (Figure 2.11). The mostly flat Pahsimeroi valley is composed of glacial and fluvial alluvium. Large alluvial fans issue from the toes of the Lemhi and Lost River Ranges covering the alluvial valley and in some locations extending to the margins of the Pahsimeroi River. This coarse alluvium is responsible for causing shallow subsurface flow that locally causes the Pahsimeroi River to flow entirely subsurface (Shumar et al., 2001). This dynamic, coupled with a large amount of agricultural diversions may be responsible for unusual trends observed in the flow duration curves (Figure 2.12) from the Pahsimeroi River.

2.10.2 Climate/Hydrology

Temperatures vary from 24 °C in the summer to a -11 °C in the winter (Mean monthly PRISM data), period of record: 1980 - 2010). These temperatures reflect the watershed-wide average. Temperatures within the alluvial valley alone are warmer in summer and less cool in winter than the watershed-wide average. The majority of precipitation falls as snow in the upper elevations during the winter months with a watershed-averaged mean annual precipitation (MAP) of 49 cm. As the study sub-watersheds are located within the upper elevation portions of the watershed MAP is higher than the watershed-wide average. Big Creek experiences 82 cm annually, Patterson 81 cm and Morse 68 cm (data source: PRISM). Because of the high rates of infiltration within the coarse valley alluvium and the high level of agricultural diversions
Figure 2.11 Longitudinal profiles of the Pahsimeroi River and major tributaries draining the Lemhi Range (Top). Location of tributaries (Bottom Left), study sub-watersheds (Black) and sample locations (Red). Longitudinal profiles exhibit a well graded concave profile with small knickpoints located around the active normal fault on the eastern edge of the alluvial valley. Channel steepness (Ksn) (Top), slope and long-term rates of sediment yield all increase along a northern gradient with fault activity (Top) while relief appears decoupled from this trend.
Figure 2.12 Flow duration curves organized by decade for the Pahsimeroi River, Idaho. Note the sharp increase in magnitude of the very highest flows (< 5% exceedance). This is most likely due to the high rates of infiltration within the valley alluvium that are overcome during the highest magnitude flow events. USGS gage: 13312005 period of record: 1984 – current.

much of the tributary runoff does not reach the Pahsimeroi river except during the highest spring snowmelt flows (Shumar et al., 2001). This is reflected in the abrupt increase evident in the highest magnitude flows (<5% exceedance) from flow duration curve analysis (Figure 2.12). Aside from this no systematic trends were identified in the hydrological analysis. The relatively short period of record (1984 – current: USGS gage 13312005 located just upstream of confluence with Salmon River) precludes an in depth analysis of flow conditions before the period of wide spread agricultural diversions.
2.10.3 Methods

River borne sand was collected from the active stream alluvium in the three sample sub-watersheds (Figure 2.12). Two samples were collected within each watershed; Patterson and Morse were sampled at lower and upper elevations while Big Creek had two samples collected just upstream from the confluence with another large tributary. Samples were analyzed for $^{10}$Be concentration at the Lawrence Livermore National Laboratory’s Center for Accelerator Mass Spectrometry in Livermore, CA. The concentration of $^{10}$Be was then converted to catchment-averaged erosion rates and ultimately sediment yield following the methods outlined in section 2.3. Topographic attributes (relief and slope) were generated using the Spatial Analyst Tools within ArcGIS 10.0 on 30 m DEMs. Channel steepness ($K_{sn}$) was generated using the Stream Profiler Tool (Whipple et al., 2007) on 30 m DEMs with a reference concavity of 0.45 and a smoothing window of 250 m.

2.10.4 Functional Relationships between Long-term Sediment Yield and Topography

Within the northwest region of the Lemhi Range, average channel steepness ($K_{sn}$) and mean watershed slope both increase in a northwestern direction following the active normal fault (Figure 2.11). This trend mirrors the rates of long-term sediment yield from the three study sub-watersheds (Figure 2.11). This suggests that the northern tip of the Lemhi fault array is relatively more active than the central portion of the fault complex. This creates a functional relationship between uplift rates, millennial-scale rates of erosion and rates of river incision. Many researchers (Riebe and Kirchner, 2001a; Riebe and Kirchner, 2001b; Montgomery and Brandon, 2002; Von Blanckenburg, 2006;
Portenga and Bierman, 2011) have found similar relationships in other tectonically active regions where tectonic uplift rates are the dominant control on rates of erosion. Tectonic uplift exposes fresh surfaces for chemical and mechanical weathering which causes millennial-scale rates of erosion (Stallard and Edmond, 1983) to increase in a predictable manner. As uplift occurs this causes a lowering of local base level and forces a readjustment of the drainage network which is consistent with our findings of increased slope and channel steepness with increased rates of erosion (Von Blanckenburg, 2006).

In contrast to the findings discussed above, watershed relief appears decoupled from millennial-scale rates of erosion and sediment yield. Densmore et al. (2004) observed in the southern portion of the Lemhi Range that relief increases within 15 km of the southern fault tip until it reaches a fairly uniform value throughout the rest of the fault array. This is due to the dynamics of fault propagation where maximum fault displacement has occurred near strike center (i.e., the central part of the range in the longitudinal direction) and decays to zero at the fault tips. He also hypothesized that the topography of the northwestern portion of the Lemhi Range, where this study was focused, is probably influenced by inherited topography from the activity of the older, Paleogene fault located in the fault footwall. In other words, relief was probably established over the million year timescale by fault activity of the older Paleogene fault. This observation is consistent with the lack of a correlation between modern relief and millennial-scale sediment yield within our study sub-watersheds. Relief in this portion of the Lemhi range appears to be driven by fault activity over the million year timescales while slope and channel steepness are tightly coupled to faulting over the millennial-scale. Densmores’ (2004) hypothesis of maximum erosion rates located near strike center
is probably still true if we consider paleo-fault activity during the Paleogene but more recently (i.e., over the millennial-scale) our data suggests that the northern tip of the fault array is relatively more active driving increased erosion and sediment yield in this region. It appears that we have identified a functional relationship between watershed topography and long-term sediment yield within this tectonically active mountain range. This is in contrast to the other study sub-watersheds (JDW, ERW, WRW and the SRW) where we found no such relationships.

2.11 Conclusions

In this chapter I have shown evidence of the variability of long-term sediment yield at multiple scales within the CRB. I have provided examples from within CRB where long-term rates of sediment yield generally are poorly predicted from simple bivariate relationships with topographic and environmental parameters. Where functional relationships exist, the nature of those relationships are scale-dependent. In other words, functional relationships at the scale of the entire CRB either disappear or reverse at the scale of the entire study sub-watershed (e.g., JDW, PRW). Rock type appears to be a poor predictor of long-term sediment yield at any scale. Hypsometric curves appear to coarsely differentiate watersheds based on the level of glaciation. While hypsometric integrals exhibit a weak positive correlation ($R^2 = 0.34$) with sediment yield for fluvial erosion dominated watersheds and a stronger negative correlation ($R^2 = 0.59$) in glacial erosion dominated watersheds. Slope and relief are weak predictors at the largest scale (slope: $R^2 = 0.40$, relief: $R^2 = 0.28$) but this relationship either disappears or changes drastically within individual study sub-watersheds (Table 2.2). It should be noted that the cosmogenic sample size (n) decreases substantially at the scale of the individual study
sub-watersheds within the JDW (n = 9) and PRW (n = 6). In these locations sample size may be too low to draw a robust conclusion about the nature of the relationships between slope, relief and long-term sediment yield. Despite this, sample size is high within the Cascades (n = 20) and the positive correlation between slope, relief and long-term sediment yield observed at the scale of the entire CRB (slope: $R^2 = 0.40$, relief: $R^2 = 0.28$) is considerably different than within the Cascade Range study sub-watersheds (slope: $R^2 = 0.273$, relief: $R^2 = (-) 0.003$).

It appears that MAP is a robust predictor ($R^2 = 0.90$) of long-term sediment yield at the scale of the entire CRB with notable exceptions at a smaller scale. Within the JDW and PRW there is a weak negative correlation (JDW $R^2 = 0.27$ and PRW $R^2 = 0.65$) but the sample size (n) is small and the range of MAP and sediment yield are also small making it difficult to draw a strong conclusion about the role of MAP in driving sediment yield within these sub-watersheds. Within the Cascade Range where the range of MAP and sediment yield is large there is a strong relationship ($R^2 = 0.85$) between MAP and sediment yield. Despite this, the equivalent rates of sediment yield from Silver Creek and Roaring Creek show that there are exceptions to the broad trend of increasing sediment yield with increasing precipitation within the Cascade Range. This provides evidence of the existence of other drivers of long-term sediment yield within the Cascade Range at smaller spatial scales. Finally, the case study from the Lemhi Range within the Pahsimeroi watershed shows the primary role of tectonics in driving millennial-scale rates of sediment yield and forcing relationships between topography and sediment yield. The lack of correlation with watershed relief also illustrates how the nature of that relationship is often more complex than traditional geomorphic assumptions might lead
us to believe. In summary, this chapter further reinforces the complexity of sediment
generation, routing and deposition over multiple spatial scales within the CRB. The
general inadequacy of predicting long-term sediment yield from simple topographic and
environmental bivariate relationships suggests that multiple drivers may be setting the
pace of long-term sediment yield in the CRB. Because of this, a more rigorous
multivariate analytical approach may be needed to accurately predict sediment yield at
multiple spatial scales in the CRB. The cosmogenically derived rates of long-term
sediment yield analyzed in this chapter could be used for that purpose.
CHAPTER 3

SPATIAL AND TEMPORAL VARIABILITY OF SEDIMENT YIELD IN THE
COLUMBIA RIVER BASIN: CASE STUDIES FROM SPECIFIC STUDY SUB-
WATERSHEDS

3.1 Introduction

In the previous chapter I examined broad trends of sediment dynamics within the
CRB and identified a number of functional relationships between topographic and
environmental parameters (e.g. slope, relief and mean annual precipitation) and long-term
sediment yield. In addition to the broadest scale, each smaller watershed (e.g., ~ 10 –
2,000 km²) has a distinct geologic, geomorphic and disturbance history that sets the
template for the modern sediment dynamics and the physical aspects of aquatic habitat.
The processes of sediment generation, routing and sediment yield operating at the finer
scale influences how a watershed responds to natural and anthropogenic disturbances and
is important to quantify and analyze.

Here I used a combination of GIS analysis, geomorphic mapping, rapid
g geomorphic assessments, grain-size analyses, optically stimulated luminescence (OSL)
and Carbon-14 (C14) dating and ^10^Be-derived CAER to unravel the processes of
sediment generation, routing and deposition at the scale of individual study sub-
watersheds. Specifically I sought to identify and quantify changes in sediment
generation, routing and deposition associated with natural (e.g., wildfire and glaciation)
and anthropogenic (e.g. grazing, agriculture and mining) disturbances. Within this
chapter, I have selectively identified small watersheds throughout the CRB that allowed
me to explore these questions to a first order. I located three sub-watersheds to the John
Day watershed (JDW) in Oregon, two sub-watersheds within the Wenatchee River
watershed (WRW) in Washington, and the Skykomish River watershed (SRW), which is
immediately adjacent to the WRW but not within the CRB. The SRW was included
because of existing CAER (Moon et al., 2011) that could be leveraged to address my
research questions.

3.2 Methods

3.2.1 Geologic and geomorphic mapping

Digital elevation models (DEM) of 10 and 30 m resolution (Source: USGS
National Elevation Dataset (NED)) were used in combination with state-wide, digital
surficial geologic layers to create geologic maps for each study sub-watershed. Individual
rock-types were classified into five, broad geologic categories (e.g., Sedimentary,
Metamorphic, Igneous Extrusives, Igneous Intrusives, Eolian) to simplify interpretation
and analysis. These maps provided geologic and geomorphic context, assisted in the
identification of rock-types and landforms relevant to the geochemical (e.g., location of
fluvial terraces and quartz bearing rock types) sampling as well as aided in analysis and
interpretation of CAER results (see Appendix C for digital geologic layer sources, and
the broad categorizations). Additionally, a 1 m resolution DEM derived from aerial Light
Detection and Ranging (LIDAR) data was used in combination with aerial imagery, and
rapid geomorphic assessments to map alluvial fans and terraces adjacent to the modern
stream within one study sub-watershed, Bridge Creek, in the John Day watershed,
Oregon. The purpose of this mapping was to differentiate between tributary alluvial fans
and main-stem fill terraces directly adjacent to the deeply incised main channel of Bridge Creek. This will be discussed in more detail within section 3.4.1.3.

3.2.2 Rapid geomorphic assessments

Field-based, rapid geomorphic assessments were conducted, totaling 100 km of stream length in eight sub-watersheds within the JDW including the three study sub-watersheds (e.g., Bridge Creek, Granite-Boulder Creek, and East-Branch North-Fork Cable Creek) following the Fluvial Audits approach (Sear et al., 1995) and the River Styles framework (Brierley and Fryirs, 2005). These surveys were intended to qualitatively survey and describe the locations and types of sediment supply, transport and storage within the study sub-watersheds and locate geomorphic reach breaks or boundaries that can identify areas of process change (e.g., a transition from an open alluvial to a confined bedrock setting). Each reach had the following attributes classified and recorded. Reach length scaled to channel width but was typically between 100 m to a few kilometers in length.

- Sediment size, distribution
- Valley and channel width
- Gradient
- Sinuosity
- Channel reach morphology (Description of the features and form of the channel reach)
  - Planform Type (e.g., single threaded, braided)
  - Degree of floodplain connectivity
  - Type, distribution and relative amount of geomorphic units (e.g., pools, riffles, point bars, etc.)
  - Presence or absence of large wood
- Presence or absence of riparian vegetation
- Dominant bank material and type
- Geomorphic unit morphology (Description of the features and form of representative geomorphic units)
  - Pool types (e.g., forced vs. scour, etc.)
  - Sediment size and distribution within geomorphic units

In addition to identifying and qualitatively describing the large-scale sediment dynamics and geomorphic transitions, these surveys informed the geochemical sampling within each study sub-watershed and aided in the development of a River Styles Framework classification effort conducted by Alan Kasprak (Utah State University PhD candidate).

3.3 Study Setting

3.3.1 Major sub-watersheds of interest

Sub-watersheds were selected in coordination with the National Oceanographic and Atmospheric Administration’s (NOAA’s) Integrated Status and Effectiveness Monitoring Program (ISEMP) and the Columbia Habitat Monitoring Program (CHaMP), who are conducting watershed-wide salmonid habitat restoration and monitoring efforts within select sub-watersheds of the CRB. From these sub-watersheds a sub-set was chosen based on existing data sources and the ability to locate anthropogenic (e.g., recent history of mining or grazing) and natural experiments (e.g., recent high-intensity wildfire) to exploit. This focused my research on five principal sub-watersheds; the John Day River Watershed in central Oregon (Figure 3.1), the Wenatchee, Skykomish (Figure 3.22) and Entiat River watersheds in central Washington and the Pahsimeroi Watershed in southeastern Idaho. The focus for this chapter will be on sub-watershed specific analyses within the JDW, WRW and SRW.
3.3.2 John Day Watershed Geographic Setting

The John Day watershed (JDW) (Figure 3.1) is located in northeastern Oregon and drains approximately 20,719 km$^2$ from the headwaters in the Strawberry Mountains (2743 m above sea level) to its mouth at the Columbia River (61 m above sea level). The watershed is comprised of four major sub-watersheds, including: the North Fork John Day, Middle Fork John Day, Upper and Lower John Day (Figure 3.1). It is the second largest undammed tributary in the western US, second only to the Yellowstone River (CBMRC, 2005). Within the context of this chapter, three study sub-watersheds will be discussed: Bridge Creek, a 4th order tributary to the lower main-stem John Day River, Granite-Boulder Creek, a 2nd order tributary to the upper Middle Fork John Day River and the East Branch of North-Fork Cable Creek, a 2nd order tributary to North-Fork Cable Creek and ultimately the upper North Fork John Day River (Figure 3.1).

3.3.3 Geology

The JDW is located between two prominent physiographic provinces in the western United States. The Columbia Plateau, a vast region of flood basalts covering portions of eastern Washington, southern Idaho and eastern Oregon, forms the northern border and to the south is the immense Basin and Range province, a region of north-south trending normal faults with faulted mountains and flat valleys that extends into Mexico. The JDW is characterized by a diversity of landforms ranging from high elevation alpine peaks in the headwaters to low elevation plateaus covered in loess.
Figure 3.1: DEM of the John Day watershed located in north-eastern Oregon. Study sub-watersheds are highlighted in yellow (A = Bridge Creek, B = Granite-Boulder, C = East-Branch North-Fork Cable Creek). Sample locations are indicated with yellow dots and precipitation gages are shown in red.
A diversity of rock assemblages occur, including: marine sediments, oceanic crust, volcanic assemblages, ancient lacustrine and fluvial deposits and recent fluvial and landslide deposits (CBMRCD, 2005). Three key stratified volcanic formations comprise the majority of the watershed (labeled Extrusives in Figure 3.2), formed in the past 55 million years: the Clarno formation (37-54 Ma), John Day formation (18-37 Ma) and the Columbia River Basalt group (12-17 Ma). The Columbia River Basalt group is the youngest of the three resulting from a series of at least 40 separate basalt flows that occurred between 12 and 17 million years ago. The modern landscape began to take shape during this period. The last major volcanic eruption within the JDW occurred seven million years ago forming the Rattlesnake stratum, though volcanoes from surrounding areas continued to affect the region into the Holocene. Quartz producing rock types are relatively scarce throughout the JDW because of the dominance of volcanic basalt flows but plutonic intrusions (Intrusives in Figure 3.2) scattered throughout the high elevation regions of the Middle Fork and North Fork John Day rivers contain quartz producing rock types (e.g., Granodiorite, Quartz Monzonite) as well as some quartz bearing volcanic extrusives (e.g., Andesite and Rhyolite) and other quartz bearing volcanic rocks (e.g., Dacite) that are scattered throughout the watershed. Figure 3.3 shows the locations of all potentially quartz bearing rock types that occur within the JDW.

### 3.3.4 Climate/Hydrology

The JDW is typified by low winter and high summer temperatures with relatively less precipitation than the rest of the Columbia River Watershed. Because of the orographic effect of the Cascade Range, the JDW experiences an average annual precipitation that ranges from 30 cm in the lower elevation portions to 127 cm in the
Table 3.1 Characteristics of study sub-watersheds within the John Day watershed. Precipitation data was collected from the PRISM Climate Group and represents the past 30 year annual averages.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>John Day</th>
<th>Bridge Creek</th>
<th>GB Creek</th>
<th>EBNFC Creek</th>
</tr>
</thead>
<tbody>
<tr>
<td>Drainage Area (km²)</td>
<td>20,719</td>
<td>697</td>
<td>31.1</td>
<td>11.5</td>
</tr>
<tr>
<td>Stream Length (km)</td>
<td>457</td>
<td>35</td>
<td>12.2</td>
<td>3.8</td>
</tr>
<tr>
<td>Max Elevation (m)</td>
<td>2743</td>
<td>2075</td>
<td>2477</td>
<td>1916</td>
</tr>
<tr>
<td>Min Elevation (m)</td>
<td>61</td>
<td>451</td>
<td>1137</td>
<td>1400</td>
</tr>
<tr>
<td>Relief (m)</td>
<td>2682</td>
<td>1625</td>
<td>1340</td>
<td>516</td>
</tr>
<tr>
<td>Mean Slope (degrees)</td>
<td>13.5</td>
<td>15.3</td>
<td>20.1</td>
<td>18.8</td>
</tr>
<tr>
<td>Mean Precipitation (cm/yr)</td>
<td>46.5</td>
<td>42</td>
<td>78.5</td>
<td>71.8</td>
</tr>
</tbody>
</table>

mountainous high elevation regions (NRCS SNOTEL Site: 357, period of record: 1978-current (gage location shown on Figure 3.1)). The majority of precipitation (90 %) falls as snow between the months of November and March, though significant summer thunderstorms have been responsible for mass wasting events throughout the JDW (Howell, 2006). The upper elevations are considered sub-humid and the lower sub-arid.

The majority of water in the John Day River originates in the upper portions of the watershed as melting snow. Due to the lack of large-scale dams, the hydrograph has a pronounced peak, typically from March through May corresponding with the height of spring snowmelt runoff. Yearly low flows typically occur from August through October. Flow duration curves were evaluated, initially on a decadal time-step and then multiple decades were combined into three hydrologically distinct eras (i.e., 1900-1920’s, 1940-1970’s, 1980 – Current) that show a small reduction in base flows (i.e., 90-100% exceedance flows) over the course of the hydrologic record (USGS gage 14048000, period of record 1904-current (gage location shown on Figure 3.1)) accompanied by an
Figure 3.2: A simplified representation of the bedrock geology of the JDW. Geologic layers were collected from the state-wide digital GIS portal and categorized into the categories shown above. For more information on the sources of data and the classification scheme see Appendix C.
Figure 3.3: Potentially quartz bearing rock types that occur within the JDW. Geologic layers were collected from the state-wide digital GIS portal and categorized based on the likelihood of containing quartz. For more information on the sources of data and the classification scheme see Appendix C.
increase in the very highest flows (i.e., < 1% exceedance) but with no real increase in the 1 – 10% exceedance flows (Figure 3.4). Additional analysis, classified by decade but not displayed, shows a large amount of non-systematic decadal variability. These trends are discussed further in the next section. Magnitude of discharge from base to peak-flow routinely varies over 30X inter-annually (5.6 – 164 cms), while peak flow can vary as much as 3-8X intra-annually (152 – 1211 cms) (CBMRCD, 2005).

3.3.5 Human perturbations in the watershed

Historic and current human activities have altered the hydrology and sediment yield within the John Day River network in a variety of ways starting in 1862 with the first Euro-American settlement of the watershed. Previous perturbations likely occurred as a result of Native American land and wildlife management, but effects of these activities is largely unknown. Water withdrawals from small dams, diversions and channelization for agricultural purposes started in the late 19th century. This activity coupled with extensive mining (e.g., hydraulic, underground, and placer mining), livestock grazing and timber extraction have likely affected the runoff regime within the JDW. As evidenced in Figure 3.4 the hydrologic curve has shifted only slightly since the 1900’s when the gaging record began with an increase in maximum flows (i.e., <1% exceedance) and a reduction in base flows (CBMRCD, 2005). While the reduction in low flows appears small, this reduction could still be problematic for aquatic ecosystems, particularly resident and migratory salmonids as the low flow period occurs during the late summer and early fall when air and water temperatures are high.
Figure 3.4 Flow duration curves separated by different eras for the John Day River, OR. Note the increase in maximum flows (<1%) (B) and the reduction in base flows (90-100%) (A) over the course of the hydrologic record (USGS gage 14048000, period of record: 1904-current). Values shown in top-right corner of each plot indicate averages over the quantiles indicated.

Excessive livestock grazing, extensive mining, timber extraction and the accompanying road construction have increased rates of sedimentation in many locations and reduced riparian vegetation impacting water quality and temperature (CBMRC, 2005). These impacts have led to the water quality impaired listing on portions of the
John Day River. Additionally, timber extraction coupled with over 100 years of fire suppression has led to changes in forest ecology (i.e., a shift towards dense stands of even-aged tree species vulnerable to parasite infestation and wildfire) (CBMRCD, 2005; Howell, 2006). Post-fire, high-intensity precipitation events have triggered mass-wasting (e.g., landslides and debris flows) which temporarily elevates the rate of sediment yield into the channel network. Even with the extensive alterations, the aquatic environment of the John Day is in relatively better condition than other portions of the CRB, largely because of the absence of large dams blocking anadromy (CBMRCD, 2005) and water and sediment flux.

3.4 Study Sub-Watersheds within the JDW

The following section will introduce each study sub-watershed in more detail; explain specific methods employed to address the research objectives, and present results and discussion specific to each sub-watershed.

3.4.1 Bridge Creek, JDW, Oregon

3.4.1.1 Introduction

Bridge Creek is a 4th order tributary to the lower John Day River, Oregon (Figure 3.1) with a drainage area of 697 km² (Table 3.1). The surficial geology is representative of what can be found in the JDW as a whole (Section 3.3.3). The lower sub-watershed is primarily underlain by the easily erodible eolian-deposited ash found in the Painted Hills Unit of the John Day Fossil Beds. A relatively large area of the mid and upper watershed composed of Cretaceous, marine sedimentary formations, which are also easily erodible (Figure 3.2).
Bridge Creek is arid compared to the JDW as a whole, with mean annual precipitation of 42 cm/yr. The runoff regime, like the rest of the JDW, is dominated by spring snowmelt but summer thunderstorms, which can easily exceed the infiltration capacity of the clay-rich soils, occasionally generate flash floods that are thought to be important, and perhaps dominant, sediment mobilizing events within portions of Bridge Creek (Peacock, 1994; Kasprak and Wheaton, 2012). The longitudinal profile of Bridge Creek and its tributaries reveal well-graded, concave-up profiles with no major knickpoints (Figure 3.5) indicating no obvious systematic trends in differential uplift rates or response to a drop in base level of the main stem of Bridge Creek or the John Day River.

The axial stream of Bridge Creek, particularly in the lower section, is deeply incised (2-5 m) with an abundance of fill terraces and alluvial fans located in close proximity to the stream. While in the mid-section there are a number of strath terraces 12 – 20 m above the height of the modern stream. The incised condition, coupled with decades of intensive cattle grazing that ended in the early 1990’s, is thought to have reduced the complexity of aquatic and riparian habitat by creating a simplified channel planform with little or no access to floodplain surfaces (Pollock et al., 2007; Demmer and Beschta, 2008).

Currently there is ongoing, watershed-wide habitat restoration work, conducted by NOAA’s ISEMP program in partnership with the National Park Service and the Bureau of Land Management, using beaver dam support structures to slow the transit of sediment and water through the channel. The goal of the restoration is to induce bed aggradation behind and adjacent to the beaver dam support structures which can increase aquatic habitat and riparian vegetation complexity and allow for better floodplain connectivity.
Figure 3.5 Longitudinal profile of Bridge Creek, shown in black, and tributaries, shown in all other colors. There appears to be small deviations apparent from the generally well-graded profiles but these are artifacts of the DEM resolution (cell resolution = 10 m) and do not represent knickpoints, aside from the knickpoint indicated on the top figure derived from a quaternary landslide deposit behind a small basalt gorge. Bridge Creek is shown in Black with major tributaries indicated with colored lines.
Within Bridge Creek, I used a combination of remote and field-based geomorphic mapping of alluvial terraces and fans immediately adjacent to the modern stream, grain-size analysis of these landforms, C-14 dating of terrace deposits, CAER from the dated deposits and modern alluvium and field-based geomorphic assessments to address five inter-related, basic and applied research objectives. Primarily I sought to provide age-constraints on the rate and style of Holocene development and abandonment (i.e., steady versus unsteady) of fill terraces of lower Bridge Creek. Specifically, I sought to determine if the most recent abandonment of terrace surfaces was large-scale, rapid, and coincided with the period of ubiquitous human perturbation in the region (i.e., within the last ~150 yrs) or if incision was spatially diverse and occurred at different times throughout lower Bridge Creek before human perturbation. Additionally, I wanted to compare CAER from fill terrace alluvium to the CAER derived from the modern alluvium to analyze the temporal and spatial variability of long-term sediment supply within Bridge Creek (at the time of writing there are no modern alluvium samples available for analysis). I also analyzed the difference in grain-size between fill terraces and alluvial fans to identify longitudinal differences in sediment source composition and relate longitudinal variability in grain-size to the River Styles classification scheme employed by ISEMP.

3.4.1.2 Methods

3.4.1.3 Geomorphic Mapping of Terraces and Alluvial Fans

The lower ~ 25 km of Bridge Creek is bounded, almost exclusively, by a series of fill terraces and alluvial fans that are difficult to differentiate by field observations alone. Because of this, and the need for accurate geomorphic context for geochronologic dating,
TCN sampling and grain-size analysis, I mapped the terraces and alluvial fans that occur within the lower ~ 25 km of Bridge Creek. High-resolution aerial imagery (20 cm) and lidar (1 m) were acquired in 2005 along the main channel of Bridge Creek by Watershed Sciences Inc.. Slope and curvature rasters were generated from the lidar DEMs using ArcGIS 10.0 Spatial Analyst tools. To identify and delineate the boundaries between fill terraces and alluvial fans within the alluvial valley, rasters and imagery were combined with stream-side photographs obtained during rapid geomorphic assessments conducted in June and July of 2011. Field verification was conducted in November, 2012.

Terraces and floodplains were initially mapped manually by identifying potential boundaries based on abrupt changes in slope or curvature using ArcGIS 10.0 Spatial Analyst. Typically, terraces on Bridge Creek exhibit a nearly flat surface profile perpendicular to the stream as opposed to alluvial fans which tend to have a low angle slope perpendicular to the modern stream channel extending towards the valley margins. Years of intensive agriculture within the alluvial valley have obscured these general trends in some locations. Drastic changes in vegetation observed from the aerial imagery were also used to identify boundaries. A small number of locations were identified where I was unable to definitively delineate a terrace or fan, and these locations were mapped in the field. Additionally, I used the TerEx tool (Stout and Belmont, 2013) to identify potential terrace surfaces. TerEx is a Python script that uses a lidar DEM with user-specified relief and extent thresholds to identify and map discrete terrace surfaces at a user-defined distance from the stream channel. I compared terraces identified using the TerEx tool to my own manually delineated terrace surfaces.
Terrace heights were extracted with the Zonal Statistics tool within the Spatial Analyst extension of ArcGIS 10.0 to obtain an average elevation of each terrace. I then calculated the height of each terrace relative to the stream layer by: 1) splitting the stream layer into 100 m segments 2) buffering the stream layer and converting it to a series of polygons 3) using Zonal Statistics to calculate the elevation of each stream layer polygon and 4) spatially joining the stream layer to the terrace layer and subtracting the stream from the terrace layer. After this I manually checked the accuracy of a sub-set of the terraces and discarded any terraces that were less than 30 cm above the stream layer.

3.4.1.4 Carbon-14 Dating and TCN Sampling of Terraces

In order to provide age-constraints on the rate and style of Holocene development and abandonment of fill terraces on lower Bridge Creek and to test the hypothesis that the most recent abandonment of terrace surfaces was large-scale, rapid, and coincided with the period of ubiquitous human perturbation in the region, it was necessary to collect $^{14}$C, OSL and TCN samples from terrace alluvium deposits. Geochronology sample locations were identified during the mapping and field-based geomorphic assessments. I chose three representative terraces of roughly equal height (~4 m) above the modern stream channel. The first terrace is located approximately six km upstream from the confluence of Bridge Creek with the John Day River (Figure 3.6). This is the farthest downstream location where terraces of this height occur. The second and third are located within the upper section of the low gradient, wide valley portion of the basin, approximately 14 and 15 km upstream of the confluence, respectively (Figure 3.6). A large swath of private property prevented a more spatially balanced selection of terraces. Despite this, the sample terraces are adequately representative of those occurring on Bridge Creek.
The TCN and OSL samples are restricted to a particular grain size fraction (~200-600um), a necessary depth below the current geomorphic surface (>1 m), and the absence of roots or other forms of bioturbation. These factors ultimately dictated the exact sample locations within the chosen terraces but for the most part samples were collected from exposed sections immediately adjacent to the axial stream at least 1.2 m below the terrace surfaces so as to minimize any post-deposition acquisition of additional $^{10}$Be from cosmic ray exposure. Carbon-14 (charcoal or bone) was sampled opportunistically within the terrace deposits because datable material does not occur ubiquitously throughout the exposed deposits. Figures 3.7-3.9 show the TCN and $^{14}$C sample locations within each terrace as well as a stratigraphic description of the terrace deposits. Initially multiple OSL samples were collected from each study terrace. Two were analyzed and the dates were considered unreliable after comparing to $^{14}$C dates from the same deposits. Bias in the OSL dates was likely due to partial bleaching, discussed in more detail in the results section. Additionally, the active channel alluvium was sampled for $^{10}$Be immediately downstream from each terrace and in additional locations along the main-stem of Bridge Creek, but results for those samples are not yet available at the time of writing.

3.4.1.5 Grain Size Analysis

To identify longitudinal differences in sediment source composition and to test if longitudinal variability in grain-size relates to the River Styles (RS) classification scheme currently employed by ISEMP, I conducted grain-size analysis of 31 fill terraces and 22 alluvial fans on lower Bridge Creek.
Figure 3.6: A portion of Bridge Creek, OR. Mapped terraces are shown in brown and alluvial fans in yellow. Inset boxes show the location of C14 dated terraces. Dates and CAER are also shown but will be discussed in the results section.
Figure 3.7: Stratigraphic column of Terrace 1 showing the C14 and TCN sample location in red. Note the alluvial fan material (Unit 5, B), terraces 1 and 2 share this unit with bed material above (Unit 4) but not terrace 3 (Figure 3.9). C14 ages will be discussed in the results section.
Figure 3.8: Stratigraphic column of Terrace 2 showing the C14 and TCN sample location in red. Note the alluvial fan material (unit 7) with bed material (unit 6) above. C14 ages will be discussed in the results section.
Figure 3.9: Stratigraphic column of Terrace 3 showing the C14 and TCN sample location in red. C14 ages will be discussed in the results section.
I was not attempting to identify a causal link between RS and grain-size. Rather, I sought to identify if the RS classification scheme was able to identify broad changes in the distribution of grain-sizes from alluvial terraces and fans along the longitudinal gradient of lower Bridge Creek. Geomorphic assessment of lower Bridge Creek indicates that there are three dominant RS (i.e., Forced Meander, Imposed Form Sinuous, Imposed Form Straight), one of which has two variants based on the level of lateral valley confinement (i.e., Imposed Form Sinuous Partly Confined and Imposed Form Sinuous Unconfined). Grain-size analysis of near-channel fill terraces and alluvial fans stratified by the four RS allows for comparison of grain-size distribution curves and associated metrics (e.g., D16, D50 and D84) among the different RS and between the two landforms (i.e., fill terraces and alluvial fans) within each RS. Table 3.2 shows the number of grain-size samples collected in each RS and Figure 3.10 illustrates the locations of grain-size samples relative to each RS.

Table 3.2: Grain-size samples from fill terraces and alluvial fans of each River Style occurring within lower Bridge Creek. The number of occurrences and percent length of each River Style is also shown. Since the Imposed Form Sinuous Partly Confined River Style covered the largest area it has the highest number of samples for both alluvial fans and terraces. The total channel length in the study area is 24.5 km.

<table>
<thead>
<tr>
<th>River Style/Valley Confinement</th>
<th>Landform</th>
<th>% Study Length</th>
<th>Individual Occurrences</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Terraces (n)</td>
<td>Fans (n)</td>
<td></td>
</tr>
<tr>
<td>Forced Meander Confined</td>
<td>3</td>
<td>2</td>
<td>10</td>
</tr>
<tr>
<td>Imposed Form Sinuous Prt.</td>
<td>16</td>
<td>14</td>
<td>45</td>
</tr>
<tr>
<td>Confined</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Imposed Form Sinuous Unconfined</td>
<td>6</td>
<td>4</td>
<td>33</td>
</tr>
<tr>
<td>Imposed Form Straight Prt.</td>
<td>6</td>
<td>2</td>
<td>12</td>
</tr>
<tr>
<td>Confined</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total (n)</td>
<td>31</td>
<td>22</td>
<td></td>
</tr>
</tbody>
</table>
Figure 3.10: Bridge Creek, Oregon grain-size sampling locator map. Sample locations are shown in gold relative to River Style reach types.
I intended to sample an equal number of terraces and fans per occurrence of each River Style with a sampling interval of no less than 1 km, but this was not possible in some locations due to logistical considerations (e.g., private property). Individual samples were intended to be representative of the entire landform and the landforms selected for sampling were intended to be representative of all such features within that occurrence of each River Style. Each sample consisted of a mix of grain-size fractions exposed in the near-channel outcrop, roughly in proportion to what was present. The mass of sample material collected ranged from 8 – 45 kg depending on the variability of the exposed grain-sizes, in an attempt to ensure that the largest grain did not exceed 1% of the total sample mass, though this was not always feasible. Figure 3.11 illustrates the field sampling method. Samples were collected with a shovel and five gallon buckets, and field sieved into the following categories: >64 mm, 31.5 - 64 mm, 16 - 31.5 mm, 2 – 16 mm, < 2 mm. The < 2mm size fraction was sampled, then sieved in the laboratory using a Ro-Tap© sieve shaker into 0.63 – 2 mm and < 0.63 mm fractions. All field samples were weighed with a hanging weight scale. A digital scale was used in the lab for the fine size fractions.

After examining the grain-size data, (n = 53: terraces = 31, alluvial fans = 22) two fans and two terraces were excluded as being unrepresentative of their respective River Styles. The excluded fans were considerably coarser than all other samples from the same River Style while the terraces were excluded for being much finer than all other terraces from their respective River Styles. Though these outliers may simply illustrate the local variability of sediment sources along Bridge Creek, they are not useful to consider in the context of establishing trends in grain-size distributions amongst
individual River Styles. Cumulative grain-size distribution curves were generated and D16, D50 and D84 metrics were calculated using standard procedures (Bunte and Abt, 2001). Additionally, two sample T-tests were performed to: 1) Compare the D50’s of all terraces versus all fans regardless of River Style; 2) Compare the D50’s and D84’s of alluvial fans within confined valley settings (e.g., Forced Meander and Imposed Form Straight) to the less confined valley settings (e.g., Imposed Form Sinuous); 3) Compare the D50’s and D84’s of terraces versus fans within each River Style.

**Figure 3.11:** Grain-size sampling methods. Coarse grained alluvial fan (Top left). Grain-size fractions sampled from the fan (Bottom left) (from top left: >64 mm, 31.5-64 mm, 16-31.5 mm, 2-16 mm, <2 mm). Top right. Low (~1m) inset terrace sample location. Bottom right. Grain-size fractions from the terrace sample. Center. Locator map of an imposed form straight partly confined River Style reach with the sample locations shown in yellow.
3.4.1.6 Results from Bridge Creek, OR

3.4.1.7 Geomorphic Mapping and Grain-Size Analysis of Fill Terraces and Alluvial Fans

Field verification confirmed the mapped landforms are accurate for their intended purpose of providing context for geochronologic dating, TCN sampling, grain-size analysis and further restoration planning. Terrace heights above the stream are mean values of each individual terrace polygon relative to a 100 m long stream segment with a mean stream elevation averaged over the 100 m. As such, point elevations may vary from the mean values. Additionally, the terrace shapefile contains surfaces that are most likely part of the active floodplain, in that they are on average below one meter in height above the stream and will probably be inundated during high flows. Terrace heights (n = 322) ranged from 0.31 – 25 m with a mean of 2.5 m and standard deviation of 2.8 m. Eight strath terraces that occurred in the upper confined valley section (i.e., ~16-24 km from the mouth) are the tallest surfaces ranging in height from 8 to 25 m. No strath terraces were mapped in lower Bridge Creek. This was expected given the relatively wider valley setting of lower Bridge Creek compared to the more confined upper section.

Alluvial fans occupy more than five times the area (e.g., 490 Ha) of the alluvial valley compared to fill terraces (e.g., 81 Ha). Fill terraces occur exclusively as a thin strip within 70 m of the main channel while the remainder of the alluvial valley is composed of alluvial fan deposits. Since we were unable to dig a trench perpendicular to Bridge Creek extending to the valley margins it is not possible to verify if any terrace alluvium deposits occur further from the current location of the main-stem but surface slope and curvature analysis suggests that this is not the case. Fill terraces exhibit a nearly flat surface slope perpendicular to the main-channel while alluvial fans display a constant,
low angle slope perpendicular to the main-stem extending to the valley margins. The large difference in area between the two landforms is an important finding given the difference in grain-size composition between fill terraces and alluvial fans.

In general, alluvial fans are composed of finer-grained material compared to fill terraces regardless of RS. The difference in grain-size distribution between fill terraces and alluvial fans grouped across all River Styles is shown in Figure 3.12. A Two-Sample T-test reinforced this comparison with significant differences between the D50’s and D84’s of all terraces and all fans (D50: P-value = .01, D84: P-value = 0.01).

**Figure 3.12:** Cumulative grain-size distribution curves comparing all alluvial fans to terraces across all four River Styles. The D50 and D84 are shown with solid black lines.
Figure 3.13: Cumulative grain-size distribution curves comparing alluvial fans from less confined valley settings (e.g., Forced Meander and Imposed Form Straight River Styles) to the less confined valley settings (e.g., Imposed Form Sinuous River Style). The D50 and D84 are shown with solid black lines.

Alluvial fans from confined valley settings are considerably coarser than alluvial fans from less confined valley settings within lower Bridge Creek. Figure 3.13 illustrates the magnitude of this difference. D84 comparison (P-value = 0.06) shows that in confined settings, where there is a direct coupling of hillslopes to the channel, the D84 is composed of very coarse gravel and cobble. In a less confined alluvial valley setting the D84 of alluvial fans is medium to coarse gravel. D50 comparison reveals a similar trend but with a less distinct difference (P-value = 0.13).

Interesting trends emerge when comparing grain-size distributions between fill terraces and alluvial fans within and amongst each RS. Figures 3.14 and 3.15 illustrate that within each RS, fill terraces exhibit a coarser grain-size distribution than alluvial fans.
but the magnitude of that difference is variable. The Imposed Form Sinuous, Partly Confined RS had the strongest statistical difference between alluvial fans and fill terraces (D50: P-value = .03, D84: P-value = .01). It is important to note this RS had the largest sample size for both alluvial fans and terraces and occurred most frequently within the study area (Table 3.2) (n = 13 fans, n = 16 terraces). There is also a difference in grain-size distribution when comparing landforms amongst RS. The IFSU alluvial fans and fill terraces are substantially finer than both landforms from the IFST RS and the terraces alone from the IFSU RS are considerably finer than all other terraces. The FM RS has the largest difference between the alluvial fans and fill terraces within all RS.

**Figure 3.14:** Cumulative grain-size distribution curves for each landform type within each River Style. The sample size (n) for each River Style is shown at right. Note that within each River Style the D50 of the alluvial fans are finer than the terraces.
Figure 3.15: Boxplots comparing terraces and alluvial fans A: D50, B: D16, C: D84 (Note y-axis differs between A,B,C). Data are summarized by River Styles reach type. Black arrows represent standard error.
3.4.1.8 Geomorphic Mapping and Grain-Size Analysis Discussion

The large difference in area (e.g., 5 X) between the fill terraces and alluvial fans that compose the alluvial valley of lower Bridge Creek suggests that historically alluvial fan inputs have overwhelmed the main-stem’s ability to transport tributary inputs and exerted a strong influence on the location of the main-stem within the alluvial valley. This is particularly evident in a ~10 km section of lower Bridge Creek where the valley is widest, alluvial fans are largest in area and appear to have forced the main-stem against the western edge of the alluvial valley. This area is shown in Figure 3.11 by the green centerline of the IFSU RS. The dominance of alluvial fans in this area is probably due to the erosive properties of the fine-grained volcanic tuffs and mudstone that serve as source material for the alluvial fans. As expected, because of the fine-grained source material, grain-size analysis reveals that alluvial fans have a finer grain-size distribution than fill terraces irrespective of RS (Figure 3.12).

Though the general trend is for fine-grained alluvial fan deposits, there are notable exceptions particularly within the Imposed Form Straight Partly Confined sections where valley confinement is high. Here, occurrences of intrusive basalt dikes and coarse grained conglomerate and sandstone deposits erode into relatively coarser size fractions, rather than the fine-grained ash and tuffs found in the less confined valley setting. This is also reflected in the relatively coarser terrace deposits within this River Style (Figures 3.14 and 3.15). Where the coarse grained alluvial fan deposits occur (i.e., primarily upstream and the lowest ~4 km) the valley width is confined which provides tighter coupling between hillslopes and the alluvial valley.
Through geomorphic mapping and grain-size analysis of fill terraces and alluvial fans I was able to identify longitudinal differences in sediment source composition that appears to be primarily driven by valley width and rock type. In the widest valley setting, where fine grained ash and mudstone deposits occur and alluvial fans attain their maximum area, deposits of both alluvial fans and fill terraces are finest. Within the confined and partly-confined valley settings both landforms are coarser. Valley width is a first order parameter in determining the suite of RS within a watershed and because of this the RS classification system currently employed in Bridge Creek appears to be geomorphically significant in terms of grain-size.

3.4.1.9 Carbon-14 Dating and TCN Sampling of Terraces

3.4.1.10 Results and Discussion

Table 3.3: Results from Carbon-14 dating and CAER from study terraces on lower Bridge Creek, OR
Initially we expected a wide range of ages within individual terraces and amongst the sample group of terraces in Bridge Creek. Our hypothesis was a sediment regime characterized by unsteady, spatially variable, alluviation and incision, indicative of a dynamic river that had undergone many periods of localized, temporary blockages from tributary fan inputs producing debris flows that temporarily changed local base level with subsequent incision to evacuate the debris flow material. A modern example of this dynamic can be found within the Pat’s Cabin reach (River km 5) where a large tributary debris flow within the last five years contributed a significant amount of coarse sediment to the main stem of Bridge Creek causing a local rise in base level. Similar (and larger) debris flows occurring throughout the Holocene could have provided the temporary increases in local base level that would be necessary to form the observed fill terraces. If this dynamic was prevalent we would expect a wide range of ages from the different fill terraces and within the individual terraces reflective of the spatially variable processes of deposition and subsequent incision. Our $^{14}$C dates do not show evidence of this dynamic.

Carbon-14 dating of terraces yielded a relatively narrow range of ages (70 – 430 years BP). This relatively small range (Table 3.3), especially in light of the fact that some C14 samples were collected from near to the top of the geomorphic surface (e.g., Terraces 1 and 2, Figures: 3.7 and 3.8) and others from the bottom of the exposure (e.g., Terrace 3, Figure 3.9) suggests rapid (i.e., likely on the order of years to decades) alluviation followed by large-scale (i.e., > 10 km within lower Bridge Creek) and rapid (i.e., years to decades) incision. It appears that the relatively young terraces were deposited rapidly (i.e., years to decades) within older established alluvial fan deposits.
We lack absolute age constraints from any alluvial fans. However, cross cutting relationships between the terraces and fans indicates that fans are relatively older. Though the previous statement is true, I also observed a few local examples of inter-fingering of presumed alluvial fan deposits with Bridge creek derived terrace alluvium (Figure 3.16). This indicates that in some locations alluvial fan deposition and main-stem Bridge Creek alluvial deposition occurred concurrently. Further, frequent occurrences of well-preserved Mazama ash deposits (Figure 3.16) in the alluvial fans show that many of the alluvial fans pre-date eruption of Mount Mazama (7 ka; Peacock, 1994). After extensive field observations and mapping of lower Bridge Creek I have never found a deposit of Mazama ash in a fluvial terrace. Further evidence of the age difference, is the fact that fill terrace deposits from Terraces 1 and 2 were deposited on top of pre-existing alluvial fan deposits (Terrace 1: Unit 5, Terrace 2: Unit 7) (Figures 3.7 and 3.8).

In addition to the C14 ages from the terraces, two other lines of evidence are consistent with rapid deposition of the terraces within existing alluvial fans. First, each study terrace possesses stratigraphic units that exhibit a lack of distinct bedding and poor fluvial sorting indicative of rapid deposition. Second, OSL samples were collected and analyzed from two of the sample terraces and were considered unreliable due to the effects of partial bleaching. Partial bleaching can occur in rapid transport/deposition environments when inadequate exposure to light causes only a partial removal of the radioactive charge from the sensitive electron traps within quartz grains in some or all of the sampled sand (Truelsen and Wallinga, 2003; Rittenour, 2008).
Figure 3.16: Terrace 2 (above) illustrates the spatial relationship of the dated inset terrace to the pre-existing alluvial fan deposits with a well-preserved outcrop of Mazama Ash outlined in yellow. Terrace 1 (below) shows another example of the inset terrace deposits intermixed with pre-existing, older fan deposits.

This results in an age over-estimation for the deposit. For example, my OSL samples returned ages an order of magnitude older (7.64 +/- 3.4 Ka and 3.68 +/- 0.63 Ka) than the ages recorded from the C14 samples from the same units. In this case, single-grain dating would be necessary to better constrain the timing of deposition. Single-grain OSL dating was not used here because of logistical constraints.

3.4.1.11 Why Rapid Deposition of Terraces and Subsequent Rapid Incision (i.e., can we blame humans?)

It is not possible to definitively show that rapid deposition and incision of terraces was caused by human land-use alone. Investigating drivers of rapid alluviation and
incision within arid climates in the western United States is an active subject of geomorphic inquiry (Cooke and Reeves, 1976; Bull, 1997; Hereford, 2002; Pederson et al., 2006). Other possible explanations of rapid deposition and incision have been linked to long-term shifts in climate with associated vegetative shifts (Rodbell et al., 1999; Anders et al., 2005; Fuller et al., 2009), glacial activity (Hallet et al., 1996; Rittenour et al., 2007; Moon et al., 2011), and changes in base level through tectonic activity (Holbrook and Schumm, 1999; Von Blanckenburg, 2006) or change in sea-level (Schumm, 1993; Merritts et al., 1994). Additional causes of incision could be related to internal drivers whereby a convexity threshold (Schumm and Hadley, 1957; Wescott, 1993) is reached forcing vertical incision.

Climate reconstruction based on tree-ring chronologies, extending back to 1705 for the southern Blue Mountains region of eastern Oregon (Heyerdahl et al., 2002) provides evidence of a relatively stable climate, in terms of mean annual precipitation, with high decadal variability. Within Bridge Creek there is no evidence of late Pleistocene or Holocene glaciation. Longitudinal profile analysis shows no sign of systematic trends in base level suggesting no substantial and recent tectonic activity, though the existence of old strath terraces suggests that at some point in the geologic past there was a prior vertical incision event not driven by anthropogenic stressors. Longitudinal profile analysis of Bridge Creek and major tributaries also reveals concave profiles which suggests that a convexity threshold is not a driver of incision. Though the resolution of the DEMs used for longitudinal analysis may be too coarse to identify finer scale trends in convexity.
It appears from our $^{14}$C dates that the period of rapid (i.e., years to decades) deposition and subsequent rapid (e.g., years to decades) incision could have coincided with the era of wide-spread Euro-American settlement and land-use, though the dates do not preclude the possibility that deposition and incision occurred before 1860. In the JDW and elsewhere in the region, the late 19th century brought about a host of anthropogenic changes in land-use including: large-scale livestock grazing, substantial reduction of previously abundant beaver populations, timber harvest, placer and dredge mining, and diversions/channelization for agriculture. In the early 1900’s a small town in the region, Shaniko, was one of the world’s largest exporting centers of wool (Sedell, 1994). Early photographs of the region show such a high density of sheep that at first look the hills appear to be covered in snow (Sedell, 1994).

High density livestock grazing is thought to have been a major driver in shifting the hillslope vegetative community from native bunchgrass to cheatgrass (Mack and Thompson, 1982). Cheatgrass has been shown to affect watershed hydrology by decreasing the recurrence interval of wildfires leaving bare slopes with decreased infiltration rates and increased runoff (Williams et al., 2011). Other research (Boxell and Drohan, 2009) indicates that regardless of wildfire, cheatgrass can change surface soil physical and hydrological properties to decrease infiltration and increase runoff. Other investigators have shown, through comparisons of historic and modern survey data (McIntosh et al., 2000) and modeling (Wondzell et al., 2007) that Euro-American settlement within the JDW dramatically changed riparian vegetation and channel conditions with subsequent detrimental effects to the type and distribution of aquatic habitat. These changes in land-use are consistent with higher magnitude peak flows and
lower base flows. Regardless of the driver(s) (i.e., anthropogenic or climate) hydrological analysis is consistent with this trend, though the reduction in base flows appears to be modest (Figure 3.4). Higher magnitude peak flow events coupled with the vegetative shift and loss of beaver dams (which slow the transit of the increased flow and entrained sediment through the channel network) increase the potential for geomorphically significant sediment mobilization and incision.

Field observations and grain-size analysis reveal that lower Bridge Creek has a high abundance of near channel sediment sources (e.g., fine-grained alluvial fans and fine-grained terraces), compared to the JDW as a whole. Coupled with generally low discharge, these readily accessible near channel sediment sources result in a transport-limited condition. Catchment-averaged erosion rates sampled from two terraces represent the highest (i.e., Terrace 2) and fourth highest (i.e., Terrace 3) erosion rates measured within the JDW (n = 9) (Figure 3.17). This indicates that long-term rates of sediment supply are relatively high compared to the rest of the JDW. This information coupled with the nature of these sediment sources (i.e., fine-grained ash deposits) and clay-rich soils with low infiltration capacity support the assertion that large storm events have the potential for significant scour and deposition.

Costa (1987) conducted a study of the hydraulics and basin characteristics of the twelve largest floods recorded, up to 1987, in the conterminous United States, including a 1956 flash flood generated from a convectional summer thunderstorm within a tributary to Bridge Creek, called Meyers Canyon. He estimated a maximum flood discharge of 1847 cms which is more than three orders of magnitude higher than base flow conditions.
**Figure 3.17:** Catchment-averaged erosion rates within the JDW (n = 9). Bridge Creek terraces samples are the highest and fourth highest of all samples indicating a relatively high, long-term supply of sediment.

Additionally, Peacock (1994) provides anecdotal evidence from interviews with elderly, life-long residents of Mitchell, OR, a small rural town located on Bridge Creek, and observations from the local cemetery, where there are numerous gravestones indicating residents killed by floods of the late 19th century, that there have been multiple occurrences of intense flash floods generated by convectional summer thunderstorms since Euro-American settlement. Flood events like these would be capable of scouring and depositing dramatic amounts of sediment within a short time (i.e., years to decades).

The size, abundance and position of alluvial fans on lower Bridge Creek provides evidence that pre-historically tributary alluvial fan inputs overwhelmed the main-stem’s transport capacity and exerted a strong influence on channel planform and the location of the main-stem within the alluvial valley (Figure 3.6). The relative age difference between
fans and terraces is consistent with this dynamic. The $^{14}$C ages from fill terraces indicate rapid deposition and large-scale incision. This suggests that during the modern era (e.g., 400 years ago to present) the role of alluvial fans on lateral forcing has diminished and the main-stem may be regaining control of channel planform and its location within the alluvial valley.

When viewed in its entirety my data suggests that anthropogenic changes in land-use that have changed patterns of runoff coupled with intense storm events may have been responsible for the rapid deposition and subsequent rapid vertical incision on lower Bridge Creek.

### 3.4.2 Granite-Boulder Creek, North Fork JDW

#### 3.4.2.1 Introduction

Granite-Boulder Creek (GBC) was selected as a study sub-watershed for four reasons. First, quartz-bearing rock types are well-represented within the watershed. Second, GBC experienced recent glacial erosion, which is relatively unique within the JDW. Third, there are existing sampling efforts underway by ISEMP and CHaMP assessing salmonid habitat and population abundance. And lastly, to achieve a spatially balanced sampling strategy within the broader JDW. Here I investigated how rates of hillslope erosion have changed over time as evidenced by comparing the CAER from an archived, dated terrace deposited before Euro-American settlement began, to the CAER from the modern alluvium post settlement.

Granite-Boulder Creek (GBC) is a 2nd order tributary to the Middle Fork John Day (MFJD) watershed, with a drainage area of 31 km$^2$ (Table 3.1) The bedrock geology underlying the watershed is quite different than the lower elevation portions of the JDW...
(e.g., Bridge Creek) in that there are very few exposures of volcanic extrusives. The upper reaches of GBC are primarily composed of the Jurassic, resistant, quartz-bearing igneous intrusives: granodiorite and quartz-diorite. The middle and lower-sections of the watershed are primarily composed of Paleozoic and Mesozoic, less-resistant, marine sedimentary argillite and sandstone outcrops with two small occurrences of Paleozoic and Mesozoic serpentanite (Figure 3.18).

Unlike the lower elevation portions of the JDW upper GBC was glaciated during the Quaternary. There are abundant glacial deposits in this area, including boulder-sized glacial erratics. Glaciation does not appear to have been as severe compared to other portions of the CRB (e.g., North Cascades) but still left its characteristic imprint on the watershed, carving broad U-shaped valleys and a stair-step profile into the upper portion of the stream longitudinal profile (Figure 3.18). Below the zone of glaciation the stream maintains a high gradient until transitioning into a broad alluvial setting. Within the steep section, valley confinement is high with direct input of coarse sediment and wood from the steep hillslopes. Direct hillslope inputs appear to be the dominant sediment source within the steep upper reaches. Cobble to boulder-sized clasts and large woody debris (LWD) exert a strong influence on planform and gradient locally. There is very little storage within this zone and no floodplain. The stream is primarily single-thread with very low sinuosity until the alluvial reach where gradient is lower and valley setting is unconfined in the 2 km before the confluence with the Middle Fork John Day River. Here a floodplain is present with evidence of past mining activity throughout the floodplain.
Figure 3.18: Longitudinal profile of Granite-Boulder Creek (above) with knickpoints illustrated (crosses). Below the same knickpoints are shown in blue superimposed on the bedrock geology. Note all knickpoints occur within the glacial deposits. The longitudinal profile downstream of the zone of glaciation exhibits a ‘normal’ concave-up profile.
Mining intensity in the alluvial valley of GBC appears to have been high based on the abundant, large (1-3 m tall) piles of alluvium scattered throughout the lower alluvial valley. Placer mining for gold began in the upper Middle Fork John Day region in 1862, followed by dredge mining beginning in the early 20th century (Sedell, 1994). This activity disturbed the riparian area and alluvial valley of GBC by channelization and excavation of alluvium. This degraded the aquatic habitat in the mining area, dramatically altering the stream’s flow path and liberating and redistributing fine sediment originally deposited in the floodplain (Sedell, 1994; McIntosh et al., 2000).

**3.4.2.2 Methods: TCN Sampling and OSL Dating of GBC Terrace**

Through field reconnaissance and GIS analysis I identified a small (1.5 m) terrace suitable for OSL dating located in the steep, confined section upstream from the alluvial valley where mining took place (Figure 3.19). The OSL sample was collected ~80 cm below the top of the geomorphic surface in a sand lens, which was also sampled for TCN. It would have been optimal to sample ~ 1.2 m below the geomorphic surface to assure that the TCN sample did not continue to receive cosmic ray exposure post-deposition but this was not possible due to the height of the terrace and the grain-size necessary for TCN. In addition, modern alluvium was sampled within the lower, alluvial section of GBC (Figure 3.19).

**3.4.2.3 Results and Discussion**

The OSL age returned from the terrace deposit establishes the time of deposition as prior to Euro-American settlement and mining (Table 3.4). The high overdispersion value associated with the OSL age is relatively common from Holocene aged terrace
Figure 3.19: TCN and OSL sample locations from Granite-Boulder Creek shown above.

deposits and represents partial bleaching during transport prior to deposition but does not mean that the age is unreliable. Catchment-averaged erosion rates from the terrace deposit versus the modern alluvium are essentially the same, within one standard deviation.

Calculating a significant hillslope stripping depth for CAER

We developed the following method to determine if the magnitude of a perturbation to the sediment generation and routing system within GBC was significant
Table 3.4: OSL age and CAER from terrace deposit and modern alluvium, Granite-Boulder Creek.

<table>
<thead>
<tr>
<th>Sample Location</th>
<th>Sample Depth</th>
<th>OSL Age (ka)</th>
<th>Overdispersion (%)</th>
<th>Scaling Factor</th>
<th>Be$^{10}$ Production Rate (atoms/g/yr)</th>
<th>TCN Derived CAER (mm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Terrace</td>
<td>80 cm</td>
<td>0.31 +/- 0.06</td>
<td>44.6 +/- 12.8</td>
<td>4.58</td>
<td>20.63</td>
<td>0.054 +/- 0.004</td>
</tr>
<tr>
<td>Modern Alluvium</td>
<td>Surface</td>
<td>NA</td>
<td>NA</td>
<td>4.32</td>
<td>19.43</td>
<td>0.048 +/- 0.004</td>
</tr>
</tbody>
</table>

1. Error on age is 2-sigma standard error.
2. Overdispersion represents scatter in Do, equivalent dose beyond calculated uncertainties in data. OD > 20% is considered significant.
3. 10Be production rate scaling factor generated using Stone (2000). Shielding factors not included in this table but were accounted for.
4. Sea level high latitude (SLHL) Be production rate of 4.5 at/g/yr (half-life 1.36 My) multiplied by scaling factor.
5. Error represents 1 Std. Dev. due to AMS measurement uncertainty. See appendix A for more information about all parameters included in the calculation of CAER.

enough to be revealed by comparison of CAER from the active channel alluvium to a dated terrace alluvium deposit. We determined a depth of hillslope stripping necessary to observe a noticeable difference in CAERs using the following two equations. Following the approach of Brown et al., (1998) and Belmont et al., (2007) equation 3.1 allows us to calculate the site specific Be$^{10}$ production rate at any depth below the geomorphic surface.

$$P_z = P_o e^{-z/\Lambda}$$ \hspace{1cm} (3.1)

where $P_z$ is the production rate at depth, $z$ in cm, $P_o$ is the total production rate at the surface, $\Lambda$ is the attenuation path length (160 g cm$^{-2}$) and $\rho$ is either a soil density (1.5 g cm$^{-3}$), sapprolite density (1.8 – 2.2 g cm$^{-3}$) or rock density (2.5 g cm$^{-3}$). We can then use equation 3.2 with the production rates we calculated at 5 cm increment depths to determine the Be$^{10}$ concentration within a hillslope depth profile extending to 160 cm from the surface. Equation 3.2 is a modified version of an equation developed by Dunai (2010).
\[ C_z = \frac{P(z)}{\lambda + \rho \varepsilon / \Lambda} \]  

where \( C_z \) is equal to the \(^{10}\text{Be} \) concentration at depth, \( z \) in cm, \( \lambda \) is the \(^{10}\text{Be} \) decay constant and \( \varepsilon \) is the surface erosion rate. Using \( C_z \) from equation 3.2, we calculated erosion rates from depth to estimate the minimum hillslope stripping depth necessary to observe a quantifiable difference in CAERs. Figure 3.20 shows \(^{10}\text{Be} \) concentrations as a function of depth.

**Figure 3.20:** Modeled erosion rates as a function of depth below a geomorphic surface. Compared to the surface at 20 cm (shown in red) there is an observable difference in erosion rates when uncertainty is considered. As \(^{10}\text{Be} \) production attenuates rapidly with depth and density, at 30 cm there is a drastic reduction in \(^{10}\text{Be} \) production, thus a much higher erosion rate compared to the surface.

The fact that the CAERs from the terrace and modern alluvium are essentially the same is evidence of relatively steady erosion averaged over the millennial time-scale over
which CAERs integrate. This does not necessarily mean that there wasn’t a temporary increase in sediment yield, to the stream, associated with mining activity. What can be inferred from the equivalent rates is that there was not a significant hillslope stripping event (e.g., hillslope stripping <20 – 25 cm) associated with mining activity. Previous research has shown that comparison of CAERs alone can capture changes in sediment yield associated with anthropogenic activity but the scale of disturbance has to be substantial for rates to reflect this change (Brown et al., 1998; Hewawasam et al., 2003).

Based on the comparisons of the CAERs it appears that over the millennial-scale anthropogenic activity has not profoundly altered the sediment generation and routing within Granite-Boulder Creek.

3.4.3 East-Branch North-Fork Cable Creek

3.4.3.1 Introduction

East-Branch North Fork Cable Creek was included as a study sub-watershed primarily because of the recent (i.e., 1996), high intensity wildfire that when coupled with subsequent storm events, generated debris flows that are still visible today in the alluvial valley. Secondly, the sub-watershed was included to establish a spatially balanced sampling strategy within the broader JDW. Finally, its inclusion was due to ongoing sampling efforts by ISEMP and CHaMP assessing salmonid habitat and population abundance. Here, research questions were: 1) how have rates of sediment yield changed over time as evidenced by CAER measured in a dated terrace deposit versus modern alluvium? 2) Will the temporary, substantial increase in sediment yield associated with the Tower Fire significantly affect CAER?
This small (i.e., drainage area: 11.5 km\(^2\)) sub-watershed is a 2\(^{\text{nd}}\) order tributary to North-Fork Cable Creek which drains into Camas Creek before entering the North-Fork John Day River (Figure 3.1). The sub-watershed is primarily underlain by easily weathered Eocene, Caldera-fill tuff of the Tower Mountain Caldera (Figure 3.21). There are also intrusions of more resistant porphyrctic and aphyritic dacite and rhyolite masses, which contribute the quartz necessary for cosmogenic analysis.

Quaternary landslides are mapped within the sub-watershed (Figure 3.21). Deep seated landslides comprising a significant fraction of a watershed could bias CAER if the landslide material was not exposed to cosmic rays (Binnie et al., 2006). This will be discussed further but appears to have not influenced the CAER in this watershed. Field surveys did not find evidence of these mapped landslides either because they were incorrectly identified during mapping or because they were old enough to no longer be visible on the landscape. Either way, CAERs from the sub-watershed do not seem to be affected by dilution of \(^{10}\)Be poor quartz. Longitudinal profile analysis of the stream does not reveal any systematic trends and exhibits a relatively concave up profile.

The largest documented fire on the Umatilla National Forest burned through the upper reaches of the Cable Creek watershed including the East-Branch of North-Fork Cable Creek (EBNFC) from mid-August through early September, 1996 (Powell and Erickson, 1997). The Tower Fire (20,640 ha) was one of four large fires in the upper John Day River basin and much of the area burned at moderate to high severity. Intense spring storms in 1997 and 1998 triggered large floods, landslides, and debris flows that affected streams within and downstream of the burned area (Howell, 2006), including EBNFC.
Within EBNFC, evidence of these debris flows are still visible 17 years later as broad, unsorted gravel and cobbles sheets deposited in the alluvial valley, as deep as 1.5-2 m in some places (Figure 3.21). Forest regrowth has been slow with mostly bare slopes in the upper study area and dense conifer regrowth in the lower basin. The more mobile fine sediment fraction has primarily been evacuated from the upper reaches and redeposited in a lower gradient, broader valley setting downstream of the gravel sheets. Here the flow is shallow, sub-surface and diffuse forming a marshy section for ~200 m. In addition, there are abundant, charred Large Woody Debris (LWD) jams emplaced during post-fire debris flows located above the height of the modern channel. Alluvial fans are abundant in the upper reaches with the dissected toes composed of loose, unsorted angular gravels and cobbles with no fluvial sorting apparent, implying rapid deposition in a debris flow.

At some point after the initial period of debris flow deposition in the alluvial valley the main-stem vertically incised one to two meters. This incision has reached bedrock in many locations within the upper reaches of EBNFC. Tributary response to main-stem incision is visible in many alluvial fans where the lower portions of the fans are vertically incised to match the new base level. Additionally, the toes of hillslopes and tributary fans adjacent to the main-stem are oversteepened resulting from the main-stem incision and have failed in many places by shallow slumping. This is most prevalent where the toes of hillslopes are composed of the loosely-packed, fine-grained tuff. These events are small-scale and do not appear to be actively contributing large amounts of material into the channel.
Figure 3.21: (Center) Bedrock geology of EBNFC, OR. (Upper Left Photo) Terrace sample location. C14 sample from Unit 1: A mix of moderately sorted silty-sand with abundant charcoal scattered throughout. Interpreted as overbank deposits with evidence of past fire. Unit 2 is a mix of sand and sub-angular gravels, interpreted as bed material. TCN sample came from this unit. (Upper Right Photo) Debris flow remnants 1.4m tall from upper watershed being reworked by modern stream. (Bottom Right Photo) Gravel sheets covering pre-Tower fire soil horizon. (Bottom Left Photo) Sample location of modern alluvium TCN sample.

3.4.3.2 Methods: TCN Sampling and C14 Dating of Terrace

Through field reconnaissance and GIS analysis I located a suitable 1.5 m alluvial terrace located ~ 0.5 km from the confluence with North-Fork Cable Creek (NFC) and found multiple pieces of charcoal for C14 analysis ~ 0.8 m from the terrace surface at the bottom of a layer of silty-sand overbank deposits (Figure 3.21). Only one piece of charcoal was sent to AMS facilities for analysis. I also sampled the layer of bed material composed of sand and small sub-angular gravels directly beneath the overbank deposits for TCN analysis. Additionally, I sampled a sandy-gravel bar for TCN located upstream
from the confluence of EBNFC and NFC. Finally, I completed a rapid geomorphic assessment of EBNFC (July 2012) and NFC (June 2011).

3.4.3.3 Results and Discussion

The C14 date on the terrace deposit (1880 years BP) was older than expected but seems reasonable given the very slow CAERs from both the terrace and modern alluvium, which were essentially the same (Table 3.5). Regardless of the sediment pulse associated with the Tower Fire, I originally hypothesized that CAER from both the terrace and modern alluvium would be high based on the erosive tuff that comprises the majority of the basin’s bedrock. To the contrary, both CAERs were in fact the second and third lowest of all samples collected in the JDW (Figure 3.22). This is a compelling case for how, in many locations, rock type cannot be used to infer rates of long-term (i.e., millennial-scale) sediment yield.

Table 3.5: C14 age and CAER from terrace deposit and modern alluvium, East-Branch North-Fork Cable Creek

<table>
<thead>
<tr>
<th>Sample Location</th>
<th>Sample Depth(s)</th>
<th>C14 Convent.(^1) Age (yrs BP</th>
<th>2-Sigma Calib.(^2)</th>
<th>Scailing (^3) Factor</th>
<th>Be(^{10}) Production Rate(^4) (atoms/g/yr)</th>
<th>Be(^{10}) Derived CAER (^5) (mm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Terrace</td>
<td>C14: 80 cm</td>
<td>1880 +/- 30 BP</td>
<td>AD 60 - 180</td>
<td>3.73</td>
<td>16.78</td>
<td>0.029 +/- 0.002</td>
</tr>
<tr>
<td></td>
<td>BE10: 120 cm</td>
<td></td>
<td>AD 190 - 210</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Modern Alluvium</td>
<td>Surface</td>
<td>NA</td>
<td>NA</td>
<td>3.73</td>
<td>16.78</td>
<td>0.027 +/- 0.002</td>
</tr>
</tbody>
</table>

\(^1\) Conventional age represents measured radiocarbon age corrected for isotopic fractionation, calculated using the delta \(^{13}\)C.
\(^2\) Error on age is 1 relative Std Dev based on combined uncertainty from sample, background, and modern reference standards.
\(^3\) 2-Sigma Calibration calculated from the conventional radiocarbon age.
\(^4\) \(^{10}\)Be production rate scaling factor generated using Stone (2000). Shielding factors not included in this table but were accounted for.
\(^5\) Sea-level high latitude (SLHL) base production rate of 4.5 at/g/yr (half-life 1.36 My) multiplied by scaling factor.
\(^5\) Error represents 1-Std. Dev. due to AMS measurement uncertainty.

See appendix A for more information about all parameters included in the calculation of CAER.
Figure 3.22: Catchment-averaged erosion rates within the JDW (n = 9). EBNFC samples are the second and third lowest of all samples indicating a steady and relatively low, long-term supply of sediment despite the history of substantial pulses of sediment associated with fire.

What can the CAERs tell us about the fire-related sediment pulse and past fire regimes?

Given the field evidence of a large pulse of sediment delivered to the stream after the Tower Fire and the evidence of past fires in the dated terrace deposit it seems that fire related temporary increases in sediment yield have been a part of the sediment regime within this watershed for at least the last ~ 2,000 years. But it appears that CAERs average over a sufficiently long time period to average out post-fire erosion pulses. This was unexpected given the depth of the post-Tower fire deposits in the modern alluvial valley (i.e., 1 – 2 m). This depth implies that hillslopes could have been stripped enough (e.g., 20-25 cm) to provide $^{10}$Be poor material which translates into high CAER. In some locations it is possible that post-fire hillslope erosion exceeded 20 - 25 cm in depth.
However, either these deeply scoured areas were very rare or they were sufficiently averaged out in time and space. The equivalency of the CAERs implies that over the millennial-scale, sediment generation and delivery within EBNFC is relatively steady, and low, despite occasional, substantial pulses of sediment after wildfires.

*Are the CAERs representative of watershed-wide sediment generation and delivery given the lack of quartz-bearing rock types within EBNFC?*

Quartz bearing rock types are scarce within the watershed and may be more erosion resistant than the tuff. Thus CAERs recorded from the quartz-bearing rock types could be lower than erosion rates from the tuff alone. CAERs record the average rate of sediment generation from the quartz-bearing rock types alone and because of this the CAER do not strictly apply to the entire watershed. Regardless of this, the relationship between the CAERs derived from the terrace alluvium compared to the active channel alluvium is accurate and records relatively steady sediment generation and delivery when averaged over the millennial-scale.

*Why are the rates so low?*

Cable Creek watershed is geographically located in a low-relief, tectonically inactive landscape generated from a series of Tertiary lava flows. Thus, the low rates are probably reflective of this lack of tectonic activity which has been shown to be a primary driver of rapid long-term erosion (Riebe and Kirchner, 2001a; Von Blanckenburg, 2006).
3.5 Wenatchee and Skykomish Watersheds

3.5.1 Introduction

The Wenatchee and Skykomish watersheds in the north Cascade Range of western Washington experienced a profound series of glaciations during the Pleistocene and Holocene epochs (Porter, 1978). The last glaciation ended ~ 14,000 years ago (Collins et al., 2002). Glaciation left a considerable imprint on watershed morphology (e.g., steep hillslopes with broad U-shaped valleys). Sediment generated during the period of glacial erosion is still present within these watersheds underlying alluvial valleys and occurring as outwash terraces (5 – 30 m in height) located along the modern channels of both watersheds.

The recent glacial history of these watersheds represents a natural experiment to address study objectives of understanding how long-term rates of sediment yield vary spatially and temporally throughout the CRB. Specifically we seek to quantify how rates of sediment yield have changed over time in a relatively undisturbed (by humans) watershed with a recent geologic history of glaciation, as evidenced by $^{10}$Be concentrations from archived, glacial outwash terrace deposits and active channel alluvium. Additionally this natural experiment allows us to illustrate the importance of geomorphic consideration when using $^{10}$Be concentrations from the active channel alluvium to calculate CAER (Granger and Kirchner, 1996; Niemi et al., 2005) and long-term rates of sediment yield.

The Wenatchee River (WRW) and Skykomish River (SRW) watersheds are located in North-central Washington and drain a portion of the North Cascade Range. The WRW covers 3439 km$^2$ and contains six primary sub-watersheds while the SRW has
a drainage area of 2160 km$^2$ with two major sub-watersheds (Figure 3.23). The mainstem stream length of the WRW is 85 km and the Skykomish is 47 km. Within the WRW, elevation ranges from 2856 m in the headwaters to 185 m at the confluence with the Columbia River and in the Skykomish elevation ranges from 2429 to 21 m where it joins the Snohomish River (Table 3.6).

**Figure 3.23**: DEM of the Wenatchee and Skykomish watersheds located in central Washington. Study sub-watersheds are highlighted in yellow (TCN samples). Sample locations from this research are shown with yellow dots. Moon et al., (2011) TCN samples are shown in red and gage locations are shown with red squares.
3.5.2 Geology

There are five primary bedrock units mapped within the WRW (Tabor, 2005). Precambrian Swakane Biotite Gneiss can be found in the northeastern portion of the watershed with Precambrian Easton Schist composing the southwestern portion. The rest of the bedrock of the WRW is composed of a mix of Mesozoic Ingalls Tectonic Complex (serpentine and serpentinized peridotite, gabbro and diabase), the Mesozoic Mount Stuart Batholith (biotite, hornblende, quartz diorite, and granodiorite), Tertiary nonmarine sedimentary and volcanic rocks, and Miocene basalt flows constituting a part of the Columbia River Basalt Group (Tabor, 2005). The Tertiary, right-lateral strike-slip Straight Creek fault complex is a major Pacific NW structure that extends from central WA into Canada (Figure 3.24) and occurs within the SRW. The fault mostly separates the easily erodible un-metamorphosed and low-grade metamorphosed oceanic rocks on the west, (e.g., Mesozoic marine meta-sedimentary rocks) low elevation portion of the SRW from medium to high-grade metamorphic rocks on the east, (e.g., banded gneiss) high elevation portion of the watershed. There are also major, resistant, Neogene batholith intrusions composed of granodiorite and the Cretaceous Mt. Stuart batholith composed of tonalite that underlie the mid to upper elevation portions of the watershed (Tabor et al., 1993).

Glacial moraines, outwash and lacustrine deposits combined with loess records provide evidence of the three most recent Pleistocene glaciations that have occurred within the study area. Presumably, earlier glaciations also sculpted the landscape but the most recent glaciations (~ 14 ka) had profound impacts on the current morphology of both watersheds. Within the WRW many portions of the alluvial valley are underlain
with glacial outwash deposits 50 m thick and at least 90 m thick near the confluence of Icicle Creek with the main-stem Wenatchee River. These glacial outwash deposits also occur as sand and gravel terraces along portions of the Wenatchee and Skykomish Rivers (NPCC, 2004) (Figure 3.24) and are important sources of sediment for the modern rivers.

3.5.3 Climate/Hydrology

Because of the orographic effect, there is a factor-of-ten precipitation gradient (Moon et al., 2011) within the Cascades based on topography and proximity to the range crest. Within the WRW, precipitation is greatest near the crest with over 380 cm occurring annually in some locations. Snow depth in the winter months along the Cascade Range crest varies from 3 - 7.5 m in depth. Air temperature in the winter at high elevations averages -3 to 4 °C and 15 to 27 °C in summer (NRCS SNOTEL Site: 21B01S, period of record: 1981-current (gage location shown on Figure 3.23)). In contrast, lower elevations of the WRW are semi-arid with the mouth of the WRW experiencing 23 cm of precipitation annually and summer temperatures of 38°C. Within the SRW, precipitation is also greatest near the crest, but because the SRW is on the windward side of the Cascade Range low elevation portions of the watershed still receive 281 cm annually.

Like much of the CRB, the WRW is primarily a snowmelt driven system producing 80% of the total runoff from the watershed. Peak flows typically correspond to timing of peak snowmelt in the late spring and early summer and range from 254 to 134 cms (USGS gage: 12457000, Wenatchee R. at Plain WA, period of record: 1910-current current (gage location shown on Figure 3.23)).
Figure 3.24: A simplified representation of the bedrock geology of the WRW and SRW, WA. Note the lack of extrusives and dominance of resistant intrusive and metamorphic lithologies. Also shown within the upper portion of the SRW is the Evergreen Fault, the eastern boundary of the Straight Creek Fault Complex. There are abundant glacial outwash terraces found within the lower portions of both watersheds adjacent to the modern stream. Geologic layers were collected from the state-wide digital GIS portal and categorized into the categories shown above. For more information on the sources of data and the classification scheme see Appendix C.
Table 3.6 Watershed characteristics of study sub-watersheds within the Wenatchee and Skykomish watersheds.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Wenatchee River</th>
<th>Nason Creek</th>
<th>Chiwawa River</th>
<th>Skykomish River</th>
</tr>
</thead>
<tbody>
<tr>
<td>Drainage Area (km²)</td>
<td>3439</td>
<td>285</td>
<td>474</td>
<td>2160</td>
</tr>
<tr>
<td>Stream Length (km)</td>
<td>85</td>
<td>36</td>
<td>60</td>
<td>47</td>
</tr>
<tr>
<td>Max Elevation (m)</td>
<td>2856</td>
<td>2465</td>
<td>2742</td>
<td>2429</td>
</tr>
<tr>
<td>Min Elevation (m)</td>
<td>185</td>
<td>601</td>
<td>601</td>
<td>21</td>
</tr>
<tr>
<td>Relief (m)</td>
<td>2671</td>
<td>1864</td>
<td>2141</td>
<td>2408</td>
</tr>
<tr>
<td>Mean Slope (degrees)</td>
<td>24.3</td>
<td>24</td>
<td>25.3</td>
<td>24.5</td>
</tr>
<tr>
<td>Mean Precipitation (cm/yr)¹</td>
<td>130.1</td>
<td>167.2</td>
<td>134</td>
<td>281.2</td>
</tr>
</tbody>
</table>

¹ Precipitation data was collected from the PRISM Climate Group and represents the past 30 year annual averages.

Lowest flows are typically in September and range from 8 to 46 cms (NPCC, 2004). Flow duration curve analysis classified by decade does not show any systematic trend, but does indicate a moderate amount of decadal variability (Figure 3.25).

Because of the precipitation gradient and ongoing glacial retreat, runoff patterns within the SRW are more complex. The SRW experiences two distinct runoff pulses with peak flows occurring during spring runoff (i.e., May and June) and another peak occurring during late fall (i.e., October and November) due to rain events. Pelto (2011) conducted hydrologic analysis comparing the period, 1950 to 1985 to 1985 to 2009 (USGS gage: 12134500, period of record: 1950 - 2009) and found that during the recent period summer streamflow (i.e., July – September) has declined by 26%, spring runoff (i.e., April – June) declined 6% and winter runoff (i.e., November – March) has increased 10%. He attributes the large decline in summer low flows to a significant reduction in glacial runoff as the principal alpine glaciers in the watershed have declined in area by 10, 35, 60 and 90%.
**Figure 3.25**: Flow duration curves organized by decade for the Wenatchee River, WA. Note the lack of any systematic trends, and the evidence of natural, decadal variability (USGS gage 12457000 period of record: 1910 - current) Gage location shown on Figure 3.23.

### 3.5.4 Human Perturbations

The majority of the WRW is federally owned (80 %), 99 % of which is managed by the Wenatchee National Forest. Most federal land is designated as wilderness and as such, much of the aquatic habitat of the upper WRW is in good condition. Private land accounts for 18.2 % of the total area and is primarily found in lower elevations within the alluvial valley. Much of this land is irrigated for orchard production. Only 1.4 % of the total area is owned by state, county, or municipalities (Bugert, 1997). Hydrologic diversions from the Wenatchee and its tributaries have taken place since 1891, primarily for orchard irrigation. Two small dams remain in operation (Tumwater Canyon Dam...
since 1907 and Dryden Dam since ~1904) on the Wenatchee River but are small enough that they do not have a substantial impact on streamflow.

Similar to the WRW, the majority of the SRW is federally owned > 70% and managed by the USFS. Most of this land occurs in the forested upper reaches of the watershed. The Washington Department of Natural Resources owns and manages the majority of the remaining watershed area, which comprises the low elevation portions of the watershed. A minimal amount of development is present (4%) and mainly occurs at low elevations. The lowlands have been heavily impacted by the history of LWD removal that occurred throughout the Puget Sound in the late 19th and early 20th centuries. Collins et al. (2002) estimates current wood abundance in the Skykomish as one or two orders of magnitude less than pre-Euro-American settlement. Changes to wood abundance and size have substantially changed the morphology and abundance of aquatic habitat across all spatial scales (e.g., from channel unit to entire alluvial valley). Collins et al. (2002) also found that the Skykomish historically had more abundant and deeper pools and suggests that LWD aggregates were responsible for maintaining a dynamic anastomosing river planform with good access to its floodplain.

3.5.5 Longitudinal Profile Analysis

Longitudinal profile analysis of the Wenatchee River and tributaries shows two distinct types of profiles. Upper elevation tributaries experienced substantial Pleistocene and Holocene glaciation and the abundant knickpoints create a ‘stair-step’ morphology that reflects this glacial history (Figure 3.26). For example, the Chiwawa River, one of the study sub-watersheds in the WRW, shown in blue in Figure 3.26, exhibits a
Figure 3.26: Longitudinal Profiles of the Wenatchee River and major tributaries (Top). Location of tributaries (Bottom Left). Study sub-watersheds in red (Bottom Right). Longitudinal profiles appear to be decoupled from tectonic activity instead they reflect the Pleistocene and Holocene glaciation with prominent knickpoints reflective of tributary and mainstem confluences.
prominent knickpoint towards the bottom of the profile that corresponds to the terminal moraine of a valley glacier.

Below this knickpoint the river maintains a continuous gradient to the confluence with the Wenatchee River. The upper knickpoint occurs at a major tributary confluence and is likely an artifact of a tributary glacier’s confluence with the main-valley glacier. The other study sub-watershed in the WRW, Nason Creek, shown in dark red in Figure 3.26 exhibits a similar profile to the Chiwawa River with knickpoints occurring primarily at tributary junctions. The other type of profile evident in the WRW tributaries has a well-graded, mostly knickpoint free, concave-up morphology. These occur in low elevation tributaries below Tumwater canyon that drain low elevation hillslopes that did not experience profound glaciation during the Pleistocene and Holocene glaciation.

Longitudinal profile analysis from the Skykomish River and major tributaries reveal similar trends to the upper, glaciated WRW (Figure 3.27). Abundant, prominent knickpoints occur in the upper portions of the main-stem and all major tributaries. Again, knickpoint location appears to coincide with tributary, main-stem glacial confluences.

3.5.6 Methods

Here, I leveraged an existing data set of CAER derived from $^{10}$Be concentrations from active channel alluvium collected by Moon et al., (2011) within the SRW and WRW. My sampling strategy consisted of locating glacial outwash terrace deposits within the study area functioning as sediment sources to the modern channel. Consideration was given to the location of these terraces relative to Moon et al., (2011) sample locations. Specifically, I sampled glacial outwash terraces for $^{10}$Be concentrations
Figure 3.27: Longitudinal Profiles of the Skykomish River and major tributaries (Top). Location of tributaries (Bottom Left). Bedrock geology (Bottom Right). Longitudinal profiles appear to be decoupled from tectonic activity instead they reflect the Pleistocene and Holocene glaciation with prominent knickpoints reflective of tributary and mainstem confluences.

upstream of Moon et al., (2011) active channel alluvium samples. Sample terrace heights ranged from 5 – 20 m. Sample locations are shown in Figure 3.23 relative to Moon’s sample from modern alluvium. Nason Creek and the Chiwawa River were selected within the WRW due to the prevalence of glacial outwash terraces functioning as sediment sources. The SRW was included because of the abundance of existing CAER data and the prevalence of significant (20 -30 m tall) glacial outwash deposits. OSL samples were also
collected from a sub-set of the terraces but were not analyzed because I determined that age constraints from the last glaciation (14 ka) were adequate for my analysis.

3.5.7 Results and Discussion

Dilution of CAERs with glacial sediments within the WRW and SRW, WA

Glacial erosion has been shown to be considerably more rapid than erosion dominated by fluvial processes alone (Gurnell et al., 1996; Hallet et al., 1996). This is generally the case except in regions with extremely high rates of tectonic uplift (e.g., Himalayas and Taiwan). In these locations rates of river incision are high enough to match rates of glacial erosion (Koppes and Montgomery, 2009). Because of the recent history of glaciation in the WRW and SRW, $^{10}$Be concentrations from glacial outwash terrace alluvium are generally much lower (i.e., indicating faster erosion) than $^{10}$Be concentrations from active channel alluvium alone (Table 3.7 and Figure 3.28). Within this section I chose to discuss results in terms of $^{10}$Be concentrations instead of catchment-average erosion rates because two key assumptions of the CAER method appear to have been violated (Dunai, 2010). Specifically, it is possible that recent glacial erosion violates the steady-state erosion assumption and field observations suggested that a substantial portion of channel alluvium is currently being sourced from near-channel glacial deposits rather than hillslope erosion.

The discrepancy in $^{10}$Be concentrations between active channel alluvium and glacial terrace alluvium can be explained when we consider the contribution of $^{10}$Be-poor glacial alluvium from terraces that are actively contributing sediment to the channel. For example, active channel alluvium samples from the upper elevation portions of the SRW (Figure 3.28: samples: 7, 8, 9) record the highest $^{10}$Be concentrations within the SRW
Table 3.7: TCN sample locations, \(^{10}\)Be concentrations and CAER’s from the WRW and SRW, Washington.

<table>
<thead>
<tr>
<th>Sample Location</th>
<th>Sample Type /Number</th>
<th>Terrace Height</th>
<th>Sample Depth</th>
<th>Scaling Factor</th>
<th>(^{10})Be Production Rate (^{2})</th>
<th>(^{10})Be Concentration (^{3})</th>
<th>(^{10})Be Derived CAER (^{3})</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chiwawa R.</td>
<td>Terrace 1</td>
<td>5m</td>
<td>~ 3.5m</td>
<td>4.00</td>
<td>18.01</td>
<td>2.73 +/- 0.26</td>
<td>0.48 +/- 0.06</td>
</tr>
<tr>
<td></td>
<td>Terrace 2</td>
<td>5m</td>
<td>~ 2.5m</td>
<td>3.83</td>
<td>17.25</td>
<td>3.21 +/- 0.33</td>
<td>0.39 +/- 0.05</td>
</tr>
<tr>
<td></td>
<td>Alluvium 3*</td>
<td>NA</td>
<td>Surface</td>
<td>*</td>
<td>15.34</td>
<td>2.58 +/- 0.07</td>
<td>0.43 +/- 0.03</td>
</tr>
<tr>
<td>Nason Crk.</td>
<td>Alluvium 4*</td>
<td>NA</td>
<td>Surface</td>
<td>*</td>
<td>14.42</td>
<td>3.63 +/- 0.10</td>
<td>0.28 +/- 0.02</td>
</tr>
<tr>
<td></td>
<td>Terrace 5</td>
<td>7m</td>
<td>~ 2.5m</td>
<td>3.20</td>
<td>14.39</td>
<td>2.80 +/- 0.24</td>
<td>0.52 +/- 0.07</td>
</tr>
<tr>
<td></td>
<td>Alluvium 6</td>
<td>NA</td>
<td>Surface</td>
<td>3.04</td>
<td>13.69</td>
<td>3.36 +/- 0.22</td>
<td>0.31 +/- 0.03</td>
</tr>
<tr>
<td>Skykomish R.</td>
<td>Alluvium 7*</td>
<td>NA</td>
<td>Surface</td>
<td>*</td>
<td>13.31</td>
<td>2.18 +/- 0.07</td>
<td>0.43 +/- 0.03</td>
</tr>
<tr>
<td></td>
<td>Alluvium 8*</td>
<td>NA</td>
<td></td>
<td></td>
<td>12.32</td>
<td>2.18 +/- 0.06</td>
<td>0.41 +/- 0.03</td>
</tr>
<tr>
<td></td>
<td>Alluvium 9*</td>
<td>NA</td>
<td></td>
<td></td>
<td>12.87</td>
<td>1.87 +/- 0.05</td>
<td>0.46 +/- 0.03</td>
</tr>
<tr>
<td></td>
<td>Alluvium 10*</td>
<td>NA</td>
<td></td>
<td></td>
<td>11.47</td>
<td>1.62 +/- 0.05</td>
<td>0.54 +/- 0.04</td>
</tr>
<tr>
<td></td>
<td>Terrace 11</td>
<td>20m</td>
<td>~ 2m</td>
<td>2.66</td>
<td>11.97</td>
<td>1.05 +/- 0.25</td>
<td>0.90 +/- 0.24</td>
</tr>
<tr>
<td></td>
<td>Alluvium 12*</td>
<td>NA</td>
<td>Surface</td>
<td>*</td>
<td>11.30</td>
<td>1.49 +/- 0.05</td>
<td>0.57 +/- 0.04</td>
</tr>
</tbody>
</table>

\* Samples collected by Moon et al., 2011. Scaling factors presented in a different manner but with the same methods employed here. See Moon et al., 2011 Supplemental information in Appendix A for complete information pertaining to TCN samples.

1 \(^{10}\)Be production rate scaling factor generated using Stone (2000). Shielding factors not included in this table but were accounted for.

2 Sea-level high latitude (SLHL) base production rate of 4.5 at/g/yr (half-life 1.36 My) multiplied by scaling factor.

3 Error represents 1-Std. Dev. due to AMS measurement uncertainty. See appendix D for more information about all parameters included in the calculation of CAER. Only two decimal points of precision were reported from Moon et al., 2011 thus the level of precision shown above.

(i.e., slowest erosion) \((^{10}\text{Be})\) range: 1.87 – 2.18 X 10\(^4\) atoms/g. CAER range: 0.41 -0.46 +/- 0.03 mm/yr). Within this portion of the SRW there is a minimal amount of storage of glacial sediment accessible to the modern channel. In contrast, \(^{10}\)Be concentrations from active channel alluvium in the lower elevation portions of the SRW (Figure 3.28: samples: 10 and 12) are the lowest of all samples in this study (i.e., fastest erosion) \((^{10}\text{Be})\) sample 10: 1.62, sample 12: 1.49 X 10\(^4\) atoms/g). Within this portion of the SRW abundant 20-30 m tall glacial outwash terraces are actively contributing \(^{10}\)Be poor terrace alluvium to the stream. One such terrace (Figure 3.28: sample: 11) has the lowest \(^{10}\)Be concentration in this study (1.05 X 10\(^4\) atoms/g) indicating erosion nearly two times faster than any active channel alluvium samples.
Figure 3.28: Rates of sediment yield generated from CAER and corresponding $^{10}$Be concentrations within the WRW and SRW, WA. Terrace alluvium derived rates are shown in blue and active channel alluvium in red. Boxes (Above) enclose the different study basins and the numbers illustrate the sample locations within the study area.
This is compelling evidence of dilution of active channel alluvium samples with $^{10}\text{Be}$-poor glacial outwash terrace alluvium. This dynamic artificially lowers the $^{10}\text{Be}$ concentration of these samples and artificially elevates perceived catchment-averaged erosion rates.

The same dynamic occurs in Nason Creek where the terrace-derived $^{10}\text{Be}$ concentration (2.80 X 10$^4$ atoms/g) is much lower than both samples of channel alluvium (3.36 and 3.63 X 10$^4$ atoms/g). The dilution signal is not as obvious compared to the SRW, because there are no active channel alluvium samples collected higher in the sub-watershed for comparison. Also, there are relatively fewer (compared to the rest of the SRW and WRW study area) actively contributing sources of glacial sediment to dilute the active channel alluvium. This could explain why the $^{10}\text{Be}$ concentrations from the active channel alluvium are the highest in the SRW and WRW study area. Within the Chiwawa River, the broad, relatively low gradient U-shaped valley allows for a relatively larger amount of sediment storage than the other study sub-watersheds. During field reconnaissance I observed many 5 – 30 m glacial outwash terraces actively contributing sediment to the river. Because of the high rates of sediment supply from $^{10}\text{Be}$ poor glacial sediment, the active channel alluvium sample ([$^{10}\text{Be}$] sample 3: 2.58 X 10$^4$ atoms/g) is artificially low, inferring an erroneously high apparent erosion rate. The increased rate of dilution within the Chiwawa River compared to Nason Creek could account for the difference between the active channel alluvium sample from Chiwawa ([$^{10}\text{Be}$] sample 3: 2.58 X 10$^4$ atoms/g) versus Nason Creek ([$^{10}\text{Be}$] sample 4: 3.63 sample 6: 3.36 X 10$^4$ atoms/g).
Can these rates be explained in the context of the strong precipitation gradient?

Moon et al., (2011) interpreted the spatial variability of erosion rates within the Cascades as being primarily driven by the strong precipitation gradient created by the orographic effect of the Cascades (i.e., the rates are higher on the wetter, west-side of the Cascades). Specifically, the over-steepened landscape of the Cascades, which was sculpted by glacial processes, is out of equilibrium with the fluvial processes that are currently driving erosion in these watersheds which causes an increased rate of landslides. This is unlike many other regions where climate has been shown to be decoupled from millennial-scale erosion rates (Von Blanckenburg, 2006; Portenga and Bierman, 2011). We believe that this interpretation discounts the contribution of $^{10}$Be poor sediments from glacial sediments.

If the precipitation gradient has been causing an increased rate of landslides on glacially oversteepened topography then one could expect that watershed slope would strongly correlate with the precipitation gradient. The sub-set of Moon et al., (2011) samples included in our analysis exhibited only a weak ($R^2 = 0.137$) positive correlation. This is not to say that the precipitation gradient is not influencing the rates reported here, particularly because the extent of glaciation during the last glacial maximum was probably influenced by the same precipitation gradient that still exists (Moon et al., 2011) (i.e., larger glaciers that extended further into lowlands on the west-side of the Cascades). This is consistent with the $^{10}$Be concentrations observed in the SRW terrace compared to WRW terraces. Despite the probability that the precipitation gradient may be influencing long-term erosion in the Cascades, we found evidence that the contribution of glacial sediments to modern streams within this area is high. This causes a dilution in the
concentration of $^{10}$Be measured in the active channel alluvium. Discounting this contribution will lead to erroneous or incomplete analysis of the drivers of long-term sediment generation and delivery within this region.

3.6 Conclusion

Synthesis of the work from this chapter reinforces the complexity of sediment generation, routing and deposition over multiple temporal and spatial scales. Further, results from this chapter provide examples of how $^{10}$Be-derived CAER can be used to identify broad trends of variability. The case study from Bridge Creek, OR showed how CAER can be coupled with C14 dates and stratigraphic analysis to understand timing and duration of channel cut and fill cycles. The fact that CAER remained relatively steady through aggradation and degradation in Bridge Creek suggests that such phenomena can occur in the absence of changes in sediment supply. However, it is possible that the change in sediment supply was simply below the level of detection of our method (20-25 cm, Figure 3.20). The case studies from WRW and SRW illustrate how the geomorphic history of a watershed (e.g., glaciation) can have a profound influence on modern sediment dynamics. Further, the WRW and SRW case studies emphasize the need to consider degradation of near-channel alluvial deposits when interpreting $^{10}$Be-derived CAER. This example also illustrates how CAER can be used to effectively quantify changes in sediment supply at the watershed level if those changes are drastic (e.g., transition from glacial dominated to fluvial dominated erosion) but appears to be too coarse in the absence of near-term rates (i.e., yearly to decadal sediment yield) to effectively quantify the temporary pulse of sediment associated with wild-fire (e.g., Tower fire, EBNFC, JDW) if the scale of disturbance is not sufficiently large.
Finally, even when the scale of disturbance appears to be large (e.g., Tower fire in EBNFC) and has temporarily overwhelmed aquatic ecosystems locally, CAER can show that the disturbance did not profoundly change the watershed-wide sediment yield. My data-set is a useful benchmark for any subsequent short-term sediment yield studies within these watersheds and can provide important geomorphic context for aquatic ecosystem monitoring and restoration.
4.1 Introduction

Both fine (i.e., wash load) and coarse (i.e., bed material load) sediment are natural constituents of river systems and together form the geomorphic template for aquatic ecosystems (Palmer et al., 2000; Yarnell et al., 2006). As such, salmonids and aquatic invertebrates have evolved with specific substrate requirements that differ at various life stages (Gregory and Bisson, 1997; Montgomery, 2004). Thus, when any component (i.e., coarse or fine load) of the sediment load is in excess or deficit, the aquatic ecosystem can be negatively impacted (Waters, 1995).

Many researchers have analyzed the negative effects of excess fine sediment on aquatic ecosystems (Miller et al., 1989; Frissell, 1993; Suttle et al., 2004) but a deficit of fine sediment can also be problematic (Castro and Reckendorf, 1995). For example, the Yampa River, a large, unregulated tributary to the Green River, has high natural rates of fine sediment supply with high turbidity. As a consequence, the aquatic ecosystem tends to favor biota that do not rely solely on visual predation (including some rare and endangered species, e.g., Colorado pikeminnow). Within the Yampa, the endangered Colorado pikeminnow has a relatively healthier population compared to the rest of the regulated Green River watershed that has experienced profound changes to the sediment and flow regime (Van Steeter and Pitlick, 1998). Additionally, coarse sediment inputs in excess of transport capacity (e.g., post-wildfire debris flows) can temporarily, and locally,
overwhelm aquatic ecosystems, though there is evidence that some aquatic ecosystems are resilient to this type of disturbance (Pitlick, 1993; Howell, 2006).

In order to move beyond conceptual associations between long-term sediment supply and aquatic habitat (as outlined in Chapter 1 section 1.4) and towards quantitative functional relationships it is necessary to identify an appropriate scale of inquiry. Cosmogenically derived rates of long-term sediment supply averaged over an entire watershed suggest that a watershed-wide metric for aquatic habitat is an appropriate scale of analysis. Here, I conducted a first-order analysis of the relationship between long-term, watershed-wide sediment yield and a channel-type classification scheme employed by the National Oceanographic and Atmospheric Administration’s (NOAA) Columbia Habitat Monitoring Program (CHaMP). Specifically I sought to determine whether or not a relationship exists between the occurrences of alluvial (self-formed) channel types (CT) and long-term sediment yield. To be clear, I was not attempting to identify a causal link between long-term rates of sediment yield and alluvial channel types rather to identify an association between the two parameters.

4.1.1 Channel Classification and Aquatic Habitat

River researchers have long recognized the utility of geomorphically significant stream classification systems (Pennak, 1971; Frissell et al., 1986; Kondolf, 1995; Montgomery and Buffington, 1998). Currently there exist many different stream classification systems (Powell, 1875; Gilbert and Dutton, 1877; Schumm, 1977; Frissell et al., 1986; Hawkins et al., 1993; Rosgen, 1994; Montgomery and Buffington, 1998; Brierley and Fryirs, 2005) that attempt to link channel morphology to the geomorphic and hydrologic processes that are responsible for creation and maintenance of modern
channel form. Accurate classifications of CT can be used to develop spatial linkages between watershed-wide sediment generation, routing and deposition process zones. This is essential to assess channel condition, predict channel change from disturbance and to understand the historical trajectory of channel change.

From an ecological perspective geomorphically significant channel classification allows for identification of CT’s that may be associated with potentially high quality aquatic habitat. For example, Beechie and Imaki (2014) and others (Knighton and Nanson, 1993) recognize that lateral migration rates vary, in part, based on channel form (i.e., typically straight channels have the lowest lateral migration rates while braided channels have the highest) (Beechie et al., 2006a). Beechie and Imaki (2014) elaborate on this concept to identify the highest physical and ecological diversity associated with moderate or intermediate levels of disturbance represented by intermediate lateral migration rates. Intermediate lateral migration rates maintain high age diversity of floodplain surfaces, diversity of riparian species and a high diversity of aquatic species (Ward et al., 2002; Naiman et al., 2010). By this rationale, the identification of alluvial channel types associated with intermediate levels of lateral migration (e.g., Pool-Riffle and Island-Braided) also identifies potentially high quality aquatic habitat.

4.2 Methods

Channels were classified for each study sub-watershed by Beechie and Imaki (2014) using a multivariate predictive model based on established relationships of slope, discharge, valley confinement, sediment supply and caliber. These parameters were generated using readily available geospatial datasets. Essentially, six channel and landscape variables were ultimately chosen as model parameters. Three model
parameters were estimated from digital elevation data (channel slope, discharge and valley confinement) and the remaining three were chosen as surrogates for sediment supply and size. These included relative reach slope (reach slope compared to the slope of the immediately upstream reach), percent of watershed in bare alpine land cover and the percent of the watershed comprised of fine-grained erosive sediments. For a complete description of the model parameters and data-sets employed see Beechie and Imaki (2014). In Beechie and Imaki (2014), they generated four alluvial CT’s (e.g., straight, meandering, anabranching and braided) and one non-alluvial CT (confined) for streams with a bankfull width greater than 8 m. The CT’s used here were generated with the same model employed in Beechie and Imaki (2014) with the addition of Montgomery and Buffington's (1998) CT’s (e.g., pool-riffle, plane-bed, step-pool, cascade) for streams less than 8 m in bankfull width. These additional CT’s were predicted by simple slope thresholds and subsequently compared to reference sites for validation (personal communication, Tim Beechie). Each channel reach was at least 200 m long (mean = 210 m) with geomorphically similar reaches aggregated into longer reaches. The longest reach within our study sub-watersheds was 12 km.

Within the total population of CT’s I chose to identify alluvial channel types as being the most likely to have a direct relationship between long-term sediment supply and channel morphology. Theoretically, alluvial CT’s would have the necessary lateral accommodation space and low channel slopes to allow for sediment deposition. This would occur if the relationship between discharge and sediment supply is such to create a transport-limited condition. Alternatively, non-alluvial CT’s were considered to be less responsive to long-term rates of sediment supply as the channel slope and valley
confinement (defined as the ratio between floodplain width and bankfull width (VCR)) would not accommodate deposition even if the reach was in a supply limited condition. Additionally, high-slope, high-confinement CT’s would be more sensitive to channel form forcing by large woody debris inputs and immobile instream boulders. Finally, alluvial CT’s were analyzed because these CT’s are considered potentially higher quality aquatic habitat (Yarnell et al., 2006; Beechie and Imaki, 2014). This is because of the increased rates of lateral migration (compared to non-alluvial CT’s) forcing a higher rate of floodplain/channel exchange that creates high habitat heterogeneity as explained above.

Some CT’s are unequivocally alluvial because of the criteria used to delineate them (e.g., Pool-Riffle, Island-Braided and Straight) while others occur across a range of channel slopes and valley confinements (e.g. Plane-bed, Confined) (Figure 4.1). This means that a sub-set of the Plane-bed and Confined CT’s could potentially be alluvial if they meet a slope and valley confinement criteria. A slope threshold of 0.03 was used as a threshold between supply limited and transport limited reaches (i.e., alluvial vs. non-alluvial) following the rationale of Montgomery and Buffington (1997) and Beechie and Imaki (2014). In addition to a slope threshold it was also necessary to identify a level of valley confinement that would allow the lateral accommodation required for alluviation. Beechie et al., (2006b) and Hall et al., (2007) determined that it is extremely rare within the CRB for channels with limited floodplains (valley confinement ratio < 4) to develop alluvial channel patterns. Because of this I chose a VCR of 4 as an additional threshold for potentially alluvial channel reaches (Figure 4.2).
**Figure 4.1:** Frequency distribution of channel slopes for all reaches of all channel types (CT’s) within the study sub-watersheds. (Left) complete distribution of channel slopes with the colored boxes delineating the slope thresholds of channel types (Cascade = light-blue, Step-pool = light-green). (Right) same plot as left but at a finer scale to display only the slope values from 0 to 0.05 (Pool-Riffle = salmon-colored, Island Braided = dark-green, Straight = lime-green, Confined = yellow, Plane-bed = brown). Solid red line indicates the slope threshold for a potentially alluvial reach.
Figure 4.2 Box and whisker plots showing the valley confinement ratios for each channel type. Solid black lines are the median value, boxes enclose the interquartile range (IQR) and whiskers are 1.5 times the IQR. The dashed red line indicates the VCR threshold of 4.

The combined criteria of channel slope < 0.03 and VCR > 4 identified a population of potentially alluvial reaches within each study sub-watershed. I then calculated the proportions of each CT (alluvial and non-alluvial) that comprised the total channel length for each study sub-watershed. The proportion of potentially alluvial CT’s for a particular watershed were then plotted against the long-term rate of sediment yield derived from the same watershed. Sediment yield was derived from cosmogenic analysis and is described in detail in Chapter 2, section 2.2.
4.2.1 Model Validation

Beechie and Imaki (2014) present the error analysis involved in model development in detail in their publication. Essentially this consisted of error analysis of individual model parameters by comparison with field measurements. Sediment supply and caliber were derived from surrogate variables (as explained above) and could not be independently verified. Additionally they evaluated the model accuracy with a classification error matrix and compared alternative models using cross validation. They were able to compare model predicted CT’s to a population of test data that had been field verified. The most accurate model (Model 56) had an over-all accuracy of 82%.

In addition to this I visually compared the type and distribution of model predicted CT’s to my field-based geomorphic assessments conducted within three study sub-watersheds in the JDW. Generally there was good agreement between the sub-set of alluvial reaches I identified within the broader population of all model predicted CT’s and my own field-based delineations. Differences were observed within the channel lengths of the model predicted reaches and my own delineations (i.e., model predicted CT reaches were generally shorter than their field-verified counterparts) but I did not identify any locations where a non-alluvial CT was mis-classified as alluvial. Similarly, no model predicted alluvial CT was misclassified as non-alluvial except in Bridge Creek, Oregon where there were a few discrepancies. Within the vertically incised lower mainstem section of Bridge Creek the model mis-classified some reaches as confined because of the presence of fill terraces immediately adjacent to the modern stream despite the fact that alluvial channel patterns were observed in these locations during field assessments.
Additionally these discrepancies can be partly explained by recent channel change associated with an ongoing beaver-assisted restoration project.

**4.3 Results**

Plotting the distribution of CT’s from within each study sub-watershed, organized by categories of sediment yield (i.e. < 100, 100 – 200, 200 – 400, > 400 Mg/km²/yr) reveals several interesting trends (Figure 4.3). The total channel length of each study sub-watershed are dominated by the non-alluvial CT’s (Cascade, Step-pool and Plane-bed) regardless of the amount of sediment yield or drainage area (Figure 4.3). These CT’s typically occur in the higher gradient, supply-limited upper sections of each watershed. The dominance of these CT’s is expected given that the cosmogenic sample location of many of the study sub-watersheds were from, high-gradient, mountainous regions (See Chapter 2 Figure 2.11 and Chapter 3, Figure 3.1 for cosmogenic sample locations).

Figure 4.3 also shows that the high sediment yield (> 400 Mg/km²/yr) watersheds have a larger distribution of all CT’s compared to the lower yield watersheds which are dominated by the non-alluvial CT’s. Very few of the alluvial CTs (specifically pool-riffle, island-braided, straight) occur in drainages with less than 400 Mg/yr/km2). Similarly, the higher sediment yield watersheds (>400, 200-400 Mg/km²/yr) have a greater proportion of total watershed channel length composed of alluvial CT’s compared to the lower (<100, 100-200 Mg/km²/yr) sediment yield watersheds. A potentially misleading trend in Figure 4.3 is the absence of low sediment yield watersheds (<100 Mg/km²/yr) from the Island-braided, Straight, and Confined CT’s. This is an artifact of the classification system which only considers channels with a bankfull width > 8 m for potential inclusion into those channel types.
Figure 4.3 Proportion of each study sub-watershed’s stream length composed of different channel types (CT). Each individual box represents the portion of an individual study sub-watersheds (JDW, PRW, Cascades) stream length composed of that CT. Total channel length was used regardless of cosmogenic sample location. CTs are organized along a continuous axis from alluvial (Pool-Riffle) to non-alluvial (Cascade). Watersheds were classified into different colors based on rates of sediment yield (Mg/km²/yr).

The low sediment yield watersheds are from small, (drainage area < 50 km²) mountainous drainages in the North Fork John Day Watershed and have channels with bankfull widths less than eight meters. So even if the < 8 m bankfull width drainages met all other model criteria for inclusion into the Island-braided, Straight and Confined CTs they would not be included in Figure 4.3. Within these small watersheds, potentially alluvial reaches (e.g. slope < 0.03 and VCR > 4) were still identified within the Plane-bed CT but are not included in the Plane-bed CT in Figure 4.3. Instead, the potentially alluvial channel types (slope < 0.03, VCR > 4) from study sub-watersheds with small channels (< 8 m bankfull width) were identified from the Plane-bed CT and included in the Alluvial Channel Types parameter shown on the X-axis in Figure 4.4. This enables a
**Figure 4.4** Percent of total channel length within study sub-watersheds composed of alluvial channel types compared to long-term rates of sediment yield. (Top) All samples (Middle) Cascade Range, WA samples (i.e., Wenatchee and Entiat watersheds), and (Bottom) John Day Watershed, OR samples. (Top) Dashed line indicates a possible threshold in percent alluvial channel length after which a positive trend is observed amongst the Cascade Range samples.
When all samples are considered, Figure 4.4 (Top) shows a slight positive trend in increasing rates of long-term sediment yield with increasing proportion of channel length composed of alluvial CTs. This is particularly evident beyond approximately 13-14% of total channel length composed of alluvial CTs (Figure 4.4, Top dashed line). With a few exceptions, this threshold discriminates between small mountainous drainages (drainage area < 50 km²) with few alluvial reaches to larger watersheds with more of the watershed composed of low gradient, relatively wider alluvial valleys. When the two study areas are viewed separately, we find that the relationship is primarily driven by the watersheds within the Cascade Range (Figure 4.4, Middle). Here a strong relationship is evident with minimal scatter. In contrast, there appears to be no discernable relationship between long-term sediment yield and percent alluvial CTs within the JDW, OR (Figure 4.4, bottom).

4.4 Discussion

E.W. Lane (1954) (Chapter 1, Figure 1.1) and others (Church, 2002) have developed the conceptual model for alluvial channels that describes channel form resulting from the interaction between the quantity and caliber of sediment and water supplied to a stream. This indicates that conceptually long-term rates of sediment supply will influence channel form. Despite this, can we assume that modern channel form, as described by the model-predicted channel types, would be responsive to long-term (millennial-scale) rates of sediment yield? The following discussion examines issues
related to a tight coupling of modern channel form with long-term rates of sediment yield.

Problems associated with the CT model itself could potentially obscure any relationship existing between channel form and long-term sediment yield. Misclassification of channel type potentially introduces error into any further analysis. Here, that is thought to be a minimal source of error because the model performed well (82% model accuracy) and was field verified in a sub-set of study watersheds that were considered representative of the channel diversity of the entire CRB (Beechie and Imaki, 2014). Additionally, misclassification errors were minimized by independently selecting a population of all potentially alluvial reaches (slope < 0.03, VCR > 4) from all stream reaches regardless of channel type (Figure 4.4).

The lack of a model parameter related to riparian vegetation and large woody debris inputs is also a potential source of error. Riparian vegetation combined with the cohesive properties of the sediment composing the bank affect bank strength which influences lateral migration rates (Millar, 2000). Additionally, wood loading has been shown to be a primary determinant of channel form in many forested settings (Abbe and Montgomery, 1996; O'Connor et al., 2003; Hassan et al., 2005). Beechie and Imaki (2014) acknowledge the lack of these model parameters and recognize that discounting the influence of wood loading probably results in some level of misclassification. This parameter was omitted from the model because currently there are no available data-sets related to wood loading that cover the entire CRB. Beechie and Imaki (2014) also suggested that the role of riparian vegetation is insignificant for streams > 8 m bankfull width because larger streams possess sufficient shear stress to maintain lateral migration.
despite the increase in bank strength from riparian root structures (Beechie et al., 2006b). However, several recent studies demonstrate that riparian vegetation exerts strong control over channel morphology, even in large rivers. For example, Parker et al. (2007) suggested that riparian vegetation was a primary control on large sand and gravel bed rivers. Souffront (in press) and others (Dean and Schmidt, 2011; Manners et al., 2014) showed that riparian vegetation can have a significant effect on rivers as large as 75-100 m wide. However, the CT model performed reasonably well despite omission of these important factors.

Another potential problem in evaluating the relationship between channel type and long-term sediment yield relates to the time-scales of comparison. Specifically, the rates of sediment yield are derived from the concentration of the cosmogenic isotope $^{10}$Be that accumulates over the millennial time-scale (for a detailed discussion see Chapter 2 section 2.2) and represents the average rate of sediment yield over that time. The averaging time-scale for our sampled catchments varies as a direct function of the erosion rate, ranging from 1 to 20 ka in our study sub-watersheds (with faster erosion rates averaging over shorter time-scales). This potentially presents a problem because the channel types are delineated from modern topographic, geologic and climatic parameters. Channel form can change considerably over shorter time-scales in response to changes in base level, sediment supply and caliber, or water discharge (Anderson and Anderson, 2010). Over the millennial-scale, climatic parameters are probably the most likely (of the three previously listed parameters) to experience change potentially affecting channel morphology. Presumably over thousands of years channel form has changed within the
study sub-watersheds. This leads to the question of which channel types are more likely to persist over time?

The broad classes of alluvial versus non-alluvial channel types may be less sensitive to changes in channel form over the millennial-scale than the full list of model-predicted channel types. Valley width (e.g., VCR) should be relatively insensitive to change over the millennial-scale as the processes of landscape evolution occur over longer time-scales (millions versus thousands of years). Similarly, channel gradient in non-alluvial settings (i.e., bedrock) would be relatively unresponsive to changes over the millennial-scale except in active tectonic settings. Stream channel longitudinal analysis (Chapters 2 and 3) show little to no active tectonic forcing in the majority of the study sub-watersheds. Channel gradient in alluvial reaches could be more responsive to changes in sediment and water discharge over the millennial-scale but it is unlikely that a reach would shift from fully non-alluvial (slope > 0.03, VCR < 4) to fully alluvial (slope < 0.03, VCR > 4) over the millennial-scale.

Finally anthropogenic and natural perturbations to the modern stream channel could cause deviations from model predicted channel type. Though this may be a problem locally (i.e., individual stream reaches) the scale of disturbance would have to be intense and spatially extensive for alluvial reaches to be classified as non-alluvial and vice-versa. In sum, alluvial channel types should be accurately identified by the model and insensitive enough to change over the millennial-scale to permit a first order analysis of the relationship between the proportion of alluvial CT’s and long-term sediment yield.

Figure 4.4 (Middle) suggests that there is a strong positive relationship between long-term sediment yield and percent of alluvial channel types within the study sub-
watersheds in the Cascade Range. This is in contrast to the JDW where no such relationship was observed. This discrepancy may be partially explained by the difference in the glacial history of the JDW compared to the study sub-watersheds within the Cascade Range. The Cascade Range experienced a profound glaciation during the Pleistocene. Chapter 3 (Section 3.5) describe how high rates of sediment yield associated with glaciation are still influencing modern processes of fluvial erosion. Essentially, the highest rates of long-term sediment yield were recorded in watersheds that have abundant near-channel glacial outwash terraces actively contributing sediment to the stream. Presumably the areas that these terraces occur possess attributes that facilitate deposition (slope < 0.03, VCR > 4) and would be classified as alluvial CT’s. It should be noted, that in Chapter 3 (Section 3.5) I identified a process of dilution of $^{10}$Be-poor sediment sourced from the actively contributing glacial outwash terraces in some of the study sub-watersheds within the Cascade Range. This process may be responsible for artificially elevating the perceived long-term rates of sediment yield derived from $^{10}$Be analysis. To minimize this potential bias, I excluded the study sub-watersheds where this dynamic appeared to be the most pronounced (i.e., the Skykomish River Watershed samples).

The presence of terraces in the high sediment yield watersheds is evidence of a shift in sediment supply and/or discharge since glaciation that caused abandonment of the terraces. The modern streams are likely in a phase of evacuation of the abundant Pleistocene glacial deposits. Despite this, the streams have been capable of routing and depositing large amounts of sediment in the past which may be evidence of their relatively high rates of long-term sediment yield. This is in contrast to a smaller stream that either currently or historically has been supply-limited and has not created of
maintained alluvial reaches. This could be indicative of relatively lower rates of sediment yield. Another potential explanation for the positive trend of percent alluvial channel types and long-term sediment yield in the Cascade Range are the effects on valley morphology from glacial erosion. Glacial erosion typically carves a characteristic U-shape to valley profiles where the valley gradient can be low and valley width is high. These are the parameters that facilitate alluvial channel conditions and are highly represented in the watersheds within the Cascade Range. This is in contrast to the JDW and PRW.

Mechanistically there may be several reasons why there appears to be no discernable relationship between sediment supply and percent alluvial CT in the John Day Watershed. Many of the study sub-watersheds were sampled from the high gradient and high valley confinement sections of the drainage network where these factors do not facilitate wide-spread alluviation. If samples were collected lower in the watersheds there likely would have been a higher percentage of alluvial channel types though this would not have changed the rates of sediment yield. Additionally, the presence of wood likely exerts a stronger influence on channel morphology within these small streams. The potential increase in wood-forced alluvial channel types would not have been captured by the CT model, thus alluvial CTs could have been erroneously underrepresented. Finally, comparing cosmogenically derived millennial-scale erosion rates from dated terrace deposits to the same rates derived from modern channel alluvium within a subset of study sub-watersheds in the JDW shows evidence of relatively steady and low rates of erosion and sediment yield. This is described in detail in Chapter 3. The steady rates of sediment supply would facilitate sufficient adjustment time for streams to attain a well-graded
condition in terms of sediment routing where little alluviation would be expected. The well-graded, concave-up longitudinal profiles from within the JDW are consistent with this interpretation.

4.5 Future Work

A larger sample size would be needed to further untangle the relationship between alluvial channel form and rates of long-term sediment yield across the diverse CRB. Including a wood loading parameter to the model predicted CTs would be beneficial in ensuring that all potentially alluvial reaches were included in small, high-gradient, forested streams. It would also be informative to decrease the scale of inquiry to the reach-scale. At this scale metrics more directly linked to aquatic habitat (e.g., pool frequency, width/depth ratios, pool tail fines, etc.) could be compared to isolate the role of sediment supply on metrics relevant to aquatic habitat. For instance, two channel reaches that are predicted to have the same channel form based on slope, valley confinement, discharge and other parameters have very different morphologies. Could the rate of long-term sediment yield help to explain that difference and are certain aquatic habitat metrics more responsive to differences in sediment yield?
5.1 Concluding Comments

Predicting the mechanisms, rates and timing of sediment production and storage within a watershed are central challenges in geomorphology and Earth surface sciences (Trimble, 1977; Wolman, 1977; Walling, 1983; Schaller et al., 2001; Council, 2010; National Research Council, 2010). Sediment generation and storage within a watershed are difficult to quantify because of ‘natural’ and human-induced variability in both time and space (Kirchner et al., 2001; Ferrier et al., 2005; Smith et al., 2011). This non-uniformity is produced by a variety of interrelated hydrologic, geomorphic and environmental processes that can be affected by human land-use (Walling, 1983). The mechanisms of sediment generation, transport and deposition are highly non-linear in regards to hydrologic flux and sediment availability. Additionally, sediment can be temporarily stored and remobilized many times during transport from primary source to ultimate deposition. The proportion of total sediment efflux that experiences temporary storage can range from 0 to 100%, with storage lasting from a few days to thousands of years (Allen, 2008; Skalak and Pizzuto, 2010; Smith et al., 2011). Because of this temporal variability, conventional approaches for predicting sediment yield (e.g., suspended and bed-load sediment yield gages and sediment accumulation rates in reservoirs) provide incomplete, and potentially misleading information (Schaller et al., 2001; Ferrier et al., 2005). Further, the time and cost involved in establishing and maintaining an effective watershed-wide gaging network (decades of work and $10^5$ -
$10^6$) is substantially greater than the time and cost required for a small number of TCN samples to derive a catchment-averaged erosion rate (CAER; one year and $\sim$10$^3$ per sample).

Ultimately many techniques are available to analyze sediment dynamics within a watershed (e.g., sediment gaging, dam/reservoir filling rates, sediment budgets, geomorphic change detection and geochemical methods) and the specific tools employed will be dependent on the study objectives and logistical constraints. For example, even a well-established gaging network alone may yield misleading information. Trimble (1999) provides an excellent case study illustrating the complexity of sediment yield dynamics over time in Coon Creek, an agricultural watershed in Wisconsin. Trimble constructed a sediment budget (from gage data and other sources) accounting for 140 years (1853-1993) since Euro-American settlement and found that total sediment efflux leaving the watershed remained relatively steady over time 39,000 T yr$^{-1}$, despite the fact that terrestrial erosion decreased from 441,000 to 117,000 T yr$^{-1}$, due to changes in agricultural practices. Large shifts in sediment sources and sinks within the watershed were not reflected in the total efflux leaving the watershed. Another example from Kirchner et al. (2001), working in 32 mountainous catchments in central Idaho, analyzed sediment yield over decadal, millennial and million year timescales using a combination of gage data and geochemical methods. He found that long-term yield (millennial and million year time scales) was roughly spatially uniform, and on average 17 times higher than decadal yields. This data suggests that modern sediment delivery from mountainous terrain can be relatively low, with long-term averages significantly higher due to low-frequency episodic delivery via high magnitude mass-movement events following forest
fires. These two examples illustrate how multiple lines of evidence are often necessary to unravel the complexity of sediment generation, routing and deposition over multiple temporal and spatial scales.

Quantifying long-term (i.e., millennial-scale) sediment yield by measuring the concentration of Berillium-10 ($^{10}\text{Be}$) in river-borne sand can help establish the ‘natural’ or reference conditions, in terms of sediment yield, that aquatic ecosystems have experienced. It should be noted that such $^{10}\text{Be}$ derived catchment-averaged-erosion-rates (CAER) do not allow for the quantification of individual processes of erosion or sediment routing (e.g., rill erosion, individual debris flows or landslides, etc.) but instead identify broad trends of variability (Portenga and Bierman, 2011). CAER are also sensitive to recent, very large changes in catchment erosion rates (e.g., removal of ~ 30 cm of material). Long-term erosion rates can identify erosional hotspots within a landscape, which can help guide reach-scale salmonid restoration and habitat monitoring efforts. Further, these rates can be interpreted relative to modern sediment gaging techniques (e.g., dam in-filling rates, in-stream sediment transport measurements, etc.) and/or the CAER measured in dated terrace deposits to quantify the degree of recent perturbation or disturbance within a watershed, which could provide useful context for restoration actions (Brown et al., 1998; Hewawasam et al., 2003). Finally, CAERs are effective at quantifying the rates and variability of sediment yield at a variety of scales and either alone or in combination with other methods, can provide important geomorphic context for watershed and aquatic ecosystem management (Montgomery et al., 2000; Belmont et al., 2011).
5.2 Management Implications

My work has shown an order of magnitude spatial variability in the rates of millennial-scale sediment yield at the scale of the entire Columbia River Basin. At this broad scale long-term rates of sediment yield generally are poorly predicted from topographic and environmental parameters. A notable exception is the observed positive correlation between mean annual precipitation and sediment yield. This was surprising given the finding of Von Blanckenburg (2006), Portenga and Bierman (2011) and others (Riebe and Kirchner, 2001a; Riebe and Kirchner, 2001b) who showed globally climate is poorly correlated with millennial-scale rates of sediment yield. Where other functional relationships exist, the nature of those relationships are scale and situation-dependent. In sum, watershed managers and restoration practitioners should not solely rely on models that predict sediment yield based on traditional geomorphic assumptions from readily available topographic and environmental parameters. Instead, multiple lines of evidence including the use of CAER could be used to quantify and estimate rates of sediment yield at multiple scales.

In addition to the broad scale trends discussed in Chapter 2, each smaller study watershed (e.g., ~ 10 – 2,000 km²) has a distinct geologic, geomorphic and disturbance history that sets the template for the modern sediment dynamics and the physical aspects of aquatic habitat. It is essential to consider this history when planning watershed management and restoration actions. In order to differentiate between degraded conditions and the natural range of variability in channel form for a particular watershed a historical understanding of sediment flux is critical. For example, the case studies from WRW and SRW illustrate how the geomorphic history of a watershed (e.g., glaciation)
can have a profound influence on modern sediment dynamics. Another example from Bridge Creek, Oregon illustrates how the historic large-scale removal of beaver coupled with the degradation of riparian vegetation from grazing and agriculture has changed the water and sediment routing dynamics within this watershed. This has resulted in a degraded condition of the modern aquatic habitat. Further, the CAER from the CRB that comprise my data-set are a useful benchmark of natural, background conditions of sediment yield which can be combined with subsequent short-term sediment yield studies to quantify the degree of perturbation within particular study-sub-watersheds.

I also illustrated how CAER can be used to effectively quantify changes in sediment supply at the watershed level if those changes are drastic (e.g., transition from glacial dominated to fluvial dominated erosion or large-scale transition from primary forest to intensive agriculture (Hewawasam et al., 2003)) but appears to be too coarse in the absence of near-term rates (i.e., yearly to decadal sediment yield) to effectively quantify the temporary pulse of sediment associated with wild-fire (e.g., Tower fire, EBNFC, JDW) if the scale of disturbance isn’t sufficiently large. Further, even when the scale of disturbance appears to be large (e.g., Tower fire in EBNFC) and has temporarily overwhelmed aquatic ecosystems locally, CAER can show that the disturbance did not profoundly change the long-term, watershed-wide sediment yield. It should be noted, that I do not discount the importance of smaller-scale perturbations to the aquatic biota but it was not the goal of this study to specifically quantify perturbations at that scale. Finally, in Chapter 4 I identified a positive correlation in the Cascades, WA between long-term sediment yield and potentially alluvial reaches. This is a first-step in establishing
quantitative relations between long-term sediment yield and channel form, an important physical aspect of aquatic habitat.

In summary, the Columbia River Basin is an extremely diverse physiographic region. This diversity has important implications in setting the pace of sediment yield within the watershed. Traditional, ‘cookie-cutter’ approaches to watershed management and restoration and rehabilitation actions should be informed by long-term sediment yield data and may yield more positive outcomes.
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APPENDICES
APPENDIX A

Calculating Production Rates and Catchment Averaged Erosion Rates

A. Calculating the effect of altitude and latitude on $^{10}$Be production rates across an entire watershed. The following tools, scripts, and methods used in this section were written by Luke J. Reusser at the University of Vermont.

Below are the steps we take to summarize the effective $^{10}$Be production of a drainage basin in a point, which can then be used as input parameters to the CRONUS calculator or to Cosmocalc (Vermeesch, 2007) to determine catchment averaged erosion rate for an entire basin. It should be noted that these steps may not be the most efficient way of producing the desired results, but it works for us. Details of folder hierarchy can be modified; again, it is just the way that I find to be most useful and organized.

Steps used to summarize $^{10}$Be production of a basin at an effective elevation for CRONUS or Cosmocalc erosion rate calculation. The following document was written for ArcGis v10.

1. Delineate your basins in ArcGIS.
2. Create a folder called “ASCII_files” in your workspace and within that folder, create individual folders for each basin. The extracted ASCII files will be stored here after being created in the next few steps.
3. Download and extract the Model_ascii folder. IMPORTANT: All files within this folder must remain there in order to function properly. The folder can be moved around, but don’t move any of the files to other folders!
4. In the “ASCII Prep For CRONUS” toolbox, open the “Extract_ASCII_Files” tool.
5. In the command window, select the DEM the basins were delineated from and the individual basin shapefile for the first two Inputs. For the final three inputs (long.txt, elev.txt, and lat.txt) navigate to the specified basin’s folder within the “ASCII_files” folder you created in Step 5. NOTE: Text file names must not be changed. The Matlab script calls for text files specifically called “long.txt,” “lat.txt,” and “elev.txt.” This is why .txt files for each basin must be stored in individual folders.
6. Once all five inputs are selected, push “OK” to run the ASCII extraction tool.
7. Repeat this for each basin you’ve delineated. I find it useful and easier to run the tool in Batch mode. If you do this, make sure each folder pathway is correct. The process will go even faster if your DEMs are clipped to the extent of each basin, but that is personal preference.
8. ArcGIS 10 seems to create a .prj file associated with each text file. You won’t need these and it is fine to delete them.
9. Paste a copy of the erate_v2.m file in each basin folder in the “ASCII_files” folder.
10. Double-click the .m file to open the script in Matlab. In line 26 of the script, change the sample name to your desired basin.

11. In line 28, make sure the value corresponds to the resolution of your DEM.

12. Hit F5 to run the script and if prompted, select “Change Directory.” By default, the output will show a map of production rates within your basin. Smaller basins will display blue lines crisscrossing the map; these are just artifacts and do not affect the outcome. Other results are displayed in the Command Window. Default results are: Effective Elevation (EFFelevkm), Mean Basin Elevation (meanelevkm), Mean Latitude (meanlat), and Mean Longitude (meanlong).

13. Record the values for EFFelevkm, meanlat, and meanlong and use these values as your input parameters for the CRONUS calculator or to Cosmocalc.

Folder Hierarchy:

NOTE: Folder/file names are examples of the last bit of work that I did. Obviously, you may change them however you prefer. Text in Blue indicates a Folder and text in Red indicates a file.

- SAfrica_Basins
  - ASCII_Files
    - Basin1
      - long.txt
      - elev.txt
      - lat.txt
      - erate_v2.m
    - Basin2
      - long.txt
      - elev.txt
      - lat.txt
      - erate_v2.m
    - Basin3
      - long.txt
      - elev.txt
      - lat.txt
      - erate_v2.m
  - Rasters
    - SAf90mSRTM
  - Shapefiles
    - Individual_Basins
      - Basin1.shp
      - Basin2.shp
      - Basin3.shp
    - AllBasins.shp
B. Topographic Shielding

The following section is a workflow to calculate topographic shielding for an entire basin using a modified version of the matlab script tlmshield.m which is in turn a modification of the Balco Shielding.m script. Our modified version of tlmshield.m is called tlmshieldV2.m. Tina Marstellar created the tlmshield.m script while calculating erosion rates for her master’s thesis: Investigating Sediment Source to Sink Processes in a post-Orogenic Landscape. This section is a modified version of Appendix A: Calculating Erosion Rates for Basins from Marstellar’s thesis.

NOTE: The Marstellar document (Appendix A) is a modification of Balco, 2001 found at http://depts.washington.edu/cosmolab/P_by_GIS.html and includes some reference to matlab files that can be found here: http://hess.ess.washington.edu/math/al_be_v22/functionlist.html

The matlab script tlmshieldV2.m requires the following ascii files:

elv.txt, wsheds.txt, xgrid.txt, ygrid.txt, screen.txt

As such, the following workflow is an abbreviated version of Martse\lller’s modification of Balco’s original workflow to generate the previously listed ascii files using Arcinfo and Arcdesktop version 9.3. Again it should be noted that this is not the most efficient method but worked for us.

1) Get a DEM of your watershed (There are many potential sources (USGS 30m, SRTM)

To use tlmshieldV2.m the DEM must be in either:

Projected coordinate system name: NAD_1983_UTM_Zone_name
(Or if your watershed covers more than one UTM zone use…..)
Geographic coordinate system name: GCS_North_American_1983

2) Clip to boundaries of watershed of interest (if you don’t have a shapefile to do this, there are multiple ways to delineate a watershed see Marstellar’s Section 2. From her Thesis Appendix A Calculating Erosion Rates for Basins, for a method using arcinfo and grid.

3) Use the FILL tool in Arc desktop and name this DEM: basinname_grd

4) Now we need to create Basinname_ws raster. This is necessary to define the pixels of interest within your watershed but first we need to the following:

Open Arcinfo v9.3 and type:

Arc: &station 9999
Arc: grid

In grid:
Grid: &workspace <pathname of directory/folder>
Grid: basinname_fd = flowdirection(basinname_grd)
Grid: basinname_acc = flowaccumulation(basinname_fd)
Grid: basinname_cov = gridpoint(basinname_acc)
Grid: q (this returns you to Arc:)

Back in Arc:

Arc: Pointgrid basinname_cov basinname_pnt #
(note the number is the basin number and will need to be modified for each item. And then modified later in the matlab script tlmshieldV2.m)
Arc: Cell Size (square cell): 90
(This is the size of your DEM, mine was 90m)
Arc: Convert the Entire Coverage(Y/N)?: y
Arc: Enter background value (NODATA | ZERO): nodata
Arc: grid

(Note: Depending on your pixel cell resolution you might get an error message at this point saying basinname_pnt# exceeds 10000 and number of unique values exceeds 500. Please use Buildvat if a vat is required…I have not needed to create a VAT and have ignored the error message with no problems.)

In Grid type:

Grid: Basickname_ws =   watershed(basinname_fd,basinname_pnt)

You now have a grid called basickname_ws in which defines pixels we are interested in. We can now kill other files.

In grid:
Grid: kill basinname_fd
Grid: kill basinname_cov
Grid: kill basinname_pnt

5) Next we need grids containing the x and y coordinates of each pixel. This can be done in MATLAB if you know how your elevation grid is georeferenced but is possible to do in ARC. In GRID, issue

Grid: &describe basickname_ws
Grid: setcell basickname_ws
Grid: setwindow basickname_ws
Grid: ygrid# = (%$grd$ymax% - (%$grd$dy% / 2)) - ( $rowmap * %grd$dy% )
Grid: xgrid# = (%$grd$xmin% + (%$grd$dx% / 2)) + ( $colmap * %grd$dx% )

Note: type these exactly, include spaces, or you’ll get an error message.
6) **One more grid is also required.** Most pixels in the image are not on ridgelines and will not significantly contribute to topographic shielding of most other points. In the shielding calculation for each pixel we consider two groups of other pixels in the landscape: one, pixels near the pixel of interest, and two, pixels on ridgelines. We blow off all the other pixels. This greatly reduces execution time. The "ridgeline" pixels are most easily obtained by taking only those pixels which have a flow accumulation value of zero:

In grid:

\[
\text{Grid: screen = con((basinname_acc EQ 0),1,0)}
\]

It's easiest to get ARC grids into MATLAB by exporting ASCII files:

\[
\begin{align*}
\text{Grid: elv.txt = gridascii(int(basinname_grd))} \\
\text{Grid: wsheds.txt = gridascii(basinname_ws)} \\
\text{Grid: xgrid.txt = gridascii(int(xgrid#))} \\
\text{Grid: ygrid.txt = gridascii(int(ygrid#))} \\
\text{Grid: screen.txt = gridascii(screen)}
\end{align*}
\]

7) **Then use notepad to:**

A) Remove the first six lines of each text file. This is header information required by ARC. Note: You might want to save the header information in another notepad file so you can convert your final ascii’s from matlab back into arc rasters.

B) Use the find and replace function to replace all -9999’s with NaN. You shouldn’t have to do this with the xgrid.txt and ygrid.txt files.

8) **Now open Matlab** and make sure that your current folder is set to the same folder you have already been working out of. At the very least you need the elv, wsheds, xgrid, ygrid and screen.txt files as well as the tlmshieldV2.m script in this current folder.

We now need to modify our wsheds.txt file so there are only 1’s representing the area of the watershed. Type the following:

```
>> Load wsheds.txt
>> wsheds(wsheds > 0) = 1;
>> dlmwrite('watersheds.asc', wsheds, ' ')
```

(Note: this writes your modified wsheds variable to an ascii file called watersheds.asc)

Now open watersheds.asc in notepad and save as wsheds.txt. This will overwrite your old version of wsheds.txt and is now ready for use by the tlmshieldV2.m script.

9) **Run the tlmshieldV2.m script.** In the first line you will need to define your watershed number per the number you used above in step 5 (basinname_pnt # ). Depending on the size and cell resolution of your DEM this may take anywhere from 10 minutes to 18 hours.
10) **If everything worked** you should have a variable called `s_factor` that represents the topographic shielding factor for each pixel in your watershed except the last bounding cell on the edge of your watershed boundary (Note: this is because of the way the script is currently written but could be modified if you are interested in the values from those cells).

To visualize this matrix in matlab type the following:

```matlab
>> imagesc(s_factor)
>> colorbar
```

Here is an example of what the image should look like:

![Image](image.png)

11) **If you want to load s_factor** back into arc to get a mean value for topographic shielding from your whole watershed type the following in matlab:

```matlab
>> dlmwrite('Filename.asc', s_factor, ' ')
```

Now you need to reformat the ascii so arc will recognize it.

1) Open notepad and paste the first 6 lines of header information that you previously saved from step 8 into the Filename.asc file.
2) Find and replace all NaN’s with -9999.
12) **Open Arc desktop** and use the ascii to raster conversion tool on the Filename.asc file (make sure to choose the FLOAT option for the output. You now need to define the projection of this raster using the same projection you initially used in step 1. You also will want to clip this raster using the polygon of your watershed shape so that when you get the mean value of topographic shielding it doesn’t include areas outside of your watershed of interest.

Now you can get a mean value from your raster for the topographic shielding across your entire basin.

13) **If using 10Be as the cosmogenic nuclide in a quartz poor landscape, clip the raster of the shielding factors for the whole watershed** by non-quartz bearing rock-types to remove portions of the watershed that are not contributing to quartz production. You can now also clip the raster to subbasins within the broader watershed depending on the location of your samples.

14) **Rejoice!**
APPENDIX B

Downloading Oregon State University (OSU) Climate Group’s Parameter-elevation Regressions on Independent Slopes Model (PRISM) data

1) Navigate to the PRISM website: [http://www.prism.oregonstate.edu/](http://www.prism.oregonstate.edu/) choose the Products tab from the drop down menu.

2) There are three primary data types available (Graphics, Grids, Explorer). The following information pertains to the ArcInfo ASCII data GRIDS.

   Data grids in ArcInfo ASCII format are available for all base climate parameters for the period 1895-Present. Use the data in your own GIS application to perform your own analysis or calculations.

   There are multiple raster cell resolution options available for download the following pertains to the 30-arcseconds (800 m) data available under the Products tab.

   1) Under the <800 m Data> <Climatology Normals> <1971-2000> section click on the products matrix
   3) The download procedure for each data type is the same. After clicking on the desired data type you will be taken to a page of summary metadata with information related to the time period, spatial format, resolution, units, scale factor, time interval, projection, etc. (this varies for each data product). If more detailed information is required there is a link to additional metadata on this page. Header information related to the data required by ArcMap is also displayed on this page.
   4) Click on the link to Download Data at the bottom of the page.
   5) Use a File Manager program like 7-Zip to extract and transfer the download to the desired location. Before extraction and transfer the file will be a .gz format and after extraction will be in the .asc format.
   6) Load the .asc file in ArcMap, it will not display correctly because it doesn’t have a projection defined.
   7) Use the Define Projection tool in ArcMap, I used WGS84 but your projection may differ depending on the spatial extent of your area of interest.
   8) Use the Ascii to Raster tool in ArcMap to convert the .asc to a raster. I choose the Float format instead of Integers.
9) At this point you should see a black and white raster of the conterminous USA with the climate data displayed. Now you can assign whatever symbology you desire and clip the raster to the area of interest.
Appendix C

Classifications of broad-scale lithology and quartz-bearing rock types. Asterisk denotes that quartz is present but usually < 5%.
<table>
<thead>
<tr>
<th>Rock Type</th>
<th>Sedimentary</th>
<th>Quartz Bearing</th>
<th>Metamorphic</th>
<th>Quartz Bearing</th>
<th>Extrusives</th>
<th>Quartz Bearing</th>
<th>Intrusives</th>
<th>Quartz Bearing</th>
<th>Eolian</th>
<th>Quartz Bearing</th>
</tr>
</thead>
<tbody>
<tr>
<td>Conglomerate</td>
<td>Y</td>
<td>Greenstone</td>
<td>N</td>
<td></td>
<td>Lava Flows (Basalt)</td>
<td>N</td>
<td>Dacite</td>
<td>Y</td>
<td>Loess</td>
<td>Y</td>
</tr>
<tr>
<td>Sandstone</td>
<td>Y</td>
<td>Limestone/ Marble</td>
<td>N</td>
<td></td>
<td>Andesite</td>
<td>Y</td>
<td>Diorite</td>
<td>Y*</td>
<td>Eolian/Fluvial</td>
<td>Y</td>
</tr>
<tr>
<td>Mudflow Breccia</td>
<td>Y</td>
<td>Marble</td>
<td>N</td>
<td></td>
<td>Rhyolite</td>
<td>Y</td>
<td>Quartz Diorite</td>
<td>Y</td>
<td>Eolian/Lacustrine</td>
<td>Y</td>
</tr>
<tr>
<td>Claystone</td>
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<td>Metacarbonate</td>
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Sources of digital geologic layers:


**Washington:** Washington State Department of Natural Resources: Geosciences Data URL: [http://www.dnr.wa.gov/ResearchScience/Topics/GeosciencesData/Pages/gis_data.aspx](http://www.dnr.wa.gov/ResearchScience/Topics/GeosciencesData/Pages/gis_data.aspx)


**Utah:** Utah Geological Survey URL: [http://geology.utah.gov/maps/gis/index.htm](http://geology.utah.gov/maps/gis/index.htm)