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The Torrential and the Mundane: Climate Controls on Hillslope Weathering, Channel Bed Material, and Landscape Evolution in the Colorado Front Range

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The torrential and the mundane: Climate controls on hillslope weathering, channel bed material, and landscape evolution in the Colorado Front Range

Thesis directed by Professor Gregory E. Tucker

Climate shapes the surface of the earth through processes both torrential and mundane. I quantify the geomorphic impact of climate fluctuations in the Colorado Front Range by combining field work and numerical modeling on a variety of spatial scales. This research ranges from determining how water delivery controls the spatial development of saprolite on sub-alpine hillslopes, to how sediment delivery to streams causes channels to migrate across a landscape.

Through the delivery of water in snowmelt, climate should govern the rate and extent of saprolite formation in snow-dominated mountain watersheds, yet the mechanisms by which water flows deeply into weathered rock are largely unexplored. Measurements of snow pack thickness and soil moisture reveal strong contrasts between north- and south-facing slopes in both the timing of melt-water delivery and the duration of significant soil wetting in the shallow vadose zone. Results from a 2D numerical model of vadose zone dynamics suggest that thicker soil and more deeply weathered rock on north-facing slopes reflect greater moisture supply and weathering intensity.

Sediment that is produced on the hillslopes eventually arrives in stream channels. In the summer of 2013, I measured grain size, lithology, and channel geometry on several streams. Shortly after the conclusion of this field campaign, record-shattering rainfall caused severe flooding in the Front Range, including all of the study streams. Following the flood, half of the originally sampled sites were re-surveyed. This data set offers a unique opportunity to study empirically how a torrential flood event changes the size and composition of channel bed material and how the shape of the channel itself changes.

Modulation of sediment supply or transport capacity associated with climate change related to glacial-interglacial cycles has been suggested as a possible driver for the repeated aggradation
and abandonment of strath terraces that flank the Front Range. In this study, I use a landscape evolution model to determine whether changes in glacially driven sediment flux, changes in hillslope sediment flux or changes in transport capacity of the stream, in isolation or in combination, are sufficient to explain the observed rates and patterns of terrace formation and abandonment. The models indicate that i) in the absence of a large addition of sediment to the streams, variations in stream power are necessary to allow channel aggradation and the planation of bedrock surfaces, and ii) increased sediment flux from hillslopes is necessary to match observations of increased denudation rates during deposition of terrace-capping gravels.

Strath terraces are evidence of periods of time when the lateral erosion of bedrock channels dominated over vertical incision in bedrock channels. I present a physically-based theory for the lateral migration of bedrock channels and explore climate controls on lateral erosion rates and extent of valley widening. The model predicts that weaker bedrock results in wider bedrock valleys and more channel mobility, which is a fundamental factor for developing and maintaining a bedrock valley that is several times wider than the channel it holds. Increased channel mobility and wider flat bottomed valleys in the model under transport-limited conditions suggest that sediment cover on the bed is an effective way to slow vertical incision and amplify the effect of lateral erosion. This theory for the lateral erosion of bedrock channel walls and the numerical implementation of the theory in a catchment-scale landscape evolution model is a significant first step towards understanding the factors that control the rates and spatial extent of wide bedrock valleys.
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Chapter 1

Introduction: Climate controls on a multitude of geomorphic scales

1.1 What is climate control and how does climate control geomorphic processes

Climate controls the rates at which many, if not most, geomorphic processes sculpt the landscape. These processes range from the rapid and rare, such as glacial outburst floods or stream capture and drainage network reorganization, to the slow and commonplace, such as bedrock weathering and fluvial sediment transport. In this dissertation, the two most important drivers of geomorphic evolution controlled by climate are changes in the amount of water added to a landscape and changes in the amount of sediment added to a landscape.

A geomorphic system’s response to changes in the amount of water added to a landscape can be provoked through changes in rain intensity [Tucker and Bras, 2000], changes in spatial distribution of precipitation [Roe et al., 2002], or changes in precipitation phase such as, primarily snowfall to primarily rainfall [Nearing et al., 2004]. A landscape can respond to variation in runoff without changes in the magnitude of precipitation, through changes in evapotranspiration rate [Farmer et al., 2003], infiltration capacity of the soil [Horton, 1941], or vegetation distribution [Castillo et al., 1997]. Climate not only controls how much water is delivered to a system, but also how much sediment is available for transport in a system. Climate control is often invoked to explain increases in sediment supply through the manufacture of new sediment on hillslopes [Anderson et al., 2013], the re-mobilization of stored sediment in catchments [Church and Slaymaker, 1989], the more rapid removal of sediment from hillslopes due to decreased vegetation cover [Istanbulluoglu
and Bras, 2005], or more rapid downhill sediment transport due to climate influences on animal populations and habitats [Winchell et al., 2013]. Increases in sediment removal from hillslopes can also be attributed to climate-controlled changes in mass wasting [Fuller et al., 2009] or climate-controlled wildfire frequency and intensity [Pierce et al., 2004].

1.2 Approach: field work and numerical modeling of the critical zone

The research presented here was conducted in the Boulder Creek Critical Zone Observatory (CZO) in the Colorado Rocky Mountains. The critical zone is the near-surface layer extending from the groundwater to the tops of the trees, where bedrock is shaped and transformed by interactions among water, air, and living organisms. The Boulder Creek CZO includes the entire watershed of Boulder Creek and surrounding watersheds, but research within the watershed is focused on four intensely instrumented sub-catchments. The sub-catchments are located in distinct climate and geomorphic zones in the larger watershed and are instrumented with rain gages, snow depth sensors, soil moisture and water potential sensors, time lapse cameras, and groundwater monitoring wells, to name a few used in this research. The amount and variety of data collected in the Boulder Creek CZO make it an ideal location to develop and test hypotheses about the climate controls of geomorphic processes on a range of spatial and temporal scales.

The results presented in this dissertation were obtained through a combined approach of both field work and numerical modeling. Field work provides us with primary insights about how geomorphic systems function, helps hone our process intuition, and inspires creative thinking. Field work and measurements are also used in this research to provide the necessary data to support numerical models. Numerical modeling becomes desirable or necessary when the system that is studied becomes too complex to predict, for instance because it is too large, timescales are too long, or too many processes interact [Tucker and Hancock, 2010]. Models provide the virtual laboratories in which we can test hypotheses about the response of a landscape to a particular forcing.
1.3  Mundane climate controls on hillslope processes

I have studied climate controls on several spatial scales in the Front Range. Hillslopes are relatively small in scale, on the order of tens to hundreds of meters, but they are fundamental building blocks of landscapes. Understanding the processes that work at the hillslope scale is important for tying together the story of a landscape. The process of weathering fresh bedrock into sediment in the channel begins meters below the surface. Before a rock ever sees the surface of the earth, it already feels its effects and begins to weaken through chemical and physical weathering [Anderson et al., 2007]. In the Front Range of Colorado, the first step in breaking down bedrock is the transformation of bedrock into saprolite. Saprolite is chemically weathered rock that is augerable and often disintegrates at the touch, but still retains its rock structure. In dry environments like that of the Front Range, the chemical weathering of bedrock is limited by the amount of water that can reach the bedrock. In the current climate, water infiltrating from the surface rarely flows deeper than a meter or two, although this was not always the case. Deep-seated saprolite many meters below the surface occurs throughout the Front Range [Dethier and Lazarus, 2006], from the low elevation foothills (1600 m) to the high sub-alpine region (2500 m). How was bedrock chemically altered to such an extent? We see the evidence of altered bedrock, but do not understand how such a process could happen in the current climate. I present results on climate-controlled water flux on hillslopes and potential links to chemical weathering in Chapter 2. The next step in getting sediment to the channels is the detachment of saprolite and the downhill movement of its derivative: regolith. While my work does not focus specifically on the processes of how saprolite detaches and moves downslope, climate controlled processes are undeniably important in controlling the rate of hillslope sediment transport as well [Oehm and Hallet, 2005; Anderson et al., 2013].

1.4  Enigmatic climate controls on sediment supply to streams

Unless channel bed material is plucked directly from the channel bedrock, sediment carried by the channels is supplied from the hillslopes. Erosion rates of bedrock channels are controlled
by the extent of sediment covering the bed and the transport rates of the river that moves the eroded materials downstream [Gasparini et al., 2007]. Large flood events provide the stream power necessary to move the bed sediment and abrade the bedrock, and the size distribution of the bed sediment determines the transport threshold that must be exceeded in order to set the sediment in motion and start eroding the underlying bedrock [DiBiase and Whipple, 2011; Snyder et al., 2003b]. Characterizing spatial and temporal changes in bed sediment size and lithology gives insight into bedrock erosion rates and ultimately landscape response to changes in climate. In Chapter 3, I present the results of two field surveys of channel geometry and channel bed material of streams draining the Front Range. The second survey was conducted shortly after a historic flood event hit Front Range communities. This unique data set from before and after an extreme event allowed me to analyze how a single torrential event shapes the landscape.

While glacial sources of changing sediment flux are well studied [Hallet et al., 1996] and are often invoked as the primary source of increased sediment supply during glacial intervals, evidence from the Front Range shows that glaciers were not the only source of sediment that significantly impacted channel migration and terrace formation on the Front Range. Coal Creek and Lefthand Creek formed broad strath terraces during glacial intervals; however, these streams are unique in that neither of them shows evidence of past glaciation within their current watershed boundaries [Riihimaki et al., 2006; Dühnforth et al., 2012]. This suggests that climate processes related to glaciation and deglaciation, but not glaciers directly, caused the changes in sediment supply and/or carrying capacity of the stream necessary to switch from a transport-limited stream able to laterally bevel a valley hundreds of meters wide, to a detachment-limited stream that rapidly incised into the shale bedrock, and abandoned the terrace surfaces that were occupied for so long. In Chapter 4, I attempt to answer fundamental questions related to this effect. Can the channels shift from primarily terrace-forming to primarily incising through changes in sediment supply from the hillslopes alone? Or does the carrying capacity of the stream also have to change? What are the mechanisms that result in a smooth bedrock surface? Is the lateral migration of channel walls necessary? Or can large discharge events incise through the sediment cover to bite at the soft
bedrock, attacking from the top rather than from the side?

1.5 Torrential climate controls on channel migration

Understanding how a bedrock river erodes its banks laterally is a frontier in field-based studies, experimental studies, and modeling studies. If we understand the controls on the rate and extent of valley widening, we have made an important step towards understanding strath terrace development and gaining more insight into channel-hillslope interactions. Inversely, understanding of rates and extent of valley widening would allow us to finally use the large number of strath terrace sequences that have been mapped around the world to derive parameters for landscape evolution models. In a real channel, rates of lateral channel wall erosion depend on the shear stress directed at the channel walls and the resisting strength of the bedrock. Shear stress directed at the channel walls is a function of channel curvature and discharge magnitude [Stark et al., 2010]. The ultimate width of a bedrock valley is related to the lateral erosion rate and the duration over which the lateral erosion occurred [Suzuki, 1982]. Sediment supply in the stream plays an important role by providing tools to abrade the walls and cover to shield the bed from erosion [Sklar and Dietrich, 2004]. Both lateral erosion rate and the duration of the period where lateral erosion rates exceed vertical incision rates are often strongly tied to changes in climate, but can also result from changes in tectonic forcing. The development of wide bedrock valleys is also strongly linked to the erodibility of the underlying bedrock [Montgomery, 2004; Brocard and Van der Beek, 2006], which is not directly tied to climate, but can be influenced by climate through changes in weathering processes. Chapter 5 presents the results from one of the first applications of a physically-based theory for the lateral erosion of bedrock channel walls in numerical model on a drainage basin scale.

In my final chapter I will summarize the findings presented in this body of work and discuss some possibilities for future research.
Chapter 2

Evidence for climatic and hillslope-aspect controls on vadose zone moisture and saprolite weathering

Abigail L. Langston, Gregory E. Tucker, Robert S. Anderson, Suzanne P. Anderson

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2.1 Abstract

Through the delivery of water in snowmelt, climate should govern the rate and extent of saprolite formation in snow-dominated mountain watersheds, yet the mechanisms by which water flows deeply into weathered rock are largely unexplored. In this study we link rainfall, snow depth, and water content data from both soil and shallow saprolite to document vadose zone dynamics in two montane catchments over two years. Measurements of snow pack thickness and soil moisture reveal strong contrasts between north- and south-facing slopes in both the timing of melt-water delivery and the duration of significant soil wetting in the shallow vadose zone. To help interpret these observations, we use a 2D numerical model of vadose zone dynamics to calculate the expected space-time moisture patterns on an idealized hillslope under two wetting scenarios: a single sustained recharge pulse versus a set of short pulses. The model predicts that the duration of the recharge event exerts the strongest control on the depth and residence time of water in the upper unsaturated zone. Model calculations also imply that water should move more slowly through the subsurface and downward water flux should be substantially reduced when water is applied in several pulses rather than in one sustained event. The results suggest that thicker soil and more deeply weathered rock on north-facing slopes reflect greater moisture supply and weathering intensity. They also suggest how hillslope hydrology and chemical weathering rates may change as climate alters both the delivery of snow and the timing of its melt delivery to the subsurface.

2.2 Introduction

The mobile regolith that blankets nearly all of Earth's land area is largely derived from the underlying bedrock. Mobile regolith is often produced not from pristine bedrock, but rather from saprolite, in situ rock that has been chemically altered and mechanically weakened by chemical weathering. The weathering of rock to saprolite sets the stage for the production of mobile regolith and its subsequent transport downslope [Anderson et al., 2007]. The degree and rate of weathering in the saprolite also play important roles in the chemical weathering fluxes in a landscape and
the rate that mobile regolith is transported down slope [Dixon et al., 2009a]. Understanding the processes that transform bedrock into saprolite is therefore fundamental to understanding sediment production on hillslopes, soil development, and watershed biogeochemical cycles. The chemical reactions that produce saprolite are mediated by water. Therefore, the flux of moisture to the weathering zone should exert a first-order control on the rate of saprolite evolution. Models of bedrock weathering often assume a steady and uniform downward flux of water. One implication of these models is that, all else being equal, the potential for chemical alteration is greatest near the surface, where infiltrating water is furthest from chemical saturation. As water moves progressively deeper in the profile, it approaches chemical saturation and the weathering rate approaches zero [Brantley and White, 2009]. In semi-arid regions, however, water movement through the shallow subsurface is more complex than the simple picture of a steady, uniform downward flux suggests. If the near surface soil is relatively dry, the water input from a rainstorm or snow-melt event may be trapped in the upper few centimeters of the soil profile, and subsequently released back to the atmosphere through evapotranspiration. Consequently, the overall flow of water through the near-surface may be substantially greater than in deeper areas of the unsaturated zone. Moreover, the extent to which water can penetrate below the upper few tens of centimeters may depend on both the magnitude and the relative timing of rain or snow-melt events. One might hypothesize, for example, that brief pulses of water infiltration are less likely to penetrate deeply than a single, sustained pulse, which could produce higher near-surface moisture and create a higher effective hydraulic conductivity.

These complexities associated with water input raise some interesting questions regarding the role of shallow subsurface hydrology in saprolite genesis. What are the frequency and magnitude characteristics of water delivery to the soil-saprolite interface? How do water fluxes and peak moisture vary with depth? To what extent are space-time patterns of water delivery influenced by the timing of rainfall and snow melt? Can timing be significantly influenced by slope aspect? And is there a correlation between aspect, weathering extent, and water-input timing? Here, we seek to build a foundation for addressing these questions by (1) collecting and analyzing data
on snow depth, soil moisture, and shallow saprolite moisture in two semi-arid catchments in the Colorado Front Range, and (2) using a two-dimensional model of vadose-zone dynamics to explore the influence of water-input timing on space-time moisture patterns in the upper few meters of soil and saprolite.

It has long been known that the transformation of parent rock into a developed soil depends on a complex interaction of lithological, topographical, climatic, and biological factors [Jenny, 1941]. More recently, studies have recognized that the production rate and spatial distribution of saprolite, a transitional material between rock and soil, depend on a complex interaction of climatic, lithological, erosional, and tectonic factors (summarized by Brantley and Lebedeva [2011]). Locations with warm, wet climates experience higher rates of chemical weathering than cold or dry climates [White and Blum, 1995; White et al., 1998; Riebe et al., 2004]. Higher temperatures result in more rapid chemical weathering rates both in laboratory experiments and in natural systems [White et al., 1999; Millot et al., 2003], although access to water can limit the amount of chemical weathering that can occur in warm climates [Dixon et al., 2009b; Ferrier et al., 2012]. Recent studies on chemical weathering in soil and saprolite on hillslopes found a correlation between climate, especially precipitation, and chemical weathering rates in systems where the chemical weathering rate is limited by the mineral reaction kinetics [Dixon et al., 2009a; Rasmussen et al., 2011a; Ferrier et al., 2012].

2.3 Background

Several studies have used numerical modeling to explore the chemical evolution of bedrock to saprolite and mobile regolith (summarized by Brantley and White [2009]), and some modeling studies are beginning to acknowledge the direct role that hydrological processes play in chemical weathering [Lebedeva et al., 2007, 2010; Maher, 2010]. Maher [2010] recognized that chemical weathering rates are strongly dependent on the residence time and flow rate of water in the subsurface, with faster chemical weathering rates associated with faster fluid flow rates. Lebedeva et al. [2010] developed a model that included an erosion component and considered both advective and
diffusive solute transport for moving the chemical weathering front. They found that small changes in either fluid velocity or erosion rate can significantly impact the thickness of the weathering zone. One of the important underlying assumptions in the analyses of Lebedeva et al. [2010] is that the weathering zone experiences a spatially uniform downward flux of water. While this is a reasonable simplification in some landscapes, or for the exploration of model sensitivity, studies of vadose zone dynamics reveal a more complicated picture.

Slope angle, soil depth, and bedrock permeability all play a role in determining lateral and vertical flow paths in the hillslope subsurface [Hopp and McDonnell, 2009]. Antecedent soil moisture also exerts an important influence on the subsurface flow paths in hillslopes; only following the largest storm events do antecedent conditions cease to exert a strong control on subsurface flow paths [Woods and Rowe, 1996]. The questions of how much recharge infiltrates into the bedrock, how this deep seepage component scales from hillslope to catchment scale, and what role deep seepage plays in the hydrological function of hillslopes are being addressed [Graham et al., 2010], but integrating hillslope hydrology and chemical weathering remains a frontier that is largely unexplored.

Numerous studies have recognized that fractures in bedrock can play a major role in both water transport and the weathering by water beneath hillslopes in small, steep catchments [e.g., Montgomery et al., 1997; Anderson et al., 1997; Kosugi et al., 2006; Ebel et al., 2007]. Infiltrating storm water or snow melt that flows through the vadose zone weathers bedrock en route to the water table [Anderson et al., 2002]. Hillslope aspect can exert strong control on weathering rates in soils [Egli et al., 2006], on snow depth and soil moisture [Williams et al., 2008], and on ground temperatures [Anderson et al., 2013]. In the northern hemisphere, north-facing hillslopes tend to be colder and their snow packs more persistent, whereas south-facing hillslopes tend to be warmer and have shorter-lived snow packs.

To explore the impact of aspect and climate on the delivery of water into the soil and uppermost saprolite and potential chemical weathering in the saprolite and deeper subsurface, we use measurements of rainfall, snow depth, and water content in soil and uppermost saprolite to inform
a numerical model of unsaturated flow in a two-dimensional idealized hillslope. Characterizing differences in hydrology is a crucial first step for understanding differences in the development of the weathering front. In addition to matrix flow through the soil and upper saprolite, the model also includes fractures in the saprolite, which are potentially important avenues for routing water deep in the subsurface [Beven and Germann, 2013]. The goal of this theoretical analysis is to explore how the duration of recharge, independent of the magnitude of recharge, affects water fluxes and subsurface flow paths through the saprolite.

2.4 Field Site

The influence of variations in climate on subsurface flow paths and saprolite formation is studied in the Boulder Creek watershed, located in Front Range of Colorado (Figure 2.1a). The Boulder Creek watershed ranges in elevation from 1480 m to 4120 m and bedrock includes Precambrian Boulder Creek granodiorite and metamorphic sillimanite gneiss. The mean annual precipitation and percent of precipitation that falls as snow increase with elevation, while mean annual temperature decreases with elevation [Cowie, 2010]. We present rainfall, snow depth, and water content data from 2010 and 2011 that were collected in two small catchments within the Boulder Creek watershed. The Betasso catchment (Figure 2.1b), located at 1900 masl, receives 47 cm of precipitation per year and has a mean annual air temperature of 10 °C [Cowie, 2010]. This catchment is representative of a dry foothills montane ecosystem and is largely vegetated by ponderosa pine trees (*Pinus ponderosa*). Gordon Gulch (Figure 2.1c), located at 2600 masl, is an east-west trending catchment with mean annual precipitation of 55 cm and mean annual air temperature of 6 °C [Cowie, 2010].

The north- and south-facing hillslopes of Gordon Gulch offer an opportunity to explore the effects of slope aspect on the hydrology and geomorphology of the catchment. The hillslopes at Gordon Gulch are convex upward in profile and have a relatively thin mantle of soil (15–100 cm) overlying saprolite. The north-facing slope is densely forested by lodgepole pine trees (*Pinus contorta*) and retains a seasonal snow pack from late fall to mid spring. The south-facing slope is
vegetated by grasses, shrubs, and sparsely distributed ponderosa pine trees, similar to the Betasso catchment. The south-facing slope is generally free of snow during the winter, except for a few days immediately following snow events. The depth to the weathered bedrock and the thickness of augerable saprolite is greater on the north-facing slope of Gordon Gulch than on the south-facing slope. Soil is also thicker on the north-facing slope and the underlying saprolite appears more intensely weathered [Anderson et al., 2011]. Experiments to determine the tensile strength of rock cores recovered from 1-2 meters depth on the north- and south-facing slopes show that rock is weaker and presumably more weathered on the north-facing slope than on the south-facing slope [Kelly, 2012]. Shallow seismic refraction surveys in Gordon Gulch reveal that low velocities associated with saprolite extend to approximately 8 meters depth on the north-facing slope, compared to about 4 meters depth on the south-facing slope [Befus et al., 2011].

Drilling logs from groundwater wells drilled at Gordon Gulch (Figure 2.1c) offer a glimpse into the deep subsurface on the north- and south-facing slopes. These corroborate the shallow geophysical survey results, suggesting that the interface between saprolite and fresh bedrock is deeper on the north-facing slope. On the north-facing slope, the depth to the saprolite-fresh bedrock boundary is \(~14.6\) m whereas on the south-facing slope, the depth to this interface is \(~7.4\) m. Drilling notes also indicate that drilling in the bedrock on the south-facing slope was slow compared to the north-facing slope, suggesting that rock on the south-facing slope is less weathered.
Figure 2.1: a) Shaded relief map at 10 m resolution showing Boulder Creek catchment from the Continental Divide (4000 m) to the High Plains (1600 m). The inset maps show the two instrumented sub-catchments, b) Betasso catchment and c) Gordon Gulch. Instrument locations in each sub-catchment are shown. d) Cross section of Gordon Gulch with sensor locations on each slope indicated by arrows.
2.5 Data Collection and Analysis

Precipitation and water content data from the north- and south-facing slopes of Gordon Gulch and the lower elevation Betasso catchment allow us to explore climatic controls on hillslope hydrology. We analyze rainfall, snow depth, soil moisture, and matric potential measurements in these two catchments that were collected over a period of two years to quantify the magnitude and timing of water delivery to the subsurface.

2.5.1 Instrumentation

In order to determine how hillslope aspect affects soil moisture dynamics, soil moisture sensors were placed in a 300 m long transect that spanned the north- and south-facing hillslopes at Gordon Gulch (Figure 2.1d). Twenty-four soil moisture sensors (Campbell Scientific CS616) were placed in 12 locations at 5 cm below the surface and ~25 cm below the surface [Hinckley et al., 2012] (Figure 2.2a). Soil moisture was measured every 10 minutes. Snow depth, which can change significantly within hours in this sub-alpine catchment, was measured using 16 Judd sonic snow depth sensors that were placed in four locations along the same north-south trending transect. The sensors recorded snow depth with an accuracy of ±1 cm every 10 minutes.

Soil moisture sensors (Decagon EC-5) and matric potential sensors (Decagon MPS-1) were installed in the Betasso catchment (Figure 2.2b). Three sets of paired sensors were installed in vertical profiles at two locations (the Gully site and the Borrow Pit site), ranging in depth from 15 cm to 110 cm. Four of the six sensor pairs at Betasso were installed in saprolite. Snow depth was determined using five Judd sonic snow depth sensors located in the Betasso catchment. Again, snow depth was measured with an accuracy of ±1 cm every 10 minutes. Rainfall was measured and recorded every 10 minutes with a tipping bucket rain gage located on a ridge about 140 m and 330 m, respectively, from the two sensor sites.
Figure 2.2: Photographs of representative sensor sites at Gordon Gulch and Betasso. a) south-facing slope of Gordon Gulch prior to sensor installation. Arrows mark approximate location of sensors. Total depth of the pit is 25 cm. b) Gully Site at Betasso with soil moisture and matric potential sensor pairs shown. Total depth of the pit is 80 cm.
2.5.2 Snow Depth and Soil Moisture Observations

2.5.2.1 Gordon Gulch

To characterize how temporal variability of shallow soil moisture is influenced by snow melt in Gordon Gulch, we analyze soil moisture and snow depth data from two years of monitoring records. Snow depth and duration of snow cover is strongly controlled by hillslope aspect. During the 2009–2010 winter, the north-facing slope had a seasonal snow pack that reached a maximum depth of 70 cm in March 2010 (Figure 2.3a). During the 2010–2011 winter, the north-facing slope was covered in snow from late October 2010 to late April 2011 and reached maximum snow depth of 35 cm in February. The north-facing slope snow depth sensors measured snow pack depth of 5–10 cm from late 2010 to early 2011, which is consistent with weekly manual snow depth measurements. Late spring snow events up to 30 cm deep occurred in May of both 2010 and 2011. The south-facing slope received the same amount of snowfall as the north-facing slope, but snow cover generally did not persist more than several days, presumably due to higher solar radiation relative to the north-facing slope (Figure 2.3a). Maximum snow depth on the south-facing slope during the study period reached ~30 cm in February 2011 and the longest duration of snow cover on the south-facing slope was eight days during early 2011.

Soil moisture in Gordon Gulch tends to remain elevated for longer periods on the north-facing slope. Soil moisture tends to be more dynamic on the south-facing slope, with rapid increases in moisture followed by relatively rapid decreases (Figure 2.3b). Soil moisture on the north-facing slope experienced a rapid increase in response to melting snow packs in April 2010 and 2011 (Figure 2.3b). Soil moisture on the north-facing slope remained elevated during the early summer in response to rain events, but began to decline during early fall 2010. Soil moisture on the north-facing slope reached its lowest values during the winter of 2010–2011. Soil moisture on the south-facing slope was lower than on the north-facing slope, except during the winter. Like on the north-facing slope, soil moisture on the south-facing slope increased in response to early spring snow melt, but values dropped much more quickly than those on the north-facing slope (Figure 2.3b). By May of 2010, soil
moisture on the south-facing slope was generally low (~5–10%), but responded quickly to summer rain events. In February 2011, soil moisture increased sharply in response to the largest single snow-melt event of the study period on the south-facing slope. Soil moisture on the south-facing slope remained elevated during the spring, increased in response to two late spring snow events, and began to decrease during the summer months, reaching very low levels (~5%) by late summer 2011.

Figure 2.3: Snow depth and soil moisture measurements from representative sensors on the north- and south-facing slopes of Gordon Gulch. a) Continuous snow depth measurements show the snow pack on the north-facing slope that lasts from late fall to early spring and maximum snow depth of 68 cm. Most events on the south-facing slope are smaller and remain on the ground for shorter durations. b) Soil moisture measurements at 25 cm depth (sensor $GGL-NF-SP4-R4-CS616-25$) show that on the north-facing slope, soil moisture increased following spring snow melt and remained elevated through much of the summer and fall. On the south-facing slope, soil moisture (sensor $GGL-SF-SP9-R2-CS616-25$) increased following discrete snow melt events and declined rapidly during the summer and fall.
2.5.2.2 Betasso Catchment

During the study period (July 2010–January 2012), 60 cm of precipitation fell as rain at the Betasso Catchment and 55 cm of precipitation fell as snow (Figure 2.4a). The shallowest sensors in the Betasso catchment showed the most variability in water content and matric potential in response to precipitation events (Figure 2.4). Here, we refer to matric potential in terms of the absolute value of pressure, so that a large negative pressure is called a high potential and a small negative pressure is a low potential. Water content increased at the upper sensor following a snow event in February 2011. Matric potential at the upper sensor simultaneously decreased dramatically, indicating a decrease in water tension. Following this snow-melt event, water content in the upper sensor declined rapidly, but matric potential remained low and increased much more slowly. A rain and snow event in April 2011 caused water content to increase sharply at all of the sensors, but to a smaller magnitude at the lowest sensor. Matric potential fell to the same level at all of the sensors, ~10 kPa, following this rain and snow event. Over the next 30 days, water content declined rapidly at the upper sensor and remained steady at the middle and lowest sensors, while matric potential remained steady or increased slightly at the lowest sensor. A rain event in early May 2010 caused water content at the upper and middle sensors to increase sharply, but the lower sensor showed no response.

Two weeks following this precipitation in early May, a rain event in late May caused water content in the upper and middle sensors to increase further and caused water content in the lower sensor to increase to the highest recorded value during the study period. Matric potential remained very low through these events. Over a period of 10 days following the rain events in May 2011, water content declined and matric potential increased rapidly in the upper sensor and changed more slowly in the middle and lower sensors. A rain event in June 2011 caused rapid changes in water content and matric potential at the upper sensor, followed by rapid decay over 7 days. By July 2011, water content had decreased and matric potential had increased significantly at the upper sensor. Water content data are unavailable after July 2011, when the upper sensor malfunctioned.
2.5.2.3 Summary of Snow Depth and Soil Moisture Observations

Measurements of snow pack thickness and soil moisture in Gordon Gulch reveal strong contrasts between the north- and south-facing slopes. The snow pack on the north-facing slope accumulates all winter, then melts in one spring event, producing a deep, long-duration wetting event in the subsurface. The south-facing slope is intermittently covered by snow, which melts in the days following the event. As a result of these small, frequent melt events, soil moisture on the south-facing slope is more variable with time. Soil moisture and matric potential observations at Betasso show that longer periods of recharge are necessary to sufficiently wet the upper part of the profile and drive water deep into the subsurface. These results motivate modeling efforts in which we attempt to understand how differences in snow-melt timing influence the quantity and depth distribution of moisture in the soil and upper saprolite.
Figure 2.4: Precipitation, soil moisture, and matric potential measurements from the Betasso catchment. a) Daily averages of rain and snow depth measurements. Snow depth measurements were converted to SWE using a snow density of 333 kg/m$^3$. b) Water content measurements from the profile at the Borrow Pit. All of the sensors in this profile were installed in the saprolite, not the overlying soil. As the sensor at 40 cm depth malfunctioned in July 2011, data are unavailable after this date. c) Absolute value of matric potential measurements from the Borrow Pit profile. Matric potential sensors are paired with water content sensors.
2.6 Modeling Unsaturated Zone Flow

The aspect-related differences in snow cover, shallow soil moisture, and water content response in saprolite are used to guide two-dimensional unsaturated zone flow models on an idealized hillslope. These models inform us on the conditions necessary to route water deep into the saprolite through both its matrix and its fractures. The complex architecture of the unsaturated zone necessitates a model capable of capturing variable permeability to represent the different hydrologic properties in the soil, saprolite, and fractures. The two-dimensional distribution of water content in partially saturated porous media is described by the Richards equation:

\[
\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial x} \left[ K(\theta) \left( \frac{\partial \psi}{\partial x} \right) \right] + \frac{\partial}{\partial z} \left[ K(\theta) \left( \frac{\partial \psi}{\partial z} + 1 \right) \right]
\]

where \( \theta \) is volumetric water content (VWC), \( K \) is hydraulic conductivity [L/T], \( \psi \) is the capillary pressure head [L], \( x \) is the horizontal direction [L], and \( z \) is depth below the soil surface [L]. Computing two-dimensional flow dynamics in a medium with spatially and temporally varying \( K \), \( \theta \), and \( \psi \) requires a numerical solution. Here we used VS2DI (version 1.3), a program developed by the USGS for modeling flow in the unsaturated zone [Healy, 2008]. VS2DI uses a finite-difference method to solve the Richards equation for flow in a variably saturated medium. VS2DI uses the van Genuchten equations [van Genuchten, 1980] to describe the non-linear relationship between pressure head and moisture content and hydraulic conductivity. The relation between water content and pressure head is given by:

\[
\theta (\psi) = \theta_r + \frac{\theta_s - \theta_r}{\left[1 + (\alpha|\psi|)^n\right]^{1/n}}
\]

where \( \theta (\psi) \) is soil water content as a function of capillary pressure head, \( \theta_r \) is residual soil water content, \( \theta_s \) is saturated soil water content, and \( \alpha \) and \( n \) are parameters related to air entry suction and pore size distribution, respectively. The relation between conductivity and pressure head is given by:
where \( K(\psi) \) is unsaturated hydraulic conductivity as a function of capillary pressure head.

### 2.6.1 Model Setup

Our aim was to calculate the moisture dynamics that are characteristic of a setting like that of the Boulder Creek watershed, with convex upward hillslopes that receive most of their moisture from seasonal snow melt. To that end, the models are set up as an idealized hillslope that represents the lowermost portion of the hillslopes at Gordon Gulch. The two-dimensional grid was 5.5 m high by 10 m long with cell spacing of 4 cm. This spacing is fine enough to resolve the fractures and the soil-saprolite interface, while allowing reasonable computation time. The scaled-down model grid also allows us to evaluate two-dimensional flow dynamics near the stream as well as higher on the hillslopes within a geologically reasonable framework. A model on the scale of the hillslopes at Gordon Gulch (~300 m in length) is not necessary to capture the unsaturated flow dynamics in the top several meters.

In the model, 70 cm of soil overlies several meters of saprolite, which is cut through by three vertical fractures (Figure 2.5). Because the spatial distribution and connectedness of fractures and the relative importance of fracture flow in the shallow, highly weathered saprolite and in the deep, unweathered bedrock in the hillslopes is not known, a simple continuum model approach was appropriate for this study rather than a discrete fracture model [Glass et al., 1995; Berkowitz, 2002]. Fractures in the model are simply meant to show how preferential flow paths can influence moisture distribution and potential chemical weathering in a hillslope. Spacing between fractures in the model ranges from 3 m to 0.3 m in order to capture the range of fracture spacing (7 m to less than 10 cm) measured in outcrops at the Betasso catchment [Dengler, 2010]. Each model fracture is represented by one column of cells 2 cm wide that extends several meters into the saprolite, starting at the soil-saprolite boundary. The fractures in the model have a porosity of 0.1, making
their effective aperture 2 mm. The fractures are thus approximated as highly conductive porous media reflecting the likelihood that fractures in saprolite will have rough walls and possibly contain rock fragments.

The hydraulic properties of the soil and saprolite are determined from a combination of laboratory experiments, field data, and previously documented values (Table 2.1). Soil cores taken from the four soil pits in Gordon Gulch where the soil moisture sensors were installed, were analyzed for various hydraulic properties [Hinckley et al., 2012]. Saturated hydraulic conductivity, porosity, and van Genuchten parameters from the soil core analysis were used in the model to characterize the hydraulic properties of the soil layer.

As detailed analysis of the hydraulic properties of the saprolite was not available, we make estimates for these values from field data and previously published values. Saprolite hydraulic conductivity is estimated from wetting front travel times between soil moisture sensors in saprolite at the Betasso catchment. Travel times of a pulse of water from two rain events between two sensors spaced 30 cm apart were 18.5 hours and 9.3 hours, yielding estimated hydraulic conductivities of 0.39
m/d to 0.77 m/d, respectively. This matches previously published values of hydraulic conductivity in weathered granite in the Idaho batholith, which range from 0 m/d to 1.7 m/d [Megahan and Clayton, 1986]. Hydraulic conductivity measurements derived from falling-head tests on weathered granite cores from Japan range from 0.015 m/d to 0.03 m/d, providing an estimate for matrix hydraulic conductivity [Katsura et al., 2009]. The saprolite matrix is assigned a saturated hydraulic conductivity on the lower end of the range of measured hydraulic conductivities in weathered granite (0.02 m/d), while the fractures are assigned a hydraulic conductivity on the higher end of the range (2 m/d) (Table 2.1). The sensitivity of model results to the assigned hydraulic conductivity value is tested by decreasing $K$ by one half and increasing $K$ by a factor of two and by a factor of four in additional model runs.

van Genuchten parameters for the saprolite were estimated by comparing the results of 110 model runs with moisture content data from sensors installed in saprolite at the Betasso catchment. These models were run using the assigned representative $K$ value for saprolite (0.02 m/d) and spanned a large range of plausible $\alpha$ and $n$ values. A one-year model calculation was set up in VS2DI using recharge inputs from the Betasso meteorological station for the year 2010. Model output points at which VWC was recorded were placed in the saprolite at the same depths as the soil moisture sensors in the Betasso Borrow Pit. VWC values at the three model output points in the 110 model runs were compared with the VWC data from the sensors over 200 model days that included three precipitation events and a drying period of 100 days. The combination of $\alpha$ and $n$ parameters that best matched the data was chosen for the saprolite van Genuchten parameters.

<table>
<thead>
<tr>
<th>Material</th>
<th>$K$ (m/d)</th>
<th>$\phi$</th>
<th>$\alpha$ (1/m)</th>
<th>$n$</th>
</tr>
</thead>
<tbody>
<tr>
<td>soil</td>
<td>7.6</td>
<td>0.55</td>
<td>5.35</td>
<td>1.31</td>
</tr>
<tr>
<td>saprolite</td>
<td>0.02</td>
<td>0.2</td>
<td>0.75</td>
<td>7</td>
</tr>
<tr>
<td>fracture</td>
<td>2</td>
<td>0.1</td>
<td>0.75</td>
<td>7</td>
</tr>
</tbody>
</table>

The fractures cutting through the saprolite (Figure 2.5) are assigned the same $\alpha$ and $n$ parameters as the saprolite, and the hydraulic conductivity of the fractures is two orders of magnitude
larger and porosity was half of that in the saprolite. These values are chosen to capture the apparent increase in the hydraulic conductivity of saprolite with increasing scale due to preferential flow paths [Megahan and Clayton, 1986; Katsura et al., 2009].

A no-flux boundary condition is assigned to the upper hillslope side of the model. Seepage face boundaries are assigned to the bottom and lower hillslope side of the model. At seepage face boundary cells, total head is set to elevation head and fluxes are calculated. If the water fluxes are zero or directed out of the model, the simulation proceeds. If the calculated flux at a cell is directed into the model, the cell is set as a no-flux boundary [Lappala et al., 1987]. The horizontal seepage face produces a horizontal water table. Although a horizontal water table is not realistic, it is suitable for our purposes because the depth from the model surface to the water table is similar to that found in Gordon Gulch. For periods of snow melt, the top boundary condition is assigned as a specified flux boundary, at which recharge was applied at a given rate. For intervals between snow-melt events, the top boundary is an evaporation boundary and the potential evaporation rate is set to 0.001 m/d. This relatively low value for potential evaporation rate is chosen to represent potential evaporation during the late winter and spring months, rather than a higher value of potential evaporation during the summer months. In order to isolate the effect of the episodicity of recharge, the same potential evaporation rate was assigned to all model runs, despite differences in soil temperature and vegetation cover on the north- and south-facing hillslopes in Gordon Gulch. Because early spring snow-melt recharge precedes the onset of significant transpiration in this area [Sacks et al., 2006], transpiration is not included in the model calculations. The sensitivity of model results to the assigned evaporation rate is tested by changing the evaporation rate by a factor of two in additional model runs. Prior to beginning the model runs that are analyzed in this study, all models were run until they reached periodic steady state by applying 29 cm of recharge over 24 days followed by 322 days of evaporation for two cycles.
2.6.2 Gordon Gulch Models

The first set of models is driven by snow-depth measurements from the north- and south-facing slopes in Gordon Gulch. The two-year record of snow depth is converted into snow water equivalent by using the average late-spring density of the snow pack in Gordon Gulch. Measurements of snow density made in annual snow-pit surveys at Gordon Gulch indicate that the average snow density between mid-March and early May is 333 kg/m$^3$. Water is added to the model at the rate at which snow pack decreased. Six snow-melt events are applied to the north-facing slope model and 20 snow-melt events are applied to the south-facing slope model over two model years. Over the two-year model runs, the two hillslope models receive similar amounts of precipitation on the north- and south-facing slope models (59 cm and 54 cm, respectively), but the north-facing slope model experiences 515 days of evaporation while the south-facing slope experiences 636 days of evaporation (Table 2.2). Five model output points are placed in the model at various locations to record VWC and degree of saturation (Figure 2.5). The model output points are placed in the saprolite both to mimic field instrument set up and to understand the role played by fractures in routing water deep into the subsurface. Model output points are placed in the soil at 10 cm and 70 cm below the hillslope surface and in the saprolite at 90 cm and 145 cm below the ground surface and 15 cm away from the base of a fracture (Figure 2.5). In the northern hemisphere, north-facing slopes experience lower temperatures, less solar radiation, and are more likely to have a seasonal snow pack than south-facing slopes. Recognizing this, one can equate the terms used here, ‘north-facing slope’ and ‘south-facing slope’, with ‘pole-facing slope’ and ‘equator-facing slope’ (respectively) so that the terms are applicable to hillslopes in the southern hemisphere.

2.6.2.1 Idealized Models

We expect that the sunny south-facing slope would experience more evaporation in nature; the south-facing slope model experiences more days of evaporation because a seasonal snow pack is not present. To determine the extent to which subsurface flow paths depend on the timing of recharge,
we also conduct a set of idealized modeling calculations that differ in neither the magnitude of input fluxes nor the number of days of evaporation. The model domain, parameters, and output points for these models were identical to the Gordon Gulch hillslope models; the model experiments differ from each other only in the timing of the application of 58 cm of recharge (approximately equal to the recharge applied to the north-facing slope model). A concentrated recharge model scenario is run for two more years with one large pulse of water per year, applied over 24 days, at a rate of 0.012 m/d, followed by 322 days of evaporation. An episodic recharge model scenario is run for two years in which repeated cycles of recharge and evaporation are modeled, with four days of recharge, added at 0.012 m/d, followed by 24 days of evaporation. After six recharge-evaporation periods, 202 days of evaporation followed (Table 2.2).

Table 2.2: Recharge and evaporation summary for model runs.

<table>
<thead>
<tr>
<th>Model Run Name</th>
<th>total recharge (cm)</th>
<th>number of recharge events</th>
<th>days of evaporation</th>
</tr>
</thead>
<tbody>
<tr>
<td>north-facing slope</td>
<td>59</td>
<td>6</td>
<td>515</td>
</tr>
<tr>
<td>south-facing slope</td>
<td>54</td>
<td>20</td>
<td>636</td>
</tr>
<tr>
<td>concentrated recharge</td>
<td>58</td>
<td>2</td>
<td>644</td>
</tr>
<tr>
<td>episodic recharge</td>
<td>58</td>
<td>12</td>
<td>644</td>
</tr>
</tbody>
</table>
2.6.3 Model Analysis

2.6.3.1 Aspect Control

To understand how the contrasting snow-melt patterns on north- and south-facing slopes influence flow dynamics and water input to the saprolite, we use VS2DI to calculate flow in the unsaturated zone in response to the estimated timing and magnitude of the observed melt events at Gordon Gulch. The models of the north- and south-facing hillslopes show distinct patterns in water content and flow paths through the hillslope (Figure 2.6).

After all of the recharge for one model year is added to both the north- and south-facing slope models, the soil is saturated to a similar degree. However, the wetting front penetrates $\sim 3$ m into the saprolite on the north-facing slope and only penetrates $\sim 1$ m into the saprolite on the south-facing slope (Figure 2.6a). Moisture content in the saprolite on the north-facing slope is higher than that on the south-facing slope after the last recharge was applied. Ten days following the last recharge, water starts to drain from the upper saprolite in both the north- and south-facing slope models and move through the hillslope (Figure 2.6b). Ten days after the last recharge, the wetting front reaches the water table in most places on the north-facing slope, while on the south-facing slope, the wetting front is still $\sim 1.5$ m above the water table. In both models, the saprolite matrix near the fractures is drier than the rest of the saprolite, while the bases of the fractures are more saturated than saprolite at comparable depth. After 45 days of evaporation, the top several centimeters of soil and the $\sim 0.5$ m below the soil-saprolite boundary is significantly dried in both models (Figure 2.6c). In the north-facing slope model, the wetting front reaches the water table and the saprolite is $\sim 50\%$ saturated, while in the south-facing slope model, the wetting front is $\sim 0.5$ m from the water table and is $\sim 30\%$ saturated. At the end of the model year, after 225 days of evaporation, the soil is dried to the level of residual soil moisture, the only exception being at the base of the hillslopes where the soil intersects with the water table. Where the saprolite is above the water table, it is dried to the level of residual soil moisture in both hillslope models (Figure 2.6d). Water content at the top of the saprolite is lower than in adjacent soil because of
the difference in the $n$ parameter, which indicates a wider distribution of pore sizes in the soil.

Time series of model output at five points in the model domain show how modeled water content varies over the two-year model runs (Figure 2.7). Water content in the soil in both models increases to similar levels by the end of the recharge periods. However, water content in the north-facing slope model increases more rapidly and remains elevated for a longer period of time, while water content in the south-facing slope model varies more widely and takes longer to reach high levels in response to repeated wetting and drying. In both models, the response to recharge at the deep soil model output point (70 cm below the surface) is damped compared to the response in the shallow soil, and the deep soil dries more slowly than the shallow soil.

The water content at the three model output points in the saprolite varies between the north- and south-facing slope models. On the north-facing slope, water content in the shallow saprolite increases rapidly a few days after the deep soil is wetted (Figure 2.7a). Ten days after the sharp increase in water content in the shallow saprolite, water content increases rapidly in the deep saprolite, nearly equalling that in the shallow saprolite. The shallow saprolite dries more quickly than the deep saprolite, reaching its residual soil moisture value 100 days after the last recharge pulse, while the deep saprolite takes 200 days to dry to residual soil moisture.

On the south-facing slope, the water content in the shallow saprolite increases quickly, but less abruptly than on the north-facing slope (Figure 2.7b). Twenty days after the large increase in water content at the shallow saprolite model output point, water content in the deep saprolite increases, but less quickly than in the shallow saprolite. As on the north-facing slope, the shallow saprolite dries more quickly than the deep saprolite. The steady-state water content at the model output point near the base of a fracture is higher in both models than at other saprolite model output points because of proximity to the water table (Figure 2.7). Water content near the base of the corresponding fracture on the north-facing slope increases rapidly and remains elevated for 50 days following the last recharge event before slowly declining to the initial water content. Water content at the comparable south-facing slope site increases modestly and gradually, then declines gradually to the initial water content.
During periods of recharge, when both models showed maximum water content in the saprolite, the degree of saturation on the north-facing slope model is nearly double that of the south-facing slope model (Figure 2.8). The recharge pulse moves through the saprolite to the water table in both models, but water content is consistently higher and the water pulse moves more quickly through the north-facing slope (Figure 2.8a). In the north-facing slope model, the recharge pulse from the surface reaches the water table ~15 days following the last recharge event, compared to ~75 days in the south-facing slope model (Figure 2.8b).
Figure 2.6: Degree of saturation of the entire model domain for the north- and south-facing slope model runs based on recharge input inferred from snow depth records at Gordon Gulch. The model domains are the same size (5.5 m high, 10 m wide). The three vertical lines in the model domain are fractures in the saprolite. The water table developed 1 m above the bottom boundary and did not fluctuate significantly during the entire model period. a) Models on the last day of recharge in the first model year (model day 100) for both model runs. b) Model domains 10 days after the last recharge event. c) Model domains 45 days after the last recharge event. d) Model domains at the end of the first model year, 225 days after the last recharge event.
Figure 2.7: Degree of saturation recorded at five output points in the model domain. Bars at the top of each plot show recharge rate for the two model runs driven by precipitation data from Gordon Gulch.

Figure 2.8: Average degree of saturation with depth below the surface for the 84 columns closest to the hillslope ridge in the Gordon Gulch models. a) north-facing slope b) south-facing slope
2.6.3.2 Water-input timing

Data from the north- and south-facing slopes of Gordon Gulch show a clear contrast: north-facing slopes experience prolonged water input from melt of a seasonal snow pack, while south-facing slopes experience smaller and more frequent melt water inputs (Figure 2.3a). A similar contrast in melt frequency and magnitude occurs along an altitudinal transect as well: higher altitude sites have seasonal, sustained melt water pulses [Cowie, 2010]. It is of interest, therefore, to consider how differences in the timing of melt water input, independent of differences in amount, influences subsurface moisture patterns in a mountainous, snow-dominated catchment. To address this issue, we run a set of model calculations in which flow in the model hillslope is driven by recharge input in two different scenarios: a concentrated recharge model in which all recharge occurs during a single prolonged seasonal melt (representing a site with a seasonal snow pack) and an episodic recharge model in which the same amount of recharge is distributed among six melt events spaced 24 days apart (representing a site with periodic snow-melt or rain events). This approach allows us to isolate the role of recharge timing.

The degree of saturation of the model domain during the recharge and drying periods differs significantly between the concentrated recharge model and the episodic recharge model (Figure 2.9). After all recharge is added to both models (that is, at the end of the snow-melt season), the wetting front reaches approximately the same depth in the saprolite in both models, but the saprolite in the concentrated recharge model is saturated to ~90% while the saprolite in the episodic recharge model is saturated to ~40% (Figure 2.9a). Ten days following the last recharge event, water begins to drain out of the soil into the saprolite. In the concentrated recharge model, the saprolite near the fractures is drier than the surrounding saprolite at the same depth, because the fractures draw water in from the surrounding matrix (Figure 2.9b). Twenty-five days after the last recharge (Figure 2.9c), the top of the saprolite begins to dry in both models. While the wetting front reaches the water table in most of the domain in the concentrated recharge model, the wetting front is still ~1.5 m from the water table in the episodic recharge model. Forty-five days after the
last recharge event, the top of saprolite is similarly dry in both models, but the lower saprolite
in the concentrated recharge model is still $\sim 50\%$ saturated, and the wetting front in the episodic
recharge model has not yet reached the water table (Figure 2.9d). By the end of the model year,
the wetting front has reached the water table in both models and the soil and saprolite are dried
to approximately the same extent.

As before, five model output points are placed in the model domain to monitor the degree
of saturation throughout the model runs. In the concentrated recharge model run, water content
increases immediately following recharge in the shallow soil and increased in the deep soil after a
few days of delay (Figure 2.10a). The shallow saprolite, deep saprolite, and fracture saprolite model
output points all show a large, rapid increase in water content early in the model run, before the
recharge period ends. Following the recharge period, the soil dries more slowly than the saprolite,
and the deeper locations in both the soil and the saprolite dry more slowly than shallow sites.

In the episodic recharge case, water content in the shallow soil rapidly increases and decays
with each precipitation-evaporation cycle (Figure 2.10b). The wetting and drying response in the
deep soil is initially damped and out of phase with the shallow soil response. By day 90 of the
model run, the deep soil water content comes into phase with that in the shallow soil. On day 90 of
the model run, the wetting front reaches the shallow saprolite and water content increases rapidly.
Over the inter-storm evaporation period following day 90, water content in the soil declines little,
but moisture declines rapidly in the shallow saprolite. Despite the drying of the shallow saprolite
between recharge pulses, water content increases in the shallow saprolite with each recharge pulse
following day 90 of the model run. The wetting front reaches the deep saprolite after day 115,
increases slowly and responds to the remaining recharge events. Water content at the base of the
monitored fracture experiences a gradual increase, followed by a gradual decline. The maximum
saturation in the upper 3.5 m of saprolite in the concentrated recharge case is nearly double that of
the episodic recharge model (Figure 2.11). In the concentrated recharge model, the recharge pulse
tends to flow quickly through the saprolite as a large pulse with a distinct wetting front, while the
wetting front in the episodic recharge is more diffuse and moves more slowly. In the concentrated
recharge model, the wetting front is 3.5 m below the surface 25 days after the last recharge pulse. Water that is recharged at the hillcrest of the model, and therefore furthest from the water table in these simulations, reaches the water table at 4.5 m below the surface 30 days after the last recharge (Figure 2.11a). In the episodic recharge model, the wetting front is 2.5 m below the surface 25 days after the last recharge, and water recharged at the hillcrest does not reach the water table until at least 75 days following the last recharge (Figure 2.11b).
Figure 2.9: Degree of saturation of the entire model domain for the concentrated and episodic model experiments. The model domains same size (5.5 m high, 10 m wide). The water table developed 1 m above the bottom boundary and did not fluctuate significantly during the entire model period. a) Models on the last day of recharge in the model year. b) Model domains 10 days after the last recharge event. c) Model domains 25 days after the last recharge event. d) Model domains 45 days after the last recharge event.
Figure 2.10: Degree of saturation recorded at five output points in the model domain. Bars at the top of each plot recharge rate for idealized modeling experiments: a) concentrated recharge, b) episodic precipitation.

Figure 2.11: Average degree of saturation with depth below the surface for the 84 columns closest to the hillslope ridge in the modeling experiments. a) concentrated recharge b) episodic recharge
2.7 Discussion

2.7.1 Interaction between soil moisture and water flow in the saprolite

Gordon Gulch is an ideal location to explore the effects of variable recharge input on hillslopes because slope aspect dramatically influences the timing of snow pack melt. Ideally, data would be available from the top of the soil to the water table in order to definitively demonstrate the link between elevated soil moisture in the shallow subsurface and the flow of that water into the deeper subsurface. However, the consistently elevated water content in the soil on the north-facing slope during the spring suggests that any recharge events on the north-facing slope during this period would allow water to flow relatively effectively through the soil and into the underlying saprolite. Higher water content in the unsaturated soil or saprolite allows higher hydraulic conductivity in the medium and faster fluid transport. The wetting and drying on the south-facing slope caused by small, repeated snow-melt events, interspersed with periods of evaporation, results in lower water content and lower effective hydraulic conductivity, which impedes water flow deep into subsurface. Soil that dries during periods of evaporation must be re-wetted to some degree with each new precipitation event to allow water to flow, leaving less water to flow into the saprolite.

The soil moisture and matric potential sensors at the Betasso catchment sites were installed directly into the saprolite, unlike the sensors at Gordon Gulch. The magnitude of the precipitation event does not seem to exert primary control on the soil moisture response in the middle and lowest sensors; large precipitation events in the fall of 2010 and early winter of 2011 that caused increased water content at the upper sensor did not produce a response in the lower sensors (Figure 2.4). The water content and matric potential at the middle and lowest sensors responded only to precipitation events in the spring, after snow melt had saturated the upper parts of the profile and matric potential remained low at the upper sensor. In April 2011, the addition of new water when matric potential at the upper sensor was low, allowed water to flow deeper into the saprolite, both increasing the water content and decreasing the matric potential dramatically at the middle and lowest sensors (Figure 2.4b,c). Water content at the middle and lowest sensors increased very little
following a rain event in late June 2011, although matric potential was low at these locations (Figure 2.4b,c). Water content was low and matric potential was high at the upper sensor immediately prior to the June 2011 precipitation event. Therefore, the recharged water first had to re-wet the upper part of the profile before flowing deeper into the saprolite, but this recharge event was not large enough to allow flow deeper into the profile. Frequent wetting and drying inhibits fluid flow deep into the subsurface because the soil or saprolite loses moisture to evaporation between precipitation events and must be re-wetted following drying; a large or sustained recharge event is necessary order to drive water into the deep subsurface.

2.7.2 Impact of the Timing of Recharge

Jones and Banner [2003] show that the distribution of rainfall throughout the year, rather than the average annual precipitation, is the primary control on recharge rates in a tropical karst aquifer. Our results suggest that the timing of melt delivery is also important in mountainous, snow-dominated catchments. Model calculations based on precipitation data from the north- and south-facing slopes of Gordon Gulch imply that the seasonal pattern of snow melt on a hillslope can exert a significant control on subsurface flow paths, even when the magnitude of total snow melt over the year is similar. These models indicate that a large sustained recharge pulse, such as the melt of a seasonal snow pack, forces more water deeper into the saprolite matrix and the fractures than repeated, smaller recharge events. The contrast in concentrated recharge and episodic recharge models show that even when the magnitude of precipitation added to the model, and the number of days of evaporation applied, are the same, the timing of the recharge exerts a surprisingly strong control on the flow paths and degree of saturation in the deep subsurface.

Although the concentrated and episodic recharge models have the same total number of days of evaporation, the episodic recharge model loses 60% more water to evaporation over two years than did the concentrated recharge model (47 cm of evaporation vs. 29 cm of evaporation). Evaporation in the episodic recharge model occurs during several intervals when the soil was wet and water can be removed easily, while evaporation in the concentrated recharge model begins only
after all recharge is applied to the model. Mass-balance calculations from the model runs show that recharge that did not leave the model through evaporation flowed out of the base of the model. Fifty-eight cm of recharge is applied to both models over the course of each two-year model run. Thus, the concentrated recharge model shows an effective recharge (inflow minus evaporation) to the water table of 29 cm over the two-year run, while the episodic recharge model has an effective recharge of 11 cm over two years. Mass-balance calculations for the Gordon Gulch models show that the north-facing slope model has an effective recharge of 31 cm over the two-year model run, while the south-facing slope model has an effective recharge of 13 cm. These calculations show that using mean annual precipitation as the sole climate metric for water that flows through the subsurface can be misleading if the timing of the precipitation is not considered. The timing of precipitation exerts a strong control on how deeply into the subsurface water can flow and whether precipitation recharges the deep subsurface at all.

Not surprisingly, fractures in the saprolite allow water to flow more deeply into the subsurface than matrix flow alone. Where fractures are closely spaced, more water flows quickly into the deep subsurface (Figure 2.6, 2.9). In landscapes where fractures are closely spaced, flow in the unsaturated zone may be dominated by fractures rather than the matrix of the saprolite, resulting in faster and deeper flow through the unsaturated zone through these preferential flow paths. In landscapes with large fracture spacing, water will flow uniformly through the saprolite matrix where the saprolite matrix is permeable enough to allow water to flow. A granitic landscape with few initial fractures may not form a permeable saprolite. Water will tend to flow through the soil and along the soil-rock boundary when the recharge rate exceeds the infiltration rate of the rock [Weiler and McDonnell, 2007; Flint et al., 2008]. Prolonged contact of water and rock at the soil-rock boundary may form a weathered, more permeable layer in the rock. More flow through fractures may increase dissolution along these flow paths and increase the aperture of the fracture, resulting in a positive feedback between fluid flow and permeability. Increased water flow and potentially more chemical weathering near fractures suggests an intriguing feedback between permeability contrasts that route water in the subsurface and the development of permeability through chemical
weathering. On the other hand, precipitation of secondary minerals along preferential flow paths may decrease chemical weathering and permeability by coating minerals along the flow path or blocking the fracture to flow [Megahan and Clayton, 1986].

### 2.7.2.1 Model Sensitivity

Because there is a large range in published saprolite hydraulic conductivities [Megahan and Clayton, 1986; Katsura et al., 2009], the sensitivity of model results to the selected hydraulic conductivity, $K$, for saprolite is evaluated by running the concentrated and episodic recharge models with a range of plausible saprolite hydraulic conductivities. The models are run with the saprolite $K$ decreased by a factor of two from the base runs (0.01 m/d) and increased by a factor of two (0.04 m/d) and by a factor of four (0.08 m/d) from the base runs. Decreased saprolite $K$ results in higher saturation at saprolite model output points, higher average saturation with depth, and slower movement of the water pulse through the subsurface. Increased saprolite $K$ results in lower saturation in the models and faster movement of water through the subsurface, with the greatest effects seen in the concentrated recharge models. Mass balance calculations show that effective recharge to the water table in the model runs was not significantly affected by changes to saprolite $K$.

There are large fluctuations in evapotranspiration rate throughout the year and between the north- and south-facing slopes in Gordon Gulch. The sensitivity of model results to evaporation rate is evaluated by decreasing the evaporation rate by a factor of two and increasing the evaporation rate by a factor of two from the base run (0.0005 m/d and 0.002 m/d, respectively) for the concentrated and episodic recharge models. Model results for the concentrated recharge model are not sensitive to the evaporation rate. Saturation at model output points and average saturation with depth do not change significantly when evaporation rate was increased or decreased by a factor of two (Table 2.3).

The episodic recharge model is sensitive to changes in evaporation rate. The magnitude of the average saturation with depth increases slightly when evaporation is decreased by half, but the
Table 2.3: Effective recharge to the water table in model sensitivity analysis.

<table>
<thead>
<tr>
<th>Model Recharge</th>
<th>evaporation rate (m/d)</th>
<th>recharge (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>concentrated</td>
<td>0.0005</td>
<td>34</td>
</tr>
<tr>
<td>episodic</td>
<td>0.0005</td>
<td>27</td>
</tr>
<tr>
<td>concentrated</td>
<td>0.001</td>
<td>28</td>
</tr>
<tr>
<td>episodic</td>
<td>0.001</td>
<td>10</td>
</tr>
<tr>
<td>concentrated</td>
<td>0.002</td>
<td>22</td>
</tr>
<tr>
<td>episodic</td>
<td>0.002</td>
<td>-6</td>
</tr>
</tbody>
</table>

rate at which water flows through the model is greatly increased (Figure 2.12b). When evaporation is decreased, water reaches the water table ∼50 days faster than in the base model. Mass balance calculations show that when the evaporation rate is decreased by half, effective recharge to the water table over two model years nearly triples from 10 cm to 27 cm (Table 2.3). When the evaporation rate is increased by a factor of two, water flow through the saprolite becomes almost non-existent (Figure 2.12d). With a high evaporation rate, effective recharge to the water table becomes negative with more water leaving the model over two years than is added over the same time (Table 2.3).

Where evaporation rates are low, the importance of the episodicity of water influx to flow in the deep subsurface begins to decline and model results from concentrated recharge models and episodic recharge models become more similar. However, evaporation rates on south-facing hillslopes are likely to be higher than on north-facing hillslopes due to higher temperature, higher radiation, and in many mountain catchments in the western U.S., less shade. When evaporation rates are high, episodicity of water influx becomes much more important to flow in the deep subsurface. When the evaporation rate is high, a concentrated recharge pulse allows water to flow quickly into the deep subsurface and to be sequestered from the high evaporation rate at the surface. Episodic recharge pulses saturate the top of the profile, but that water is quickly removed by high evaporation before it can flow into the deep subsurface.
Figure 2.12: Average degree of saturation with depth below the surface in the model sensitivity runs. a,b) Evaporation rate was increased by a factor of two for the concentrated and episodic recharge runs, resulting in little change in saturation with depth for the concentrated model run, but a significant increase in saturation and flow velocity for the episodic run. c,d) Evaporation rate was decreased by a factor of two, resulting in little change in saturation in the concentrated recharge run and significantly decreased saturation and flow velocity in the episodic recharge model.
2.7.3 Implications for Chemical Weathering

Larger water fluxes through model hillslopes after a large, sustained recharge pulse may help explain differences in soil depth and weathering intensity between the north- and south-facing slopes observed at Gordon Gulch. The north-facing slope has deeper soil, more weathered saprolite, and greater depth to fresh rock compared to the south-facing slope [Befus et al., 2011; Anderson et al., 2011; Kelly, 2012]. Sustained elevated water content on hillslopes following a concentrated recharge pulse suggests that more chemical weathering may occur on hillslopes with a seasonal snow pack compared to hillslopes with intermittent snow. The melt of a seasonal snow pack allows water to flow faster and more deeply into the subsurface, where the transformation of rock into saprolite can occur. Water delivered to the soil or saprolite after a short recharge event moves more slowly through the subsurface, resulting in slower chemical weathering rates and a more shallow weathering front [Maher, 2010].

Geochemical modeling [Lebedeva et al., 2010] shows that the thickness of weathered material increases with pore fluid velocity, which is assumed to be proportional to mean annual precipitation. In recent studies, low chemical weathering rates have been attributed to lack of water in the subsurface [Dixon et al., 2009a; Ferrier et al., 2012]. Our model results suggest that chemical weathering rates are strongly controlled not just by the mean annual precipitation, but also by the duration over which this precipitation enters the subsurface. This finding is important for snow-dominated mountain catchments, and may imply a transition in behavior as the seasonal snow pack becomes more marginal.

2.7.4 Impacts of Climate Change

The results of this study have implications for how the rate and primary areas of bedrock weathering may change with a changing climate, and how rates and distributions of chemical weathering may have been different in the past. In the western United States, snow melt contributes a disproportionately larger amount to groundwater recharge than average annual precipitation
would predict [Simpson et al., 1970; Winograd et al., 1998]. A warming global climate is expected to increase the amount of precipitation that falls as rain, and to accelerate snow pack melting both in the Colorado Rockies [Rasmussen et al., 2011b] and throughout the Northern Hemisphere [Schlaepfer et al., 2012]. Climate change that results in decreased snow pack, even with constant annual precipitation, is predicted to result in decreased groundwater recharge [Earman et al., 2006]. Our model calculations suggest that reducing the area that is covered by a seasonal snow pack will result in reduced water flux to and through the deep unsaturated zone. This will therefore reduce chemical weathering rates in the saprolite and diminish groundwater recharge in small mountain catchments.

Decreasing the temporal or spatial distribution of the snow pack in the future may cause changes in chemical weathering in the saprolite and groundwater recharge. Conversely, increasing the area covered by a seasonal snow pack on a mountain landscape would result in more area with larger water fluxes through hillslopes. During the most recent glacial interval, the elevation of tree line was lowered by 500-800 m in the Colorado Rockies [Birkeland et al., 2003]. The fraction of the landscape with a seasonal snow pack would have been greatly increased. The increased area of hillslopes with a seasonal snow pack suggests that more water may have been routed through the saprolite during glacial intervals, allowing more weathering of the bedrock, despite lower mean annual temperature. Currently, when snow melts and recharges the underlying soil, the temperature of the water is slightly above 0 °C. This would have been the case for snow melt during glacial intervals, although the temperature of the rock would have been lower due to a decrease in mean annual temperature. Ferrier et al. [2012] suggest that in a kinetically limited weathering regime, temperature only weakly influences weathering rates if precipitation is also limited. That is, only when water is already available in the subsurface can temperature strongly influence the rate of chemical weathering.

Chemical weathering of the bedrock in areas in the Boulder Creek watershed that are covered by an intermittent snow pack, such as the south-facing slope of Gordon Gulch and Betasso, seems to be kinetically limited based on the presence of plagioclase grains in the upper soil. A decrease
in temperature and an increase in fluid flow through the bedrock at such locations may result in increased chemical weathering, although the presence of permafrost in hillslopes during glacial intervals would inhibit the flow of water. Dühnforth and Anderson [2011] suggest that the mean annual temperature during the last glacial maximum in the Boulder Creek watershed was at least 4.5 °C lower than present. This drop in temperature would suggest the presence of permafrost during LGM in many high-altitude catchments in the Boulder Creek watershed, and discontinuous or no permafrost in lower elevation catchments.

Physical weathering of bedrock may also have increased during glacial intervals through the growth of ice lenses in the bedrock [Walder and Hallet, 1985; Anderson, 1998; Hales and Roering, 2007; Anderson et al., 2013]. A limiting factor for the growth of ice lenses in the bedrock is the availability of water. Increased water content in the hillslopes during glacial intervals would supply water for ice lenses following the spring snow melt. Liquid water for frost cracking can also be accessed by upward capillary flux from the water table, which may have been closer to the surface than it is at present due to increased recharge and lower evaporation during glacial intervals [Hales and Roering, 2007].

2.8 Conclusions

The episodicity of recharge, independent of magnitude, can strongly influence water fluxes in the subsurface of mountain hillslopes. Data collected in the Boulder Creek watershed demonstrates that in locations with a seasonal snow pack (e.g., pole-facing hillslopes), soil moisture increases following the spring snow melt and remains elevated for several months. In locations with intermittent snow melt events (e.g., equator-facing hillslopes), soil moisture is more variable through time and water can only flow into the deeper subsurface on rare occasions when the shallow soil is sufficiently saturated.

Model calculations imply that the primary control on the speed and extent of water flow below the surface is the episodicity of the recharge. In a model scenario driven by a prolonged period of recharge, meant to mimic the melt of a seasonal snow pack, more water moves deeply through the
hillslope and recharges the water table. In a model scenario with the same magnitude of recharge spread out over several shorter-duration events, water moves through the hillslope more slowly and recharge to the water table is reduced by more than 50%. These findings have implications for both chemical weathering and water chemistry in the subsurface and changes in groundwater recharge in the past as well as the future.

Recent studies have found that water supply is of critical importance to chemical weathering rates in soils and saprolite [Dixon et al., 2009a; Maher, 2010; Rasmussen et al., 2011a; Ferrier et al., 2012]. Our results suggest that locations with a seasonal snow pack may experience more chemical weathering in the subsurface than locations with the same amount of snow melt spread over several events. Our results highlight a potential pitfall in using mean annual precipitation as the sole climate metric for evaluating the relationship between chemical weathering and climate. Mean annual precipitation alone is not a faithful proxy of how much water fluxes to the weathering front if the episodicity of the precipitation is not considered.

Our results suggest that during glacial intervals, when snow pack in the Colorado Front Range was likely deeper and more likely to melt during a single spring event, more water moved through hillslopes allowing more chemical weathering. On the other hand, sub-surface water flow in high-elevation catchments may have been impeded during glacial intervals if spatially continuous permafrost was present. Increased water inputs to the hillslopes may have also allowed more vigorous development of ice lenses that physically weather bedrock. More flux through hillslopes during glacial intervals would also suggest more groundwater recharge in sub-alpine mountain catchments. Reducing the fraction of the landscape covered by a seasonal snow pack will likely reduce water fluxes through the hillslopes, decrease groundwater recharge and slow rates of chemical weathering in the subsurface.

2.9 Acknowledgments

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project, especially Nathan Rock and Eve-Lyn Hinckley.
Chapter 3

Evaluating channel and bed sediment response in Front Range streams to an extreme flood event

Abigail L. Langston
3.1 Introduction

In this study, I seek to understand how channel geometry, bed sediment size, and bed sediment composition change in response to an extreme flood and how those changes varied with bedrock lithology. I present the results from two data sets collected before and after the historic flood that occurred in September, 2013. The first field campaign took place in the summer of 2013, when I measured channel geometry and conducted point counts of bed sediment rocks at 61 locations along five streams surrounding Boulder, CO. The original goal of this campaign was to better understand how channel properties and bed sediment properties change moving from a high energy environment in the resistant lithologies of the Rocky Mountains to a lower energy environment in the easily eroded lithology of the High Plains. Just one month after the summer field campaign ended, from September 11–14, 2013, Boulder and the surrounding communities in the Front Range were devastated by massive flooding produced by rain that the National Weather Service described as “biblical” in amount and intensity. After the flood, I repeated the measurements of channel geometry and bed sediment size and lithology at half of the original sampling locations. These two data sets provide a unique opportunity to determine not only how the channel and bed sediment adjust across a lithological boundary, but also how channel shape and bed sediment respond to an extreme flood event. Results indicate that downstream fining occurs at the study’s largest spatial scales of 10–30 km, but over shorter distances, grain size does not show a pattern in downstream fining. Grain size can coarsen downstream over some stream segments. After the flood, grain size in the streams generally coarsened, although some segments fined. The strongest response to flooding in the bed sediment characteristics was a distinct shift in bed sediment lithology towards the lithology distribution of the total watershed. Before the flood, bed sediment in most locations was primarily composed of granitic material. After the flood, the streams were more likely to have metamorphic, volcanic, or sedimentary rocks in the bed sediment, reflecting more input from more non-local hillslopes compared to before the flood.
3.2 Study Area

The Colorado Front Range, at the eastern edge of the Rocky Mountains, is shaped by streams flowing from the glacially sculpted Continental Divide, through steep, incising canyons, and onto the low-relief High Plains. The core of the Front Range in this area is made up of the resistant 1.7 Ga Boulder Creek granodiorite and an older sillimanite gneiss [Lovering and Goddard, 1950]. Volcanic outcrops in the mountain core are remnants of the ignimbrite flare-up, a period of high volcanic activity in the western United States about 40–25 million years ago [Tweto and Sims, 1963]. The Precambrian rocks were overlain by Paleozoic to Mesozoic-aged sedimentary units, primarily the Fountain Formation (300–280 Ma), Lyons Sandstone (280–250 Ma) and Dakota Sandstone (100 Ma) [Kellogg et al., 2008; Cole and Braddock, 2009], that were later uplifted and tilted, forming the iconic Flatirons over the skyline of Boulder. These sedimentary units are part of the resistant units through which streams draining the mountains must flow before emerging onto the low relief plains, which are underlain by easily eroded Cretaceous-aged sedimentary units, particularly the Pierre shale (Figure 3.1).

In the past ∼5 Ma, the erosion of ∼500 m of sediment from the High Plains caused base level of the streams to fall, causing the series of knickpoints that are currently making their way into the granitic core of the mountain range [Anderson et al., 2006]. The headwaters of many of the sampled streams were repeatedly glaciated during the Pleistocene and glaciers covered ∼10% of the watershed area during the last glacial maximum [Madole et al., 1998]. Glacial carving resulted in U-shaped valleys and relatively low channel slopes in the upper reaches of the catchment. Above the knickpoint on Boulder Creek, the landscape is a high elevation, gently rolling surface and channel slope remains relatively low. Below the knickpoint, the streams have carved steep canyons and channels. Locally the canyons can be as narrow as 50 m and as deep as 300 m. At the sharp transition from the core of the mountains to the adjacent High Plains, the streams move from the canyons onto the plains, slope decreases in the channels and the width of the flood plain increases dramatically.
3.3 Methods

In order to determine how channel properties and bed sediment characteristics change as streams flow from the mountains to the plains, I measured channel properties and conducted point counts at 61 locations along five creeks in the study area: Lefthand Creek, Fourmile Creek, North Boulder Creek, Middle Boulder Creek, and South Boulder Creek (Figure 3.2). Fourmile Creek and North Boulder Creek are situated entirely in the mountains and drain into Middle Boulder Creek. Both Boulder Creek and South Boulder Creek were dammed in the early-mid 20th century and have reservoirs that provide flood control and generate electricity. The canyon sections of the streams
are usually accompanied by paved roads beside the streams, possibly influencing stream geometry. Downstream, towards and in the plains, diversions and other water works, such as bridges, might influence stream geometry and flow regime.

Figure 3.2: Elevation map of study area, study streams, and sampling locations. Glacially carved valleys are visible in the west of the figure. A high elevation, low relief surface occurs at ~2500 m, around the highest elevation sampling points. Streams flow through deeply incised canyons just before flowing onto the gently sloping High Plains. Pre-flood sampling locations are shown in yellow dots and locations that were also sampled post-flood are shown in red dots. Watershed boundaries for Lefthand Creek and Boulder Creek are outlined in black.

The first field campaign was conducted during nine days of sampling in the period between July 26, 2013 and August 13, 2013. Sampling locations were spaced with fairly regular intervals of about 2 km, to capture most of the reaches of the channels within the Front Range and parts of the High Plains. Sampling locations were limited to places where reaching and crossing the creeks were possible, which has introduced sampling bias towards the wider and calmer parts of the creeks. The boundary between the Front Range and the High Plains was defined as the eastern boundary of the Fountain Formation sedimentary rocks (Figure 3.1). Near this boundary, sampling locations
were spaced at shorter intervals of 500 to 1000 m to more accurately capture the transition of the channels from a confined, mountainous environment to the relatively flat, open environment of the High Plains.

At each location, I measured channel width, bankfull channel width, slope, cross-sectional profile and the size and lithology of at least 100 randomly chosen rocks in the channel and on the channel banks. Measured channel width was the width of flow on the measurement day. Bankfull channel width was determined based on the level of permanent vegetation or channel bank scour [Whittaker et al., 2007]. Slope of the water surface was measured with a laser range finder over a distance of about 20–30 meters, centered on the sampling location. The bed sediment material sampled at each location was randomly selected using the Wolman pebble count method [Wolman, 1954]. The measured size classes ranged from $< 8 \text{ mm}$ to $> 300 \text{ mm}$ with half $\phi$ ($\log_2$) intervals. $\phi$ is a common unit in grain size analyses (Wentworth, 1922) and provides equal differences between size classes, based on consecutive halvings of the diameter. Five lithological classes were distinguished in the pre-flood survey: granitic, quartzite, metamorphic, sedimentary, and volcanic rocks. Before the flood, the sedimentary class was made up entirely of sandstone and conglomerates. These classes were chosen because they represent the major lithologies in the catchments and because they are readily distinguished in the field. No lithology was recorded for the particles $< 8 \text{ mm}$. 
3.3.1 Flood Event

Flood sirens along Boulder Creek began to sound on the night of Wednesday, September 11, and evacuation of local residents, including my neighbors, began that night. Most of the rain fell during a 36 hour period beginning the night of September 11; 27.5 cm of rain fell during this period at the meteorological station maintained by the Boulder Creek Critical Zone Observatory at Betasso (Figure 3.3a), and peak rain intensity was over 30 mm/hr. Annual exceedance probability for maximum rainfall amounts over 24 hours, 48 hours, and 7 days measured at the Boulder Justice Center are less than 1/1000, also known as a 1000-year rainfall event [Hydrometeorological Design Studies Center, 2013]. Before the flood, Colorado was experiencing a moderately severe drought and pre-flood discharge on Fourmile Creek and Boulder Creek was below normal for September (Figure 3.3b,c). Decreasing downstream discharge in Boulder Creek before the flood demonstrates the effect of water diversions from Boulder Creek needed to sustain local agriculture during the drought. Discharge in the streams increased rapidly in response to the continuous heavy rainfall. Within several hours on September 11, discharge in Boulder Creek in the canyon (Orodell) increased from 115 cfs to 980 cfs and peaked at 1520 cfs. Discharge in Boulder Creek farther downstream on the plains (75th St.) increased from 275 cfs to 5080 cfs over the same time period. The gage on Boulder Creek at the bridge at Broadway only gave reliable measurements up to 1050 cfs and did not record the peak flood discharge (Figure 3.3c). Fourmile Creek was the hardest hit of the study streams. Discharge on Fourmile Creek increased from 0.1 cfs at Orodell on September 9, to 1200 cfs on the night of September 12 (Figure 3.3b), turning a trickling mountain stream into a destructive torrent of water. The discharge data on Fourmile Creek is probably still somewhat reliable at least until September 13. After September 17, no data are available for the gage.

At the beginning of November, as soon as the flood waters retreated and it was safe to re-enter the streams, I repeated the channel measurements and point counts at 30 of the original 61 sampling locations. Instead of bankfull channel width, after the flood, I measured maximum flood channel width, and measured channel cross sections across this width. After the flood, I also
Figure 3.3: Rain and discharge data from September 8, 2013 to September 23, 2013. a) Hourly rainfall and cumulative rainfall from September 1, 2013, obtained from the meteorological station at Betasso maintained by the Boulder Creek CZO. b) Discharge on Fourmile Creek at Orodell, near the confluence with Boulder Creek from the gage maintained by the U.S. Geological Survey. There is no flood data for this gage after September 17. c) Discharge on Boulder Creek from gages at Orodell, above the confluence with Fourmile Creek (light blue line), Broadway St. bridge in Boulder (green line), and 75th St. in Boulder (pink line). The gages at Orodell and Broadway are maintained by the Colorado Division of Water Resources. The gage at 75th St. is maintained by the U.S.G.S.

recorded five lithological classes: granitic, metamorphic, volcanic, and sandstone as before, with the addition of shale and the removal of quartzite. In the pre-flood survey, quartzite was only found in South Boulder Creek, which was not re-surveyed after the flood. These measurements
from before and after a major flood offer a unique opportunity to learn how channel geometry, grain size, and bed sediment lithology change in response to a flood, as well as the opportunity to determine whether persistent features of the streams are caused by infrequent, extreme events.
3.4 Results

The long profiles of the five study streams show a transition in channel character from the mountains to the plains (Figure 3.4). Downstream of the mountain front, channels are low slope and concave up, while upstream of the mountain front, the channels are steep and cut narrow, deep canyons through the resistant crystalline mountain core. The long profiles reveal a series of knickpoints that have been moving up the stream channel since base level fall occurred in the Plio-Pleistocene [Dethier, 2001; Anderson et al., 2006]. In North and Middle Boulder Creek, the knickpoint has moved 15 km upstream of the mountain front, and in South Boulder Creek the knickpoint has moved 20 km upstream of the mountain front. There is also a knickpoint on South Boulder Creek 1 km upstream of the mountain front where the stream flows through a very resistant quartzite unit. Lefthand Creek shows a less obvious knickzone about 13 km from the mountain front and the knickzone on Fourmile Creek manifests as a subtle convexity about 11 km from the mountain front.

Before the flood, bankfull channel width in Boulder Creek increased moving downstream towards the mountain front, but decreased as streams move onto the plains, presumably due to water diversion for agriculture (Figure 3.5a). In the mountains of Boulder Creek, maximum measured flood channel width was only marginally greater than pre-flood bankfull width at all but one location, but flow depth during the flood was much higher. This indicates that in the mountains of Boulder Creek, the flood flow was accommodated by the existing channel, rather than spilling out onto a floodplain. On the plains, the maximum flood channel width was much greater than the pre-flood bankfull width in many locations, but only marginally greater in some locations, especially locations where the channel sides were stabilized by large boulders. In these locations where the channel was engineered, flood waters over-topped the channel and often resulted in flood waters spreading widely outside of the channel, for example at Eben G. Fine Park and the Boulder High School practice fields on the north side of the creek.

Maximum flood channel width on Fourmile Creek was much larger in all cases than pre-flood
bankfull width, and was five times larger at the farthest downstream location, 3.8 km upstream from the mountain front (Figure 3.5b). Where maximum flood channel width was closer to pre-flood bankfull, the channel did not have a floodplain, and flow was confined to the channel. Maximum flood channel width on Lefthand Creek was much greater than pre-flood bankfull width, often 3 times greater (Figure 3.5c). Where this was not the case, at the sampling location 6.4 km downstream of the mountain front, the shape of the channel was relatively unaffected, but the flood waters rose out of the banks and spread out over much of the surrounding area. This location was also just downstream of a bridge.
Figure 3.5: Pre-flood bankfull width and maximum flood channel width plotted against distance from mountain front a) Boulder Creek, b) Fourmile Creek, c) Lefthand Creek
3.4.1 Cross-sectional Shape

Where it was possible, measured cross sections from before and after the flood were compared in order to evaluate changes in cross-sectional shape and determine the extent of bed or wall erosion. As pre-flood cross section points were not surveyed or marked, post-flood cross section locations were selected based on GPS points, photos of pre-flood locations, and memory. On the plains section of Boulder Creek the most significant change in cross sectional shape was due to channel aggradation. In the months following the flood and the sampling period, sediment on Boulder Creek and Fourmile Creek was excavated from the channels in order to increase the conveyance capacity of the channels before the spring snow melt runoff [Meltzer, 2013]. Following the flood, a coarse cobble- and boulder-sized sediment bar was emplaced where there was formerly a deep pool at the sampling site closest to the mountain front on Boulder Creek (Figure 3.6a,b). At this sampling location on the plains section of Boulder Creek, the before and after cross sections were taken in the same location, and showed 1.5 meters of aggradation (Figure 3.6c). The channel narrowed from 19 to 9.5 m, the width of water flow on the day of sampling. This nicely demonstrates the reduced conveyance capacity of the channel and flood plain due to sediment deposition during the flood. The banks of Boulder Creek were well engineered, with large meter-sized boulders lining much of the channel through the city of Boulder (Figure 3.6a,b). Although flooding around Boulder Creek occurred when the stream jumped its banks, there was little measurable change of the channel banks in the study reach.

In most sampling locations on Fourmile Creek, the creek filled nearly all of its valley width, as opposed to channel width, during flood stage, causing massive damage to local homes and infrastructure. By the time it was safe to enter the Fourmile Creek area following the flood, major excavations in the creek and repairs to the road next to the creek had begun. At the pre-flood sampling location 5.5 km upstream of the mountain front, large boulders had been placed along the channel walls and the channel shape looked much as it did before the flood, with the exception that the vegetation had been completely stripped. This location was not measured or sampled
Figure 3.6: a) Pre-flood photo at the Boulder Creek sampling location closest to the mountain front. Photo is looking upstream. Before the flood, this location was a large, deep pool, lined with boulders on the north bank. b) Post-flood photo of same location looking upstream, showing bar of cobble and boulder-sized grains emplaced where there was formerly a pool. c) pre- and post-flood cross sections for Boulder Creek, plotted going from north to south. The boulders lining the bank were unmoved during the flood. Post-flood cross section shows \( \sim 1.5 \) m of aggradation at this location.
for grain size. In some re-sampled locations on Fourmile Creek, there was little obvious change to the channel shape, but in others there was evidence of bedrock channels being swept clean or large amounts of aggradation. All sampling locations on Fourmile Creek were largely stripped of vegetation that lined the channel before the flood, including large trees (Figure 3.7a,b). At the farthest upstream re-sampling location ~12 km from the mountain front, there was about 1.5 m of aggradation on the north side of the channel. The flood deposit consisted of a mix of fine and coarse grained material (Figure 3.7c) and in this location, I sampled 200 rocks instead of 100 to better capture the grain size distribution of the flood deposit and the channel material. In this location, the channel thalweg shifted ~4 m to the south, eroding both the channel bed and bank (Figure 3.7c), but flow width remained ~3 m, as it was before the flood.

On Lefthand Creek, the pre- and post-flood cross sections were taken 30 m apart at the sample location 1 km downstream from the mountain front (cross section 1), and comparison of the cross sections show that the channel had widened by ~3 m (Figure 3.8a). The heavily vegetated channel banks were totally stripped of vegetation by the flood, suggesting that the significant widening indicated by the cross section comparison is not solely the result of downstream variation in channel width. The pre- and post-flood locations for the sample location 1.7 km downstream of the mountain front (cross section 2) were taken 80 m apart. Despite this, the two measured cross sections are still compared in order to corroborate the lateral erosion of channel walls observed in the field, as evidenced by exposed shale bedrock on the channel walls (Figure 3.9). The comparison of the pre-and post-flood cross sections suggests a possible 1.5 meters of lateral channel erosion in this location (Figure 3.8b). At the sampling location 2.4 km downstream of the mountain front (cross section 3), large sections of the flood plain massively aggraded. Comparison of the pre-and post-flood cross sections gives an estimate of 1.6 m of vertical incision into the flood deposits (Figure 3.8c). There was little change in channel shape at the sample location 6.4 km downstream from the mountain front (cross section 4), and little observed evidence of the flood except for shale cobbles that were not observed in the pre-flood survey (Figure 3.8d). At the farthest sampling location from the mountain front on Lefthand Creek (cross section 5), there was little change in channel
Figure 3.7: a) Pre-flood photo of Fourmile Creek at the farthest upstream post-flood sampling location. Photo is looking upstream. b) Post-flood photo at farthest upstream post-flood sampling location, looking upstream. Photo shows total stripping of the vegetation, bank erosion on the south bank (left side of photo) and aggradation on the north bank (right side of photo). c) Close up of flood deposit at farthest upstream post-flood sampling location on Fourmile Creek. Flood deposits consist of angular, coarse grained material and well sorted fine grained material. d) Pre-and post-flood cross section plotted from north to south. Note that the post-flood cross section does not show the entire maximum flood channel width. Cross section shows erosion of south bank and 1.5 m of aggradation on the north bank.

shape, but evidence of flow over the wide flood plain (Figure 3.8e).
Figure 3.8: Pre- and post-flood cross sections for Lefthand Creek plotted from north to south. Cross sections are numbered 1-5 going downstream. Note the difference in x distance for each cross section. a) Lefthand Creek cross section closest to mountain front, 1 km downstream of mountain front. Possible lateral erosion of channel bank on north bank. b) 1.7 km downstream from mountain front. Lateral erosion on north bank, supported by observation of shale bedrock at this location. c) 2.4 km downstream from mountain front, flood plain aggradation. d) 6.4 km downstream from mountain front, little change in cross sectional shape. Possible vertical incision. Flood waters at this location likely spilled out of the channel into surrounding low-lying areas. e) 8 km downstream from mountain front, little change in cross sectional shape from before and after, but evidence of flow on the flood plain.
Figure 3.9: Lefthand Creek 1.7 km downstream of the mountain front (cross section 2), looking downstream before and after the flood. a) Lefthand Creek before the flood was a small mountain stream with densely vegetated banks. b) After the flood, most of the trees lining the banks were ripped out and bank cover on the north bank of Lefthand Creek was stripped to expose shale bedrock.
3.4.2 Grain Size Distribution

Pre-flood $D_{50}$ grain size in Boulder Creek decreased downstream, going from $D_{50}$ of 90 mm in the upper reaches to $D_{50}$ of 22.6 mm at the sampling location farthest downstream (Figure 3.10a). A rigorous statistical analysis of the patterns of downstream fining is conducted in Menting [2014]. Following the flood, $D_{50}$ in the mountain section of Boulder Creek increased in two locations ∼15 km upstream of the mountain front and decreased in one location ∼7 km upstream of the mountain front, but remained unchanged for the remaining sampling locations. Post-flood $D_{50}$ grain size increased for six of the ten sampling locations on the plains section of Boulder Creek, remained the same for three sampling locations, and decreased for one location (Figure 3.10a). On Fourmile Creek, $D_{50}$ grain size increased at four of the five sampling locations, dramatically in some cases, and remained the same at one location (Figure 3.10b). Pre-flood $D_{50}$ grain size in Lefthand Creek showed an unexpected downstream coarsening [Menting, 2014] and the post-flood $D_{50}$ was equally mysterious (Figure 3.10c). After the flood, $D_{50}$ grain size decreased at four of the five post-flood sample locations on Lefthand Creek and increased by four size fractions at one location.

Figure 3.11 shows the distribution of fine, medium, and coarse grained material for each sampling location before and after the flood. The fine fraction is made up of grain sizes from < 8 mm to 32 mm, the medium fraction is made up of grain sizes from 33 mm to 90 mm, and the coarse fraction is made up of grain sizes from 91 mm to 301 mm. The pre-flood grain size distribution in the mountain section of Boulder Creek was made up of ∼20–30% fine-grained material and ∼35–50% coarse grained material. Starting at the first sampling point after Boulder Creek leaves the mountains, fine grained material in the plains segment increased to ∼40–60% of the bed material and coarse grained material decreased to ∼5–25% of the bed material (Figure 3.11a). After the flood, the mountain section of Boulder Creek showed little change in distribution of grain sizes on the channel bed, except for a decrease in fine grained material in the upper reaches (Figure 3.11b). Post-flood, Boulder Creek retained the same size distribution of very coarse, medium, and fine grained material going from the lower mountain locations (∼5 km upstream of mountain front) to
Figure 3.10: Pre- and post-flood $D_{50}$ grain sizes plotted against distance from mountain front. a) Middle Boulder Creek, both mountains and plains section. b) Fourmile Creek c) Lefthand Creek

The plains for $\sim 4$ km downstream of the mountain front. As the coarse fraction unchanged, the overall coarsening in the plains segment reflects increased medium-sized bed material and reduced fine grained material. At the two sampling locations farthest from the mountain front, the medium size fraction decreased significantly compared to pre-flood grain size distribution and increased in fine grained material.

Grain size distribution on Fourmile Creek before the flood was generally composed of $\sim 40$–$60\%$ fine grained material and $\sim 20\%$ coarse grained material (Figure 3.12a). Following the flood,
the fraction of coarse grained sediment changed little, except at one location where the coarse fraction increased from 15% to 40% (Figure 3.12b). As in Boulder Creek, the fraction of medium sized sediment increased at nearly all locations at the expense of the fine fraction, which was reduced to $\sim$20–30% of bed material.

Lefthand Creek is again rather mysterious. Before the flood, the fraction of coarse grained sediment was quite high, around 35% in the downstream reaches. The fraction of medium sized grains was also around 35%, with the fine sediment taking up about 30%. Neglecting the farthest downstream sampling location, pre-flood grain size distribution on Lefthand Creek shows down-
Figure 3.12: Grain size distribution for Fourmile Creek divided into fine, medium, and coarse grained fractions. Coarse fraction of bed sediment is below the red line. Medium size fraction is between the red and blue lines. Fine fraction is above the blue line. a) pre-flood grain size distribution, b) post-flood grain size distribution.

Stream fining (Figure 3.13a). After the flood, the fraction of medium sized sediment remained about the same (25–30% of the total), but the coarse grained fraction decreased and the fine fraction increased (Figure 3.13b). The exception to this trend was at the sampling location 1.7 km downstream of the mountain front, where I sampled exclusively in the channel, not on the flood plain. It is interesting to note that the decrease in overall grain size at the sampling location 6 km downstream of the mountain front and the increase in grain size at the sampling location 8 km downstream of the mountain front are present before and after the flood.
Figure 3.13: Grain size distribution for Lefthand Creek divided into fine, medium, and coarse grained fractions. Coarse fraction of bed sediment is below the red line. Medium size fraction is between the red and blue lines. Fine fraction is above the blue line. a) pre-flood grain size distribution, b) post-flood grain size distribution
3.4.3 Lithological Composition of the Bed Sediment

The bedrock geology of Boulder Creek, Fourmile Creek and Lefthand Creek are composed of metamorphic, granitic, and volcanic lithologies in the mountains, and resistant sandstone and shale in the plains. The lithological composition of the bed sediment in Boulder Creek is a combination of contributions from upstream sources and local sources. I compare the lithology of the bed sediment before and after the flood to determine how composition of the bed sediment changes.

The mapped boundary between the metamorphic bedrock of the upper section of catchment and the granitic bedrock of the middle of the catchment is at Barker Reservoir, near the town of Nederland, CO (Figure 3.1). Metamorphic lithology composes 40–60% of the upstream lithology for the sampling locations in the mountains section of Boulder Creek, decreasing as the creek flows downstream through granitic lithology. Local hillslope sources (within 200 m) of the sampling locations on Boulder Creek consist entirely of granitic material, according to the geological map (Figure 3.1). It should be noted that the geologic map is not detailed enough to capture small scale variations in lithology that are present in the watershed. Before the flood, bed sediment lithology in Boulder Creek in the mountains locations was composed of 85–95% granitic material, with metamorphic material making up the majority of rest of the bed sediment lithology (Figure 3.14a). At the three locations farthest downstream in the mountain section, beginning at 6 km from the mountain front, volcanic lithology was first observed, composing ~5% of bed sediment material. It is appealing to attribute the appearance of volcanic material in the bed sediment to contributions from Fourmile Creek, which contains more volcanic material, but Fourmile Creek enters Boulder Creek 2.2 km above the mountain front so that only the last sampling location in the mountains is downstream of Fourmile Creek.

As Boulder Creek flows onto the plains, the pre-flood lithological composition changed significantly. At this point, Boulder Creek flows through 500 m of resistant sedimentary units, which can become part of the bed sediment. At the first three sampling locations downstream of the mountain front, granitic rocks continued to compose 80–90% of the bed sediment, with metamor-
phic and volcanic material making up the rest. One kilometer downstream of the mountain front, granitic material made up only 60% of the bed sediment, with a large spike in volcanic material. The first tributary to Boulder Creek in the plains is Gregory Creek, which joins Boulder Creek 1.5 km from the mountain front, downstream of where this change in lithology occurred. Moving farther downstream the decrease in the fraction of granite and increase in fractions of other lithologies continued: granitic material composed 50–70% of the bed sediment, metamorphic and volcanic materials both composed 10–20% of bed sediment, and sandstone increased slowly to make up 10% of bed sediment material at the farthest downstream sampling point.

Figure 3.14: Lithological composition of the bed sediment plotted as percent of total bed sediment against distance downstream for Boulder Creek. a) Pre-flood bed sediment lithological composition. b) Post-flood bed sediment lithological composition.

After the flood, the biggest change in the mountain sampling locations was in the upper
most section, where granitic material decreased to make up 60% of bed sediment and metamorphic material increased to make up 40% of bed sediment. Moving downstream into the canyon, bed sediment lithology again was dominated by granitic rocks, with metamorphic rocks composing the rest of the bed sediment. On the plains following the flood, the bed sediment composition did not change significantly compared to before the flood. As before, the bed sediment composition at the sampling locations on the plains was much more diverse compared to the mountain catchments. The most significant change in bed sediment lithology following the flood was the appearance of cobble-sized shale in the bed sediment beginning 1 km downstream of the mountain front. I observed many shale rocks on the bars and in the stream channel following the flood, although the point counts reflect shale making up a maximum of 5% of the bed lithology. The shale is likely very short-lived as bed sediment, because as soon as the shale is removed from the water, it begins to dry and crumble (Figure 3.15), and no shale was observed in the pre-flood survey Boulder Creek.

Fourmile Creek watershed is composed of granitic, metamorphic, and volcanic bedrock, as is Boulder Creek upstream of the mountain front, but the Fourmile Creek watershed has a larger mapped area of volcanic material. Before the flood, bed sediment lithology in the upper reaches of Fourmile Creek was comparable to the lithology distribution of the watershed, and became more dominated by granitic rocks moving downstream (Figure 3.16a). Granitic material increased from 60% of bed sediment lithology 21 km from the mountain front to 90% of bed material 3.7 km from the mountain front. In the upper-most section of Fourmile Creek, metamorphic material made up ~30% of the bed sediment, decreasing to 5% at the farthest downstream point. Volcanic material made up ~20% of the bed sediment in the upper-most sections, and also decreased to 5% at the farthest downstream point. Following the flood, six of the ten original sampling locations on Fourmile Creek were re-surveyed. Between 8 and 13 km from the mountain front, in the uppermost re-sampled locations, post-flood bed sediment composition did not change significantly from before the flood (Figure 3.16b). But in the downstream reach, between 3 and 7.5 km from the mountain front, post-flood lithology became more proportional to the distribution of lithology in the watershed compared to pre-flood composition. Granitic material was reduced to 75% of the
Figure 3.15: Shale cobbles found during post-flood survey on Boulder Creek. a) Shale cobble on Boulder Creek that is still quite cohesive. b) Shale cobble on Boulder Creek disintegrating after exposure to post-flood wetting and drying.

bed sediment lithology and volcanic and metamorphic material increased to 18% and 8% of bed sediment lithology, respectively.

In the upper reaches of Lefthand Creek, the pre-flood bed sediment lithology was generally
dominated by granitic material, with up to 30% contribution from volcanic or metamorphic rocks. 600 m downstream of the mountain front, granitic material decreased below 80% and the contribution from sandstone increased to 10–20% (Figure 3.17a). Contribution to bed sediment from metamorphic rocks remained constant at $\sim$10% from the mountains into the plains. After the flood, only five locations were re-surveyed, all of them on the plains section. After the flood the lithological composition of the bed sediment decreased in granitic material, going from 70–80% to 50–65% of bed material, and increased in all other lithologies (Figure 3.17b). Metamorphic and volcanic materials were slightly more abundant in the post-flood bed sediment lithology, but the largest increases were in sandstone and shale. Before the flood, sandstone composed 10–15% of
the bed sediment; after the flood, the amount of sandstone increased at every sampling point and made up 15–20% of the bed sediment. Before Lefthand Creek reaches the plains to the east of the mountains, it bends sharply to the north and flows on weak sedimentary rock between resistant sedimentary units for 5 km. This is 10 times the distance Boulder Creek flows through these units, which could explain why sandstone was far more abundant in Lefthand Creek than in Boulder Creek. Shale was not observed in the bed sediment during the pre-flood sampling campaign. After the flood, shale made up 3–5% of the bed sediment at four of the sampling locations, but was not sampled or observed in the location farthest downstream, 8 km from the mountain front. Shale cobbles and bedrock was observed on Lefthand Creek as far as Hover Road, 19 km downstream of the mountain front (S.P. Anderson, personal communication). Lefthand Creek was the only location where shale bedrock was observed at any point in the sampling period (Figure 3.9b), although later in the winter following the flood, I also observed shale bedrock along Boulder Creek.
Figure 3.17: Lithological composition of the bed sediment plotted as percent of total bed sediment against distance downstream for Lefthand Creek. a) Pre-flood bed sediment lithological composition. b) Pre-flood bed sediment lithological composition, close up of plains section from mountain from to 10 km from mountain front. c) Post-flood bed sediment lithological composition on the plains.
3.5 Discussion

Why does the diversity of lithology increase in all plains locations and in Fourmile Creek and Lefthand Creek following the flood? As streams move onto the plains, there is no longer a local source lithology for granitic, metamorphic, and volcanic rocks, as the streams flow over the shale bedrock. There are a few possible explanations for the decrease of granitic material in the bed sediment. One possibility is that the granitic bed sediment is abraded into finer and finer sizes as it is transported downstream, ultimately dropping below the identification threshold (8mm), resulting in a decreased representation of granitic material on the plains. This idea requires that the granitic material is more erodible than metamorphic or volcanic material since all bed sediment will be abraded as it is transported downstream. The source for granitic material is much closer to the plains than the source for metamorphic material, which is 25 km upstream from the plains segment of Boulder Creek (Figure 3.1). Although this study does not determine fining rates for individual lithologies, the granitic material is likely at least as resistant to abrasion as metamorphic material or volcanic material. Results from an experimental study on abrasion rates of various lithologies found that the abrasion rate for a sillimanite gneiss, like the metamorphic rock found in the study area, fined 3 times faster than intrusive granitic rock [Attal and Lave, 2006]. This makes the downstream abrasion of granitic rock and preferential preservation of metamorphic and volcanic rock an unlikely explanation for the increased lithological diversity observed on the plains. A more plausible explanation is that in the absence of steep granite canyon walls, and the associated high influx of exclusively granitic bed sediment, the stream returns to a bed sediment composition that is more representative of the lithology of the entire watershed. All of the metamorphic and volcanic material sampled on the plains segment must have passed through the mountains; at one sample location in Boulder Creek, 5 km upstream of the mountain front, the pre-flood bed sediment composition was not dominated by granitic material, but had significant contributions from volcanic and metamorphic material. Grain size of the bed sediment material fined as the streams flow onto the plains, and the transport capacity of the stream decreased. This results in
the preferential transport of finer material that has been abraded as it was transported downstream and the loss of coarse grained material from the bed sediment that has been abraded over a shorter distance.

As decreased fraction of granitic material in bed sediment after the flood in Fourmile Creek (mountains) and Lefthand Creek (plains) may have different causes, I will address each individually. After the flood, both grain size and the fractions of metamorphic and volcanic material increased at the farthest downstream sampling point in Fourmile Creek, so that the bed sediment composition at this location looked more like its upstream neighbors in terms of lithological composition, while the upstream locations did not change appreciably. In this case, the flood distributed sediment with the same characteristics along Fourmile Creek. This suggests that before the flood, the bed sediment delivered to the farthest downstream location on Fourmile Creek was small compared to the influx of granitic material from the canyon walls.

In Lefthand Creek, the majority of the increased diversity came from sandstone and shale lithologies, and smaller increases in metamorphic and volcanic material. Clearly the effect of the flood was to erode, carry, and deposit more sedimentary material in Lefthand Creek, suggesting that in normal flow years relatively small amounts of sedimentary material were eroded and transported. Given the amount of sediment carried by the flood waters, it is likely that the additional sandstone came from lateral erosion of the channel walls rather than vertical incision of the channel bed. Since it is known that the shale does not last long in the channel, how long sandstone persists in the channel following the flood is an interesting topic for further research.

3.6 Conclusions

This data set offers a unique opportunity to study the impact of an extreme flood event on the channel and bed sediment characteristics of three streams in the Colorado Front Range. Due to the inability to re-measure precisely cross sections from before and after the flood, conclusions of how the channel shape changed as a result of the flood are tentative. The engineering of many channel banks in the city of Boulder significantly decreased the chances of lateral migration in
the channel, to the good fortune of the residents of Boulder. Lefthand Creek was generally less engineered and lateral migration of the channel occurred in a few sample locations, but because of the distance between the before and after flood cross sections, I cannot provide a figure for the magnitude of lateral migration. The most significant observable change in channel form was due to large amounts of aggradation on the flood plains, which is a significant hazard itself. Aggradation on the flood plain reduced the conveyance capacity of the channel, potentially resulting in flooding during subsequent spring snow melt runoff.

The amount of information to be gleaned from the data set about bed sediment size and composition is limited by the lack of sites that were re-sampled following the flood. As winter rapidly approached, 30 locations were re-sampled during the month of November and into the first days of December. Following a hard freeze in early December, persistent ice in the canyons prevented continued re-sampling. The conclusions that can be reached from the sparse data set are further limited by the variability in data among neighboring sampling locations, especially on Lefthand Creek. However, future work could focus at higher scale levels, such as individual reaches or sedimentary links, where the data set conceivably allows an assessment of the relative importance of preferential transport and abrasion in causing downstream fining, and of the magnitude of hillslope contributions to the bed material. Menting [2014] took on this question for the pre-flood data set, suggesting that preferential transport is more important than abrasion in causing fining in the mountain sections, and that local hillslopes need to be over 30% steep before they become substantial sources of material to the channel.

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Chapter 4

Interpreting climate-modulated processes of terrace development along the Colorado Front Range using a landscape evolution model

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4.1 Abstract

Flights of terraces that flank range fronts throughout the Rocky Mountains record episodic stream incision over at least the past 1.5 Ma. Studies dating terraces in the Denver Basin along the Colorado Front Range suggest that these high surfaces were formed during glacial intervals and were rapidly incised and abandoned during interglacials. Modulation of sediment supply or transport capacity associated with climate change related to glacial-interglacial cycles has been suggested as a possible driver for the repeated aggradation and abandonment of these high surfaces. Potential mechanisms for increasing sediment supply and transport in rivers include variations over time in (1) sediment flux from intermittently glaciated major valleys, (2) the efficiency of hillslope sediment transport, and (3) the magnitude and timing of rainfall intensity and stream flow. Lower temperatures during glacial intervals may increase rates of both physical weathering and downslope transport of regolith through frost cracking and frost creep, respectively, thus supplying the upland rivers with sediment that can be delivered to the basins. Cold-period aggradation of the high surfaces in the basins and in mountain channels could be linked to enhanced sediment flux derived from glaciers in the headwaters of the catchment. Decreasing peak rain intensity from interglacial to glacial intervals can result in aggradation in streams, as the carrying capacity of the stream falls below the sediment load. In this study, we use a landscape evolution model to determine whether any of these processes, in isolation or in combination, is sufficient to explain the observed rates and patterns of terrace formation and abandonment along the Colorado piedmont. We also use the model to determine whether the presumed rapid incision rates of terraces during interglacials can be attributed to reduced sediment supply alone or whether changes in hydrology must also be invoked. We employ an idealized numerical catchment in which the upper half lies on resistant rock, representing the crystalline mountain range, and the lower half lies on soft rock, representing the adjacent sedimentary basin. Cycles of channel aggradation and incision are apparent in the models when either mean rain intensity is varied or a glacial source of sediment is added. Varying hillslope transport efficiency alone is not sufficient to produce significant trends in channel aggradation
or incision. The models suggest that i) in the absence of a large addition of sediment to the streams, changes in stream power are necessary to allow channel aggradation and the planation of bedrock surfaces, and ii) increased sediment flux from hillslopes is necessary to match observations of increased denudation rates that coincide with the deposition of terrace-capping gravels.

4.2 Introduction

Strath terraces record the incision history of rivers in response to variations in climate or base level. The mechanisms through which tectonic uplift can trigger incision of strath terrace surfaces are well understood [e.g. Merritts et al., 1994]. But it is unclear how climate fundamentally drives strath terrace formation and how strong climate forcings must be in order for strath terraces to develop.

Strath terraces are formed by lateral planation of a bedrock surface by a river, followed by rapid vertical incision by the river, leading to the abandonment this surface, that is often capped with fluvial sediment. In order to develop strath terraces, the ratio of lateral bedrock planation to vertical incision must change through time [Gilbert, 1877; Merritts et al., 1994; Hancock and Anderson, 2002]. Bull [1990] and Hancock and Anderson [2002] suggest that periods of lateral planation occur when bedrock exposure is limited by an inability of the stream to transport all of its sediment load. During these periods, the bed of the river is protected from vertical incision by sediment armoring of the bed, and erosion in the channel is focused towards the sides of the channel [Johnson and Whipple, 2010; Hancock and Anderson, 2002]. A transition from lateral bedrock beveling to vertical incision can occur when sediment supply decreases or effective discharge increases. When the carrying capacity of the stream again exceeds the sediment load in the stream, sediment on the bed of the stream is stripped, allowing renewed vertical incision into the bedrock and abandonment of the strath terrace. Lateral bedrock erosion and strath terrace development are more likely to occur in soft bedrock, such as shale or weakly indurated sandstone [Montgomery, 2004].

A number of processes have been invoked to explain the shifts between lateral erosion and
vertical incision that develop strath terraces. These include terrace abandonment due to meander cutoffs [Finnegan and Dietrich, 2011], transition from a single thread stream to a braided stream [Finnegan and Balco, 2013], increased sediment load from glacial sources [Hancock and Anderson, 2002; Brocard et al., 2003], increased sediment supply from intensified hillslope processes [Wegmann and Pazzaglia, 2002], increased stream power from changes in rain intensity [Hanson et al., 2006], and a combination of both changes in sediment flux and transport capacity [Pierce et al., 2011]. Because we do not fully understand how various climate processes alter the balance of sediment load versus stream power, different studies often invoke the same climate forcing mechanism and predict opposite results. For example, it is difficult to judge on the basis of intuition alone whether an increase in precipitation should produce incision due to increased hydraulic power or aggradation because of increased sediment yield from headwaters [Tucker and Slingerland, 1997].

The formation of strath terraces around the world has been tied to Pleistocene shifts between glacial and interglacial climate regimes, but various authors find that terraces form at the peak of glaciation [Molnar et al., 1994; Hancock et al., 1999], at the transition from a glacial to an interglacial climate [Pan et al., 2003], or during interglacial intervals [DeVecchio et al., 2012]. The widely varying ages of strath terrace formation can be attributed to the proximity of the terraces to glaciers and the different processes that influence sediment supply to streams in glaciated areas and unglaciated areas that feel the effects of climate shifts [Church and Slaymaker, 1989].

Few studies have modeled climate processes that result in strath terrace development in one-dimensional models [Hancock and Anderson, 2002]. Hancock and Anderson [2002] found that changing the inputs of sediment load and water discharge allows the stream to shift from primarily eroding laterally, to incising vertically, resulting in flights of strath terraces. Finnegan and Dietrich [2011] proposed flights of strath terraces can develop independently from variations in climate or tectonic uplift through meander cutoffs that incite knickpoint migration, vertical incision, and abandonment of the terrace surface. These previous modeling efforts laid a foundation for determining the mechanisms that drive terrace planation and abandonment, but did not address the basin-wide contribution of water and sediment to terrace genesis. In this study, we use a two-dimensional
landscape evolution model to determine whether climate-driven changes in sediment flux from a glacier, sediment flux from hillslopes, or mean rain intensity, either in isolation or in combination, are sufficient to explain observed rates and patterns of strath terrace formation and abandonment. This is the first use of a two-dimensional landscape evolution model to explore formation of strath terraces at the edge of a mountain range. We are able to address the relative impact of each of these climate drivers on cyclic channel aggradation and incision and lateral movement and valley widening of the channel.

4.2.1 Field Area

Flights of strath terraces flank the streams that drain the Colorado Front Range (Figure 4.1), recording periodic swinging between lateral bedrock erosion and aggradation of fluvial gravels, followed by vertical incision of bedrock and terrace abandonment. The terraces are located at the transition from the rugged mountain core to the adjacent High Plains. The mountains of the Front Range consist of a resistant core of Paleoproterozoic-aged granitic and metamorphic rocks, while the High Plains are underlain by easily eroded Cretaceous-aged Pierre shale and poorly lithified Tertiary sedimentary units.

We focus on terraces formed by three streams that flow from the mountains onto the plains: Lefthand Creek, Boulder Creek, and Coal Creek (Figure 4.1). One of the unique characteristics of these terraces is that they are present in watersheds of varying size. Boulder Creek has a drainage area of 1158 km² and has many levels of terraces that range from 5 to >100 m above the current streams. Lefthand Creek and Coal Creek are smaller watersheds (145 km² and 91 km², respectively), and have flights of terraces, including very broad terraces (Table Mountain and Rocky Flats), in their watersheds. The headwaters of Boulder Creek were repeatedly glaciated during the late Pleistocene, but there is no evidence for significant past glaciation within the current watershed boundaries for either Lefthand Creek or Coal Creek [Madole et al., 1998](Figure 4.1).

The terraces are capped with fluvial sands and gravels that were deposited simultaneously with the beveling of the bedrock surface. The sediments capping Table Mountain are 5–10 m thick
Figure 4.1: Hillshade image showing the three focus watersheds in the study area of the Denver Basin. The streams undergo an abrupt transition from the rugged, high-relief mountains to the low-relief plains. Light green shading in the high alpine reaches shows the glacial extent at the last glacial maximum [Madole et al., 1998]. Terraces on the low relief plains are colored according to their height above modern stream level, with darker colors indicating highest terraces and lighter colors indicating lowest terraces.

and are composed of coarse-grained material, up to 30 cm in diameter, in a matrix of fine sediment. The coarse fraction consists of granites, granodiorites, sandstones, and metamorphic rocks, reflecting sourcing of this material from throughout the Lefthand Creek catchment [Dülmforth et al., 2012]. The sediment cap on the Rocky Flats surface consists of 5–10 m of fluvial gravels and interbedded sands with a grain-size distribution similar to that of Table Mountain [Riihimaki et al., 2006]. Knepper [2005] found that the bedrock surface beneath the sediment mantle on Rocky Flats is undulating and reflects the shapes of the former stream channels, rather than a smooth bedrock
Studies dating the abandonment of several terrace surfaces suggest that rivers occupied the surfaces and cut laterally for long periods during glacial intervals and rapidly incised vertically during deep interglacial intervals. Riihimaki et al. [2006] used in situ $^{10}$Be and $^{26}$Al to date two sites from the alluvium capping the Rocky Flats surface. The ages of the sites range from 380 ky to 780 ky. Vertical profiles in the Rocky Flats surface show a complex fluvial history of deposition, incision, and re-deposition and that the bedrock was beveled by 2.4 Ma. These data suggest that the Rocky Flats surface was occupied by streams for hundreds of thousands of years over this interval.

Dünnforth et al. [2012] used both in situ and meteoric $^{10}$Be to date the abandonment of several terraces, including Table Mountain. They found an abandonment age for Table Mountain of 95 ky, which is younger by an order of magnitude than previously inferred from elevation relationships of the few dated terraces in the basin [Madole, 1991]. These young dates suggest that terrace treads were occupied and beveled for long periods before being abandoned following bursts of rapid incision, at rates of up to several mm/yr [Dünnforth et al., 2012]. Foster et al. [2013] used $^{10}$Be to date a lower-elevation terrace near Table Mountain and report a preliminary abandonment age for the lower terrace that is close to the abandonment age of 95 ky for Table Mountain reported by Dünnforth et al. [2012]. Foster et al. [2013] also calculated paleodenudation rates from cosmogenic radionuclide inheritance in the sediments capping Table Mountain. They report a basin-averaged paleodenudation rate during the deposition of the Table Mountain terrace sediments that is much higher than the modern denudation rate of 2.5 cm/ky [Dethier and Lazarus, 2006], perhaps up to 2–3 times higher than modern denudation rates.

Schildgen et al. [2002] found the youngest absolute ages of any terraces dated in the study area. They dated fill terraces within Boulder Canyon at >100 ky (15–20 m above channel), 32–10 ky (4–15 m above the channel), and Holocene age (< 4 m above channel) and argued for net excavation of sediment during transitions from glacial to interglacial intervals. They interpreted the decline in the height of the terraces as evidence for rapidly changing stream load and/or stream
4.3 Methods and Model Setup

We use the CHILD landscape evolution model [Tucker et al., 2001] to construct a scaled model landscape that includes the sharp transition from the high relief Front Range to the low relief High Plains (Figure 4.2). The model measures 800 m by 1600 m with 20 m cell spacing. This downscaling was required for computational efficiency at the necessary cell resolution. In order to model the effect of hillslope processes on sediment supply to the rivers of the Front Range, the cell size in the model needs to be no larger than about 20 m. We use a stochastic storm generation model [Tucker and Bras, 2000] that iteratively draws storm intensity, storm duration and interstorm duration from a Poisson distribution, given mean values of each. We initially set mean precipitation intensity to 10 m/yr (1.14 mm/hr), storm duration to 0.1 years, and interstorm duration to 0.9 years. The mean storm duration 0.1 years, or 877 hours, is about two orders of magnitude longer than typical storm lengths. In nature, individual storm events last on the order of minutes to hours [Eagleson, 1978], with some exceptional storms lasting days. The values for storm and inter-storm duration used in the model (0.1 years and 0.9 years, respectively) capture the seasonal variability in climate seen in locations where the majority of the yearly precipitation occurs during one season. Variability in the magnitude of precipitation occurs on inter-annual timescales, for example due to the El Niño-Southern Oscillation pattern, resulting in periods of higher or lower than normal precipitation for some regions [Ropelewski and Halpert, 1987]. Climate variability on centennial and millennial timescales in the Holocene resulted in changes in the frequency and magnitude of flood events in the American southwest [Ely, 1997]. Systematic climate variability during the Holocene on similar millenial timescales is also present in the central Rocky Mountains, resulting in higher flood magnitudes [Carson et al., 2007] and episodes of lateral beveling and sediment aggradation followed by incision [Pierce et al., 2011]. Thus, although the modeled storm-interstorm intervals substantially overstate the duration and timing of real storms, their mean values are compatible with seasonal variability and their extremes with interannual variability.
Figure 4.2: Model domain representing a scaled-down version of the crystalline core of the Front Range (high-relief area, \( y > 800 \) m) and the low-relief High Plains (low-relief area, \( y < 800 \) m). The color overlay shows grain-size distribution, with warm colors representing a high percentage of fine-grained material and the cool colors representing coarse-grained material. The coarse material is sourced from the mountains and has been spread on to the plains by the river as it moved across the plains. The river carries coarse sediment out of the mountains, and sediment in the stream fines as the river moves further away from the mountains. Location of sediment influx point for the glacial sediment model is indicated by a black circle, and location of cross sections is indicated by the white line.

The model consists of two types of material, bedrock and regolith. Regolith is produced from bedrock by weathering and by erosion and re-deposition of rock by fluvial processes. Two grain-size fractions are represented in the model, a coarse fraction (5 cm in diameter) and a fine fraction (1...
mm in diameter). In the model, both diffusive creep and fluvial erosion are modeled in all cells; however, for the purposes of this study, we consider convex areas that erode primarily through diffusive creep to be the hillslopes and concave areas that are eroded primarily by fluvial processes to be the channels. The fluvial erosion component only allows vertical incision and aggradation; there is no mechanism that allows the channel to erode laterally.

Evolution of regolith thickness, $h$, is described by the following mass-conservation equation:

$$\frac{\partial h}{\partial t} = P_0 e^{-h/h^*} + K_d \nabla^2 \eta - \nabla q_c$$

(4.1)

where $h$ is regolith thickness, $P_0$ is the maximum regolith production rate, $h^*$ is a scaling depth, $K_d$ is hillslope diffusivity, $\eta$ is the land surface elevation, and $\nabla q_c$ is excess transport capacity of running water. The first term on the right describes production of regolith from bedrock weathering, which assumes that bedrock weathers iso-volumetrically to regolith [Heimsath et al., 1997]. The second term describes the diffusion of regolith downslope, which moves as a function of hillslope curvatures and $K_d$, a coefficient that describes the efficiency of the downslope movement [e.g., Tucker and Hancock, 2010]. The third term describes the transport of regolith by fluvial processes.

Land surface elevation is described by:

$$\frac{\partial \eta}{\partial t} = U + K_d \nabla^2 \eta - F$$

(4.2)

where $U$ is vertical motion relative to a given baselevel, and $F$ is a fluvial erosion function.

The fluvial erosion function represents erosion rates from running water for both a transport-limited case and a detachment-limited case:

$$F = \nabla q_c \text{ where } D_c > \nabla q_c$$

$$F = D_c \text{ elsewhere}$$

(4.3)

The detachment capacity of bedrock is calculated as a power law function of shear stress with a threshold term:

$$D_c = K_b (\tau^{3/2} - \tau_c^{3/2})$$

(4.4)
where $K_b$ is the bedrock erodibility coefficient, $\tau$ is shear stress exerted on the bed, and $\tau_c$ is critical shear stress needed to detach a given grain size [e.g., Whipple et al., 2000].

The volumetric transport capacity of the flow for sand and gravel size fractions $[L^3/T]$ is given by:

$$Q_{cs} = K_f W f_s \tau^{3/2} (1 - (\sqrt{\tau_c/\tau}))^{1.5}$$
$$Q_{cg} = K_f W f_g \tau^{3/2} (1 - (\tau_c/\tau))^{1.5}$$

(4.5)

where $K_f$ is a transport efficiency factor, $W$ is width of the channel, $f_s$ and $f_g$ are the fractions of sand and gravel, and $\tau_{cs}$ and $\tau_{cg}$ are the critical shear stress for the entrainment of sand and gravel, respectively [Gasparini et al., 1999].

The model represents terrain as an irregular grid of cells with surface water flow routed from cell to cell following the path of steepest descent [Tucker et al., 2001]. The numerical approximation for excess transport capacity $[L/T]$ in a cell $i$ for a given grain size, $j$ is:

$$\nabla q_{cij} = Q_{sij} - \Sigma_{n=1}^{N} Q_{cnj} / \Lambda_i$$

(4.6)

where $Q_{sij}$ is total fluvial sediment flux of grain size $j$ into a cell and $\Lambda_i$ is the area of the cell. When $\nabla q_{ci}$ is negative, the stream does not have the capacity to carry its sediment load, and deposition occurs in a cell.

Total fluvial sediment flux into a cell is the sum of the volumetric erosion or deposition rate in each cell upstream of a given cell, where $N$ is the number of upstream cells that contribute water and sediment to cell $i$.

$$Q_{sij} = \Sigma_{n=1}^{N} F_{nj} \Lambda_n$$

(4.7)

The relief of the mountains in the model is created by assigning low bedrock erodibility ($K_b = 2.5 \times 10^{-4}$ kg/yr$^2$m) and by dictating that regolith produced in the mountains is 90% coarse-grained material (5 cm in diameter) and 10% fine-grained material (1 mm in diameter). The adjacent plains are assigned a much higher bedrock erodibility ($K_b = 1$ kg/yr$^2$m), and regolith produced on the plains is 90% fine-grained material and 10% coarse-grained material (Figure 4.2).
These parameters serve to represent the strong granodiorite core of the Front Range and the weak, friable shale bedrock of the High Plains. Long-term lowering of regional baselevel is represented by applying a steady, uniform rate of vertical motion (0.1 mm/yr) to the interior model nodes, while leaving the bottom boundary fixed at zero elevation. This configuration means that the model runs in the frame of reference of a steadily eroding surface (Figure 4.2).

We allow the model to evolve into a condition of roughly uniform erosion over 5 M model years, after which we run model scenarios to investigate whether any of the suggested climatic drivers, either alone or in combination, results in strath terrace development and abandonment. The climate perturbations in the model are created using 782 ky of a $\delta^{18}$O curve [Lisiecki and Raymo, 2005], as a proxy for either glacial sediment flux, hillslope transport efficiency, or rainfall intensity. In each case, the variable in question is assumed to scale linearly with the isotopic value. We explore the terrace-forming potential of (1) sediment flux from a point source, (2) variations in hillslope transport efficiency, (3) variations in rain intensity, and (4) a combination of changes in both hillslope transport efficiency and rain intensity. We also use a control model run, in which hillslope diffusivity and storm intensity are held constant and there is no sediment input from a point source, in order to distinguish between fluctuations that arise from extrinsic model forcing and those that reflect internal variability in response to stochastic rainfall variation.

### 4.3.1 Variations in Glacial Sediment Flux

In the first model run, we vary sediment influx at a point on the upstream end of the model (Figure 4.2) to represent sediment sourced from a glacier in the headwaters of the model catchment. Sediment flux from the point representing the glacial outlet is controlled by setting the slope and drainage area of this point; the sediment flux at that point for each storm is then set equal to the carrying capacity, $Q_c$, of the flow at the sediment inlet. We set the drainage area of the inlet to 4000 m$^2$ and allowed the sediment flux to change by varying the slope between zero and 0.74 following the $\delta^{18}$O curve. We use this range of slopes so that the mean sediment flux during peak glacial intervals equals $1.5 \times 10^5$ m$^3$/yr. This maximum sediment flux in the model was determined from
the area covered by glaciers during the Last Glacial Maximum in the Boulder Creek watershed (147.6 km$^2$) multiplied by a maximum erosion rate in the area covered by glaciers ($E_{gl}$) of 1 mm/yr [Ward et al., 2009] so that $Q_{sglacier}$ is $1.47 \times 10^5$ m$^3$/yr.

### 4.3.2 Hillslope Diffusivity

In order to investigate the role played by variation in sediment influx from hillslopes in downstream aggradation and incision, we vary the efficiency of hillslope sediment flux by changing the hillslope diffusivity coefficient, $K_d$. Published values of $K_d$ range from about $10^{-4}$ to $10^{-2}$ m$^2$/yr, with higher values of $K_d$ associated with colder or wetter climates [Oehm and Hallet, 2005]. Numerical modeling of soil creep through frost heave shows that hillslope diffusivity ranges from 0 to 0.04 m$^2$/yr, with most efficient transport when mean annual temperature is -6°C [Anderson et al., 2013]. $K_d$ in the model run varies with climate following the $\delta^{18}$O curve, with values ranging between 0 and 0.03 m$^2$/yr; $K_d$ in the control run was held constant at 0.015 m$^2$/yr. The imposed changes in hillslope diffusivity could represent any number of climate-controlled processes that change the efficiency of sediment transport down hillslopes, including changes in vegetation cover [Istanbulluoglu and Bras, 2005], changes in permafrost and overland flow [Bogaart et al., 2003], changes in temperature that cause faster or slower frost creep [Delunel et al., 2010; Anderson et al., 2013], or increased sediment transport from landsliding [Fuller et al., 2009].

### 4.3.3 Rainfall Intensity

In the final model run, we prescribe changes in mean rain intensity so that low rain intensity occurs during glacial intervals and high rain intensity occurs during interglacial intervals, but total yearly rainfall remains the same throughout the model run. Jarrett and Costa [1983] showed that rainfall-dominated catchments have the potential for much higher peak flood discharge than snowmelt-dominated catchments. Currently, streams draining the Front Range have characteristics of both snow-melt dominated and rainfall-dominated catchments, and are susceptible to large peak flood discharges as a result of intense rain storms. With an increase in snow coverage during glacial
intervals, these streams would tend to shift to more snow-melt dominated catchments. Carson et al. [2007] used reconstructed cross-sectional areas of abandoned channels in the Uinta Mountains, to show that peak flood discharge varied by as much as 15–20% for much of the Holocene, and that higher peak flood flood discharge coincided with warmer intervals during the Holocene. In order to represent such variations at the glacial-interglacial scale, we change the mean precipitation intensity from 10 m/yr (1.14 mm/hr) as in the control run, to precipitation intensity that varies from a cold-period minimum of 5 m/yr (0.6 mm/hr) to a warm-period maximum of 15 m/yr (1.7 mm/hr).

4.4 Results

We employ four metrics in the models to determine whether the applied climate perturbations result in the formation of terrace surfaces similar to those that flank the Colorado Front Range: (1) episodes of channel aggradation, which presumably coincide with long periods of terrace surface occupation, (2) rapid channel incision and surface abandonment, (3) increased paleo-denudation rates during terrace sediment deposition, and (4) broad flights of strath terraces along the modeled channels.

4.4.1 Variations in Glacial Sediment Flux

Adding a point that represents glacial sediment flux to the model results in cycles of aggradation and incision that smoothly follow the model forcing (Figure 4.3). Figure 4.3a shows mean channel elevation on the plains for the control model run (black) and for the glacial sediment flux model run (gray). Here mean channel elevation refers specifically to the elevation of the largest stream that flows across the plains (Figure 4.2). Channel elevation in the control run fluctuates between 1.5 meters below the starting value and two meters above the starting value over the course of the 782 ky model run. Many low-amplitude changes in sediment flux (such as from 450 to 500 ky and 670 to 710 ky in the model run) result in synchronous changes in channel elevation in the glacial sediment flux run, but the greatest magnitude sediment influx does not result in the greatest
magnitude of channel aggradation. For example, from 460 to 560 ky in the model run, sediment load rises, peaks, and declines. The peak sediment influx during this interval is not at the highest level during the model run, nor is higher sediment flux particularly long lasting during this interval. However, during this interval the channel aggrades to the highest level during the model run, up to 8 m above the control model value, and its elevation remains relatively high.

The depth of interglacial intervals directly influences the magnitude of channel incision, while the magnitude of channel aggradation due to sediment flux from the inlet is controlled by both glacial stage, through the slope of the inlet, and by the random duration and intensity of storms. During extended glacial intervals, the channel elevation shows up to 4 m of high-frequency variability over periods of a few thousand years. This pronounced variability in channel elevation only occurs when there is ample sediment supply; channel elevation is generally more steady during periods of incision.

Figure 4.4a shows the long-term rate of landscape lowering (i.e., denudation rate) averaged over the high-relief area of the model domain for the glacial sediment flux model and the control model. During periods of low and moderate sediment flux, denudation rates in the glacial sediment flux model are largely similar to denudation rates in the control run, about 0.1 mm/yr. During periods of high sediment flux, abrupt fluctuations in denudation rates in the mountains, ranging from 0 to 0.2 mm/yr, are the result of sediment filling the mountain channels, which decreases the slope between the channel and the hillslopes, slowing denudation from the hillslopes. The opposite occurs when mountain channels are flushed of sediment, resulting in a large pulse of rapid denudation. This effect is generally most pronounced near the sediment influx point.

Figure 4.5 shows cross sections of regolith and bedrock elevation across the model domain in the plains, 100 m beyond the edge of the mountains (Figure 4.2), for the glacial sediment flux run. Four different time steps are shown, demonstrating the effect of fluctuating glacial sediment flux on channel geometry, channel aggradation, lateral movement of the channel, and bedrock planation by the channel. Figure 4.5a shows the glacial sediment flux model at 658 ky into the model run, during a period of very low sediment flux. The valley is fairly wide (~80 m) and there is very
little sediment coverage in the valley. The channel has incised into terrace fill that is \( \sim 10 \text{ m} \) thick. After an extended period of high sediment flux, at 746 ky, the valley has not widened significantly, but the channel has shifted slightly and there is a continuous cover of sediment on the valley floor (Figure 4.5b). At maximum sediment flux at 758 ky, the bedrock valley is filled with \( \sim 5 \text{ m} \) of sediment (Figure 4.5c). At 782.4 ky, following a period of rapid sediment decrease over 18 ky, the thick blanket of sediment in the bedrock valley has been stripped away and the bedrock valley has widened by about 30 m (Figure 4.5d).
Figure 4.3: Mean elevation of the largest stream in the plains area of the model domain (Figure 4.2) for the model runs perturbed by climate changes (gray line) and the control model run (black line) divided by the beginning elevation of the control run. Dashed line shows climate forcing over the final 482 ky of the model run. a) glacial sediment flux, b) hillslope diffusivity, c) rain intensity, d) combined rain intensity and hillslope diffusivity
Figure 4.4: Mean denudation rates in the mountain area of the model domain for the model runs perturbed by climate changes (gray line) and the control model run (black line), along with the climate forcing (dashed line) over the final 482 ky of the model run. The control model run is in a state of quasi-equilibrium and mean denudation rate in the mountains is steady at $\sim 0.1$ mm/yr, equal to rate of uplift applied to the model. a) glacial sediment flux, b) hillslope diffusivity, c) rain intensity, d) combined rain intensity-hillslope diffusivity
Figure 4.5: Cross sections at y=700 m in model domain (Figure 4.2) for glacial sediment flux run at four different stages of sediment flux (a-d) and time series showing when model cross sections were taken (e). 

a) During periods of very low sediment flux, the valley is wide, with negligible sediment cover. 
b) Following an extended period of high sediment flux, the channel has shifted and the valley has filled with sediment. 
c) At maximum sediment flux, the valley is filled with \( \sim 5 \) m of sediment. 
d) Following rapid decrease in sediment flux, the valley is stripped of sediment and has widened by 30 m. 
e) Time series showing inlet slope (a proxy for sediment flux in the model) for the last 180 ky of the model run and when each cross section was taken.
4.4.2 Variations in Hillslope Transport Efficiency

Changing the efficiency of sediment transport on the hillslopes, $K_d$, from the control run value of 0.015 m$^2$/yr to values ranging from between zero and 0.03 m$^2$/yr has little net effect on aggradation and incision in the channel (Figure 4.3b). The channel has the potential to aggrade higher than the control model elevation during periods of high $K_d$, and to incise during periods of low $K_d$ (e.g., incision at 658 ky, Figure 4.3b), but decreasing $K_d$ does not always result in channel incision (e.g., little elevation change at 374 ky, 564 ky, and 586 ky, Figure 4.3b). Increasing $K_d$ tends to result in more variability in channel elevation, by up to 5 meters, during a cold period (500 ky - 540 ky model time) compared to a warm period (550 ky to 590 ky model time). There is no discernible lag between the forcing and channel response; if such lag is present, it is masked by the variability in the response signal.

Figure 4.4b shows mean denudation rate for the mountain area of the model domain for the model with variations in hillslope transport efficiency. Hillslope denudation rates follow the changes in $K_d$ very smoothly and without a lag. During periods of high $K_d$, when sediment is transported very efficiently down slope, denudation rates increase from 0.1 mm/yr up to 0.17 mm/year. In order for denudation rates to fall below the long term average, hillslope diffusivity must be very low, approaching zero; the lowest denudation rate for the model run is 0.05 mm/yr. In the final 20 ky time interval of the model run, the mean denudation rate is nearly three times higher during the peak of the final glacial interval compared to the denudation rate during the final deep interglacial.

4.4.3 Variations in Rain Intensity

In the rain intensity model, we change mean rain intensity from the control run value of 1.14 mm/hr to mean rain intensity that varies from 0.57 mm/hr to 1.71 mm/hr, while adjusting the storm duration so that the same amount of precipitation falls during the control run and the rain intensity model run. Note that the peak rain intensity from storm to storm can be significantly higher than the mean. Periods of low mean rain intensity during glacial intervals result in the largest magnitude
of channel aggradation seen in any of the model runs (Figure 4.3c). During periods of low mean rain intensity, the channel aggrades because there is less capacity to transport the sediment load. There is a threshold-like response in channel elevation to changes in rain intensity. For example, from 374 ky to 402 ky, rain intensity gradually decreases, but channel elevation remains low during the period. After rain intensity falls below ∼1.25 mm/hr, the channel begins to aggrade. Channel incision commences again when mean rain intensity rises above ∼1.25 mm/hr. The scale of channel aggradation is not very sensitive to the magnitude of the rain intensity, as long as it is below the threshold, nor does the magnitude of incision depend on the magnitude of rain intensity, as long as mean rain intensity is above the threshold. Channel elevation varies by as much as 5 m during periods of aggradation, as observed in other model runs.

Mean catchment-averaged denudation rates in the mountains for the rain intensity run are initially three times higher than the long term average (Figure 4.4c), due to the increased rain intensity that rapidly denudes the landscape. By 750 ky into the model run, mean denudation rates are nearly equal to the long term average and the model has returned to a state such that erosion is equal to baselevel lowering. Despite the departure from the long term average, it is apparent that during periods of low rain intensity, which correspond to glacial intervals, average denudation rates are lower than during interglacial intervals, which is contrary to evidence that paleo-denudation rates were higher during periods when terrace-capping gravels were deposited [Fuller et al., 2009; Foster et al., 2013; Dühnforth et al., 2012].

Figure 4.6 shows cross sections through the model domain at four different times, demonstrating the effect of changing rain intensity on channel geometry, channel aggradation, lateral movement of the channel, and bedrock planation by the channel. Figure 4.6a shows the channel at 696 ky into the model run, during a period of relatively high rain intensity. The valley is relatively narrow and V-shaped and there is very little sediment covering the bottom on the channel. Figure 4.6b shows the channel at 728 ky, during a period of low rain intensity. The valley has widened significantly and the channel has moved laterally by ∼50 m. Much of the valley is blanketed in 2-3 meters of sediment, but during this interval, the channel also occasionally bites down into
the bedrock, forming a wide, flat valley bottom. At 764 ky, during a period of very low mean rain intensity, the channel has moved again by 60 m and is completely filled with ~4 meters of sediment (Figure 4.6c). Rain intensity increases sharply over the next 18 ky, and by 782 ky, the channel has incised vertically into the bedrock, forming a narrow, V-shaped valley leaving paired, sediment-capped terraces 7 meters above the channel (Figure 4.6d).
Figure 4.6: Cross sections at y=700 m in model domain (Figure 4.2) for rain intensity run (a-d) and time series showing when model cross sections were taken (e). Note that the landscape relief in the rain intensity run is much lower than the glacial sediment flux run and the combined rain intensity - hillslope diffusivity run. a) During a high rain intensity period, the valley is narrow and V-shaped, with little sediment cover. b) Following a low rain intensity period, the valley has widened and the channel has shifted by 50 m. c) During a period of very low rain intensity, the channel has shifted again by 60 m, and the valley is filled with \( \sim 4 \) m of sediment. d) Following a sharp increase in rain intensity, the channel has incised vertically, leaving small, sediment-capped terraces 7 m above the channel. e) Time series showing rain intensity for the last 180 ky of the model run and when each cross section was taken.
4.4.4 Simultaneous Variations in Rain Intensity and Hillslope Transport Efficiency

In the final model run, we adjusted the climate with changes in both rain intensity and hillslope diffusivity. The opposing effects of lower hillslope diffusivity during periods with high rain intensity prevent over-smoothing of the model domain to the point that relief contrast between the mountains and the plains is negligible, as occurs in the rain intensity model (Figures 4.5, 4.6, 4.7). Peaks in channel aggradation correlate with peaks in hillslope diffusivity, and general periods of high channel elevation correlate with extended periods with rain intensity below $\sim 1.1$ mm/hr (Figure 4.3d). The periods of channel aggradation are shortened in the rain-diffusion model compared to the rain intensity model. The rain intensity threshold for aggradation is lower in the combined model, with aggradation only occurring when mean rain intensity falls below 1.1 mm/yr. The rain intensity-diffusion model has less variability during incisional periods than the rain intensity model run, but is equally variable in the aggradational periods.

The combined rain intensity-hillslope diffusion model run has long periods of elevated denudation rates during glacial periods, with the catchment-averaged denudation rate up to 1.9 times higher during glacial intervals than during interglacial intervals. Figure 4.4d shows average catchment-wide denudation rates in the mountains for the combined rain intensity-diffusion model run. Denudation rates for the combined model run smoothly follow changes in hillslope diffusivity, except when $K_d$ is very low and rain intensity is very high. During these periods of low diffusivity and high rain intensity, denudation rates for the combined run flatten out at or above the long-term average rate and do not fall below the long-term average, as in the hillslope transport efficiency run (Figure 4.4b,d). This is because during these periods, the higher rain intensity increases fluvial erosion, balancing the effect of very low or zero hillslope transport through diffusion.

Cross sections in the plains for four time steps in the combined rain intensity/hillslope diffusivity model show that during periods of high rain intensity and low hillslope diffusivity (representing interglacial periods), the channel is narrow and V-shaped and there is negligible sediment in the channel (Figure 4.7a). Fourteen thousand years later in the model run, at 716 ky, rain intensity
has decreased, hillslope diffusivity has increased, and the channel has shifted and begun to erode the sides of the bedrock valley (Figure 4.7b). The bedrock valley walls in the model cannot be eroded laterally; in the model, the bedrock valley widens as the channel changes position through aggradation and avulsion, followed by renewed vertical incision in the new channel location. This process of channel aggradation, avulsion, and vertical bedrock incision results in effective bedrock planation and valley widening. Following 50 ky of low rain intensity and high sediment flux from the hillslopes, by 766 ky, the channel has widened the bedrock valley further to 80 m, and the valley is filled with 7–8 meters of sediment (Figure 4.7c). Over the final 18 ky of the model run, rain intensity increased while hillslope diffusivity decreased dramatically, resulting in vertical incision of the bedrock and the abandonment of one small sediment-capped terrace (Figure 4.7d).
Figure 4.7: Cross sections at y=700 m in model domain (Figure 4.2) for the combined rain intensity-hillslope diffusivity run (a-d) and time series showing when model cross sections were taken (e). a) During a period of high rain intensity and low hillslope diffusion, the valley is narrow and V-shaped, with negligible sediment cover. b) Following a shift to low rain intensity and higher hillslope diffusion, the channel has shifted and begun to erode the sides of the bedrock valley. c) During an extended period of low rain intensity and high hillslope diffusion, the valley has widened by 80 m and is filled with ∼7-8 m of sediment. d) Following a sharp increase in rain intensity and decrease in hillslope diffusion, the channel has incised vertically through the sediment fill and bedrock, leaving a small, sediment-capped terrace above the channel. e) Time series showing rain intensity (solid line) and hillslope diffusivity (dashed line) for the last 180 ky of the model run and when each cross section was taken.
4.5 Discussion

Three of the four models, glacial sediment flux, rain intensity, and combined rain intensity/hillslope diffusion, match the observation that periods of channel aggradation correspond to glacial intervals, while periods of vertical channel incision correspond with interglacial intervals. Only one of the model scenarios, the combined rain intensity/hillslope diffusion model, also shows increased denudation rates in the mountains during periods of aggradation, which is necessary to agree with observed higher paleo-denudation rates during the deposition of terrace-capping sediment [Fuller et al., 2009; Foster et al., 2013; Dühnforth et al., 2012].

The glacial sediment flux model has the appropriate pattern of long periods of channel aggradation and rapid channel incision (Figure 4.3a), but lacks a coherent signal of changes in denudation rates (Figure 4.4a). Increased denudation rates during glacial intervals do not occur during the glacial sediment model run, but glacial erosion on a real landscape could show increased rates of catchment-averaged denudation from rapidly eroding glaciers. This element of reality is not present in our model. Although the glacial sediment flux model shows channel aggradation and incision associated with the formation of strath terraces, this model cannot explain strath terraces along the western edge of the High Plains that are present in small watersheds that were not significantly glaciated during the last glacial maximum, such as Coal Creek and Lefthand Creek. The presence of strath terraces in these catchments and other unglaciated catchments in the American West [e.g. Hanson et al., 2006] calls for a climate-driven process that does not require direct input of sediment from glaciers [Anderson et al., 2012].

The hillslope diffusivity model does not show a coherent signal of aggradation and incision in the channels (Figure 4.3b), but does show denudation rates that are nearly three times higher during glacial intervals than during interglacial intervals (Figure 4.4b). Despite the increase in hillslope transport efficiency by a factor of three over the duration of the model run, the amount of sediment added to the stream is too small in relation to the transport capacity of the stream, $Q_c$, to cause significant channel aggradation. Aggradation and incision patterns in the detachment-limited
channels of the hillslope diffusivity model are largely controlled by the variability in rain intensity and storm duration, allowing the channels to fill with sediment during quiescent periods and then rapidly flush the sediment out during intense storms.

The rain intensity model shows aggradation and incision in the channels (Figure 4.3c) resulting from changes in transport capacity, $Q_c$. The periods of intense rainfall during this model run reduced the relief of the landscape by vigorously eroding the hillslopes as well as incising the channels, potentially allowing the channel to migrate to new positions and widen the bedrock valley (Figure 4.6). The rain intensity model shows decreased denudation rates during glacial intervals, due to lower rain intensity and decreased erosion of hillslopes from running water (Figure 4.4c). Denudation rates are higher during interglacial periods due to intense rainfall, which efficiently strips the hillslopes of available sediment. Despite the increase in sediment flux in the channels during periods of intense rainfall, vertical incision still occurs because the increase in transport capacity is proportionately larger. This pattern of decreased denudation rates during glacial intervals is the opposite of what is observed [Fuller et al., 2009; Foster et al., 2013; Dühnforth et al., 2012], leading us to conclude that variation in rain intensity is not sufficient by itself to explain the genesis of the terraces flanking the Front Range.

Combining the climate forcing mechanisms from the rain intensity model and the hillslope diffusivity model resolves the weaknesses of both models. The combined scenario matches three of the four markers of terrace development that we have identified: channel aggradation, channel incision, and increased denudation rates during terrace planation (Figures 4.3d, 4.4d). The combined rain intensity-hillslope diffusivity model can also explain the presence of strath terraces in catchments that have no evidence of past glaciation, such as Lefthand Creek and Coal Creek. The combined rain intensity-hillslope diffusivity model demonstrates that in the absence of a large influx of sediment, such as glacially sourced sediment, to a detachment-limited stream, the primary mechanism for channel aggradation is reducing the transport capacity of the stream.
4.5.1 Model Variability

We observe small, rapid variations around a long-term mean in channel elevation (Figure 4.3) and denudation rates (Figure 4.4) in all of the model runs, including the control run, which has no extrinsic forcing apart from stationary stochastic variability in rainfall input. We consider the variability we observe in the models on centennial to millennial timescales to be a reasonable representation of how natural systems operate. The dynamic fluctuation in the model between model output intervals (1200 years), requires us to observe patterns over several thousand years of the model runs to discern long term trends. This becomes especially important during aggradational periods when the ratio of sediment load to carrying capacity is high and the magnitude of variability is amplified. The variability in channel elevation driven by random storms of varying size suggests that large storms that drastically increase carrying capacity can result in episodes of significant channel incision, even during intervals of net channel aggradation.

Terraces in the Lefthand Creek watershed suggest rapid incision during a glacial interval. Table Mountain and a lower adjacent terrace are offset by 20 vertical meters (Figure 4.1) and have very close reported abandonment ages [Dünnforth et al., 2012; Foster et al., 2013]. The amplified variability in channel elevation we observe in the model during aggradational periods provides an explanation for the apparently rapid incision by Lefthand Creek at \( \sim 90-95 \) ky, during a period that should generally have been characterized by high sediment load in the streams, strath terrace beveling, and fluvial sediment deposition.

4.5.2 Mechanisms for Bedrock Incision

Several authors [e.g. Gilbert, 1877; Hancock and Anderson, 2002] attribute the increase in the ratio of lateral to vertical erosion rates required to create strath terraces to armoring of the river bed due to increased sediment supply. Our models demonstrate this effect and show that widening of channel valleys and the creation of bedrock surfaces can occur due to episodic channel aggradation, avulsion, and punctuated bedrock incision. During periods of increased sediment flux or decreased
stream transport, the model channels tend to fill with sediment, protecting the underlying bedrock from vertical incision. But they also avulse to occupy a lower position in the valley (Figure 4.6c). The stochastic nature of the storms applied to the model means that even during periods when the channel bed is protected by overlying sediment cover, large storms can occur that expose the underlying bedrock to incision, and potentially, produce lateral bedrock valley widening (Figure 4.6b). Our models show that a bedrock strath can be generated without an explicit lateral erosion rule. During long periods of aggradation, an avulsion of the channel allows brief episodes of bed scour and vertical bedrock incision. This concept of punctuated biting of the bedrock strath rather than continuous lateral planation of the bedrock may explain the undulating bedrock topography of the Rocky Flats surface [Knepper, 2005]; however, the lack of broad, drainage-spanning surfaces in the models suggests that this mechanism fails to explain fully the extensive strath terraces observed adjacent to the Front Range.

4.5.3 Model Limitations

A chief shortcoming of these models is their inability to create terraces that are broad enough to span between adjacent drainage outlets and that extend up to the edge of the crystalline mountain front, as they do along the Front Range (Figure 4.1). The theory outlined in model description provides no mechanism for lateral erosion. In the numerical model, the only way for a stream to move laterally is for the stream bed to aggrade above the channel banks so that water is rerouted by avulsion on the landscape. We hypothesize that lateral channel erosion is the essential missing ingredient needed to form broad strath terraces like those that flank many Rocky Mountain ranges. CHILD, like most other landscape evolution models, does not include a mechanism for lateral erosion of the bedrock channel, although there are several generations of landscape evolution models that capture meandering in alluvial rivers [e.g., Lancaster and Bras, 2002; Coulthard et al., 2006] and models that take adjustments in channel width into account [Turowski et al., 2008; Yanites and Tucker, 2010]. So far the only landscape evolution model that attempts to capture lateral bedrock erosion is the 1-D model of Hancock and Anderson [2002], who do so through an ad hoc
lateral erosion rule based on stream power and ratio of channel width to total valley width in order to explain the observation that beveling a bedrock strath requires much higher rates of lateral planation compared to vertical incision. Understanding the processes critical to lateral erosion and implementing them into landscape evolution theory and computational model remains a frontier in geomorphology.

4.6 Conclusions

Three of the four model scenarios driven by different climate forcings show vertical bedrock incision during periods of low sediment flux or intense rainfall, and aggradation during the opposite phases of a climate cycle. Increasing hillslope diffusivity by a factor of four did not produce cyclic aggradation and incision, because the amount of sediment added to the rivers was too small in comparison with the carrying capacity of the stream. Only the hillslope diffusivity model and the rain intensity-hillslope diffusivity models produced higher erosion rates in the mountains during glacial intervals, leading us to conclude that processes that increase hillslope sediment flux must play a role in increasing the sediment load that both promotes lateral valley widening and deposition of sediment on terraces during glacial intervals. An alternate explanation, that increased sediment flux during glacial intervals is the result of a wetter climate, appears to be insufficient to produce both channel aggradation and increased paleo-denudation rates. In nature, the unusual breadth of the strath terraces that flank the Front Range arises from efficient lateral erosion on easily eroded rocks of the High Plains. None of the model scenarios produced flights of broad strath terraces, suggesting that an explicit lateral bedrock erosion rule is indeed required to explain the remarkable breadth of these range-bounding terraces.

4.7 Acknowledgments

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Chapter 5

Developing and evaluating a theory for the lateral erosion of bedrock channels in landscape evolution models

Abigail L. Langston
5.1 Abstract

Understanding how a bedrock river erodes its banks laterally is a frontier in field-based studies, experimental studies, and modeling studies. Theory for the vertical incision of bedrock channels is widely implemented in our current generation of landscape evolution models. However, in general existing models do not seek to implement the lateral migration of bedrock channel walls. This is problematic, as modeling geomorphic processes such as terrace formation and hillslope-channel coupling depend on accurate simulation of valley widening. We have developed and implemented a theory for the lateral migration of bedrock channel walls in a catchment-scale landscape evolution model. In a real channel, rates of lateral channel wall erosion depend on the shear stress directed at the channel walls and the resisting strength of the bedrock. Shear stress directed at the channel walls is a function of channel curvature and discharge magnitude. Sediment supply to the stream, which provides tools to abrade the walls and cover to shield the bed from erosion, is a critical component of lateral channel migration. Two model formulations are presented, one representing the slow process of widening a bedrock canyon, the other representing undercutting, slumping, and rapid downstream sediment transport that occurs in softer bedrock. Modeling experiments were run with a range of values for bedrock erodibility and runoff and varying magnitudes of sediment flux and water discharge in order to determine the role each plays in the development of wide bedrock valleys.

Results from modeling experiments show that this simple, physics-based theory for the lateral erosion of bedrock channels produces bedrock valleys that are several model cells wide. The model predicts wider bedrock valleys in weaker bedrock, as many have observed in natural landscapes. Weaker bedrock also results in more channel mobility, which is a fundamental factor for developing and maintaining a bedrock valley that is several times wider than the channel it holds. Increased channel mobility and wider, flat-bottomed valleys in the model under transport-limited conditions suggest that sediment cover on the bed that is present under transport-limited conditions is an effective way to slow vertical incision and amplify the effect of lateral erosion. This theory for
the lateral erosion of bedrock channel walls and the numerical implementation of the theory in a catchment-scale landscape evolution model is a significant first step towards understanding the factors that control the rates and spatial extent of wide bedrock valleys.

5.2 Introduction

Understanding the processes that control the lateral migration of bedrock rivers is fundamental for understanding the genesis of landscapes where valley width is many times the channel width. Where valleys are much wider than the channels they hold, hillslope processes are decoupled from stream networks. Strath terraces are a clear indication of a landscape that has experienced an interval where lateral erosion has outpaced vertical incision [Hancock and Anderson, 2002]. Broad strath terraces that are many times wider than the channels that beveled them are found in mountainous and hilly landscapes throughout the world [e.g. Chadwick et al., 1997; Lavé and Avouac, 2001; Dühnforth et al., 2012] and provide many clues about the nature of their evolution.

Theory for the vertical incision of bedrock channels has advanced considerably since the first physics-based bedrock incision models were presented in the early 1990’s. For example, bedrock incision models now include theories for adjustment of channel width [Wobus et al., 2006; Yanites and Tucker, 2010], the role of sediment size and bed cover [Whipple and Tucker, 2002; Sklar and Dietrich, 2004], and thresholds for incision [Tucker and Bras, 2000; Snyder et al., 2003b]. These theories are assimilated in our current generation of landscape evolution models [Tucker and Hancock, 2010]. However, existing models do not seek to implement lateral erosion of bedrock channel walls and the consequential migration of the channel, in no small part because of the lack of a rigorous understanding of the processes that control lateral erosion of bedrock channel walls. If this theoretical hurdle can be cleared, an algorithm for lateral erosion must be applied within a framework of models that currently only erode and deposit vertically. To our knowledge, this study is the first attempt at incorporating a physics-based algorithm for lateral bedrock erosion and channel migration on a drainage basin scale to a two-dimensional landscape evolution model. Until now, landscape evolution models have lacked a generic mechanism for allowing channels to
migrate laterally and widen bedrock valleys, as well as incise bedrock valleys. While advances in controls on bedrock valley width have been made using meandering models, the representation of a sinuous channel doesn’t describe all rivers, and often such models are constructed on a channel scale rather than on a drainage basin scale. In this study, we develop a theory for the lateral migration of bedrock channel walls and implement this theory in an existing landscape evolution model. We seek to explore the processes that control the morphology of bedrock valleys and the rate of bedrock valley widening using a series of numerical experiments.

5.3 Background

In order for a wide bedrock valley to form, lateral erosion rates must exceed vertical incision rates [Hancock and Anderson, 2002]. Lateral erosion rates depend on the hydraulic characteristics of the eroding channel, specifically shear stress applied to channel walls, and the resistance of the bedrock to erosion. The width a bedrock valley is ultimately able to achieve is related to the lateral erosion rate as well as the duration of the interval of lateral erosion [Suzuki, 1982].

Several studies have found a link between wider bedrock valleys and less resistant bedrock, especially weak sedimentary rock, such as shale and marine sedimentary rock [Montgomery, 2004; Snyder and Kammer, 2008]. The frequency of intense rain is also correlated with higher channel sinuosity and lateral erosion rates [Stark et al., 2010]. The connection between intense rainfall and lateral erosion is supported by one of the few studies that reports bedrock wall retreat rates. Hartshorn et al. [2002] showed that channel widening in the LiWu River in Taiwan occurred during rare extreme events, while moderate, yearly floods were primarily responsible for vertical bed incision. Turowski et al. [2008] determined that the channel wall erosion first reported by Hartshorn et al. [2002], resulted from armoring of the bed during the extreme flood event, which suppressed vertical bed erosion and enhanced lateral erosion. Sediment cover on the bed that suppresses vertical incision and allows lateral erosion to continue unimpeded is a critical element for the development of wide bedrock valleys [Hancock and Anderson, 2002; Brocard and Van der Beek, 2006].

Experimental work has given some insight into how bedrock channel width responds to
changes in sediment and water flux. In flume experiments, increasing sediment supply resulted in a wider area of bed load transport and erosion on the margins of the channels [Finnegan et al., 2007]. Another flume experiment showed that with increasing sediment supply, the fraction of the bed covered in sediment grew, causing incision in a narrow inner channel to slow, while total flume-averaged erosion continued to increase [Johnson and Whipple, 2010]. This suggests that as streams reach carrying capacity, and the inner channel becomes covered in sediment, incision on the bed slows or halts and erosion of the channel walls begins.

Rivers respond to changing boundary conditions by adjusting both slope and channel width [Lavé and Avouac, 2001; Duvall et al., 2004; Snyder and Kammer, 2008] and landscape evolution models must be able capture both of these responses if we are to fully describe the behavior and function of landscapes. Research on bedrock channel width gives important insights into the larger scale problem of bedrock valley widening.

Studies suggest that lateral erosion of channel walls occurs either when erosive shear stress is partitioned from the channel bed to the channel walls, when the channel bed is protected from vertical incision by sediment cover, or through a combination of both mechanisms [Stark, 2006; Wobus et al., 2006; Turowski et al., 2008, 2009; Yanites and Tucker, 2010]. In particular, the effects of sediment cover on the bed play an important role in the evolution of channel cross-sectional shape because sediment cover on the bed can slow or halt vertical incision [Sklar and Dietrich, 2004; Turowski et al., 2007], while allowing lateral erosion to continue. Models of channel cross-sectional evolution predict that increasing sediment supply to a steady-state stream results in a wider, steeper channel for a given rate of base level fall [Yanites and Tucker, 2010].

Lateral migration of bedrock channel walls has only been implemented into landscape evolution models in a few specialized studies [Lancaster, 1998; Hancock and Anderson, 2002; Clevis et al., 2006a; Finnegan and Dietrich, 2011; Limaye and Lamb, 2013]. Hancock and Anderson [2002] reproduce valley widening using a 1-D stream power model for vertical incision and assume that valley widening rates depend on stream power. They note that the width of the valley floor is related to the duration of the graded period of the river, as theorized by Suzuki [1982]. This model
is based on the key observation that lateral erosion exceeds vertical incision when the channel is carrying the maximum sediment load dictated by the transport capacity. By varying sediment supply to the channel, their model predicts the development of a series of strath terraces. Strath terrace sequences have also been produced by coupling a meandering model with a river incision model [Finnegan and Dietrich, 2011]. Lateral migration of a meandering channel has been implemented in several landscape evolution models. Clevis et al. [2006a] modeled meandering channels in a valley section using a 2-D landscape evolution model and an adaptive grid approach. A vector-based approach to modeling lateral migration of meandering streams in heterogeneous bed material has been used to reproduce a range of bedrock valley forms [Limaye and Lamb, 2014], but this model is primarily a channel-scale model. While each of these studies model lateral migration of bedrock channel banks, they all operate with a meandering model that is not applicable to lateral migration in non-sinuous channels.

In this study, we apply a simple process-based theory for the lateral migration of channels to a 2-D landscape evolution model for the first time. We aim to create a first-order model that can reproduce the qualitative aspects of lateral erosion and bedrock valley widening in a broadly applicable landscape evolution model. Through a set of numerical experiments, we seek to answer the following set of questions:

- Does a set of simple rules based on shear stress applied to banks qualitatively match bedrock valley widths observed in natural systems?
- How do bedrock strength and runoff influence valley width and channel migration? Does weaker bedrock result in more lateral migration?
- Does a more transport-limited system result in more lateral migration?
- How do changes in sediment flux and water discharge affect cross sectional shape, valley width, and channel migration?
- Which of these factors is most important in determining rate and magnitude of channel
migration?

In the following sections we outline our theory for lateral channel wall migration and explain the two algorithms we have developed to apply this theory to an existing model. We then present the results from our set of numerical experiments and discuss how well the model describes the formation of wide bedrock valleys.

5.4 Theory

We have deliberately chosen the most simple formulation possible for deposition and erosion, while still capturing the role of sediment. We do this in order to focus on developing the lateral erosion component of our model. Evolution of the height of the landscape, $\eta$, through time is described by deposition rate minus erosion rate, plus a constant rate of uplift.

$$\frac{\partial \eta}{\partial t} = -e + d + U$$ (5.1)

Deposition rate in the landscape is calculated by

$$d = \frac{\nu_s d_s Q_s}{RA}$$ (5.2)

where $\nu_s$ is the average settling velocity of grains in the water column, $Q_s$ is the volumetric sediment concentration in the water, $d_s$ is a dimensionless number describing the vertical distribution of sediment in the water column, which is equal to 1 if sediment is equally distributed through the flow, $R$ is the runoff rate, and $A$ is drainage area [Davy and Lague, 2009]. $\nu_s$, $d_s$, and $R$ are lumped into a single dimensionless parameter, $\alpha$, that represents the potential for deposition.

$$\alpha = \frac{\nu_s d_s}{R}$$ (5.3)

A larger $\alpha$ means greater deposition, either because settling velocity, $\nu_s$, is high and sediment is quickly lost from the flow, or because runoff rate, $R$ is low and there is little water in the channels to move the sediment. A smaller $\alpha$ represents slower settling velocity, or more intuitively, greater runoff. $\alpha$ can be thought of as a detachment/transport number: when $\alpha << 1$, the model tends
towards detachment-limited behavior and when $\alpha >> 1$, the model tends towards transport-limited behavior.

5.4.1 Vertical Erosion Theory

Vertical erosion rate is derived from the rate of energy dissipation on the channel bed, which is given by

$$\omega_v = \rho g \frac{Q}{W} S$$  \hspace{1cm} (5.4)

where $\rho$ is the density of water, $g$ is gravitational acceleration, $Q$ is water discharge, $W$ is channel width, and $S$ is channel slope. The rate of vertical erosion scales as

$$E_v = K'_v \frac{\omega_v}{C_e}$$  \hspace{1cm} (5.5)

where $K'_v$ is a dimensionless vertical erosion coefficient and $C_e$ is bank cohesion. Substituting $RA$ for $Q$ and $k_w Q^{1/2}$ for $W$ in equation 5.4, and combining equations 5.4 and 5.5 gives:

$$E_v = \frac{K'_v \rho g R^{1/2}}{k_w C_e} A^{1/2} S$$  \hspace{1cm} (5.6a)

$$E_v = K_v A^{1/2} S$$  \hspace{1cm} (5.6b)

where $K'_v$ is a dimensionless bank erosion coefficient and $k_w$ is a bank width coefficient. Lumping several parameters gives $K_v$, a dimensional vertical erosion coefficient, which consists of known or measurable quantities, and one unknown dimensionless parameter, $K'_v$.

5.4.2 Lateral Erosion Theory

Lateral erosion rates depend on energy expenditure on channel walls [Suzuki, 1982; Lancaster, 1998; Hancock and Anderson, 2002]. We hypothesize that the lateral erosion rate is proportional to the rate of energy dissipation per unit area of the channel wall created by centripetal acceleration around a bend. Centripetal force is $F_c = m \frac{v^2}{r_c}$, where $m$ is mass, $v$ is velocity, and $r_c$ is radius...
of curvature. The centripetal force of a unit of water can be found by replacing \( m \) with \( \rho LHW \), where \( \rho \) is the density of water, and \( L, H, \) and \( W \) are unit length, water depth, and channel width, respectively. Centripetal force of water around a bend can be expressed in terms of centripetal shear stress, which is analogous to bed shear stress, by dividing both sides by \( HL \) giving:

\[
\sigma_c = \frac{\rho W v^2}{r_c}
\]

Centripetal shear stress can be turned into a rate of energy expenditure by multiplying by velocity, giving:

\[
\omega_c = \frac{\rho W v^3}{r_c}
\]

To express this in terms of discharge, \( Q \), instead of velocity, we follow the Darcy-Weisbach equation, giving \( v^3 = gqS/F \), where \( q \) is discharge per unit width and \( F \) is a friction factor, which yields

\[
\omega_c = \frac{gQ S}{r_c F}
\]

We hypothesize that lateral erosion rate scales with energy dissipation rate around a bend according to

\[
E_l = K'_l \frac{\omega_c}{C_e}
\]

where \( K'_l \) is a dimensionless lateral erosion coefficient. Combining equations 5.9 and 5.10 gives

\[
E_l = K'_l \frac{\rho g R AS}{C_e F r_c}
\]

\[
E_l = K_l \frac{AS}{r_c}
\]

where \( K_l \) is a dimensional erosion coefficient for lateral erosion, which is composed of known or measurable quantities, and one unknown dimensionless parameter, \( K'_l \).
If $K_l'$ is equal to $K_v'$, we find a ratio between $K_l$ and $K_v$, given by

$$\frac{K_l}{K_v} = \frac{R^{1/2}k_w}{F}$$

(5.12)

which consists of runoff rate, $R$, bank width coefficient, $k_w$, and friction factor, $F$. We can measure or make reasonable estimates of each of these parameters in order to determine what the ratio of lateral to vertical erodibility should be. Runoff rate can vary widely, but a higher runoff intensity will lead to a higher $K_l/K_v$ ratio and more lateral erosion, as suggested by field observations of lateral erosion in bedrock channels [Hartshorn et al., 2002] and correlation of increased sinuosity and storminess of climate [Stark et al., 2010]. A bank width coefficient of $10 \text{ m}/(\text{m}^3/\text{s})^{1/2}$ is reasonable for a range of natural rivers [Leopold and Maddock, 1953]. If $k_w$ is lower, then the channel is more narrow and water is deeper, and more vertical incision should occur. The friction factor, $F$, is the Darcy-Weisbach friction factor, which can range from 0.01-1.0 for natural rivers [Gilley et al., 1992; Hin et al., 2008]. With a lower friction factor, representing smooth walls, the lateral erosion ratio would be higher due to less energy being dissipated on the walls, leaving more energy available for eroding the banks.

5.5 Numerical Implementation

One challenge in representing vertical and lateral erosion lies in the representation of topography. Normally, landscape evolution models use a numerical scheme in which the terrain is represented by a grid of points whose horizontal positions are fixed and whose elevation represents the primary state variable in the model. Such a framework does not lend itself to the motion of near-vertical to vertical interfaces (such as stream banks and cliffs), and for this reason, incorporating lateral stream erosion in a conventional landscape evolution model requires a modification to the basic numerical framework. A vertical rather than horizontal grid [Kirkby, 1999] can be used for near-vertical landforms in isolation, but is inappropriate when one wishes to represent vertical interfaces that are inset within a larger landscape. Grid-node movement combined with adaptive re-gridding [Clevis et al., 2006b,a] provides a possible solution, but is computationally expensive,
and particularly difficult to implement when multiple branches of a drainage network may undergo lateral motion. Here, we adopt a simpler approach in which valley walls are viewed as sub-grid-scale features that migrate through the fixed grid. Rather than tracking the position of these vertical interfaces, we instead track the cumulative sediment volume that has been removed from a given grid node as a result of lateral erosion. When that cumulative loss exceeds a threshold volume, the elevation of the grid node is lowered.

More specifically, at each node in the model, we calculate a vertical incision rate at the primary node and a lateral erosion rate at a neighbor node. The lateral neighbor node for the primary node is chosen on the outside bank of two stream segments, the largest upstream flow direction to the primary node and the downstream flow direction (Figure 5.1). If the two segments are straight, then a neighbor node of the primary node is chosen at random. The radius of curvature for the stream segments is calculated based on the angle of the intersecting stream segments, which can be equal to 0°, 45°, or 90° on a square grid. We assume that for two straight stream segments, the angle difference between the segments can range from +22.5 to -22.5, giving $1/r_c$ of $0.23/dx$, where $dx$ is grid spacing. For a 45° bend, the last stream segment could be positioned anywhere from half way to 0° to half way to 90°, resulting in $1/r_c$ of $0.67/dx$. Following the same principle for a 90° bend, $1/r_c$ is equal to $1.37/dx$.

The volumetric rate of material eroded laterally for each lateral node is calculated by $E_l \times dx \times H$, where $dx$ is the cell size in the flow direction and $H$ is the water depth. Water depth at each node is calculated by $H = 0.4Q^{0.35}$ [Andrews, 1984]. The volume of sediment eroded laterally per time step is sent downstream along with any material eroded from the primary cell. Volumetric erosion rate is multiplied by the time step to get the volume eroded at the lateral nodes, and the cumulative volume eroded from each lateral node is tracked throughout the entire model run.

We have implemented two ways of determining whether enough lateral erosion has occurred to erode the lateral node. The first method dictates that the entire volume of the lateral node above the elevation of the downstream node must be eroded before its elevation is changed (Figure 5.1a,b). This formulation assumes that the bank material being eroded is very resistant and blocky.
The second method dictates that only the volume of the water height on the bank times the cell area must be eroded for the elevation to change (Figure 5.1c,d). This model represents lateral erosion on a bank that has been laterally undercut and the remaining material slumps into the channel and is transported away as wash load, and assumes that the bank material slumps easily and rapidly breaks down into small grains that are easily transported. With these two end member models, we address whether lateral erosion rate should scale with valley wall height. In the first method, the total block erosion model, lateral migration is dependent on bank height so that taller banks experience less lateral migration, as all of the volume of the lateral node must be eroded for the valley to widen [Lancaster, 1998]. On the other hand, if all of the material that has been undercut by the channel is also swept away by the channel, then lateral erosion rate is independent of bank height. However, this undercutting-slump model is not appropriate for landscapes with very hard bedrock (low erodibility), as evidenced by overhanging cliffs along many rivers and persistent blocks of collapsed material following slumping.
Figure 5.1: Conceptual figure of model nodes showing the stream segments (in light blue) from the upstream node to the primary node (in green), to the downstream node. Vertical erosion ($E_v$) occurs at the primary node. The neighbor node (in pink) where lateral erosion ($E_l$) occurs is located on the outside bend of the stream segments. The height over which lateral erosion occurs, $H$, is shown in the dashed blue line. a) For the total block erosion model, the volume that must be laterally eroded before elevation is changed is $(Z_n - Z_d)dx^2$, the difference in elevation between the neighbor node and the downstream node (indicated with black arrow) times the surface area of the neighbor node. b) Elevation of the lateral node is changed after after the entire block is eroded and flow can be rerouted. c) In the undercutting-slump model, the volume that must be laterally eroded (representing bank undercutting) before elevation is changed is $(H - Z_d)dx^2$. $H - Z_d$ is the difference in elevation between the water height and the elevation of the downstream node, indicated with black arrow. d) When the neighbor node has been undercut, elevation is changed, allowing water to be re-routed, while the slumped material is transported downstream as washload.
5.6 Model Description

In order to explore the effect of bedrock erodibility and transport-limited and detachment-limited conditions on lateral channel migration, we ran several sets of models with a range of bedrock erodibility values and $\alpha$ values using both the total block erosion model and the undercutting-slump model. All models were spun up to a condition of approximately uniform erosion rate with no lateral erosion component, then run for 200 ky with one of the lateral erosion models. The erodibility in the models ranges from $5 \times 10^{-5}$ to $5 \times 10^{-4}$. The $\alpha$ values range from 0.1 to 2, which represent a detachment-limited system when $\alpha < 1$ and a transport-limited system when $\alpha > 1$. Runoff rate for the different sets of model runs is set to 36 mm/hr, 14 mm/hr, and 9 mm/hr, making $K_t/K_v$ ratio for these runoff rates equal to 1.5, 1, and 0.8.

In order to develop insight into model sensitivity to changes in runoff without changes in the settling velocity component of $\alpha$, we also ran a set of model runs where $v_s$ is held constant. Runoff rate ranges from 2 mm/hr to 36 mm/hr, and $K_t/K_v$ ratio for these runoff rates ranges from 0.35 to 1.5. This set of model runs shows how the model responds to conditions that are transport-limited due to low runoff rate rather than large grain size, i.e., high $\alpha$ values with low $K_t/K_v$ values.

5.6.1 Effects of Lateral Erosion on Model Landscapes

5.6.1.1 Channel Mobility

Channel mobility distinguishes models with lateral erosion from models with only vertical incision. At steady state, channels do not migrate across the model domain in models of bedrock incision without lateral erosion. Additionally channel mobility is an important and necessary factor in creating broad strath terraces that are many times wider than a single channel width. In order to describe channel mobility in a single term, we calculate the cumulative channel-averaged migration distance over a 100 ky time interval in the model runs. The resulting number, called $\lambda$, indicates how far the channel has moved on average from its starting position. It is also an indicator for the maximum lateral extent occupied by the channel during the model run. That is, the extent
of x positions occupied by the channel is \( \lambda \) at a maximum, but could be lower as the channel migrates over the same area repeatedly. Bedrock erodibility and \( K_l/K_v \) ratio have the strongest control on channel migration distance (Figure 5.2). Increasing \( K \) results in a monotonic increase of channel mobility. Increasing the \( K_l/K_v \) ratio from 0.8 to 1.5, results in 5.5–2.5 times more channel mobility, with the largest increases for the lowest \( K \) values (Figure 5.2a). For model runs with the same bedrock erodibility, but different \( \alpha \) values, the \( K_l/K_v \) ratio is the most important factor for channel mobility (Figure 5.2b). When \( K_l/K_v \) ratio is constant at 1.5, \( \lambda \) is between 40–50 m for detachment-limited (<1) \( \alpha \) values. As \( \alpha \) increases towards transport-limited (>1) values, \( \lambda \) increases to over 100 m.

Channel-averaged migration distance \( \lambda \) for the undercutting-slump model runs (Figure 5.2c,d) is several times higher than \( \lambda \) for the total block erosion model, but follows the same general pattern. Channel migration is highest in models with high bedrock erodibility and in models with high \( K_l/K_v \) ratio. Models runs with \( K_l/K_v \) ratio equal to 1.5 show increasing channel mobility with increasingly transport-limited model runs, as with the total block erosion models, and models with \( K_l/K_v \) ratio equal to 1 have less channel mobility that peaks when \( \alpha = 1.5 \), but less channel mobility with further increases in \( \alpha \). The models that have the largest increase in \( \lambda \) using the undercutting-slump model, are the runs that had the least channel migration in the total block erosion models, especially the models with low \( K \) values (Figure 5.2a,c). However, it is these models that have resistant bedrock (low \( K \)) that are least suitable for the undercutting-slump model. This model assumes that after the lateral bank has been sufficiently undercut, then the rest of the material will slump into the stream and be carried away. In order for this to be a good assumption about how nature works, the bed material must be able to break up into small pieces that are easily transported away.

Figure 5.3a shows \( \lambda \) plotted against runoff rate for the model runs where \( v_s \) is held constant. For the total block erosion models, as runoff increases from 1.8 mm/hr to 36 mm/hr (\( K_l/K_v = 0.3–1.5 \) and \( \alpha = 2.0–0.1 \)), \( \lambda \) increases from 3 m to 40 m. For the undercutting-slump model, \( \lambda \) increases from 47 m to 205 m over the range of runoff rates. Comparing transport-limited, low runoff runs from Figure 5.3a with transport-limited, high runoff runs from Figure 5.2d demonstrates the effect
Figure 5.2: Cumulative channel-averaged migration, $\lambda$, over 100 ky of the model runs. a,c) show $\lambda$ plotted against bedrock erodibility, $K$ for the total block erosion models (a) and the undercutting-slump models (c). b,d) show $\lambda$ plotted against $\alpha$ for the total block erosion models (b) and the undercutting-slump models (d).

of runoff rate and $K_l/K_v$ ratio on model runs with the same $\alpha$ value. There is less channel mobility in a stream that is transport-limited by low runoff, giving low $K_l/K_v$ ratio (Figure 5.3a), and more channel mobility in a channel that is transport-limited by grain size with a high runoff rate and high $K_l/K_v$ value (Figure 5.2d).

The effect of bedrock erodibility and runoff on channel migration through time for both model
versions is shown in Figure 5.4. Channel migration over 200 ky is shown for six selected runs that span the range of bedrock erodibility and $\alpha$ values for the two different model formulations: the undercutting-slump model where $K_l/K_v=1.5$ and the total block erosion model where $K_l/K_v=1.5$. In all model runs, the total block erosion model produced more confined channels compared to the undercutting-slump model. The undercutting-slump model produces more dynamic channel migration over the model domain, especially in the high $K$ model. In both models formulations, the high $K$ and high $\alpha$ runs have widest extent of channel migration and the low $K$ and low $\alpha$ runs have the most restricted channel migration.
Figure 5.3: Channel mobility, $\lambda$ (a), valley width based on height (b), and flat-bottomed valley width (c) plotted against runoff rate for the total block erosion and undercutting-slump model runs. Bedrock erodibility, $K$, is the same in all of the model runs.
Figure 5.4: Channel position over 200 ky for the undercutting-slump model (blue lines) and total block erosion model (red lines). a-c) show channel positions through time for models with decreasing bedrock erodibility and medium $\alpha$, for both the total block erosion runs and the undercutting slump models. d-f) show channel positions through time for model runs with medium bedrock erodibility and decreasing $\alpha$. 
5.6.1.2 Valley Width

Valley width is an important indicator of lateral erosion; a wide bedrock valley tells us that significant lateral erosion has occurred. There are several ways a valley can be defined. Many studies use low slope areas of a DEM to determine valley width [e.g., Brocard and Van der Beek, 2006; May et al., 2013]. This gives the width for the flat valley that has been shaped by channel processes, but excludes areas in rapidly responding landscapes that have been recently shaped by channel processes and then reworked by hillslope processes. Another way to measure valley width is by determining the width of the valley at a certain height above the channel. This simple metric is often used for finding valley width in the field, for example using eye height above the channel [e.g., Snyder et al., 2003a; Whittaker et al., 2007]. Valley width in our model runs is measured using both metrics. Flat valley width is the width of the area next to the main channel, where slope is characteristic of the channel rather than hillslopes for a given bedrock erodibility and $\alpha$ value. Valley width based on height is found by taking the nodes adjacent to the channel that are lower in elevation than 2.5 times the maximum water depth, $d$, for the models ($d=0.63$ m). Figure 5.5 shows mean valley width averaged over the entire model run and averaged over the 20 furthest downstream nodes in the models. To ensure that using elevation lower than 2.5$d$ as the criterion for a valley in all model runs gives valley width resulting from lateral erosion, and not valley width inherent in the model, we first use this criterion to measure valley width for the spin up models that include no lateral erosion component. Valley width for the spin up models is consistently 10 m, the width of one model cell. Where valley width for the spin up model deviates significantly from 10 m using this elevation criterion, for example in the model runs with the highest erodibility, valley width is not evaluated.

Valley width based on height in the models increases with increasing bedrock erodibility and $K_l/K_v$ ratio. For the total block erosion model, valley width increases only marginally for the low bedrock erodibility models with a $K_l/K_v$ ratio of 0.8 (Figure 5.5a). In the total block erosion model, valley width exceeds 20 m in the high erodibility run with $K_l/K_v$ of 0.8. When $K_l/K_v$ ratio
is increased to 1.5, valley width across the range of bedrock erodibilities increases by 30–45%.

When $\alpha$ is varied and $K$ held constant (Figure 5.5b), valley width based on height depends on the $K_t/K_v$ ratio. Models with $K_t/K_v=1.5$ have a constant valley width of about 16 m for this particular bedrock erodibility value. The constant valley width based on height across a range of $\alpha$ values when $K_t/K_v$ is high is unexpected, given the many observations suggesting that valley widths should be wider in transport-limited streams [e.g. Brocard and Van der Beek, 2006], so we look to the flat bottom valley width. Figure 5.6b shows the flat valley width for the total block erosion models. With a high $K_t/K_v$ ratio, as the tendency towards transport-limited conditions increases ($\alpha>1$), the flat bottom valley width increases. An increase in flat valley width means that not only has the valley widened, but the valley has also changed in morphology, from a V-shaped valley to a valley with a flat bottom and steep sides. Flat valley width does not increase for any of the model runs in which $K$ is varied (Figure 5.6a). This indicates that the valley has remained V-shaped, although the hillslopes surrounding the channel have lowered and valley width based on height has increased.

Mean valley widths based on height above the channel for the undercutting-slump model are significantly greater than in the total block erosion model (Figure 5.5a,c), particularly for the model runs with low bedrock erodibility. Valley width increases with increasing bedrock erodibility, ranging from 18 m in the low $K$ model to 34 m in the high $K$ model with $K_t/K_v$ ratio equal to 0.8. When $K_t/K_v$ ratio is increased to 1.5, valley width increases by 20–60%, with the largest increases in the lower $K$ cases (Figure 5.5c). When $\alpha$ is varied and $K$ held constant for the undercutting-slump model, the $K_t/K_v$ ratio controls valley width (Figure 5.5d). With $K_t/K_v$ ratio held constant at 1.5, valley width is approximately constant at 32 m for this $K$ value. When the $K_t/K_v$ ratio is held constant at 1, valley width based on height gradually decreases with increasing $\alpha$. Constant valley width across a range of $\alpha$ values in the undercutting-slump models suggests another look at flat valley width. Figure 5.6d shows flat valley width for the undercutting-slump models with a constant $K$ value. When $K_t/K_v \geq 1$, increasingly transport-limited streams have larger flat valley width, indicating a transition from a V-shaped valley to a valley with a flat bottom.
Flat valley width in the undercutting-slump model for changing bedrock erodibility shows a somewhat counter-intuitive signal (Figure 5.6c). When using the undercutting-slump model, flat valley width increases with harder bedrock. However, this trend reflects use of the undercutting-slump model, which is inappropriate for hard bedrock wall erosion in natural systems. With the undercutting-slump model, only a very small volume threshold must be overcome for lateral erosion to occur, and the rest of the node material is transported downstream as wash load. The total block erosion model is more appropriate for the erosion of hard bedrock channels. Flat valley width is an indicator of valley shape and in order to maintain a flat valley over a period of time, the model must be transport-limited or have hard bedrock. Figure 5.7 demonstrates how model runs that have low $\alpha$ values tend to have V-shaped valleys (Figure 5.7a) and model runs with high $\alpha$ values tend to have flat-bottomed valleys (Figure 5.7b).

Figure 5.3b shows mean valley width measured using the elevation criterion plotted against runoff rate for the model runs in which $v_s$ was held constant. For the total block erosion models, as runoff increases from 1.8 mm/hr to 36 mm/hr ($K_l/K_v = 0.3–1.5$ and $\alpha= 2.0–0.1$), valley width increases from 10 m to 17 m. For the undercutting-slump model when runoff rate is lower than 9 mm/hr ($K_l/K_v$ ratio less than 1), valley width is only marginally wider than the spin up model runs with no lateral erosion. With a high runoff rate (high $K_l/K_v$ ratio and low $\alpha$), valley width measured using the elevation criterion increases to over 30 m. Increases in flat bottom valley width for these detachment-limited runs using both model formulations is more subdued (Figure 5.3c).
Figure 5.5: Mean valley width calculated from height above channel averaged over 100 ky of the model runs. a,c) mean valley width vs. bedrock erodibility, $K$ for the total block erosion models (a) and the undercutting-slump models (c). b,d) mean valley width vs. $\alpha$ for the total block erosion models (b) and the undercutting-slump models (d).
Figure 5.6: Mean flat-bottomed valley width averaged over 100 ky of the model runs. Flat-bottomed valley width is the width of the valley that has a slope that is more channel-like than hillslope-like. a,c) show mean valley width plotted against bedrock erodibility, $K$ for the total block erosion models (a) and the undercutting-slump models (c). b,d) show mean valley width plotted against $\alpha$ for the total block erosion models (b) and the undercutting-slump models (d).
Figure 5.7: Surface topography and cross section at y=500 for model runs with different $\alpha$ values using the total undercutting-slump model, showing valleys with the same width using height above channel (Figure 5.5d), but different flat valley widths (Figure 5.6d). a) low $\alpha$, tendency towards V-shaped channel. b) high $\alpha$, tendency towards flat-bottomed valley.
5.6.2 Adding Complexity: water flux, sediment flux, terrace formation

5.6.2.1 Effects of increased discharge on lateral channel migration

In the total block erosion and undercutting-slump models in which \( v_s \) is constant, we introduce increased discharge at a point in the upstream end of the model. Using drainage area as a proxy for discharge, increasing water flux in the model represents how a larger stream on the same landscape will influence valley width. Increasing drainage area also allows us to observe the extent of landscape change and how rapidly the different model runs respond to an event such as stream capture. The drainage area at this input point is increased from 20,000 m\(^2\) to 160,000 m\(^2\) and sediment load is set to the carrying capacity of the new drainage area. For a typical model run, the additional drainage area approximately doubles the drainage area at the outlet of the main channel in the model domain.

Lateral erosion scales with drainage area (Equation 5.11), while vertical incision scales with the square root of drainage area (Equation 5.6), and therefore we expect that increasing drainage area will increase lateral erosion and valley width in every case for the undercutting-slumping model, where the threshold for lateral erosion is much smaller than in the total block erosion model. In the total block erosion model, lateral erosion will temporarily stall because of the volume threshold that must be exceeded before lateral erosion occurs. There is no threshold for vertical incision, which will speed up when additional water flux is added to the model.

In the total block erosion models, increased water flux results in more channel mobility in all of the model runs, with only marginal increases in channel mobility in the low erodibility runs and the low \( K_l/K_v \) ratio runs. Since mean valley width is averaged over the channel in the downstream half of the model domain, widening in the upper section of the channel is not captured, where most valley widening occurs for the increased water flux models. In the total block erosion model runs, valley width increases by \( \sim 25\% \) for most models, except those with low \( K_l/K_v \) ratio.

For models with low \( K_l/K_v \) ratio, increasing water flux results in increased channel mobility, but not in increased valley width in the downstream channel segments. For these models, the
increased potential lateral erosion from the increased drainage area is not enough to force valley widening with a low $K_l/K_v$ ratio and steep hillslopes with a large volume threshold to overcome before they are eroded. In these models, the vertical incision response is stronger than the lateral erosion response. For the models with a high $K_l/K_v$ ratio, introducing more discharge results in more lateral erosion because the increased drainage area and $K_l/K_v$ ratio greater than 1 both work to produce higher lateral erosion rates.

In the undercutting-slump models, all of the runs show a significant increase in channel mobility with additional water flux, except the runs with low $K_l/K_v$ ratio. The largest increases in valley width occur in the runs with low bedrock erodibility. In these model runs, valley width increases by ~40–50% while valley width in most of the other runs increases by ~10%.

Figure 5.8 shows cross sections through time for the low and medium erodibility runs using the total block erosion model. It demonstrates the effect of valley deepening, then widening in response to increased water flux. After water flux increases at 100 ky, both the low and medium $K$ runs incise, deepening and narrowing the valley. After 20 ky of increased water flux and increased vertical incision, incision in the channel and uplift are equal and channel elevation is stationary. Only after this period of incision is over does lateral erosion begin to widen the valleys. The higher erodibility runs begin to erode laterally more quickly than in the low erodibility runs and valley width is greater in the higher erodibility model.

One of the few studies that has been able to report bedrock valley widening through time is from a unique case in Death Valley [Snyder and Kammer, 2008]. Stream capture increased the drainage area of a small basin by 75 fold in the 1940’s and channel response over the following 60 years was mapped by aerial photos. Snyder and Kammer [2008] found that mean valley width in a channel segment with weak bedrock increased by 9 meters in 60 years. In contrast, in channel segments in hard bedrock, they found vertical channel incision and the development of knickpoints. They attribute the difference in response to lithological differences and suggest that the presence of sediment on the bed in the weak bedrock channel segments protects the bed from incision, allowing the valley walls to migrate laterally. This difference in response is similar to the behavior of the end-
member models presented here: the total block erosion model shows rapid incision and narrowing in response to increased water flux, whereas the undercutting-slump models show incision and valley widening.

Figure 5.9 shows surface topography and cross section across the model domain for two time steps in the low erodibility model run using the total block erosion model. Before water flux is increased, the channel is narrow and has steep valley walls (Figure 5.9a). After 40 ky of increased water flux, the entire channel has incised, especially in the upper valley. At y=420, the position of the cross section, the channel has been incised by 3 m, but the valley walls are still steep and no lateral erosion has occurred at this position in the model (Figure 5.9b). This is to be expected when the threshold for lateral erosion is high because of the hard bedrock and steep slopes. Figure 5.10 shows topography and cross sections for two time steps in the low erodibility model run using the undercutting-slump model. Before water flux is increased, the channel is significantly wider than in the total block erosion model. The cross section shows a wider valley spanning two model cells, and lower slopes on the neighboring interfluves, indicating that these areas were shaped by the lateral erosion from the channel as well. After 40 ky of increased water flux, the valley is much wider across the entire model domain, especially at the upstream segments of the channel. At y=420, the position of the cross section, the valley is broad and gently U-shaped.
Figure 5.8: Cross sections across model domain at \( y=120 \) during period of increased water flux for the total block erosion model. Cross sections over 36 ky show vertical incision and increasing relief between the channel and hillslopes initially. After equilibrium is reached, lateral erosion can begin at an increased rate compared to before the additional water flux. a) Low erodibility model \( (K=5 \times 10^{-5}) \). b) Medium erodibility model \( (K=10^{-4}) \).
Figure 5.9: Surface topography and cross section at y=420 during period of increased water flux for the total block erosion model, low $K$ run ($K=5\times10^{-5}$). a) 100 ky, before the increase in water flux. Note that this model looks similar to the spin up model runs with no lateral erosion. b) 140 ky, after 40 ky of increased water flux. Cross section shows incision in the channel and increased relief between the channel and the hillslopes. No obvious lateral erosion has occurred at this position.
Figure 5.10: Surface topography and cross section at $y=420$ during period of increased water flux for the undercutting-slump model, low $K$ run ($K=5 \times 10^{-5}$). a) 100 ky, before the increase in water flux. b) 120 ky, after 20 ky of increased water flux, the channel is slightly lower elevation than before the addition of water flux and the valley is much wider. Valley widening occurred along the length of the main stream.
5.6.2.2 Effects of increased sediment flux on lateral erosion

In order to explore how the addition of sediment to a stream affects lateral erosion and valley widening, we added sediment to the influx point at the top of the model. Before additional sediment flux was added, the sediment flux at the input point equaled to the carrying capacity of the stream, which is equal to $UA$. During periods of increased sediment flux, five times more sediment flux was added, forcing all of the streams to aggrade initially. Adding $Q_s$ increases the deposition term, which decreases the vertical erosion term if the model is in steady state, that is $e - d = U$. While vertical erosion minus deposition is less than uplift rate, the channel will aggrade. When the channel slopes become steep enough to increase the vertical erosion term so that $e - d = U$ again, then channel is again in steady state. Decreasing the vertical erosion term does not change the lateral erosion term, but gives lateral erosion a chance to outpace vertical incision. The addition of sediment flux is expected to have the greatest effect on valley widening in the total block erosion models and in the undercutting-slump models with low $K_l/K_v$ ratio. In these cases, lateral erosion can only outpace vertical incision in order to make a wide valley if vertical incision is suppressed. These models are limited by two different mechanisms. The total block erosion models are limited by a high volume threshold that must be exceeded in order to erode laterally. The low $K_l/K_v$ models have a lower lateral erosion rate compared to the vertical incision rate, and so require a longer period of suppressed vertical incision to widen the valleys.

In the undercutting-slump model, channel mobility increases slightly with increased sediment flux in several model runs, but the largest increases in channel mobility occurred when $K_l/K_v=1.5$. Channel width increases slightly after adding sediment flux for the undercutting-slump models. Figure 5.11 shows the high $\alpha$, high $K_l/K_v$ run with added sediment flux. This run typifies many of the model runs; the upper valleys tend to fill with sediment, but not erode the valley walls. At 50 ky in the model run, the channel is about 30 m wide with a low slope hillslope on one side and a steep valley wall on the other side (Figure 5.11a). After 15 ky of sediment flux, the channel has aggraded by 10 m at the location of the cross section, and the channel has aggraded by 20 m at the
sediment influx point. The aggradation in the valley has widened the valley to the extent that the valley is flatter than prior to the addition of sediment. But the valley walls were not eroded during this period. During the period of aggradation in this model run, the channel tends to straighten, which decreases the lateral erosion rate due to its dependence on the radius of channel curvature.

Figure 5.12 shows the medium erodibility, high $K_l/K_v$ run with added sediment flux. This run is one of the few runs that shows true increased valley widening with added sediment flux. At 50 ky in the model run before the additional sediment is added, the valley in the upper half of the model domain ($y=240$) is flat and about 30 m wide (Figure 5.12a). Over 50 ky, sediment is added to the model and the channel aggrades for $\sim$25 ky before it comes into steady state, i.e., its slope is steep enough to carry the additional sediment load and aggradation stops. During the 25 ky of aggradation, this model run shows both retreat of the valley walls and channel aggradation. By 72 ky in the model run, the channel has aggraded by 5 meters and the valley is 40 m wide (Figure 5.12b). During this period of 22 ky, the channel has migrated 50 m, eroding the hillslope and forming steep valley walls. This run makes wide valleys with the addition of sediment because it has none of the factors that limited the other runs that failed to produce valley widening. The $K_l/K_v$ ratio is high enough and the bedrock erodes enough to produce widening in the relatively long period of aggradation.
Figure 5.11: Topography and cross section at y=240 during period of increased sediment flux for the undercutting-slump model with $\alpha=2.0$ and $K_l/K_v=1.5$. a) before the additional sediment flux at the input point (indicated with the arrow), relief is high between the main channel and the steep hillslopes. b) After 15 ky of increased sediment flux, the channel has aggraded by 10 m, and relief between the channel and valley walls is reduced, but there is no new erosion of the valley walls.
Figure 5.12: Topography and cross section at y=240 during period of increased sediment flux for the undercutting-slump model with medium erodibility and $K_v/K_e=1.5$. a) before the sediment flux is introduced at input point, indicated with the arrow. b) after 22 ky of increased sediment flux, the channel has aggraded by 5 m and has eroded the valley wall by 50 m.
5.6.3 Cycling $K_l/K_v$ ratio

In an effort to determine whether the theory for lateral erosion captures the processes of valley widening followed by incision and terrace abandonment, we ran models that cycle between high $K_l/K_v$ ratio and low $K_l/K_v$ ratio. Drainage area at a cell at the top of the watershed was set to $1.6 \times 10^5 \text{ m}^2$, representing a large stream coming in to the top of the model domain. These model runs have a high bedrock erodibility ($K = 2.5 \times 10^{-4}$) and $\alpha$ values that tend toward transport-limited behavior ($\alpha=2.0$). The models were run for 200 ky with two cycles of 50 ky of high $K_l/K_v$ ratio (1.5) and 50 ky of low $K_l/K_v$ ratio (0.35). Because $K_l/K_v$ ratio, which is controlled by runoff, was changed in these models and $\alpha$ was held constant, this means that $v_s$, a proxy for grain size that is not explicitly set, must have changed. With a lower runoff rate and same $\alpha$ value, $v_s$ must decrease. There is some evidence to support increased grain size during glacial intervals [Pierce and Scott, 1983], when strath terrace formation is believed to have occurred, but the intent of these model runs was to test lateral erosion response to changing $K_l/K_v$ ratio, not changing grain size and $K_l/K_v$ ratio.

Figure 5.13 shows mean valley width from height above the channel and mean flat valley width through time for the model run. The two measures of valley width through time are quite close during periods of high $K_l/K_v$ when maximum valley width is 40–50 m. Both measures of valley width change rapidly in the transition from high to low $K_l/K_v$ ratio. During low $K_l/K_v$ periods, flat valley width tends to be about 5 m lower than valley width based on height.

Figure 5.14 shows the model during the high $K_l/K_v$ interval. At 45 ky, the flat-bottomed valley is $\sim 70 \text{ m wide}$ and is eroding the left valley wall (Figure 5.14a). At 47 ky, the channel has shifted by 60 meters and is eroding the right valley wall, but it is already clear that the left side of the valley is aggrading and will not remain flat without lateral erosion from the channel (Figure 5.14b). Figure 5.15 shows the same model run just after $K_l/K_v$ ratio has dropped from 1.5 to 0.35. At 51 ky, the channel immediately begins to incise and the sides of the valley are uplifting and beginning to form a V-shaped valley (Figure 5.15a). After 8 ky of low $K_l/K_v$ and channel
Figure 5.13: Mean valley width and mean flat bottom valley width through time for the model run where $K_l/K_v$ ratio is cycled between high and low values. High $K_l/K_v$ ratio is between 0–50 ky and 100–150 ky. Low $K_l/K_v$ ratio is between 50-100 ky and 150–200 ky. Mean valley width and mean flat bottom valley width are similar during periods with high $K_l/K_v$ ratio, but flat valley width is lower and decreases more quickly during periods with low $K_l/K_v$ ratio.

incision, the channel still largely incises vertically, and what was formerly the valley floor is now indistinguishable from the neighboring hillslopes.

The valleys do not retain a flat-bottomed shape when they are not being actively eroded by the channel because the lateral erosion from the channel has over-flattened the valley bottoms with respect to the rest of the landscape. These over-flattened valley bottoms are corrected through both sediment deposition and erosion rates lower than the uplift rate. The steep valley walls deposit sediment at the base of the slopes and relief decreases between the valley walls and valley bottom if the channel is not actively eroding the valley. Valley slopes that are too low to erode material at the same rate as the landscape is uplifting, increase in slope until they are again in steady state with uplift. In nature, valley floors get filled in from steep valley walls to some extent, but the flat valley bottom can still be uplifted and preserved for a period of time that is inversely proportional to uplift rate [Anderson et al., 1999].
Figure 5.14: Topography and cross section at y=420 for model run cycling between high and low $K_l/K_v$ ratios with high $K_l$ and $\alpha=20$.

a) During period of high $K_l/K_v$ ratio, the valley is 70 m wide and the channel is eroding at the left wall. b) 2 ky later, still during the period of high $K_l/K_v$ ratio, the channel shifted by 60 m and is eroding the right wall.
Figure 5.15: Topography and cross section at y=420 for model run cycling between high and low $K_l/K_v$ ratios with high $K$ and $\alpha=2.0$.  

a) 1 ky after the beginning of the low $K_l/K_v$ ratio period, the channel begins to incise in the wide valley.  
b) 7 ky later, during the period of low $K_l/K_v$ ratio, the channel has incised even further and the wide valley is nearly indistinguishable from the hillslopes.
5.7 Discussion

This simple theory for lateral bedrock channel erosion combined with a landscape evolution model produces valleys that are several times wider than the channels they hold. The development of wide valleys is strongly dependent on both bedrock erodibility and runoff rate. The model predicts that landscapes with highly erodible bedrock and high runoff rate with transport-limited conditions will produce the most channel mobility and the widest flat-bottomed valleys. The model also captures transient responses in valley width to changing sediment and water flux. The two model formulations presented here offer an explanation for different widening responses in hard and soft bedrock.

In both model formulations presented, easily erodible bedrock allows the development of wider bedrock valleys, as observed in many natural systems [e.g., Montgomery, 2004; Brocard and Van der Beek, 2006]. This occurs in the models because the threshold for lateral erosion is lower in low relief landscapes with easily eroded bedrock. The model predicts significantly more channel mobility in models with weaker bedrock. Channel mobility is a critical factor in the development of wide bedrock valleys, because all of the erosion of the valley must be accomplished through erosion by the channel [e.g., Tomkin et al., 2003]. The model also predicts more channel mobility and wider flat valleys in models with transport-limited behavior. In natural systems, wide bedrock valleys are considered a diagnostic feature of transport-limited streams [Brocard and Van der Beek, 2006].

The total block erosion model shows how landscapes with hard bedrock and detachment-limited conditions respond to increased discharge by first incising the channel bed, increasing the relief between the channel and the hillslopes (Figure 5.8). Only after the channel has come into equilibrium can lateral erosion begin. In the model, the lateral erosion is limited by a volume threshold based on the height of the neighboring hillslope that must be exceeded before lateral erosion occurs. When the channel is incising and relief between the channel and hillslopes increases, the volume threshold grows more quickly than lateral erosion, temporarily stalling lateral erosion. This behavior is similar to narrowing and incision of bedrock channels in response to increased...
uplift or increased discharge [Duvall et al., 2004]. The model predicts that not only will channels in easily eroded bedrock reach equilibrium more quickly than channels in resistant bedrock, but channels in easily eroded bedrock will begin to widen valleys faster than in more resistant bedrock [Lavé and Avouac, 2001].

Two end member approaches to modeling the lateral erosion of bedrock channels are presented. The total block erosion model, in which the entire volume of a neighboring node must be eroded before lateral erosion can occur, best describes the behavior of resistant bedrock. In the undercutting-slump model, the neighboring node need only be undercut over the area of the model cell before the remainder of the node is transported out of the model as wash load, and more accurately reflects behavior from weakly cohesive bedrock that tends to weather into small pieces, such as shale. The behavior of the models varies significantly based on which model is selected, although the same general trends are seen in both models. In nature, lateral erosion will not follow either one of these end members perfectly, but will operate on a continuum between the two [Lancaster, 1998]. Tomkin et al. [2003] presented two end member relationships between channel erosion and valley erosion that are similar to the models presented here, but they found similar behavior between their two models.

Understanding the model behavior in response to detachment- vs. transport-limited behavior, $K_l/K_v$ ratio, runoff rate, and grain size is complex. $\alpha$ is a lumped parameter that describes detachment- or transport-limited behavior in the model and is set by $v_s$, a proxy for grain size, and runoff rate, $R$, although neither of these terms needs to be explicitly set. An $\alpha$ value that appropriately captures detachment- or transport-limited behavior can be set instead. When $K_l/K_v$ ratio is set, runoff rate is explicitly set in order to determine a $K_l/K_v$ ratio that is realistic for natural systems. Once a runoff rate for $K_l/K_v$ ratio is set, by extension the lumped parameter $\alpha$ also has a set runoff rate and $v_s$ value. The high runoff rate in a transport-limited system requires that grain size must be larger in this case compared to a detachment-limited system with the same runoff rate. The models predict more channel mobility and wider flat-bottomed valley development under transport-limited streams with a high runoff rate compared to detachment-limited streams.
with the same runoff rate (Figure 5.2b,d, Figure 5.6b,d). As \( \alpha \) increases, the deposition term increases, and a steeper slope is needed to maintain the landscape in steady state. Potential lateral erosion also increases. The high slopes on the hillslopes in the runs with high \( \alpha \) should balance the additional lateral erosion from the increased channel slope. But the high \( \alpha \) runs create broad valleys that are not rapidly filled in, decreasing the volume needed to erode laterally and allowing the channel to migrate more rapidly. The models also predict much more narrow valleys and less channel mobility in streams that are transport-limited with low runoff rates and small grain size because of the difference in the \( K_l/K_v \) ratio (Figure 5.3). This means that in the model, a landscape with the same \( \alpha \) value and different runoff rates will have vastly different responses in lateral erosion. The limitations that arise from lumping runoff and grain size into \( \alpha \) are discussed below.

While the model captures several important markers of lateral bedrock erosion, the model did not develop broad, smooth, valleys that are many times the width of their channel and that are sustained over many years, as we see in the Front Range of Colorado, for example. The model also did not show a strong relationship between increased sediment flux and protection of the channel bed and increased lateral erosion of valley walls. The major limitations of the model are related to the simplifications made in model formulation, specifically: 1) runoff and grain size are lumped in a single parameter, 2) sediment is not explicitly accounted for in the model, and 3) runoff rate is constant and not applied stochastically.

5.7.1 Critical Role of Sediment

In order to focus on implementing the equations for lateral erosion into the model, the simplest possible erosion-deposition model was used. This erosion-deposition model (Equation 5.1) has the advantage of not requiring the calculation of transport capacity and prevents potential problems with abrupt transitions from erosion to deposition, but does so at the expense of losing the details of runoff rate and grain size, which are lumped into the parameter \( \alpha \). In this model, detachment- or transport-limited behavior is set through \( \alpha \), which works well for general model exploration,
but becomes problematic when exploring specific model responses to changes in runoff rate and sediment size. Setting runoff and grain size explicitly is an important next step for determining how these factors independently impact bedrock valley width and channel mobility.

Another limitation of the current model is that sediment is not treated explicitly, but rather is tracked in the model through the $Q_s$ term. No distinction in erodibility is made between sediment and bedrock. In the current model, when the landscape is in steady state, vertical erosion plus deposition is equal to the uplift rate. Increasing sediment flux, $Q_s$, in the deposition term immediately results in channel aggradation. In model formulations that use the concept of transport capacity of a stream, adding sediment to a river that is far below transport capacity will not cause aggradation, but will easily carry the sediment load downstream. If sediment is continually added to a such a stream, the ratio of sediment flux, $Q_s$, to transport capacity, $Q_t$, will increase until $Q_s/Q_t=1$ and the stream becomes transport-limited [Willgoose et al., 1991]. As $Q_s/Q_t$ for a stream increases, the bed of the stream is progressively covered by more sediment, protecting the underlying bedrock from further incision [Sklar and Dietrich, 2004]. Under these kinds of scenarios, adding sediment to a detachment-limited stream eventually reduces vertical incision, and allows lateral erosion to widen the bedrock channel walls while the bed remains stationary [Hancock and Anderson, 2002].

The addition of sediment in this model does not lead to increased sediment cover on the bed, as bedrock and sediment are not differentiated in the model, but rather results in immediate channel aggradation. This channel aggradation in the model certainly indicates that vertical incision has stopped, allowing lateral erosion to become the primary erosive agent, even in models where $K_l/K_v$ ratio is low or in the total block erosion models. This predicted increase in lateral erosion during periods of aggradation does not occur in most model runs because the temporal and spatial extent of the period of aggradation is limited. For most models, the channel aggradation only occurs in the upper half of the model (Figure 5.12) and aggradation lasts on the order of a few thousand years in model runs with narrow valleys and little lateral erosion. In these model runs, where one might expect to see the most significant effect of reducing vertical incision and allowing lateral
erosion to widen the channel, several thousand years is not enough time to widen the valley. When the channel has aggraded to the point that channel slopes are steep enough to carry the increased sediment load, vertical incision again becomes the dominant mode of channel erosion.

In not differentiating between sediment and bedrock explicitly in this model, the different erodibilities of sediment and bedrock are not accounted for. In most cases, sediment in a channel should be much easier to erode than the bedrock in a channel. But in some cases, sediment in a soft bedrock channel can be composed of coarse grained, resistant lithology sourced from upstream. For example, the streams that drain the Colorado Front Range flow from hard, crystalline bedrock onto soft, friable shale bedrock. The granitic cobbles that armor the channel bed in stream segments underlain by shale bedrock, take much more energy to move than it does to transport the friable flakes of shale that line the walls of the channel.

5.7.2 Variation in Runoff

Channel mobility and valley width are controlled in large part by runoff, $R$, which sets the $K_l/K_v$ ratio. The current model formulation does not allow testing of the impact of stochastic storms and changing $K_l/K_v$ ratios over time on lateral erosion and valley widening. In the model runs that cycle between high and low $K_l/K_v$ ratio by changing runoff (Figure 5.14, 5.15), the $\alpha$ value did not change, which means that grain size must have decreased. In order to keep grain size the same while changing $K_l/K_v$, $\alpha$ must also change. For the terrace model runs, that means that during periods of low runoff, $\alpha$ would be set to 40 instead of 2, resulting in much steeper channel slope and hillslopes. Model runs where changing runoff set both $\alpha$ and $K_l/K_v$ ratio did not run to completion after several days. In order to test the effects of stochastic runoff rate, $R$ and $v_s$ must be set independently.

Variability in runoff rate is hugely important to how lateral erosion works in nature [Hartshorn et al., 2002; Stark et al., 2010]. As the $K_l/K_v$ ratio is proportional to $R^{1/2}$, a distribution of rain storms about a mean would result in a lower overall $K_l/K_v$ ratio compared to a constant storm size. But if the $K_l/K_v$ ratio for a given mean runoff rate is less than 1, then periods of time where
the $K_l/K_v$ ratio is significantly higher could result in valley widening, despite the overall lower average $K_l/K_v$ ratio. Exploring the effect of stochastic storms on thresholds for lateral erosion in the model is an appealing next step.

Let us now briefly return to the terrace-like model runs and describe the necessary conditions needed to develop terraces through lateral erosion as understood through this model. The first key factor for very wide bedrock valleys is easily erodible bedrock. More resistant bedrock undergoes transient responses for longer periods, which can suppress lateral erosion, and the rate of widening is slower for resistant lithologies. High sediment flux and nearly constant transport-limited conditions that inhibit vertical incision of the bedrock are crucial elements for valley widening, especially in the absence of a $K_l/K_v$ ratio that is not $>>1$. Runoff rate may not be critically important, but only if the stream is transport-limited. If a stream is transport-limited and the bed is protected from incision, then even with a low $K_l/K_v$ ratio from a low runoff rate, the valley walls will erode faster than the bed and the valley can widen.

5.8 Conclusions

We have shown that a simple, physics-based theory for lateral bedrock channel migration, when combined with a landscape evolution model, produces several interesting behaviors observed in natural systems. During transient channel incision, lateral erosion in the model temporarily stalls until channel equilibrium is re-established. Following a transient disturbance, wide bedrock valleys develop more quickly in weaker bedrock. The model predicts wider bedrock valleys with easily erodible bedrock, as many have observed in natural landscapes [Montgomery, 2004; Brocard and Van der Beek, 2006]. Weaker bedrock also results in more channel mobility, which is a fundamental factor for developing and maintaining a bedrock valley that is several times wider than the channel it holds [Tomkin et al., 2003]. Increased channel mobility and wider flat-bottomed valleys under transport-limited conditions in the model, suggests that sediment cover on the bed that is present under transport-limited conditions is an effective way to slow vertical incision and amplify the effect of lateral erosion [Hancock and Anderson, 2002]. However, the model lacks some important
elements of reality, especially variations in runoff and separate handling of bedrock and sediment in the channels. Our theory for the lateral erosion of bedrock channel walls and the numerical implementation of the theory in a catchment-scale landscape evolution model is a significant first step towards understanding the factors that control the rates and spatial extent of wide bedrock valleys.
Chapter 6

Summarizing, reflecting, and looking forward

This dissertation set out to answer how climate controls geomorphic processes on a range of scales and in various subsystems in the Colorado Front Range. We have seen that climate controls the rates of the studied geomorphic processes at work in the Front Range, although the time scales of the climate controls can vary from hours or days (torrential flooding) to thousands of years (lateral planation of a strath terrace). In this final chapter I summarize the main conclusions of this work, discuss some of the questions it raises, and review some possibilities for future work that builds on the results presented.

6.1 Climate controls on water fluxes and implications for chemical weathering

In Chapter 2, I showed that the episodicity of recharge, independent of magnitude, can strongly influence water fluxes in the subsurface of mountain hillslopes. Data collected in the Boulder Creek watershed demonstrates that in locations with a seasonal snow pack (e.g., pole-facing hillslopes), soil moisture increases following the spring snow melt and remains elevated for several months. In locations with intermittent snow melt events (e.g., equator-facing hillslopes), soil moisture is more variable through time and water can only flow into the deeper subsurface on rare occasions when the shallow soil is sufficiently saturated. Model calculations imply that the primary control on the speed and extent of water flow below the surface is the episodicity of the recharge. In a model scenario driven by a prolonged period of recharge, more water moves
deeply through the hillslope and recharges the water table, while in a model scenario with the same magnitude of recharge spread out over several shorter-duration events, water moves through the hillslope more slowly and recharge to the water table is reduced by more than 50%. These findings have implications for both chemical weathering and water chemistry in the subsurface and changes in groundwater recharge in the past as well as the future.

These results suggest that locations with a seasonal snow pack may experience more chemical weathering in the subsurface than locations with the same amount of snow melt spread over several events. The link between aspect-controlled differences in water flux and chemical weathering rates has yet to be proven and is a tantalizing, if not necessarily low-hanging, fruit waiting to be picked.

The results from Chapter 2 also suggest that during glacial intervals, when snow pack in the Colorado Front Range was likely deeper and more likely to melt during a single spring event, more water moved through hillslopes allowing more chemical weathering. On the other hand, subsurface water flow in high elevation catchments may have been impeded during glacial intervals if spatially continuous permafrost was present. Increased water inputs to the hillslopes may have also allowed more vigorous development of ice lenses that physically weather bedrock. More flux through hillslopes during glacial intervals would also suggest more groundwater recharge in subalpine mountain catchments. Reducing the area of the landscape covered by a seasonal snow pack will likely reduce water fluxes through the hillslopes, decrease groundwater recharge and slow rates of chemical weathering in the subsurface.

6.2 Extreme climate and effects on channel properties

In Chapter 3, I turned from studying the hillslopes to the channels that bound them and characterized channel geometry and bed sediment grain size and lithology. The two data sets collected before and after an extreme flood event allow the exploration of channel response to spatial changes in lithology, as well as in response to a rare flood event. The engineering of many channel banks in the city of Boulder significantly decreased the chances of lateral migration in the channel during the flood. Lefthand Creek was generally less engineered and lateral migration of
the channel occurred in a few sample locations, but because of the distance between the before and after flood cross sections, it was not possible to give a figure for the magnitude of lateral migration. The most significant observable change in channel form from before and after the flood was large amounts of aggradation on the flood plains, which is a significant hazard itself.

Results from this data set indicate that downstream fining occurs at the study’s largest spatial scales of 10-30 km, but over shorter distances, downstream fining was not observed. Grain size even coarsens downstream over some stream segments. After the flood, grain size generally coarsened, although some segments fined. The strongest response to flooding in the bed sediment characteristics was a distinct increase in the diversity of bed sediment lithology. Before the flood, bed sediment in most locations was primarily composed of granitic material. After the flood, the streams were more likely to have metamorphic, volcanic, or sedimentary rocks in the bed sediment, reflecting more input from more non-local hillslopes compared to before the flood and different temporal scales for the addition of different bed sediment lithologies. Combining this data set with other data sets collected following the flood (e.g., LIDAR, landslide mapping, hydrodynamic modeling, and additional grain size collection) could help expand the impact of these results.

6.3 Evaluation of climate controls on terrace development

Over longer timescales, I tested the effects of climate-controlled variation in discharge and sediment supply on terrace genesis with landscape evolution model simulations. In Chapter 4, I set out to answer the question, “can the streams draining the Front Range transition from detachment-limited to transport-limited solely from changes in sediment supply from the hillslopes or does the carrying capacity of the stream also have to change?”, and concluded by asking the question, “how does the beveling of a terrace surface happen mechanistically and is the lateral migration of channel walls necessary?” The results from this chapter showed that three of the four model scenarios driven by different climate forcings show vertical bedrock incision during periods of low sediment flux or intense rainfall, and aggradation during the opposite phases of a climate cycle. Only the hillslope diffusivity model and the rain intensity-hillslope diffusivity models produced higher
erosion rates in the mountains during glacial intervals, leading to the conclusion that processes that increase hillslope sediment flux must play a role in increasing the sediment load that both promotes lateral valley widening and deposition of sediment on terraces during glacial intervals. An alternate explanation, that increased sediment flux during glacial intervals is the result of a wetter climate, appears to be insufficient to produce both channel aggradation and increased paleo-denudation rates. In nature, the unusual breadth of the strath terraces that flank the Front Range arises from efficient lateral erosion on easily eroded rocks of the High Plains. None of the model scenarios produced flights of broad strath terraces, suggesting that an explicit lateral bedrock erosion rule is indeed required to explain the remarkable breadth of these range-bounding terraces.

In Chapter 5, I presented a theory for the lateral erosion of bedrock channels through shear stress directed at the channel walls and implemented this theory into a landscape evolution model at a drainage basin scale for the first time. Results show that this simple, physics-based theory for lateral bedrock channel migration, when combined with a landscape evolution model, produces bedrock valleys that are several model cells wide. The model predicts wider bedrock valleys with more easily erodible bedrock, as many have observed in natural landscapes. Weaker bedrock also results in more channel mobility, which is a fundamental factor for developing and maintaining a bedrock valley that is several times wider than the channel it holds. Increased channel mobility and wider flat bottomed valleys under transport-limited conditions in the model suggest that sediment cover on the bed is an effective way to slow vertical incision and amplify the effect of lateral erosion. During transient channel incision, lateral erosion in the model temporarily stalls until channel equilibrium is re-established. This theory for the lateral erosion of bedrock channel walls and the numerical implementation of the theory in a catchment-scale landscape evolution model is a significant first step towards understanding the factors that control the rates and spatial extent of wide bedrock valleys. This first effort lacks some important elements of reality, particularly regarding variations in runoff and the separate handling of bedrock and sediment in the channels, leaving the possibilities for future work on this model especially exciting.
6.4 Future Work

This research raises as many questions as it answers, which is a relief to budding scientists who fear that there will be nothing left to study by the time they are ready to make a unique contribution to the scientific community. Here I outline some of the questions raised by my research and possible directions for future work.

6.4.1 Linking climate processes on hillslopes with sediment production on hillslopes

Future work that comes as a natural extension of this body of work should include linking climate processes on hillslopes with sediment production on hillslopes and downslope sediment transport. In Chapter 2, I present evidence for deeper water flow in the vadose zone following the melt of a seasonal snowpack. The next logical step is to test the hypothesis that weathering is deeper and more intense on north-facing slopes or in wetter climates. In order to test the depth of weathering on north- and south-facing slopes, one needs at least two sample profiles from the deep subsurface with the same parent lithology through the entire profile. One of the limitations faced on the north- and south-facing slopes in Gordon Gulch when attempting to answer this question was the heterogeneity of the parent material. Future work will include the continuation of a study I began on weathering profiles exposed from road cuts. Road cuts give a broad, deep view of the subsurface and expose the entire weathering profile from the top of the soil, through saprolite and down into fresh, unaltered bedrock (Figure 6.1). I collected 12 samples and made Schmidt hammer measurements where possible along an 8.2 m deep profile on a sunny, south-east facing road cut on Sugarloaf Road, just off of Boulder Canyon. The samples were prepared and run for XRD mineral analysis. The goal of the project was to link degree of chemical weathering with reduction of rock strength. The next step in this project is to repeat the profile measurements on a north-facing slope at the same altitude, perhaps on the other side of the canyon. The two sample sets would have the same mean annual air temperature and precipitation, but as suggested in Chapter 2, I
expect that the north-facing profile would have a deeper weathering depth and a greater reduction in strength.

Figure 6.1: Vertical weathering profile sampled at 10–50 cm intervals on a road cut on Sugarloaf Road. Profile shows transition from unaltered granite to saprolite to soil.

Another remaining question is how do catchments provide enough sediment to form terraces during glacial intervals with no direct input from glacial sediment sources? Chapter 4 gives some insight into the question with numerical modeling, but more advances could be made with additional field work. This question could be answered with a detailed study of the sediment capping the terraces. Where in the catchment is sediment sourced from? How does grain size carried by
streams during terrace forming periods compare to current grain size in streams? What can we learn about flow conditions during glacial intervals compared to interglacial intervals from paleoflow reconstructions? Currently, other researchers are making valuable contributions by attempting to establish paleodenudation rates for the interval when terrace sediments were deposited and ages for the terraces.

6.4.2 Lateral erosion modeling: the next steps on a long journey

Chapter 5 presented groundbreaking work on modeling the lateral migration of bedrock channels. The implementation of the lateral erosion algorithm is robust and should be modified to work with hexagonal and non-uniform grids as well in order to test its sensitivity to these model procedural choices. Research on this topic will continue with the next steps focused on testing other erosion-sediment transport models that represent the physical processes necessary to accurately describe lateral erosion, especially sediment cover effects on the bed.
Bibliography


