

**Facing reality: Late Cenozoic evolution of smooth peaks, glacially ornamented valleys, and deep river gorges of Colorado’s Front Range**

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**ABSTRACT**

Thirty to forty m.y. of post-Laramide degradation of the southern Rocky Mountains likely produced relatively low-relief topography within the crystalline cores of the ranges, and capped the adjacent sedimentary basins with easily eroded sediments. We focus on the modern, more dissected topography of these ranges, reflecting late Cenozoic evolution driven by fluvial and glacial exhumation, each of which affects different portions of the landscape in characteristic ways. Ongoing exhumation of the adjacent basins, in places by more than 1 km, is effectively lowering base level of streams draining the crystalline range cores. The streams have incised deep bedrock canyons that now cut the flanks of the range. Over the same time scales, glaciation of the headwaters of the major streams has modified the range crests. We utilize the topography of the northern Front Range of Colorado to explore the response of a Laramide range both to the exhumation of the adjacent basin and to glaciation in the high elevations. We break the problem of whole landscape evolution into three related, one-dimensional problems: evolution of the high smooth summit surfaces; evolution of the longitudinal profiles of adjacent glacial troughs; and evolution of the fluvial profiles downstream of the glacial limit. We review work on the high summit surfaces, showing quantitatively that they are steady-state features lowering at rates on the order of 5 µm/yr, and are entirely decoupled from the adjacent glacial troughs. Glaciers not only truncate these high surfaces, but greatly alter the longitudinal profiles of the major streams: major steps occur at tributary junctions, and profiles above the glacial limit are significantly flattened from their original fluvial slopes. We extend existing models of glacial valley evolution by including processes that allow headwall retreat. This serves to enhance the headward retreat of east-facing valleys, and explains the asymmetric truncation of the high smooth surfaces that form the spine of the range. Fluvial profiles downstream of the glacial limit commonly display a prominent convexity inboard of the range edge. Stream-power–based numerical models of profile evolution of specific rivers demonstrate that this reflects a transient response of the streams to base-level lowering. This response varies significantly with drainage basin area. We explore the degree to which this differential response controls the...
location of major remnants of pediments on the edge of the Great Plains, such as the prominent Rocky Flats and adjacent surfaces.

Keywords: glacial valleys, river incision, erosion, landscape evolution.

INTRODUCTION

The Laramide Ranges of the western United States were generated by compressional tectonics in the late Cretaceous and early Tertiary (e.g., Dickinson et al., 1988; Bird, 1998; Erslev, 2001). These crystalline-cored ranges are bounded by vast basins filled with syn- and post-tectonic sediments that partially buried this region tens of millions of years ago. Because the region has been relatively quiescent since the early Tertiary, its post-tectonic history has been inferred largely from qualitative arguments about how a suite of distinctive and characteristic landforms might have developed. These landforms include widespread surfaces of low relief within the ranges, narrow, glaciated “spines” at the range crests, deeply incised fluvial canyons at the range edges, and remnants of former river valley bottoms preserved as benches and surface fragments in the sedimentary basins. To clarify the post-tectonic history of the Laramide region, we consider it essential to couple conceptual models with quantitative models of Laramide landform evolution. In this paper, we use a set of one-dimensional (1D) numerical models to simulate the evolution of several key landforms. The models are simple but incorporate many essential details of real systems. Our efforts aim to complement previous empirical work and to place additional constraints on the timing, magnitude, relative importance, and linkages of the geomorphic processes that have sculpted the Laramide landscape for the last few million years.

The most recent significant geomorphic event affecting the Laramide region involves the exhumation of the range bounding, sediment-filled basins. Many of these basins have been eroded to depths of 1 km or more (e.g., McMillan et al., 2002; Leonard, 2002). The exhumation began in the late Cenozoic, although the timing is poorly constrained and very likely varies from basin to basin (e.g., Dethier, 2001; Reiners and Heffern, 2002; Rihimaki, 2003). The Laramide landscape also reflects the influences of both regional and local geophysical and tectonic processes. Broad-scale tilting of late Cretaceous seaway deposits, and of gravel sheets emplaced in the mid-Cenozoic is thought to reflect dynamic topography driven by mantle flow (Mitrovica et al., 1989; Heller et al., 2003), while down-to-the-east tilting of paleochannels of mid-Cenozoic age in deposits of exterior basins to the east is claimed to reflect both local tectonics and exhumation of the basin deposits (McMillan et al., 2002; Leonard, 2002). Possible mechanisms driving local tectonic activity include the propagation of the Rio Grande Rift into Colorado from the south (Chapin and Cather, 1994; Erslev, 2001), and the arrival of the Yellowstone hotspot in the NW of the province (e.g., Smith and Braile, 1993).

Despite these local effects, we expect the post-tectonic history of the Laramide ranges to be characterized largely by topographic decay. Low-temperature geochronology reveals that no more than a few kilometers of erosion have occurred within the crystalline cores of the ranges since the Laramide orogeny or shortly thereafter; apatite fission-track ages, documenting exhumation from roughly 120 °C, are early Tertiary (e.g., Crowley et al., 2002; Kelley and Chapin, 1997; Kelley and Duncan, 1986). Calculated mean Tertiary erosion rates are typically less than ~0.1 mm/yr, although there is significant scatter. More recently, the ranges have experienced topographic rejuvenation linked to basin exhumation (e.g., Izett, 1975). The mechanism driving exhumation in the Laramide region is not known, and we do not address that question in this paper. Rather, we focus on the response of the adjacent ranges, in particular the Front Range, to basin exhumation. In effect, the basin exhumation, driven by the incision of the master streams within them, serves to lower of base level for the fluvial system draining the adjacent range. We may therefore treat this portion of the southern Rockies as a natural experiment that allows us to probe the response of a partly buried landscape to both lowering of base level and to climatically driven erosion processes (e.g., glaciation).

In part, this paper constitutes a review and assembly of our past work (e.g., on high surfaces [Anderson, 2002] and glacial valleys [MacGregor et al., 2002]). Herein we extend our models and our discussions of these features, we address the linkages among them, and we address for the first time the fluvial system as it responds to the base-level lowering of the South Platte. The features we aim to explain in the Front Range specifically, and for the Laramide ranges in general, are described below.

Introduction to Front Range Physiography

The Front Range is roughly 50 km from the base of the Flatirons at the mountain front to the continental divide (Fig. 1). The range is bordered by the very flat floor of Denver Basin at 1500 m elevation, into which a number of low-gradient smooth surfaces (pediments, pediments buried by alluvium, and more narrow fingers of alluvium) protrude from the Front Range. Collectively, these geomorphic surfaces and deposits define former levels of the Denver Basin as it has been exhumed. Among these surfaces, the most prominent is the Rocky Flats pediment between Boulder and Golden (e.g., Scott, 1960, 1975). The range steps upward abruptly by 800 m from the adjacent plains. If one avoids the canyon floors, a walk from plains to crest would then encounter a rolling surface at 2300–3000 m elevation that stretches toward the spine of the range. This physiographic feature has been called the subsummit surface, or the Rocky Mountain surface (e.g., Epis and Chapin, 1975). The topography then steps up by another 1000 m to the narrow summit spine that forms the drainage divide (Continental
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Divide) between the Colorado and Mississippi Rivers. The few-kilometer-wide range crest is dominated not by glacial arêtes, but by smooth tundra-covered surfaces, with occasional bedrock knobs poking through shallow regolith. It is easy walking; most of the work of the climb to the summits is done getting to the ridge. A walk up the river canyons is significantly different. Following any of the gently sloped rivers on the plains, all tributaries of the South Platte, into the mountain front, one ascends through a steepening, deep bedrock gorge with walls that at first are quite narrow and steep. Above an inflection in the profile, many of these rivers become less deeply incised into the subsummit surface. Further upstream one encounters the end moraines from the glaciers that have periodically occupied the headwaters. Thereafter, the character of the valley profile alters significantly. Major steps and flats dominate the profile up to the tip of the drainage, where the channel form halts, and one is left standing in a cirque basin or a bench just below the divide.
It is these major features (Fig. 2), held in common among those Laramide ranges that have not experienced late Cenozoic tectonism, for example tectonism associated with the Rio Grande Rift, that we wish to explain quantitatively: the high smooth surfaces of the summit spine, the glacial canyons that bound them, the fluvial canyons downstream of the glacial troughs, and the pediments on the edge of the plains. Below we describe the conceptual framework upon which our numerical modeling is based.

Conceptual Model of Late Cenozoic Landscape Evolution

It is likely that in the roughly 40 m.y. between the end of the Laramide orogeny and the initiation of basin exhumation, the topography of the range had decayed significantly, although perhaps not monotonically (e.g., Evanoff, 1990; Mears, 1993). The relief between the crest of the range and the basin floor has become diminished as the products of the weathering of the range have been transported to the nearby depositional basin. The basin floor used to be significantly higher than it is now, as witnessed by the remnants of older surfaces on the fringe of the plains. The local relief between river and ridgeline was likely small as well. By the late Cenozoic, both the subsummit surface and the summit spine had likely evolved into low-relief surfaces adjusted to transport the products of weathering of the crystalline bedrock of the core of the range.

With the initiation of exhumation of the adjacent Denver Basin driven by incision of the South Platte River, the Front Range landscape was rejuvenated. Tributaries of the South Platte then responded by incising into the crystalline core, and are presently in a state of transient response to the exhumation event. The hillslopes of the subsummit surface remain relatively intact, showing only minor relief where they exist beyond the reach of this transient response. The exceptions are the steep canyons of the front of the range. That the exhumation occurred within the last few m.y. means that the climate was undergoing strong glacial-interglacial cycles. This resulted in major glacial incision of the headwaters of these tributary streams, which in turn both increased the local relief of the spine, and fed the fluvial system downstream with glacial sediment.

In the following sections, we describe in detail our current quantitative understanding of the geomorphic processes that generated key Laramide landforms, beginning with the high surfaces in the montane interior and working basinward to the pediments carved into sediment fill.

HIGH SURFACES

A large fraction of the spine of the Front Range is a very smooth surface. A walk on the Continental Divide is for the most part a stroll in alpine meadows, only occasionally interrupted by glacially carved aretes. We first briefly summarize the set of geomorphic features and processes acting on these high surfaces, and the recently published numerical model of them (Anderson, 2002). We then apply this model to demonstrate how morphologic asymmetry of the range crest can be produced by asymmetric (east versus west) headward retreat of glacial canyons in the Front Range.

Features of the High Surfaces

Aerial photos and digital elevation models (DEMs) (Figs. 3 and 4) show that these surfaces are truncated by glacial headwaters, more extensively on their eastern than western sides. This strong glacial asymmetry of the range has long been noted. The high surfaces of the Front Range, like those of the Wind River range described in Anderson (2002), have the following features in common: broad convex-up parabolic profiles, thin (roughly 1 m) but uniform regolith cover, occasional tors, which are commonly found along the crestline, and bedrock edges that separate the surface from bounding glacial headwalls. Some of these surfaces are more than 1 km wide. The smooth quality arises not only from the lack of significant roughness on the surface, but also from the low curvature, typically $0.5-8.0 \times 10^{-4}$ m/m (see Fig. 5). The tors are often surrounded by block fields, in which large blocks on the order of 1 m diameter dominate the surface of the otherwise smooth surface.

At present, these surfaces lie at elevations of roughly 3–4 km, and experience mean annual temperatures that subject them to periodic freezing. They display the activity of periglacial processes. Sorted nets and polygons are occasionally present, and frost heave of individual blocks is common. Both the bedrock of the tors and the blocks derived from them are often riven by frost shattering.
Figure 3. Mid-November photograph of summit spine of the Front Range, looking north from above Berthoud Pass. The highest summit surfaces are white from a skiff of recent snow; the subsummit (Rocky Mountain) surface is dark green with forest cover. The floor of the Denver Basin at the eastern edge of the Great Plains is in the distance on the right.

Figure 4. Photograph of representative high surfaces on the Continental Divide. James Peak is the most prominent peak on the divide in the scene. Note abrupt truncation of surfaces on several sides by glacial valley walls and headwalls. The high surface in the foreground slopes gently toward the viewer, and displays several regolith-mantled fingers that are bounded by tor-studded ridges.
The rate of weathering in the landscape is difficult to measure. However, application of cosmogenic radionuclides to this problem has yielded constraints that are consistent in crystalline rocks from the Sierras to the Laramide ranges (Small et al., 1997). Bare bedrock weathering rates are commonly 3–15 μm/yr. Working in the Wind River Mountains on analogous high smooth surfaces, Small et al. (1997) showed that these rates match well with the rates derived from bedrock samples beneath the regolith. Small et al. (1997) argued that this implies that the surfaces in the Wind River Mountains are in steady state; most importantly for the evolution of local relief, they indicate lowering at rates on the order of 5 μm/year (5 m/m.y.). The present profiles of the Front Range crest surfaces (Fig. 5), and the few measured bare bedrock rates from the range crest, imply that these surfaces are eroding at comparably low rates, and have achieved steady-state forms. We review below the quantitative basis for this conclusion.

Model of High Surface Evolution

Conservation of regolith demands that:

$$\frac{dh}{dt} = \frac{P}{\rho_b} w - \frac{dQ}{dx}, \quad (1)$$

where $h$ is the thickness of regolith, $t$ is time, $Q_r$ is flux of regolith, $w$ is the rate of production of regolith by weathering, $\rho_b$ is the density of rock, and $P$ is the bulk density of the regolith. Here we have made use of the axial symmetry of these surfaces in order to cast the problem in one dimension, in distance from the divide, $x$. We have also assumed that the deposition of dust is negligible, and that chemical dissolution of regolith is a minor contribution to the evolution of the surface (see discussion in Anderson, 2002). This equation is coupled to the complementary equation for the lowering of the bedrock interface:

$$\frac{dz_b}{dt} = -w, \quad (2)$$

where $z_b$ is the elevation of the regolith-bedrock interface, and the topographic surface is simply $z = z_b + h$. Solution of these equations requires rules for both the generation of regolith by weathering, $w$, and transport of regolith, $Q_r$.

In order to explain the existence and location of tors in the landscape, Anderson (2002) argued that the weathering rate must be a function of regolith thickness, which displays a strong maximum at a finite regolith thickness. One such rule is (Fig. 6, inset):

$$w = \frac{dz}{dt} = \min \left[ A_b + bh, A_c e^{-\frac{h}{\zeta}} \right], \quad (3)$$

where $\min$ refers to the minimum of the two terms. The first equation linearly increases from a minimum rate of $A_c$ on bare bedrock ($h = 0$), where $b$ is the rate of increase with regolith thickness, while the second equation decays exponentially with regolith thickness, where $A_b$ is the bare bedrock weathering rate, and $h$, is the length scale for decay. The maximum weathering rate occurs at the regolith thickness at which the two curves cross.

We assume frost creep dominates the transport of regolith downslope (Anderson, 2002), although similar formulation of a regolith transport rule could be carried out for other dominant processes. Following the field and theoretical work of Matsuoka and Moriwaka (1992), and acknowledging that regolith discharge should not occur in the absence of regolith(!), the discharge of regolith per unit width of hillslope becomes

$$Q = f \beta \left[ \xi_x^2 - e^{-k_c \zeta} \left( h_x^2 + \xi_x^2 \right) \right] \frac{dz}{dx} = k \frac{dz}{dx}, \quad (4)$$

where $f$ is the annual number of frost events, $\beta$ is the soil strain associated with ice lensing, and the probability distribution of frost penetration depth decays exponentially with depth, with a characteristic depth scale of $\zeta$. Here $k$ is an effective landscape diffusivity associated with the frost-creep process. Importantly, the discharge smoothly feathers to zero as the regolith approaches zero thickness (Fig. 6, inset), and approaches the maximum discharge appropriate for the slope, $dz/dx$, as $h >> \zeta$. Note that in the case of uniform regolith, the efficiency of transport (represented by $k$) is uniform, and discharge becomes linear in slope. This should result in diffusive behavior of the hillslopes.

If the hilltop is at steady state, the regolith thickness must be uniform, so that the weathering rate of the bedrock interface is uniform. Such hilltops should display uniform curvature,

$$c = \frac{\partial^2 z}{\partial x^2} = \frac{\rho_b w}{\rho_b k}, \quad (5)$$

meaning that their profiles should be parabolic. Therefore, if the hill crest has achieved a steady-state form, one should be able to fit topographic profiles across the hilltops using

$$z = z_{top} - \frac{c}{2} \left( x - x_{top} \right)^2, \quad (6)$$

where the crest of the surface lies at $(x_{top}, z_{top})$.

High Surfaces of the Continental Divide

In Figure 5, we show four examples of curve fits from the summit surfaces of the Front Range, using 30-m-resolution DEMs available from the U.S. Geological Survey. While the curvature varies by roughly an order of magnitude among these examples, all may indeed be well fit using a parabolic profile. We therefore argue that these surfaces are in steady state. If so, and if the lowering rate associated with weathering is on the
order of 5 m/m.y., then the topographic diffusivity $k$ may be constrained to lie between 0.2 and 0.02 m$^2$/yr (assuming $R_r/R_b \approx 2$). The variation in curvature can be attributed to variation in both the weathering rate, $w$, and the efficiency of frost creep, $k$, with microclimate.

That the surfaces may be in steady-state may seem surprising given that the present profiles often display abrupt truncation by headwalls and valley sides that are clearly of glacial origin (Figs. 4 and 5). Anderson (2002) argued that these surfaces are fully decoupled from the glacial valleys beside them. The decoupling occurs at the bare bedrock rims that bound the surfaces. The rims are kept bare by the strong curvature there (regolith is removed down the steep headwall more rapidly than it is delivered to the rim), and therefore they decrease at rates dictated by the (very slow) bare bedrock weathering rate. This is illustrated in Figure 6, in which we numerically simulate the evolution of a high surface adjacent to a glacial valley. Glacial valley deepening and extension (here simply imposed by dictating a boundary

Figure 6. Simulation of high surface profile evolution. Inset: model rules for weathering rate, and for the transport of regolith as a function of regolith thickness. Transport rate feathers to zero for zero regolith thickness. Freeze event frequency = 5/yr; characteristic freeze depth = 0.25 m; slope = 0.10. (A) Profile evolution over 20 m.y.; time steps of 1 m.y. shown. Blue line is final profile. Glacial canyon evolution is crudely mimicked by both deepening and backwearing of valley at the right boundary for the last 3 m.y. of simulation. Smooth parabolic profile is simply truncated by backwearing. A small bare bedrock tor adorns the crest. (B) Evolution of regolith thickness profile. Original uniform thin cover (0.1 m) evolves to a uniform cover of roughly 1 m at end of simulation (blue line). This corresponds to the thickness of regolith needed in the weathering rule such that the bare bedrock edge of the surface and the bedrock beneath the regolith are lowering at the same rate (see arrow in inset in A), as required for a steady-state form.
retreat rate on the right-hand side of the simulation) are allowed to occur only within the last 3 m.y. of the 20 m.y. simulation. As expected, the surface itself is “ignorant” of the evolution of the glacial valleys adjacent to it. It shows no signs of reacting to the rapidly eroding boundaries.

**Tors with Block Fields—A Role for Lightning?**

One feature of the high surfaces bears further discussion. Anderson (2002) noted that when they exist, tors (bare bedrock knobs) are largely confined to the crests of these surfaces. He argued that the weathering rule of the sort used here can result in tors, while a rule in which weathering rate simply exponentially decreases with regolith thickness cannot. As can be seen in two of the profiles presented in Figure 5, and in the photograph of Figure 4, knobs do ornament the crests of some of these surfaces. Some of these tors are indeed isolated bedrock outcrops. In many instances, the tors are surrounded by block fields (felsenmeer), consisting of blocks ranging up to >1 m in diameter. The blocks have clearly originated from the outcrop or tor. These block fields display an open framework, with little if any fine-grained regolith in the interstices between blocks, at least in the top meter.

The origin of these block fields is difficult to explain. Repeated frost-cracking is commonly invoked (e.g., Washburn, 1979), which can certainly explain some fraction of the degradation of the outcrop (tor) upslope of the block field. However, in more than one instance, one of us (Anderson) has observed a block of roughly 1 m diameter that is resting on the bedrock surface of a tor at a distance of at least 1 m from a joint-bounded hole from which it must have come. No frost action that we know of could have hoisted the block out of its hole and placed it on the surface so far away. We posit that lightning may be an important process on these surfaces, inspired by the following observations. (1) Every climber of these mountains knows to get off the summits by mid-afternoon in the summer, as thunderstorms pound the mountains, charging the atmosphere with ice-axe buzzing electricity. (2) Evidence of lightning strikes is common. Paleomagnetists who sample rocks from mountainous areas know that summits are places to avoid, as it is difficult to find samples not magnetically reset by lightning (S. Bogue, 1996, personal commun.; R. Coe, 1996, personal commun.). (3) Walking the Continental Divide in this same area in the summer of 1969, one of us (Anderson) witnessed 11 sheep being thrown roughly 2 m into the air by lightning. In addition, the ball lightning dug a trench through the regolith roughly 50 m long, 10–20 cm wide, and 5–10 cm deep (Fig. 7). Many eyewitness accounts exist of heavy objects being thrown into the air by lightning (e.g., Uman, 1969). We therefore hypothesize that lightning strikes on these slowly eroding surfaces may be responsible for both cracking apart outcrops and moving the resulting blocks of rock. The mechanism by which this might operate, propelling rocks this massive such distances, could involve water flashing to steam within joints between blocks. While there is no doubt that enough power exists in lightning strikes to accomplish this work, one might reasonably ask whether lightning occurs with sufficient frequency and density to be worthy of serious consideration. The world map of lightning strike density (http://science.nasa.gov/headlines/images/lightning2/lightningmap_large.gif) shows that lightning strikes roughly 20–30 times per year per km² in the Laramide region. As the resolution in these maps is low, we would expect that the enhancement in mountainous areas renders this a minimum estimate. Assuming that this frequency is appropriate over long time scales, assuming it is spatially random, and ignoring the likely enhancement factor, every 1 m² of the high surface should be struck about every 50 k.y. If a 1 m block is indeed removed (eroded) from the surface upon impact, then the lowering rate would be 1 m/50,000 yr, or 20 µm/yr. This is clearly in the range of the 5 µm/yr lowering rate measured using cosmogenic radionuclides (Small et al., 1997), implying that this process occurs with sufficient frequency to be a viable candidate for the degradation of tors and the origin of the felsenmeer on these particular surfaces. Clearly, this process will not operate
in arctic environments, where other spectacular tors exist, meaning that one must appeal to other processes there. Further study is needed; our hope in including this speculative section is to inspire such study.

### Summary of High Surfaces

In summary, the high smooth surfaces that characterize the summit spine of the Front Range are likely steady-state surfaces that operate in a manner not unlike that described in the word-picture of G.K. Gilbert (1909). The processes of weathering and of downslope transport are periglacial and have likely remained so throughout the late Cenozoic. We speculate that lightning plays an important role in detaching blocks on the crests of these surfaces. They are eroding at rates that will likely have lowered them only 10 m over the entire Quaternary, while the neighboring glacial valleys have eroded at rates several orders of magnitude faster, averaged over the many glacial cycles of the Quaternary. That these valleys currently truncate the high surfaces, leaving in some cases less than half of the original surface intact, implies that the glacial valleys have both deepened and migrated headward over the Quaternary. It is therefore inevitable that the local relief in the headwaters of the Front Range has increased dramatically since the inception of late Cenozoic glaciation. The ornamentation, and the aesthetic drama of the high parts of the range are due to the operation of glaciers.

### GLACIAL CANYONS

Glaciers leave a distinctive signature in the landscape, both in cross valley (e.g., Harbor et al., 1988; Harbor, 1992) and longitudinal valley (MacGregor et al., 2000; 2002) profiles. The Front Range is particularly instructive, because the glacial limit (Madole et al., 1988) lies roughly halfway between the summit spine and the range front. One can therefore distinguish the signatures between fluvial and glacial erosion on the landscape. In contrast, for example, the glaciers of the Wind River Range were so extensive that they exited the range and gouged deeply into the easily eroded rock of the adjacent basin floors.

Consider the profiles of Clear Creek and its tributaries (Figs. 8 and 9). While the lower portions of the trunk valley profile are smooth, the upper reaches display multiple steps and flats, many of which occur at tributary junctions. In a following section, we discuss the broad convexity of the fluvial profile, but focus here on the steps and flats in the glacially impacted headwaters. MacGregor et al. (2000) argued that this longitudinal valley signature of glacial occupation is dictated by the long-term discharge of ice through the system. Numerical simulations of longitudinal valley profile evolution showed patterns that were robust against the details of the erosion rule used. In the simulations they reported, the first manifestation of glacial erosion is the flattening of the original stream profiles occupied by the glacier. The slope-area relationship so characteristic of fluvial systems is overwritten.

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Figure 8. Digital elevation model (DEM) of the Front Range centered on Clear Creek, showing prominent tributaries. See Figure 10 for longitudinal profiles of creeks. B—Berthoud Pass, GT—Grays and Torres peaks, E—Mt. Evans, L—Loveland Pass. Clear Creek exits the mountain front to run between the North (N) and South (S) Table Mountains at Golden.
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as the more efficient glacial erosion lowers the valley floor. The simplest explanation for effects of glacial occupation of a valley is as follows. Model results incorporating subglacial erosion rules of various types have all shown that erosion rate is roughly proportional to the ice discharge (MacGregor et al., 2000; Fig. 10). Recall that the instantaneous ice discharge in a glacial system reaches a maximum at the equilibrium line elevation (ELA), and feathers to zero at both the headwall and the terminus. The resulting pattern of erosion should mimic the pattern of ice discharge. The erosion should lead to steepening of the original profile between the ELA and the headwall, and flattening of the valley floor between the ELA and the terminus. Over many glacial cycles of varying duration and intensity, this pattern of erosion shrinks and swells. The integration of the erosion pattern over many cycles yields the pattern of total erosion.

Figure 9. Profiles of several creeks draining the east side of the northern Front Range. Note prominent convexity in each main profile, reflecting capture of the river in the midst of a transient response to lowering of base level on the South Platte. (A) St. Vrain Creek, (B) North and Middle Boulder Creeks, (C) Clear Creek, and (D) Bear and Turkey Creeks. Tributaries reaching high elevations show steps and flats associated with glacial modification (e.g., above join of BP and LP in Clear Creek; and above triangle [glacial limit] in Boulder Creek profiles). uK—upper Cretaceous; T—Tertiary; LP—Loveland Pass; GT—Grays and Torreys; WC—West Chicago Creek; FR—Fall River; BP—Berthoud Pass; P, N—Pine and North Clear Creek; BB—Beaver Brook; GGG—Golden Gate Gulch.

Figure 10. Simulation of glacial longitudinal valley evolution. (A) Evolution of longitudinal profiles from initial fluvial profiles (gray) to final glacial profiles (black) in the face of repeated glacial occupation of the valley. Inset shows history of deep-sea δ18O that serves as a proxy for the climate history forcing precipitation and melt histories over the 400 k.y. simulation. The step in the valley floor corresponds to a junction with a 10-km-long tributary valley. The tributary valley is left hanging above the trunk valley floor. (B) Profiles of total ice discharge and of total erosion over the simulation. The erosion pattern closely mimics that of ice discharge. The increase in total erosion downstream of the junction reflects the added ice from the tributary (after MacGregor et al., 2000).
The glacial signature is strong at tributaries as well. Long-term ice discharge and hence total erosion jump abruptly at tributary junctions (Fig. 10). In the main valley, the increase in ice discharge leads to greater erosion and a step in the long profile of the valley. In addition, the disparity in ice discharge between small tributaries and the trunk valleys leads to the hanging of tributary valleys (Fig. 10). The height of the hang should be more dramatic the greater the disparity in long-term ice discharge, while the amplitude of the step in the main valley will decrease (MacGregor et al., 2000). We note the correspondence of the major steps in the Clear Creek profiles with tributary junctions.

While useful in addressing first-order issues of glacial longitudinal valley morphology, these early simulations leave several issues unresolved. The simulations were unable to produce overdeepened cirques. As there were no supraglacial erosional processes included in the model, glacial headwall retreat could not occur. Instead, headwalls simply continued to steepen as valley erosion continued. This runs counter to the suggestion that considerable headwall backwearing has occurred in some alpine settings (e.g., Brocklehurst and Whipple, 2002; Whipple et al., 1999). The subglacial erosion rules used were very simple, either invoking the empirically derived relation between sliding and erosion rate (Humphrey and Raymond, 1994), or a theoretical relation between sliding and abrasion (Hallet, 1979).

In an attempt to target these issues, and to make the simulations more relevant to situations akin to those found in the Laramide ranges, MacGregor (2002) updated the model rules. The modifications included the presence of a smooth high surface (or plateau) above the headwall of the glacier and a suite of headwall processes, including blowing of snow off the high surface, avalanching of snow down steep headwalls, frost-cracking and erosion of the bare rock headwall above the glacier, and a more process-specific set of subglacial erosion rules. We merely outline these alterations here, and focus on the model results.

**New Processes Included in Glacial Valley Profile Simulations**

Inclusion of the smooth high surface (plateau in Fig. 11) allows the high surface to act as an additional source of snow for the head of the glacier. This reflects the present-day observation that these high surfaces do not accumulate great depths of snow in the winter (e.g., the February photograph of Fig. 4). Rather, the snow blows off the surfaces to accumulate as large seasonal cornices that overlook the glacial cirques below, largely to the east (downwind). We assume that this same eolian sweeping of the surfaces occurred over the glacial epoch, and that the cornices acted as an additional source of snow. We incorporate avalanching of the snow onto the valley floor simply, by assuming a threshold slope for complete, annual avalanching. In addition, we enforce conservation of cross-sectional area of the snow, and dictate a reasonable run-out pattern onto the glacier. This serves two purposes: to augment the positive mass balance of the glacier near its headwall, and to remove snow cover from the steep headwall, exposing it to subaerial geomorphic processes (Fig. 11). The bare headwall is allowed to erode at a rate dictated by freeze-thaw activity (e.g., Matsuoka and Sakai, 1999). The bedrock particles freed by this process are then delivered either to the ice surface or the bed of the glacier via the bergschrund. The subglacial erosional rules include explicit treatment of both glacial quarrying and abrasion. The processes are linked in that quarried material supplies the tools for the abrasion of bedrock down-valley. The only other source of tools is clasts derived from erosion of the headwall, which encounter the glacial bed either immediately (through the bergschrund) or some distance downstream from the headwall (e.g., Lliboutry, 1994). In addition, the population of clasts is allowed to evolve down-valley. Clasts wear down as they are employed to abrade the bed, for a distance that is dependent on their original size. Finally, the entire valley is allowed to rise isostatically as rock is removed from the valley. This calculation is performed assuming either that the unloading driving the rebound is (1) only local, driven by the glacial erosion pattern, or (2) more regional, principally driven by exhumation of the adjacent sedimentary basin (e.g., Small and Anderson, 1998).
Inclusion of these processes in the longitudinal valley evolution simulations results in more realistic deepening and extension of the glacial valleys (Fig. 12), and can in some instances generate glacial cirques (MacGregor, 2002). Model runs targeting the details at the valley head (as opposed to tributary junctions) show that significant headwall retreat is expected to occur; this could not happen in the earlier model simulations, as no headwall processes were incorporated. In the model run depicted in Figure 12, the climate forcing is taken to be an asymmetric triangular pattern through time (e.g., Harbor, 1992), reflecting slow decay into glacial climates, and rapid emergence from them. During simulations of 400 k.y. periods, headwalls retreat by ~1 km, enough to truncate significantly the high smooth surfaces described above. While isostatic response to the removal of rock from these valleys drives tens of meters of uplift, it is insufficient to counter the erosional downwearing that effectively lowers the valley relative to the climatically determined ELA. This feedback, first appearing in Oerlemans’ (1984) simulations, would yield smaller and smaller glaciers through uniform climate swings in the absence of either a deepening of the glacial cycles, or considerably more uplift than the glacial incision alone can generate.

**Sediment Output History: The Fluvial Transition**

One of the most interesting elements in an alpine geomorphic system is the glacial-fluvial interface. Glacial erosion produces vast amounts of both fine and coarse sediment, which is subsequently handed over to the fluvial system. While the initial impact of this sediment delivery to the fluvial system is immense, the timing of sediment delivery may dictate the timing of river downcutting and terrace formation far downstream of the glacial-fluvial interface (Hancock and Anderson, 2002). The sediment output history from glacial headwaters strongly varies over the glacial cycle. Averaged over a relatively short time scale associated with subglacial sediment storage, sediment output from the glacier must be the integral of the erosion rate over the entire subglacial footprint.

![Figure 12. Glacial valley profile evolution, including headwall retreat (after MacGregor, 2002). (A) Climate history used in the 400 k.y. simulation: temperature decreases linearly for 80% of a 100 k.y. cycle, stays cold for 10% (glacial maximum), and warms rapidly during the final 10%. The total amplitude of the temperature change is 4 °C. (B) Time series of sediment and water output from the glacier, showing very strong modulation of sediment output over a glacial cycle. (C) Valley profile evolution after 400 k.y. The headwall is dramatically steepened and lengthened, and retreats by roughly 1 km. The lower 10 km of the valley floor flattens. Successive glaciers become shorter as the erosion lowers the valley into warmer microclimates. Long-term mean location of the equilibrium line altitude (ELA) is shown, along with its down-valley location. (D) The cumulative patterns of abrasion (gray) and quarrying (black). Quarrying is most important near the valley headwall, while abrasion dominates down-valley and is closely tied to the glacier size.](image-url)
In Figure 12B, we show the sediment output history from the simulation of valley evolution already described. The sediment output varies with the size of the glacier and with climate, as expected. We note that the impact on the fluvial stream will be largely derived from the bedload component of the sediment load, as this affects the channel morphology and the thickness of alluvium in a reach. While the glacial flour (fine silt in suspension) emanating from a glacial terminus stream is quite visible, it is not appropriate to assume, as is commonly done, that the bedload is a small fraction of the suspended load (usually taken to be 10% in nonglacial settings). In recent work on a small valley glacier in Alaska, it has been shown (Riihimaki, 2003; Riihimaki et al., 2005) that when it can be measured in the system, bedload comprises almost 30% of the load. In addition, recent work on a glacially impounded lake into which glacially derived sediments are debouched demonstrated that the volume of coarse load trapped in the delta to the system is approximately equivalent to the volume of fine-grained silt deposited across the basin floor (Loso et al., 2004).

**Summary of Glacial Impacts on the Landscape**

In summary, glaciation of the headwaters of the east-flowing valleys has left its mark in several ways. Not only do the valleys display glacial signatures in cross valley and longitudinal profiles, but the glaciers likely produced sediment at rates high enough to impact the fluvial system downstream significantly. We note several possible feedbacks. For example, headwall retreat into the high surfaces of the summit spine may result in the diminution of their own snow/avalanche supply, which in turn might stall headward erosion and allow relict high surfaces to be preserved. As headwalls steepen, supraglacial processes become important in the evolution of the summit spine.

**FLUVIAL CANYONS**

The bedrock fluvial canyons of the Front Range are sandwiched between glacially influenced headwaters and the alluvial reaches of the western portion of the Great Plains. Their profiles reflect this position in the landscape. Steps and flats in the headwaters represent the strong signature of glacial occupation, as discussed above. Below the glacial limit, the profiles are smooth, but commonly display a convexity within the crystalline mountain front that interrupts the profile (Fig. 10). All of these profiles smoothly grade to the South Platte River, which acts as the lower boundary for these streams. In modeling the evolution of these river profiles, we must therefore acknowledge the control exerted by both the headwaters and the main trunk stream.

In this section, we will only attempt to model the portions of the fluvial system in which bedrock erosion dominates. More specifically, we will not count against the model any misfit of the profiles out of this domain. We will specify the lowering rate of the master stream as the lower boundary condition, and refer to ongoing work at the regional scale for understanding of the long-term evolution of these master streams (Riihimaki, 2003). We discuss in a separate section the reality of the headwaters connection, and how we have treated it in the present model.

**Bedrock Incision**

Stream incision is accomplished by a combination of processes that include quarrying of rock blocks from the bed, abrasion of bedrock, and minor dissolution in susceptible rock types (e.g., Hancock et al., 1998; Whipple et al., 2000). The efficiency of these processes varies strongly with lithology, the pace of incision dictated by the most resistant rock in a reach. In some cases, the switch from dominance of one to another process is frequent, as both the character of the rock or the flow characteristics that are relevant to incision are very local. We acknowledge that there are many formulations of the problem, some casting it as a function of local shear stress, others as a function of local stream power (see reviews in Whipple et al., 2000; Whipple and Tucker, 1999; Tucker and Whipple, 2002). Most researchers will also acknowledge that the incision process has a threshold, either or both because the sediment mantling the bed must be entrained to make way for erosion of the bedrock beneath it (Hancock and Anderson, 2002) or because this sediment is itself the tool of abrasion (e.g., Lavé and Avouac, 2001; Sklar and Dietrich, 2001). This assumption has led researchers to incorporate a probability distribution of discharges into landscape evolution models, as only some fraction of discharge events will incite erosion (Snyder et al., 2003; Tucker and Bras, 2000; Baldwin et al., 2003; Molnar, 2001).

We choose a middle ground for now, in which we appeal to the stream-power formulation of river incision, but incorporate an incision threshold. Importantly, we also explicitly address the drainage topology in that the spatial pattern of water discharge reflects the pattern of drainage area as a function of position in the basin. We also include a first-order approximation of the orographic effects that lead to a nonuniform and evolving pattern of precipitation within the landscape. We do not smooth the drainage area into a power-law distribution, but instead represent the interesting discrete jumps of area at tributary junctions.

Specific stream power (defined as power per unit area of bed) may be written:

\[ \omega = \rho_w g QS/W, \]  

where \( \rho_w \) is the density of water, \( g \) is the acceleration due to gravity, \( Q \) is the water discharge, \( S \) is the local slope of the channel \( (dz/dx) \), and \( W \) is the channel width. Vertical incision is then cast as a function of excess stream power over a threshold stream power, \( \omega_c \):

\[ E = \frac{dz}{dt} = k(\omega - \omega_c) = k[\rho_w g QS/W - \omega_c]. \]  

In this formulation, the constant \( k \) reflects primarily the erodibility of the channel. The threshold reflects the need to
transport sediment of the caliber that exists on the local bed in order to accomplish the scouring of the local sediment to reach the bedrock in any particular year. In addition, there is likely a threshold involved in both the abrasional and quarrying processes themselves.

**Linkage to the Glacial Headwaters**

Glaciers impact the downstream fluvial system through their prodigious production of sediment. As discussed at length in Hancock and Anderson (2002), the coarse end member of this sediment load can result in major aggradation of the river downstream, which can reduce or eliminate access of the river to its bedrock bed for large fractions of the glacial cycle. We have already depicted a reasonable time series of sediment delivery from the glacial headwaters. The formal coupling should therefore involve modeling of bed aggradation and degradation through formulas for conservation of sediment and for bedload sediment transport (e.g., Hancock and Anderson, 2002). The upper boundary condition for this transport includes the location of input of sediment (the glacial terminus), and the strength of the source of sediment. This leads to prolonged periods of aggraded conditions that limit vertical incision, but that allow lateral planation. It is in these periods that the river can produce broad strath surfaces in reaches of weak bedrock (e.g., the Wind River as it passes through the Wind River Basin [Chadwick et al., 1997; Hancock and Anderson, 2002; Hancock et al., 1999]). While Hancock and Anderson (2002) showed that wide straths of the Wind River system can be explained by this mechanism, fluvial systems in harder rock will not easily carve laterally. In these cases, the signature of the glacial sediment input variation will likely be fill terraces rather than strath terraces.

Alternatively, one can address this issue by modulation of an armoring function (Sklar and Dietrich, 2001; see also Riihimaki, 2003) through the glacial-interglacial cycle. In simplest terms, this reduces to the inclusion of a factor, which varies from 0 to 1, by which the calculated instantaneous bedrock incision rate is multiplied.

**Orographic Response**

The precipitation pattern dictates the spatial pattern of stream discharge. Following the development of Roe et al. (2002, 2003), the stream discharge is taken to be

\[ Q = \int_0^S P(x) \frac{dA(x')}{dx'} dx' , \]  

where \( P(x) \) is the runoff (precipitation-evaporation) pattern, and \( A(x) \) is the spatial distribution of drainage area, as derived from a DEM of the drainage basin. The far-field atmospheric speed, \( V \), the local slope of the topography presented to the atmosphere, and the temperature of the atmosphere dictate the precipitation pattern. We follow the work of Roe et al. (2002, 2003) who defined the convergence of air-column moisture flux, \( C \), as

\[ -\nabla \cdot F = C = \left( \alpha_e + \alpha_i \left( V \frac{dz}{dx} \right) \right) e_w(T) , \]  

where \( F \) is moisture flux, \( e_w \) is the saturation vapor pressure at the surface, which is a strong function of temperature, \( T \), and \( \alpha_e \) and \( \alpha_i \) are coefficients. The term in square brackets represents the rate of vertical lift of the air mass. Roe et al. (2002) acknowledged a spatial lag between the location of hydrometeor formation in the air mass at height \( H \) above the ground and the location of precipitation on the ground by incorporating a spatial lag that is distributed as a Gaussian. The resulting precipitation pattern is

\[ P = \frac{L}{2} \int_S^0 C e^{-(x-x')^2/2} dx' . \]  

The length scale, \( L \), may be scaled using the fall speed of the precipitation, \( w_{00} \), the height of formation of droplets, and the airspeed, \( V \), giving \( L = \Gamma(H w_{00}) \). The surface temperature is taken to be \( T = T(z_{basin}) - \Gamma(z_0) \), where \( \Gamma \) is the lapse rate, 6.5 °C/km, and \( T(z_{basin}) \) is the temperature in the basin adjacent to the range.

Given the axial symmetry of the range, and the relatively narrow valleys incised into it, we assume that gradients in the mean elevation of the range profile, and not those in the stream profile itself, control the pattern of atmospheric flow relevant to generating the precipitation pattern. The important storms that deliver precipitation to the present range are upslope storms, meaning that they are not driven by the mean westerly flow, but back up into the range from the Great Plains to the east. We posit a long-term pattern of precipitation in the Front Range utilizing the records of the Mountain Research Station, which provides a range-normal transect of precipitation.

**Initial Condition for the Fluvial Profile Model**

Any numerical model of river profile evolution must start with a specified profile: the initial condition. We choose to initiate the model using a steady profile (\( dz/dt = 0 \) everywhere) in which the incision is perfectly balanced by a uniform but very slow rock uplift rate, \( U_c \). This uplift rate is meant to signify the regional isostatic response to slow uniform erosion, the rate of which was set by the weathering rate in the landscape, on the order of a few microns per year. Following the work of Whipple and Tucker (1999), we solve the erosion equation for river incision under these circumstances to yield the local slope of the stream. This is then integrated to yield the initial profile. The slope of the channel, \( S \), is

\[ S = \frac{W (U_c + \omega)}{\rho k Q} . \]  

Note that we must know the pattern of river discharge, \( Q(x) \), in order to perform this calculation, meaning that we must also assume an initial precipitation pattern. Here we appeal to a pattern of precipitation, \( P(x) \), that is reasonable given the initial topography and the orographic effects described above. The river
discharge is then calculated using equation 9, incorporating the real distribution of drainage area, \( A(x) \). The initial profile, \( z(x) \), is then obtained by integration:

\[
z = z_o - \int_0^x S(x) \, dx ,
\]

where \( z_o \) is the elevation at the headwaters of the system. (Equivalently, one could integrate this profile from the junction with a master stream, \( x = x_{jct} \), upward:

\[
z = z_{jct} - \int_{x_{jct}}^x S(x) \, dx
\]

where \( z_{jct} \) is the elevation at the junction with the master stream, \( x_{jct} \). Given an initial guess at the river profile and using an initial guess of uniform precipitation, \( P_o \), one can assess the slopes, \( S(x) \) and then integrate. This profile can then be used to correct for orographic effects on precipitation to generate a new \( P(x) \); this alters \( Q(x) \), and hence \( S(x) \) and \( z(x) \), and the process can be repeated until the profile converges on one that accommodates both a more realistic precipitation pattern and the requirements of uniform, steady rock uplift.

**Fluvial Model Results**

In Figure 13, we present a simulation of the evolution of Middle Boulder Creek over 3 m.y. A lowering rate of 0.15 mm/yr was imposed at the outer edge of the calculation, mimicking incision of the South Platte River (Dethier, 2001). A single hardness contrast of 4-fold was used at the mountain front. Isostatic rebound in response to the unloading of both the soft bedrock of the basin and the crystalline bedrock valleys within the range was employed, using mantle and bedrock densities of 3300 kg/m³ and 2700 kg/m³, respectively. We assumed that cross valley profiles of rock removal were very broad in the basin fill and had 20° bounding slopes in the crystalline core, a value derived from averaging slopes on the DEM near the present mountain front. The response to lowering of the boundary was rapid; a subtle convexity in the profile within the rock of the basin migrated 60 km to arrive at the mountain front within 1 m.y. Thereafter, the migration of the convexity in the profile was more prominent, and slowed, progressing 20 km in the remaining 2 m.y. of the simulation. The position and amplitude of the convexity at the end of the model run roughly coincided with that in the modern Middle Boulder Creek profile. The base of the convexity is now at Boulder Falls, essentially at the junction of the North and Middle forks of Boulder Creek.

Reported late Pleistocene and Holocene rates of incision of Boulder Creek (Schildgen et al., 2002) are much higher than those in the model run at these times. This likely reflects the reliance on dates from fill deposits to constrain river incision. As our target is the incision of bedrock channels, the appropriate datum is the bedrock beneath the fill. In addition, incision rates based

![Figure 13. Middle Boulder Creek 3 m.y. model results. Initial condition, shown in green and calculated from assumption of steady very low uplift rate (see text), is subjected to steady lowering of base level on the left side of the calculation space at a rate of 0.15 mm/yr. Lithology contrast at mountain front is assumed to be vertical, and is reflected in persistent break in slope in the profile. Migration of response upriver is shown in 20 profiles evenly spaced at 150 k.y. intervals. Final profile is shown in red, while modern river profile is shown in blue. The break in slope migrates rapidly across Cretaceous bedrock of the basin floor, and slows at the crystalline mountain front, to be found at 3 m.y. roughly 15 km into the front. Isostatic rebound is driven not only by local incision, but by exhumation of the much wider basin at the mountain front. The calculation is expected to fail above the glacial limit, as glacial erosion is not captured in the fluvial model; pC—Precambrian.](image-url)
uppon the time interval since the last glacial can be misleadingly fast (Hancock and Anderson, 2002), as the access of the river to channel bedrock in this interval is likely greater than the mean over a glacial-interglacial cycle (see also Wegmann and Pazza-glia, 2002).

The evolution of the abrupt mountain front enhances orographic precipitation, and hence alters the expected pattern of river discharge through time (Fig. 14). The enhancement of precipitation occurs near the range front, as the mean slope of the topography changes most dramatically there, increasing from a minor step to one on the order of many hundred meters.

The profile of modeled local relief up the creek (Fig. 14) shows the strong increase in relief at the mountain front, and decay of relief into the range. Above the convexity, where little erosion has taken place, the relief remains subdued, reflecting the rolling nature of the subsummit surface. This captures the essence of the DEM-derived cross valley profiles depicted in Figure 15. We note that in reality, the mountain front has become more crenulated near the exit of Middle Boulder Creek, reflecting the launching of incision waves up the smaller tributaries in response to lowering of the trunk stream.

**Stream Evolution and the Preservation of the Rocky Flats Surface**

We have begun to explore the variability of the response of the various streams draining the Front Range to the lowering of the trunk stream, which in this area is the South Platte.

[Figure 14. Orographic feedback in Middle Boulder Creek simulation. (A) Drainage area distribution in Middle Boulder Creek, assumed not to change through the simulation. (B) Width distribution of Boulder Creek channel, assumed to go as $x^{0.5}$. (C) Evolution of the pattern of precipitation, calculated according to the orographic precipitation rules discussed in the text. As the topographic step at the mountain front becomes more prominent, the precipitation generated by it increases. The amplitude of the final precipitation pattern is consistent with modern records. (D) Evolution of mean annual river discharge as the precipitation pattern changes. Enhancement in the lower reaches is more than 10%.

Figure 15. Profiles of hillslopes adjacent to Boulder Creek, aligned such that Boulder Creek is centered at 0. Uppermost profile 1 shows obvious U-shaped cross section. Profiles 2 and 3 display roughness typical of the subsummit surface, with tributaries of Boulder Creek only slightly incised. Profile 4 is centered over the present position of the knickpoint, at Boulder Falls, and shows significant incision into the surface, and steep bounding slopes. Profiles 5 and 6 show very steep canyon walls maintained well after passage of the knickpoint, and more complexity associated with the response of smaller tributaries to the incision of Middle Boulder Creek.]
While we have illustrated the transient nature of the response of Middle Boulder Creek to the lowering of the South Platte River, other nearby streams appear to have been much less effective in their response. We focus on two of these, Coal Creek and Ralston Creek. These streams cross and bound, respectively, the prominent pediment named the Rocky Flats surface (e.g., Scott, 1960; Fig. 16). Importantly, neither of these streams reaches the Continental Divide, and hence they fail to tap glacial headwaters. Their drainage areas at the exit from the mountain front (Ralston Creek, 55 km²; Coal Creek, 44 km²) are much smaller than Boulder Creek (300 km²) to the north, or Clear Creek (1800 km²) to the south. Their headwaters are limited by the anomalous topography of the 3-km-tall Thorodin Mountain. This interrupts the otherwise low-relief Rocky Mountain, or “subsummit,” surface roughly halfway between the range crest and the range front. Coal Creek rises on the east side of this massif, while Ralston Creek snakes around it to the south and taps a portion of its western side. Finally, Coal Creek crosses a very prominent band of strong Precambrian quartzite (Coal Creek quartzite), clasts of which form 90% of the clasts on the Rocky Flats surface (Shroba and Carrara, 1996).

The profiles of Coal Creek and Ralston Creek merge smoothly with the profiles of the adjoining pediments. Rather than becoming more steep outbound of the mountain front, as does Middle Boulder Creek, for example, Coal Creek steepens. Coal Creek is still coupled to the Rocky Flats surface at the mountain front, and becomes progressively more incised with distance into the basin. One should therefore expect that the age of this surface, meaning the time of its abandonment by the stream contributing sediments to it, will depend upon distance from the mountain front, i.e., it is expected to be diachronous (Riihimaki, 2003). Ralston Creek, with greater drainage area, has incised more deeply into both the mountain front and the basin. In the process of this incision, it has left two terraces lower than the Rocky Flats surface (Verdos and Slocum surfaces of Scott, 1960). A slight convexity occurs a few kilometers into the crystalline rock of the range, indicating that the effect of base-level lowering has propagated into the range front. The net result is that Ralston and South Boulder Creeks, at the southern and northern edges of the Rocky Flats surface, respectively, are deeply incised into the surface, while Coal Creek is still un-incised at the exit from the mountain front. In essence, then, the preservation of the Rocky Flats surface has been allowed by the presence of the Thorodin Mountain massif, which most likely represents the decayed remnant of a topographic high within the post-Laramide landscape, and which in turn limits Coal Creek to a small drainage area. The remaining Rocky Flats surface resides in something like an erosional shadow downstream of the topographic high.

Three factors dictate the timing and intensity of the response of a stream to exhumation in the Laramide basins: (1) the position of the stream within the fluvial network, (2) the drainage area of the stream, and (3) the geologic and geomorphic nature of the headwaters (Riihimaki, 2003). We scale this timing as follows. The stream-power formulation of the fluvial incision problem is a kinematic wave equation, in which the celerity of the wave may be thought of as the rate of upstream propagation of a knickzone (or convexity) (e.g., Anderson, 1994; Whipple and Tucker, 1999; Crosby and Whipple, 2002). Ignoring the erosion threshold, the rate is proportional to both the discharge of water, \( Q \), and the erodibility of the rock, \( k \), and is inversely proportional to the channel width:

\[
c = k \left[ \frac{\rho g Q W}{k} \right]. \tag{15}\]

In general, both \( k \) and \( Q \) vary upstream, so that only in uniform rock and in allochthonous rivers (\( Q \) uniform) is the wave celerity a constant. The time, \( T \), of arrival of a knickpoint at a position \( x \) within the system is then

\[
T(x) = \int_{x_o}^{x} \frac{1}{c} \, dx. \tag{16}\]

Here \( x_o \) is taken to be the junction with the trunk stream casting off the knickpoint. Alternatively, and more relevant to the Front Range natural experiment, the present position of the knickpoint (convexity) in each drainage will be set by

\[
L = \int_{0}^{T} \frac{c \, dt}{}, \tag{17}\]

where \( T \) is the time since the lowering began on the master stream (South Platte) at the base of the system. Because the celerity is related to the erodibility of the rock, the knickpoints should migrate rapidly across the portion of the network dominated by soft Cretaceous sediments, and slow markedly as they enter the mountain front (see Zapradowski et al. [2001] for an example of knickpoint migration in channels beside the Black Hills). In addition, as the knickpoint migrates higher into the network, loss of drainage area will dictate loss of discharge, and hence slow the progress. This will be especially true at tributary junctions, where discharge declines in discrete and sometimes large jumps.

**CONCLUSIONS**

The Front Range is ideal for separating and understanding the development and functioning of each of its geomorphic components. The range is ornamented at its crest by glacial erosion and along its front by fluvial incision. That the glaciers of the Front Range only reached halfway to the range front allows us to discriminate among the signals of each portion of the system. In addition, at the scale of an individual range, there has been little tectonic activity for the last 40 m.y., providing us the opportunity to investigate the operation of geomorphic processes in the absence of tectonics and associated feedbacks.

The various geomorphic components of the alpine landscape presented by the Front Range can be accounted for in several straightforward 1D models. The high surfaces and their counterpart at mid-elevations, the subsummit surface or Rocky Mountain surface, are dominated by slow weathering and transport rates that lower these portions of the landscape at
Figure 16. Digital elevation model (DEM) and profiles of the Rocky Flats surface and associated streams (see inset with boxed area for larger fluvial context). (A) DEM of entire drainage area of Coal Creek as it rises on the east side of Thorodin Mountain. Ralston Creek rises to the west of Thorodin Mountain, but does not access the glaciated terrain of the Continental Divide. (B) Coal Creek and Ralston Creek profiles, along with profiles of Rocky Flats, Verdos, and Slocum surfaces. (C) Drainage basin area profiles for each creek. Names of creeks that add significantly to the drainage basin areas are shown vertically. Profiles extend beyond the junction of Coal Creek with the South Platte. Coal Creek is only slightly etched into the Rocky Flats surface at the mountain front. Ralston Creek exits the crystalline range at an elevation 150 m below that at which Coal Creek exits.
rates on the order of 5–10 μm/yr. The glacial troughs adjacent to these within ~10 km of the crest are decoupled from the high surfaces. They have deepened significantly, perhaps by large fractions of a kilometer in places, over the late Cenozoic epoch, producing significant local relief along the spine of the range. In addition, significant headward retreat of glacial headwalls has truncated the high surfaces. Headwall retreat, driven largely by westward growth of glacial valleys east of the divide, results in a highly asymmetric range crest, with intact long, smooth slopes from the west, meeting glacial headwalls at or near the crest of the surfaces. This glacial deepening and extension serves to ornament the crest, and has transformed what was likely a boring alpine landscape in mid-Cenozoic time into one that merits National Park status.

The fluvial system in the Front Range is sandwiched between these glacial headwaters and an incising trunk stream, the South Platte River. The fluvial channel is supplied with both meltwater and sediment from the glacial system, the location of the transition being the glacial terminus. Because both the terminus position and the maximum ice discharge (and along with it, the maximum erosion rate) will co-vary in a glacial cycle, sediment delivery (both fine and coarse) will be highly pulsed over time scales of 10^3 to 10^6 yr. The montane fluvial system is currently caught in the midst of a transient response to base-level lowering on the South Platte River, and its ability to incise is likely highly punctuated by pulses of aggradation associated with glacial sediment delivery. Each stream draining the mountain front is in a different state within this transient response. The large streams draining the glaciated crest (e.g., Boulder Creeks, Clear Creek) display strong convexities well up into the crystalline core of the range, while the knickzones of some smaller streams have yet to reach the mountain front (e.g., Coal Creek). This differential response is dictated by both the location of the stream in the South Platte drainage network, which sets the timing of the initiation of base-level lowering, the distance to the mountain front from this confluence, and the water discharge of the tributary stream. This latter is set largely by drainage area, which is strongly affected by anomalies in the topography (e.g., Thorodin Mountain) within the ancient post-Laramide landscape. Consequently, the streams on the edge of the Denver Basin are presently at different levels relative to the prominent pediments, and will have experienced different incision histories. This raises a cautionary flag for correlation of surfaces within the basin (Riihimaki, 2003). The position of the stream within the fluvial network, and the drainage characteristics of each stream must be taken into account when attempting to deduce a basinwide history of incision from sparsely dated surfaces.

Although we have shown that the glacial troughs and high surfaces are effectively decoupled from one another, the same cannot be said of other components in the system. Future work should address (model) the formal linkage between the glacial and fluvial portions of the landscape, and between channels and hillslopes. Important and intriguing questions about the evolution of the Front Range remain. The timing of the initiation of exhumation of the Denver Basin (or any other basin in the Laramide province) remains poorly constrained. Identification and dating of scraps of fluvial deposits on the subsummit surface may aid in this. We have not addressed the cause of the incision of the South Platte River, which has produced the clear response within its tributaries on which we have focused. Debate still revolves around the roles of climate change and of tectonics in inciting this broad exhumation of the Laramide basins. While local tectonic events may be invoked in the NW of the Laramide province (arrival of the Yellowstone hotspot [e.g., Smith and Braile, 1993]), or in the south (northward propagation of the Rio Grande Rift [Chapin and Cather, 1994; Ersliev, 2001; Formento-Trigilio and Pazzaglia, 1998; Leonard, 2002; MacMillan et al., 2002]), no single regional geophysical event has been identified that would incite regional exhumation in the late Cenozoic. The primary long-wavelength geophysical mechanism to which one might appeal is the dynamic topographic response to the movement of the Farallon slab beneath this region (e.g., Mitrovica et al., 1989; Heller et al., 2003). On the other hand, significant changes in late Cenozoic climate (e.g., Zachos et al., 2001) may have driven changes in the probability distribution of storm sizes and hence in stream discharges to which the fluvial system will have been subjected (Zhang et al., 2001). In either case, the temporal and spatial pattern of exhumation appears to demand that the fluvial signal is propagating up the system. While this alone does not discriminate between tectonic and climatic (discharge of water and sediment) forcing mechanisms, we presently favor a climatic driver (Riihimaki, 2003). The primary data needed to solve this riddle will come from more dating of depositional and erosional surfaces within the basins and down the trunk streams.

ACKNOWLEDGMENTS

This research was supported by two grants from the National Science Foundation (NSF) (EAR-0003604 and OPP-9818251), an NSF graduate fellowship (to Riihimaki), a National Aeronautics and Space Administration (NASA) graduate fellowship (to MacGregor), and a University of California Presidential postdoctoral fellowship award (to Safran). The manuscript was improved significantly by the careful reviews of Frank Pazzaglia and Eric Kirby. Finally, we thank the organizers of the Penrose conference for a spectacular and informative meeting.

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Printed in the USA