An evaluation of stream characteristics in glacial versus fluvial process domains in the Colorado Front Range

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A B S T R A C T

Many of the conceptual models developed for river networks emphasize progressive downstream trends in morphology and processes. Such models can fall short in describing the longitudinal variability associated with low-order streams. A more thorough understanding of the influence of local variability of process and form in low-order stream channels is required to remotely and accurately predict channel geometry characteristics for management purposes, and in this context designating process domains is useful. We define process domains with respect to glacial versus fluvial valleys and lateral confinement of valley segments. We evaluated local variability of process domains in the Colorado Front Range by systematically following streams, categorizing them into stream morphologic type and process domain, and evaluating a number of channel geometry characteristics. We evaluated 111 stream reaches for significant differences in channel geometry among stream types and process domains, location and clustering of stream types on a slope–drainage area (S–A) plot and downstream hydraulic geometry relationships. Although individual channel geometry variables differed significantly between individual stream types in glacial and fluvial process domains, no single channel geometry variable consistently differentiated all stream types between process domains. Hypothetical S–A boundaries between bedrock- and alluvial-bed channels proposed in previous studies did not reliably divide bedrock and alluvial reaches for our study sites. Although downstream hydraulic geometry relationships are well-defined using all reaches in the study area, reaches in glacial valleys display much more variability in channel geometry characteristics than reaches in fluvial valleys, less pronounced downstream hydraulic geometry relationships, and greater scatter of reaches on an S–A plot. Local spatial variability associated with process domains at the reach scale (10^1–10^3 m) overrides progressive downstream relationships in low-order mountain streams of the Colorado Front Range.

1. Introduction

Conceptual models developed for river networks that emphasize progressive downstream trends in channel morphology and processes include the river continuum concept (Vannote et al., 1980), downstream hydraulic geometry (Leopold and Maddock, 1953), and slope–drainage area (S–A) relationships (Hack, 1957; Sklar and Dietrich, 1998). Although these models are useful for large or lowland rivers, they may not be as applicable for low-order (first- to third-order) streams in mountainous areas because of abrupt downstream changes in geomorphic history (e.g., glaciations: Arp et al., 2007; Brardinoni and Hassan, 2007; and history of landslide-producing variations in valley width: May et al., 2013), geology (e.g., Adams and Spotila, 2005; Wohl, 2005; Thompson et al., 2008), and climate (Wohl, 2010b) within mountainous terrains. These abrupt downstream changes create segmented longitudinal profiles and spatial variability in valley and channel geometry and disturbance regimes over short (10^1–10^3 m) distances in mountain streams, which can have limited ability to readily adjust channel morphology to spatial variation in substrate resistance and sediment supply (Wohl, 2010b). The absence or weak development of progressive downstream trends in mountain river networks indicates the need to focus on reach-scale patterns. We define a reach as a length of channel at least several times the average channel width that has consistent gradient and channel geometry, typically 10^1–10^2 m along low-order mountain streams.

Mountain streams, which are commonly the headwaters for larger river systems, have been less extensively studied than their low-gradient counterparts. Low-order streams typically compose over two-thirds of total stream length of a drainage basin (Freeman et al., 2007), and their abundance and influence on the river network as a whole can be underestimated and inadequately acknowledged from a management perspective (Gomi et al., 2002). The location and spatial abundance of mountain streams make them important sources of sediment, water, nutrients, and organic matter for downstream portions of the river network (Milliman and Syvitski, 1992; Benda et al., 2005). In addition, the small drainage areas and variation in roughness elements associated with mountain streams lead to at least
temporary storage of organic matter, which in turn provides essential food sources and habitats for the base of the food chain (Gomi et al., 2002). Mountain streams include wide ranges of gradient, light, temperature, water chemistry, substrate, food sources, and species composition, which combine to form a wide variety of habitats (Meyer et al., 2007). All of these features of low-order mountain streams collectively indicate disproportionately high physical and ecological significance of mountain streams in the context of an entire watershed.

Mountain streams are different from their lowland counterparts in their typically steeper gradients and in the influences of local and the in fluvial and hyporheic hydromorphology, elevation-related spatial variation in hydrologic regime (snowmelt-dominated versus precipitation-dominated hydrograph), and the strong influence exerted directly on rivers by hillslope sediment dynamics, disturbance regimes, and differences in rock resistance. Lower gradient reaches of mountain streams tend to be transport limited with respect to fine sediments and are response reaches in which changes in sediment supply are likely to cause changes in channel morphology (Montgomery and Buffington, 1997). In contrast, high-gradient reaches of mountain streams tend to have high transport capacity relative to sediment supply because of their steeper bed gradients, so that they are supply limited with respect to pebble-sized and finer sediments (Montgomery and Buffington, 1997). Sediment dynamics in mountainous headwaters are directly related to the diverse morphology in mountain streams. The persistence of a specific stream morphology is maintained by roughness and energy dissipation influenced by sediment dynamics and by larger clasts that are only moved during extreme events (Montgomery and Buffington, 1997; Flores et al., 2006; Thompson et al., 2008).

The widely used channel classification system developed by Montgomery and Buffington (1997) for mountain streams focuses on reach-scale channel geometry. Channel geometry is categorized in terms of dominant bedform (cascade, step-pool, plane-bed, pool–riffle, dune–ripple), and we refer to these categories as stream types. This classification is widely used in part because much resource management focuses at the reach-scale (Wohl et al., 2007), where stream type can be used to predict aquatic communities and habitats or areas of hyporheic exchange (e.g., Montgomery et al., 1999; Buffington et al., 2004; Buffington and Tonina, 2009; Bellmore and Baxter, 2013). Stream type correlates with reach-scale gradient (Montgomery and Buffington, 1997; Wohl and Merritt, 2005, 2008; Wohl et al., 2007) and with gradient combined with an index of specific stream power based on drainage area (Flores et al., 2006). This is particularly useful in a management context because reach-scale gradient can be readily extracted and mapped from remote data such as digital elevation models (DEMs), which facilitate mapping the spatial distribution and abundance of stream type (e.g., Buffington et al., 2004). Individual categories of stream type differ in their response to changes in water and sediment yield, as well as types and abundance of aquatic and riparian habitats (Wohl et al., 2007). Sklar and Dietrich (1998) proposed that consistent correlations exist between stream substrate type and channel slope–drainage area (S–A), such that reaches of similar substrate will plot distinctly in S–A space.

Individual stream types can occur within diverse process domains. A process domain is a spatially discrete area that is characterized by a distinct geomorphic history and assemblage of geomorphic processes, which together create distinctive landforms and disturbance regimes (size, frequency, duration of floods, and debris flows) (Montgomery, 1999). Process domains can be used to understand and predict sediment input, transport, and storage in streams, as well as ecological structure along and within stream segments (Wohl and Merritt, 2005; Wohl, 2010a; Polvi et al., 2011; May and Lisle, 2012). Disturbance regimes can physically modify existing or progressive downstream trends by influencing sediment and water dynamics, which in turn dictate channel morphology and stream type.

Process domains identified for the Colorado Front Range emphasize lateral valley-bottom confinement and geomorphic history (Polvi et al., 2011; Wohl et al., 2012). Lateral valley-bottom confinement is differentiated into confined, partly confined, and unconfined valleys based on the ratio of active channel width to valley-bottom width. Geomorphic history is differentiated into glacially formed valleys above ~2430 m elevation and fluvially formed valleys at lower elevations. Glacially and fluvially formed valleys also have distinct hydroclimatology and disturbance regimes. Glacially formed valleys have snowmelt floods and long (ca. 300–400 y) recurrence intervals for fires (Veblen and Donnegan, 2005) and associated debris flows. Fluvially formed valleys have snowmelt floods and rainfall flash floods resulting from convective storms. Fire recurrence intervals are much shorter (ca. 40–100 y) (Veblen and Donnegan, 2005), and debris flows resulting from fires and from convective storms are more common than at higher elevations. Glaciation widened and deepened valleys, created steep valley walls and headwalls, and flattened the lower portions of glacial valleys (Anderson et al., 2006; Amerson et al., 2008). These effects on previously glaciated valleys have decoupled hillslope processes from inner stream valleys so that stream channels flow through valleys that are not necessarily adjusted to current fluvial sediment, water, and disturbance regimes but to inherited glacial terrain and characteristics (Brardinoni and Hassan, 2007; Arp et al., 2007; Klaar et al., 2009). In contrast, streams in unglaciated valleys likely have maintained their coupling with hillslopes and have created and maintained their own channels according to historical and current sediment, water, and disturbance regimes.

Both glacial and fluvial valleys in the Colorado Front Range include all three levels of valley confinement and also contain diverse stream types. As a result of the differences in hillslope–channel coupling and disturbance regime between glacial and fluvial valleys, we hypothesize that individual stream types have statistically different geometry between glacial and fluvial process domains. In other words, even though step–pool channel segments are present in glacial and fluvial valleys, and are clearly different than cascade or pool–riffle channel segments, step–pool channels in glacial valleys will be a statistically distinct population from step–pool channels in fluvial valleys. To test this hypothesis, we systematically examine whether consistent differences in channel geometry and gradient within a stream type exist between glacial and fluvial process domains in the Colorado Front Range.

Because we expect geomorphic history to strongly influence stream geometry, we also hypothesize that (i) the S–A relations proposed by Sklar and Dietrich (1998) will not adequately describe observed patterns of channel substrate type in the Front Range, and (ii) downstream hydraulic geometry relations will differ significantly between glacial and fluvial process domains. Each of the three hypotheses reflects our expectations that local geomorphic history will influence contemporary channel geometry strongly enough to create significant differences between process domains.

Much of the previous research regarding trends in channel geometry of mountain streams either (i) comes from climate and tectonic regimes that differ from the semiarid, tectonically stable Colorado Front Range, such as the U.S. Pacific Northwest region (Montgomery and Buffington, 1997; Montgomery, 1999; Buffington et al., 2004; Brardinoni and Hassan, 2007; Buffington and Tonina, 2009) or southeastern Australia (Thompson et al., 2008), (ii) was not designed to include the entire range of channel types we examine here (e.g., Wohl and Merritt, 2005, 2008), or (iii) does not explicitly evaluate how correlations between channel type and potential control variables differ between process domains.

2. Study area

The mountainous portions of the four catchments (North St. Vrain Creek, Glacier Creek, Big Thompson River, Cache la Poudre River) surveyed in this study begin at the eastern side of the continental divide in Rocky Mountain National Park, Colorado (Fig. 1). Streams within each
network flow eastward through the park into Roosevelt National Forest and beyond the Colorado Front Range to the South Platte River (Supplementary data). Most study sites for this project were located from near the headwaters (~3100 m) to ~1945 m in elevation, where North St. Vrain Creek enters Ralph Price Reservoir. Field sites were chosen because of the lack of flow diversion or regulation structures, land development, or recent land use impacts so that reaches have been largely unaltered by human activity. In addition, no sites within the study area had undergone major disturbances such as fire, blowdowns, and major floods in ~30 years or more at the time of study, and we assume that sites had recovered since the last disturbance. Most stream segments were second- to fourth-order streams.

Precambrian Silver Plume granite crystalline rocks, which are composed of granite with some biotite schist and granodiorite, are the dominant core of the study area (Braddock and Cole, 1990). Since the end of the Tertiary, relatively little tectonic activity has occurred in the Front Range (Anderson et al., 2006). The upper portions of the catchments studied were glaciated during the Pleistocene epoch. Pinedale glaciation extended to ~2430 m elevation in the study area, where the terminal moraine is located (Wohl et al., 2004; Polvi et al., 2011). This elevation divides the study area between the glacial and fluvial (nonglacial) process domain types. Below the narrow, summit spine of the continental divide, from ~3000 to 2300 m elevation, lies a low-relief and widespread surface called the subsummit surface; this portion of the Front Range contains deeply incised fluvial bedrock canyons, which become more deeply incised downstream because of continued exhumation of the Denver Basin below the mountain front (Anderson et al., 2006).
Glaciation substantially changed the topography of the upper portion of the study area, widening and deepening valleys and creating steep valley walls and headwalls, in addition to flattening the lower portions of glacial valleys. An assessment of valley width, valley height, and valley cross-sectional area for glacial versus fluvial valley segments in the study area indicates significant differences in valley width and valley cross-sectional area, with glacial valleys wider and larger than fluvial valleys. When the data are plotted as normalized valley width and cross-sectional area (i.e., valley width/drainage area and cross-sectional area/drainage area), even larger differences between glacial and fluvial valley segments emerge, with glacial sites having larger values and a much greater range in values (Fig. 2). Glaciers also created steps in the longitudinal profile (e.g., hanging valleys) at tributary junctions as a result of differences in ice volume between valleys (Anderson et al., 2006). Streams located in the glaciated portions of the study area could thus have inherited valley characteristics to which modern channels continue to adjust (Brardinoni and Hassan, 2007; Arp et al., 2007; Klaar et al., 2009). Conversely, streams located in unglaciated portions of the study area are more likely to have adjusted their channel and valley geometries to historical and current water and sediment regimes that include past glacial outwash and meltwater.

Bedrock jointing patterns in the study area strongly influence valley width and canyon evolution. Differential weathering throughout the Colorado Front Range has created canyons with substantial downstream variability in valley width, regardless of glacial history. Wider valley segments tend to have more closely spaced bedrock joints, lower gradients, and minimal stream–hillside coupling; whereas narrow, bedrock-confined segments tend to have more closely spaced joints, steeper gradients, and maximum stream–hillside coupling (Ehlen and Wohl, 2002).

Mean annual precipitation is 70–80 cm in the upper elevations of the study area, and streams have a snowmelt-dominated hydrograph that peaks during May–June. This portion of the study area has subalpine forest of Engelmann spruce (Picea engelmannii), subalpine fir (Abies lasiocarpa), and lodgepole pine (Pinus contorta) with stand-killing fires that recur ca. 300–400 years (Veblen and Donnegan, 2005). Debris flows are rare. Mean annual precipitation in the lower elevations is ~36 cm. The lower elevations have montane forests of ponderosa pine (Pinus ponderosa), lodgepole pine, and Douglas-fir (Pseudotsuga menziesii) with stand-killing fires at intervals of 40 to 100 years (Veblen and Donnegan, 2005) and more frequent debris flows than the subalpine zone. Below ~2300 m, summer convective storms create the largest peak flows, which recur much less frequently than the annual snowmelt peak flows (Wohl et al., 2004; Wohl, 2011b). Peak unit discharge does not exceed 1.1 m³/s/km² above 2300 m elevation, whereas peak unit discharge approaches 40 m³/s/km² below 2300 m (Jarrett, 1990). The peak flows created from thunderstorms at lower elevations are disturbances that maintain depositional and erosional features of streams located in the fluvial domain (Wohl, 2011b). A gaging station within the glacial portion of the North St. Vrain Creek catchment (1926–2011) has a mean annual peak discharge of 20 m³/s and peak unit discharge of 0.24 m³/s/km² (Wohl and Beckman, 2014). A USGS gaging station (40211405350101) located on the Big Thompson River below Moraine Park (1926–2012) and within the glacial process domain has a mean annual peak discharge of 15 m³/s and peak unit discharge of 0.15 m³/s/km² (USGS National Water Information System, NWIS, 2013).

3. Methods

3.1. Field methods

Surveyed channel segments were primarily located within the North St. Vrain Creek catchment, with additional surveyed reaches on Glacier Creek, the Big Thompson River, and the uppermost portions of the Cache la Poudre River. Site selection began based on accessibility and prior knowledge of streams in the study area; the uppermost accessible portion of a stream represented the upper boundary of the first reach on that stream. Each stream was then followed continuously and divided into reaches until (i) the stream became inaccessible; (ii) the stream entered a confluence or water body; or (iii) continuous reaches were all of the same morphology type and thus moving was necessary to diversify reach types. The end of the continuously followed reaches represented the lower boundary of the last reach on that stream. Reaches were field-selected as a continuous length of channel at least 10 times the bankfull width having consistent and uniform gradient, substrate, and stream type (cascade, step–pool, plane-bed, pool–riffle; Montgomery and Buffington, 1997), which was determined visually. An additional channel type, riffle–run, was added for low-gradient reaches with interspersed riffle and run sections displaying characteristics that did not fit within the designated Montgomery and Buffington (1997) stream types. This morphology type typically had an intermediate gradient and substrate size between step–pool and plane-bed morphology and visually resembled a very low-gradient cascade. Woodforced and debris flow morphology types were also considered in the study and were noted, although reaches were still classified by stream type according to organization of bed substrate. (We found only one wood-forced reach and two reaches in which channel morphology continued to be directly influenced by past debris flows.) For each reach, endpoints were mapped using a handheld GPS device (Garmin eTrex, typically 3–5 m horizontal accuracy). We assigned each reach

**Fig. 2.** Plots of normalized valley width and cross-sectional area (valley width/drainage area and cross-sectional area/drainage area) show significant differences between glacial and fluvial valley segments. The line within each box indicates the median value, box ends are the upper and lower quartile, and whiskers are the 10th and 90th percentiles. Numerical value near each box indicates the mean, with standard deviation in parentheses. A Student t-test was used to test significance and both tests have p-values < 0.05.
to a process domain based on elevation and confinement. We categorized confinement as confined: valley bottom width less than two times the bankfull width; partly confined: valley bottom width approximately two to eight times the bankfull width; or unconfined: valley bottom width greater than eight times the bankfull width. Valley bottom width was determined in the field with a laser rangefinder (TruPulse 350B, horizontal accuracy ± 0.1 m) using indicators such as change in slope, vegetation, and area likely to be inundated in higher flows.

Lengths of individual reaches were variable, having a minimum of 10 times the bankfull width and a maximum based on the length over which morphology and channel geometry characteristics were consistent and uniform. In order to maximize the number of reaches studied (i.e., sample size), channel and valley geometry characteristics were measured only once at one location for each reach. Once the endpoints of a reach were determined, a section of reach that appeared representative of the entire reach was chosen for sampling.

A random-walk pebble count (Wolman, 1954) of bed substrate was conducted for each reach, measuring the intermediate length of 100 clasts; measurements were recorded to the nearest 5 mm. Values of D50 and D90 for each reach, which refer to the grain size diameter where 50 and 84% of grains are finer, respectively, were derived from the grain size distributions. In addition, if the bed substrate of a reach was predominantly bedrock with lesser alluvial coverage, the reach substrate was categorized as bedrock regardless of grain size distribution in alluvial cover (Montgomery et al., 1996).

Field indicators such as changes in bank geometry, sideslope, or vegetation; areas of organic debris collection; and/or stains on rocks derived from 10-m DEMs using a laser range finder (TruPulse 350B, horizontal accuracy ± 0.1 m) was measured with a laser range finder. This was measured over which morphology and channel geometry characteristics were measured only once at one location for each reach.

Field-delineated stream types were plotted by predominant substrate (bedrock, coarse-bed alluvial, bedrock, coarse-bed alluvial, coarse-bed alluvial). To visualize these distributions, we made boxplots by stream type and process domain (glacial, fluvial).

Data were divided into fluvial versus glacial reaches. We developed classification trees, which choose the best subset of control variables and their associated threshold values that partition differences in categorical variables, to classify all stream type by each process domain using the following variables as potential control variables: confinement, S, A, Q, ω, W, D, W/D, D50, and D90. We chose the tree and control variables of best fit using cross validation, lowest residual mean deviance, lowest misclassification error rate, and lowest number of end nodes that produced the lowest partitioning of channel types.

Field-delineated stream types were plotted by predominant substrate (bedrock, coarse-bed alluvial, fine-bed alluvial) and process domain on a diagram using slope and drainage area. We plotted the hypothetical transition between coarse-bed alluvial and bedrock channels as estimated by Sklar and Dietrich (1998) and Addy et al. (2011) and visually assessed sites for their location relative to this line. We also developed regression power relationships for each of the predominant substrate types to compare among coefficients, exponents, and correlations.

4. Results

4.1. Channel geometry differences between process domains

Significant differences are varied between the channel geometry characteristics and between stream types (Table 1), as are the directions of differences between the process domains. Tables 2 and 3 show the

\[ \omega = \omega / \left( p g R D_{50} \right)^{3/2} \]
results of tests of differences in channel geometry and mean values for each channel geometry variable between glacial and fluvial process domains, respectively. For example, Table 2 shows that $D_{50}$ and $D_{84}$ are not typically significantly different for a stream type between glacial and fluvial valleys, with the exception of plane-bed streams, which have a greater mean $D_{50}$ in glacial streams (Table 3). In all stream types except for pool–riffle reaches, mean substrate size is greater in the glacial process domain. Gradient measured in the field is significantly different between glacial and fluvial step–pool streams regardless of statistical method and between glacial and fluvial cascade streams and glacial and fluvial plane-bed streams, depending on the statistical method. Other stream types do not have significant differences in gradient between glacial and fluvial valleys. Significant differences in $D$, $W$, and $W/D$ between glacial and fluvial valleys typically occur with cascade and plane-bed streams, but not for other stream types. Pool–riffle streams display no significant differences between glacial and fluvial valleys for any of the channel geometry characteristics. No stream type displays significant differences for every variable.

With respect to lateral valley confinement, $W$ and $W/D$ are not significantly different across confinement categories (Livers, 2013). Variables $D_{50}$, $D_{84}$, $D$, and $S$ differ significantly between confined streams and others. Partly confined streams do not differ significantly from unconfined streams. These results suggest that the primary effect of lateral valley confinement is to create streams that are steeper and coarser grained than stream reaches in less-confined valley segments.

In summary, the results generally support our hypothesis that individual stream types have statistically different geometry between glacial and fluvial process domains.

### 4.2. Variables that distinguish stream type in process domains

Fig. 3 displays the two classification trees built for glacial sites and fluvial sites. Variables used in construction of the glacial tree are $S$, $D$, $\omega$, and $D_{50}$. Variables used in construction of the fluvial tree are $S$, $A$, and $W$. Misclassification error rates for the glacial and fluvial reaches are 0.24 and 0.18, respectively. Stream types in glacial valleys with lower gradients (pool–riffle and plane-bed) are delineated using $S$, $D$, and $\omega$, whereas stream types with higher gradients (step–pool and cascade) are delineated with only $S$ and $D_{50}$. Riffle–run reaches are more closely related in the tree to the higher gradient stream types. In comparison to streams in glacial valleys, stream types in fluvial valleys are very easily classified, with only five terminal nodes corresponding to the five stream types; riffle–run reaches are more closely related to lower gradient streams in the fluvial classification tree.

### 4.3. Channel distribution on S–A diagram between process domains

The hypothetical bedrock-coarse-bed transition proposed by Sklar and Dietrich (1998) takes the form of

$$S = 0.07A^{-0.5}$$

where $S$ is stream gradient in m/m, and $A$ is drainage area in km$^2$. This line was plotted on an S–A diagram with all of our study reaches plotted by the predominant substrate. Neither of the transition lines proposed by Sklar and Dietrich (1998) or Addy et al. (2011) reliably divided bedrock from coarse-bed reaches in our study sites (Fig. 4A). We visually estimated the boundary between bedrock and alluvial substrates for our data, and then fit an equation to this line. The boundary associated with our data has an intermediate intercept with a gentler slope than that proposed by previous studies. Bedrock (with alluvium) reaches are surrounded by boulder reaches and cobble reaches, meaning that coarse-bed alluvial reaches do not plot separately from bedrock reaches, although the size of the predominant substrate does appear to grade from bedrock to pebble reaches toward the lower left corner on Fig. 4. The one sand-bed reach is anomalous, as it was located in a morphology forced by instream wood. Fig. 4A also reveals that substrate types for glaciated sites are not well discriminated by the S–A curve. A plot with only fluvial sites (Fig. 4B) also does not discriminate effectively by substrate types, but does indicate that the slope of the bedrock–alluvium line is very similar between our study area and the lines from Sklar and

<table>
<thead>
<tr>
<th>Table 1</th>
</tr>
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<tbody>
<tr>
<td>Summary of all field sites subdivided by the number of reaches with respect to stream type, valley history and confinement, and predominant substrate.</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Stream morphology</th>
<th>Valley history</th>
<th>Confinement</th>
<th>Predominant substrate</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cascade</td>
<td>18</td>
<td>14</td>
<td>4</td>
</tr>
<tr>
<td>Step–pool</td>
<td>52</td>
<td>30</td>
<td>22</td>
</tr>
<tr>
<td>Rifle–run</td>
<td>13</td>
<td>7</td>
<td>6</td>
</tr>
<tr>
<td>Plane-bed</td>
<td>9</td>
<td>6</td>
<td>3</td>
</tr>
<tr>
<td>Pool–riffle</td>
<td>19</td>
<td>15</td>
<td>4</td>
</tr>
<tr>
<td>Total</td>
<td>111</td>
<td>72</td>
<td>39</td>
</tr>
</tbody>
</table>

* Denotes a wood-forced morphology.

### Table 2

Summary of $p$-values for tests of significant differences in variables between glacial and fluvial process domains by stream type.

<table>
<thead>
<tr>
<th>Stream morphology</th>
<th>Slope (field)</th>
<th>Slope (DEM)</th>
<th>Width (m)</th>
<th>Depth (m)</th>
<th>W/D</th>
<th>$D_{50}$ (mm)</th>
<th>$D_{84}$ (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cascade</td>
<td>0.08</td>
<td>0.17</td>
<td>0.11</td>
<td>&lt;0.01</td>
<td>0.01</td>
<td>0.22</td>
<td>0.60</td>
</tr>
<tr>
<td>Step–pool</td>
<td>WRS</td>
<td>ST</td>
<td>WRS</td>
<td>ST</td>
<td>WRS</td>
<td>ST</td>
<td>WRS</td>
</tr>
<tr>
<td>Rifle–run</td>
<td>ST</td>
<td>WRS</td>
<td>ST</td>
<td>WRS</td>
<td>ST</td>
<td>WRS</td>
<td>ST</td>
</tr>
<tr>
<td>Plane-bed</td>
<td>WRS</td>
<td>ST</td>
<td>WRS</td>
<td>ST</td>
<td>WRS</td>
<td>ST</td>
<td>WRS</td>
</tr>
<tr>
<td>Pool–riffle</td>
<td>ST</td>
<td>WRS</td>
<td>ST</td>
<td>WRS</td>
<td>ST</td>
<td>WRS</td>
<td>ST</td>
</tr>
</tbody>
</table>

$p$-values in bold indicate statistically significant differences at alpha = 0.10.

* The student’s t (ST) test could only be performed on variables with normal distributions; the nonparametric Wilcoxon Rank Sum (WRS) test was performed on all variables.

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Dietrich (1998) and Addy et al. (2011). The Colorado Front Range line, however, is shifted upward relative to the other lines.

Fluvial bedrock reaches are more clearly located on steeper reaches (above the line) than most other fluvial reaches, with the exception of a few boulder and cobble sites. The pattern of grading to finer substrate occurs as it did using all reaches. For glacial reaches, however, bedrock reaches plot within an area that also contains many boulder and cobble reaches. The pattern of grading to finer substrate is less pronounced for the glacial reaches. The results thus support our hypothesis that the S–A relations proposed by Sklar and Dietrich (1998) do not adequately describe observed patterns of channel substrate type in the Front Range. In addition, the similarity in the slope of the bedrock–alluvial bed boundary line between other fluvial data sets and the fluvial process domain in the Front Range, and the difference in the slope of the boundary line between other fluvial data sets and the entire data set from the Front Range, support our expectations that local geomorphic history influences contemporary channel geometry strongly enough to create significant differences between process domains.

### 4.4. Downstream hydraulic geometry relationships

Table 4 lists the summary of the results of downstream hydraulic geometry power relationships for W vs. Q as fit for the different groups of data. The basic equation that corresponds to this relationship is

\[ W = aQ^b \]  

where \( W \) is bankfull width of the channel in meters, \( Q \) is 2-year peak discharge in \( \text{m}^3/\text{s} \), and \( a \) and \( b \) vary according to the best fit of the data. All the power relationships have good correlation \( (R^2) \) values, meaning there is a well-defined, positive relationship between \( W \) and \( Q \). Although the relationship with all data is very good \( (R^2 = 0.7365) \), a pattern emerges when the reaches are divided into subgroups and fits to a power relationship.

When reaches are divided into glacial sites versus fluvial sites, the corresponding correlation value for fluvial sites is higher than the correlation value for all sites, whereas the corresponding correlation value for glacial sites is lower. When reaches are divided into confinement types, correlation values are equal to or greater than correlation values associated with all data, with unconfinned reaches displaying the greatest correlation between \( W \) and \( Q \). This likely reflects the relative lack of constraints on adjustment of channel width to discharge in unconfinned reaches.

Table 5 lists the summary of the results of downstream hydraulic geometry power relationships for \( D \) vs. \( Q \) as fit for the different groups of data. The basic equation that corresponds to this relationship is

\[ D = cQ^f \]  

where \( D \) is bankfull depth of the channel in meters, \( Q \) is 2-year peak discharge in \( \text{m}^3/\text{s} \), and \( c \) and \( f \) vary according to the best fit of the data. The relationship between \( D \) and \( Q \) is well defined \( (R^2 = 0.572) \), but not as well defined as the relationship between \( W \) and \( Q \). Much like \( W \) vs. \( Q \), \( D \) vs. \( Q \) becomes more correlated in all fluvial sites and less correlated in all glacial sites in comparison to the relationship using all data. When reaches are divided into confinement types, confined reaches display the best relationship between \( D \) and \( Q \), presumably because overbank flows that complicate stage–discharge relations are less likely

![Classification tree for all glacial reaches versus all fluvial reaches. If condition listed at split is true, continue left from split. Note the different variables used to delineate stream types in the different valleys. Misclassification error rates for the glacial and fluvial reaches are 0.24 and 0.18, respectively. 'DA' refers to drainage area in km² and 'Dim.St.Pwr' refers to dimensionless stream power; \( n \) values refer to the number of sites partitioned out in that terminal node.](image-url)
in confined reaches. The results thus support our hypothesis that downstream hydraulic geometry relations differ significantly between glacial and fluvial process domains.

5. Discussion

5.1. Channel and valley geometry differences between process domains

Our analyses, which consisted of determining differences in seven mean channel geometry characteristics between glacial and fluvial process domains and between three different levels of confinement, indicate a lack of consistent differences in channel geometry characteristics among Montgomery–Buffington (1997) stream types across process domains in the Colorado Front Range, in that no single variable differs consistently within a stream type in glacial and fluvial process domains. However, patterns emerge in comparing channel geometry characteristics between process domains. Cascade and plane-bed morphologies display the greatest differences in channel geometry between glacial and fluvial process domains and at least one geometric variable differed significantly between glacial and fluvial process domains for each stream type except pool–rifflfe channels (Tables 2 and 3). Stream types in glacial valleys have typically greater mean substrate sizes and gradients and lower \( W, D \), and \( W/D \) than streams in fluvial valleys. Streams located in confined valleys typically have significantly steeper gradients, coarser substrate, and greater depths than streams located in other levels of valley confinement.

Reaches in glacial terrain have a typically greater range of values for channel geometry characteristics than fluvial reaches (Fig. 5; Supplementary data), greater valley widths and cross-sectional areas than fluvial reaches (Fig. 2), and significantly different mean values of some

Fig. 4. (A) S–A diagram with all reaches. Solid shapes indicate glacial sites; open shapes indicate fluvial sites. The solid gray line represents the Sklar and Dietrich (1998) hypothetical bedrock–coarse-bed transition \( (y = 0.07x^{-0.5}) \); the dashed gray line represents the Addy et al. (2011) bedrock–coarse-bed transition \( (y = 0.15x^{-0.47}) \); the white line with solid gray outline represents the bedrock–coarse-bed transition for this study \( (y = 0.1x^{-0.275}) \). (B) S–A diagram showing only fluvial reaches. The bedrock–coarse-bed transition for fluvial sites only is \( y = 0.2x^{-0.375} \).
channel geometry characteristics of stream morphologies between valley types. These differences equate to glacial reaches having greater diversity in channel geometry characteristics than fluvial terrain, perhaps because streams in the glacial process domain continue to adjust to inherited glacial terrain, whereas streams in the fluvial process domain are better adjusted to current water and sediment regimes.

Gradients associated with specific stream morphologies in this study do not necessarily match the stream gradients predicted by Montgomery and Buffington (1997) for streams in the U.S. Pacific Northwest or the stream gradient range identified for specific stream types in other regions of the world (e.g., Thompson et al., 2006, 2008; Brardinoni and Hassan, 2007) (Fig. 5). Gradient ranges for the Front Range rivers tend to be larger for each stream type than those reported in Montgomery and Buffington (1997), for example, which may reflect the larger data set from the Front Range and the inclusion of glacial and fluvial process domains.

5.2. Variables that distinguish stream type in process domains

Classification of stream reaches in this study into Montgomery and Buffington (1997) stream types was based on visual approximation of predominant bedforms with the assumption that different stream types would be distinguished by specific gradients and channel geometry characteristics. The classification tree analysis (Fig. 3) indicates that no single variable or succinct combination of variables consistently discriminates among all stream types between process domains, but, as would be expected, reach-scale gradient appears to be the most successful channel geometry parameter in delineating stream types. Stream types are distinguished differently according to process domain: fluvial sites do not require measures of substrate or stream power to delineate between stream types, whereas glacial sites do. Streams in glacial valleys also require more channel geometry parameters than streams in fluvial valleys to distinguish among stream types. The higher misclassification rate in delineating glacial sites and multiple end nodes of single morphology types such as plane-bed and cascade are other indicators of the difficulty in classifying stream types in glacial terrain, presumably from the variability in channel geometry characteristics.

5.3. Channel distribution on S–A diagram between process domains

The data collected in this study appear to have a bedrock–coarse-bed transition that has an intermediate intercept and gentler slope in comparison to the relations proposed by Sklar and Dietrich (1998) or Addy et al. (2011). The semiarid Colorado Front Range has less discharge per drainage area, and steeper stream gradients per drainage area, than the empirical data set from Scotland (Addy et al., 2011). We interpret this to indicate that higher gradients are needed in the Front Range to create enough stream power to initiate bedrock incision. Lower discharge per unit drainage area in the Front Range than in many other mountainous regions may also help to explain the preponderance of very steep channel types (cascade and step–pool reaches) within the study area (e.g., Flores et al., 2006).

Although the transitions between substrate types may have different locations at larger drainage areas or steeper gradients on the Front Range and hypothetical S–A diagrams, the pattern of substrate fining that we observe is similar to Sklar and Dietrich’s diagram. Table 1 also illustrates substrate fining as S decreases, as predicted in the Montgomery and Buffington (1997) classification. The boundaries separating streams of diverse substrate types are much less pronounced than the Sklar and Dietrich diagram implies, however, as illustrated by the coarse-bed channel reaches that plot well above the bedrock S–A threshold in Fig. 4. At least some outliers occur in both directions: cobble-bed channels with small S and large A values, and pebble-bed channels with large S and small A values. The numerous outliers relative

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**Table 4**

<table>
<thead>
<tr>
<th>Coefficient (a)</th>
<th>Exponent (b)</th>
<th>R²</th>
<th>Significance</th>
</tr>
</thead>
<tbody>
<tr>
<td>All data</td>
<td>2.38</td>
<td>0.63</td>
<td>0.74</td>
</tr>
<tr>
<td>Glacial</td>
<td>2.15</td>
<td>0.69</td>
<td>0.67</td>
</tr>
<tr>
<td>Fluvial</td>
<td>2.67</td>
<td>0.58</td>
<td>0.87</td>
</tr>
<tr>
<td>Confined</td>
<td>2.62</td>
<td>0.58</td>
<td>0.75</td>
</tr>
<tr>
<td>Part. Confined</td>
<td>0.51</td>
<td>1.06</td>
<td>0.73</td>
</tr>
<tr>
<td>Unconfined</td>
<td>0.81</td>
<td>1.11</td>
<td>0.89</td>
</tr>
<tr>
<td>G Confined</td>
<td>2.40</td>
<td>0.60</td>
<td>0.64</td>
</tr>
<tr>
<td>G Part. Confined</td>
<td>2.44</td>
<td>0.72</td>
<td>0.59</td>
</tr>
<tr>
<td>G Unconfined</td>
<td>1.28</td>
<td>0.80</td>
<td>0.92</td>
</tr>
<tr>
<td>F Confined</td>
<td>3.18</td>
<td>0.53</td>
<td>0.89</td>
</tr>
<tr>
<td>F Part. Confined</td>
<td>2.36</td>
<td>0.59</td>
<td>0.75</td>
</tr>
<tr>
<td>Cascade</td>
<td>2.12</td>
<td>0.78</td>
<td>0.80</td>
</tr>
<tr>
<td>Step–pool</td>
<td>2.17</td>
<td>0.63</td>
<td>0.78</td>
</tr>
<tr>
<td>Riffle–run</td>
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<td>0.65</td>
<td>0.94</td>
</tr>
<tr>
<td>Plane-bed</td>
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<td>0.57</td>
<td>0.75</td>
</tr>
<tr>
<td>Pool–riffle</td>
<td>3.35</td>
<td>0.46</td>
<td>0.41</td>
</tr>
</tbody>
</table>

**Fig. 5.** Comparison of Montgomery and Buffettong (1997) (MB) stream type slope ranges versus slope values found in this study for the Colorado Front Range (CFR). CFR refers to lumped (glacial and fluvial) data. Extremely large slopes (>-20°) come from short stream segments located on glacial terminal moraines or associated with bedrock valley segments.
to predicted patterns, especially with regard to glaciated reaches, suggest that local controls beyond S and A strongly influence stream substrate in the Front Range. An example of such a local control could be coarse sediment relict from glacial processes at higher elevations or coarse sediment introduced by rockfall within the fluvial process domain. Furthermore, Fig. 4 illustrates the variation between process domains in that reaches located in glacial terrain have much greater spread on an S–A diagram than reaches in fluvial terrain, in the context of S–A space in the diagram and per substrate type.

5.4. Downstream hydraulic geometry relationships

Although previous studies found weak or limited downstream hydraulic geometry relationships in mountain rivers (Wohl, 2004; Wohl et al., 2004), the greater sample size associated with this study indicated significant downstream hydraulic geometry relationships for the Colorado Front Range. Exponent values for depth and width relationships are within the range proposed by Park (1977), but the average exponent for width (0.63) is higher and the average exponent for depth (0.28) is lower than those proposed by Leopold and Maddock (1953), indicating that W/D increases downstream relatively rapidly in Colorado Front Range streams.

Although downstream hydraulic geometry relationships indicate progressive downstream trends in the study area, glacial history complicates the relationship between width or depth and discharge. Fluvial sites have less variable downstream hydraulic geometry relationships than glacial sites, suggesting greater adjustment to contemporary discharge. Unconfined streams have well-developed width versus discharge relationships, likely as a result of stream self-adjustment of width with changing discharge. Confined streams have well-developed depth versus discharge relationships, as stream depth rises more rapidly with increased discharge. Downstream width relationships, but not depth relationships, are well-developed when reaches are divided into stream type, except for pool– riffles, indicating that width of a stream may be more influenced by discharge, whereas depth is more influenced by confinement.

6. Conclusions

Channel geometry characteristics of individual stream types typically differ significantly between glacial and fluvial process domains in the Colorado Front Range, but no single variable consistently expresses these differences. In general, glacial stream reaches visually identified as a particular stream morphologic type (e.g., step–pool) have a greater range of values for each channel geometry parameter such as reach-scale gradient. Within a process domain, classification trees to distinguish stream types use gradient and at least two other variables, which differ between process domains. Streams within the glacial process domain have a higher misclassification rate and require more end nodes.

An S–A diagram fails to reliably differentiate bedrock and coarse-bed alluvial reaches in the study area, and the lower boundary of bedrock channel occurrence on this diagram differs from those previously proposed based on theory and empirical data from other regions. We interpret this to result from the relatively low unit discharges in the semiarid Colorado Front Range. The lower boundary defining the occurrence of bedrock streams within the fluvial process domain has a similar slope to boundaries delineated for wetter regions, but higher intercepts, which we also interpret to result from lower unit discharges in the Front Range.

Downstream hydraulic geometry relations for width and, to a lesser extent depth, are well developed in the entire data set. Splitting the data by process domain indicates stronger relationships within laterally unconfined valleys and within reaches in the fluvial process domain. In summary, our results indicate significant differences between low-order streams in glacial and fluvial process domains in the Colorado Front Range. Streams in glacial valleys flow over inherited terrain, resulting in greater diversity in channel geometry characteristics because of continuing adjustment to glacially influenced sediment and water dynamics. This variability is expressed in larger ranges of channel geometry characteristics, more variable downstream hydraulic geometry relationships, and greater scatter on an S–A diagram. The presence of significant differences in channel geometry characteristics in relation to process domains delineated based on geomorphic history and lateral valley confinement suggests that local controls, as reflected in process domains, limit the development of progressive longitudinal changes in channel geometry. The disproportionately high importance of low-order mountain streams relative to total stream length, biological productivity and habitat, and sediment production — combined with the difficulty of remotely predicting stream morphologic type because of local variability in valley type and sediment dynamics — indicate the need for field-calibration and verification in glacial and fluvial process domains of stream type as inferred from DEM-derived S or S–A relations in low-order mountain stream networks.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.geomorph.2014.12.003.

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