Downstream changes in the channel geometry of a large gravel bed river

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1. Introduction

The problem of predicting downstream trends in channel geometry (width, depth, and slope) remains central to the understanding and modeling of river behavior. Changes in channel geometry are determined both by changes in water discharge [Leopold and Maddock, 1953], and by changes in sediment load and bank properties [Schumm, 1960]. The problem therefore is to predict the equilibrium width, depth and slope of a river given representative values of discharge, sediment load and grain size. Approaches to this problem have been largely empirical, with a heavy emphasis on discharge as the primary controlling factor [Ferguson, 1986; Knighton, 1987; Church, 1992]. Discharge clearly governs sediment transport capacity; however, differences in sediment supply, as well as processes of downstream fining and bank erosion can force longitudinal changes in channel properties without much change in discharge [Schumm, 1960; Brierly and Hickin, 1985; Ferguson and Ashworth, 1991]. Thus, in addition to the equations for continuity, flow resistance, and sediment transport capacity, the solution to the problem of channel geometry requires a criterion for bank erosion to model the change in width [Henderson, 1966; Parker, 1979; Ferguson, 1986; Chang, 1988], or a “rule” that governs the downstream change in grain size to model the change in sediment load [Pizzuto, 1992].

This paper focuses on channel adjustments in a gravel bed river where the downstream change in discharge is small in relation to the change in sediment supply. The study objectives and approach draw on theory developed by Parker [1978, 1979] to explain the equilibrium morphology of rivers with mobile beds and stable banks. Parker [1978, 1979] reasoned that gravel bed rivers will adjust their bank-full width and depth to provide a boundary shear stress, , that is higher than the threshold for bed load transport, , but not so high as to cause bank erosion and widening. The geometry of a gravel bed river should thus bear a consistent relation to the excess shear stress, , and the long-term bed load flux, which scales with . In the downstream direction, changes in width, depth, slope and/or grain size should nudge the channel in a direction that maintains equilibrium transport. This approach is fundamentally different from that taken by Leopold and Maddock [1953] and many others, because flow and sediment transport are coupled directly to the processes of erosion and deposition. The theory has been refined and tested in a number of laboratory experiments [Ikeda et al., 1988; Diplas, 1990; Cao and Knight, 1998; Macky, 1999], and in a few field studies [Andrews, 1984; Pitlick and Van Steeter, 1998].

The data presented in this paper provide a relatively rigorous test of the hypothesis that gravel bed rivers adjust their bank-full channel geometry to carry bed load at shear stresses not far above critical. We test this hypothesis with closely spaced measurements of width, depth, slope, and bed-material grain size along a 260-km reach of the Colorado River in arid regions of western Colorado and eastern Utah. The study segment is characterized by repeated transitions in valley form related to hard and soft sedimentary rocks, and it includes two major tributaries: the Gunni-

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channel is sufficient to maintain a gravel bed throughout the study area. Together, the scale and characteristics of the field setting provide a somewhat unique perspective on the relation between discharge, sediment load, and channel geometry in gravel bed rivers.

2. Study Area

The present study focuses on a nearly contiguous alluvial segment of the Colorado River between approximately Rulison, Colorado, and Moab, Utah (Figure 1). In this area the Colorado River flows in a southwesterly direction, bisecting the Roan Mesa and Paradox Basin physiographic provinces of the Colorado Plateau [Liebermann et al., 1989]. As the river traverses this area, it flows through a series of Jurassic- and Cretaceous-age sedimentary rocks which vary in their resistance to erosion. Where the bedrock consists of shale, the river flows within broad alluvial valleys, and where the bedrock consists of sandstone, the river is more confined. This setting is typical of the large rivers in the region [Andrews, 1986; Allred and Schmidt, 1999; Grams and Schmidt, 1999]. The differences in valley form are the basis for subdividing the study area into 10 separate reaches, which we designate as “fully alluvial” or “quasi-alluvial” (Table 1). Fully alluvial reaches are those where the river is free to migrate laterally, and where a wide (0.2–1.0 km) floodplain is present. Quasi-alluvial reaches are more incised and partially bounded by bedrock, but the presence of floodplain segments along one or both banks indicates the river is free to adjust its width as necessary.

Changes in channel properties of adjacent alluvial and quasi-alluvial reaches are not as pronounced in the Colorado River as in some other rivers. Exceptions to this generalization occur in Westwater Canyon, where the river cuts through crystalline bedrock, and in the reaches between Dewey and Moab, Utah, where the river flows across a series of salt-cored anticlines [Doelling, 1985]. Uplift along these anticlines has been occurring for at least the last 2.5 Myr [Colman, 1983]; effects include a steepening of the profile through Professor Valley, and incision into resistant sandstones through the Big Bend reach. Aside from these two reaches, the channel characteristics of this segment of the Colorado River are not strongly influenced by transitions in reach type or junctions with major tributaries (this point is pursued in detail later).

The annual hydrograph of the Colorado River is dominated by spring snowmelt with most of the runoff originating in the mountains of central and southwestern Colorado. Streamflows are regulated by a series of storage reservoirs and water diversions upstream. These structures are small in comparison to main stem reservoirs farther downstream (e.g., Lake Powell or Lake Mead), but their operations significantly affect the timing and magnitude of peak snowmelt flows. Since 1950 instantaneous peak discharges of the Colorado River and its major tributary, the Gunnison River, have decreased by 29–43% [Van Steeter and Pitlick, 1998]. Annual sediment loads have decreased by similar amounts (30–40%), but the contribution of sediment from erosive basins in western Colorado and eastern Utah remains high. Our analysis of streamflow and sediment data from main stem gauging stations indicates that the annual suspended sediment load of the Colorado River increases markedly downstream in comparison to the discharge (Table 2). Silt and sand constitute a large part of the annual suspended sediment load; we estimate that at least 95% of the total load is carried in suspension. Gravel is a minor component of the total load, however, this material forms the bed of the channel; thus it has a major influence on channel form. Bank materials grade from predominantly gravel with a thin veneer of overbank fines in upper reaches to gravel capped with 1–2 m of silt and fine sand in lower reaches. Channel adjustments in the lower reaches are perhaps limited by the cohesion of fine-grained bank sediment and riparian vegetation, but almost everywhere the bed of the channel and toe-slopes of the banks consist of gravel which is mobile under normal flood flows.

Given the history of flow regulation on the Colorado River, it is important to establish whether the study reach is still adjusting to the altered flow regime. In a previous study based on repeat aerial photography, Van Steeter and Pitlick [1998] found that the width of the Colorado River has decreased by an average of 20 m (15%) since the late 1930s. Evidence for narrowing can sometimes be seen in the field in the form of inset benches which lie about 0.5 m below an older floodplain surface. These benches are very discontinuous, and it is not clear whether they reflect long-term changes in sediment transport capacity caused by reservoir operations or deposition in response to large floods in 1983 and 1984. To determine if channel adjustments are continuing, we examined recent trends in width and bed elevation at the U.S. Geological Survey (USGS) gauging stations at Cameo, Colorado, and Cisco, Utah. Discharge measurements are made at each of these gauges almost every month; thus, if channel change is continuing, it should be evident in the records of width and bed elevation, as shown in Figure 2. The data in the upper panel (Figure 2a) indicate that there has been no significant change in width at either gauge since 1984, when the last major flood occurred. The data in the lower panel (Figure 2b) show an irregular pattern of scour and fill at the Cameo gauge, and slight scour at the Cisco gauge. The step-like increases in bed elevation at Cameo occurred during moderate floods in 1993 and 1995; similar changes did not occur at the Cisco gauge, nor at any of our previously measured cross sections [Van Steeter and Pitlick, 1998]. The slight scour at the Cisco gauge could be natural or dam-related, but whatever the cause, the total change (0.1 m in 16 yr) is no more than about 2% of the bank-full depth. Thus, if the Colorado River is continuing to adjust to the effects of flow regulation, it is not apparent in the recent record.

3. Methods

Field measurements and data characterizing the geomorphology of this segment of the Colorado River were obtained at closely spaced intervals from Rulison, Colorado to Moab, Utah (Figure 1) (distances are given here in river kilometers, rkm, measured upstream from the Green River confluence). Limited access to short segments in DeBeque Canyon and Westwater Canyon prevented us from collecting data in portions of these two reaches.

Cross sections of the main channel were surveyed at 1.6-km intervals from rkm 366 to rkm 77 with the exception of the two segments noted above that were difficult to access and a few sites that could not be surveyed for
logistical reasons (e.g., they were located across rapids or densely vegetated islands). Cross sections were surveyed using a total station and a motorized raft outfitted with a depth sounder. To survey a cross section, the total station was set up over one of the endpoints. Distance readings were then taken along the line of the cross section by targeting a reflecting prism on the raft; at the same time, the person operating the raft would record the depth and relay the reading by radio to the person on shore. Bank-full levels were identified in the field by the break in slope between the active channel and the most recent floodplain surface. In cases where this boundary was indistinct, we sometimes relied on differences between mature and young vegetation, and in cases where there was more than one level (e.g., due to the presence of an inset bench), we referenced the lowest surface which was not part of the active channel.

Table 1. Geomorphic Characteristics of Specific Reaches of the Colorado River

<table>
<thead>
<tr>
<th>Reach Name and Location</th>
<th>Reach Type</th>
<th>w, m</th>
<th>h, m</th>
<th>slope</th>
<th>D50, mm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rulison-DeBeque (rkm 365–328)</td>
<td>A</td>
<td>114</td>
<td>2.45</td>
<td>0.00196</td>
<td>58</td>
</tr>
<tr>
<td>DeBeque Cyn. (rkm 327–300)</td>
<td>QA</td>
<td>77</td>
<td>3.12</td>
<td>0.00150</td>
<td>52</td>
</tr>
<tr>
<td>15-mile reach (rkm 298–275)</td>
<td>A</td>
<td>134</td>
<td>2.54</td>
<td>0.00175</td>
<td>58</td>
</tr>
<tr>
<td>18-mile reach (rkm 274–246)</td>
<td>A</td>
<td>175</td>
<td>3.01</td>
<td>0.00130</td>
<td>54</td>
</tr>
<tr>
<td>Ruby-Horsethief Cyn. (rkm 245–206)</td>
<td>QA</td>
<td>129</td>
<td>3.64</td>
<td>0.00100</td>
<td>44</td>
</tr>
<tr>
<td>Westwater Cyn. (rkm 205–182)</td>
<td>B</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Cisco-Fish Ford (rkm 180–153)</td>
<td>A</td>
<td>147</td>
<td>4.49</td>
<td>0.00258</td>
<td>38</td>
</tr>
<tr>
<td>Dewey (rkm 151–140)</td>
<td>QA</td>
<td>132</td>
<td>5.14</td>
<td>0.00047</td>
<td>35</td>
</tr>
<tr>
<td>Professor Valley (rkm 138–126)</td>
<td>A</td>
<td>203</td>
<td>4.61</td>
<td>0.00149</td>
<td>69</td>
</tr>
<tr>
<td>Big Bend (rkm 124–113)</td>
<td>B</td>
<td>106</td>
<td>6.43</td>
<td>0.00098</td>
<td>63</td>
</tr>
<tr>
<td>Moab (rkm 111–105)</td>
<td>A</td>
<td>151</td>
<td>5.13</td>
<td>0.00034</td>
<td>25</td>
</tr>
</tbody>
</table>

Locations of reaches are given in river kilometers (rkm) upstream of the Green River confluence.

A indicates alluvial reach; QA indicates quasi-alluvial reach; B indicates bedrock-bounded reach.

No measurements were taken in Westwater Canyon.
Subsurface samples were taken by collecting 100–150 kg of the sediment underneath the surface layer. These samples were nearly always large enough to ensure that the coarsest particle represented no more than 5% of the total weight. The coarse fraction of this sediment was sieved in the field, while the fine fraction was sieved in the laboratory. A total of 78 surface samples and 27 subsurface samples were taken in the 260-km study reach.

The average channel slope of the 10 individual reaches was measured with a global positioning system (GPS) supplemented in a few places by measurements from topographic maps. Readings of the water surface were taken every 0.8-km with a mapping-grade GPS (Trimble Pathfinder Pro-XR). These data were subsequently corrected with differential post-processing techniques, yielding vertical positions with errors of ±0.5–0.3 m. These errors tend to be random and are small in comparison to the total drop in elevation through most reaches (25–50 m). As a further check on accuracy, we compared slopes derived from the GPS measurements with those derived from topographic maps and found that they were essentially the same.

The hypothesis stated in the introduction implies that gravel bed rivers adjust their bank-full width \(B\), depth \(H\), and slope \(S\) to transport bed load at shear stresses slightly above the threshold for motion. If so, the width, depth, slope and grain size should change simultaneously downstream to give a constant bank-full dimensionless shear stress \(\tau^*\). The dimensionless shear stress, or Shields stress, is defined as \(\tau^* = \tau \left[ \frac{\rho_s}{\rho} g D \right]^{1/2}\), where \(\tau = \rho g H S\) is the bank-full shear stress, \(\rho_s\) and \(\rho\) are the densities of sediment and water, respectively, \(g\) is the gravitational acceleration, and \(D\) is the grain size. We formulated individual values of \(\tau^*\) for each cross section using the measured bank-full depth, the reach-average slope, and the reach-average median grain size of the surface sediment, \(D_{50}\). Reach-average values of \(S\) and \(D_{50}\) were used largely for practical reasons: we could not measure the surface grain size or slope at every cross section, thus we combined measurements in each of the 10 reaches. In our experience reach-average slopes do not differ greatly from local slopes at bank-full flow [Pitlick and Van Steeter, 1998], and variations in grain size have as much to do with the local topography and sedimentology of individual bars as with systematic downstream fining or tributary inputs.

Standard statistical tests were used to evaluate the significance of regression relations between \(\ln\) transformed values of grain size, width, depth, \(\tau\), and \(\tau^*\), and distance downstream. We examined trends within individual reaches as well as overall trends. T test comparisons were used to assess differences in local versus regional trends in grain size and to assess the correlation between downstream changes in grain size and shear stress. Analysis of covariance was used to examine differences in channel geometry between alluvial and quasi-alluvial reaches [Kleinbaum and Kupper, 1978].

4. Field Observations and Results

4.1. Channel Slope

The longitudinal profile of this portion of the Colorado River is made up of four concave segments separated by three breaks in slope (Figure 3). The breaks in slope occur in...
DeBeque Canyon, Westwater Canyon and Professor Valley. The first of these is caused by a series of three low-head diversion dams which divert water into the Grand Junction area. These dams alter the profile but do not reflect natural processes. The breaks in slope through Westwater Canyon and Professor Valley are related to local geologic and tectonic processes. The segments in between these points define smooth profiles, which are not strongly influenced by transitions in reach type (alluvial to quasi-alluvial) or by junctions with tributaries. The segment between DeBeque Canyon (rkm 300) and Westwater Canyon (rkm 205) for example, includes a major tributary, the Gunnison River at rkm 274, and a transition in reach type at rkm 245. Neither of these features appears to affect the profile in a strong way. Elsewhere we observe similar trends, indicating that transitions between hard and soft sedimentary rocks have little influence on the overall form of the longitudinal profile. Similar to Rice and Church [2001], we find that the concave segments of the profile are better fit by quadratic functions than exponential functions.

4.2. Bed Material

[16] The bed material of this segment of the Colorado River grades from cobbles and large gravels in the upper reaches to medium gravels in the lower reaches. Plots of specific percentiles of the surface grain-size distributions ($D_{84}$, $D_{50}$, and $D_{16}$) define a weak trend of downstream fining (Figure 4). Locally high values in two of the lower reaches (Professor Valley and Big Bend) skew the data somewhat, but even with these values excluded, the surface sediment fines at relatively slow rates. The trend lines shown in Figure 4 (and subsequent figures) are fitted exponential relations, $ln y = ln a + bx$, where $ln a$ is the $y$ intercept, $b$ is the slope of the line, and $x$ is the distance, measured upstream with respect to the Green River (rkm 0). In this case the coefficient $a$ represents the grain size at $x = 0$, and $b$ represents the rate of downstream fining. These values are listed in Table 3, along with relevant parameters for the subsurface sediment, channel morphology and shear stress. Comparison of the values of $b$ for the surface percentiles indicates that large and small sizes fine downstream at about the same rate (Table 3). In comparison to trends reported in other studies [Knighton, 1998; Rice, 1999; Gomez et al., 2001], the rate of surface-finishing in the Colorado River is relatively weak. In this case, surface particle sizes change slowly downstream because coarse material is continually supplied from local sources such as ephemeral tributaries, terraces, and valley side-slopes. Input from these sources is not large enough to overwhelm downstream trends, but apparently sufficient to replenish coarse material worn down by abrasion.

[17] A close inspection of the data in Figure 4 suggests that the rate of downstream fining within several individual segments is higher than the overall trend. Stronger fining is apparent in the reach between DeBeque Canyon and Westwater Canyon (rkm 300–200), and in the reach below Professor Valley (rkm 130–100). To evaluate the importance of these differences, we separated the data into four segments bounded by the breaks in slope discussed previously. For each segment we derived separate exponential relations for the surface $D_{50}$ (Table 4) (relations for the bank-full width, depth, shear stress and Shields stress were also derived, but, for the moment, we focus only on the downstream trends in surface $D_{50}$). The results listed in
Table 3. Parameter Estimates for Regression Relations Between ln Transformed Values of Grain Size, Width, Depth, Shear Stress, Shields Stress, and Distance

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>a</th>
<th>b</th>
<th>SEb</th>
<th>r²</th>
<th>F</th>
<th>p</th>
</tr>
</thead>
<tbody>
<tr>
<td>ln surface D64</td>
<td>78</td>
<td>57.6</td>
<td>0.0015</td>
<td>0.00050</td>
<td>0.11</td>
<td>9.4</td>
<td>0.003</td>
</tr>
<tr>
<td>ln surface D50</td>
<td>78</td>
<td>33.4</td>
<td>0.0015</td>
<td>0.00049</td>
<td>0.11</td>
<td>9.1</td>
<td>0.003</td>
</tr>
<tr>
<td>ln surface D16</td>
<td>78</td>
<td>18.4</td>
<td>0.0013</td>
<td>0.00041</td>
<td>0.12</td>
<td>10.8</td>
<td>0.002</td>
</tr>
<tr>
<td>ln subsurface D4</td>
<td>27</td>
<td>45.9</td>
<td>0.0021</td>
<td>0.00075</td>
<td>0.24</td>
<td>8.0</td>
<td>0.009</td>
</tr>
<tr>
<td>ln subsurface D6</td>
<td>27</td>
<td>16.5</td>
<td>0.0023</td>
<td>0.00081</td>
<td>0.25</td>
<td>8.3</td>
<td>0.008</td>
</tr>
<tr>
<td>ln subsurface D1</td>
<td>27</td>
<td>0.8</td>
<td>0.0034</td>
<td>0.00202</td>
<td>0.10</td>
<td>2.9</td>
<td>0.10</td>
</tr>
<tr>
<td>ln width</td>
<td>132</td>
<td>167.0</td>
<td>-0.0011</td>
<td>0.00036</td>
<td>0.07</td>
<td>9.7</td>
<td>0.002</td>
</tr>
<tr>
<td>ln depth</td>
<td>132</td>
<td>7.8</td>
<td>-0.0035</td>
<td>0.00026</td>
<td>0.58</td>
<td>180</td>
<td>&lt;0.001</td>
</tr>
<tr>
<td>ln shear stress</td>
<td>132</td>
<td>27.6</td>
<td>0.0014</td>
<td>0.00039</td>
<td>0.08</td>
<td>12.0</td>
<td>&lt;0.001</td>
</tr>
<tr>
<td>ln Shields stress</td>
<td>132</td>
<td>0.048</td>
<td>-0.00005</td>
<td>0.00026</td>
<td>&lt;0.01</td>
<td>0.03</td>
<td>0.858</td>
</tr>
</tbody>
</table>

Here n is the number of observations; a and b are parameters in the relation \( \ln y = \ln a + bx \), with x measured in km; SEb is the standard error of the coefficient b; \( r^2 \) is the coefficient of determination; F is the value of the F distribution; and p is the significance probability.

Table 4 indicate that the trends in surface \( D_{50} \) are not significantly different from zero (p > 0.05), except in the segment between DeBeque Canyon and Westwater Canyon. We would be surprised if this was not the case, given that the segment is 100-km long, and includes a 40% increase in discharge from the Gunnison River. The other segments fail the test of significance because the standard error of the regression \( SE_b \) is of the same order as the coefficient b itself. To determine if fining is indeed stronger within the DeBeque-Westwater segment, we then performed an analysis of covariance (ANCOVA), using a binary predictor variable for the separate segments (DeBeque-Westwater vs. the remainder of the data). The results of the ANCOVA indicate that there is a marginal difference in the slopes of the separate regression relations for \( \ln D_{50} \) (p = 0.08), suggesting that the rate of downstream fining within the DeBeque-Westwater segment is not much different from the trend defined by the remainder of the data. What is perhaps more important to the analysis and the main hypothesis of this study is the correlation between grain size and shear stress, which we pursue later in the section on channel geometry.

[18] Trends in the percentiles of the subsurface sediment (Figure 5) indicate that the bulk bed material of the Colorado River also fines systematically downstream. Regression analysis of these data indicates that the subsurface \( D_{64} \) and \( D_{50} \) are modeled reasonably well by exponential relations that parallel each other (Table 3); the relation for \( D_{16} \) is not as strong (p = 0.10), and the trend is slightly steeper. Comparison of the fining coefficients b for surface and subsurface \( D_{50} \) (Table 3) suggests that the bulk bed material fines downstream at slightly higher rates than the pavement. The difference reflects an increase in the proportion of sand and granules in the bed, and implies that the surface coarsens downstream relative to the subsurface. However, a plot of the ratio of surface \( D_{50} \) to subsurface \( D_{50} \) derived from paired samples at the same locations (Figure 6) shows that this trend is not very strong (\( r^2 = 0.22, p = 0.02 \)). On average the surface \( D_{50} \) is about 1.5 times the subsurface \( D_{50} \).

4.3. Channel Geometry

[19] The downstream trends in bank-full channel width and depth, derived from individual cross section measurements, are plotted in Figure 7. These data are further subdivided according to reach type (alluvial or quasi-alluvial), and fit with separate regression equations. To assess the influence of reach type on overall trends, we compared the regression estimates of the intercept a and slope b of the separate relations using an analysis of covariance (ANCOVA), with distance as a covariate and reach type as a factor. The purpose of these tests was to establish whether changes in channel geometry are driven more by one reach type than the other, and to examine potential interactions between reach type and distance.

[20] The results of this analysis are summarized in Table 5, which lists the regression estimates of a and b for each relation, standard measures of the strength and significance of the relation, and a summary of the ANCOVA results with different levels of complexity. A test of the regression equations for \( \ln \) bank-full width indicates that there is no significant difference in the slopes of the separate relations (p = 0.44); therefore reach type does not appear to affect downstream trends in bank-full width. Removing the interaction term (distance x reach type) results in a significant difference in adjusted means (p < 0.001), thus bank-full widths are consistently higher in alluvial reaches than in quasi-alluvial reaches. Tests of the regression equations for \( \ln \) bank-full depth likewise indicate no significant difference in the slopes of the separate relations (p = 0.53); thus downstream trends in bank-full depth do not appear to be affected by reach type. Removing the interaction term results in a significant difference in adjusted means (p = 0.01) suggesting that bank-full depths are, on average, lower in alluvial reaches than in quasi-alluvial reaches.

[21] Previous studies of downstream changes in channel geometry have shown that, when discharge is used as the

Table 4. Parameter Estimates and Measures of Statistical Significance of Within-Segment Relations Between Individual ln Transformed Variables and Distance

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>a</th>
<th>b</th>
<th>SEb</th>
<th>r²</th>
<th>F</th>
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</tr>
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<tr>
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<td>0.002</td>
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<tr>
<td>ln subsurface D4</td>
<td>27</td>
<td>45.9</td>
<td>0.0021</td>
<td>0.00075</td>
<td>0.24</td>
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<tr>
<td>ln subsurface D6</td>
<td>27</td>
<td>16.5</td>
<td>0.0023</td>
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<td>8.3</td>
<td>0.008</td>
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<tr>
<td>ln subsurface D1</td>
<td>27</td>
<td>0.8</td>
<td>0.0034</td>
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<tr>
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<td>&lt;0.01</td>
<td>0.03</td>
<td>0.858</td>
</tr>
</tbody>
</table>

The notation is the same as in Table 3.
scaling variable, the bank-full width typically increases faster downstream than the bank-full depth [Leopold and Maddock, 1953; Church, 1992; Knighton, 1998]. Our data exhibit the opposite trend. The slope of the overall regression equation for bank-full depth is much higher than the slope of the overall equation for bank-full width (Table 3). Over a total distance of 260 km, the bank-full depth more than doubles, whereas the bank-full width increases by only about 50% (Table 1). The downstream trends in channel geometry of the Colorado River appear to be driven by the particular combination of input variables; in this case the discharge and grain size change slowly in relation to the slope. The disproportionate increase in channel depth helps satisfy the requirements of continuity for both water and sediment, but a large increase in width is not necessary if the input variables (discharge, total bed load, grain size) are slowly varying.

[22] We conclude this section by focusing on downstream trends in the bank-full shear stress, \( \tau_b \), and the bank-full Shields stress, \( \tau'_b \). Figure 8 shows the downstream trends in \( \tau_b \) with separate segments marked as in previous figures. The overall trend indicates that there is roughly a 3-fold decrease in \( \tau_b \), from an average of 47 N/m\(^2\) in the uppermost reach to an average of 17 N/m\(^2\) in the lowermost reach. An exponential fit of the full data set gives the relation \( \ln \tau_b = 27.6 - 0.00135 X \), with \( r^2 = 0.08 \) and \( p < 0.001 \). However, this trend is clearly skewed by the high values in the lower reaches of the study area (rkm 140–115); with these values removed, the relation is considerably steeper, \( \ln \tau_b = 16.9 - 0.0030 X \), and the correlation is higher (\( r^2 = 0.47; p < 0.0001 \)).

[23] Similar to the trends in surface \( D_{50} \), there appear to be steeper trends in \( \tau_b \) within several segments. In fact, the
most notable changes in τb occur in the same segments where there is clear downstream fining (rkm 300–200 and rkm 139–100); the reverse is also true, i.e. where τb varies slowly there is also little variation in grain size (rkm 370–310 and rkm 185–140). The correlation between τb and D50 is clearly important to our main hypothesis, as it implies that downstream fining is weak, suggesting that the supply of coarse sediment from tributaries, hillslopes, and terraces is high enough to replenish material worn down by abrasion, but not high enough to force significant changes in

words, there is no significant trend in lnD50 or lnτb in either of these segments. A t test of the equations for lnD50 and lnτb in the second segment (Figure 9b) indicates that there is no significant difference in the slopes of these relations (p = 0.49). Comparison of the equations for the lower segment (Figure 9d) also indicates no significant difference in slopes (p = 0.08), however, this result is not convincing, given the scatter in the grain size data; if the two values of D50 at rkm 139 and 136 are removed, then the relations for lnD50 and lnτb are nearly parallel.

[24] The results presented above indicate that the change in τb within individual segments is roughly balanced by the change in D50. It follows that the bank-full Shields stress, τb, should be approximately constant from segment to segment. A plot of the individual values of τb (Figure 10) suggests that this is indeed the case. A least squares fit of the full data set gives the relation \( \ln \tau_b = 0.048 - 4.7 \times 10^{-05} \times \) with \( r^2 \ll 0.01 \) and \( p = 0.86 \). If the data from rkm 139–105 are excluded, there appears to be a slight negative trend in \( \ln \tau_b \), but the relation is very weak (\( r^2 = 0.03; p = 0.07 \)). Even with these data excluded, the difference in \( \tau_b \) between upper and lower reaches is very small (0.05 to 0.045), thus we have no reason to reject the hypothesis that \( \tau_b \) is constant downstream. For the reach as a whole, \( \tau_b \) averages 0.049, which is about 1.5 times typical values for the threshold for motion (0.03).

5. Discussion

[25] The results of this study show that the geomorphic characteristics of the Colorado River vary systematically downstream to maintain a constant bank-full Shields stress, \( \tau_b \). The consistency in \( \tau_b \) holds for reaches that are bounded by different sedimentary strata and potentially for reaches that are influenced by recent tectonism. The Colorado River has developed these geomorphic characteristics apparently in response to regional-scale processes which supply water and sediment in disproportionate amounts. The average annual discharge of the Colorado River roughly doubles through the study area, whereas the average annual suspended sediment load increases by a factor of almost four. Longitudinal measurements of bed material grain size show that downstream fining is weak, suggesting that the supply of coarse sediment from tributaries, hillslopes, and terraces is high enough to replenish material worn down by abrasion, but not high enough to force significant changes in

![Figure 8](image-url) Downstream trends in the bank-full shear stress of the Colorado River. Symbols are the same as in Figure 7.
channel slope. The relatively large increase in bank-full depth is thus driven by a mass balance requirement of carrying coarse sediment across a relatively low slope with little additional flow. Ferguson and Ashworth [1991] describe similar trends on the Allt Dubhaig, a small stream in Scotland that undergoes systematic changes in channel geometry with almost no change in discharge. The Allt Dubhaig is characterized by a rapid decrease in slope and grain size, and a modest increase in channel depth [Ferguson and Ashworth, 1991]. One difference here is the increasing abundance of fine sediment and bank vegetation, which help constrict the channel and promote vertical accretion during floods, similar to what Allred and Schmidt [1999] have observed on the Green River in Utah. Limits to this process are determined by bank stability and the resistance of the channel to high shear stresses during floods. Fine and coarse sediment in the Colorado River thus interact to form a channel that generates sufficient shear stress to maintain bed load transport without widening.

Our observation that the width-depth ratio of the Colorado River decreases downstream contrasts with results from many other studies [Church, 1992; Knighton, 1998]. The differences in this case are most likely the result of a small change in discharge, coupled with a small change in total bed load flux. To support this point we compare data from the Colorado River with a set of hydraulic geometry relations formulated by G. Parker [personal communication]; for purposes of discussion we will say that Parker’s data set is representative of “typical” gravel bed rivers. Using data from 62 gravel bed rivers in Canada, the UK and the USA, Parker formulated the following dimensionless hydraulic geometry relations:

\[ B^* = 4.9 Q^{0.46} \]  \hspace{1cm} (1a)

\[ H^* = 0.37 Q^{0.41} \]  \hspace{1cm} (1b)

\[ S = 0.098 Q^{0.34} \]  \hspace{1cm} (1c)

\[ \tau_b^* = 0.049 \]  \hspace{1cm} (1d)

The variables are defined as follows: \( B^* = B/D_{50} \) is the dimensionless bank-full width; \( H^* = H/D_{50} \) is the dimensionless bank-full depth; \( Q^* = Q/(\sqrt{RgD_{50}D_{50}}) \) is the dimensionless bank-full discharge; \( Q \) is the bank-full discharge; and \( R \) is the submerged specific gravity of the sediment, assumed to be 1.65.

Figure 11 shows corresponding hydraulic geometry relations for the Colorado River. In this case, values of \( Q \) were calculated for individual cross sections using the Manning equation for velocity and the measured cross sectional area. Estimates of Manning’s \( n \) were made on the basis of flow modeling results, verified by field observations [Pitlick and Cress, 2000]. The resulting hydraulic geometry equations are

\[ B^* = 39.3 Q^{0.32} \]  \hspace{1cm} (2a)

\[ H^* = 0.07 Q^{0.53} \]  \hspace{1cm} (2b)

\[ S = 0.78 Q^{-0.50} \]  \hspace{1cm} (2c)

\[ \tau_b^* = 0.049 \]  \hspace{1cm} (2d)

Comparison of the two sets of equations reveals some interesting differences and similarities. The exponent in the width relation for the Colorado River is low in comparison to the relation for “typical” gravel bed rivers, and the exponents for depth and slope are correspondingly
Both data sets support the hypothesis that \( \tau_b^* \) is independent of \( Q^* \), and the average values of \( \tau_b^* \) are identical (0.049). This last result has interesting implications for bed load transport. Typical transport equations express the dimensionless unit bed load transport rate, \( q_b^* \), as a function of the excess shear stress, \( \tau_b^* = \tau_c^* \), raised to a power >1. The results above imply that if \( \tau_b^* \) is constant, then \( q_b^* \) is also constant (assuming that \( \tau_c^* \) does not also vary). By definition, the volumetric bed load transport rate per unit width is \( q_b = q_b^* \sqrt{RgD^{3/2}} \), and the total transport rate is \( Q_b = Bq_b \). Combining these relations, we get \( Q_b = Bq_b^* \sqrt{RgD^{3/2}} \), or more generally, \( Q_b \propto BD^{3/2} \). Thus, if \( q_b^* \) is indeed constant, then the downstream change in total bed load transport depends on the change in \( B \) relative to the change in \( D^{3/2} \). Our data show that the grain size decreases more rapidly than the width increases (Table 3), suggesting that the total volume of bed load carried by the Colorado River is decreasing downstream. This interpretation is supported to a limited extent by the comparison of surface and subsurface sediment (Figure 6), which shows a progressive coarsening of the surface layer downstream. The surface layer regulates the transport of different sizes, and for particular combinations of discharge, slope and sediment supply, the surface texture may become finer or coarser downstream [Parker, 1990]. The development of a static armor downstream of a dam represents the extreme case of supply limitation, while the absence of a surface layer may indicate high sediment supply [Lisle and Madej, 1992]. The Colorado River lies somewhere in the middle of this spectrum, with hillslopes and ephemeral tributaries maintaining the supply of coarse material, but apparently at rates that diminish downstream.

6. Conclusions

[30] The segment of the Colorado River examined in this study exhibits morphologic characteristics that are, in some aspects, very different from other gravel bed rivers, but in other aspects quite similar. We find, for example, that channel properties (bank-full width, depth, slope and bed material size) change systematically downstream, and that transitions in reach type caused by local changes in bedrock geology often have little effect on overall trends. Partitioning of cross section data from alluvial and quasi-alluvial reaches shows that downstream relations for bank-full width and bank-full depth cannot be distinguished on the basis of reach type. In contrast to the trends observed in many other rivers, the bank-full width of the Colorado River increases slowly downstream in comparison to the bank-full depth. The differences in bank-full width and depth appear to be driven by competing effects of slope and grain size. In comparison to other rivers, the slope of the Colorado River decreases more rapidly downstream, whereas the grain size decreases more slowly. Continuity requires that the river form a deeper
channel, such that the shear stress is sufficient to carry coarse material across a lower slope with little additional discharge. [31] The physical mechanisms of channel adjustment in the Colorado River are strongly tied to bed load transport processes, as shown by the consistency in the bank-full Shields stress, \( \tau^*_{bf} \). Our measurements indicate that \( \tau^*_{bf} \) does not vary with distance or discharge, averaging 0.049 for the reach as a whole. The consistency in \( \tau^*_{bf} \) suggests that in spite of differences in slope and grain size, the Colorado River has adjusted its bank-full width and depth to a Shields stress that is 20–50% higher than typical thresholds for motion (0.03–0.04). The equilibrium width and depth therefore appear to be maintained by flows slightly in excess of the threshold for motion.

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