Evaluation of in-channel gravel storage with morphology-based gravel budgets developed from planimetric data

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[1] Gravel budgets developed from changes in river morphology have emerged as an important tool for exploring stream dynamics and sediment transport. In many cases, old aerial photographs are the only data available by which to evaluate past morphologic changes. Existing methods for building morphology-based gravel budgets from air photos are subject to several sources of uncertainty, including difficulty in identifying deposition within the active channel. In-channel deposition sites consist of submerged bars or general increases in bed elevation, both of which are virtually impossible to detect on air photos. However, bed deposition can be inferred when measured gravel storage losses are greater than the quantity of gravel exiting the study area at its downstream boundary. An integrated method combining measurements of gravel storage changes with a gravel routing procedure based on estimated gravel transport path lengths was developed to identify sites of bed aggradation. Inclusion of storage change results in the routing procedure reduces the uncertainty associated with the selection of appropriate transport path lengths. The method was applied to development of gravel budgets for a 50-year period in the lower Duchesne River, Utah. Areas in which the predicted bed aggradation was greatest displayed higher rates of channel activity and greater channel instability during subsequent time periods. INDEX TERMS: 1824 Hydrology: Geomorphology (1625); 1815 Hydrology: Erosion and sedimentation; 1860 Hydrology: Runoff and streamflow; KEYWORDS: channel morphology, aerial photographs, sediment budgets


1. Introduction

[2] Sediment budgets are a powerful tool in geomorphology [Reid and Dunne, 1996]. The construction of sediment budgets provides a basis for quantifying sediment transport pathways [Kondolf and Matthews, 1991], and is fundamental for identifying the critical factors affecting landscape development [Singer and Dunne, 2001]. Measurement of morphologic change in order to construct historical sediment budgets has emerged as a useful approach for quantifying stream dynamics and fluvial sediment transport [Goff and Ashmore, 1994; Martin and Church, 1995; Lane and Richards, 1995; Ashmore and Church, 1998; McLean and Church, 1999; Ham and Church, 2000; Eaton and Lapointe, 2001; Sutherland et al., 2002]. A morphology-based analysis of bed load transport in rivers consists of repeated measurement of channel morphology, from which volumes of bed material eroded or deposited during the time period between measurement dates are estimated.

[3] The morphology-based approach is applicable over a large range of spatial and temporal scales. Measurements can consist of ground surveys of regularly spaced channel cross sections [Martin and Church, 1995], more detailed ground surveys [Goff and Ashmore, 1994], digital photogrammetry [Lane, 2000], or interpretation of historical aerial photos [McLean and Church, 1999; Ham and Church, 2000]. In many historical studies, the only data available are old aerial photographs from which it is difficult or impossible to obtain precise topographic measurements, thereby limiting the resolution possible in historical sediment budgets. These limitations can be partially overcome by integrating two complementary analytical methods used in sediment budget development: the storage change [e.g., Martin and Church, 1995; McLean and Church, 1999; Ham and Church, 2000; Sutherland et al., 2002] and travel distance [e.g., Goff and Ashmore, 1994; Ham and Church, 2000; Eaton and Lapointe, 2001] approaches. Each method is based on different assumptions, and each has somewhat different data requirements and limitations. We incorporate elements of both approaches to develop a gravel budget covering a 50-year period for a stream in northeastern Utah.

[4] Our objectives are to (1) show that undetected bed level adjustments can significantly affect sediment budgets...
2. Study Area

The Duchesne River and its major tributaries drain the south slope of the Uinta Mountains in northeastern Utah (Figure 1). The flow regime is snowmelt driven with annual spring peaks. The study area covers a gravel bed portion of the Duchesne River extending from near the mouth of the Uinta River downstream to where the streambed changes to sand ~9.3 km upstream from the Green River. To facilitate the calculation of sediment budgets, the study area was divided into 15 subreaches we refer to here as budget cells (Figure 1). Each budget cell contains between one and four meander bends and ranges between 0.5 – 2.7 km in valley length. Cell boundaries were chosen at meander inflection points where the channel has been relatively stable, so that a single set of cell boundaries could be used for the study period.

The alluvial valley is ~2 km wide and is a wandering gravel bed river floodplain (order B2), using the classification of Nanson and Croke [1992]. The valley is bordered on either side by benches exceeding 100 m above the present river level. These benches are remnants of Pleistocene outwash plains capped with quartzite-rich gravel [Osborn, 1973]. Quartzite-rich gravel also underlies the valley floor. The channel has a meandering planform, a nearly constant slope with average gradient of ~0.0019, bank-full width of ~40 to 50 m, and a bank-full depth of ~2 m (Figure 2). The bed is predominately composed of gravel and cobbles. This morphology is consistent through the study area. Downstream from river km 9.3, the channel slope abruptly decreases to less than 0.0003 (Figure 3), bed material changes to sand, and channel form assumes a narrow canal-like geometry.

Streamflow has been measured at a USGS gaging station immediately downstream from the mouth of the Uinta River (Duchesne River near Randlett, station 09302000) since 1943 (Figure 1). The mean annual flow in the study area is 15.8 m³ s⁻¹, and the mean annual peak flow is ~122 m³ s⁻¹. No tributaries enter the Duchesne River within the study area.

3. Geomorphic Mapping and Attributes of the Channel and Valley

The geomorphology of the alluvial valley was mapped in the field in June 2000, using a 1997 aerial photograph base. Map units, ordered from lowest to highest derived from planimetric data sources; (2) propose a general gravel routing structure for implementing travel distance concepts in sediment budget development, and; (3) integrate sediment routing results with calculations of gravel storage changes to more fully reconstruct geomorphic history. In particular, this integrated approach provides a means for detecting and quantifying past adjustments in bed elevation from historical air photos.

Figure 1. Study area map with field sites indicated. The areas numbered from 1 to 15 are budget cells in which gravel erosion and deposition volumes were computed. The shaded areas adjacent to the channel indicate high bar surfaces, secondary channels, and chutes.

Figure 2. Example cross section showing typical topographic and stratigraphic relationships between some of the geomorphic surfaces present in the study area. Tamarisk terrace is not shown, but it is similar to high terrace in elevation and materials. The cross section shown is located within the Below Bowtie site, as indicated on Figure 1.
becomes widespread at

Model results indicated that inundation of point bar surfaces necessary because of protracted drought during our study. Surveyed cross sections, respectively. Model estimates were using a one-dimensional hydraulic model (HEC-RAS, U.S. Above Pipeline, and Wissiup detailed study sites (Figure 1), 3.1. The Active Channel
during the intervals between each photograph series. Changes in the distribution of these map units were the basis for determining areas of erosion or deposition during the intervals between each photograph series.

3.1. The Active Channel
[9] The active, or bank-full, channel includes the wetted channel and emergent low bars, which are defined as unvegetated gravel bars within the bank-full channel (Figure 2). We considered low bars as part of the active channel to avoid problems that arise with different discharges at the time of each photograph series.

[10] We estimated bank-full discharge at the 24-hour, Above Pipeline, and Wissiup detailed study sites (Figure 1), using a one-dimensional hydraulic model (HEC-RAS, U.S. Army Corp of Engineers) and field surveys of 14, 5, and 10 surveyed cross sections, respectively. Model estimates were necessary because of protracted drought during our study. Model results indicated that inundation of point bar surfaces becomes widespread at ~113 m² s⁻¹. Aerial photographs taken 4 June 1999, when discharge was 110 m³ s⁻¹, support this estimate of bank-full discharge.

[11] The bed in the study area is primarily gravel with some sand patches in pools and along the inside banks of meander bends, especially at base flow. Surface bed material sizes were measured using standard point count procedures on two riffles at the three study sites described above, plus point counts on one riffle each at km 20.5, the Bowtie site, and the Below Bowtie site (Figure 1). The D₅₀ ranges between ~30 mm and 75 mm and does not display systematic downstream fining (Table 1); the smallest gravels occur in the vicinity of the Bowtie.

[12] The bed subsurface has interstitial sand, based on one subsurface sample collected at four sites (Table 1). Measures equivalent to the surface point counts were obtained using the mass-by-volume procedures suggested by Church et al. [1987]. In all cases, the largest particle accounted for no more than 2% of the total sample mass. At three of the four subsurface sites, the ratio of the surface to subsurface median particle size is less than 1.6, indicating that the bed is lightly armored [Liste and Church, 2002]. At the 24-hour site, surface and subsurface median particle sizes are approximately equal. The sand content of the four subsurface samples ranged between 13% and 20% by mass.

3.2. The Active Floodplain
[13] We distinguished two components of the active floodplain: high bars and floodplains. High bars are point bars and other lateral bars or islands (Figure 2). Their surfaces are generally sandy with scroll ridge topography and relatively sparse vegetation consisting of tamarisk (Tamarix spp.), Russian olive (Elaeagnus angustifolia), and occasional cottonwood (Populus spp.) seedlings. Gravel is locally exposed in scoured areas and chute channels. Areas mapped as floodplain occur at similar elevations, support dense stands of tamarisk and Russian olive, and are typically covered with a layer of silt and organic material several cm thick. Floodplains are generally separated from adjacent high bars by well-developed chute channels. Measurements in cutoff chutes, bank exposures, excavated pits, and auger holes indicate that both the high bars and floodplains are composed of a basal layer of gravelly channel deposits overlain by a layer of sand ~1 m thick (Figure 2).

3.3. Terraces
[14] Three fill terraces comprise more than 75% of the alluvial valley. The elevation of the lowest terrace, which we informally name the cottonwood terrace, averages ~0.7 m above the floodplain (Figure 2). The cottonwood terrace is composed of gravelly channel deposits overlain by a layer of sand or silt averaging ~1.5 m in thickness, and supports several species of xeric shrubs and scattered mature or dead cottonwood trees. A higher surface, designated the high terrace, occurs primarily on the west side of the river at an elevation more than 3 m above the active floodplain. Its surface slopes toward the valley center and contains scattered eolian sand dunes. The high terrace supports only xeric vegetation, and is composed predominantly of fine-grained material. Gravel is generally absent in

Table 1. Surface and Subsurface Particle Sizes

<table>
<thead>
<tr>
<th>Site Name</th>
<th>River Km</th>
<th>D₅₀</th>
<th>D₉₀</th>
<th>D₅₀</th>
<th>D₉₀</th>
<th>% Sand</th>
</tr>
</thead>
<tbody>
<tr>
<td>24-hour</td>
<td>22.0</td>
<td>55</td>
<td>110</td>
<td>52</td>
<td>70</td>
<td>13</td>
</tr>
<tr>
<td>Km 20.5</td>
<td>20.5</td>
<td>48</td>
<td>87</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>At Bowtie</td>
<td>17.9</td>
<td>30</td>
<td>59</td>
<td>–</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>Below Bowtie</td>
<td>16.3</td>
<td>44</td>
<td>93</td>
<td>30</td>
<td>42</td>
<td>15</td>
</tr>
<tr>
<td>Above Pipeline</td>
<td>13.2</td>
<td>75</td>
<td>127</td>
<td>45</td>
<td>64</td>
<td>20</td>
</tr>
<tr>
<td>Wissiup</td>
<td>10.1</td>
<td>52</td>
<td>93</td>
<td>35</td>
<td>43</td>
<td>18</td>
</tr>
</tbody>
</table>

Sizes in mm. Surface sizes are based on pooled data for sites with more than one surface count. Subsurface sizes are based on the gravel fraction only (>2 mm). Subsurface sand content is expressed as percent by mass.
bank exposures above the level of the modern channel bed. A surface of similar elevation found at the extreme downstream end of the study area consists of a nearly featureless plain dominated by dense tamarisk thickets, and is referred to as the tamarisk terrace.

3.4. Longitudinal Changes in the Elevation of Gravel Horizons in the Alluvial Valley

[15] The elevations of the tops of the gravel deposits underlying floodplains and terraces were mapped at all cutbanks. These measurements allowed us to estimate the thickness of gravel involved in deposition or erosion throughout the study area. Gravel horizon elevations were measured relative to a reference water surface stage, and were subsequently adjusted to a sloping datum defined by the reach-averaged thalweg elevation (Figure 3). Georeferenced water surface elevations were obtained by differential GPS, and the average thalweg depth below the reference water surface was determined at detailed study sites using topographic survey data. We estimated the mean height of the streambed above the thalweg as one half the relief of the gravel bed, based on the roughly triangular geometry typical of most channel cross sections (Figure 2).

[16] Measurements of gravel horizon elevations in the cottonwood terrace unit were made upstream from budget cell 10 and downstream from cell 8. No cottonwood terrace gravel horizons were measured in budget cells 8, 9, and 10, because all terrace cutbank exposures in this portion of river are in the high terrace unit. Gravel horizon elevations in the cottonwood terrace for all budget cells were estimated using a least squares linear fit through the measured horizon elevations. The standard deviation of the residuals about this regression line were 0.33 m, and we used this value to assign error margins to our estimates of the volume of gravel involved in erosion or deposition during different time periods. Gravel horizon elevations decline slightly downstream between the upstream study area boundary and cell 4. The cottonwood terrace grades into the tamarisk terrace downstream from cell 4, and the elevation of the gravel horizon declines rapidly (Figure 3). Scattered gravel exposures in the high terrace and tamarisk terrace cutbanks upstream from budget cell 2 indicate that horizon elevations in those units are approximately equal to the mean bed level. A lack of terrace gravel exposures near the downstream end of the study area indicates that terrace gravel horizons are beneath mean bed level near the downstream boundary.

[17] The elevation of the top of gravel deposits in high bars and floodplains maintains a consistent height above the channel through cell 4, then decrease downstream from cell 4. These units terminate at the downstream boundary; bars downstream of the study area consist of narrow lateral benches composed of sand.

4. Creation and Analysis of a GIS Database

[18] Geomorphic maps were digitized into a geographic information system (GIS), registered to a common coordinate system, and assigned database attributes. Average channel width, areas of erosion and deposition, and volumetric changes in sediment storage for each budget cell were determined.

Table 2. Positional Errors for Individual Coverages and Uncertainties in the Measured Areas of Erosion and Deposition Polygons Produced by Spatial Union

<table>
<thead>
<tr>
<th>Photo Date</th>
<th>RMS Error, m</th>
<th>Digitizing Error, m</th>
<th>Total Linear Error, m</th>
<th>Uncompensated Overlay Error, %</th>
</tr>
</thead>
<tbody>
<tr>
<td>1948</td>
<td>17.0</td>
<td>3.6</td>
<td>17.4</td>
<td>1948–1961: 11</td>
</tr>
<tr>
<td>1961</td>
<td>4.6</td>
<td>3.6</td>
<td>5.8</td>
<td>1961–1969: 6</td>
</tr>
<tr>
<td>1997</td>
<td>4.5</td>
<td>3.6</td>
<td>5.8</td>
<td></td>
</tr>
</tbody>
</table>

*Uncertainties are expressed as a percent of the measured areas of erosion and deposition.

[19] Individual GIS coverages are subject to positional errors arising from digitizing errors, mapping errors, and photograph distortion. We assumed a mean digitizing error of 3.6 m, which corresponds to one half the width of the 0.5-mm pencil line used for mapping on our 1:14,500-scale photo enlargements. This assumption is consistent with those of Gurnell et al. [1994], who experimentally determined a magnitude for digitizing error of ~2 m for boundaries drawn on 1:10,000 scale maps, and with assumptions of Ham and Church [2000]. Errors in mapping channel bank locations were possible where the transition between low bars and high bars was gradational, but these are generally steep with relief of 0.5 m or more (Figure 2) and can be distinguished and mapped on air photos when viewed in stereo.

[20] Distortion of the original aerial photographs produces errors in coordinate registration on the digitized coverages. The magnitudes of registration errors associated with GIS coverages are usually expressed as the square root of the mean square of linear errors in the digitized positions of ground control points [Gurnell et al., 1994]. Root-mean-square (RMS) errors averaged 6.4 m for the 1961, 1969, 1980, 1987 and 1997 coverages (Table 2). The initial RMS error for the 1948 coverage was 33.9 m, because these photos are subject to significant distortion caused by aircraft tilt. Registration errors were reduced by a rubber-sheeting procedure in which a coordinate system transformation was performed to reposition the control points at their correct coordinate locations. We conservatively estimate that the procedure reduced the initial RMS error by 50%, giving an estimated RMS error for the corrected 1948 coverage of ~17 m. The total positional error for a single coverage contains contributions from all types of contributing error: digitizing error, mapping error, and registration errors. As these three sources of error are independent of one another, we have combined them as the square root of the sum of their squares [Benjamin and Cornell, 1970].

4.1. Determination of Areas of Erosion and Deposition

[21] Areas of erosion and deposition between sequential coverages were determined for all budget cells by performing a spatial union in the GIS to produce a new coverage composed of all map unit boundaries from both input coverages. The coverage produced by each spatial union contained new polygons with database attributes inherited from both input coverages, referred to as “change polygons.” Attributes of change polygons include the map unit that existed in the polygon area on the earlier photo (the
“from surface”) and the map unit that existed in the polygon area on the later photo (the “to surface”). Each “change polygon” was classified as an area of erosion, area of deposition, or an area of no change, according to its polygon area on the later photo (the “to surface”). We considered erosion to have occurred wherever a polygon changed from a high elevation category to a lower category, such as a change from cottonwood terrace to channel. We considered deposition to have occurred if the opposite trend occurred, such as a change from channel to high bar.

[22] The quality of the resulting overlay coverages depends on the accuracy with which the individual input coverages were coregistered to a common coordinate system. The relative positional error between two input coverages was estimated as the square root of the sum of squared total positional errors associated with each input coverage. Relative positional errors were propagated to uncertainty in the areas of change polygons as described below.

4.2. Determination of Volumes of Gravel Erosion and Deposition

[23] Gravel erosion ($V_{E}$) and deposition ($V_{D}$) in each budget cell $i$ are reported in terms of bulk volumes of deposited gravel (Table 3). Volumes were estimated by multiplying the area of each change polygon by the difference in gravel horizon elevation ($H$) for the from surface and the to surface. For each erosion polygon, $H$ was calculated as the difference between the mean channel bed elevation and the erosion volume of the top of the gravel deposits in the from surface. For each deposition polygon, $H$ was calculated as the difference between the elevation of the top of the gravel deposit in the to surface and the mean channel bed elevation.

[24] The sand component of the bed was not included in our analysis. Sand beneath the gravel horizon was found only in the interstitial spaces between gravel particles, and so does not contribute to the bulk volume of the bed sediments. All field evidence indicates that gravel deposits are almost entirely framework supported to the downstream boundary. Our farthest downstream subsurface sample, located within 1 km of the downstream boundary, contained 18% sand by mass. The complete transition of a gravel-sand mixture to a matrix-supported condition is associated with sand contents approaching 27% [Wilcock et al., 2001].

4.3. Errors in Estimated Gravel Erosion, Deposition and Storage Change Volumes

[25] Uncertainty in estimated changes in gravel storage is produced primarily by uncertainty in the elevations of gravel horizons in the alluvial deposits. Planimetric errors have little effect on the calculation of changes in gravel storage, regardless of the magnitude of actual channel change. For their morphology-based analysis of gravel transport on the Chilliwack River, Ham and Church [2000] assumed that planimetric errors produced by spatial overlay are compensating, and can be neglected. A local positional error that produces a false erosion polygon on one river bank also produces a false deposition polygon on the opposite bank (Figure 4). The magnitudes of total local erosion errors and local deposition errors are similar, so that net differences between the reported areas of erosion and deposition to have occurred if the opposite trend occurred, erosion to have occurred wherever a polygon changed from a high elevation category to a lower category, such as a change from cottonwood terrace to channel. We considered deposition to have occurred if the opposite trend occurred, such as a change from channel to high bar.

Table 3. Measured Volumes of Gravel Erosion, Deposition, and Storage Changes in Each Budget Cell for All Time Intervals

<table>
<thead>
<tr>
<th>Phases</th>
<th>Volume (m$^3$)</th>
<th>Erosion</th>
<th>Deposition</th>
<th>Changes in Storage</th>
</tr>
</thead>
<tbody>
<tr>
<td>1948 to 1961</td>
<td>10.2 (±2.6)</td>
<td>0.7 (±0.4)</td>
<td>9.5 (±2.2)</td>
<td>0.5 (±0.9)</td>
</tr>
<tr>
<td>1961 to 1969</td>
<td>24.3 (±8.3)</td>
<td>16.2 (±6.7)</td>
<td>19.4 (±17.1)</td>
<td>11.5 (±4.3)</td>
</tr>
<tr>
<td>1969 to 1980</td>
<td>28.2 (±7.1)</td>
<td>18.1 (±7.5)</td>
<td>18.3 (±10.1)</td>
<td>4.0 (±1.2)</td>
</tr>
<tr>
<td>1980 to 1987</td>
<td>14.3 (±17.1)</td>
<td>8.6 (±2.4)</td>
<td>13.9 (±4.6)</td>
<td>5.3 (±5.2)</td>
</tr>
<tr>
<td>1997 to 1997</td>
<td>70.1 (±18.8)</td>
<td>50.8 (±16.8)</td>
<td>31.7 (±14.7)</td>
<td>15.4 (±4.9)</td>
</tr>
</tbody>
</table>

Uncertainty margins are shown in parentheses. $V_{E}$ is erosion volume; $V_{D}$ is deposition volume; and $\Delta V$ is storage change volume.
deposition or deposition are not affected by the planimetric error. Uncertainty in the estimated volumes of erosion or deposition used to determine gravel storage changes was calculated for each erosion or deposition polygon as the product of polygon area ($A$) and the uncertainty in the gravel horizon elevation in the alluvial deposit ($dH$):

$$\delta V = A dH. \quad (1)$$

The value of $dH$ is estimated to be ±0.33 m for all floodplain and terrace map units, based on standard deviations of gravel elevations measured in the field. This value may underestimate the uncertainty in the elevation of terrace gravel in cells 1–3, where the terrace gravel horizons are below the streambed and were not observed directly (Figure 3). Terrace cores to determine the depth to gravel were not performed for logistical reasons. Instead, we assumed that the elevation of terrace gravel in cell 1 is equal to the elevation of the channel thalweg. This is based on the assumption that incision of the present channel in cells 1–3, which proceeded rapidly after the 1948 avulsion described below, slowed or stopped after gravel was encountered at depth. However, it is possible that sufficient aggradation has occurred since the cessation of incision that the channel thalweg is now significantly higher than the in situ terrace gravel in cell 1.

Uncertainty regarding gravel horizon elevations exists independently for the eroded sediments and for the deposited sediments. Both sources of uncertainty must be considered when evaluating changes in gravel storage. The total uncertainty for estimated storage changes on the scale of the budget cell is then the combined uncertainties associated with the estimated volumes of erosion and deposition:

$$\delta \Delta S_i = \sqrt{(\delta V_{Ei})^2 + (\delta V_{Di})^2}, \quad (2)$$

where $\delta \Delta S_i$ is the uncertainty in the change in storage in cell $i$, $\delta V_{Ei}$ is the volumetric uncertainty summed for all erosion polygons in cell $i$, and $\delta V_{Di}$ is the volumetric uncertainty summed for all deposition polygons in cell $i$. Values of $\delta \Delta S_i$, $\delta V_{Ei}$, and $\delta V_{Di}$ are listed as plus or minus uncertainty margins in Table 3.

For some purposes, it is useful to report gross erosion and deposition volumes, rather than net differences between erosion and deposition. The assumption of complete planimetric error compensation adopted for the reporting of net storage changes is inappropriate for reporting gross volume changes. Compensation of the planimetric errors that affect gross measurements requires that real channel change occurred during the time interval between photo dates, and the magnitude of the compensation is a nonlinear function of the magnitude of real change (Figure 4). Volumetric uncertainties reported for gross erosion or deposition estimates must therefore incorporate any uncompensated planimetric error. For any individual polygon, the volumetric uncertainty reported for gross measurements ($\delta V_g$) is estimated as the square root of the sum of squared contributing errors and uncertainties:

$$\delta V_g = \sqrt{\left(\varepsilon_{un}H\right)^2 + (A \delta H)^2}, \quad (3)$$

where $\varepsilon_{un}$ is the uncompensated error in polygon area and the other variables are as previously defined. The uncertainty in the gross erosion (or deposition) reported for each budget cell was calculated as the sum of $\delta V_g$ for all erosion (or deposition) polygons in the cell.

We evaluated the magnitude of $\varepsilon_{un}$ using computer simulations to generate independent linear displacement of two sinusoidal curves about a stationary curve (D. Gaeuman et al., unpublished analysis, 2003). The displacement of one curve represented a relative linear error with respect to the stationary curve, and displacement of the other curve represented a real change in channel position from the position of the stationary curve. The area between the curve representing real change and the curve representing planimetric error quantifies the net uncompensated error in polygon areas generated by one stream bank. We found that the probable values of $\varepsilon_{un}$ ranged between ~6% and 13% for the five overlays used in this study (Table 2).

4.4. Determination of Channel Width

Average channel width within a cell was calculated by dividing the channel area by the channel length. Because these area measurements do not require the accurate coregistration of different coverages, the effects of the linear planimetric errors are relatively modest. Errors in polygon area caused by photograph distortion have been empirically determined to be much less than 3% except when mapping near the edges of photographs where distortion is most exaggerated [Van Steeter and Pitlick, 1998]. Van Steeter and Pitlick [1998] concluded that all sources of error produce no more
than an 8% error in the average channel widths measured from historical air photos.

5. Gravel Budget Calculation Using the Storage Change Approach

[30] A morphology-based gravel budget for the study area was developed using a storage change approach and an assumed zero-transport boundary at downstream end of the study area. The storage change approach involves the construction of a bed material budget directly from measured changes in the volume of bed material stored within a series of budget cells. For each budget cell during any specified time interval:

\[ I_i - E_i = \Delta S_i, \]

where \( I_i \) is the bed material volume transported into budget cell \( i \), \( E_i \) is the bed material volume transported out of budget cell \( i \), and \( \Delta S_i \) is change in bed material storage within budget cell \( i \) (Figure 5). \( \Delta S_i \) is determined from morphologic data as the difference between the estimated erosion and deposition volumes that occurred within cell \( i \) during a given time interval (Table 3). Bed material fluxes at each cell boundary can be derived from these storage changes, provided that an input or output flux (\( I \) or \( E \)) is known or estimated for at least one budget cell in the study area. This known flux represents a boundary condition from which the fluxes at all other cell boundaries can be determined by applying equation (4).

[31] In some cases, a boundary condition of zero gravel transport can be assumed. Ham and Church [2000] assumed an upstream zero-transport boundary where their study reach exits a lake, whereas Sutherland et al. [2002] defined a zero-transport boundary at a debris dam located in the middle of their study reach. Downstream boundaries of zero gravel transport were used on the gravel bedded Fraser River [McLean and Church, 1999] and the Vedder River [Martin and Church, 1995] at points where the gravel bed changes to sand.

[32] We have assumed a downstream boundary condition of zero gravel transport for our study reach of the Duchesne River. Gravel bars abruptly disappear and the channel slope rapidly decreases by an order of magnitude downstream from cell 1. The channel itself narrows to \( \sim 35 \) m, and the bed material transitions to sand over a distance of \( \sim 1 \) km (Figure 1). The narrowness of the channel and its entrenched position \( 3-4 \) m below the adjacent terraces indicate that the reach downstream from cell 1 has not stored significant quantities of gravel in the recent past. Gravel is found downstream from the gravel-sand transition area only where the river channel locally erodes into the higher gravel-capped benches. Although it has been shown that gravel transport rates can increase by orders of magnitude as the bed sand content increases [Wilcock et al., 2001], significant gravel throughput to the Green River is unlikely. The Green River is sand-bedded downstream from its confluence with the Duchesne River, and remains so for a distance of more than 50 km.

[33] Particle abrasion is unlikely to account for significant losses in gravel or introduce significant error into the budget. Assuming a diameter diminution coefficient of 0.0017 km\(^{-1}\) for quartzite [Shaw and Kellerhals, 1982] and an average transport distance of 8 km (one half the total channel length), we calculated that a clast 45 mm in diameter would be reduced in diameter by \( \sim 0.6 \) mm, and reduced in volume by \( \sim 4\% \). The diminution coefficient used in this calculation was obtained from measurements of downstream fining in rivers, and so may reflect the effects of particle sorting as well as abrasion. Diminution coefficients obtained from laboratory studies are generally much smaller.

6. Gravel Routing Using the Travel Distance Approach

[34] The travel distance approach was used as a complement to the storage change approach in order to estimate the proportion of sediment deposited on the bed in each cell. This approach is based on the specification of a characteristic distance that eroded bed material is transported downstream during a time interval [Lane and Richards, 1995; Ashmore and Church, 1998]. In general, volumes of bed material eroded at individual source areas along the stream channel are calculated from changes in channel morphology and routed downstream using an assumed transport dis-
tance, or “path length” (Figure 5). Local bed material fluxes can be determined by summing the total volume of material from all upstream source areas passing defined boundaries, and local changes in storage can be back-calculated from the boundary fluxes using equation (4) (Figure 5). This approach was pioneered by Neill [1971], who assumed the characteristic transport path length to be one half the meander wavelength for regularly meandering rivers. This method has the advantages that no information regarding boundary transport conditions is needed and only erosion volumes need be measured. Erosion is usually a more spatially focused process than is deposition, and so is more readily quantified through morphological measurements [Ashmore and Church, 1998].

[15] Application of the travel distance approach necessitates selection of an appropriate transport path length and application of some type of sediment routing procedure. Methods for estimating transport path lengths in complex channels are poorly developed. Most estimates have been scaled as a function of channel width, or related to the spacing of other channel elements whose scale depends on channel width [Goff and Ashmore, 1994; McLean and Church, 1999; Beechey, 2001]. Path lengths may depend on the frequency and magnitude of floods during a time interval, such that estimates based on excess stream power have also been proposed [Hassan et al., 1992].

[16] Although sediment budgets developed using the travel distance approach have been published over the past decade [e.g., Lane and Richards, 1995; Eaton and Lapointe, 2001], detailed description of the methods used to route eroded gravel to the appropriate deposition areas are generally absent. Here, we describe a simple and explicit gravel routing procedure for estimating the total volume of gravel deposition in each budget cell during each time period. Objective criteria for selecting appropriate path lengths are also described.

[17] Rather than assuming all gravel eroded at a given location moves the same distance as a coherent slug, we propose that the average transport distance (L) corresponds to the mean value of an exponential probability density function. The choice of an exponential distribution was made for simplicity and because it has been reported to fit downstream distributions of tracer particles observed in the field [Mosley, 1978]. Although a handful of studies have explored the applicability of various probability distributions to gravel transport [Einstein, 1937; Hubbell and Sayre, 1964; Kondolf and Matthews, 1986; Hassan et al., 1992; Hassan and Church, 1992], no single distribution describing particle path lengths has been definitively identified [Pycke and Ashmore, 2003]. No studies have considered the problem at the scale of multiple meander wavelengths relevant to this study.

[18] Bed load transport in a natural stream is widely regarded as being mediated by channel morphology [Schmidt and Ergenzinger, 1992; Ashmore and Church, 1998; McEwan et al., 2001], i.e., coarse particles are preferentially deposited on riffles or other bed forms characteristic of channel architecture. Our routing model explicitly recognizes this morphological control by expressing L in units of half meander wavelengths. Casting L in terms of simple distance would be unnecessarily indirect, and would require assumptions regarding the spacing of channel elements and its spatial variability. The number of half-meander units in each budget cell is given by the variable uj, where the subscripts refer to the budget cell that is j cells downstream from cell i (Figure 6). Thus u0 indicates budget cell i, and u1 indicates the cell immediately downstream from cell i. The cumulative number of half-meander channel units downstream from the midpoint of any cell i is

\[ n_j = u_j / 2 + \sum_{j=1}^{i-1} u_j. \]

The volume of gravel eroded within a time interval in budget cell i is given by \( P_i \). The fraction of \( P_i \) deposited in each cell (fj) is determined according to an exponential decay curve with its origin at the center of cell i, and its tail extending downstream (Figure 6). For each cell upstream from cell 3, fj is calculated by integrating the exponential function over the appropriate interval:

\[ f_j = \frac{1}{L} \int_{n_j}^{n_0} e^{-x/L} dx, \quad 0 < j < (i-3). \]

We defined cells 1 through 3 as a zone of gravel accumulation. Early aerial photographs show that virtually all gravel visible in cells 1 through 3 has accumulated there since about 1948, when an avulsion initiated the development of a new channel with large gravel bars (Figure 7). All gravel not assigned to cells 4 through 15 by equations (6) and (7) was assigned to this accumulation zone as the fraction of each \( P_i \) equal to 1 minus the sum of all upstream fractions (Figure 6).
It should be noted that the subscripts $i$ and $j$ increment in opposite directions. The subscript $j$ denotes the cell that is $j$ cells downstream from cell $i$, and thus increases in the downstream direction, whereas $i$ corresponds to the upstream numbering scheme for budget cells shown in Figure 1. The total gravel deposition in budget cell $i$ ($G_i$), which is equal to the sum of all fractions of upstream $P_i$ routed to cell $i$, is therefore given by

\[ G_i = \sum_{c=0}^{k-i} P_{i+c}f_{i+c}, \]  

where $c$ signifies the number of cells upstream from cell $i$ where gravel fractions deposited in cell $i$ originated.

Results from the gravel routing model allow the volume of bed deposition within individual budget cells to be estimated by subtracting the volume of deposition measured on air photos from the total deposition volume predicted by the model. When divided by the channel area in each cell, these volumes of bed deposition yield the average change in bed elevation for each cell during each time interval. Gravel volumes are routed only during the time periods in which they were eroded. The model contains no provisions for remobilizing and routing gravel eroded from the banks and deposited on the channel bed during earlier time periods.

The value of $L$ used for each time interval was selected according to two criteria. First, $L$ was constrained so that, in each time period, the routing procedure delivered a volume of gravel to the accumulation zone equal to the volume increase in gravel storage in cells 1 through 3, as measured directly on the air photos. The volume of gravel routed to the accumulation zone was allowed to range within the uncertainty margins estimated for the storage changes, which are ~100% of the reported storage changes for all time periods (Table 3). In time periods with low streamflows and small gravel transport rates, a large value of $L$ delivers too much gravel to the accumulation zone compared with the increases in storage measured on the air photos. This criterion effectively places an upper limit on the value of $L$ that can reasonably be assumed during any period.

Second, $L$ was chosen to optimize the correlation between calculated changes in cell bed elevations and measured changes in cell channel widths for each period. This criterion was based on the observation that the wider locations along the channel in the gravel bed portion of the study area were associated with sites of local aggradation, such as large active riffles and shoals. Channel widening in any cell is, by definition, caused by net bank erosion, which delivers some volume of gravel to the active channel. If some portion of the introduced volume of gravel fails to move downstream, bed aggradation will have occurred in the cell where the widening occurred. If the introduced gravel is transported downstream, bed aggradation takes place downstream. This aggradation presumably reduces channel capacity and triggers bank erosion at the deposition site [Leopold and Maddock, 1953]. Local channel widening is therefore assumed to indicate either the local production
of or deposition of a gravel slug in the active channel, and in many cases, both.

7. Results and Discussion

7.1. Flood Discharge and Gravel Activity

Streamflow for the periods between air photos is summarized in terms of mean annual flood discharge. Mean annual flood discharge is defined in this study as the cumulative flow volume in excess of the bank-full discharge, and is expressed in units of m$^3$ s$^{-1}$ days per year. By far the wettest interval in terms of flood discharge spanned water years (WYs) 1981 through 1987 (Figure 8). This interval included the flood of record (326 m$^3$ s$^{-1}$ on 20 June 1983) and the highest annual flow total on record. WYs 1962 through 1969 also had a relatively high mean annual flood discharge, and included the second largest flood on record (292 m$^3$ s$^{-1}$ on 13 June 1965). Flood discharge volumes were small in WYs 1970 through 1980 and WYs 1988 through 1997. WYs 1968 through 1997 were especially dry, and included the driest 7-year period on record from 1988 through 1994. However, the period also included two relatively large, but brief high magnitude events in 1995 and 1997. These floods peaked at 203 and 152 m$^3$ s$^{-1}$ and have recurrence intervals of $\sim$5.9 years and $\sim$3.2 years, respectively.

We defined an index of total gravel activity as the total gravel erosion and deposition volumes during each time interval divided by the total channel length and the number of years in the time interval. A similar index has previously been suggested for comparing rates of geomorphic activity between different time periods [Ham and Church, 2000]. Temporal variation in the total gravel activity correlates well with mean annual flood discharge (Figure 8). The differences in activity between the two wet periods and the two dry periods exceed the margins of uncertainty.

7.2. Losses in Gravel Storage Between 1948 and 1969

Decreases in measured gravel storage indicate that significant bed aggradation occurred during two of the five time intervals evaluated for this study (Table 3 and Figure 9). Large gravel storage losses were calculated for most budget cells during 1948–1961 and 1961–1969, and the net losses for the study area exceed the uncertainties associated with all measurement errors. The greatest losses occurred between 1961 and 1969 and exceeded 125,000 m$^3$, a quantity of gravel equivalent to a bar 1 m high and 8 channel widths in length and width.

Although local occurrences of storage losses within a stream reach are physically reasonable, a cumulative loss in storage summed upstream from a downstream boundary of zero transport is not. When local gravel storage changes within individual budget cells are summed upstream from a downstream zero-transport boundary, the cumulative storage change at each local budget cell boundary represents the gravel flux across that cell boundary (Figure 9). If the cumulative storage losses exceed the cumulative storage gains, the calculated gravel fluxes at cell boundaries will be negative. Such a result implies that the assumption of zero gravel transport at the downstream boundary is violated, or that undetected aggradation has occurred. Martin and Church [1995] calculated negative gravel fluxes for some locations on Vedder River and concluded that the zero-transport assumption at the downstream boundary was violated in at least some years. They therefore applied a nonnegative criterion to the calculated bed load fluxes, wherein the most negative flux for a given time interval was increased to equal zero, and fluxes at all other cross sections for the same time interval were adjusted accordingly.

We propose a different interpretation for the negative gravel fluxes calculated in our study area. Our data suggest that gravel losses measured on the Duchesne River between 1948 and 1969 reflect gravel deposition within the active channel and widespread aggradation of the streambed.
Temporal changes in channel widths indicate that the measured storage losses are closely linked to changes in channel geometry. Increases in mean channel width in cells 5 through 15 between 1948 and 1969 show that bank erosion was a dominant process at the time, and support an interpretation that gravel stored in terrace and floodplain units was transferred to storage on the bed of the active channel. Average channel width in budget cells 5 through 15 increased by 11% between 1948 and 1961, and by 1969 had increased by 37% over the 1948 value (Figure 10). A narrowing trend began in these cells after 1969, and continues to the present. Narrowing during the three most recent time periods coincided with total study area changes in gravel storage that were positive or near zero, suggesting a link between channel narrowing and decreased bed deposition.

[48] It is possible that a portion of the measured losses could be attributed to undetected aggradation in the gravel accumulation zone near the downstream boundary. Significant aggradation in cells 1 and 2 would have escaped detection if the channel that formed after the 1948 avulsion had scoured to a greater depth than the present thalweg. Our budget measurements do not include any gravel deposited below the elevation of the present thalweg. Assuming that the terrace gravel horizon is actually 1 m below the channel thalweg throughout all of cells 1 and 2, the amount of gravel storage in cells 1 and 2 between 1948 and 1969 would have been \( \approx 101,500 \text{ m}^3 \), rather than our estimate of \( \approx 43,000 \text{ m}^3 \). Although this would constitute a local increase in gravel storage of \( \approx 135\% \), it could account for only \( \approx 65\% \) of the total storage losses measured between 1948 and 1969. Such large gravel depths are inconsistent with field observations of terrace gravel well above the thalweg elevation at least as far downstream as cell 2.

[49] A large depth of fill near the downstream boundary is an unlikely explanation for the measured gravel losses on the grounds of the required transport distances as well. In both periods between 1948 and 1969, gravel storage losses were measured in every cell upstream from cell 4 (Table 3 and Figure 9). Consistent losses over consecutive cells are likely to be maintained only if the erosion volumes measured in successive cells consistently increase downstream, or the bulk of the gravel eroded in the upstream cells passes through the downstream cells without being deposited. The first possibility is inconsistent with the data (Table 3), and the second implies that gravel eroded far upstream in cells 9 through 15 was transported to the downstream end of the gravel accumulation zone or even out of the study area. This would require that gravel eroded in the upstream cells was transported distances in the range of \( \approx 200 \) to \( 400 \) channel widths during each time interval. Such large travel distances are unlikely.

[50] Consideration of the flood hydrology during the study period casts further doubt as to whether such long transport distances can account for the gravel storage losses. If losses were related to significant gravel transport to the downstream boundary, it would be reasonable to expect a positive relationship between storage loss and flood discharge. No such pattern is apparent. Positive storage changes were greatest between 1980 and 1987 when flood discharge was by far the greatest, and presumably had the greatest potential to transport gravel past the downstream boundary. Large losses in storage were recorded between 1948 and 1961, at a time when mean annual flood flows were little more than a third those of the wettest period.

7.3. Bed Aggradation Predicted by the Travel Distance Approach

[51] Results from the travel distance analysis indicate that bed aggradation between 1948 and 1969 was greatest in the middle part of the study area, particularly in cells 8 through 11. Comparison with historical channel adjustments in those cells indicates that this persistent bed aggradation can be linked to subsequent channel instability.

[52] Values for the average gravel transport distances \((L)\) for each time interval were selected according to criteria that include the strength of the correlation between predicted changes in cell bed elevations and measured changes in cell channel widths. These correlations were used to provide objective guidance in the selection of gravel transport distances for the various intervals, and should not be interpreted as statistically meaningful tests of model performance. The strongest correlation \((R^2 = 0.76)\) was obtained for 1961–1969 (Figure 11). This period was characterized by channel widening, large losses of gravel storage, and implied bed aggradation. A moderately strong correlation was also determined for 1948–1961, which was also a time of channel widening and storage losses. The correlations were weakest for 1980–1987 and 1987–1997, when average channel widths for most budget cells decreased and positive cumulative gravel fluxes at the upstream boundary indicate that bed aggradation was minor or absent. These results are consistent with the hypothesis that local channel widening is linked to local bed aggradation, but only during the periods in which bed aggradation occurs. The strength of the correlation for 1969–1980 \((R^2 = 0.69)\) is related to the very small flood discharges (Figure 8) and corresponding low gravel fluxes \((L = 0.4)\) during that period, such that nearly all eroded gravel was deposited locally in the same budget cell where it originated.
gravel erosion reflects bank erosion, very low transport distances necessarily result in deposition in locations where the channel is most likely to have widened, or to have narrowed least.

The values of $L$ obtained for the five time periods show $L$ to generally increase with increasing total flood discharge (Figure 12a). The larger value of $L$ for 1961–1969 compared with the value obtained for 1948–1961 is reasonable, even though a larger total volume of flood discharge was recorded during the earlier period. Total flood discharge shown in Figure 12a was summed over 13 years (WYs 1949–1961), whereas total flood discharge for 1961–1969 was summed over just 8 years (WYs 1962–1969). However, the mean annual flood discharge was greater for WYs 1962–1969 than for WYs 1949–1961. When plotted against mean annual flood discharge, the plotting order for the 1948–1961 and the 1961–1969 data are reversed and a strong log-log relationship results (Figure 12b). Although both hydrologic measures incorporate flood magnitude and duration, mean annual flood discharge is a better measure of the intensity of individual flood events than is total flood discharge. It has previously been suggested that flood magnitude is a better predictor of particle travel distance than flood duration, and that the form of the magnitude-distance relationship is log linear [Hassan et al., 1992].

The value for $L$ of 4.0 obtained for 1980–1987 may be anomalously low. As noted above, the correlation

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**Figure 11.** Plots of changes in channel widths versus modeled bed aggradation in budget cells 5 through 15 for five time intervals, with linear regression lines. Values of $L$ used to predict bed aggradation for each interval are indicated.
between channel widening and predicted bed aggradation used as a selection criterion was very poor for this time interval, leaving gravel delivery to the accumulation zone as the only meaningful criterion by which to select $L$. For $L = 4.0$, the predicted increase in accumulation zone storage nearly matched the measured increase and is well within the uncertainty margins for the measurements. $L$ could be increased to as much as 5.5 without exceeding the range of uncertainty around the measured storage increase. Such an increase in $L$ would virtually eliminate the discrepancy between $L$ for this period and the remainder of the data, as indicated on Figure 12. However, our original selection criteria provided no impetus for selecting the higher value.

When parameterized with the $L$ values, the gravel routing model projects that bed aggradation was greatest in the central part of the study area during both 1948–1961 and 1961–1969 (Figure 13). The locations of peak aggradation were in cell 9 in 1948–1961 and in cell 11 in 1961–1969. Aggradation between 1948 and 1980 is projected to have increased steadily downstream from the upstream boundary to cells 9 and 10 (Figure 13). In this region, the model indicates that gross aggradation reached $\sim 55$ cm, or more than 25% of the bank-full channel depth. As the model makes no allowance for remobilization of gravel deposited on the bed during earlier time periods, the actual net increase in bed elevation was probably less than the gross figure. The model also indicates that gross aggradation between 1948 and 1980 decreased rapidly downstream from cell 8, such that bed degradation is indicated in cell 7.

Moderate levels of aggradation are indicated farther downstream in cells 5 and 6.

The validity of the locations of significant bed aggradation can be evaluated by comparison with channel adjustments between 1980 and 1987, when several large floods occurred. Large-scale channel adjustments were concentrated in cells 8 through 10, in and immediately downstream from the cells of maximum predicted bed aggradation during the preceding years (Figure 14). Gravel activity measured for the period between 1980 and 1987 also mimics the pattern of prior bed aggradation (Figure 15). Gravel activity was highest in budget cells 8 through 10, reaching its largest value in cell 9 where the total bed aggradation between 1948 and 1969 was also greatest.

### 7.4. Integration of the Storage Change and Travel Distance Approaches

The storage change and travel distance approaches to constructing sediment budgets are based on different assumptions and slightly different measurements, and therefore have different strengths and limitations. We have mitigated the deficiencies associated with each of these general approaches with a hybrid approach that incorporates elements of both.

Gravel budgets developed using the storage change approach are sensitive to uncertainty in the elevations of gravel deposit horizons in the various map units. Changes in gravel storage are calculated as the difference between erosion and deposition volumes, and are often small compared with the gross volumes of erosion and deposition. However, uncertainties in net storage change estimates are based on the measurement errors associated with the much larger gross volumes. Error margins for storage change measurements are thus large on a percent basis, often exceeding 100%. More importantly, adjustments in bed elevation can produce large errors that are virtually impossible to detect on air photos. We have shown that unmeasured bed deposition can result in impossible gravel budgets that imply negative gravel transport rates.

![Figure 12](image1.png)

**Figure 12.** Relationship between estimated gravel path lengths ($L$) for each time interval and (a) total flood discharge (volume of flow exceeding the channel-forming threshold) and (b) mean annual flood discharge. The plotting positions for an alternative estimate for $L$ for 1980–1987 ($L = 5.5$) is shown by the open triangle.

![Figure 13](image2.png)

**Figure 13.** Calculated gross bed aggradation before 1980 in budget cells 5 through 15, as determined from the gravel routing algorithm. Error margins for total aggradation between 1948 and 1980 are the sums of the uncertainties in bed elevation changes resulting from uncertainty in the input volumes of erosion and deposition for each of the three preceding periods. The error margins shown do not include model error or error resulting from gravel transport into the study area from upstream.
The travel distance approach to building a morphology-based sediment budget requires the quantification of erosion volumes only, such that the identification of deposition sites is unnecessary. However, this approach requires a method for determining the characteristic distances that gravel is transported during specific time intervals. The basic understanding of reach-scale sediment transport necessary to reliably predict bed load transport distances in natural channels remains undeveloped, so that sediment budgets developed using travel distance estimates also incorporate a high degree of uncertainty.

Our combined approach uses estimated changes in gravel storage in the accumulation zone during each time period to calibrate a procedure for determining the spatial distribution of bed deposition throughout the study area. This procedure includes a gravel routing model based on travel distance concepts, and incorporates several assumptions whose validity has not been demonstrated. It is assumed that the transport distances of gravel particles can be represented by an exponential probability density function specified by a spatially constant mean path length \( L \). It is further assumed that local bed aggradation is spatially correlated with local channel widening. The model lacks provisions to remobilize gravel predicted to have been deposited in the active channel during earlier time intervals, and thus yields a gross, rather than net, estimate of bed aggradation. Refinement of the current routing model and a better understanding of its theoretical basis are needed. Nonetheless, this model is, to our knowledge, the only fully specified implementation of the travel distance approach to appear in the literature. Improved spatial resolution, in the form of smaller budget cells, or an alternative travel distance distribution function could be readily integrated into the existing model structure.

In spite of its simplicity, the longitudinal pattern of relative bed aggradation predicted by the model is consistent with observations of subsequent channel behavior. Areas in which our analysis indicated substantial bed aggradation had occurred by 1980 displayed higher levels of gravel activity and channel instability during subsequent time periods than areas where less bed aggradation was predicted. We were thus able to extract information on the

**Figure 14.** Channel changes in budget cells 8 through 10 between 1980 and 1987.

**Figure 15.** Total gravel activity in budget cells 5 through 15 between 1980 and 1987. Activity after 1980 was highest in and just downstream from cells where gravel routing predicted the greatest bed aggradation before 1980.
spatial distribution of past adjustments in bed elevation from planimetric data only.

7.5. General Implications of the Integrated Results

The integrated gravel budgets presented above show that two contrasting modes of adjustment in channel geometry have occurred in the lower Duchesne River since 1948. Between 1948 and 1969, adjustments were dominated by bank erosion and bed aggradation. This mode of adjustment prevailed during periods in which mean annual flood discharges were moderate. The trajectory of channel adjustment reversed after 1980, when a series of extreme peak events, including the 1983 flood of record, coincided with a phase of channel narrowing that continued until at least 1997. Gravel budget results indicate that this narrowing was associated with the cessation of bed aggradation, and possible reincision of the streambed. The shift from the widening/aggradation mode of adjustment to the narrowing/incision mode may have been related to the capacity of these large floods to fully mobilize the bed surface and break up any armor layer that may have existed, as has been suggested for a similar shift in the Kemano River of western Canada [Church, 1995]. Although periods of channel widening and channel narrowing can be readily identified from air photos alone, our integrated gravel budget provided perhaps the only means for inferring the vertical component of the adjustments and a mechanistic explanation for the shift in adjustment style in the Duchesne River.

The ability to determine bed load transport path lengths would constitute an important step toward characterizing stream dynamics at the segment-scale, and our integrated gravel budget analysis provides new data to that end. Path lengths emerging from our analyses are comparable to the path lengths determined by tracer studies, and may therefore be more realistic than path lengths estimated by other indirect methods. When converted to units of channel widths (~24 channel widths per meander wavelength) and divided by the number of years in each time period, mean gravel path lengths estimated for the Duchesne River ranged between 0.4 and 6.9 channel widths per year. If \( L \) for 1980–1987 is assumed equal to 5.5, as discussed above, path lengths of nearly 10 widths per year result. By comparison, most fractions of painted and magnetically tagged tracer particles relocated after an 8-year period in Allt Dubhaig traveled a total of between 10 to 40 channel widths [Ferguson et al., 2002, Figure 3]. This corresponds with annual distances of ~1 to 5 widths per year. Results of several published tracer studies summarized by Beechie [2001] yielded annual path lengths that were typically near 10 widths per year. However, path length estimates based on several indirect methods, including tracking the translation rate of maximum channel width or the midpoint of sediment waves, imply considerably longer path lengths, often exceeding 20 channel widths per year [Beechie, 2001].

8. Conclusions

Historical gravel budgets calculated from gravel storage changes measured on air photos can contain significant errors caused by changes in the elevation of the streambed. Gravel deposition in submerged areas and dif fuse deposition resulting in a general increase in bed elevation are difficult or impossible to detect directly. Failure to include bed aggradation in a storage change budget can produce impossible results, such as negative gravel transport rates when missing gravel volumes are accumulated upstream from a known transport boundary.

Difficulties in measuring gravel deposition volumes can be largely avoided by use of a travel distance approach for constructing gravel budgets. Application of this approach requires the selection of appropriate gravel transport path lengths and some type of gravel routing procedure. We have presented a fully specified algorithm for distributing eroded gravel downstream according to a probability density function specified by a single parameter, the mean transport path length.

Uncertainty in estimating mean transport path lengths was reduced by integrating storage change results and measured changes in channel width into the travel distance analysis. The mean path length for each time interval was constrained so that the volume of gravel routed to a zone of gravel accumulation near the downstream end of the study area was equal to the measured increase in local gravel storage. Path lengths were further adjusted to improve correlations between local bed aggradation predicted by the gravel routing algorithm and local channel widening measured on air photos. Mean gravel transport distances in the lower Duchesne River were found to average ~7 channel widths per year during wet periods and ~1 channel width per year or less during drought periods.

Results of this integrated morphological analysis indicate that a period of channel widening lasting about 20 years was accompanied by persistent streambed aggradation. Maximum aggradation was predicted in a portion of the study area that later underwent significant channel reconstructions, including multiple meander cutoffs and rapid bend extension. These reconstructions, which occurred during a period of large flood events, marked a transition in the style of channel adjustment from one of widening and aggradation to one of narrowing and possible incision.

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References


