High Flow Requirements for Channel and Habitat Maintenance on the Lower Duchesne River between Randlett and Ouray, Utah

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EXECUTIVE SUMMARY

The objectives of this study were to (1) characterize the geomorphic attributes of the present channel and alluvial valley of the lower Duchesne River between the mouth of the Uinta River and the Green River; (2) evaluate the flows, sediment transport, and channel processes that formed and maintained the present channel/valley system; (3) evaluate the degree to which the present channel/valley system is in a state of dynamic equilibrium with recent stream flows; and (4) determine the discharges necessary to insure that existing geomorphic and habitat conditions can be maintained in the future.

Our strategy for meeting these objectives employed an historical analysis of geomorphic change in the study area, in addition to more traditional methods of characterizing modern fluvial geomorphic processes. Analysis of historical aerial photography in a geographic information system (GIS) was used to quantify past channel responses, and the record of channel change was compared to changes in water and sediment discharge. This comparison allowed us to determine the role of decade-scale periods of high and low flow in forming and maintaining the channel and valley of the Duchesne River. The form and functioning of the present river was quantified through extensive field measurements of channel and valley characteristics, and through the use of computational methods. Main findings from this work are:

1. The lower Duchesne River consists of four distinct zones with differing morphologies and histories. Channel morphology and response to flow varies in time and between zones.
2. Channel-forming discharge on the lower Duchesne River is about 4,000 ft^3/s. Gravel mobilization and inundation of high bar surfaces occur at this discharge.
3. Little channel activity occurs during periods when the volume of stream flow in excess of the channel-forming discharge of 4,000 ft^3/s is less than 7,000 ft^3/s-days per year. Physical habitat is created and maintained during decades when the volume of stream flow in excess of 4,000 ft^3/s is greater than 7,000 ft^3/s-days per year. A channel-forming flow regime to maintain channel activity at rates sufficient to support all habitat components is proposed.
4. The recurrence of daily mean discharges of 4,000 ft^3/s is about 2.2 years for the period from 1943 through 1971. The recurrence of this discharge has increased to 3 years since completion of the Bonneville Unit of the CUP in water year 1972.

5. The increase the recurrence period for daily mean discharges of 4,000 ft^3/s or greater since 1971 has contributed to a consistent trend of channel narrowing since 1969.
6. Fine sediment accumulation related to a 50-percent reduction in stream flow after the 1920s and an increase in the local sediment supply resulted in significant channel narrowing, the loss of side channel habitat, and large-scale avulsions on the lower Duchesne River.
7. Modest flushing flows to prevent further accumulation of fine sediment in the lower Duchesne River are proposed in addition to the channel-forming flow regime.
8. Existing measurements of suspended sediment concentrations in the lower Duchesne River are inadequate for making well-constrained estimates of suspended sediment loads during high discharge periods. An extended sampling program to monitor suspended sediment concentrations in the lower Duchesne River during peak flow events should be undertaken.

Past and Present Characteristic of the Study Area

The channel of the lower Duchesne River has a meandering planform, and a mixed bed of cobbles, gravel, and sand upstream from about river km 9. The river has historically been active through much of this gravel-bed portion, and remains so to the present day. The character of the Duchesne River changes abruptly downstream from river km 9. Channel gradient flattens, and the channel assumes a deep, narrow geometry. Bed material changes from gravel to sand, and the channel becomes fully sand-bedded by river km 7. The pool-riffle channel morphology with wide point bars and a complex shoreline found in the upstream part of the study area is replaced by a simple canal-like channel with steep, well-vegetated banks.

The 20th century geomorphic history of the lower Duchesne River includes complex adjustments to changes in both sediment supply and water discharge. The nature of the adjustments has varied both spatially and temporally over a period of at least 65 years, and continues to influence river morphology to the present day. This history can be condensed into a few periods of consistent trends and processes. These are 1) channel narrowing, filling of side channels, and avulsions before 1950, 2) channel metamorphosis involving extreme widening of a short reach downstream from the Pipeline between 1948 and 1987, 3) bend extension with frequent chute cutoffs throughout the middle part of the study area, and 4) relative stability in the upstream part of the study area.

Changes in Stream Flow in the Study Area

Total annual runoff and flood magnitudes have been highly variable over the period of record. The years with total annual runoff in the upper quartile of the measured record averaged 5.8 times the mean annual flow of the years with total annual runoff in the lowest quartile of the distribution. The magnitudes of 1.5-year, 2-year, and 5-year floods, and the flows exceeded 10, 50, and 90 percent of the time are all at least five times greater in upper-quartile years than in lower-quartile years. Wetter and drier years appear to cluster in the latter half of the 20th century, when three relatively dry periods lasting roughly a decade were separated by two relatively wet periods, each lasting about 5 years.

Mean annual runoff for water years 1972 through 2000, after the Bonneville Unit of the Central Utah Project (CUP) began operations, declined by approximately 10 percent from the mean annual runoff for the pre-project water years of 1943 through 1971. The impact of the CUP is most pronounced during years with total annual flow in the lowest quartile of the record. For these years, mean annual flow is 33 percent less since 1972 than for before 1972. The frequencies and magnitudes of large discharges decreased less than the frequencies and magnitudes of moderate discharges. The magnitudes of 1.5-year and 2-year floods and of the flow exceeded 50 percent of the time decreased by more than 25 percent after 1972, while the magnitude of the 5-year flood decreased by only about 5 percent.

A significant decline in stream flow through the study area occurred in the mid-1920s, as determined from statistical extension of the hydrologic record. The total annual flow, magnitude of floods, and the magnitude of flow of various durations all were much greater for the period between 1912 and 1924 than for the period between 1925 and 2000.

Discharges Necessary to Access High Bars and Secondary Channels

One-dimensional hydraulic models developed at three detailed study sites indicate that floods ranging from the 2-year to the 2.6-year events are required to initiate significant flow onto high bar surfaces. The average discharge that inundates these surfaces is 4,000 ft^3/s. Somewhat smaller magnitude events with discharges of approximately 3,000 ft^3/s and a recurrence interval of 1.7 years are sufficient to produce flow into the main chute channels. Flows capable of inundating the floodplains and higher bar ridges are larger than the 3.2-year flood and may be approximated by the 6-year event.

Discharges Necessary to Entrain Gravel

The discharges necessary to entrain gravel were estimated by comparing values of average boundary shear stress derived from our hydraulic models with the critical shear stress needed to mobilize gravel particles of the sizes found on the surface of riffles at detailed study sites. Results indicate that reach-averaged shear stress approaches the threshold of entrainment over riffles and runs at discharges of approximately 4,000 ft^3/s. These results are consistent with field observation of gravel entrainment during the spring 2001 peak, which briefly reached 2,900 ft^3/s. Little gravel entrainment was observed to occur as a result of this discharge.

Channel-forming Discharges

Our gravel entrainment and inundation analyses converge on a threshold discharge of about 4,000 ft^3/s for mobilizing a significant portion of the bed and inundating some overbank surfaces. This discharge level has a long-term daily mean recurrence interval of about 2.4 years and is exceeded about 1.6 percent of the time. Comparison of the history of channel changes and activity rates with the historic hydrology demonstrate that the level of channel activity is low when flows above this critical discharge threshold are less frequent. Two channel-forming flow hydrographs are suggested. It is proposed that a moderate channel-forming hydrograph be implemented in 30 percent of all years, and a larger channel-forming hydrograph be implemented in 10 percent of all years.

Discharges Necessary to Flush Fine Sediment

Loss of fine sediment transport capacity has historically resulted in the loss of habitat in the lower Duchesne River. Fine sediment transport capacity adequate to prevent further habitat loss can be maintained by a combination of the proposed channel-forming flow regime and additional flushing flows. The proposed flushing flows consist of 7 days with discharges greater than 3,000 ft^3/s in 30 percent of all years, and 7 days with discharges greater than 2,500 ft^3/s in and additional 30 percent of all years.

INTRODUCTION

The objectives of this study were to (1) characterize the geomorphic attributes of the present channel and alluvial valley of the lower Duchesne River between the mouth of the Uinta River and the Green River; (2) evaluate the flows, sediment transport, and channel processes that formed and maintained the present channel/valley system; (3) evaluate the degree to which the present channel/valley system is in a state of dynamic equilibrium with recent stream flows; and (4) determine the discharges necessary to insure that existing geomorphic and habitat conditions can be maintained in the future. This study complements biological studies being conducted by the Ute Tribe, U.S. Fish and Wildlife Service, and state of Utah concerning the distribution and life history of two species of endemic endangered fish, the Colorado Pikeminnow and the Razorback Sucker, that occupy the lower Duchesne River. The collective results of these studies will be used in development of a comprehensive recommendation for the minimum flows necessary to maintain the role of the lower Duchesne River as habitat for endangered fish.

The research described in this report utilizes a strategy for determining channel-maintenance flow requirements employing an historical analysis of geomorphic change in the study area, in addition to more traditional methods of characterizing modern fluvial geomorphic processes. We determine the nature of past responses to changes in water and sediment discharge on a decadal time scale, and thereby place present channel processes and conditions within an historical context.

High Flows for Channel and Habitat Maintenance

Stream channel morphology is a function of water discharge, the type and amount of sediment being transported, and the character of the materials making up the channel bed and banks. Changes in any of these variables may cause the stream channel to reconfigure its geometry or plan form so that the imposed water and sediment loads can be transported through the system (Schumm 1969). Decreases in water discharge or increases in sediment load can potentially cause channel narrowing or bed aggradation, both of which reduce the capacity of a channel to carry peak flows (Warner 1994). Large flows during infrequent wet years then have an increased potential to result in destructive flooding or trigger complex channel adjustments that can take decades to complete. It is therefore critical for channel maintenance that the sediment transport capacity through a river reach is sufficient to transport the sediment load supplied to the reach.

Loss of sediment transport capacity also impacts the quality and availability of physical habitat used by aquatic biota. Gravel substrates can become choked with fine sediment, impacting macroinvertebrate production and the availability of spawning sites for some fish species (Giller and Malmqvist 1998), while increased sediment deposition in backwaters and side channels can reduce channel complexity and habitat diversity (Van Steeter and Pitlick 1998). Flow regimes needed to maintain these elements of physical habitat must exceed certain discharge thresholds, in addition to transporting the imposed sediment load. Maintenance of a productive gravel substrate requires discharges sufficient to mobilize the stream bed so that fine sediment can be flushed from the subsurface (Petts and Maddock 1996). Maintenance of backwater and side channel habitats requires discharges capable of accessing and scouring sediment from these areas.

Numerous authors have stressed the importance of the natural flow regime, including relatively large floods, for maintaining stream and riparian health (Hill et al. 1991; Richter et al. 1997; Poff et al. 1997). Higher flows are necessary to maintain channel and flood plain form and function (Kondolf and Wilcock 1996; Stanford 1996; Milhous 1998; Trush et al. 2000). Discharges adequate to cause bank erosion and re-distribute bed sediment are needed to create disturbance and maintain the structural elements that provide the basis for diversity in aquatic and riparian habitat (Stanford et al. 1996). The ongoing deformation of channel pattern that accompanies active meander migration may lead to the formation of meander cutoffs, i.e., the main flow cuts a new path that bypasses part or all of a meander bend (Lewis and Lewin 1983). Growing point bars may develop chutes and swales through a variety of processes, some of which are discussed by Bridge et al. (1986), Hooke (1986), and Howard (1996). The resulting islands, backwaters, side channels, and oxbow lakes comprise the elements of a complex aquatic habitat.

Such physical complexity has been shown to be important factor for habitat use among Colorado pikeminnow (*Ptychocheilus lucius*), an endangered endemic species found throughout the lower Duchesne River. Osmundson (2001) reported that 84 percent of adult pikeminnow located by radiotelemetry during spring runoff in the upper Colorado River were observed in protected off-channel areas. Backwater habitats were also observed to serve as nursery habitat for young pikeminnow (Osmundson and Kaeding 1991; Osmundson 2001). Loss of off-channel habitat has had a detrimental effect on razorback sucker (*Xyrauchen texanus*), another endangered native fish species known to use the lower Duchesne River (Osmundson 2001).

Off-channel water bodies are also frequently sites of enhanced primary production, which is then exported to the main channel (Eckblad et al. 1984; Saunders and Lewis 1988; Tockner et al. 1999). Loss of high flows can cause the desiccation and death of existing riparian vegetation, or eventually eliminate certain species because of failure to regenerate (Rood and Mahoney 1990; Poff et al. 1997). Riparian vegetation, in turn, is a significant factor controlling the evolution of channel morphology. Riparian vegetation along stream banks helps prevent high rates of bank erosion and traps sediment to assist in building new bars and floodplains (Hicken 1984).

The Necessity for an Historical Perspective

Any imposed changes that modify stream flow or sediment flux in a stream reach, such as climatic fluctuations, water development such as dams and diversions, or other land use changes, may cause the stream channel to adjust. However, predicting the response of a given stream to a particular environmental trigger is far from clear cut. Each stream system is unique and may exhibit a singular response to imposed changes that cannot be readily integrated into any general model (Carling 1988). Rivers are historical systems, in that present form is influenced strongly by variables passed down through geomorphic and geologic time (Schumm and Lichty 1965). For example, the size of bed material available for transport may be dictated by the presence of relict alluvium in the floodplain, or the potential of a stream to adjust its slope is limited by the slope of the valley through which it flows. The response of a particular stream to external changes may be determined by autogenic factors internal to the stream system itself (Hooke and Redmond 1992). Such internal factors can be described in terms of intrinsic geomorphic thresholds (Schumm 1977) or chaotic system dynamics in which multiple equilibrium conditions exist so that, when perturbed, the stream system may evolve to a new equilibrium condition rather than returning to its initial state (Hooke and Redmond 1992). These issues complicate the problem of precisely identifying the relationship between cause and response, and underscore the need for an historical context in which to consider channel changes on an individual river.

Reconstruction of the 20th century geomorphic history of the lower Duchesne River allows us to characterize the unique responses of this system to historical fluctuations in discharge and sediment supply, and to identify the key geomorphic components controlling those responses. This information enables us to predict future channel changes with a greater degree of confidence. In addition, an historical perspective is necessary to evaluate the current trajectory of channel change and address the question as to whether the channel is currently adjusted to the prevailing hydrologic conditions.

DESCRIPTION OF THE STUDY AREA

The Duchesne River drains most of the Uinta Basin, a broad structural and topographic depression between the Uinta Mountains and the Tavaputs Plateau in eastern Utah. All of the Duchesne River's tributaries that have significant stream flow have their headwaters on the south slope of the Uinta Mountains. These tributaries flow in a southerly direction and join the Duchesne River on the south side of the Uinta Basin. The Duchesne River itself flows east and southeast along the basin axis. Its confluence with the Green River is at Ouray, Utah (Figure 1). The east-flowing portion of the river traversing the Uinta Basin is incised into Tertiary sandstones and shales. Remnants of Pleistocene outwash plains form scattered benches 100 to 200 m above the present river level (Osborn 1973).

The Duchesne River and its tributaries have been impacted by water development projects, including trans-basin tunnels that divert water to the west slope of the Wasatch Range. The Strawberry Tunnel began diverting up to 60,000 acre-feet per year of Uinta Basin water out of the basin in 1915. A second trans-basin tunnel, the Duchesne Tunnel, started diverting an additional 37,000 acre-feet annually in 1954 (Brown 1991). Currently, continued development of a system of reservoirs and canals known as the Central Utah Project is intended to deliver additional stream flow from a number of Duchesne River tributaries to Strawberry Reservoir and on to the Wasatch Front. As described in this report, reduced stream flow as a result of these and other development projects has altered the geomorphology of the alluvial valley and channel of the Duchesne River and has the potential to play a role in the loss of native fish in the upper Colorado River basin.

The lower Duchesne River extends from the mouth of the Uinta River near Randlett, Utah, to the confluence of the Duchesne with the Green River 27 km downstream (Figure 1). The reach extending 4 km upstream from the Green River has been designated as critical habitat for razorback sucker (*Xyrauchen texanus*), an endangered native fish species (Fed. Reg./Vol. 59, No. 54, March 21, 1994). Razorback sucker and another endangered native fish, the Colorado pikeminnow (*Ptychocheilus lucius*), have been documented to utilize habitats at the mouth of the Duchesne River (Modde 1997), and Colorado pikeminnow have been observed in areas throughout the lower 24 km of the river (Duchesne River Fisheries Study Progress Report:

1999). A USGS gaging station immediately downstream from the Duchesne River -Uinta River confluence, Duchesne River near Randlett (USGS station number 09302000), provides a continuous record of stream flow dating to 1943. Mean annual flow at the gage is 558 ft^3/s and the mean annual peak flow is approximately 4,310 ft^3/s for the period 1943 to 2000. The maximum recorded instantaneous peak discharge was 11,500 ft^3/s, and occurred on June 20, 1983. Although no significant tributaries enter the Duchesne River downstream from the gage, irrigation diversions and return flow ditches are present in this segment of the river.

The lower Duchesne River valley is approximately 2 km wide and is bounded on either side by gravel-capped bluffs in excess of 100 m in height. The relatively flat valley floor contains three distinct terrace surfaces, some of which functioned as active floodplains within historical time, as indicated by extant vegetation, early aerial photography, and early cadastral surveys. Through much of the study area, the valley bottom is best described as a wandering gravel-bed river floodplain (order B2), using the classification system of Nanson and Croke (1992). The active floodplain, which now occupies less than a quarter of the full valley width, is composed primarily of sandy scroll bar platforms with areas of dense vegetation farther from the main channel. Chute channels commonly define the boundary between the bars and vegetated floodplain surfaces. Numerous abandoned channels and swales are present on the floodplain, as well as on the lower terrace levels. Riparian vegetation consists primarily of dense tamarisk thickets (*Tamarix* spp.), russian-olive (*Elaeagnus angustifolia*), and young cottonwood (*Populus* spp.). The lower terrace levels contain decadent cottonwood galleries mixed with sagebrush (*Artemesia* spp.), rabbitbrush (*Chrysothamnus nauseosus*), squaw bush (*Rhus trilobata*), and greasewood (*Sarcobatus vermiculatus*). The river has a meandering planform, a mixed bed of cobbles, gravel, and sand, with an average gradient of about 0.0019. This general morphology extends through the upstream portion of the study area to approximately river km 9, where the channel slope abruptly decreases to less than 0.0003, bed material changes to sand, and channel form assumes a narrow canal-like geometry of low sinuosity. The floodplain in this downstream portion of the study area is a comparatively featureless plain dominated by expansive tamarisk thickets. It is best described as either a lateral migration/backswamp floodplain (order B3c) or a laterally stable, single-channel floodplain (order C1).

Three reaches within the study segment were selected for detailed geomorphic study and hydraulic modeling. These are the 24-hour Camp reach, the Above Pipeline reach, and the

Wissiup Return reach (Figure 2). Surveys of channel geometry were also conducted at three additional study sites (Figure 2, sites 4, 5, and 6). River locations within this report are sometimes referred to in terms of local landmarks, such as the Bowtie, Pipeline, Grey Bluff, and the Oil Shack (Figure 2).

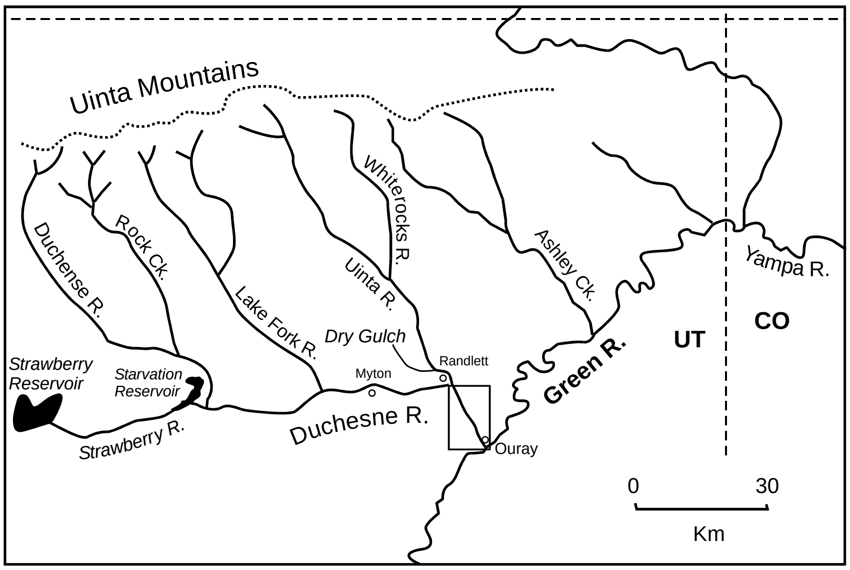
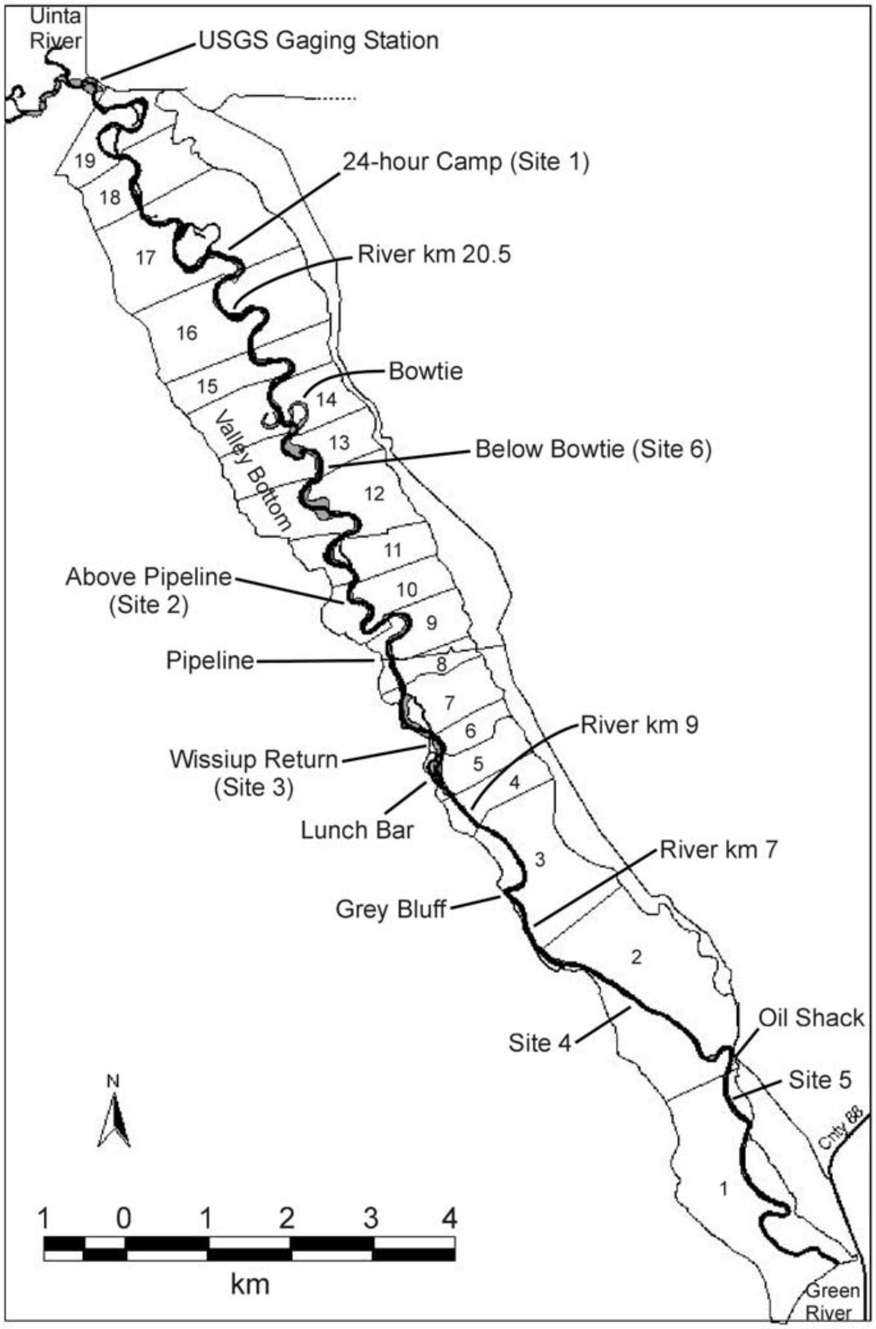


Figure 1: Map of Uinta Basin showing the location of the lower Duchesne River.



2: Map of the study area showing numbered subreaches, locations of detailed study sites, and landmarks.

METHODS

Hydrology

Historical stream-flow gaging records were obtained from US Geological Survey online databases. USGS gaging station 09302000 provides a continuous record of daily mean discharge for the lower Duchesne River from water year 1943 to the present. We extended this record back in time to water year 1912 by correlation with the USGS gaging station 09295000 (Duchesne River at Myton, UT). The Myton gage, which is located approximately 35 km upstream from the Uinta River, has been in operation since July 1911. We generated synthetic daily mean stream-flow records for the Duchesne River near Randlett prior to water year 1943 using MOVE1 methodology (Hirsch 1982) applied to overlapping data from the two gages spanning October 1, 1942, to September 30, 2000. Hirsch (1982) demonstrated that maintenance of variance extension (MOVE) techniques perform significantly better than ordinary least squares regression in producing synthetic stream-flow records with statistical properties similar to those of the longer-running time series.

Daily mean discharge measured at the gage near Randlett is linearly correlated with corresponding daily mean discharge at the Myton gage ($R^2 = 0.9539$), although a subset of the larger discharges at Randlett increase at a faster rate relative to this relationship, probably because of large inflow from the Uinta River in some years (Figure 3). Transformation of the data does not substantially improve the relationship, hence extension analysis was performed on untransformed daily mean data. Many of the predicted daily mean discharge values near Randlett are less than the lower boundary of the 90-percent confidence envelope for flood events greater than about 7,000 ft^3/s, suggesting that the magnitude of the largest events may be underestimated in the synthetic time series (Figure 4).

We used the extended daily mean discharge record for the gage near Randlett to generate flow duration curves, average hydrographs, flood frequency analyses, and an annual flood discharge time series. Annual flood discharge for each water year was calculated as the total discharge in excess of a given flood discharge threshold. Thus, days for which the daily mean discharge was less than the threshold discharge contribute nothing to annual flood discharge. Days for which the daily mean discharge is greater than the threshold contribute the difference between the daily mean discharge and the threshold discharges in units of ft^3/s-days.

Multiplication by the number of seconds in a day converts ft^3/s-days to water volume in cubic feet. Flood frequencies and the magnitude of floods of specific recurrence intervals were calculated for several time intervals by fitting the plotting positions of annual maximum daily peaks to a log Pearson Type III distribution. Maximum daily peaks were chosen for this analysis, because instantaneous peak data cannot be computed for the synthetic portion of the record prior to 1943. Average hydrographs for specific time intervals consist of the average of all daily mean discharge records within the interval for each day of the water year. All records for February 29 were removed from the time series before construction of average hydrographs.

Suspended-Sediment Transport

The USGS measured suspended-sediment concentrations in the Duchesne River at the gaging station near Randlett on an approximately monthly basis between July 1975 and March 1989. We analyzed these 136 concentration measurements to produce sediment-rating relations for estimating suspended-sediment discharge in the Duchesne River. Measurements were sorted into high flow and base flow classes. Base-flow measurements for this analysis were defined as all concentration measurements taken at times when discharge was less than 600 ft^3/s. Measurements below this threshold exhibit no discernable relationship between concentration and discharge (Figure 5). Discharge events greater than 600 ft^3/s generally occurred during the snowmelt runoff season, and were separated into rising and falling limb flows according to whether they occurred before or after that year's peak date. Discharges greater than 600 ft^3/s were recorded during fall and winter months on four occasions and were excluded from subsequent high-flow analysis. Suspended-sediment measurements taken on the rising limb and falling limbs of the annual spring peak showed distinct relationships between concentration and discharge (Figure 5). Ratings relations were developed as follows:

$$C = C_{mx} - \frac{C_{mx} - C_{mn}}{1 + \exp\left[k_s (Q - Q^*)\right]} \qquad (1)$$

where Q is discharge in cubic meters per second, C_{mx} is a maximum concentration occurring at large discharge, C_{mn} is the average concentration at base discharge, k_s scales the discharge range over which the transition from C_{mn} to C_{mx} occurs, and Q^* is the midpoint of that range in cubic meters per second. Equations of this form were fit by eye to base-flow plus rising limb data and base-flow plus falling limb data to yield the following suspended sediment concentration ratings relations (Figure 5):

$$C = 2800 - \frac{2680}{1 + \exp[0.08(Q - 55)]}\text{; rising limb} \qquad (1a)$$

$$C = 800 - \frac{680}{1 + \exp[0.07(Q - 65)]}\text{; falling limb} \qquad (1b)$$

where discharge is in cubic meters per second and concentration is expressed in milligrams per liter.

This functional form reflects reasonable constraints that are consistent with these data and with better-measured concentration time series on comparable streams. Among the simplest, and perhaps most common, functional form used for suspended sediment rating relations is the power function. In this data set, suspended sediment concentration increases with discharge in a manner that can be well represented by a power function only through only a small range of flows that occur in the Duchesne River. These data show that no systematic correlation between concentration and discharge is apparent under base flow conditions, such that base flow concentrations are best represented by a mean value. Only two measurements are available to represent concentrations at discharges approaching or exceeding bankfull flow. However, these measurements indicate that there is an upper limit to the suspended sediment concentration. Simply extrapolating a power function to higher flows would not capture this phenomenon. Instead, extrapolation would produce extremely high concentrations that are inconsistent with this data and with observations elsewhere. These constraints – low-flow concentrations that are insensitive to discharge and an upper bound on concentration – define the relations we fitted to the data.

Total annual suspended-sediment discharge was estimated for time periods before 1925 and after 1943, and for upper-quartile, lower-quartile, and middle-quartile years within the period of measurement from 1943 to 2000. Upper-quartile years were defined as years with total annual discharge in the upper quartile of all years in the measurement period. Water years with total annual discharge in the lowest quartile of all years in the measurement period were defined as lower-quartile years, while water years with total annual discharge in the middle two quartiles were defined as middle-quartile years. Average annual suspended sediment fluxes were estimated by separating the records from each water year into rising and falling limbs of the annual spring peak, multiplying each daily discharge by the daily load computed from the appropriate ratings relation, and dividing by the number of water years in each interval or quartile subset.

Geomorphic Mapping

The distribution of geomorphic surfaces in the alluvial valley of the Duchesne River between Ouray and Randlett was mapped in the field in June 2000 onto a 1997 aerial photograph base. The river configuration depicted on the 1997 photo base was generally consistent with observed field conditions, except in a short reach known as the Bowtie (Figure 2) where there were significant channel changes in 1998 and 1999. Major types of geomorphic surfaces defined for this study include channel, bar surfaces, floodplain surfaces, and terrace surfaces. Each of these major categories was further subdivided into several sub-types (Table 1). Mapping units were distinguished in the field and on the photo base by elevation above the stream, surface topography, vegetation, and surficial geologic materials. Unit boundaries were precisely drawn on mylar overlain on the photo base in the field and subsequently adjusted in the laboratory with the aid of stereoscopic viewing and standard photogeologic techniques. Map units and geomorphic photo interpretations developed in the field and with the 1997 photographs were applied to seven additional sets of aerial photographs taken between 1936 and 1993. Mapping unit boundaries were drawn on mylar overlays for each photo set while viewing the photos through a stereoscope. Because the degree of uncertainty in classification of surface units increases on older or poor-quality photo bases, a system of inheriting map units from younger photo bases was implemented. Once a map unit classification was determined for a map unit observed on a given photo series, the same classification was assumed for the same surface on the preceding photo series, unless there was compelling evidence of change. This method allows surface classifications verified by field observation to be propagated to older photomaps for which field verification is impossible. In applying this method, we recognize that the character of some geomorphic surfaces change over time even in the absence of fluvial reworking. For example, we believe some areas classified as cottonwood terrace in 1997 were active floodplain in 1936.

Creation and Analysis of GIS Databases

Geomorphic map overlays drawn on the aerial photographs were digitized into a geographic information system (GIS) and assigned database attributes. These data were examined to evaluate historical changes in channel form and size, and changes in the distribution of other alluvial surfaces. Completed GIS coverages were developed from mapping using photographs taken in 1936, 1948, 1961, 1969, 1980, 1987 and 1997. A coverage from photos taken in 1993 showed no apparent change from the 1987 coverage in preliminary comparisons, and so was not included in subsequent GIS development.

<u>Digitizing</u>

Arc/Info® GIS software and a Calcomp 9100® digitizing tablet were used for all digitizing. Control points were identified for each photo base using a variety of quasi-stable features such as road intersections and individual bushes or trees that could be located on USGS orthophoto maps. Identification of adequate positional control for this study area proved difficult, as cultural features are few and natural features change over time. Most roads available for control in the area are dirt tracks that also may change in position over time. Nonetheless, root mean square (RMS) errors were modest for most scenes digitized, averaging 6.5 m for the 1936, 1961, 1969, 1980, 1987 and 1997 coverages. Reported RMS errors were considerably greater for the 1948 photos at 33.9 m (Table 2). These photos, believed to have been taken during military training flights, are subject to significant distortion cause by aircraft tilt. The errors associated with the finished GIS coverage were reduced from the reported RMS error by an unknown amount by a rubber-sheeting procedure, described below, which was applied to the digitized scenes.

During digitizing, the GIS software places the specified control points within the coordinate system according to the given control point coordinates. Any positional error in control point locations on the photographs will cause the actual control point locations in the digitized coverage to be placed at slightly different coordinate locations than the coordinate locations specified. The rubber-sheeting procedure used in this study consists of digitizing points representing control point locations for each scene, then linking the actual control point locations to their proper specified coordinates and moving them to their proper coordinates with Arc/Info's "adjust" command. This command causes the program to apply a polynomial interpolation based on the displacement specified for the control points to reposition all features in the coverage. This procedure has the potential to significantly reduce the reported RMS error associated with distortion of the original photographs. However, we have no basis to determine the final positional accuracy of the rubber-sheeted coverage. In the case of the 1948 coverage, we conservatively estimate that it reduced the initial RMS error by 50 percent. This coverage then has an estimated RMS error of approximately 17 m. Portions of cadastral survey maps from 1875 and 1882 were also digitized using intersections of section lines as control points. The resulting coverage shows the approximate river bank location in the late 19th century.

<u>Analysis of Geographic Data</u>

The study area was divided into 19 subreaches between 0.5 – 2.7 km in length, measured down the valley axis (Figure 2). These subreaches were established to facilitate the analysis and reporting of longitudinal patterns of channel form and adjustment. Subreach boundaries were drawn normal to the channel to the extent permitted by temporal changes in channel configuration. Metrics of channel change were obtained for each subreach, and areas of similar channel behavior were identified. Subreach metrics include subreach channel width, areas of erosion and deposition, volumes of gravel erosion and deposition, and changes in gravel storage.

<u>Channel Width</u>

Two width metrics were defined for each subreach: low-flow channel width and floodway channel width. The low-flow channel was defined as the region comprised of the water surface and low bars. Dividing the area of this region within a subreach by the channel length within the subreach yielded the average low-flow channel width for the subreach. The floodway channel was defined as the low-flow channel plus adjacent high bars, i.e., it is the near-channel area that is essentially free of vegetation. Width of the floodway channel was computed by dividing the area of water plus both high and low bars within each subreach by the channel length within the subreach. The channel length used in these computations was measured along the centerline of the floodway channel for each photo series analyzed, and varied from year to year with changes in channel pattern. The area of a 10-km-long secondary channel that was active in the middle part of the study area in 1936 was excluded from the main channel width calculations.

Error in calculating subreach channel width stems from error in the determination of channel and bar polygon areas arising from three primary sources: mapping error, digitizing error, and photo distortion. Error in mapping channel area can occur where differences in discharge between various photo series produces the illusion of greater or lesser width of the active channel. This is a potential source of error in this study, as discharge at the time photographs were taken ranged from 8 ft^3/s to 2,430 ft^3/s (Table 2). However, the mapping scheme employed in this study was designed to minimize this source of error. The 'low bar' mapping unit was applied to the emergent portions of lateral bars and riffles visible at low discharges and other low-elevation bar surfaces. These low bars are progressively submerged at higher discharge so that increasing stage results in both an increase in water area and a corresponding decrease in the area of low bars. As the low-flow channel was defined to include low bars, these changes have no effect on the area assigned to the low-flow channel as long as flow is contained within the low-flow channel banks. These banks, which separate the low-flow channel area from the higher bar surfaces, are steep and well-defined, with relief of 0.5 m or more, and can be distinguished on aerial photographs when viewed through a stereoscope. We demonstrate below that bankfull discharge on this portion of the Duchesne River, when the water surface would be expected to cover significant portions of the high bar surface, is greater than 3,000 ft^3/s in most locations. Discharges associated with all aerial photographs used in this study are well below this threshold. It was assumed that error in distinguishing high bar surfaces from the low-flow surfaces due to fluctuations in discharge constituted less than 10 percent of the low-flow channel width in most subreaches. A notable exception occurred in subreach 2 on the 1948 photographs where the river had abandoned a long section of its channel and was in the process of incising a new channel farther to the west. Relatively high discharge of the Duchesne River at the time, coupled with backwater effects from the Green River, may have produced a significant exaggeration of channel width through this area of avulsion (Figure 6). The abandoned channel is indicated with a dashed line in the figure.

Mapping error associated with the placement of unit boundaries between high bars and other terrestrial units cannot be separated from the subjective nature of these boundaries, which often pass through gradations in vegetation density or elevation lacking sharp breaks. Variability in the perceived locations of these boundaries reflects the changes in vegetation density and fluvial activity being evaluated by these metrics.

Digitizing error is the result of the inability of a GIS technician to exactly trace mapped lines during the digitizing process. The effect of this error on polygon area depends on polygon shape, polygon size, and the magnitude of the digitizing error. Sondossi (2001) empirically determined the effect of random linear errors of 2 m or less for polygons with an average length-to-width ratio of 5:1 and a range of areas ranging across five orders of magnitude. He reported that the percent error in polygon area for a given linear error of 2 m declined with polygon size according to the relationship:

$$E_A = 27.827A^{-0.464} \qquad (2)$$

where E_A is the percent error in polygon area and A is polygon area. This relationship should hold for all polygons of similar shape. As Sondossi's polygons also represented fluvial surfaces along a river corridor and have a similar relationship between polygon perimeter and area [Sondossi reported $P = 3.36A^{0.59}$ while a test of this data set using polygons representing channel, high bar, and floodplain areas from the 1980 coverage yielded $P = 3.364A^{0.565}$], it was assumed that the Sondossi relationship between percent error in area due to a given linear error and polygon area hold for this data set as well. Sondossi also determined that for a given polygon size and shape, percent error in area scales linearly with the random linear error. In other words, doubling the magnitude of the random linear error results in a doubling of the percent error of a given polygon.

Following Sondossi, random linear errors associated with this study were evaluated in terms of photograph scale. The ability of a GIS technician to accurately trace and digitize unit boundaries is limited by the width of the lines used to map unit boundaries on air photos A 0.5-mm pencil was used for mapping in the present study, and the smallest-scale photo enlargements were 1:14,500. A 0.5-mm accuracy limit for drawing and digitizing unit boundaries on our enlargements therefore incorporates maximum positional errors about 7.2 m on the ground. Low-flow channel polygons used in this study average approximately 41,000 m^2 in area while floodway channel polygons are considerably larger. Application of the Sondossi relationships to the average areas of polygons used to calculate channel width in this study using a random linear error of 7.2 m results in an insignificant expected error of less than 1 percent.

Error in polygon area caused by photograph distortion has been empirically determined to average about 3 percent when working near the edges of photographs where distortion is most exaggerated (Van Steeter and Pitlick 1998). As we mapped near the centers of photographs as much as possible, error in polygon area from photograph distortion is probably well below this value. Large-scale registration errors like those described by reported RMS errors have a muted effect on errors in polygon area because these deviations from true coordinates tend to displace all portions of a polygon in a similar direction and magnitude. The result, especially for small polygons, is a simple translation of polygon location with little effect on polygon shape or area. Taken together, we assumed that all sources of error produce no more than a 10 percent error in the average channel width reported in this study. An exception was made for subreach 2 from the 1948 coverage, where the reported channel width is unreliable.

Areas of Erosion and Deposition

Areas of erosion and deposition between sequential coverages were obtained for subreaches 5-19 by performing a spatial union in Arc/Info and classifying the resulting change polygons as areas of erosion, deposition, or no change (Table 3). Important potential sources of error in the reported erosion/deposition areas included mapping errors and positional errors due to systematic shifts in the coordinate system. Mapping error in this context consisted of incorrect or inconsistent placement of unit boundaries due to the gradational nature of many boundaries. An example would be a floodplain surface that gradually grades into a terrace surface over an extended slope. Variation in placement of the boundary between the two units on successive photo series could result in an apparent change from floodplain to terrace where no real change occurred. Errors of this type were located and eliminated by visual inspection of the relevant change polygons. It should be pointed out that transitions between high bars and floodplains were considered to represent no change with respect to erosion or deposition because these surfaces are found at similar elevations in this study area. Their differences lay primarily in the extent of vegetation colonization and in lateral distance from the active channel.

Positional errors associated with the overlay of polygons from coverages for different years are more problematic than are errors within a single coverage. This is because feature shifts due to photograph distortion or inadequate control points in one coverage are independent of feature shifts associated with the other coverage. Consider two successive coverages, each associated with a positional error described quantified by the RMS error reported by the digitizing software. Actual errors at specific points on each coverage are vectors whose magnitudes and directions are spatially variable. The RMS error for each of the two coverages therefore represent a large number of individual error vectors that are independent of the error vectors on the other coverage, such that the errors for the two coverages can be combined as the square root of the sum of their squares (Benjamin and Cornell 1970) to give the average total displacement error for the overlay coverage ($\overline{D_T}$). Average total displacement errors for successive pairs of coverages range from 9.4 to 19.2 meters (Table 5).

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 self-compensating, and ignored them. While positional errors produce false areas of erosion and deposition on the overlay coverage, they also conceal similar areas of real erosion and deposition. This type of error compensation requires that actual channel movement producing real areas of erosion and deposition occurred during the time period being analyzed. For periods when the areas of real erosion and deposition are less than the potential errors, only a portion of the potential error can be compensated. For periods when no real channel change occurred, all apparent change is due to positional error. It is therefore necessary to evaluate the magnitude of the potential area of false erosion and deposition relative to the measured areas of erosion and deposition to determine what portion of the planimetric error is likely to be compensated.

We evaluated the average effect of independent positional shifts by determining the net relative displacement of polygons from successive coverages and using a sine wave to model the sliver area generated by error displacement of a meandering channel. The channel bank is represented by a sine function multiplied by an amplitude coefficient that produces a sinusoidal curve with a sinuosity value equal to the sinuosity of the river channel (Figure 7). This sinusoidal curve is then translated a distance equal to the total displacement error between the two successive photo pairs being evaluated. The error generated by the curve translation is equal to the area between the two curves, and is computed according to:

$$h = \int_{-\pi/2}^{3\pi/2} \left| Sin(x - Sin(r))D_T m - Sin(x)m \right| dx \qquad (3)$$

where h is the sum of all area between the two curves, D_T is the magnitude of the displacement vector, r is the spatially-variable azimuth of the displacement vector, and m is the amplitude of the sinusoidal curves. The magnitude of D_T is equal to the root mean square of the positional errors associated with each of the two coverages. The azimuth (r) is unknown, but is equally likely to take any value between 0 degrees and 359 degrees. Because of the symmetry of the curve, only values between 0 degrees and 90 degrees need be considered. The mean value of h for all possible azimuths is determined as:

$$\bar{h} = \frac{1}{\pi/2} \int_0^{\pi/2} h \, dr \tag{4}$$

Another pair of sinusoidal curves is needed to represent the opposite river bank, so that the expected total planimetric error (ε_a) is equal to $2\bar{h}$. The average displacement error calculated in this manner represents the area of false polygons expected to result from the spatial overlay of two independent coverages (Table 5).

We experimentally evaluated the proportion of ε_a that would remain uncompensated by real change with computer simulations in which two sinusoidal curves, one representing error displacements and one representing real changes in channel position, were independently displaced. The mean and standard deviation of the uncompensated area between the two displaced curves was numerically determined for all possible combinations of displacement azimuths. Simulation results showed that the expected uncompensated planimetric errors (ε_{ac}) for the six overlays range between 1 percent and 10 percent of the measured areas of erosion and deposition (Table 5). Subtracting ε_{ac} from the measured areas of change gives an improved estimate for the actual area of change. The standard deviations of the simulated uncompensated planimetric errors were used to calculate uncertainty margins about the corrected estimates of erosion and deposition areas. These margins of error are denoted by ε_{au}, and range between about 8 percent and 20 percent for the six overlays (Table 5). A more detailed description of the methods used for calculating ε_{au} is given in Gaeuman et al. (submitted).

The areas of displacement errors calculated from the original RMS errors associated with the digitized coverages overestimate the actual probable displacement errors by an unknown amount. Actual errors are reduced in this study by manual corrections applied to the coverages before overlay. Some false polygons generated by displacement errors are obvious, and help to identify portions of coverages in need of correction. Any polygon indicating a change from an active channel to a terrace surface is clearly false. A local displacement error equal to or greater than the local width of the false polygon must exist at each point along the channel adjacent to the false polygon. Significant false polygon errors were initially detected at between one and five locations on each change coverage. For each sequentially paired set of coverages, one coverage was rectified relative to the other using a spatially-limited rubber-sheeting procedure. Near channel features on the coverage being rubber-sheeted were adjusted to a position chosen to reduce or eliminate any false polygons upon subsequent spatial overlay with the other coverage in the pair. All adjustments were limited to the immediate vicinity of false polygons by using the "limitadjust" feature of the software.

Development of a Gravel Budget

Sediment budgets were developed for subreaches 5-19 by multiplying areas of erosion and deposition for each subreach by the estimated average thickness of gravel deposits associated with particular map units. This approach required estimation of the elevations of each geomorphic surface above a datum tied to the channel bed and estimation of the elevation of the top of the gravel deposits in each surface. Reach-scale budgets for gravel were developed by establishing a downstream boundary condition where gravel transport rates were assumed to be zero, then sequentially accumulating storage changes in the upstream subreaches. The downstream zero-transport boundary was assumed to exist near the point where the channel gradient abruptly flattens to less than one-third of its upstream value and the river bed begins a rapid transition from gravel to sand.

Gravel-thickness attributes were assigned to particular map units based on field measurements. The longitudinally variable deposit heights used in calculating gravel volumes in this study are reported in Table 6. Terraces within subreaches 5-19 are composed of gravel deposits overlain by a layer of sand or silt of variable thickness. The elevations of terrace and the tops of gravel deposits underlying the terraces above a base-flow water surface level of 500 ft^3/s were observed and mapped at all cutbanks through subreaches 5-19 in October 1999. An average stage-discharge relationship was estimated using several surveyed cross sections and used to adjust subsequent observations of deposit heights to a similar water surface datum.

Measurements of gravel thicknesses in the cottonwood terrace unit were made upstream from subreach 14 and downstream from subreach 12. No measurements were made in subreaches 12, 13, and 14 because all cutbank exposures in this reach of river are in the high terrace unit, in which no exposed gravel is present. Gravel deposit thicknesses in the cottonwood terrace for subreaches 8 through 10 were estimated using a least-squares linear fit through the data point upstream and downstream (Figure 8). These data indicate that gravel thickness declines slightly in the downstream direction between the upstream study area boundary and subreach 4. The standard deviation of measurements made upstream from subreach 10 is 0.33 m, as is the standard deviation of measurements made downstream from subreach 8. In subreach 4, gravel thickness in the cottonwood terrace declines rapidly as the cottonwood terrace unit is replaced by the tamarisk terrace unit.

High bars in the study area typically consist of a gravel platform capped by sand. The mean height of high bar surfaces above the reference water surface level at detailed survey sites was determined using surveyed cross sections. Chute channels are common on point bars in the study area, and often scour away sand overburden to reveal the coarse basal platform underlying the point bar. These exposures, plus a small number of pits dug in bar and floodplain surfaces, were used to estimate the mean elevation of the gravel base at the same locations. Floodplain surface elevations were also determined from surveyed cross sections. Pit, cutbank, and side channel exposures indicated the gravel depth of floodplain surfaces is similar to that of high bars. The elevation of the top of gravel deposits in high bar and floodplain units maintains a fairly consistent relationship with the elevation of the top of gravel deposits in the cottonwood terrace upstream from subreach 4. Gravel thickness in the floodplain and high bar units also decrease in the downstream direction, maintaining an average top elevation about 0.4 m below the top of the cottonwood terrace gravels.

<u>Uncertainty in Gravel Erosion and Deposition Volumes</u>

Uncertainty in estimated volumes of erosion or deposition was calculated for each erosion or deposition polygon as the product of polygon area and the uncertainty in the thickness of gravel in the alluvial deposits (δH). The value of δH is estimated as the standard deviation of field measurements of gravel elevations, as described above. Volumetric uncertainty pertaining to gross erosion and deposition estimates therefore incorporates any uncompensated positional error, such that gross volumetric uncertainty for any individual polygon (δV_g) was estimated as the square root of the sum of squared contributing errors and uncertainties:

$$\delta V_g = \sqrt{\varepsilon_{au}^{\ 2} + \left(A \cdot \delta H\right)^2} \qquad (5)$$

where A is the area of erosion or deposition. Gross volumetric uncertainties for erosion volumes (δV_{gi}) within each channel subdivision, i, were found by summing δV_g for all erosion polygons in each subdivision, i. Gross volumetric uncertainties for deposition volumes ($\delta V'_{gi}$) were similarly found by summing δV_g for all depositions polygons in each subdivision. Gross volumes are reported as a percent of the measured erosion or deposition in each subreach (Table 7).

Uncertainty in Estimated Changes in Gravel Storage

Uncertainty in estimated changes in gravel storage is produced primarily by uncertainty in the thickness of gravel in the alluvial deposits. Planimetric errors are considered to have little effect on the calculation of changes in gravel storage (Ham and Church 2000). A local positional error that produces a false erosion polygon on one river bank also produces a false deposition polygon on the opposite bank, even in the absence of real channel movement. 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 not affected by the planimetric error. Equation 5 then reduces to:

$$\delta V = A \cdot \delta H \qquad (6)$$

where δV is the uncertainty in the volume of erosion of deposition in a specific polygon.

Uncertainty regarding gravel deposit thicknesses 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 channel subreach (Table 7) is then the combined uncertainties associated with the estimated volumes of erosion and deposition:

$$\delta \Delta S_i = \sqrt{\left(\delta V_i\right)^2 + \left(\delta V_i'\right)^2} \qquad (7)$$

where $\delta\Delta S_i$ is the uncertainty in the change in storage in subreach *i*, δV_i is the volumetric uncertainty summed for all erosion polygons in subreach *i*, and $\delta V_i'$ is the volumetric uncertainty summed for all deposition polygons in subreach *i*.

Equation 7 indicates that the magnitude of the uncertainty margins associated with storage changes are based on the uncertainties associated with the total volume of erosion plus deposition within a subreach, rather than on the net volume difference between erosion and deposition (the storage change). Where the total volume of erosion plus deposition is large but the magnitudes of erosion and deposition are similar, the net storage change is relatively small. The uncertainty margins associated with these relatively small storage changes can be comparatively large, and in some cases are much greater in magnitude than the storage changes (Table 7).

Hydraulic Modeling

HEC-RAS software (US Army Corp of Engineers 1998) was used to model one-dimensional steady flow at three detailed study sites and through an extended downstream reach of the Duchesne River to its confluence with the Green River. Models at the three detailed study sites were intended to evaluate the discharges necessary to inundate targeted geomorphic surfaces and to initiate movement of the gravel substrate. A variable number of monumented channel cross sections were surveyed and monitored through two runoff seasons to gather stage-discharge data for use in model calibration. The modeling effort in the downstream portion of the Duchesne River was intended to determine the extent to which backwater effects from the Green River control stage, and therefore the extent of flooding, in the downstream portion of the study area.

<u>Data Acquisition</u>

We established, developed, and monitored three detailed study reaches in the gravel-bedded portion of the Duchesne River during the 2000 and 2001 field seasons. The locations of these sites, 24-hour Camp, Above Pipeline, and Wissiup Return, are shown on Figure 2. Although each site was visited for monitoring purposes on numerous occasions over the course of two field seasons, relatively high flows useful for model calibration were rare during the study period. Runoff during spring 2000 was exceptionally low, making collection of high-flow calibration data impossible that year. Peak runoff in 2001 was also below normal, but peak water surface measurements at discharges up to 2,450 ft^3/s were obtained in May. Cross section geometry was not re-surveyed at high flow.

The 24-hour Camp site consists of 10 monumented cross sections and five supplementary cross sections (Figure 9). Stage observations most suitable for high-flow calibration at 24-hr Camp were made May 26, 2001 when discharge was 2,450 ft^3/s and May 22, 2001 when discharge was 1,050 ft^3/s. Five of the 24-hour Camp cross sections traverse a chute channel active at moderate discharges. The chute portions of these cross sections were separated from the main channel portions and used to construct an independent stream segment connected to the main channel at upstream and downstream junctions. Flow in the chute began when main channel discharge was about 1,050 ft^3/s, and discharge in the chute was measured at 58 ft^3/s when main channel discharge was 2,600 ft^3/s.

Five monumented cross sections were installed at Above Pipeline (Figure 10). A single stage observation above base flow was obtained for the cross sections at the Above Pipeline site on May 23, 2001 when discharge was 770 ft^3/s. In spite of the small number of cross sections at this site and the small discharge used in calibration, we believe that HEC-RAS modeling provides a more robust method of estimating stage-discharge relationships at cross sections than considering single cross sections in isolation.

At Wissiup Return, we installed seven monumented cross sections and three supplementary cross sections (Figure 11). Useful stage observations were made at this site on May 28, 2001 when discharge was 1,840 ft^3/s and May 16, 2001 when discharge was 1,040 ft^3/s. The Wissiup Return model also includes an independent chute channel segment consisting of portions of six cross sections. This chute becomes active at higher flows than the chute modeled at 24-hour Camp. The chute was conveying little flow when main channel discharge was 1,840 ft^3/s, which is the highest discharge observed at the site.

Extrapolating observed stages to larger discharges at a single cross section, either by applying a uniform flow equation such as the Manning equation or by fitting observed data to a curve, does not take into account the effects of changes in downstream hydraulic controls that may occur with changing discharge. Use of the HEC-RAS model allowed us to include the effects of downstream controls on upstream cross sections. We developed one stage-discharge relationship to serve as a downstream boundary condition for each modeled reach. Assuming roughness parameters were adequately specified, any error in that initial water surface elevation tended to diminish for predicted energy heads at more upstream cross sections.

Longitudinal variability in discharge caused by irrigation withdrawals and returns through the lower Duchesne River is a potential source of error in evaluating stage-discharge relationships at the modeling sites. The Wissiup Ditch diverts a potentially significant portion of flow from the river just downstream from the 24-hour Camp site. Part of this water is pumped out of a floodplain pond, part returns to the river at about river km 16.5, and a large portion re-enters the river at the downstream end of the Wissiup Return site. Another large ditch, the Leland Canal, diverts an unknown quantity of water upstream from the gage near Randlett. Some portion of this water re-enters the river as diffuse return flow downstream from the gage. Deviation in discharge at the modeling sites as compared to the discharge recorded at the gage near Randlett was evaluated by measuring discharge at each of the study sites. In addition, discharges calculated from 21 channel transects with slope and velocity measurements supplied by US Fish and Wildlife personnel were compared to the discharges recorded at the gage. Although interpretation of these data was complicated by problems with unsteady discharge at the gage and distance from the gage, the data suggest that longitudinal discharge variation through the study area is generally within about 70 ft^3/s of the gaged value during the irrigation season. This magnitude of variability represents a significant proportion of base flow, but is relatively small compared to peak flow.

Cross section data for building the Green River backwater model was mainly extracted from a 2-m grid of river bed elevations developed by JC Headwaters, Inc. This grid was produced using DGPS-integrated hydro-acoustic technology in June 1999. Fifty-nine cross sections extending from 0 to 9600 meters upstream from the confluence with the Green River were extracted from the grid using an Avenue script in ArcView® GIS. Two of the extracted cross sections were selected so as to coincide with a pair of cross sections we surveyed on the lower Duchesne River during the 2000 field season. The locations of these cross sections (site 4 and site 5) are indicated on Figure 2. Cross sections extracted from the grid portray the submerged bed of the river only – banks are omitted. Comparison of the extracted cross sections with surveyed cross sections allowed reconstruction of the bank heights above the channel bed. As the entire downstream sand-bedded portion of the river exhibits simple canal-like channel morphology, bank height at the two surveyed sites was assumed to be representative of bank height through the full modeled reach.

Model Calibration

HEC-RAS models were constructed for three discharges whose water surface levels were observed at the 24-hour Camp study site: 2,450 ft^3/s, 1,050 ft^3/s, and 226 ft^3/s. One calibration model was constructed for the Above Pipeline site, 770 ft^3/s. At Wissiup Return, models were constructed for two flows: 1,840 ft^3/s and 1,040 ft^3/s. Model calibration at all sites involved varying values of Manning's *n* at each cross-section station to attain a satisfactory fit between predicted and observed water-surface elevations at each cross section for each observed discharge. All modeled water-surface elevations were brought to within 3 cm of observed values using values of Manning's n appropriate for this stream type (Figure 12). In most rivers, *n* values typically decrease with discharge. Where the variation in *n* between the higher discharges at a given cross section was modest, the *n* value for the largest observed discharge was taken to be the appropriate *n* value for higher discharges. At 24-hr Camp, *n* values for 1,050 ft^3/s and 2,450 ft^3/s were very different at some cross sections. At these cross sections (stations 470, 668, and 709), the *n* value for higher discharges was estimated with the aid of a logarithmic curve fit to the *n* values determined for each of the three calibration discharges.

Cross sections were assigned their estimated values of Manning's *n* for all modeling runs extrapolating stage at higher discharges (Table 8). Values of *n* in the chute modeled at 24-hour Camp was calibrated by adjusting *n* and discharge into the chute to match water-surface elevations and discharge observed in the chute during May 2001. Discharge into the chute at modeled discharge levels greater than those observed was determined by adjusting discharge into the chute until the total energy head at the most upstream chute cross section matched the total energy head in the main channel at the chute's upstream junction. As no stage-discharge measurements were made for stage-discharge calibration in the chute channel at Wissiup Return, cross sections in this chute were arbitrarily assigned moderate values of Manning's *n* between 0.035 and 0.04. When running the model for higher discharges, it was necessary in increase Manning's *n* in some cross sections to prevent the prediction of supercritical flow at certain discharges. Discharge into the chute for high-flow extrapolations was estimated by adjusting discharge into the chute until the total energy head at the most upstream chute cross section matched the total energy head in the main channel at the chute's upstream junction. Downstream boundary conditions for modeling higher discharge were derived by fitting a curve to stages measured at the downstream cross section over the observed range of discharges.

Input data for the Green River backwater model consisted of an estimate for Manning's *n* for the Duchesne River and a downstream boundary condition comprising a stage-discharge relationship for the Green River at the mouth of the Duchesne River. Cross sections of the Duchesne River were assigned Manning's *n* values of 0.025. The lack of rigorous calibration of this model is justified on the grounds that it is intended to predict only the spatial extent of backwater effects extending up the Duchesne River from the Green River. River stage within this backwater zone is primarily a function of stage at the downstream boundary such that local channel roughness has little or no effect on local hydraulic conditions. This model is not intended to accurately predict flow velocity or depth, and has no validity beyond assessment of the influence of the Green River backwater. A stage-discharge relationship was previously developed for Old Charlie Wash 1, a site on the Green River approximately 500 m upstream from the mouth of the Duchesne, as part of the US Fish and Wildlife Recovery Program for Endangered Fishes of the Upper Colorado River (FLO Engineering 1997). We adjusted the datum used at Old Charlie Wash to better match the elevation at the mouth of the Duchesne River, and this relationship was applied as a downstream boundary condition to our model. As geographic data in our model were derived from the work of LC Headwaters, Inc., the water surface elevation at the mouth of the Duchesne River reported by LC Headwaters was used for datum adjustment. LC Headwaters reported a water surface elevation of 1,425.2 m at a time when the Green River discharge was approximately 18,000 ft^3/s.

Sediment Sampling

Two standard pebble counts were conducted on each of two riffles at each detail study site. Surface layer size distributions produced from these data allowed determination of mean gravel size for use in critical shear stress analysis at each site. Bulk subsurface bed material samples were also collected from riffles exposed at low flow at each site according to the 2-percent criterion of Church et al. (1987). These samples were sieved through 8 mm screens in the field, and the remainder was dry sieved through finer screens in the laboratory. Samples of fine sediment were collected from several locations in pools and backwaters on occasions before and after the 2001 runoff. Sandy samples were dry sieved. Silty samples were oven-dried, weighed, then wet sieved to remove the silt and clay fraction. The remaining sample was then oven dried and reweighed to determine the weight of lost fines before dry sieving (Guy 1969).

Estimation of the Discharge Necessary for Gravel Entrainment

Use of HEC-RAS modeling produces friction slopes for individual cross sections that can be used with stage and channel geometry data to estimate shear stress at the bed for various discharges. HEC-RAS reports average boundary shear stress for each modeled discharge at each cross section calculated from the Duboys equation:

$$\tau_0 = \gamma RS \qquad (8)$$

where τ_0 is the average shear stress exerted on the bed, γ is the specific weight of water, R is hydraulic radius of the flow at the cross section, and S is friction slope through the cross section. Hydraulic radius, defined as the area of flow in the cross section divided by the wetted perimeter.

The magnitude of shear stress necessary to entrain gravel depends on the sizes of gravel present on the channel bed. We used the Shields equation to calculate the dimensionless critical shear stress through riffle areas at the detailed study sites. This relationship is:

$$\tau^* = \tau_{cr}/D_i(\gamma_s - \gamma_w) \qquad (9)$$

where τ_{cr} is the critical shear stress necessary to initiate movement of bed particles, γ_s *and* γ_w are the specific weights of sediment and water, D_i is a representative sediment particle size, and τ^* is the Shields parameter expressing the dimensionless critical shear stress for particle entrainment. We evaluated the value of τ^* for a range of discharges using measured gravel particles sizes and values of shear stress determined from the output of our HEC-RAS models. For purposes of this analysis, we assumed that bed entrainment can be evaluated by entrainment of the median particle size (D_{50}), even though stream bed sediments are mixtures of a range of particles sizes.

The value of dimensionless critical shear stress for entrainment varies according to differences in the particle size distributions for different mixtures (Parker and Klingeman 1982; Wilcock 1998). Typical experimentally-determined values of τ^* for uniform mixtures are about 0.04, while the critical value drops to near 0.02 for sand and gravel mixtures (Wilcock 1998). For this analysis, we adopted a value of τ^* of 0.03. This value of the Shields parameter has been previously applied as the critical value for incipient gravel motion in the development of flow recommendations for the protection of native fish habitat elsewhere in the Colorado River basin

(Pitlick and Van Steeter 1998). We calculated values of shear stress at riffles using modeled stage-discharges relationships, then computed the corresponding values of dimensionless shear stress using the Shields relationship. We compared the computed dimensionless shear stress values with our Shields criterion of 0.03 to determine the discharges at which boundary shear stress reaches the threshold for bed entrainment at each riffle.

Substrate Mapping

We evaluated spatial and temporal variation in the distribution of bed material by mapping the channel substrate on two occasions. Mapping was discontinued at approximately river km 7 because the bed downstream from that point consists only of sand. We first mapped the substrate in October 1999, then repeated the process in August 2000. Mapping was accomplished by wading down the channel visually observing the bed materials. In areas too deep for visual inspection, bed materials were classified as sand, gravel, or cobbles by feel. The distribution of clean gravel, mixed sand and gravel, cobbles, sand patches, and fines were mapped on mylar taped over 1997 aerial photographs.

Dendrochronology

The ages and germination depths of tamarisk were examined to evaluate the rate and timing of overbank deposition at the Wissiup Return site. Several large tamarisks growing on the terrace near the river were excavated to a depth we estimated to be below the germination point of the plant. The presence, depth, and texture of any stratigraphic layers in the excavation pit were noted, and ground surface height was marked on the tamarisk stems. The plants were then taken to the laboratory where they were cut into slabs, sanded smooth, and examined under a microscope. We considered the germination point of the plant as the point where pith in the center of the stem can no longer be found. The ages of the plants were determined by counting annual growth rings at the germination point.

Table 1: Geomorphic mapping units.

Unit Type	Unit Subtype	Approximate Elevation above Water Surface in Meters	Surface Topography	Surficial Material	Vegetation
Channel					
	WS	0	Water surface	Submerged gravel, cobble, sand, or mud	None
	Low Bars	<1	Slopes toward water surface	Gravel, cobble, sand, or mud	Little or none
High bars					
	High bars	1.5	Ridge and swale	Sand	Sparse tamarisk or cottonwood seedlings, weedy, or bare sand
	Bar Chutes	1	Trough	Gravel, cobble, sand, or mud	Little or none
Floodplain					
	Floodplain	1.5	Undulating or hummocky; may be dissected by scour channels	Silt/clay or sandy	Dense tamarisk or russian-olive. Occasional young cottonwood.
	Swales and channel fills	1	Trough	Silt/clay or sandy	Open with grasses/sedges or wooded
	High Floodplain	1.5-2	Undulating or flat	Silt/clay or sandy	Dense tamarisk or russian-olive. Occasional young cottonwood.
Terrace					
	Cottonwood Terrace	2	Mostly flat	Silt/clay or sandy	Dead or decadent cottonwood with grass or shrubs
	High Scroll	1.5-2	Undulating, may be dissected by scour channels	Silt/clay or sandy	Cottonwood or russian-olive with grass or shrubs
	High Terrace	2-4	Mostly flat; sand dunes may be present	Sandy	Sage, rabbit brush, greasewood
	Tamarisk	3-4	Mostly flat	Silt/clay or sandy	Dense tamarisk

Table 2: Aerial photographs used in the study.

Year	Dates	Discharge ft^3/s	Nominal Scale	Average RMS	Source
1936*	09/07/36	460	1:40,000	7.4 m	National Archives, Wash. D.C.
	09/09/36	326			
1948*	05/22/48	2,430	1:40,000	17.0 m**	National Archives, Wash. D.C.
	06/05/48	1,710			
	10/08/48	90			
1953-5	8/11/53	241	1:80,000	N/A	?
	10/03/55	96			
1961*	08/28/61	8	1:40,000	4.6 m	USDA Aerial Photography Field Office, Salt Lake City, UT
	09/04/61	13			
1969*	10/21/69	478	1:40,000	6.4	USDA Aerial Photography Field Office, Salt Lake City, UT
1980*	06/27/80	2,050	1:40,000	7.4 m	USDA Aerial Photography Field Office, Salt Lake City, UT
1987*	08/16/87	516	1:40,000	8.8 m	USDA Aerial Photography Field Office, Salt Lake City, UT
1993	06/20/93	596	1:40,000	N/A	USDA Aerial Photography Field Office, Salt Lake City, UT
1997*	07/06/97	172	1:40,000	4.5 m	USDA Aerial Photography Field Office, Salt Lake City, UT

*GIS coverage fully developed
** After rubber sheet correction

Table 3: Erosion/deposition classes.

From Surface Type	To Surface Type	E/D Class
Terrace	Channel	E
High bar	Channel	E
Floodplain	Channel	E
Channel Fill	Channel	E
Channel	Bar	D
Channel	Floodplain	D
Channel	Channel Fill	D
Terrace	High bar	E/D
Terrace	Floodplain	E/D
High bar	Floodplain	NC
Floodplain	High bar	NC
Channel	Terrace	X
All other changes		NC

E = erosion, D = deposition, E/D = erosion followed by deposition, NC = no change, X = false polygon

Table 4: Positional errors for individual coverages.

Photo Date	RMS Error	Digitizing Error	Total Positional Error
1948	17.0* m	3.6 m	17.4 m
1961	4.6 m	3.6 m	5.8 m
1969	6.4 m	3.6 m	7.3 m
1980	7.4 m	3.6 m	8.2 m
1987	8.8 m	3.6 m	9.5 m
1997	4.5 m	3.6 m	5.8 m

*After rubber-sheet correction.

Table 4: Potential and uncompensated planimetric error for coverage overlays.

Photo Pair	Total Average Displacement	(ε_{ac}) Expected Uncompensated Planimetric Error (percent of A)	(ε_{au}) Uncertainty about A' (percent of A')
1936, 1948	19.2 m	6 percent	± 11 percent
1948, 1961	18.3 m	6 percent	± 11 percent
1961, 1969	9.4 m	1 percent	± 6 percent
1969, 1980	11.0 m	10 percent	± 12 percent
1980, 1987	12.6 m	2 percent	± 7 percent
1987, 1997	11.1 m	10 percent	± 12 percent

A = measured area of erosion and deposition polygons.

A' = estimated area of erosion and deposition polygons, corrected for uncompensated error.

Table 6: Longitudinal variation in map unit deposit thickness in meters.

Map Unit	Subreaches 19-15			Subreaches 14-12			Subreaches 11-9		
	SH	G	S	SH	G	S	SH	G	S
CT	3.5	2.2	1.3	3.5	2.1	1.4	3.6	2.0	1.6
HT	3.95	1.5	2.45	4.5	0.8	3.7	5.0	0.8	4.2
TT	n/a	n/a	n/a	n/a	n/a	n/a	n/a	n/a	n/a
HB	2.6	1.8	0.8	2.7	1.8	0.9	3.1	1.6	1.5
FP	3.1	1.8	1.3	2.7	1.8	0.9	3.1	1.6	1.5
CF	2.6	0.8	1.6	2.2	0.8	1.4	2.6	0.8	1.6
CH	0.8	0.8	0	0.8	0.8	0	0.8	0.8	0
	Subreach 8			Subreaches 7-6			Subreach 5		
	SH	G	S	SH	G	S	SH	G	S
CT	4.0	1.4	2.6	4.0	0.75	3.25	n/a	n/a	n/a
HT	5.0	0.8	4.2	5.5	0.75	4.75	5.5	0	5.5
TT	n/a	n/a	n/a	5.5	0.75	4.75	5.5	0	5.5
HB	3.1	1.6	1.5	3.0	1.25	1.75	3.0	1.25	1.75
FP	3.1	1.6	1.5	3.6	1.25	2.35	3.6	1.25	2.35
CF	2.6	0.8	1.6	3.6	0.75	2.95	3.6	0.75	2.95
CH	0.8	0.8	0	0.75	0.75	0	0.75	0.75	0

Map Units: CT = Cottonwood Terrace, HT = High Terrace, TT = Tamarisk Terrace,
HB = High Bar, FP = Floodplain, CF = Channel Fill, CH = Channel.
Headings: SH = Ground Surface Height, G = Gravel Thickness, S = Sand Thickness.

Table 7: Percent uncertainty margins (plus or minus) for reporting volumes of gravel erosion, gravel deposition, and changes in gravel storage. Uncertainties in gravel storage changes can be large on a percent basis when the magnitudes of storage changes are small.

Photo Pairs	1936 to 1948			1948 to 1961			1961 to 1969			1969 to 1980			1980 to 1987			1987 to 1997		
Subreach	δV_{gi}	$\delta V'_{gi}$	$\delta \Delta S_i$	δV_{gi}	$\delta V'_{gi}$	$\delta \Delta S_i$	δV_{gi}	$\delta V'_{gi}$	$\delta \Delta S_i$	δV_{gi}	$\delta V'_{gi}$	$\delta \Delta S_i$	δV_{gi}	$\delta V'_{gi}$	$\delta \Delta S_i$	δV_{gi}	$\delta V'_{gi}$	$\delta \Delta S_i$
5	na	na	na	na	44	44	na	48	48	67	67	86	66	58	88	67	62	187
6	na	na	na	66	66	162	66	66	86	67	67	635	66	66	78	67	67	400
7	66	66	70	66	66	107	66	66	139	67	67	127	66	66	81	67	67	103
8	51	42	46	42	41	118	46	41	363	54	43	110	51	41	254	43	43	238
9	33	40	100	30	41	52	35	41	169	37	43	51	28	41	252	39	43	135
10	34	39	629	34	41	133	32	41	88	40	435	63	32	41	151	34	43	217
11	30	41	113	32	41	255	32	41	56	33	43	3259	29	41	110	41	43	60
12	30	30	566	27	33	169	30	33	53	30	35	97	29	33	85	33	35	186
13	27	32	61	30	33	68	31	33	74	32	35	79	31	33	1152	32	35	174
14	28	32	196	27	33	131	28	33	40	31	35	228	28	33	377	32	35	205
15	33	33	476	25	33	87	27	33	38	27	35	139	25	33	163	28	35	63
16	31	28	64	28	33	45	26	33	35	28	35	109	26	33	270	33	35	64
17	27	29	1977	31	33	120	30	33	47	34	35	207	31	33	103	34	35	114
18	29	29	48	27	33	37	31	33	62	27	35	193	30	33	60	28	35	204
19	24	31	107	30	33	67	32	33	1078	31	35	97	26	33	199	32	35	89

δV_{gi} = uncertainty in erosion volume; $\delta V'_{gi}$ = uncertainty in deposition volume; $\delta \Delta S_i$ = uncertainty in storage change volume.

Table 8: Roughness parameters used at cross sections for high-flow extrapolation.

24-hour Camp		Above Pipeline		Wissiup Return	
Station	n	Station	n	Station	n
905	0.030	424	0.028	424	0.028
848	0.036	353	0.042	353	0.040
803	0.030	291	0.027	291	0.027
750	0.030	249	0.024	249	0.024
709	0.031	193	0.024	193	0.024
668	0.030	165	0.025	165	0.025
628	0.030	138	0.025	138	0.025
554	0.030	106	0.025	106	0.025
470	0.030	36	0.025	36	0.025
419	0.040	0	0.026	0	0.026
341	0.026				
250	0.029				
36	0.034				
0	0.031				

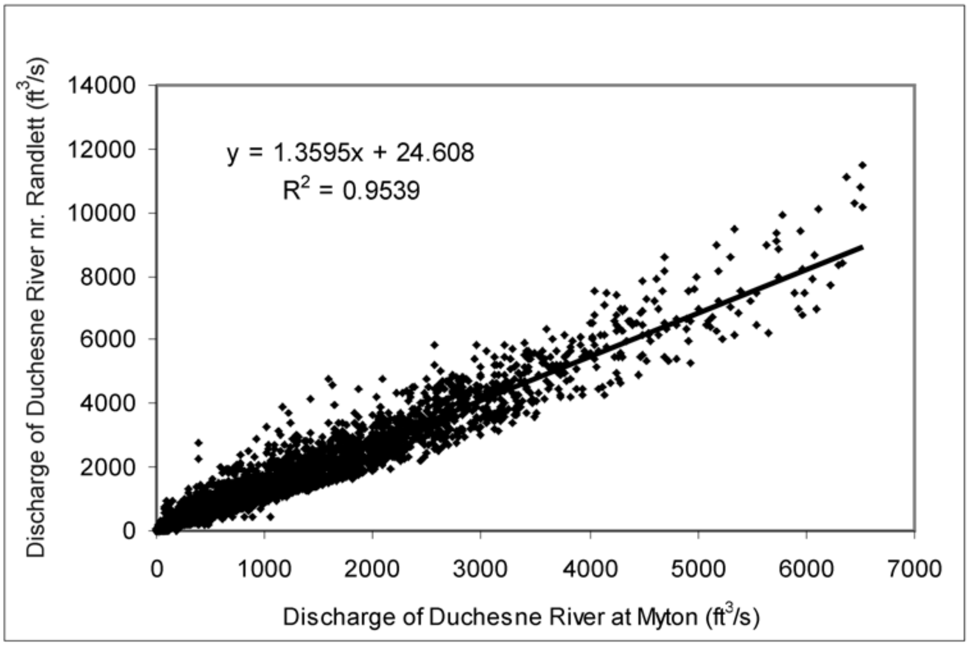


Figure 3: Graph comparing daily discharge at the Myton and Randlett gaging stations.

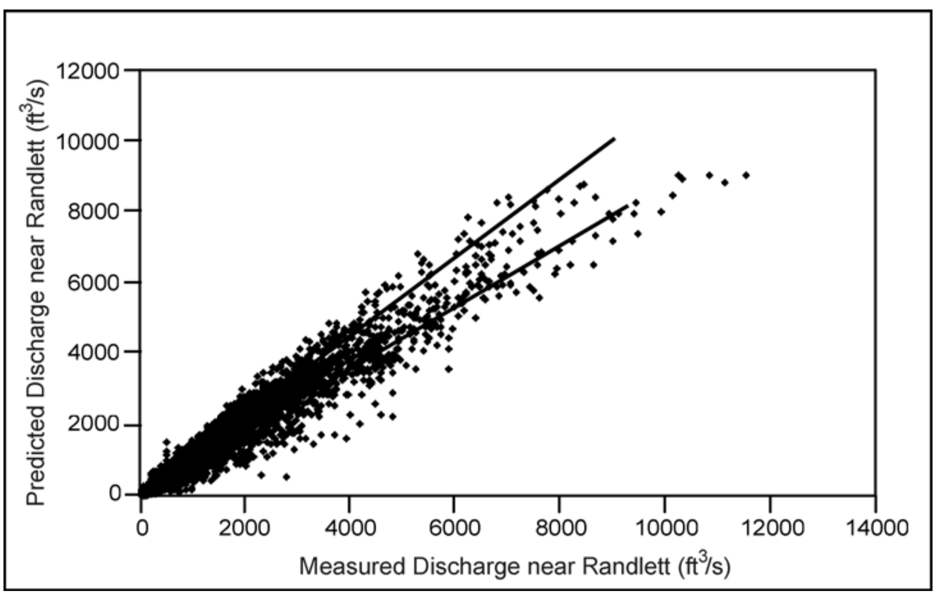


Figure 4: Graph comparing observed and predicted daily discharge values for the Randlett gage for the period 1943-2000.

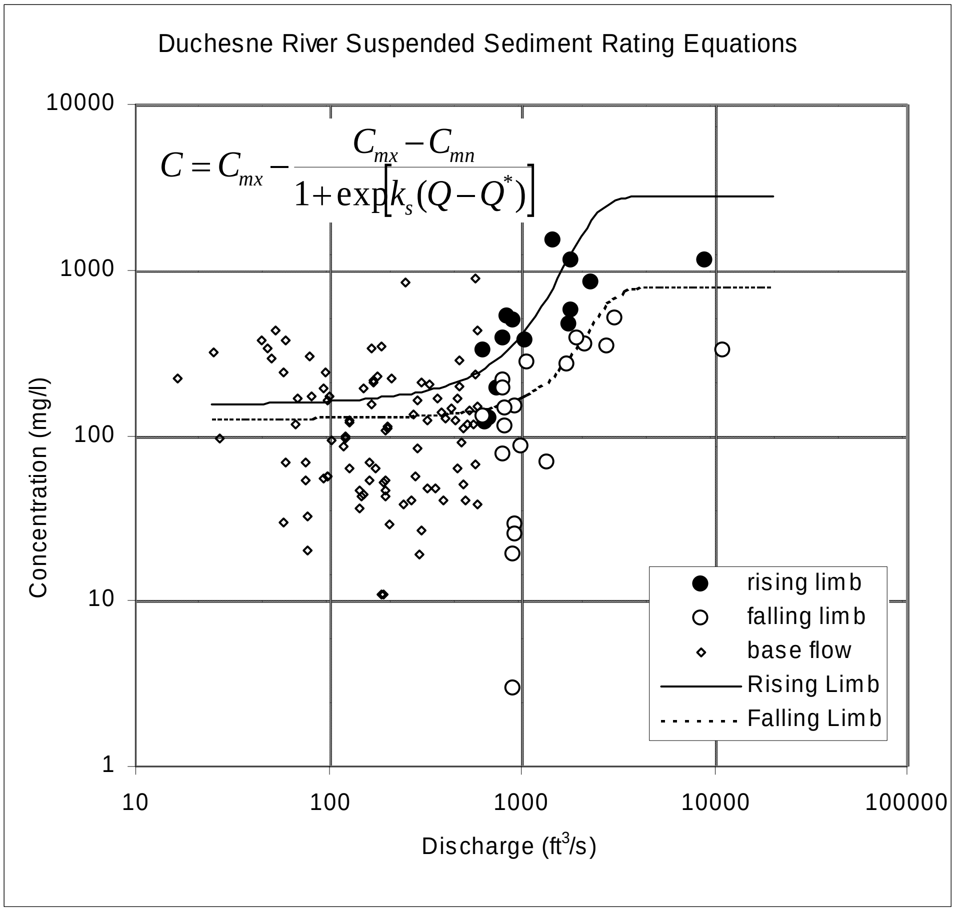


Figure 5: Graph showing suspended sediment rating curves for rising and falling limb concentrations.

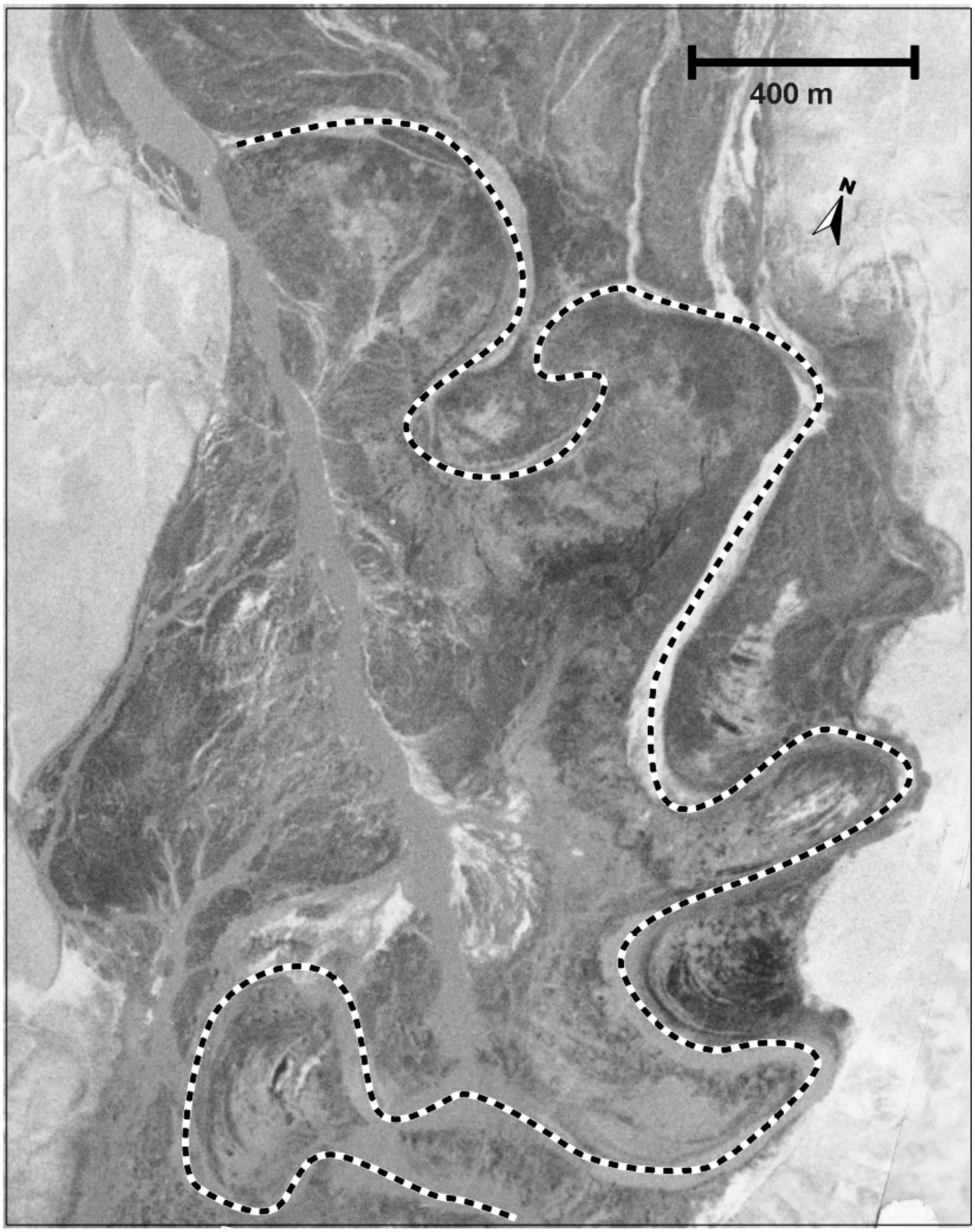


Figure 6: 1948 photograph of subreach 2 showing avulsion in progress. The pre-avulsion channel is highlighted with a dashed line.

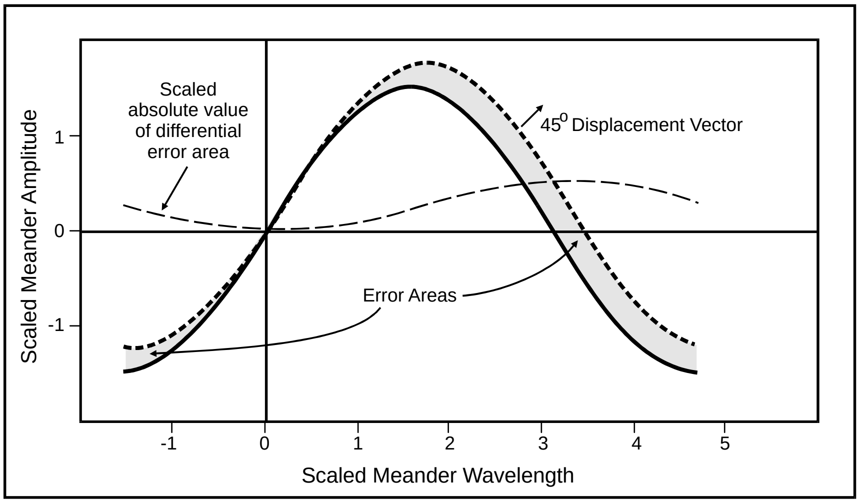


Figure 7: Diagram of error area generated by displacement error of a sinusoid curve.

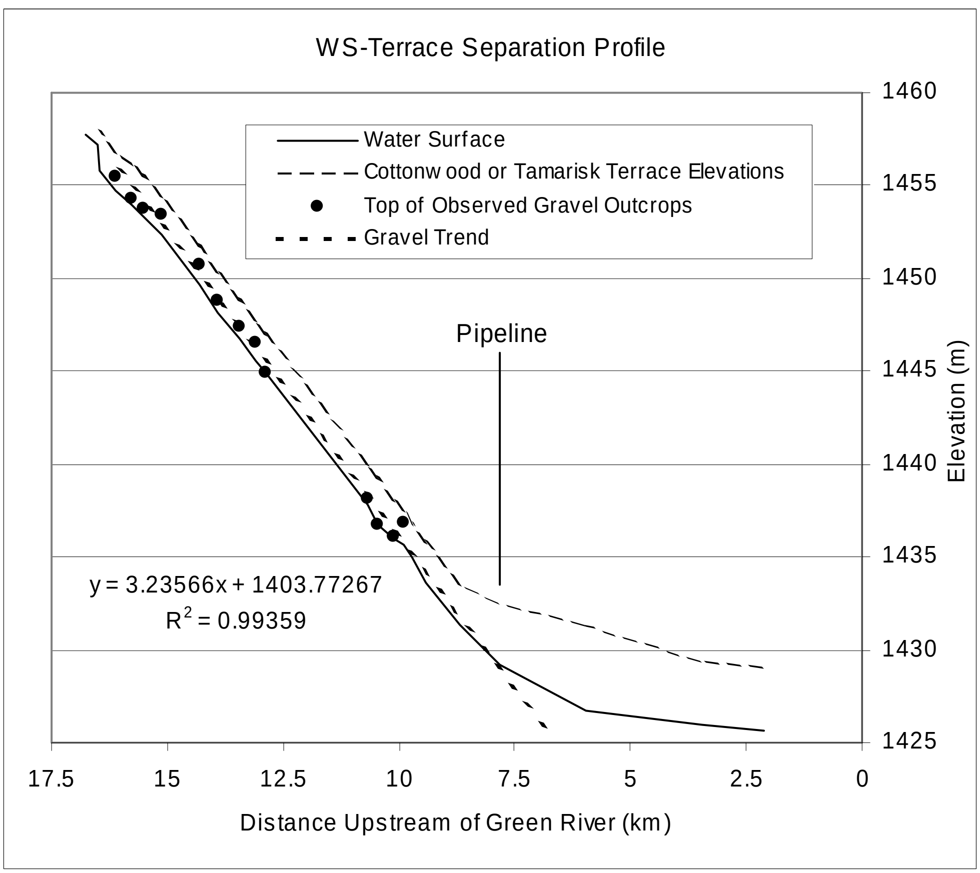


Figure 8: Longitudinal profile of bank and gravel deposit elevations. The equation and R^2 shown on the graph refers to a regression line fit to the top of observed gravel outcrops.

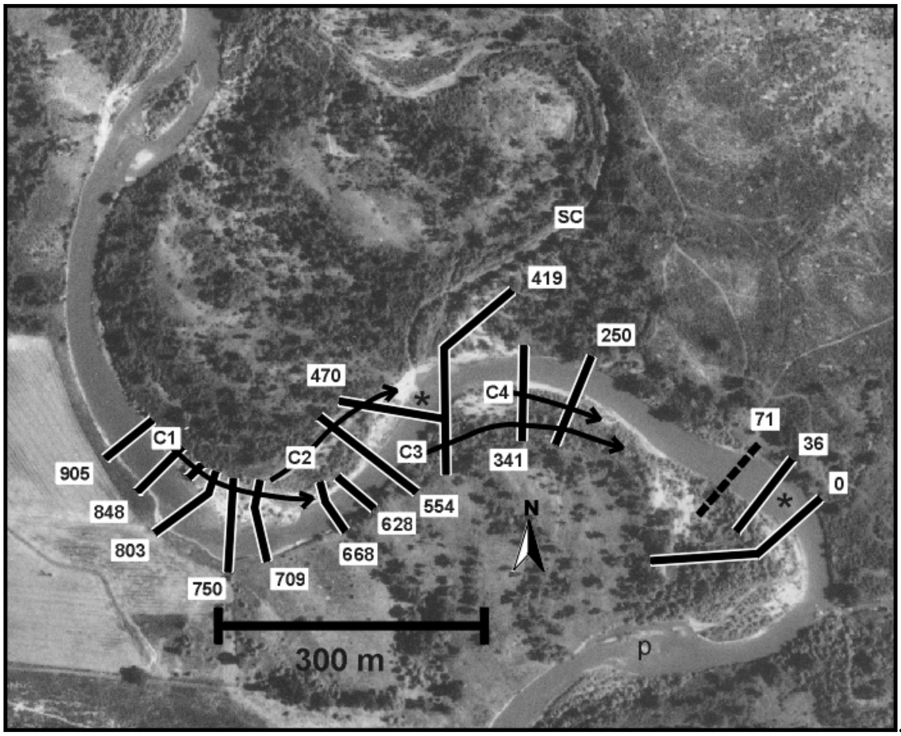


Figure 9: Photomap of the 24-hr Camp detailed site indicating cross section locations and main chutes.

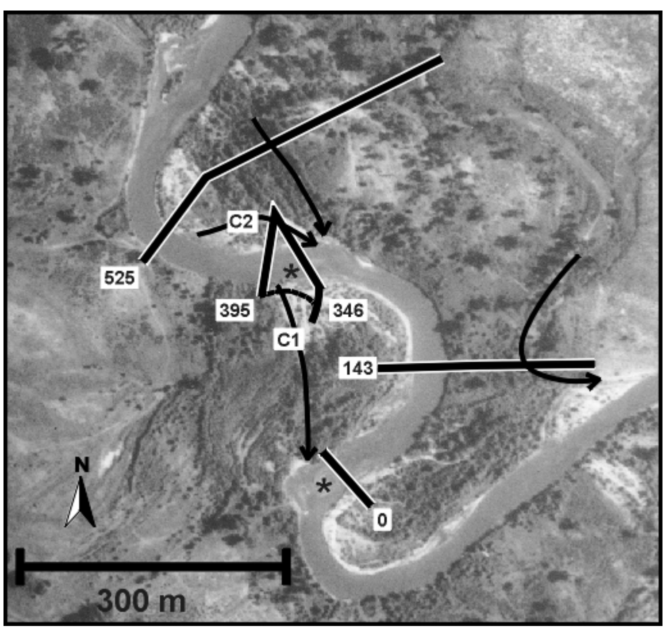


Figure 10: Photomap of the Above Pipeline detailed site indicating cross section locations and main chutes.

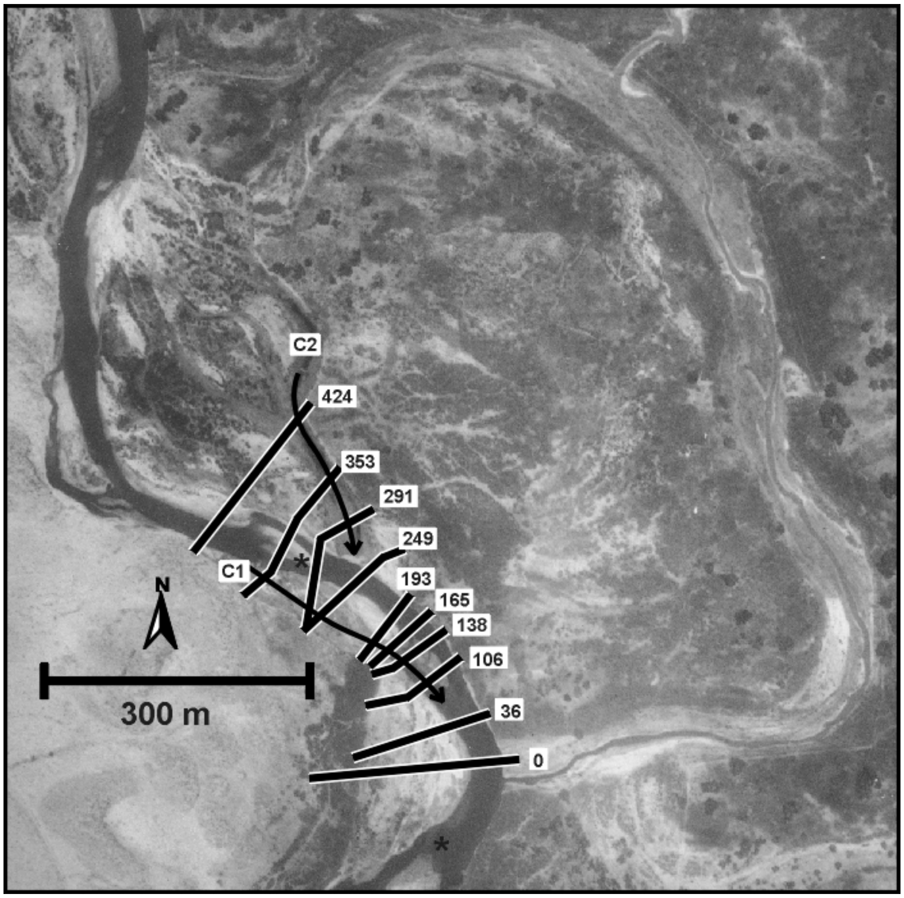
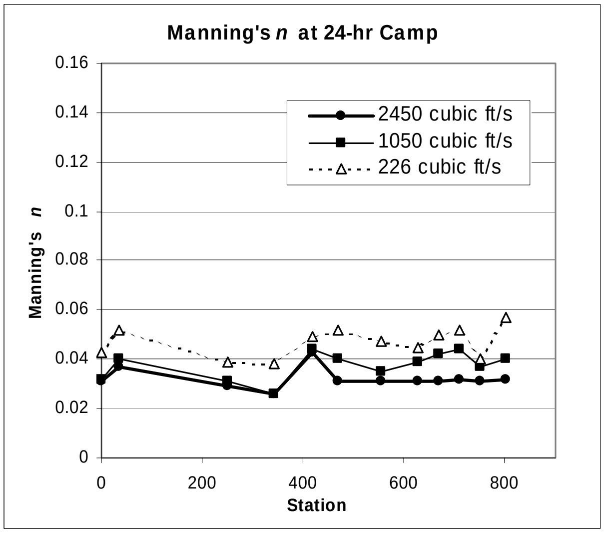
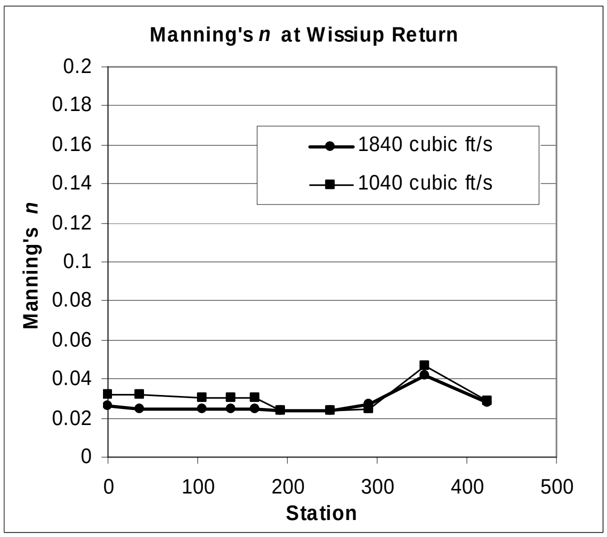


Figure 11: Photomap of the Wissiup Return detailed site indicating cross section locations and main chutes.





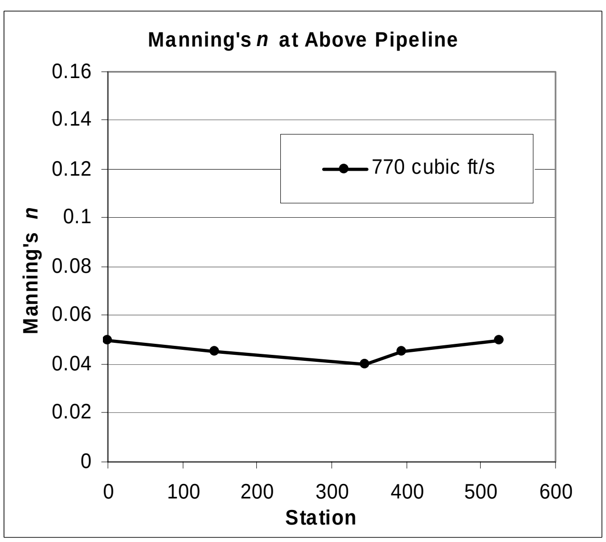


Figure 12: Graphs showing Manning's n calibration values for each station and discharge.

RESULTS

Hydrology of the Duchesne River in the Study Area

Water flow through the study area is significantly less today than it was prior to the mid-1920s (Figure 13). This finding is consistent with other analyses of gaging stations elsewhere in the Colorado River basin where flow reductions may be related to climatic factors (Stockton 1976; Graf 1978; Allred and Schmidt 1999). Some portion of the decrease in the Duchesne River basin is due to increased water diversions for the Strawberry Valley Project. This project, which was the first large-scale transbasin diversion impacting the Duchesne River, started operation in 1922 (US Fish and Wildlife Service 1998). The total annual flow, magnitude of floods, and the magnitude of flow of various durations all were much greater for the period between 1912 and 1924 than for the period between 1925 and 2000. Stream flow prior to 1943 was estimated using the MOVE1 method (Hirsch 1982), as described above. This extension of the stream-flow record was based on correlation with the record of stream flow measured at the Duchesne River gage at Myton. The estimated average total annual stream flow between 1912 and 1924 was 877,060 acre-feet, which is more than twice the average for the gaged period between 1943 and 2000. A comparison of the flow duration characteristics and flood magnitudes for the two periods, as computed from annual maximum daily mean discharges, show that the magnitude of flood flows has also decreased greatly from the early part of the 20th century. The magnitudes of the 1.5-year and 2-year recurrence floods have decreased by 62 percent and 56 percent respectively, and the magnitude of the 5-year recurrence flood has decreased by about 40 percent (Table 9). Base flows have also decreased greatly. The magnitude of the flow exceeded 90 percent of the time decreased from 390 ft^3/s to 60 ft^3/s. The decrease in stream flow after the mid-1920s occurred rapidly, so that average annual runoff and the durations and frequencies of large flows during 1925 to 1942 are similar to those of 1943 to 2000. These two periods are separated in this analysis because actual stream gaging did not begin at the Randlett gage until water year 1943.

Total annual runoff and the flood magnitudes were highly variable over the period of record. The years with total annual runoff in the upper quartile of the measured record averaged 5.8 times the mean annual flow of the years with total annual runoff in the lowest quartile of the distribution. The magnitudes of 1.5-year, 2-year, and 5-year floods, and the flow exceeded 10, 50, and 90 percent of the time were all at least five times greater in upper-quartile years than in lower-quartile years as well (Table 10). Wetter and drier years appear to cluster as a cyclic pattern in the latter half of the 20th century, when three relatively dry periods lasting roughly a decade were separated by two relatively wet periods, each lasting about 5 years (Figure 13). The dry periods occurred from 1954 to 1964, 1970 to 1982, and 1988 to 1996. Wetter conditions prevailed from 1965 to 1969 and from 1983 to 1987. Both total discharge and flood magnitudes deviated significantly from longer-term values during these wet and dry cycles. Total annual runoff during the dry periods is about 40 percent of total the annual runoff during the wet periods (Table 11). Flood magnitudes during dry periods are typically about 70 percent of the post-1943 flood magnitudes, while flood magnitudes during the wet periods are more than 150 percent of the post-1943 figures.

The Bonneville Unit of the Central Utah Project (CUP) began diverting water into Strawberry Reservoir in December 1971 (CH2M Hill 1997). Mean annual runoff for water years 1972 through 2000, after the Bonneville Unit began operations, declined by approximately 42,600 acre-ft (10 percent) from the mean annual runoff for the pre-project water years of 1943 through 1971 (Table 12). CUP diversions averaged about 33,400 acre-ft annually between 1972 and 1995 (CH2M Hill 1997). Diversion increased substantially about 1989, when the project collection system was completed. Before 1989, annual diversion averaged about 11,400 acre-ft, whereas the average annual diversion for 1989 through 1995 was 86,800 acre-ft. If the average annual diversions during the remaining five years from 1996 through 2000 were assumed similar to the average for 1989 through 1995, average CUP diversions for 1972-2000 would have been 42,600. This is nearly identical to the decrease in stream flow between the pre-CUP and post-CUP periods, and evidence that the decrease in stream flow was mainly due to diversion. The increase in CUP diversions after 1988 also coincided with a period of extremely low in-stream flow in which peak flows failed to attain the channel-forming threshold for 7 consecutive years.

The impact of the CUP is most pronounced during drier-than-average years. Of the 14 driest years during the 1943 to 2000 period of record (the lower quartile), nine occurred after 1972. Mean annual flow for these nine years since 1972 is 33 percent less than the mean annual flow for the five years before 1972. Operation of the CUP has had a smaller effect on flows during wetter years. Mean annual flow for years in the middle two quartiles of total annual flow is just 3.2 percent less since 1972 than mean annual flow for middle-quartile years before 1972.

The frequency of upper-quartile years has remained constant during the pre-project and post-project time periods, and mean annual flow actually increased by 20 percent in the upper-quartile years after 1972. The frequencies and magnitudes of large discharges decreased less than the frequencies and magnitudes of moderate discharges after 1972. The magnitudes of 1.5-year and 2-year floods and of the flow exceeded 50 percent of the time decreased by more than 25 percent, while the magnitude of the 5-year flood decreased by only about 5 percent.

Because we analyzed geomorphic change during periods defined by available aerial photography, we also calculated the characteristics of stream flow for these periods. Most photograph intervals span portions of the previously-described dry and wet periods. Stream-flow measurements divided according to photography intervals show that the periods between 1936 and 1948, 1961 and 1969, and 1980 and 1987 embraced periods of high total annual runoff, although peak discharges in between 1936 and 1948 were moderate (Table 13). The magnitudes of 10-percent exceedence flows for these three time periods were similar. The intervals between 1948 and 1961 and 1969 and 1980 represent moderate periods with high-flow durations of approximately 60 percent of those in the three wetter periods. Although the 1948 to 1961 interval was generally dry to average, the interval also includes the particularly high runoff year of 1952. Water year 1969 was the wettest year within the 1969 to 1980 interval, which otherwise corresponds with the dry cycle recognized between 1970 and 1982. Another dry cycle from 1988 through 1996 was well captured by the photograph interval of 1987 to 1993, which stands out as being exceedingly dry. The final photograph interval of 1993 to 1997 was also characterized by generally low flow conditions, as indicated by the small magnitudes of its 90- and 50-percent exceedence flows. Nonetheless, this short period contains the relatively large peak event of 1995.

Delivery of Fine Sediment into the Study Area

Sources of fine-grained sediment are abundant in the Uinta Basin. Sand- and silt-sized material is delivered to stream channels from hillslope erosion, bank erosion, gullies, and return flows from irrigation canals. The fine sediment that enters the study reach at Randlett is transported downstream towards Ouray. Channel adjustment in the study reach, as well as associated changes in aquatic habitat, are caused by changes in the balance among the capacity of the Duchesne River to transport this load, the load delivered from upstream, and local erosion and deposition in the study area.

Sediment Delivery to the Study Reach from Upstream

Estimation of the annual suspended sediment load delivered from upstream reaches provides a baseline annual sediment load against which the magnitude of channel change can be compared. We developed suspended sediment rating curves for rising and falling limbs of the annual hydrograph, based on 136 USGS measurements of suspended-sediment concentration at the Duchesne gaging station near Randlett. We determined that rising-limb suspended-sediment concentrations typically reach values approximately three times the values observed on the falling limb. When applied to the daily mean discharge records, these ratings relations provide estimates of annual suspended sediment fluxes into the study area. The average annual suspended sediment flux at the gage near Randlett, as calculated over the period from 1943 to 2000, is estimated at 405,350 tons per year. Loads calculated from flow records separated into upper-quartile, lower-quartile, and middle-quartile classes suggest that the majority of transport occurs during years with annual runoff totals in the upper quartile of the stream flow record. Total annual suspended sediment load calculated for upper-quartile years exceeded the quantity calculated for years in the middle two quartiles by a factor of 3.5, and exceeded the quantity calculated for lower-quartile years by a factor of 26 (Table 14).

Similar analyses using pre- and post-1925 hydrology, and assuming no change in the sediment rating relations, allow comparison between the potential sediment transport capacities of the Duchesne River before and after the shift from higher to lower discharges in the mid-1920s. Calculations suggest that the average annual sediment load entering the study area between 1912 and 1924 was 3.2 times greater than the average annual load calculated from gaged flows between 1943 and 2000. Average sediment flux during 1912-1924 is estimated at 1,285,250 tons, compared to 405,350 tons for the gaged record. The pre-1925 average annual sediment load appears to have been about 25 percent greater than wet-year loads during the modern flow regime.

Another decrease in the estimated mean annual sediment load transported through the lower Duchesne River occurred after the Bonneville Unit of CUP began operations in water year

1972. The estimated annual load for water years 1972 through 2000 is 347,200 tons, which is 25 percent less than the estimated annual load for years 1943 through 1971 of 463,600 tons.

Local Sources of Fine-grained Sediment

The magnitudes of fine sediment inputs from the Uinta Basin to the study reach have been large at times in the past. Dry Gulch, which enters the Uinta River approximately 5 km upstream from the study area, is one such source. The main channel of this small drainage is currently a large gully extending approximately 15 km upstream from the Uinta River to near Cottonwood Creek, just south of Roosevelt. We conservatively estimate that the gully averages 5 m in depth and 15 m in width, implying that at least 1.1 million cubic meters of material have been removed from this drainage since initiation of the gullying.

The majority of erosion in the Dry Gulch gully probably occurred during a few years sometime around 1930. Our best information concerning the timing of Dry Gulch incision comes from an eyewitness account. According to Brett Prevedel of the Roosevelt office of the Natural Resources and Conservation Service (personal communication), long-time resident Orlan Cook recalls that the gully developed within a 3-4 year period when Mr. Cook was in his late teens. Mr. Cook passed away in 2001 at the age of 85, placing the probable time of gully development at around 1930-1935. Our independent evidence corroborates this account. Historical aerial photographs clearly show Dry Gulch to be significantly entrenched in 1961. Although poorer image resolution on the 1948 photographs makes a definitive determination of the condition of Dry Gulch difficult, the incision appears to be present at that time as well. The cause of the sudden incision at Dry Gulch is unknown at this time, but use of this drainage as a conduit for irrigation return flows may have been a factor. Mature Russian-olive stands currently found along the banks of Dry Gulch indicate that the gully has been relatively stable in recent times. Assuming that the bulk of Dry Gulch incision occurred during a 10-year period, a sediment porosity of 0.3, and a delivery ratio to the Duchesne River of 0.5, the annual sediment load into the study area from Dry Gulch alone is conservatively estimated at 38,500 m^3/yr of solids. This amount of sediment constitutes about 40 percent of the entire annual sediment flux for the Duchesne River during a middle-quartile runoff year (98,800 m^3/yr), and is 2.9 times greater than the annual load calculated for lower-quartile years (13,300 m^3/yr). In short, during the time of its incision, Dry Gulch constituted a substantial local sediment source for the study area. Several other potentially significant gullies and ditches exist within the study area as well. The quantity of sediment delivered from these sources is not directly tied to the magnitude of discharge in the Duchesne River. Much of the gully erosion in the area occurred during a time of declining discharge after the 1920s.

Cutbank erosion along the Duchesne River in the reach just upstream from the Uinta River is another potentially large source of local sediment production. The river runs adjacent to 30-m-high bluffs forming the north edge of Leland Bench through much of the reach between the Uinta River and Red Bridge. Aerial photographs and field observations indicate that the stream is actively eroding into the base of the bluffs in some locations.

Uncertainty in Estimated Suspended Sediment Loads

The 136 measurements of suspended sediment concentration used to develop the rating relations described above comprise all existing suspended sediment data for the lower Duchesne River. Of these measurements, only two were taken at discharges greater than the 1.5-year flood. Therefore, significant uncertainty exists regarding the accuracy of the ratings relations for high discharges when a disproportionately large quantity of sediment is transported. Alternative methods of estimating the average suspended sediment flux in the lower Duchesne River suggest that estimates derived from the rating relations proposed here may be low.

A partial record of daily suspended sediment concentrations exists for the Green River gaging station at Jensen covering the period between 1948 and 1968. The Jensen gage is upstream from the mouth of the Duchesne River. Ashley Creek and Brush Creek are the only perennial tributaries intervening between the gage and the Duchesne River. Another partial daily suspended sediment record exists for the Green River gaging station at Ouray, located immediately downstream from the mouths of the Duchesne and White Rivers, for the period between 1957 and 1966. The average annual load for all records at the Ouray gage between August 1957 and March 1966 is 10.2 million tons, and the average annual load for all records at the Jensen gage is about 5.5 million tons. The mean annual loads were calculated by multiplying the mean of all daily sediment load records for each gage by 365. Subtracting the calculated average annual load at the Jensen gage from the load calculated for the Ouray gage implies that the total load supplied by the White River, Duchesne River, and all other ungaged areas averages about 4.7 million tons per year.

This estimate is reasonably consistent with other estimates of suspended sediment contributions to the Green River from all sources between the Jensen gage and Ouray gage. Andrews (1986) reported that an average of 5.87 million tons of suspended sediment was delivered to the Green River from the Duchesne River, White River, and ungaged tributaries between the Jensen and Ouray gages. Andrews' analysis was based on measurements made between 1947 and 1979 at the Jensen gage, 1951 and 1966 at the Ouray gage, and 1974 and 1982 at the mouth of the White River. Data reported by Iorns and others (1964) suggest an annual figure of 5.5 million tons for water years 1952 through 1955. There is reason to suspect that the estimates of both Andrews (1986) and Iorns and others (1964) somewhat overstate the average annual suspended sediment flux. Error margins for estimated suspended sediment loads were not reported in either analysis.

The differences obtained by Iorns and other (1964) were limited to only four years of overlapping records at both Green River gages. In three of the years, the reported differences were 3.4, 3.4, and 4.1 million tons. The average for the four years was greatly increased by just one of the four years, water year 1952 for which Iorns reported a suspended sediment load difference between the two gages of 11.1 million tons. Water year 1952 was extremely wet. The second highest total annual discharge ever measured at the Randlett gage was recorded during water year 1952. The year also included the fourth largest peak discharge on record, an approximately 15-year event.

Of the 5.87 million tons estimated by Andrews (1986) to enter the Green River each year between the Jensen and Ouray gages, 1.67 million tons per year were attributed to the White River, and 4.2 million tons per year were attributed to the Duchesne River and the other ungaged areas. The USGS sampled suspended sediment concentration at the mouth of the White River daily during water years 1976 through 1983. Presumably using the same USGS data used by Andrews, we calculate that the average annual suspended load discharged at the mouth of the White River during these years was about 2.4 million tons rather than the 1.67 million tons reported by Andrews. The source of this discrepancy is unknown. More importantly, it is unclear how Andrews (1986) arrived at the estimated suspended load of 4.2 million tons per year for the Duchesne River and other ungaged areas. If the estimated flux for these ungaged areas was derived as a residual, the resulting figure incorporates any error associated with the White River estimates, as well as with all other terms (Kondolf and Matthews 1991).

We assume that our estimated average annual suspended sediment load of 4.7 million tons for the combined White River, Duchesne River, and other ungaged areas, as well as our estimate of 2.4 million tons for the White River alone, are approximately correct. The Duchesne River and all other ungaged areas therefore contribute an average of 2.3 million tons per year of suspended sediment to the Green River. It is reasonable to assume that the contribution of the other ungaged areas contribute at least 550,000 tons of suspended sediment per year to the Green River. We estimated the ungaged area outside of the Duchesne River basin at 3,374 km^2 by summing the areas of the Ashley-Brush hydrologic unit (USGS Hydrologic Unit Code 14060002) and two-thirds of the Lower Green—Diamond hydrologic unit (USGS HUC 14060001). A total sediment yield of 550,000 tons per year from this area would require a unit sediment yield of 163 tons per km^2 per year. This figure is quite plausible when compared with a unit yield for the White River basin of 181 tons per km^2 per year, as calculated by dividing our estimated average annual flux at the mouth of the White River by basin area. The average suspended sediment flux at the mouth of the lower Duchesne River can then be calculated as a residual equal to about 1.75 million tons per year.

Thus, our estimated average annual suspended sediment load of 405,350 tons for the lower Duchesne River, as determined from suspended sediment concentration measurements at the Randlett gage, is at least four-fold less than other estimates. The only measurements of suspended sediment concentration at discharges greater than the 1.5-year flood discharge were made on June 2, 1986 and June 22, 1983 when discharges were 9,000 ft^3/s on the rising limb of the annual hydrograph and 11,020 ft^3/s on the falling limb, respectively. The concentrations measured on these two dates were 1,150 mg/L in 1986 and 322 mg/L in 1983. Both values are much lower than would be predicted by extrapolation of the remaining data (Figure 5). Typical suspended sediment concentrations at high discharges may be much larger than the values obtained by our rating curves. If so, these curve may significantly underestimate the quantity of sediment transported by infrequent high discharge events or during particularly wet years. An extended sampling program to monitor suspended sediment concentrations in the lower Duchesne River during peak flow events may be required to resolve this issue.

Longitudinal Classification of the River System

We divided the study area into 19 subreaches in which local metrics of channel morphology and change were calculated. We then grouped these subreaches into four longitudinal zones with similar present characteristics and histories of channel change. Zones were numbered consecutively, with Zone 1 referring to the most downstream reach near the Green River and Zone 4 referring to the most upstream part of the study area. Zone 1 was subdivided into Zones 1a and 1b (Figure 14). Zones, their constituent subreaches, and zone designation criteria are listed in Table 15. The basis for zone designations is discussed below.

Longitudinal Profile

The longitudinal profile of the Duchesne River and its valley is composed of two parts within the study area. Channel slope upstream from river km 9 (valley km 7) averages about 0.0019. Slope is about 0.00134 near the upstream end of the study area and gradually steepens to about 0.00268 at river km 9. Downstream from river km 9, the channel has a much flatter slope averaging 0.00024 (Figure 15). This eight-fold decrease in channel slope between the upstream and downstream parts of the study area occurs near the boundary between Zone 1b and Zone 2. A break-in-slope in the valley gradient occurs at nearly the same location. Valley gradient averages 0.0032 upstream from the break and 0.00033 downstream from the break. Elevations for constructing these longitudinal profiles were derived both from 1:24000 topographic maps and from data extracted from georeferenced water depth and bed elevation grids developed by LC Headwaters, Inc.

Characteristics of the Alluvial Valley

The relatively flat valley floor contains an active floodplain and point bars, plus at least three distinct terrace surfaces, some of which functioned as active floodplains within historical time, as indicated by extant vegetation, early aerial photography, and early cadastral surveys. The active floodplain now occupies less than a quarter of the full valley width. Terrace surfaces within the study area are designated, in ascending order of elevation, as the cottonwood terrace, high terrace, and tamarisk terrace (Figure 16). The cottonwood terrace is the dominant terrace surface through Zones 3 and 4. Much of this surface still has the remains of a cottonwood gallery forest. However, many of the cottonwoods on this surface are dead or dying and much of the surface is covered with xeric vegetation such as rabbitbrush, sagebrush, and squaw bush.

This surface contains frequent swales representing abandoned channels that have filled to a level of about 1 m below the general terrace elevation. The cottonwood terrace is composed of channel gravels overlain by a layer of sand or silt averaging about 1.5 m in thickness. It stands about 2 m above the low-flow water level through Zones 3 and 4. Its elevation rapidly increases downstream from the Pipeline, and the cottonwood terrace merges with the tamarisk terrace in Zone 2.

First appearing in Zone 2, the tamarisk terrace extends downstream through Zone 1 to the Green River. This terrace level is essentially flat, and stands at least 4 m above the low flow water surface. Much of its surface is covered with dense tamarisk thickets. Bank exposures of the sediments underlying the tamarisk terrace show that it is composed of fine-grained sediment. No gravel is exposed in cutbanks. Although we have designated this surface is as a terrace, it is more properly considered a floodplain at the far downstream end of the study area where backwater ponding from the Green River promotes flooding onto this surface. This area could be described as either a lateral migration/backswamp floodplain (order B3c) or perhaps a laterally stable, single-channel floodplain (order C1) using the floodplain classification system of Nanson and Croke (1992). However, flooding of this surface by river water becomes rare farther upstream in Zone 1b, and probably never occurs in Zone 2.

The third terrace surface identified, the high terrace, runs along the southwestern margin of the alluvial valley and therefore appears only on river right. Sagebrush is the characteristic vegetation of this unit through most of the study area. Sand dunes are scattered on its surface, giving it uneven topography. It has a significant component of slope toward the valley center so that its height above the river varies according to the distance each bend of the river has eroded into it laterally. On average, it is about 3 m above the river through the upstream reaches and roughly matches the elevation of the tamarisk terrace in the downstream reaches. The exposed stratigraphy of the high terrace consists of fine-grained material. The active floodplain and point bars of the Zones 2, 3, and 4 are set within these terraces at levels at approximately 1.5 m above the low-flow water surface (Figure 17). This floodplain area is best described as a wandering gravel-bed river floodplain (order B2). Chute channels commonly define the boundary between the bars and vegetated floodplain surfaces. Numerous abandoned channels and swales are present. Riparian vegetation consists primarily of tamarisk, russian-olive, and young cottonwood. Downstream from Zone 2 the active bar/floodplain surface is transformed into a narrow bench that rises in elevation downstream. By survey site 4, the bench has approached the level of the tamarisk terrace (Figure 18).

Characteristics of the Channel

The channel of the Duchesne River is alluvial, in the sense that the channel bed and banks are composed of the bedload and suspended load materials that the river transports. The river occasionally encounters fine-grained deposits that may be of colluvial origin where southwesterly-swinging meander bends extend into the high terrace. Contact with colluvium or bedrock on the northeast side of the valley is limited to the apexes of two short bends near the Oil Shack. Through Zones 2-4, the channel has a meandering planform, and a mixed bed of cobbles, gravel, and sand. We conducted surface pebble counts on nine different riffles at six sites distributed through Zones 2-4. The locations of these sites, 24-hour Camp, river km 20.5, at the Bowtie, Below Bowtie, Above Pipeline, and Wissiup Return, are indicated on (Figure 14). Particle size distributions at each site show that the median grain size of the riffles is between 30 and 75 mm (Figure 19, Table 16). The bed surface also includes patches of surface sand, which occur along the insides of bends near point bars. The point bars themselves are capped with a meter or more thickness of sand.

Bulk subsurface bed sampling at four locations indicates that bed material consists of coarse to very coarse gravel mixed with 13 to 20 percent sand, with sand content generally increasing in the downstream direction (Figure 20, Table 17). The median grain size of the gravel fraction of the subsurface is approximately the same as the median grain size of the surface pavement at 24-hour Camp. Subsurface median grain size ranges from 60 percent to 68 percent the size of the surface pavement median grain size at the other three sites.

Crossover riffles are typically spaced 8-10 channel widths apart at the inflection points between bends. Smaller riffles or shoals commonly occur within bends as well, and have led to the development of lobed or compound bends in many locations. Eddies with deep scour holes in alcoves eroded in the outer banks of bends may also occur just downstream from mid-bend riffles. Where chutes and secondary channels occur, they invariably leave the main channel just upstream from a riffle. Chute mouths are generally found just downstream from the riffles. These reentrant areas downstream from riffles commonly contain scour holes, and may also be the sites of backwater embayments or eddies. These backwater areas tend to accumulate silty sediments during periods of low flow. Samples of this bottom material taken from pools and main channel backwater areas prior to the 2001 runoff season were composed of approximately 60 percent silt and finer materials, on average.

The river has historically been active through much of the gravel-bed portion, and remains so to the present day. Outer banks at bends are eroding throughout the gravel-bed reaches, as shown by the presence of vertical cutbanks along the majority of most bends. Recent channel activity has been most pronounced near the Bowtie, a complex area consisting of cut-off meander loops on either side of the present channel. The eastern bend comprising the Bowtie was cut off between 1993 and 1997 and was subsequently reconfigured as recently as 1999. The presence of large mobile gravel bars at the Bowtie and through the first few bends downstream suggests that additional cutoffs are imminent in this reach.

The character of the Duchesne River changes abruptly near the downstream end of Zone 2. Bed material changes from gravel to sand, channel gradient flattens, and the channel assumes a deep, narrow cross section. The transition in bed material and channel form begins at approximately river km 9 where channel and valley gradients flatten, and is fully complete by river km 7 just downstream of the sharp bend at Grey Bluff. The pool-riffle channel morphology with wide point bars and a complex shoreline found in the upstream part of the study area is replaced by a simple canal-like channel with steep, well-vegetated banks standing in excess of 4 m above the low flow water surface The mean high-flow channel width in Zones 2-4 is 73 m, while the mean channel width in Zone 1 is 43.7 m. Sinuosity is low downstream from the transition, where essentially straight stretches extend for up to 2 km before being interrupted by sharp bends.

Discharges Necessary to Access High Bars and Secondary Channels

One-dimensional hydraulic modeling was used at three detailed study sites to estimate the discharges necessary to inundate the various topographic surfaces of the alluvial valley of Zones 2, 3, and 4. Several lines of evidence indicate that only rare large-magnitude floods reach the level of the cottonwood terrace that comprises the largest part of the alluvial valley. Flooding of this surface occurs primarily where lower-lying areas are connected to the channel by partly-filled abandoned channels. Thus, we restrict our model predictions to evaluation of the discharges necessary to access high bar surfaces and initiate flow through chutes and secondary channels. HEC-RAS models were constructed and calibrated for the 24-hour Camp, Above Pipeline, and Wissiup Return reaches.

24-hour Camp

Our analysis at 24-hour Camp indicates that out-of-channel flow is initiated at localized points well before the river is flowing at an overbank level throughout the reach. These localized points exist at the upstream ends of point bars where hydraulic controls in the form of crossover riffles raise the water surface sufficiently to promote flow onto the point bar surface. This point bar flow becomes progressively more channelized in chute channels and isolated from the main channel flow toward the downstream end of the bar.

Predictions of water surface elevations at 24-hour Camp were based on 15 cross sections (Figure 9). The highest observed discharge at this site for which calibration data were collected was 2,450 ft^3/s. The model was run to simulate river stage at discharge values of 3,000 ft^3/s, 3,500 ft^3/s, 4,000 ft^3/s, 4,500 ft^3/s, and 5,000 ft^3/s. Portions of five of the cross sections and one additional small cross section confined to the left bank area define the low-elevation chute 'C1' (Figure 9). In May 2001, throughflow was observed in chute C1 at a discharge of about 1,020 ft^3/s. Throughflow was also observed in the long sinuous secondary channel SC (Figure 9). The downstream end of the secondary channel was inundated, and flow just began to enter at its upstream end when discharge reached 2,850 ft^3/s. The left endpoints of cross sections 628 through 905 terminate on a high floodplain surface and their right endpoints terminate on the terrace level. The left endpoint of cross section 709 falls at the point on the floodplain where flow would enter the chute labeled C2 on the figure. Both ends of cross section 554 terminate on high floodplain. All of these termination points are well above the water surface level predicted by the highest modeled discharge of 5,000 ft^3/s, a 3.2-year event as calculated from the annual maximum daily mean discharges from 1943-2000. All floodplain surfaces are at least 0.2 m above the predicted water surfaces and all terraces are at least 0.35 m above the water surfaces at these cross sections. These results indicate that the discharge required to initiate flow onto the floodplain surface or into higher-elevation floodplain chutes is in excess of 5,000 ft^3/s. Field surveys of flood debris left by the 1999 peak flow of 7,010 ft^3/s suggest that stage at this discharge more closely corresponds with the elevation of the higher point bar ridges and floodplain surface. This is slightly larger than a 6-year flood.

Cross sections 470 and 419 traverse the head and tailwater areas, respectively, of a relatively large riffle located at a sharp bend in the low-flow channel. Modeled longitudinal water surface profiles through this reach suggest that this bend and riffle sequence function as a hydraulic control that causes an increase in the upstream water surface elevation at higher discharges (Figure 21). This backwater-induced elevation of the water surface allows flows to access the upstream margin of the point bar surface at lower discharges than would otherwise be required.

The point is illustrated by considering the sequence of cross sections from station 250 to station 470. At station 250, even a discharge of 5,000 ft^3/s fails to overtop the point bar crest (Figure 22A). Any water that may be present in the chutes shown to the right of the main channel would have had to enter the chute either from a point upstream or as backwater from the chute mouth. The far right chute shown in Figure 22A is labeled as C3 in Figure 9, while the smaller chute nearer the main channel is labeled C4. Figure 22B shows that the point bar crest at station 341 is also higher than any of the modeled water surface levels, indicating that no flow can enter the far right chute at this cross section either. We infer, however that flow would enter chute C4 at its mouth, which is located between cross sections 341 and 250, at discharges of 4,000 ft^3/s or more. The diagram of station 419 (Figure 22C) indicates that, although flow may begin entering the low-elevation chute immediately adjacent to the main channel at discharges as low as 3,500 ft^3/s, the bulk of the point bar surface is still above the 5,000 ft^3/s stage. The sharp bend in the low-flow channel between stations 419 and 470 causes these two cross sections to converge to the same right overbank area, so station 470 shares the same point bar profile shown for station 419. However, the 5,000 ft^3/s stage predicted for station 470 effectively inundates the majority of the point bar surface (Figure 22D). Flow into the far right depression shown by the station cross section begins at about 4,000 ft^3/s and is well-developed by the time discharge reaches 4,500 ft^3/s. This far right chute entrance is continuous with the far right chutes shown for the more downstream cross sections.

At this site on the Duchesne River, flow begins to overtop the upper end of the point bar at about 4,500 ft^3/s. This flood magnitude occurs with a recurrence interval of approximately 2.6 years. Flow into the main point bar chutes begins at discharges in the range of 4,000 ft^3/s. Flow into lower-elevation chutes and side channels begins over a wide range of discharges, ranging as low as 1,050 ft^3/s.

Above Pipeline

Model predictions at Above Pipeline suggest that floodplain surfaces and some portions of the high bars are rarely inundated. The point bar on river right at station 346 has a surface elevation similar to floodplain surfaces in the reach, while the point bar on river left at station 525 is somewhat lower. As a result, the upstream bar is inundated at lower discharges than the downstream bar. The higher bar and floodplain surfaces are deeply incised by well-developed chute channels that carry flow in most years.

The Above Pipeline model consists of 5 cross sections as shown on Figure 10. A single stage observation above base flow was obtained for the cross sections at the Above Pipeline site on May 23, 2001 when discharge was 770 ft^3/s. We evaluated the cross sections at stations 346 through 525 at this site. Stations 395 and 346 are located just upstream and downstream from a crossover riffle. A well-defined chute channel traversing the point bar on river right exits the main channel just downstream from station 395. Cross section 346 extends from a floodplain surface on river left, across the low-elevation chute labeled C2 on Figure 10, and terminates well up on the point bar surface on river right (Figure 23A). Model predictions suggest that both the bar crest and floodplain are perhaps 0.3 m higher than river stage at 5,000 ft^3/s at this cross section. The cross section at station 395 shows a similar situation (Figure 23B) in that river stage at the highest modeled discharge is well below the bar tops at the cross section. The heavy dashed line on the right bank of the cross section represents the projection of survey points defining the longitudinal profile of the point bar downstream from station 395 as shown by the dashed line between stations 395 and 346 on Figure 10. The mean water surface slope between stations 395 and 346 is also projected to the right of the cross section. These projections shows that substantial flow into the chute labeled C1 will occur long before river stage approaches the level of the point bar crest. Discharge into the chute probably begins when main channel discharge is in the range of 2,500 ft^3/s to 3,000 ft^3/s. The diagram of station 525 shows that the wide point bar on river left begins to flood when discharge is about 3,500 ft^3/s (Figure 23C). The bar, which begins in the diagram at a distance of 250 m from the left pin, is mostly inundated when discharge reaches 4,000 ft^3/s. However, as at the other sites, most of the floodplain remains above even the 5,000 ft^3/s stage. It is surmised that the numerous deep chute channels incised into that surface are accessed at their upstream ends by more frequently-occurring discharges.

Wissiup Return

Modeling predictions at the Wissiup Return site indicate that the tamarisk and cottonwood terrace are never inundated. High bar surfaces and the main chute channels are inundated by common floods, but the floodplain is rarely inundated.

Predictions at Wissiup Return are based on 10 cross sections shown on Figure 11. Portions of six of the cross sections define chute 'C1' (Figure 11). This chute was conveying a small unmeasured amount of flow when discharge in the main channel was 1,840 ft^3/s, which was the highest observed discharge at this site for which calibration data were collected. This model was run to simulate river stage at discharge values of 2,000 ft^3/s, 2,500 ft^3/s, 3,000 ft^3/s, 3,500 ft^3/s, 4,000 ft^3/s, 4,500 ft^3/s, and 5,000 ft^3/s. Representative cross sections from this reach showing selected water surface elevations are shown in Figure 24. The bar traversed by chute C1 is lowest and most rapidly flooded at its downstream end near stations 106 and 165 (Figure 24A and Figure 24B). The bar surface at both of these stations is entirely submerged at a discharge of 2,500 ft^3/s. At stations 193 and 249, located farther upstream, a greater discharge of 3,500 ft^3/s is required to substantially inundate the bar surface (Figure 24C and Figure 24D). The cross section at station 249 also shows the most downstream extent of the bar and floodplain surfaces on river left over which chute C2 flows. The return area of the chute is represented on the cross section by the small shelf located between 40 and 50 meters from the left pin. The elevation of this shelf suggests that backwater flooding into C2 may begin at between 3,500 ft^3/s to 4,000 ft^3/s.

The level of the modern floodplain at this cross section is represented by the higher flat surface located between 20 and 40 meters from the left pin. This surface is slightly higher than the highest modeled water surface level. A discharge of 4,000 ft^3/s is at the threshold of widespread flooding of the flat bar top on the right side of the cross section at station 353 (Figure 24E). The channel-like depression at the far right represents the entrance area to chute C1. This cross section spans the sill area leading into chute C1, so all flows shown can access this chute. The low area on the left portrays the downstream portion of chute C2. This chute is separated from the main flow by an elevated floodplain area and is therefore inaccessible to all modeled flows at this cross section. It is likely, however, that it would at least contain ponded backwater above main channel discharges of 3,500 ft^3/s. The stage and discharge at which flow would enter chute C2 at its upstream end is unknown.

These simulations show that much of the bar area traversed by chute C1 is flooded at discharges of 3,500 ft^3/s, and its inundation is essentially complete by the time discharge reaches 4,000 ft^3/s. These discharges correspond with floods equal to or slightly greater than the 2-year event. However, the floodplain surfaces in this reach remain above all modeled discharges. We suggest that the floodplain elevation in this reach of the river may be related to flood magnitudes closer to 7,000 ft^3/s, as was found in the 24-hour Camp reach. It is also clear from this analysis that only an extremely rare large flood would inundate the tamarisk and high terrace levels on which the cross section ends for this reach terminate. These surfaces are on the order of 2 m above the highest water surfaces modeled.

<u>Integration of Reach Results</u>

Hydraulic modeling at all three study reaches consistently indicate that floods ranging from the 2-year to the 2.6-year events are required to initiate significant flow onto high bar surfaces. The average discharge that inundates these surfaces is 4,000 ft^3/s. Somewhat smaller magnitude events with discharges of approximately 3,000 ft^3/s and a recurrence interval of 1.7 years are sufficient to produce flow into the main chute channels (Table 18). Flows capable of inundating the floodplains and higher bar ridges are almost certainly larger than the 3.2-year flood and probably have recurrences of about 6 years.

Discharges Necessary to Entrain Gravel

Results of our HEC-RAS modeling and gravel entrainment analysis indicate that reach-averaged shear stress approaches the threshold of entrainment over riffles and runs at discharges of approximately 4,000 ft^3/s (Table 19). The proportion of bed area where shear stress has reached the critical value increases from near zero to a large proportion with increasing discharge over a range of discharges from about 2,500 ft^3/s to more than 5,000 ft^3/s. These results are consistent with field observation of gravel entrainment during the spring 2001 peak, which briefly reached 2,900 ft^3/s. Little gravel entrainment occurred as a result of this discharge. Changes in channel morphology or disturbance of coarse deposits were also generally absent. However, some limited gravel mobilization did occur. A portion of a riffle crest at the take-out for Wissiup Ditch was painted prior to runoff. Most of the painted stones were still in place after the peak, indicating that the riffle surface layer had remained intact. A number of painted cobbles up to 85 mm in diameter had tumbled from the riffle crest a short distance down the riffle slipface. Most of the observed transported cobbles had moved less than about 3 m. The number and locations of painted particles that moved out of the immediate vicinity of the painted riffle crest are unknown. These observations show that 2,900 ft^3/s is sufficient to initiate limited particle movement in certain locations, but is insufficient to produce significant bed mobilization.

Green River Backwater Analysis

Overbank flooding in the backwater basins of the lower Duchesne River in Zone 1 of the study area is an important habitat for endangered fishes. Flooding in this area is controlled to a large degree by backwater effect associated with the stage of the Green River. Indeed, the tamarisk terrace of the Duchesne River discussed in this study may be more properly regarded as Green River floodplain in the extreme downstream reached of the study area. Use of the Duchesne River by razorback sucker is probably limited to this zone of backwater influence from the Green River.

We evaluated the spatial extent to which Green River backwater effects elevate stage in the Duchesne River over a range of discharges using a HEC-RAS model of the lower reaches of the Duchesne River. We specified water surface elevations representing various stages of the Green River at the model's downstream boundary. We then compared the resulting stage and mean velocity at upstream cross sections of the Duchesne River to their values calculated by specifying normal flow conditions at the downstream boundary. We defined the extent of backwater influence on the Duchesne River as the point in the Duchesne River where either stage is increased by at least 5 percent or mean flow velocity is reduced by more than 5 percent relative to their normal values. The upstream distance of Green River backwater effect for discharges on the Green River between 1,000 ft^3/s and 40,000 ft^3/s and discharges on the Duchesne River from 50 ft^3/s to 7,000 ft^3/s is shown in Figure 25. The approximate 2-year and 5-year flood discharges for the Duchesne (3,320 ft^3/s and 6,285 ft^3/s) and Green Rivers (22,960 ft^3/s and 29,385 ft^3/s, pre-Flaming Gorge Dam) are indicated by on the graph by bold vertical and horizontal lines. Over these ranges of discharges, Green River stage can be expected to elevate the water surface in the Duchesne River by at least 5 percent through a distance of between 5 and 8 km upstream from the Green River. These results show that Green River backwater effects historically extended upstream from the bend at Grey Bluff to near the downstream boundary of

Zone 2. However, the magnitude of the effect is small relative to the height of the river banks in the upstream part of the backwater zone. Valley bottom inundation due to Green River backwater effects is probably restricted to the reach downstream from the Oil Shack at present.

Historic Changes of the Duchesne River Channel and Alluvial Surfaces

The channel and the alluvial valley of the Duchesne River have changed greatly during the past 120 years. The topography and ecological functioning of today's channel and valley are the result of past channel migration, large-scale channel avulsions, periods of aggradation and incision, and changes in vegetation. Channel and valley evolution continues to the present day, although the rate of change has slowed in recent years (Table 20). This history can be condensed into a few periods of consistent trends and processes. These are 1) avulsions and filling of side channels before 1950, 2) two decades of channel widening throughout Zone 2 followed by nearly another two decades of bar and floodplain construction before an apparent stabilization of the area after 1987, 3) dynamic bend extension with frequent chute cutoffs throughout Zone 3 in all time periods with wet or moderate hydrology, and 4) relative stability in Zone 4 in all time periods.

<u>Qualitative Summary of Historic Channel Changes</u>

Our earliest detailed information concerning the state of the Duchesne River is from cadastral survey map from 1875 and 1882. These maps show that much of the main-stem Duchesne River channel between Myton and the Green River has narrowed by 6-20 m since 1875 (Brink and Schmidt 1996). The maps depict a sinuous multi-threaded channel system spreading across much of the alluvial valley in the upstream reaches (Zones 2, 3, and 4) of the study area. The remains of portions of this anastomosing system are now preserved as channel fills visible on aerial photography. For the most part, the filled channels were abandoned prior to the time of our earliest aerial photographs, based on comparing the cadastral surveys with the 1936 photos. There is one remaining active anabranch that is 10 km long depicted on the 1936 photos. This anabranch had been abandoned and filled by 1948. Although the precise cause of this period of channel narrowing and simplification is not yet known, its timing coincides with both a large reduction in annual stream flow in the Duchesne River and a large influx of fine sediment associated with gully incision in the Uinta Basin, as described above. In contrast with the upstream part of the study area, the 19th century maps show that the farthest downstream reach of the river in Zone 1a has remained virtually unchanged for the past 120 years.

The Duchesne River of 1936 was notably different from today's river. As previously mentioned, a 10-km-long active secondary channel meandered through the eastern half of the valley from just downstream from 24-hour Camp to downstream from the Pipeline. The Wissiup Ditch, a dredged irrigation canal, now occupies the swale left after this channel was abandoned. Multi-thread channel segments existed in several other locations in Zones 3 and 4 as well. For example, the long secondary channel loop labeled SC on the site map for 24-hour Camp (Figure 9) was the larger of a pair of anabranches that also included the current main channel.

A dominant trend in Zones 3 and 4 following 1936 was the continued loss of secondary channels. By 1948, the low-flow channel was reduced to a single thread nearly everywhere upstream from the Pipeline. In Zone 3, active river meandering occurred during this time period as well. Three chute cutoffs occurred in Zone 3. Two of them, at the location of the present Bowtie, were chutes in 1936 that captured all stream flow by 1948. The third, at the location of the present Above Pipeline site, had not yet begun to develop in 1936. Significant bank erosion occurred at several bends.

The most dramatic changes to occur between 1936 and 1948 occurred farther downstream. In 1936, Zone 1b contained a tortuously sinuous channel 9 km long. This channel was cut off by a large-scale avulsion and replaced by a post-avulsion channel only 3 km long in the late 1940s. The location of the former 1936 channel can be seen on the 1948 photography (Figure 6), which shows formation of the avulsion channel in progress. Two other long sections of channel, one located near Grey Bluff and the other at the current Wissiup Return site, were also lost to avulsion between 1936 and 1948. These avulsions may reflect channel response to a continued influx of excess fine-grained sediment. The avulsions occurred in the low-gradient reaches of the river where sand that would have passed through the steeper upper reaches would begin to aggrade and reduce channel capacity. Although the Wissiup Return area is upstream from the break in channel slope on the modern river, the high sinuosity of the pre-avulsion channel would have resulted in a gradient through the former bend half that of the modern channel. The effects of channel aggradation can also be seen in the reach between the old Wissiup bend and Grey Bluff. The 1936 channel had a moderately braided planform, with water spilling out into a long secondary channel on river left and a pair of shorter side channels near the valley margin on river right (Figure 26). In the 1948 photos, water can clearly be seen throughout the side channels on river right in spite of the fact that discharge at the time was no more than 460 ft^3/s. This same surface is now stranded 4 m above river level and vegetated with sagebrush.

The avulsions between 1936 and 1948 had different consequences in different reaches of the river. New straight channels bypassed abandoned sinuous channels in Zone 1b. The most notable difference in channel form was a change in sinuosity. Otherwise, the pre- and post-avulsion channels are both relatively narrow with simple shorelines. Since their formation, these channels have been extremely stable, showing almost no change for 50 years. The avulsion at Wissiup Return, however, triggered a dramatic river metamorphosis that stabilized only recently. The location of this avulsion is upstream from the point where valley gradient flattens, so the river nearly doubled its gradient through the reach when this avulsion occurred. In 1948, the avulsion was just beginning (Figure 27). It progressed rapidly, and photographs taken in 1955 show that the new channel was well established.

The channels abandoned during the avulsions described above were nearly filled by fine sediment by 1961. This is especially true for the channels in Zone 1. Today, it is difficult to find the filled abandoned channel fill near Grey Bluff, and the filled channel in Zone 1b is obliterated near the river. Portions of this fill near the channel margin retain some relief and are easier to locate in the field. The fill in the old channel at Wissiup Return remains about 1.5 m lower than the surrounding terrace. Other than filling of these abandoned channels, very little change occurred in Zone 1 after 1948. In Zone 2, the new channel near Wissiup Return began eroding into the surrounding terraces and widened rapidly. Widening occurred throughout Zone 2, through the entire length of the cutoff channel and extending about 400 m downstream to where a large braid bar complex we call lunch bar began to develop. Zone 3 was moderately active in the vicinity of the Bowtie during this period, and bank erosion in bends was widespread. Bend migration was less pronounced upstream in Zone 4.

Zone 4 became more active in the next time sequence, which spans 1961 to 1969. The bend immediately downstream from 24-hour Camp began to develop lobes, as did the farthest upstream bend in subreach 19. The channel near the former cableway for the gaging station near Randlett began to widen at this time as well. Widening at this site was described by Brink and Schmidt (1996). However, the pattern of changes near the cableway only occurred in the short reach near the gage and was not representative of changes in downstream reaches. In Zone 3, the area near the Bowtie continued to actively reconfigure itself and regular meander bends formed where a straight reach had been at the Above Pipeline site. The widening at Wissiup Return and lunch bar continued during this time interval.

Very little change occurred between 1969 and 1980, based on comparison of the aerial photographs. However, the period between 1980 and 1987 was a dynamic time in most areas. The area near the gaging station continued to widen, and it was during these years when the cableway was repeatedly washed out and finally abandoned. In Zone 4, the area of bare sand bars increased as vegetation was apparently scoured from bar surfaces or buried under new deposits. In Zone 3, the two bends upstream from the Bowtie extended laterally, and the Bowtie area itself was very active. A large nearly circular bend downstream from the Bowtie was cut off and the new channel eroded vigorously in the opposite lateral direction. Lateral erosion and widening at Zone 2 slowed.

Between 1987 and 1997, the pace of change slowed throughout the study area. In Zone 2, where channel widening had been consuming the terrace for 40 years, the channel finally stabilized. Riparian vegetation became established on a newly-developed floodplain surface set about 2 m below the elevation of the adjacent terraces. In Zone 3, which had been consistently active with high rates of chute cutoff and bend erosion, the eastern loop of the Bowtie was cut off, but little else happened. No obvious changes occurred in Zone 4.

<u>Changes in Channel Widths through Time</u>

Average widths of both the low-flow channel and the floodway channel were determined for each zone in the study area for the years 1936, 1948, 1961, 1969, 1980, 1987, and 1997 (Figure 28). The low-flow channel is defined as the water surface and lowest emergent bar surfaces seen on aerial photographs, while the floodway channel also includes adjacent high bar surfaces. This analysis indicates that significant narrowing occurred through much of the study area between 1936 and 1948. Widths have since remained relatively steady at the narrower post-1948 values in most areas, so that the present channel is narrower than the 1936 channel. Exceptions to this general narrowing trend include extreme widening that occurred in Zone 2 between 1948 and 1969. A significant increase in low-flow channel width is also found in Zone 3 during the period between 1961 and 1969, while Zone 4 shows an apparent increase in low-flow channel width during this period. In all cases, the increases are followed by a narrowing trend after 1969 that continued through at least 1997. Channel width in Zone 1a has not changed during the period of the study.

Channel width in Zone 1a did not change between 1936 and 1997 (Figure 28A), while the channel in Zone 1b is now significantly narrower than the 1936 or 1948 channel (Figure 28B). The average floodway channel width in Zone 1b decreased from about 58 m in 1936 to about 43 m in 1997. The apparent widening in 1948 reflects the fact that the 1948 photographs capture the development of a large channel avulsion in Zone 1b (Figure 6).

The width time series for Zone 2 shows the rapid increase in width resulting from the avulsion at Wissiup Return (Figure 28C). The new, steeper channel first appeared in 1948 and immediately began eroding laterally into the surrounding terraces. The large pre-avulsion channel was completely abandoned between 1948 and 1961, accounting for the decline in average low-flow channel width indicated for 1961. A wide braided zone with extensive mid-channel bars developed within the terrace banks by 1969, as indicated by the large widths of both the low-flow and floodway channels. The widening phase of this channel adjustment was largely complete by 1969, after which time the width of the floodway channel stabilized and the low-flow channel width declined as in-channel bars were converted to high bars. The large floods in 1983, 1984, and 1986 produced little additional terrace or bar erosion. Instead, these flood events may have served to deposit sediment on the bars, increasing their elevations and building the present floodplain. A formerly braided channel section was transformed into a well-defined channel flanked by a new floodplain surface 2 m lower than the pre-avulsion floodplain.

This stabilization is reflected by the continued decrease in the width of the low-flow channel in 1987, and by the decreased bar width that occurred as vegetation colonization converted bar to floodplain between 1987 and 1997. Only the far downstream portion of Zone 2 at lunch bar retains a large mid-channel bar area in a braided channel. Both the low-flow and floodway channels of Zone 3 (Figure 28D) narrowed by about 20 m between 1936 and 1948. The low-flow channel increased in average width by 14 m between 1961 and 1969, then slowly decreased back to the 1961 value. Floodway width has been steady since its initial decrease in 1948. Channel width in Zone 4 also decreased by about 20 m between 1936 and 1948 (Figure 28E). An increase in width of the floodway channel by about 10 m may have occurred between 1948 and 1969, although this amount of change is only marginally greater than the 10 percent margin of error assumed for these width measurements. This apparent widening was reversed after 1969, when the low-flow channel in Zone 4 returned to its 1948 width and the floodway channel reached even smaller dimensions than in 1948.

<u>Changes in channel sinuosity through time</u>

Channel sinuosity was calculated for each river zone from photographs taken in 1936, 1948, 1961, 1980, 1987, and 1997 (Figure 29). Zone 4 has consistently maintained sinuosity values ranging between 1.62 and 1.83. Sinuosity decreased slightly between 1936 and 1948, largely as a result of abandonment of the more sinuous channel of a two-channel stretch just upstream of 24-hour Camp (Figure 30). Sinuosity in Zone 4 has been steadily increasing since 1948 as meander bends slowly grow in amplitude. Sinuosity in Zone 3 was 1.57 in 1936 and grew steadily to 1.81 by 1980. A steady decline in Zone 3 sinuosity began after 1980, and by 1997 it was reduced to 1.68. Bends in Zone 3 grow more rapidly in amplitude than in Zone 4, but are frequently cut off by point bar chutes that periodically return the channel to a straighter course. The period between 1980 and 1987 was a particularly dynamic time in Zone 3, with widespread channel re-alignment (Figure 38). Bend extension and bend cutoff also occurred in Zone 3 in all other time periods except between 1969 and 1980 and between 1987 and 1993 (Table 20). No such cutoffs have occurred in Zone 4 during the study period. Gravel erosion and deposition data presented below suggests that significant bed aggradation may have triggered increased instability in Zone 3 after 1980.

Sinuosity in Zone 2 was about 1.86 prior to abandonment of a large bend near Wissiup Return. After the 1948 avulsion at this site the sinuous channel was replaced by a new channel with a lower sinuosity of about 1.2. This new channel assumed a wide braided form that persisted until about 1987. Sinuosity in this reach increased slightly after 1980 as the braid bars stabilized and the channel thalweg became more established as a permanent main channel. Sinuosity was extremely high in Zone 1b prior to the channel avulsions that occurred between 1936 and 1948. These avulsions replaced long stretches of tortuously sinuous channel with an essentially straight channel, causing sinuosity in Zone 1b to drop from 2.55 to about 1.25. The channel in this zone has remained essentially constant since the avulsions. Sinuosity in Zone 1a has not changed significantly through the study period, remaining relatively low at about 1.35.

Areas of Erosion and Deposition as an Index of River Activity

The areas of erosion and deposition within each river subreach within Zones 2-4 were calculated by overlay and analysis of GIS coverages from successive years. Zones 1a and 1b were excluded from this analysis because, other than the avulsions noted in Zone 1b between 1936 and 1948, this part of the river has been inactive. Earlier studies have used various measures of erosion and deposition during specific time periods as an index of channel activity (Ham and Church 2000). In this report, areas of erosion and deposition measured during specific periods are considered in quantifying channel activity for the time period.

In Zone 4, the period between 1936 and 1948 was the most active in terms of the total area of erosion and deposition, mainly reflecting the particularly large area of deposition that occurred during this time period. Deposition occurred over a large area between 1936 and 1948 in Zone 3 as well. However, the area of deposition was nearly as large in the period between 1980 and 1987, and the total erosion and deposition between 1980 and 1987 exceeded the total erosion and deposition during any other time period. Patterns of erosion and deposition in Zone 2 suggest that a geomorphic adjustment lasting at least until 1980 followed the 1948 channel avulsion in this reach. The periods from 1969 to 1980 and 1987 to 1997 were times of relative inactivity in all zones. Zone 4 has been relatively inactive compared with Zones 2 and 3.

Construction of high bar and floodplain surfaces and erosion of high bar, floodplain, and terrace surfaces were quantified for time intervals from 1936 to 1948, 1948 to 1961, 1961 to 1969, 1969 to 1980, 1980 to 1987, and 1987 to 1997. Temporal patterns of erosion and deposition are similar in Zones 3 and 4, although the magnitudes of erosion and deposition in Zone 3 are commonly 50 percent greater or more than in Zone 4 (Figure 31). Areas reported in Figure 31 are normalized by channel length through each zone and the number of years in each time interval. The normalized deposition rate of new bar and floodplain surfaces in Zones 3 and 4 was high between 1936 and 1948. It was during this time period when channel narrowing occurred along most of the main channel and a long secondary anabranch spanning from subreach 7 to subreach 16 on the east side of the valley bottom was completely abandoned and filled. Rates of high bar and floodplain construction then dropped to less than half of the pre-1948 values for the remainder of the study period in Zone 4. The rates of high bar and floodplain deposition in Zone 3 also dropped by more than half after 1948, reaching a clear historical low between 1969 and 1980 before rebounding to a rate nearly equal to the pre-1948 value. The interval between 1980 and 1987 showed the highest rate of all post-1948 periods in Zone 4 as well. Normalized areas of deposition in both Zones 3 and 4 dropped after 1987 to levels similar to those recorded between 1948 and 1980.

Patterns of high bar and floodplain erosion are also similar in Zones 3 and 4. Relatively low rates were recorded for the periods from 1936 to 1948, 1948 to 1961, and 1987 to 1997, while rates during the periods from 1961 to 1969 and 1980 to 1987 approximately doubled. High bar and floodplain erosion rates were lowest from 1969 to 1980 in both zones. Result of an analysis of terrace erosion in Zones 3 and 4 similarly yield maximum erosion rates for both zones during 1961 to 1969 and 1980 to 1987, and minimum rates during 1969 to 1980.

Total active area, defined as the normalized area of all erosion and deposition, shows prominent peaks for the years from 1936 to 1948 and 1980 to 1987 in both Zones 3 and 4. However, the peaks recorded for the interval between 1936 and 1948 are almost entirely due to high deposition rates. A third peak in total active area exists for the period from 1961 to 1969. The period from 1969 to 1980 is low in total active area, especially in Zone 3 where it is approximately half of the next-lowest value.

The temporal sequence of erosion and deposition in Zone 2 follows a very different pattern compared with the sequences described for Zones 3 and 4 (Figure 32). High bar and floodplain construction was relatively modest in Zone 2 between 1936 and 1948. Both high bar and floodplain deposition and terrace erosion increased between 1948 and 1961 as the new post-avulsion channel widened, consuming terrace and constructing new lower-elevation surfaces. The normalized areas of erosion increased by a factor of about 2.5 between 1961 and 1969, while the normalized area of deposition increased slightly. The total active area in Zone 2 for the period between 1961 and 1969 more than doubled over its value for the previous period, reaching the largest value recorded for any zone at any time. In the very next period, from 1969 to 1980, the area of terrace erosion dropped to near zero, bar deposition and erosion rates declined, and the total active area fell to a low level. Deposition rates in Zone 2 more than doubled during the years between 1980 and 1987 to reach a relatively high level, while erosion rates showed a moderate increase. The total normalized area of activity between 1980 and 1987 reached its second highest historical level. All area activity indices for Zone 2 dropped to low levels after 1987.

The somewhat erratic fluctuations in areas of erosion and deposition in Zone 2 can in part be explained in terms of the channel metamorphosis that occurred in the area in response to the 1948 avulsion near Wissiup Return. Prior to the avulsion, the channel in Zone 2 was a relatively inactive reach of high sinuosity and low gradient. Normalized areas of erosion increased immediately after the avulsion as the new channel began to widen, and the area of deposition increased as the old channel filled. These changes are reflected in the increasing values of normalized erosion and deposition recorded between 1948 and 1961. Activity rates continued to increase during the next period between 1961 and 1969, as terrace erosion accelerated and braid bars were constructed in the new wide channel. By 1969, the widening phase of the post-avulsion channel adjustment was largely complete, and a new braided channel had been established. Deposition continued as new floodplains were constructed in this wide braided reach. Most of this floodplain construction occurred between 1980 and 1987 when floods large enough to build higher-elevation floodplains occurred. By 1987, the braided reach had been transformed into a single meandering channel.

In spite of the differences in the history of erosion and deposition areas in Zone 2 from that of Zones 3 and 4, some common themes emerge in all three zones. All zones have significant peaks in erosion and total active area for the periods from 1961 to 1969 and 1980 to 1987. In addition, areas of erosion were exceptionally small in all three zones between 1969 and 1980.

Gravel Erosion and Deposition as an Index of River Activity

Areas of erosion from banks and redeposition as new bar or floodplain surfaces can be quantified from aerial photos. These changes in river channel planform combined with measurements of the thicknesses of gravel provide a means of calculating the volumes of gravel eroded and deposited during specific time periods (Ham and Church 2000). The magnitudes of gravel erosion, gravel deposition, and total gravel erosion plus deposition during a time period can all be considered in quantifying channel activity for the time period. These components of gravel redistribution are collectively referred to as gravel activity.

Gravel activity was calculated for Zones 2-4 for the time periods from 1936 to 1948, 1948 to 1961, 1961 to 1969, 1969 to 1980, 1980 to 1987, and 1987 to 1997. Peak activities rates in all river zones occurred between 1980 and 1987. The period between 1969 and 1980 was a time of reduced erosion activity in all river zones, and a time of low total activity in Zones 3 and 4.

Rates of all types of gravel activity were also low in all zones between 1987 and 1997. Total gravel activity was exceptionally low in Zone 2 between 1936 and 1948. The extremely low activity in Zone 2 during the first time interval can be attributed to the fact that prior to the avulsion and river metamorphosis near Wissiup Return, Zone 2 was a lower-gradient channel with very little gravel present. The pre-avulsion channel in Zone 2, shown on 1936 aerial photography, strongly resembles the downstream sand-bedded channel reaches in Zones 1a and 1b.

Temporal patterns of gravel activity in Zones 3 and 4 are to similar to those discussed for areas of erosion and deposition (Figure 33). Volumes of gravel erosion, deposition, and net activity shown in Figure 33 are normalized by channel length in each zone and the number of years in each time interval. The rate of gravel deposition between 1936 and 1948 is relatively high, and in Zone 4 is approximately 45 percent greater than the rate of gravel erosion during the same time period. This elevated rate of gravel deposition results in a rather high rate of total gravel activity, despite the fact that the gravel erosion rate during this period was low. The magnitude of active gravel transport in the study area between 1936 and 1948 was probably less than suggested by the morphological changes. Analysis of aerial photography indicates that this was a period of channel narrowing in the study area, while the reach-scale gravel budget for the period between 1936 to 1948 indicates that the quantity of gravel that went into storage in the study area was probably greater than the quantity transported into the study area during that time interval. Therefore, some portion of the apparent gravel deposition between 1936 and 1948 consisted of gravel deposited prior to 1936. The degree of channel activity during this time span, then, is probably more accurately represented by the rate of gravel erosion than by gravel deposition or total gravel activity.

Gravel erosion increased in both zones between 1948 and 1961, and the deposition rate decreased. Both erosion and deposition increased during the next time interval between 1961 and 1969, then declined markedly after 1969. Total gravel activity rates reached historical lows during the interval between 1969 and 1980. All measures of gravel activity increased to historical highs during the next time interval from 1980 to 1987, then declined to low values after 1987. The magnitudes of activity were consistently greater in Zone 3 than in Zone 4 throughout the study period, with the percent difference between the two zones increasing with increasing activity. Between 1980 and 1987 when total activity was at its peak, Zone 3 activity was 190 percent of Zone 4 activity. But between 1969 and 1980 when total activity was low in both zones, Zone 4 activity was roughly equal to Zone 3 activity.

All forms of gravel activity were low in Zone 2 between 1936 and 1948 (Figure 34). Gravel deposition activity then increased between 1948 and 1961 by a factor of 3.5 over its value in the first time interval, and by a factor of about 10 for the period from 1961 to 1969. These increases followed the 1948 avulsion at Wissiup Return, which steepened the channel gradient in Zone 2 and promoted gravel transport into the area. Between 1969 and 1980, the rate of gravel deposition in Zone 2 decreased to near its pre-1961 value. The gravel deposition rate in Zone 2 increased after 1980 to reach its greatest historical value during the interval between 1980 and 1987. Gravel erosion activity in Zone 2 was greatest between 1961 and 1969, but remained relatively low in all time intervals. The relative lack of gravel erosion in Zone 2 is a result of the fact that terraces in the area are composed primarily of fine-grained sediment, so little gravel is mobilized by terrace erosion.

When compiled over the full study area, the periods spanning 1969 to 1980 and 1987 to 1997 emerge as the least active in terms of gravel erosion, with the period from 1936 to 1948 coming in a close third (Figure 35). Erosion activity rates for these three periods are 0.7 m^3/m-yr, 1.0 m^3/m-yr, and 1.3 m^3/m-yr, respectively. When total gravel activity is considered, the periods from 1969 to 1980 and 1987 to 1997 again emerge as the times of lowest activity. Gravel erosion activity was greatest in the periods from 1961 to 1969 and 1980 to 1987. These two time periods also included the highest rates of total gravel activity.

Additional photographs taken in 1993 were examined for this study, but not formally analyzed. These photos show that little channel change occurred between 1987 and 1993, so that most of the activity measured between 1987 and 1997 occurred between 1994 and 1997. Although we did not measure actual activity rates within this shorter sub-period, we can produce rough estimates of activity for each sub-period by partitioning the measured activity rates for the full time periods between their shorter sub-periods. We conservatively assumed that 75 percent of the activity measured within the full time periods occurred during the sub-period for which supplementary aerial photographs showed significant channel change. We assigned the remaining 25 percent of the measured activity to the sub-period for which the supplementary photography showed little or no channel change. Erosion and total activity rates for sub-periods from 1987 to 1993 and 1993 to 1997 are indicated on Figure 35 by symbols placed at the sub-period mid-points and connected with dashed lines. The sub-period from 1987 to 1993 has the smallest activity rates of all time periods. This result is not sensitive to the precise proportion of full-period activity that is assigned to them.

Changes in Sediment Storage between Photo Intervals

Gravel mobilized during channel migration or other planform changes is transported downstream and redeposited. Such changes in the distribution of gravel stored in the system have the potential to alter channel form, local stage-discharge relationships, and local floodplain inundation frequencies. Changes in gravel storage were calculated for subreaches 5-19 as the difference between gravel erosion and deposition volumes within each subreach for each time periods between aerial photographs.

Reach-scale Changes in Gravel Storage

Reach-scale storage changes are the sum of the storage changes for all subreaches within a reach. Calculation of cumulative changes in gravel storage at the reach scale provides a means to estimate the rate of bedload transport into the study area and to evaluate the potential for bed aggradation. Results suggest that gravel was removed from the active channel during channel narrowing between 1936 and 1948. Transfer of gravel from the channel margins to the channel bed between 1948 and 1969 caused bed aggradation and temporary channel widening in Zone 3 and, to a lesser extent, in Zone 4. Gravel continued to accumulate at a much smaller rate in Zone 3 between 1969 and 1980, while Zone 4 became a net exporter of gravel. Zone 2 has been a persistent site of gravel accumulation through the study period.

Reach-scale storage changes represent net gravel accumulation or evacuation from a reach only if either an upstream or downstream boundary condition is known, that is, only if the gravel flux either leaving or entering the reach is known. For this analysis of Zones 2-4 of the lower Duchesne River, a downstream boundary condition of zero gravel transport is assumed. We propose that negligible gravel transport occurs past the downstream end of Zone 2 (river km 9.3) where the channel gradient abruptly flattens to less than one-sixth of its upstream value. The river downstream from this point assumes a narrow canal-like geometry with very little bar development and limited sites of gravel storage. The proportion of sand on the bed increases, and by river km 7 the channel is fully sand-bedded. Similar methods have been previously used to derive sediment budgets and estimates of bedload transport rates that compare favorably with estimates derived from intensive long-term monitoring (McLean et al. 1999; McLean and Church 1999).

A cumulative gravel budget for a reach with a downstream zero-transport boundary can either show no net change, implying that little or no gravel entered the reach from upstream, or a positive accumulation of gravel. Figure 36 (A-C) shows a schematic river reach with a zero-transport downstream boundary. The reach is composed of three subreaches. In Figure 36A, 1 unit of gravel passes from subreach 2 to subreach 1, and is stored in subreach 1. One unit of gravel is also stored in subreach 2, so 2 units of gravel had to pass from subreach 3 to subreach 2. The two units of total storage in subreaches 1 and 2 are supplied by 2 units of erosion in subreach 3, so no gravel is required from upstream to balance the reach gravel budget. In Figure 36B, all subreaches stored 1 unit of gravel, so that 3 units of gravel had to have entered the reach from upstream.

No gravel can leave a reach that has a gravel transport rate of zero at its downstream boundary. Therefore, a cumulative gravel budget of less than zero implies the physically impossible result that a negative quantity of gravel entered the reach from upstream (Figure 36C). Such a result can only occur if some portion of the gravel storage within the reach is overlooked. The -1 units of gravel shown entering the reach implies that at least 1 unit of undetected gravel storage has occurred somewhere in the reach. For our analysis, any additional undetected gravel storage may indicate storage within the channel in the form of vertical aggradation of the bed. Our method of quantifying gravel storage relies on planimetric data only. Vertical adjustments of the bed were not included in the analysis because of our inability to detect changes in bed elevation on historical aerial photography. In other words, gravel going into storage on the channel bed cannot be detected by our photogrammetric method.

Net changes in gravel storage for Zones 2-4 indicate that between zero and 18,100 m^3 of gravel passed the upstream boundary of the study area each year between 1936 and 1948 (Table 21). Some of the gravel stored in the study area between 1936 and 1948 may have already been present within the study area prior to 1936, and was subsequently removed from the active channel by channel narrowing and burial under finer-grained deposits. The much greater stream flows that characterized the hydrology of the early part of the 20th century may have deposited gravel over a wider channel area and at higher elevations than is possible under the later flow regime. As stream flow decreased by more than 50 percent after 1924 (Table 9), high elevation parts of what had been channel during the 1920s stabilized as high bars and floodplains. Meanwhile, the floodplain of the 1920s was transformed into a terrace. This type of response to reduced stream flow, in which channel narrowing occurs by the 'passive' abandonment of the higher gravel bars, occurred following dam construction on the Peace River (Church 1995). Therefore, the volume of gravel stored between 1936 and 1948 may not represent the volume of gravel that was transported into the reach during that time interval.

Little gravel entered the study area during the two time intervals spanning 1948 to 1969, when cumulative observed storage changes within the margins of uncertainty were negative or near zero (Table 21). The total annual gravel influx to the study area between 1969 and 1980 was less than 7,000 m^3. During the period from 1980 to 1987, between zero and 24,000 m^3 of gravel entered the study area. The total annual gravel influx to the study area between 1987 and 1997 was less than 6,900 m^3.

The cumulative storage changes through the study area for the periods from 1948 to 1961 and 1961 to 1969 are negative. Because no gravel can be removed from the study area due to the zero-transport condition at the downstream boundary, this suggests that significant quantities of gravel were stored on the stream bed during these intervals. Between 1948 and 1961, the gravel storage deficit for the study area, as calculated by summing all subreach changes in gravel storage, is estimated to be about 124,300 m^3. This quantity of sediment represents an average bed aggradation of about 11.7 cm through Zones 2-4. An even larger deficit is calculated to have accumulated through the study area between 1961 and 1969, and is greater in magnitude than the largest probable errors in estimating storage changes. Deficits during these time periods probably reflect the transfer of gravel from the river banks to the river bed by bank erosion, possibly with little downstream transport. Such a transfer would tend to increase bed elevation and might cause the channel to widen. Evidence that channel widening occurred in Zones 3 and 4 between 1948 and 1969 was presented in a previous section of this report.

Although the exact locations of bed storage cannot be determined from this analysis, it is probable that much of the bed aggradation occurred in Zone 3, particularly after 1961. Between 1948 and 1961 the gravel deficit accumulated at a similar rate through both Zones 3 and 4 (Table 21). Between 1961 and 1969, the deficit accumulation in Zone 3 more than doubled the deficit accumulation in Zone 4. During this time, widening of the low-flow channel in Zone 3 was particularly pronounced.

The gravel deficit in Zone 3 was much smaller during the period spanning 1969 to 1980. When the estimated error margins are taken into account, the net gravel deficit in the area may have been near zero. Minor gravel deficits also appeared in Zone 4 through the two time intervals spanning 1980 to 1997, and may represent gravel transfer to Zone 3. These deficits are small relative to the quantities of concurrent net deposition detected in Zone 3, suggesting that all gravel exported from Zone 4 could be easily absorbed as deposition in Zone 3. Local deficits during these last three time periods do not imply undetected deposition, as did the earlier deficits, because the sediment budgets for the full reach are positive. Net storage changes in Zone 2 have been positive throughout the study period, showing that Zone 2 has consistently been a site of gravel accumulation.

Longitudinal Changes in Subreach Gravel Storage

Longitudinal patterns of gravel storage changes for individual subreaches suggest that bed aggradation between 1948 and 1969 was greatest in Zone 3. Negative changes in gravel storage arise where the volume of local gravel erosion exceeds the volume of local gravel deposition. These areas generate a quantity of gravel that can be passed to downstream reaches. It is therefore reasonable to expect that the gravel supply and the potential for bed aggradation is greatest in subreaches some distance downstream from areas with large negative changes in storage.

The distance downstream that a given slug of gravel might travel from its source in a given time interval is not accurately known. However, an analysis of published data from many streams indicates that gravel typically travels less than one meander wavelength per year (Beechie 2001). The lengths of most channel subreaches defined for this study are one meander wavelength, while a few are two wavelengths long. Time intervals used in this study average about 10 years in duration, so the average channel residence time of gravel supplied to the channel during each period is about 5 years. Therefore, we assume that most gravel eroded from the stream banks in this study remains within four channel subreaches downstream from its source area during the time interval in which it was eroded.

Between 1948 and 1961, the negative changes in gravel storage were located in subreach 9 and in most subreaches upstream from subreach 13, indicating that these areas were net gravel source areas (Figure 37A). Most of the gravel eroded in subreaches 13 through 19 was deposited upstream from subreach 9, based on our assumed gravel transport distance of less than four channel subreaches. Gravel eroded in subreach 9 was deposited nearer its point of origin, moving only one or two subreaches downstream. The latter conclusion is based on the fact that relatively little gravel moved into the post-avulsion channel in Zone 2 (subreaches 5 through 7) during this period.

Every subreach from subreach 10 through 17 became a gravel source between 1961 and 1969 (Figure 37B). Thus, the potential for bed deposition was high everywhere in the study area except for subreaches 18 and 19. Maximum bed deposition might be expected just downstream from the largest source area for excess gravel centered at subreach 15. Combined with an assumed mean travel distance of 4 subreaches or less, the location of this source suggests that significant undetected bed deposition may have occurred in the vicinity of subreaches 11 through 14. Large volumes of sediment storage in Zone 2 during this period reflect the on-going channel adjustments following the 1948 avulsion, and suggest that much of the gravel mobilized in subreaches 10 and 11 was transported into Zone 2.

These data suggest that undetected bed aggradation during the two time intervals of negative gravel storage before 1969 was greatest in Zone 3, particularly in its upstream half (subreaches 12 through 15). Field observation of current channel morphology suggests that such aggradation probably occurred as large mid-channel bars or riffles that locally reduce channel capacity, rather than as a general increase in bed elevation. These local deposits might be expected to trigger channel instability in the form of bank erosion and the development of chute cutoffs at meander bends. Point bar cutoff chutes on the modern river almost invariably leave the main channel just upstream from mid-channel bars or islands. Comparison with historical channel adjustments in those subreaches indicates that this bed aggradation may be linked to subsequent channel instability.

The potential for instability in the upstream half of Zone 3 inferred from the gravel mass balance between 1948 and 1969 manifested itself as a wholesale reconfiguration of the area between 1980 and 1987, when some large floods occurred (Figure 38). A bend complex on river right near the downstream boundary of subreach 14 was cut off to form the western loop of the Bowtie. The next bend downstream on river left was deformed to produce the eastern loop of the Bowtie. The very next bend downstream extended to the west. Farther downstream in subreach 12, the nearly circular loop labeled Circle Bend was cut off, and the channel rapidly extended into a new bend to the west. The tight bend on the downstream limb of circle bend was washed out.

Dendrochronology and Stratigraphy

We investigated the history of floodplain aggradation at 24-hour Camp and at Wissiup Return by digging sediment pits, clearing cutbank exposures, and collecting living tamarisk trees for dendrochronologic analysis. Results show that about 8 cm of deposition has occurred on at one site on the tamarisk terrace near Wissiup Return since 1950, while about 19 cm of deposition has occurred along the margin of the partially-filled pre-avulsion channel at Wissiup Return since the late 1950s.

Three tamarisk trees were collected from near the Wissiup Return detailed study site in November of 2000. Two tamarisks, each about 12 cm in diameter, were taken from a section of cutbank on the downstream margin of the partially filled pre-avulsion channel on river left. This cutbank exposes a massive silty or clayey base more than a meter thick covered with three layers of sandier sediment totaling about 50 cm in thickness (Figure 39). An organic horizon exists at the top of the lowest sandy layer at a depth of about 37 cm. Examination of the tamarisks collected at this location revealed that the two trees probably germinated at a depth of about 19 cm, roughly the level of the top of the middle sandy layer labeled fms in Figure 39. The trees were about 41 years and 50 years in age at their germination points, placing the top of layer fms at the ground surface before 1960. Therefore, about 19 cm of deposition has occurred at this site since the 1960s. The third tamarisk was collected from a pit at the edge of the tamarisk terrace on river left about 100 m farther downstream. This tree was about 23 cm in diameter and 46 years of age. A germination point could not be determined for this tree because the interior of the stem was rotted through a length of about 0.5 meters near the ground surface. The sound wood below the rotted section was determined to be below the germination point. Stratigraphy preserved in this pit was similar to the stratigraphy observed in the upstream cutbank in consisting of a massive silty basal unit overlain by a sandier layer capped with an organic horizon (Figure 40). At this location, a single sand unit about 8 cm thick covers the organic horizon. We believe that the organic horizon at this pit correlates with the organic horizon at the upstream cutbank, which is dated at the upstream cutbank to be more than 40 years old. Therefore no more than 8 cm of deposition has occurred at this location since 1960. This timing suggests that the upper sand unit was deposited during the flood of 1965, which was the second largest event on record and peaked at 10,300 ft^3/s.

Two tamarisk samples were also collected from sediment pits in a floodplain surface at the margin of a partially-filled secondary channel at 24-hour Camp. However, the samples were collected from an insufficient depth to include their germination points. One sample extended to a depth of 20 cm and the other to a level of 40 cm. These trees were at least 18-20 years old, indicating that at least 40 cm of deposition has occurred at this location in the past two decades.

Synthesis of Historic Data

The 20th century geomorphic history of the lower Duchesne River includes complex adjustments to changes in both sediment supply and water discharge. The nature of the adjustments has varied both spatially and temporally over a period of at least 65 years, and continues to influence river morphology to the present day.

Channel Narrowing and Simplification in Zones 3 and 4

The response of the lower Duchesne River to declining discharge after the 1920s and the concurrent increase in fine sediment supply was spatially variable. In the upstream gravel-bed portion of the study area, most of the secondary channels that existed in the 19th- and early 20th century filled with sediment, and the average width of the main channel in the upstream half of the study area decreased by a third between 1936 and 1948. Similar reductions in stream flow and suspended sediment transport capacity caused significant channel narrowing and the loss of secondary channels elsewhere in the upper Colorado River basin (Van Steeter and Pitlick 1998; Allred and Schmidt 1999). Decreased discharge after the mid-1920s also resulted in a general decrease in water surface elevations, converting much of the pre-1920 floodplain into the terrace surface we refer to as the cottonwood terrace.

Channel widths reached minimum values in 1948 in Zones 3 and 4, and in 1961 in Zones 1a, 1b and 2. Widths then increased until 1969 in Zones 2 through 4, and until about 1980 in Zones 1a and 1b, but remained significantly less than their 1936 widths in Zones 1b and 4. The shift from channel shrinkage before 1948 to widening after 1948 in Zones 2 through 4 occurred without an obvious large change in hydrology, and may therefore reflect a decrease in the sediment supply.

Bed Aggradation and Avulsion in Zones 1b and 2

In much of the downstream sand-bed portion of the study area, the river responded to the decrease in stream flow and increased fine sediment supply with widespread avulsions. The avulsions downstream from river km 11 during the 1930s and 1940s were probably triggered by a decrease in channel capacity caused by channel aggradation. The direction of channel adjustment to changes in input variables is given by the qualitative model of Schumm (1969), which indicates that decreasing stream flow while increasing sediment supply should produce a decrease in channel depth, either an increase or decrease in channel width, and a decrease in channel sinuosity. Channel widening in Zones 1a and 1b in the 1940s and subsequent channel narrowing after 1948 may reflect the passage of a sediment wave associated with the 1930s gully incision in the adjacent uplands and a phase of channel re-incision after the avulsions.

Calculations using representative channel geometries and the Engelund-Hansen sediment transport function (Engelund and Hansen 1967) show that the fine-sediment transport capacity before 1925 was about four times greater than the transport capacity after 1943 throughout the lower Duchesne River. Calculations also indicate that the transport capacity through Zone 2 prior to the avulsion in 1948 was less than 60 percent of the capacity farther upstream, and transport capacity in Zone 1b was less than 6 percent of the transport capacity upstream from Zone 2. Thus, the reaches upstream can supply adequate fine sediment to induce deposition in the pre-avulsion channel in Zone 2 and the reaches downstream if the sediment supply to the upstream reaches is adequate. Because the change in flow regime after 1925 reduced the sediment transport capacity by 75 percent, a shift to net aggradation forcing channel adjustments in Zones 1a, 1b, and 2 is a likely outcome. The changes in channel pattern resulting from the avulsions resulted in increases in channel slope and increased sediment transport capacities though Zones 1b and 2. Because Zone 1a is subject to backwater effects from the Green River, hydraulic conditions and sediment transport in Zone 1a were less altered by the decrease in Duchesne River stream flow than were the zones further upstream. Consequently, no avulsions occurred in Zone 1a.

Channel Transformation in Zone 2

The long-term consequences of the four avulsions varied spatially. Since their formation, all three new channel reaches in Zone 1b have been stable, showing little change in planform or position in 40 years. The avulsion in Zone 2, however, triggered a channel transformation that took at least 30 years to complete and ultimately produced a channel that is very different from the pre-avulsion channel. The sequence of adjustments in Zone 2 were consistent with the model of channel response to perturbation proposed by Simon (1989), and included a period of incision and widening, followed by secondary aggradation, and eventual restabilization.

Prior to the 1948 avulsion in Zone 2, the channel was similar in general appearance to the reaches found downstream in Zones 1a and 1b. As shown on the 1936 photos, it consisted of a single large bend 2 km in length, with a relatively simple bank line. This pre-avulsion channel also lacked the wide point bar and riffle development found in the upstream reaches, and appears to have been predominantly sand-bedded. Sinuosity of the pre-avulsion channel in Zone 2 was 1.86, and the local channel gradient was about 0.0014.

The 1948 avulsion produced a new channel cutting 750 m in a straight line across the present high terrace surface (Figure 27). Local sinuosity fell to near 1.0, and the local channel slope nearly doubled, approaching the local valley slope of 0.0025. This channel straightening in Zone 2 was followed by a period of bank erosion and channel widening that took at least 20 years to complete. By 1961 the avulsion channel was well established and had widened significantly. At the same time, the pre-avulsion bend was abandoned and at least partially filled with sediment. Loss of the pre-avulsion channel produced a net decrease in the mean channel width measured in Zone 2 in 1961 compared with the 1948 width (Figure 28). However, extreme channel widening occurred between 1961 and 1969. The increase in local channel slope also enabled the transport of large quantities of gravel into Zone 2, which prior to the avulsion had contained no obvious large gravel deposits. As early as 1961, large mid-channel gravel bars were forming in the new channel, and by 1969 Zone 2 had assumed a wide, braided configuration. About 30,300 m^3 of gravel was transported into Zone 2 from upstream and deposited there between 1961 and 1969.

The area began to stabilize during a period of low stream flow after 1969. By 1980, vegetation had established on the channel bars, the channel had narrowed, and an incipient meandering planform had begun to establish. By 1987, most of the area had acquired its present configuration, with a moderately sinuous channel and an active floodplain set about 2 m below the elevation of the adjacent terraces. Much of modern surface composed of post-avulsion high bars and floodplain is well above the water levels attained by the 5-year flood, suggesting that these surfaces may have been built to their current elevations during the flood of 1983, an approximately 50-year event that peaked at 11,500 ft^3/s.

Changes in Channel Responses after 1969

Periods of relatively low stream flow, such those between 1936 and 1948 and between 1969 and 1980, have historically resulted in channel narrowing. Prior to 1969, large peak flow events and periods of high total runoff had the effect of restoring channel width after the drier periods. Since 1969, however, wet periods with large peak events have failed to reverse the narrowing that has occurred during the dry periods.

Unlike many other locations in the Colorado River basin, water development affecting the lower Duchesne River is not dominated by a single overwhelming event like the closure of Flaming Gorge Dam on the Green River or Glen Canyon Dam on the Colorado. Rather, withdrawals increased incrementally throughout the 20th-century, with the rate of increase accelerating after operation of the Bonneville Unit of the CUP began in 1971. The primary hydrologic effect of development has been to deplete stream flows during years with moderate runoff, thus increasing the frequency of years of low stream flow and the duration of droughts.

The period between 1948 and 1961 was a time of moderate hydrology and net bank erosion, as indicated by a negative change in detected gravel storage for the study area and increases in channel width in Zones 3 and 4. The interval between 1961 and 1969 was a period of relatively high total annual discharge, and included the second largest peak flow event on record (10,300 ft^3/s). It was also characterized by bank erosion and channel widening. Total erosion of gravel deposits in the study area exceeded detected gravel deposition by approximately 167,000 m^3. In contrast, the very wet years between 1980 and 1987, which included the flood of record, produced almost no widening of the low-flow channel. Channel widths continued to decrease in most areas between 1987 and 1997, even though that period included the floods of 1995 and 1997, with recurrence intervals of approximately 5.9 years and 3.2 years, respectively.

The failure of floods to widen the channel after 1969 could be due to changes in hydrology that allowed riparian vegetation to establish in the active channel. Tamarisk colonized the riparian areas along the Green River and its tributaries sometime between 1935 and 1955 (Graf 1978). Our studies of tamarisk dendrochronology described above confirm that at least some individuals had germinated along the lower Duchesne River by the early 1950s. Numerous clumps of dense tamarisk are visible on air photos of the study area taken in 1961. Although tamarisks were certainly present in the study area by the 1960s, they had little apparent effect on the channel response to the high stream flows between 1961 and 1969.

Tamarisks are considered to be an opportunistic species capable of quickly colonizing exposed channel bars during periods of low flow (Everitt 1998; Allred and Schmidt 1999). After 1971, water withdrawals decreased the frequency of moderate peak events capable of removing tamarisk seedlings. The increase in the durations of drought periods since 1971 has increased the opportunity for tamarisk to become established on bars and channel margins before an effective flow occurs. Once established, tamarisks have the potential to stabilize sediment and decrease erosion rates (Graf 1978). Subsequent floods are therefore less effective, and relatively rare events may be required to remove the vegetation. Similar encroachments of riparian vegetation have been shown to follow reductions in the frequency of disturbance by floods elsewhere (Erskine et al. 1999). The extent to which this increase in vegetation in and along the channel margin reduces the geomorphic effectiveness of later floods has not been determined.

The Question of Equilibrium

Among the issues we have been asked to consider for this project is whether the lower Duchesne River can be considered to be in equilibrium with current flow conditions. As the historical analysis presented in this report indicates, the validity of assigning a single state of equilibrium to this study area is questionable. In addition, a time domain must be specified. Over what time interval do we wish to consider the state of the system as being potentially in equilibrium? We have shown that the changes and adjustment that have occurred on the lower Duchesne River in the past century have been both large and variable in both space and time. Some changes can be regarded as temporary perturbations from which the system eventually returns to its former state, while others have irreversibly altered the form of the river and its future course of evolution.

The Duchesne River has been in a state of partial dis-equilibrium throughout the time period covered by this analysis, and remains so today. Adjustments initiated by a change in flow regime and increases in sediment loads in the first third of the 20th century appear to have stabilized. In spite of the known increases in flow diversions, stream flow in the study area has been relatively stable over the period of record. Mean annual flow for water years 1972 through 2000 is about 90 percent of mean annual flow for water years 1943 through 1970. More importantly, large magnitude flows capable of doing significant geomorphic work continue to occur. However, channel width has decreased throughout the study area since about 1970 when the Bonneville Unit of the CUP began operation. This narrowing is likely related to the decreased frequency of effective flood events, which has provided a greater opportunity for vegetation to establish within the active channel, in addition to an estimated 25 percent decrease in fine sediment transport. Declining flows since about 1970 has also resulted in a decrease in the rate of fluvial processes in the upstream part of the river in Zone 4.

Other portions of the study area have nonetheless continued to be active. A large quantity of gravel is presently working its way through Zone 3, and can be expected to trigger frequent cutoffs and erratic bank erosion in the area for some time to come. These types of local instabilities impact native fish in at least two important ways. Channel reconfigurations create new and diverse habitats, and the local widening that accompanies these events may constitute barriers for fish passage at low flows until subsequent high-flow events can re-establish a defined channel. Maintenance of adequate depth for fish passage during periods of low flow is among the criteria for defining base flow requirements for the lower Duchesne River (Haines and Modde 2003).

Relationship between Duchesne River Changes and the Hydrologic Record

Frequent bed mobilization and active channel migration are among the critical attributes necessary for the geomorphic and ecological integrity of alluvial streams (Trush et al. 2000). In gravel-bedded reaches like Zones 2-4 of the Duchesne River, these processes require discharges sufficient to mobilized and re-distribute significant quantities of gravel. Rates of gravel activity on the lower Duchesne River have been high during some time periods evaluated in this study and very low during other times periods. We have specified threshold discharge magnitudes for gravel mobilization in previous sections of this report. The durations of flow exceeding these threshold discharges necessary for maintaining channel integrity can be determined by comparing the hydrologic record with the rates of channel activity over specified time intervals (Table 22).

Bed mobilization has been shown to occur over significant portions of the channel bed at discharges near 4,000 ft^3/s in the study area. A similar discharge magnitude has been shown to be necessary to initiate flow over high bar surfaces. We therefore estimate the threshold discharge for channel maintenance in the study area as approximately 4,000 ft^3/s. The total volume of flows exceeding this threshold magnitude for a given year (i) can be expressed by the variable T_i in units of ft^3/s-days. This quantity represents the sum of the differences between mean daily discharge and the specified channel maintenance threshold for all flows greater than the threshold. T_i is calculated as:

$$T_i = \sum_{j=1}^{n} \left[\overline{Q}_{Tij} - 4000 \right]^b \qquad (10)$$

where $\overline{Q}_{Tij}$ represents the daily mean discharges in ft^3/s on the n days during year i on which the daily mean was greater than 4,000 ft^3/s, and b is assumed equal to 1. All days during the year on which the daily mean discharge was less than 4,000 ft^3/s are discarded. For example, if a daily mean discharge of 5,300 ft^3/s occurred on two separate days (n = 2) during year i, and the daily mean discharge was less than 4,000 ft^3/s on all other days of the year, the total flow volume in excess of the 4,000 ft^3/s channel-forming threshold for the year (T_i) would be 2,600 ft^3/s-days.

The value of the exponent b in equation 10 represents the possibility that geomorphic effectiveness is a non-linear function of discharge above the channel-forming threshold. Strong non-linearity in the relationship between unit discharge and unit bedload transport can arise when flow conditions are near the threshold for particle entrainment, particularly in the presence of surface armor or pavement. Surface coarsening of the bed can interferes entrainment processes, and alters the relationship between excess shear stress at the bed (τ_0 - τ_c) and the bedload transport rate (Dietrich 1989; Parker and Klingeman1982). At higher discharges when the bed is fully mobilized, this effect is reduced such that the increase in sediment transport rate is proportional to excess shear stress raised to the three-halves power (Martin and Church 2000). The manner in which the bedload transport rate changes with increasing unit discharge (q) above the discharge threshold for entrainment (q_c) is therefore determined by the manner in which shear stress changes with discharge. Assuming a wide rectangular channel, manipulation of the Manning equation shows that shear stress is proportional to discharge raised to the three-fifths power. Multiplying the exponent relating excess shear stress at the bed to the bedload transport rate (three-halves) by the exponent relating discharge to shear stress at the bed (three-fifths) yields an exponent relating excess discharge (q - q_c) to bedload transport rate of nine-tenths, or approximately 1. These results are incorporated in many sediment transport functions that express bedload transport rate in terms of excess discharge, such as the 1934 Schoklitsch equation as given by Raudkivi (1998):

$$q_B = \frac{7000}{\sqrt{D}} S^{3/2} (q - q_c) \qquad (11)$$

in which q is unit discharge, q_c is the critical unit discharge, S is slope, and D pertains to particles size. Note that there is no exponent on the quantity in parentheses.

The threshold we have employed for the Duchesne River of 4,000 ft^3/s is approximately the bankfull discharge, and corresponds to widespread gravel entrainment rather than incipient motion. Our gravel entrainment studies reported above indicated that limited gravel movement occurs at discharges equivalent to about 60 percent of bankfull in the lower Duchesne River, whereas Pitlick and Van Steeter (1998) found that incipient motion in the upper Colorado River occurs when discharge is approximately 50 percent of bankfull. Therefore it is reasonable in this case to assume that transport rate increases as a nearly linear function of excess discharge.

Channel changes and erosion activity rates were quantified for the time intervals from 1936 to 1948, 1948 to 1961, 1961 to 1969, 1969 to 1980, 1980 to 1987, and 1987 to 1997. Erosion activity rates were compared with the volumes of channel-forming flows for each time period to evaluate the volumes of flow necessary to promote dynamic channel processes. Metrics of erosion activity were selected for this comparison because erosion is associated only with sediment mobilization, whereas deposition can also be a passive process associated with periods of channel shrinkage (Church 1995). A pattern of wetter and drier cycles appeared in the stream flow record after 1950, such that substantial channel-forming flows occurred in some periods and little occurred in others.

The interval between 1987 to 1997 for which channel changes are well-quantified by GIS analyses spans both the driest five years on record from water year 1988 through 1993 and the years from 1994 to 1997, which included some large peak events. Changes seen in the interval from 1987 to 1997 then, integrate the effects of both a dry and wetter flow regime. Supplementary photographs taken in 1993 show that little channel change occurred between 1987 and 1993, so that most of the activity measured between 1987 and 1997 occurred after 1993. Activity rates for these sub-periods were estimated by assigning 75 percent of the activity measured through the full photo-analysis period to the wetter sub-period, and assigning 25 percent of the measured activity to the drier period. The precise ratio used for distributing the measured activity between these two sub-periods has little impact on the qualitative characteristics of the subperiods. The interval from 1987 through 1993 emerges as an extremely inactive time regardless of the ratio assumed, whereas the interval after 1993 remains a time of moderate activity.

The interval spanning water years 1937 through 1948 was characterized by deposition and channel narrowing. An inability of the river to adequately transport its sediment load resulted in several channel avulsions. Rates of erosion area and gravel erosion were low (Table 22). As has been pointed out in previous sections, significant channel narrowing occurred during that time period. The average annual volume of channel-forming stream flows of 3,400 ft^3/s-days per year for water years 1937 through 1948 was thus insufficient to maintain channel morphology. Channel activity measures increased by approximately 40 percent between 1948 and 1961, and some bend migration occurred. The average annual volume of channel-forming stream flows in water years 1949 through 1961 was 9,800 ft^3/s-days. During water years 1962 through 1969, both activity measures were high, reaching their highest values calculated for any time period (Table 22). The average annual volume of channel-forming stream flow during this time was 13,475 ft^3/s-days per year. Annual discharge volume in excess of the channel-forming threshold averaged only 1,630 ft^3/s-days per year in water years 1970 through 1980 when very little channel change occurred and both forms of activity were very low. A trend toward narrowing of the low-flow channel began during this time period.

The interval spanning water years 1981 through 1987 was the most active period studied in terms of channel reconfiguration, particularly in Zone 3. Erosion activity rates calculated for this dynamic time period, when the annual volume of channel-forming stream flows averaged 27,200 ft^3/s-days per year, were high (Table 22). In spite of the high rates of erosion activity measured for this period, low-flow channel widths continued to decline. The failure of the record floods of the period to restore channel dimensions is likely related to the encroachment of riparian vegetation, which had become established along the channel margin as a result of less frequent effective flood after 1972. Between 1987 and 1997, erosion activity was low and processes of bend migration and cutoff were limited to the area near the Bowtie. Channel-forming flow volumes averaged 2,810 ft^3/s-days per year for the full time period. Channel narrowing continued during this period as well.

Most of the channel change observed between 1987 and 1997 occurred after 1993, as did all occurrences of channel-forming discharge. When 75 percent of the activity measured between 1987 and 1997 is averaged over the 4 years after 1993 rather than the full 10-year time interval, the resulting annual erosion activity rates are moderate (Table 22). The average annual channel-forming stream flow volume in the wetter sub-period spanning water years 1994 through 1997 was 7,000 ft^3/s-days per year. The activity rates estimated by averaging the remaining 25 percent of the activity measured between 1987 and 1997 over the remaining six dry years spanning water years 1988 through 1983, when no channel-forming flows occurred, are exceptionally low.

In summary, average annual channel-forming stream flow volumes of 3,400 ft^3/s-days per year in water years 1937 through 1948, 1,630 ft^3/s-days per year in water years 1970 through 1980, and 2,810 ft^3/s-days per year in water years 1988 through 1997 produced little channel activity and little channel change (Table 22). Thus, an average annual channel-forming flow volume of less than 3,400 ft^3/s-days per year is insufficient to maintain channel dynamics or channel dimensions. Average annual channel-forming flow volumes of 27,200 ft^3/s-days per year, as measured over during water years 1981 through 1987, and 13,475 ft^3/s-days per year, as recorded over water years 1962 through 1969, resulted in high rates of channel activity. Average annual channel-forming flow volumes of greater than 7,000 ft^3/s-days per year, as estimated for water years 1994 through 1997, and 9,800 ft^3/s-days per year, as estimated for water years 1948 though 1961, resulted in moderate rates of gravel activity and channel change.

These data suggest that an average annual channel-forming stream flow volume of at least 7,000 ft^3/s-days per year is sufficient to promote channel migration and maintain channel integrity. Additional factors also influence the effectiveness of a given discharge magnitude and duration in re-distributing gravel and altering channel morphology. Antecedent channel conditions, such as the spatial distribution of active gravel accumulations and channel geometry, exert important controls on channel responses to particular flow events. In particular, substantially larger flows may be required to restore channel dimension after vegetation has become established along the channel than would have been required to prevent the initial colonization.

Table 9: Changes in flow regime after 1925.

	1912-1924	1925-1942	1943-2000	Percent Decrease 1912-1924 to 1943-2000
Mean Annual Runoff (acre-feet)	877,060	465,500	407,150	53.5
1.5-yr Flood	6,450 ft^3/s	3,750 ft^3/s	2,450 ft^3/s	62.3
2-yr Flood	7,630 ft^3/s	4,610 ft^3/s	3,320 ft^3/s	56.5
5-yr Flood	10,600 ft^3/s	6,520 ft^3/s	6,270 ft^3/s	40.8
90-percent Exceedence	390 ft^3/s	70 ft^3/s	60 ft^3/s	84.5
50-percent Exceedence	600 ft^3/s	413 ft^3/s	340 ft^3/s	43.2
10-percent Exceedence	2,826 ft^3/s	1,410 ft^3/s	1,120 ft^3/s	60.3

Table 10: Variability between upper-quartile, lower-quartile, and middle-quartile years from 1943 to 2000.

	Upper-Q Years	Middle-Q Years	Lower-Q Years
Mean Annual Runoff (acre-feet)	752,230	374,760	130,110
1.5-yr Flood	5,860 ft^3/s	2,950 ft^3/s	685 ft^3/s
2-yr Flood	6,535 ft^3/s	3,500 ft^3/s	865 ft^3/s
5-yr Flood	8,550 ft^3/s	5,085 ft^3/s	1,730 ft^3/s
90-percent Exceedence	200 ft^3/s	75 ft^3/s	32 ft^3/s
50-percent Exceedence	600 ft^3/s	350 ft^3/s	113 ft^3/s
10-percent Exceedence	2,370 ft^3/s	1,010 ft^3/s	400 ft^3/s

Upper-Q years: upper quartile in mean annual runoff 1943-2000
Lower-Q years : lower quartile in mean annual runoff 1943-2000

Table 11: Flow variability between wet and dry cycles since 1950.

	Dry Cycles	Wet Cycles
Mean Annual Runoff (acre-feet)	263,107	697,194
Percent of Mean Annual Runoff*	64.6	171.2
1.5-yr Flood	1,600 ft^3/s	5,200 ft^3/s
Percent of 1.5-yr Flood*	65.3	212.1
2-yr Flood	2,300 ft^3/s	6,275 ft^3/s
Percent of 2-yr Flood*	69.2	189.2
5-yr Flood	4,470 ft^3/s	9,265 ft^3/s
Percent of 5-yr Flood*	71.3	147.7

*Full period of actual record at gage nr Randlett (1943-2000)
Dry Cycles: 1954-64, 1970-82, 1988-1996; Wet Cycles: 1965-69, 1983-87.

Table 12: Changes in discharge since construction of the Bonneville Unit of the Central Utah Project.

	Pre-project 1943-1971	Post-project 1972-2000	Percent Change
Mean Annual Runoff (acre-feet)	429,498	384,821	-10.4
Mean Annual Runoff in Upper-Quartile Years (acre-feet)	692,450	835,562	+20.7
Mean Annual Runoff in Middle-Quartile Years (acre-feet)	380,442	368,118	-3.2
Mean Annual Runoff in Middle-Quartile Years (acre-feet)	165,653	110,354	-33.3
1.5-yr Flood	2,540 ft^3/s	1,840 ft^3/s	-27.5
2-yr Flood	3,700 ft^3/s	2,710 ft^3/s	-26.8
5-yr Flood	6,280 ft^3/s	5,959 ft^3/s	-5.2
90-percent Exceedence	57 ft^3/s	60 ft^3/s	+5.3
50-percent Exceedence	380 ft^3/s	280 ft^3/s	-26.5
10-percent Exceedence	1,180 ft^3/s	1,100 ft^3/s	-6.9

Table 13: Flow characteristics by photograph interval.

Photo Interval (Water Yrs)	37-48	49-61	62-69	70-80	81-87	88-93	94-97
Mean Annual Runoff (acre-feet)	475,310	390,660	491,932	353,685	665,693	106,381	327,981
Percent of Long-term Mean Annual Runoff	117	96	121	87	164	26	82
Max. Daily Mean (ft^3/s)	7,000	8,400	9,460	6,550	11,500	2,910	7,000
Average Max. Daily Mean (ft^3/s)	3,875	4,020	5,065	3,805	5,750	1,300	3,280
90-percent Exceedence (ft^3/s)	140	28	78	64	148	42	57
50-percent Exceedence (ft^3/s)	445	350	400	280	605	102	223
10-percent Exceedence (ft^3/s)	1,400	890	1,390	920	1,560	310	703
Description	wet	moderate	wet	moderate to dry	wet	very dry	dry with peak events

Table 14: Suspended sediment loads in wet, dry, and normal years.

	Qs (T/year)	Percent of Wet Year	Percent of Average
Wet Year	1,023,400	--	252
Normal Year	288,100	28	71
Dry Year	38,800	4	10
Average of all Years	405,350	40	--

Table 15: Longitudinal zones of the lower Duchesne River.

ZONE	Subreaches	Morphology				Change History
		Bed	Slope	Sinuosity	Description	
1a	1	Sand	.00014	1.36	Canal-like geometry. Straight sections and tight bends	No change.
1b	2-4	Sand	.00034	1.23		Large-scale avulsions.
2	5-7	Grav.	.00268	1.24	Semi-braided. Very high terrace banks.	Large-scale avulsion. Channel metamorphosis.
3	9-15	Grav.	.00231	1.68	Meandering.	High rates of bank erosion and frequent cutoffs.
4	16-19	Grav.	.00134	1.84	Meandering.	Relatively stable. Side channel loss.
Subreach 8 is a transition area between Zones 2 and 3. The channel has been modified and rip-rapped in association with construction of a petroleum pipeline that cross beneath the river bed at this location.						

Table 16: Surface layer particle sizes.

Site Name	River Km	D_{50}	D_{65}	D_{90}
24-hour Camp	22.0	55	69	110
Km 20.5	20.5	48	56	87
at Bowtie	17.9	30	37	59
Below Bowtie	16.3	44	53	93
Above Pipeline	13.2	75	90	127
Wissiup Return	10.1	52	66	93

Table 17: Subsurface particle sizes.

Site Name	River Km	Percent Sand	D_{50}*	D_{65}*	D_{90}*
24-hour Camp	22.0	13	52	70	152
Below Bowtie	16.3	14.5	30	42	71
Above Pipeline	13.2	20.5	45	64	105
Wissiup Return	10.1	18	35	43	64

*Calculated from gravel fraction only.

Table 18: Discharges to access topographic features of point bars in m^3/s.

Reach	Local Bar Inundation		Flow into Main Chutes	
	(ft^3/s)	R.I.	(ft^3/s)	R.I.
24-hour Camp	4,000-4,500	2.4-2.6	3,000-4,000	1.7-2.4
Wissiup Return	3,500-4,000	2.0-2.4	1,840	1.3
Above Pipeline	3,500-4,000	2.0-2.4	2,500-3,000	1.5-1.7
Study Area	4,000 ± 500	2.4	3,000 ± 1,000	1.7

Table 19: Critical shear stress for gravel entrainment.

Site	Average Bounday Shear		Estimated Q_{cr} (ft^3/s)
	τ*	Q (ft^3/s)	
24-hr Camp			
Station 554	0.030	4,000	--
Station 470	X	> 5,000	--
Station 419	0.034	2,500	--
Station 36	0.032	3,500	--
Station 0	0.031	4,000	--
Reach Mean	0.031	4,000	4,000 ± 1,400
Above Pipeline			
Station 395	0.033	2,500	--
Station 346	0.023	5,000	--
Reach Mean	0.031	4,000	4,000 ± 1,400
Wissiup Return			
Station 353	0.030	2,500	--
Station 291	X	> 5,000	--
Station 249	X	> 5,000	--
Station 193	0.030	3,500	--
Station 165	0.030	5,000	--
Station 138	0.030	2,500	--
Station 106	0.030	2,500	--
Station 36	0.033	4,000	--
Reach Mean	0.031	4,000	4,000 ± 1,400
Study Area			4,000 ± 1,400
X = τ* does not approach threshold at any modeled Q			

Table 20: Channel changes by photo period.

	1936-48	1948-61	1961-69	1969-80	1980-87	1987-93	1993-97
Hydrology	Wet.	Moderate.	Wet.	Moderate to dry.	Wet.	Very dry.	Dry with one peak event.
Zone 4	Secondary channels fill. Channel narrowing.	Minor bend erosion.	Lobing below 24-hr camp and below gage. Gage area begins to widen.	Little change.	Gage area widens. Vegetation scoured on bars.	No change.	Little change.
Zone 3	One cutoff, two more chutes complete their cutoffs. Bend extension.	Chute cutoff, and local widening near Bowtie area. Bend extension throughout.	Above Pipeline Site develops 2 meanders via 40-70 m of extension per bend. Bowtie area active.	Little change.	Extension above Bowtie. Bowtie area very active. Circle bend cut off and extended in opposite direction. Next downstream bend is washed out.	No change.	Eastern half of Bowtie cut off.
Zone 2	Large avulsion at Wissiup Return.	Widening at Wissiup cutoff and Lunch bar.	Continued widening at Wissiup cutoff and Lunch bar.	Growth of mid-channel bars.	Wissiup Return area begins to stabilize.	Little change.	Wissiup Return area continues to stabilize.
Zone 1b	Two large avulsions, reduced braiding near Grey Bluff, channel narrowing.	Abandoned channels fill.	Bend translation near oil shack.	Little change.	Bend translation.	No change.	Little change.
Zone 1a	No change.	No change.	Little change.	No change.	Slight increase in sinuousity.	No change.	No change.

Table 21: Cumulative changes in gravel storage in Zones 2-4 for five time intervals.

	1936-1948	1948-1961	1961-1969	1969-1980	1980-1987	1987-1997
Zone 2 Storage Change (m^3 x 1000)	5.8 ± 4.1	12.7 ± 12.9	30.3 ± 26.3.0	7.7 ± 9.5	25.2 ± 20.7	0.5 ± 10.2
Zone 2 Influx/yr (m^3 x 1000)	0.5 ± 0.3	1.0 ± 1.0	3.8 ± 3.3	0.7 ± 0.9	3.6 ± 3.0	0.1 ± 1.0
Subreach 8 Storage Change	11.7 ± 5.4	-4.2 ± 5.0	0.9 ± 3.4	1.6 ± 1.7	1.4 ± 3.5	-0.5 ± 1.1
Zone 3 Storage Change (m^3 x 1000)	26.3 ± 71.0	-79.0 ± 83.3	-134.2 ± 74.1	-7.3 ± 30.4	15.8 ± 85.0	8.8 ± 40.5.1
Zone 3 Cumulative Influx/yr (m^3 x 1000)	3.7 ± 6.7	-5.4 ± 7.8	-12.9 ± 13.0	0.2 ± 3.7	6.1 ± 15.4	0.9 ± 4.9
Zone 4 Storage Change (m^3 x 1000)	43.4 ± 49.3	-53.8 ± 34.7	-63.6 ± 33.8	7.3 ± 26.7	-18.8 ± 36.2	-13.5 ± 24.6
Zone 4 Cumulative Influx/yr (m^3 x 1000)	7.3 ± 10.8	-9.6 ± 10.5	-20.8 ± 17.2	0.8 ± 6.2	3.4 ± 20.6	-0.5 ± 7.4
Total Study Area Storage Change (m^3 x 1000)	87.2 ± 129.8	-124.3 ± 135.9	-166.6 ± 137.6	9.3 ± 68.3	23.5 ± 143.9	-4.6 ± 76.7
Bed Storage to Balance (cm)	--	11.7 ± 12.8	14.9 ± 12.3	--	--	--

Bed storage to balance = storage change/total area occupied by channel during time interval in reach.

Table 22: Magnitude-duration of channel-forming flows and erosion activity rates through time.

Water Years	Area Erosion Activity (m^2/m/yr)	Gravel Erosion Activity (m^3/m/yr)	Activity Level	Mean Annual ft^3/s-days
1937-48	1.15	1.18	Low	3,400
1949-61	1.71	1.62	Moderate	9,800
1962-69	3.44	2.70	High	13,475
1970-80	0.69	0.67	Low	1,630
1981-87	2.97	2.43	High	27,200
1988-97	0.92	0.86	Low	2,810
1988-93[S]	0.38**	0.36**	Very Low	0
1994-97[S]	1.72*	1.62*	Moderate	7,000
Average for 1937-1997				8,400

S = Supplementary 1993 photos show little or no change between 1987 and 1993.
* = Activity estimated as 75 percent of total activity for full time period.
** = Activity estimated as 25 percent of total activity for full time period.

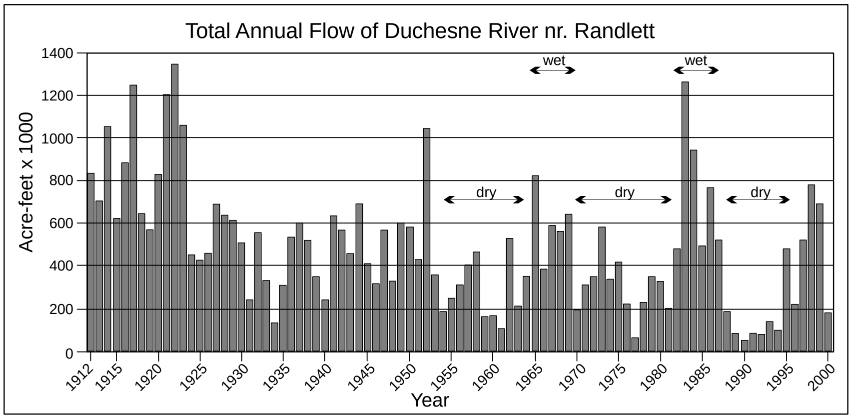


Figure 13: Graph showing the total annual flow of the Duchesne River near Randlett.

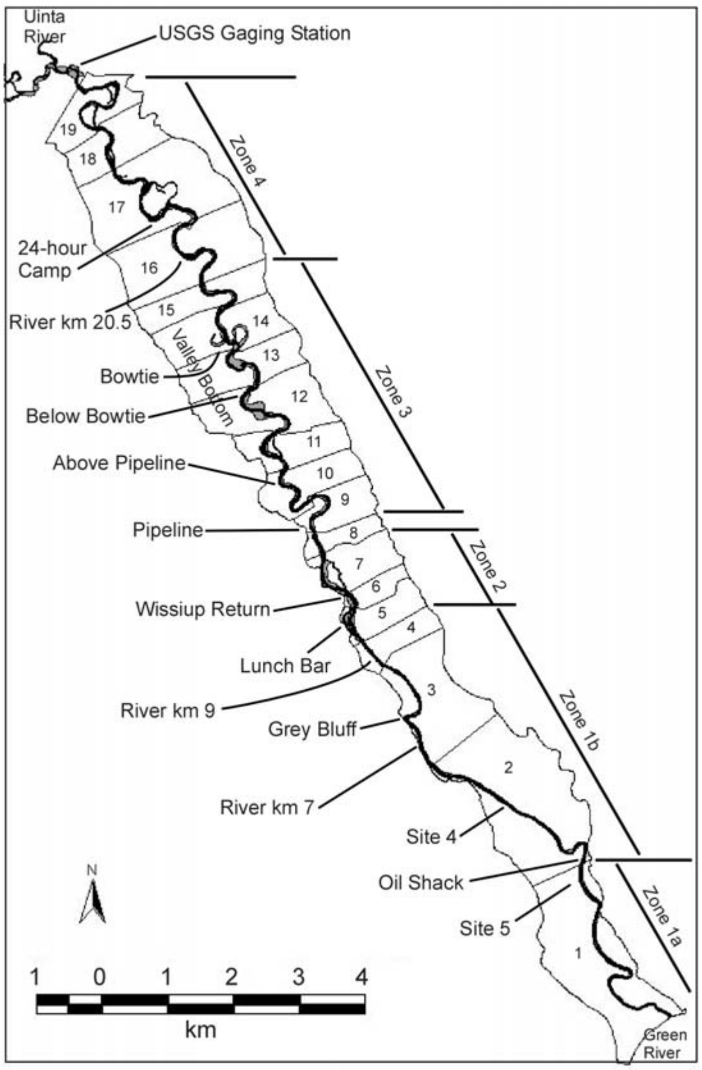


Figure 14: Map of the lower Duchesne River showing longitudinal zones, numbered subreaches, locations of detailed study sites, and landmarks.

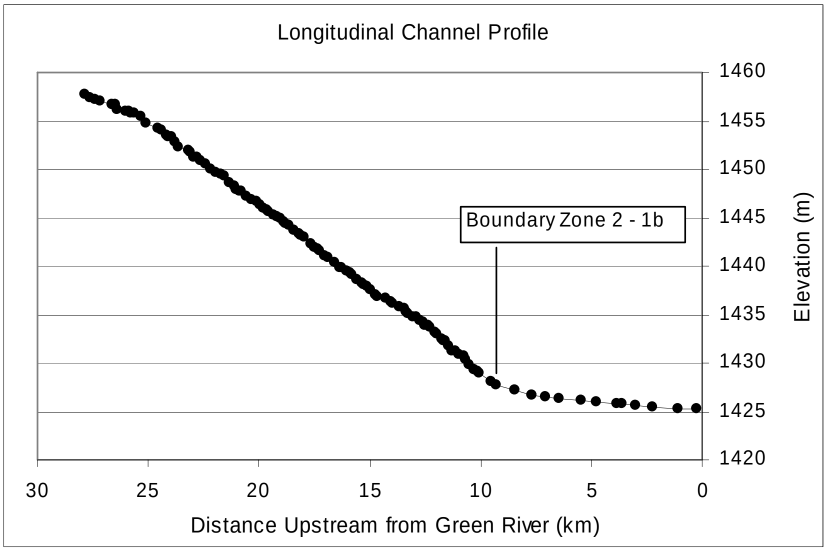


Figure 15: Longitudinal Profile of the Duchesne River Channel.

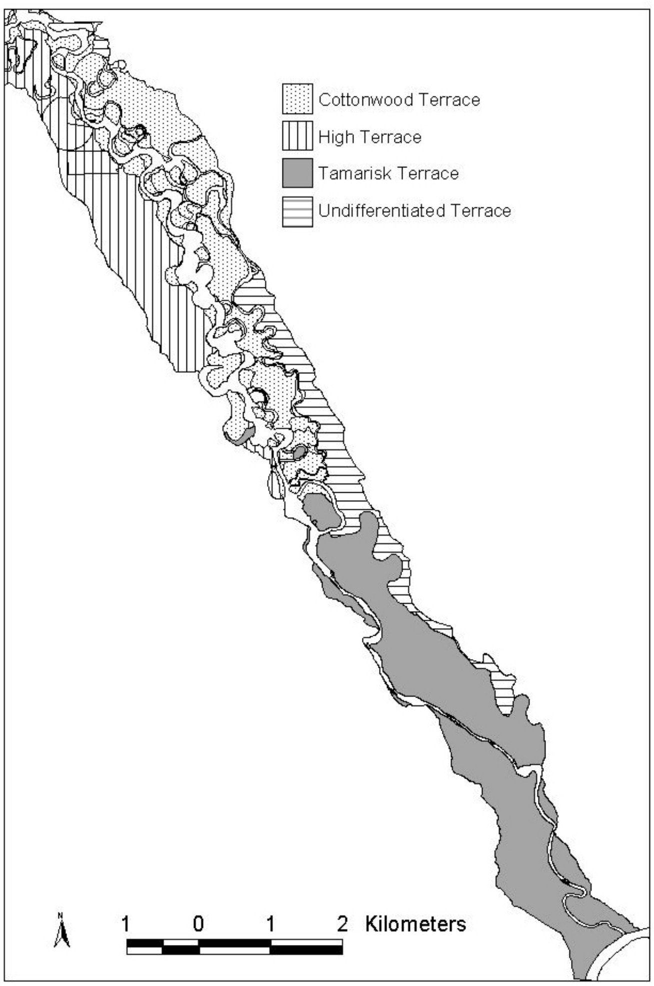


Figure 16: Map showing the distribution of terrace surfaces in the study area.

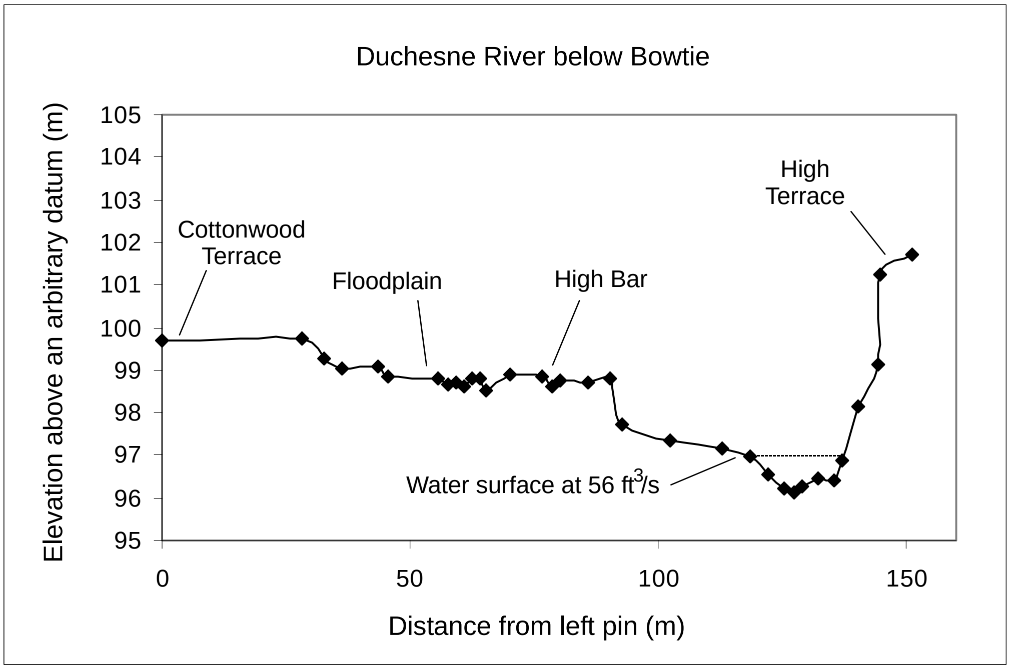


Figure 17: Cross section showing relative elevations of high terrace, cottonwood terrace, floodplain, high bars, and channel in Zone 3.

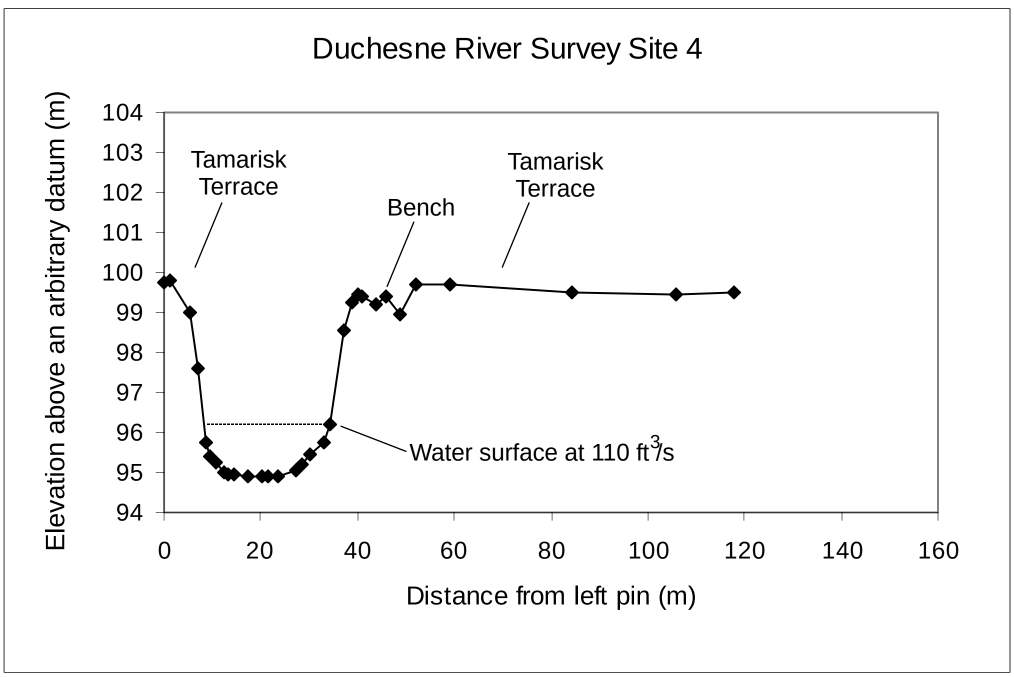


Figure 18: Cross section showing relative elevations of tamarisk terrace, bar/floodplain bench, and channel at site 4 in Zone 1b.

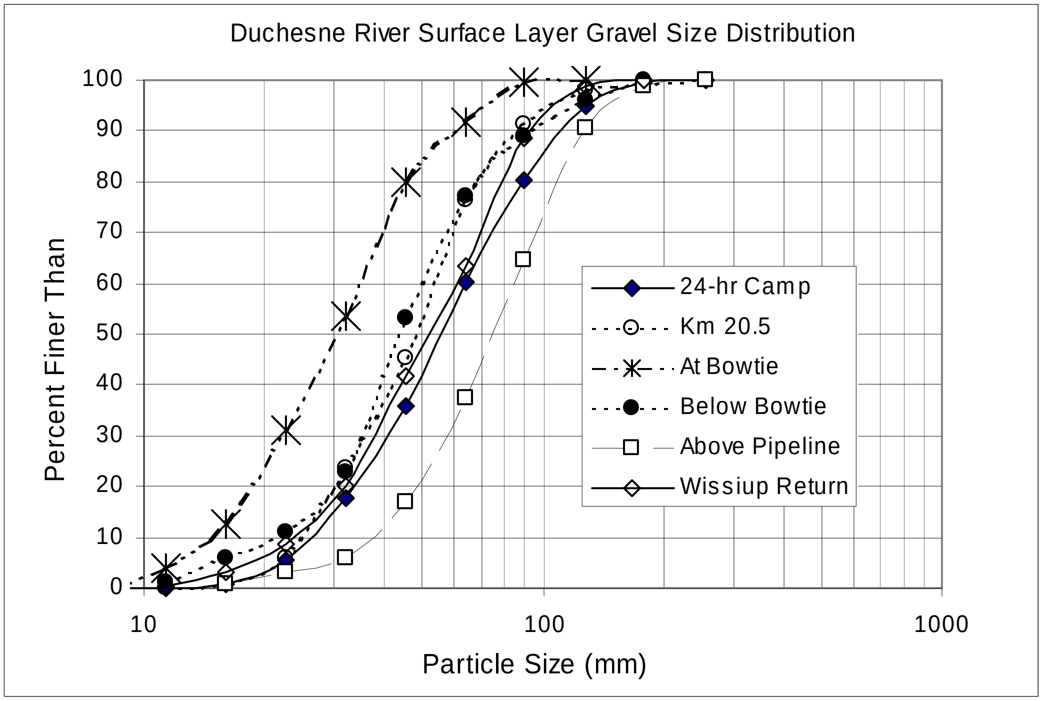


Figure 19: Surface layer particle size distributions.

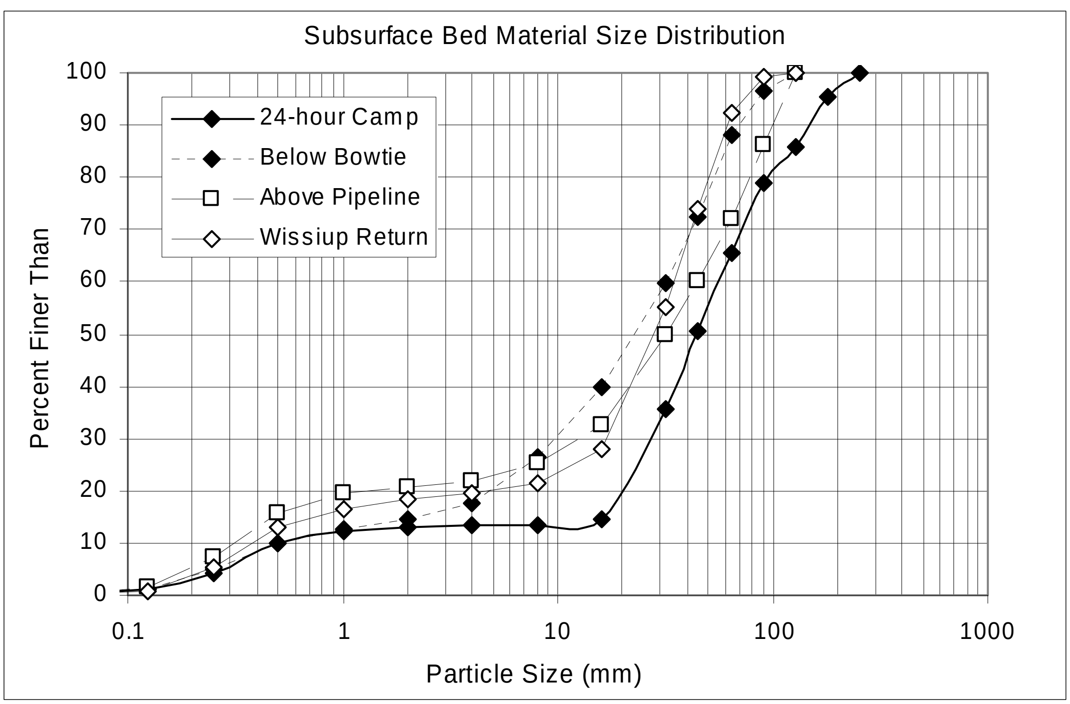


Figure 20: Subsurface particle size distributions.

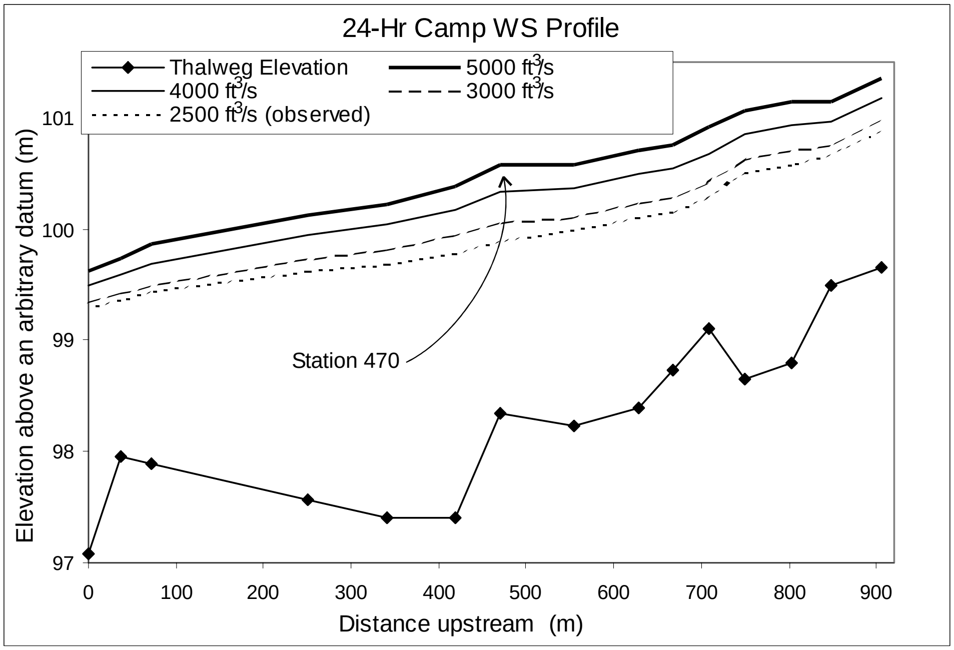


Figure 21: Longitudinal profile of bed and modeled water surface profiles at 24-hour Camp.

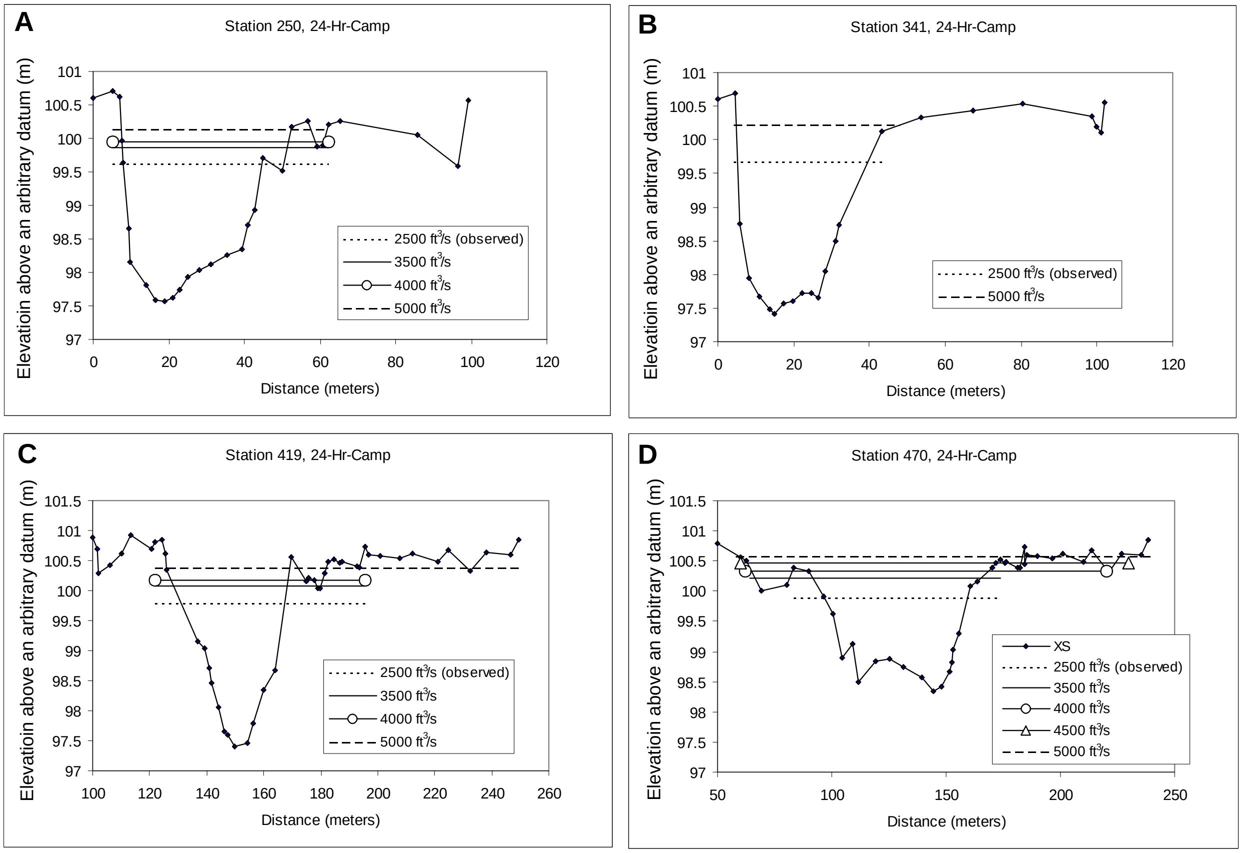


Figure 22: Cross sections and modeled water surface elevations at selected stations at 24-hour Camp.

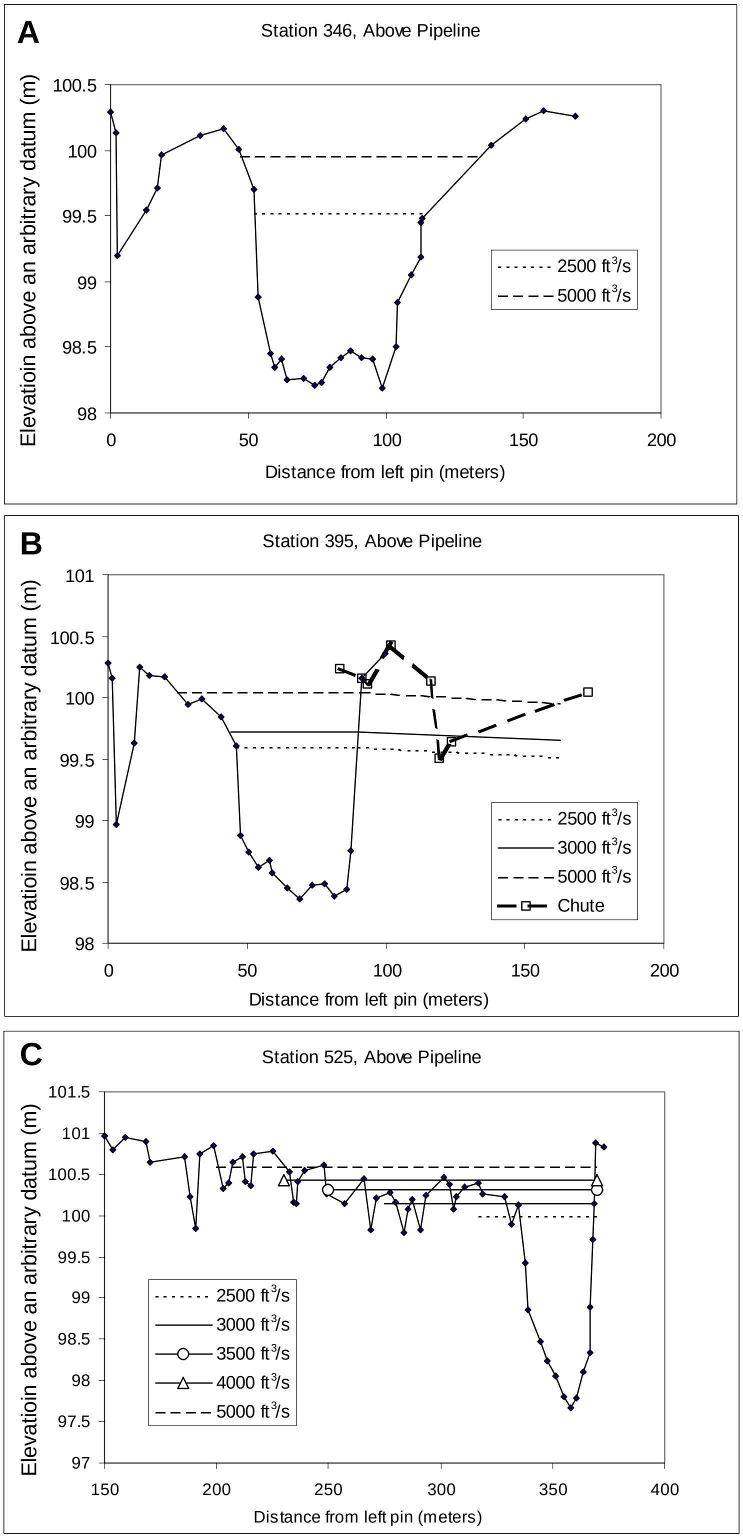


Figure 23: Cross sections and modeled water surface elevations at selected stations at Above Pipeline.

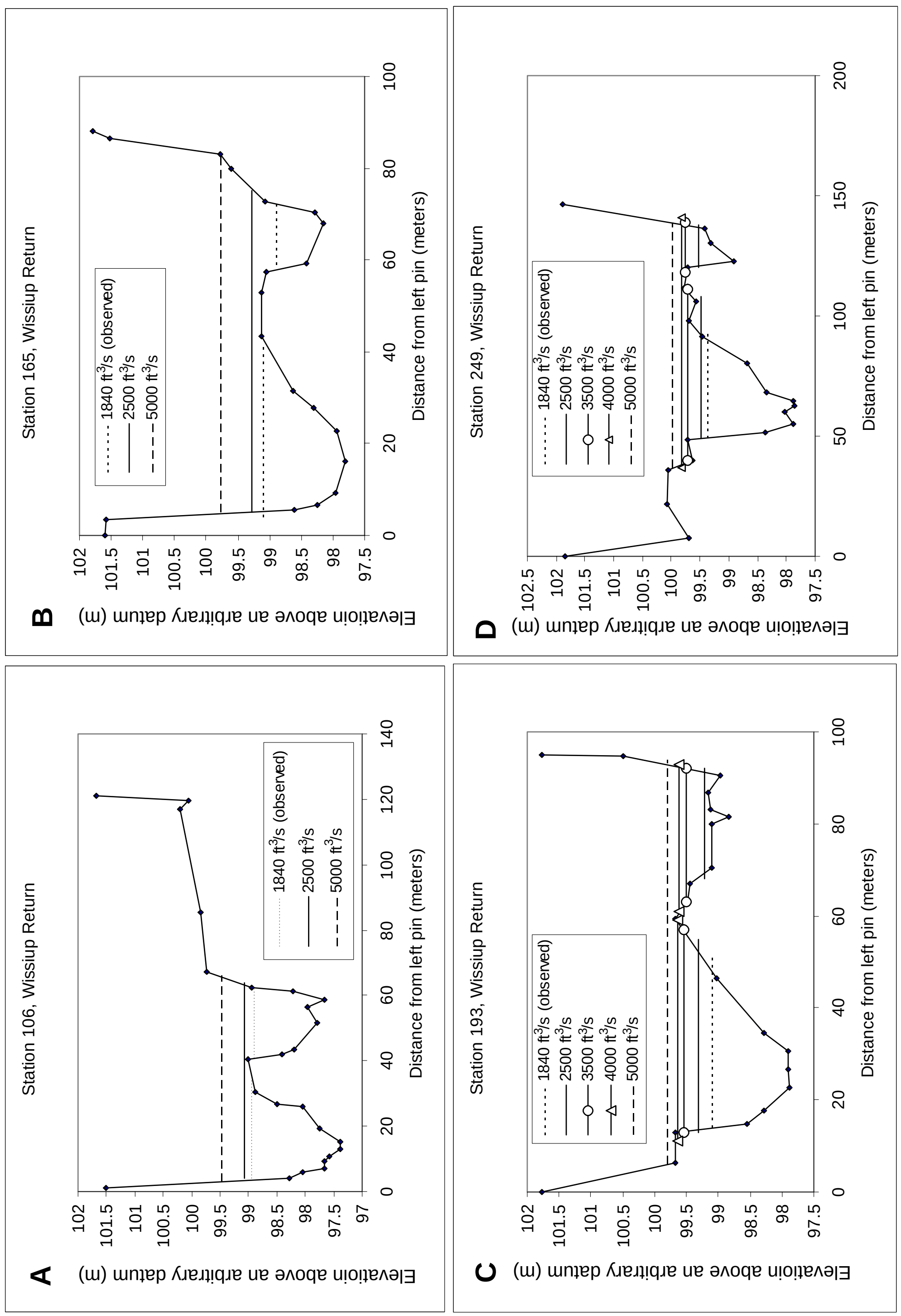
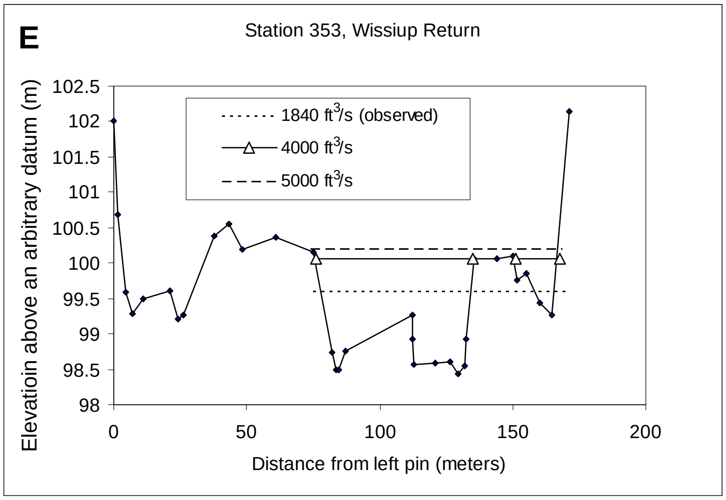


Figure 24 (A-D): Cross sections and modeled water surface elevations at selected stations at Wissiup Return.



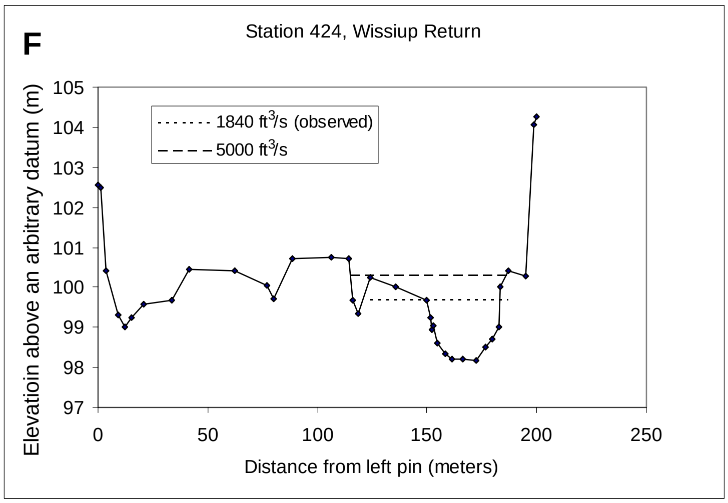


Figure 24 (E-F): Cross sections and modeled water surface elevations at selected stations at Wissiup Return.

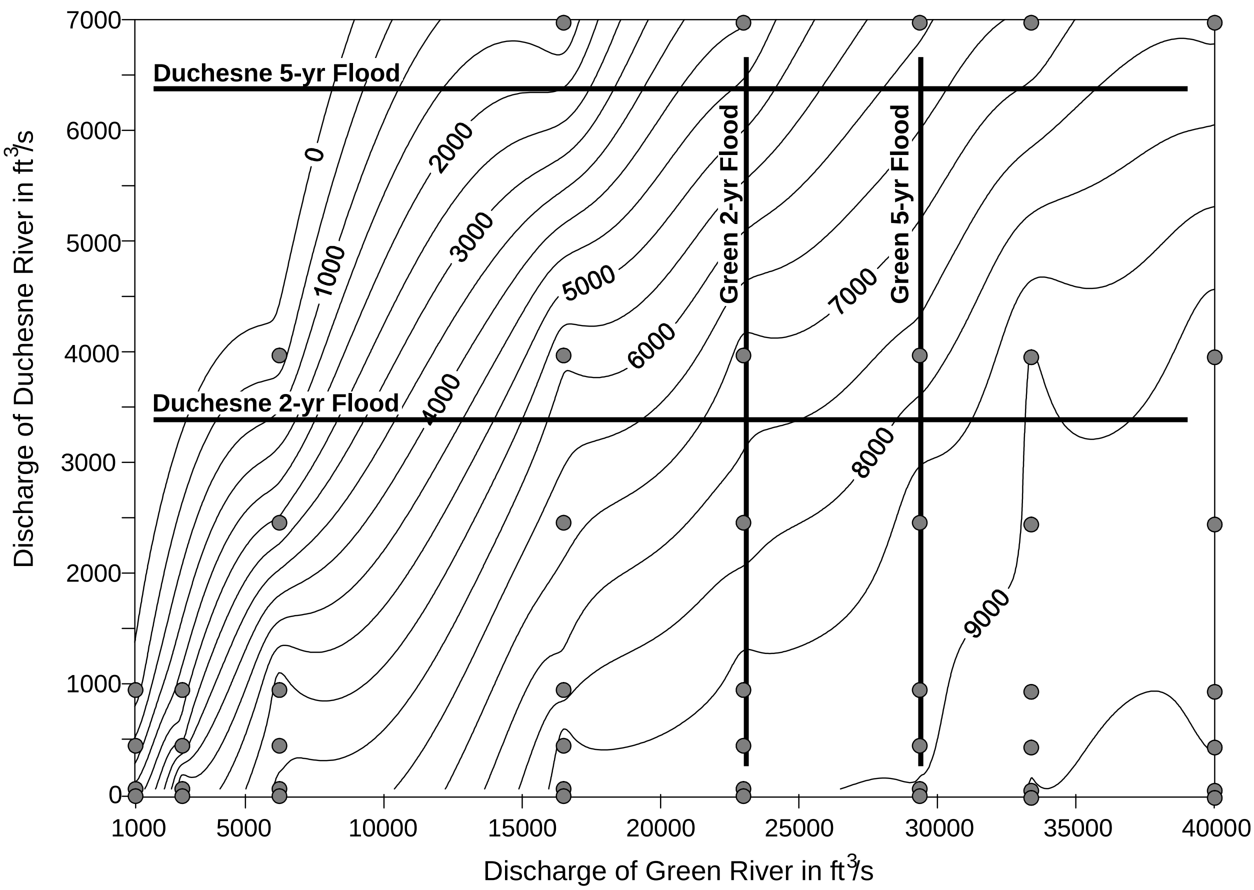


Figure 25: Contour diagram showing the upstream extent of Green River backwater on the Duchesne River. Contour lines indicate distance upstream, in m, from the Green River.

Figure 26: 1936 photograph showing the pre-avulsion bend near Wissiup Return and the reach downstream to Grey Bluff.

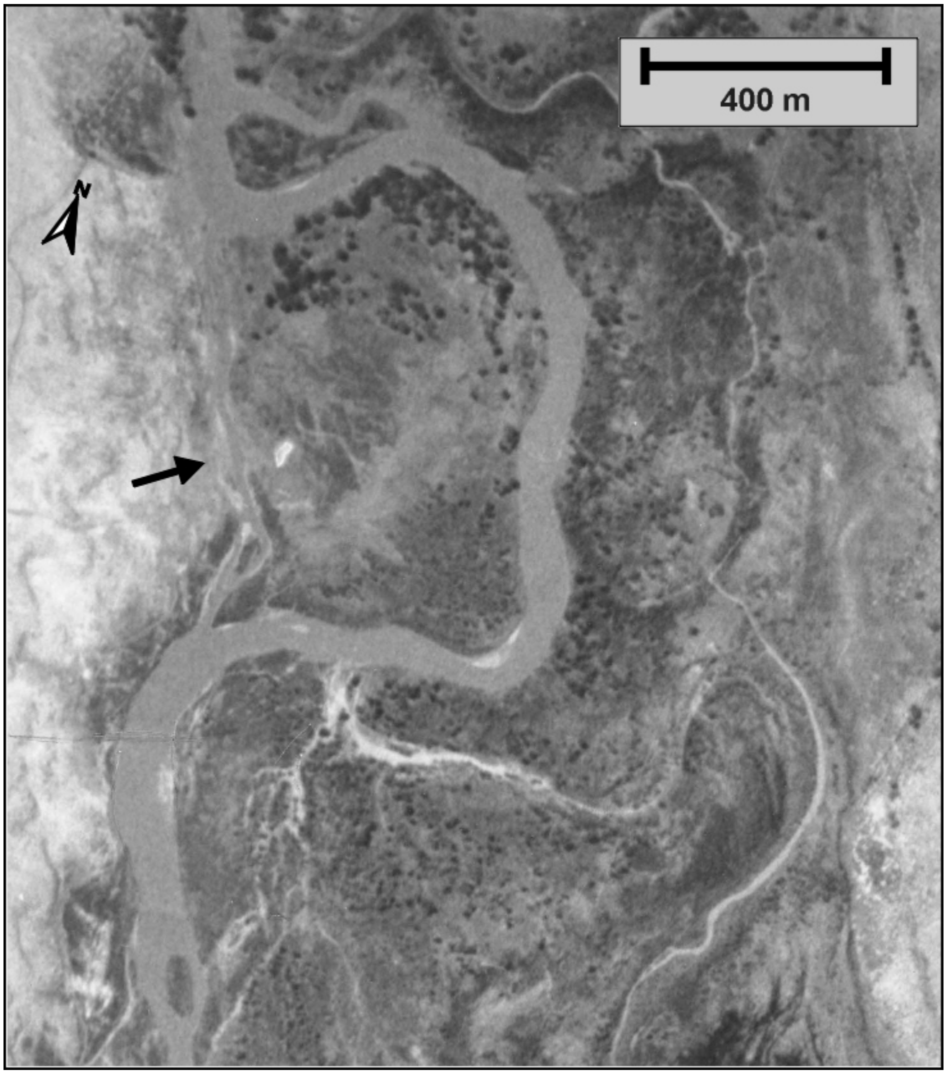


Figure 27: 1948 photograph of the bend near Wissiup Return showing avulsion developing across the bend.

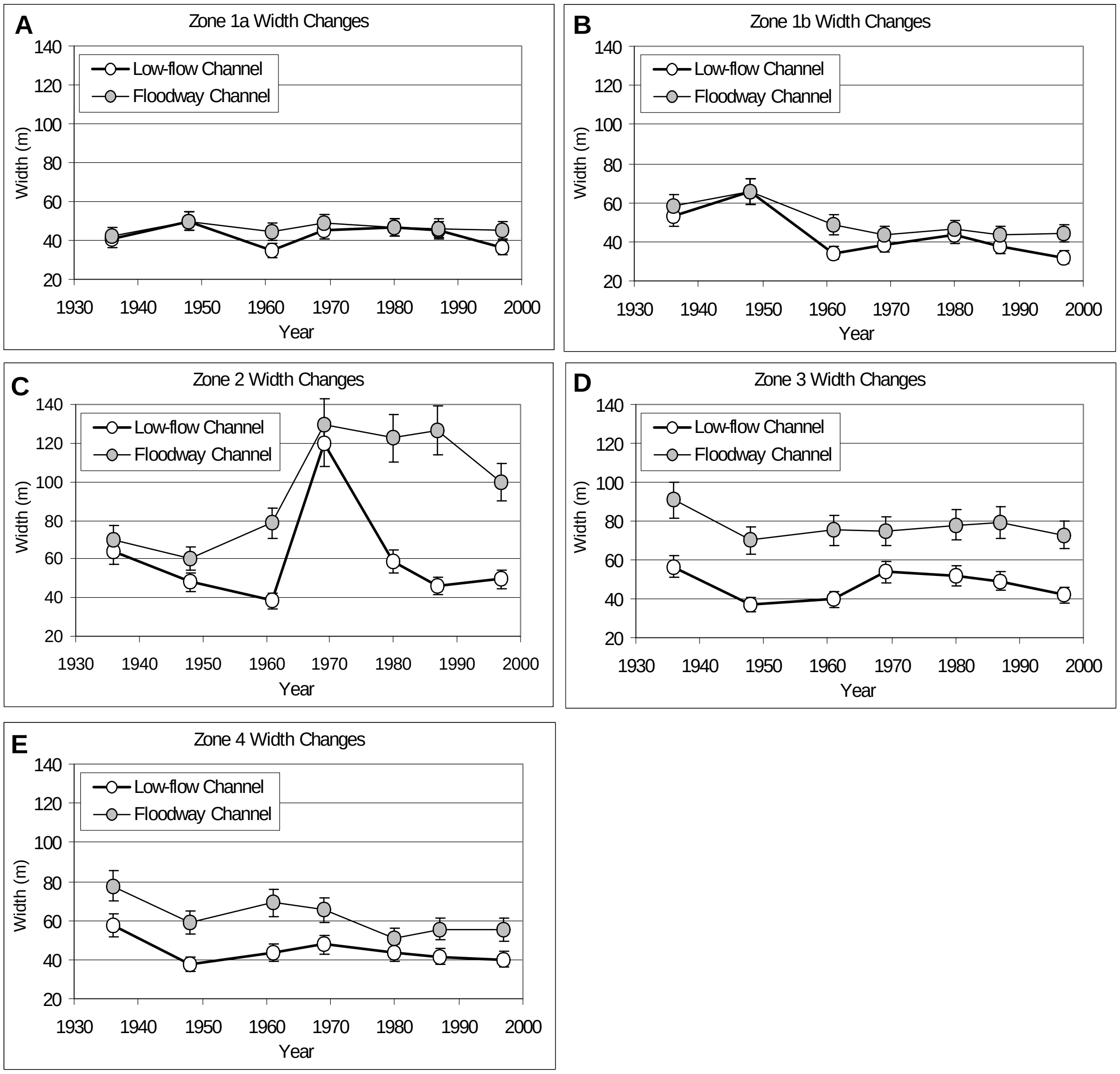


Figure 28: Graphs showing changes in channel width through time in each longitudinal zone.

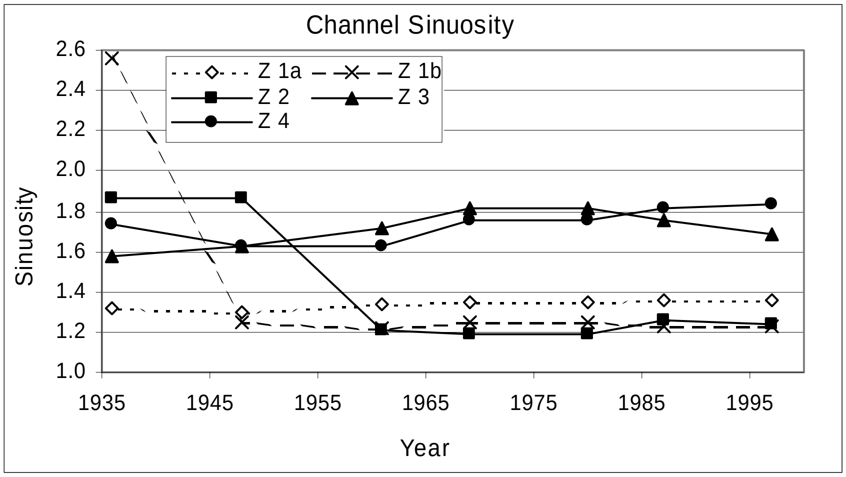


Figure 29: Graphs showing channel sinuosity through time in each longitudinal zone.

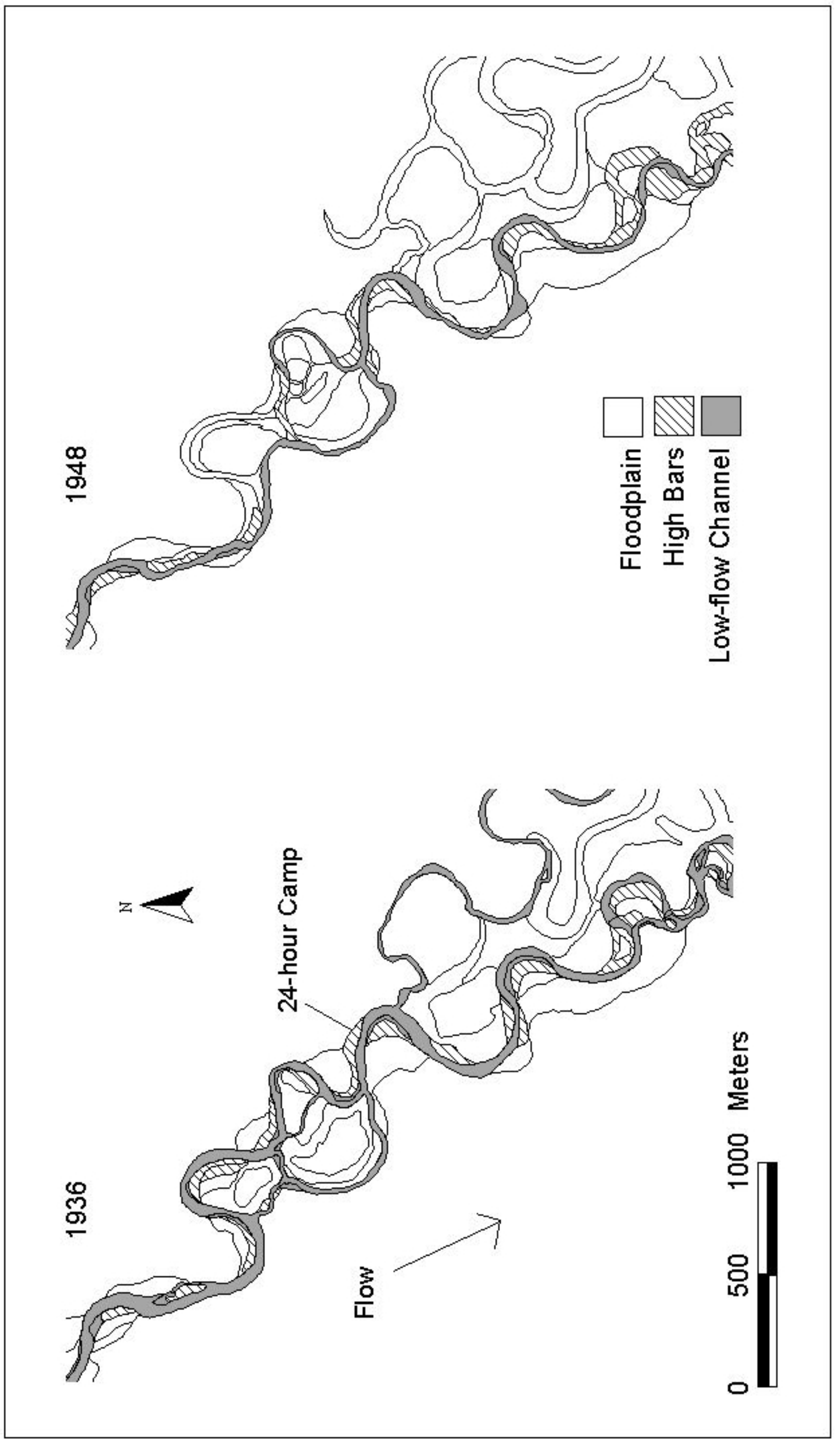


Figure 30: Maps showing the 24-hour Camp site and the middle part of Zone 4 in 1936 and 1948.

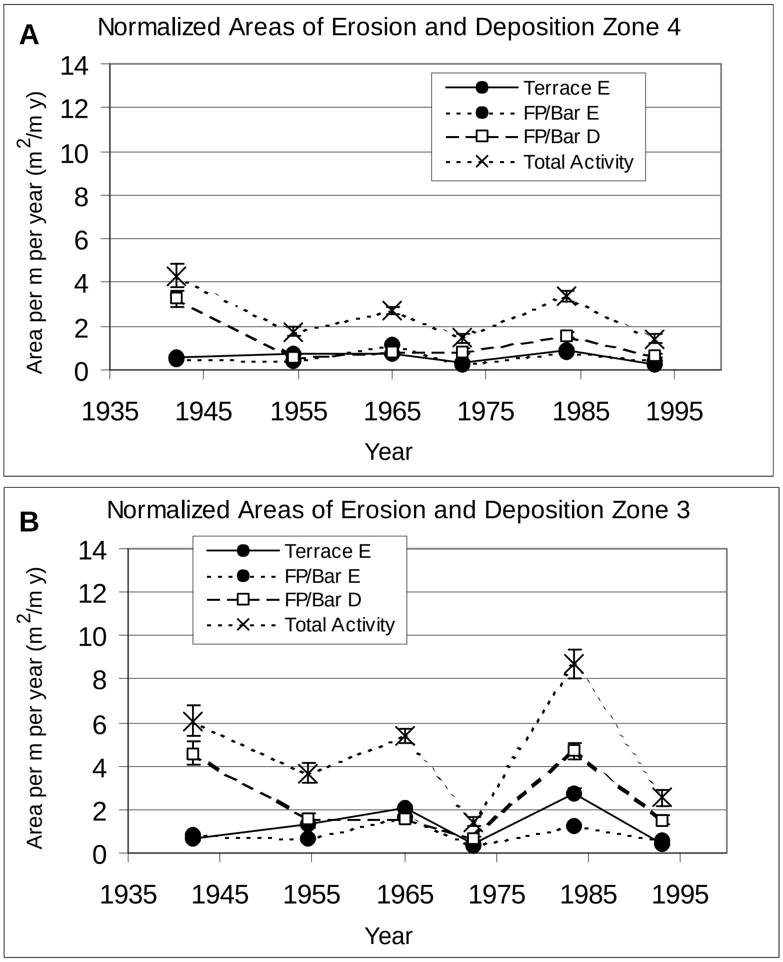


Figure 31 (A-B): Graphs showing normalized area of erosion and deposition in Zones 3 and 4 during six time intervals.

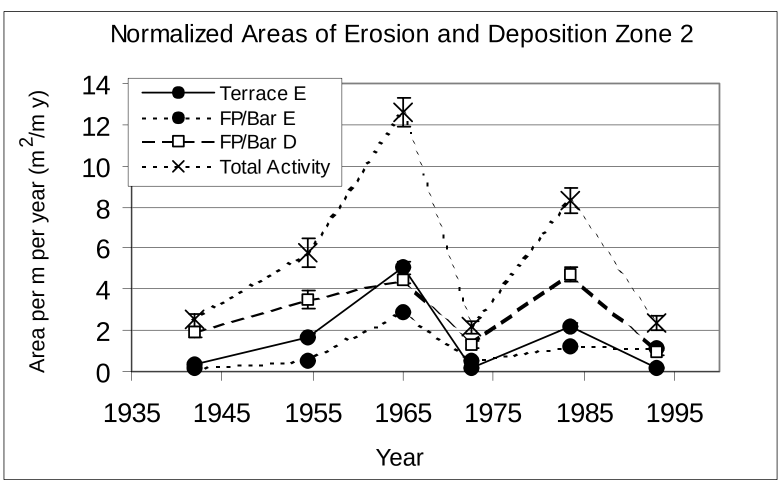
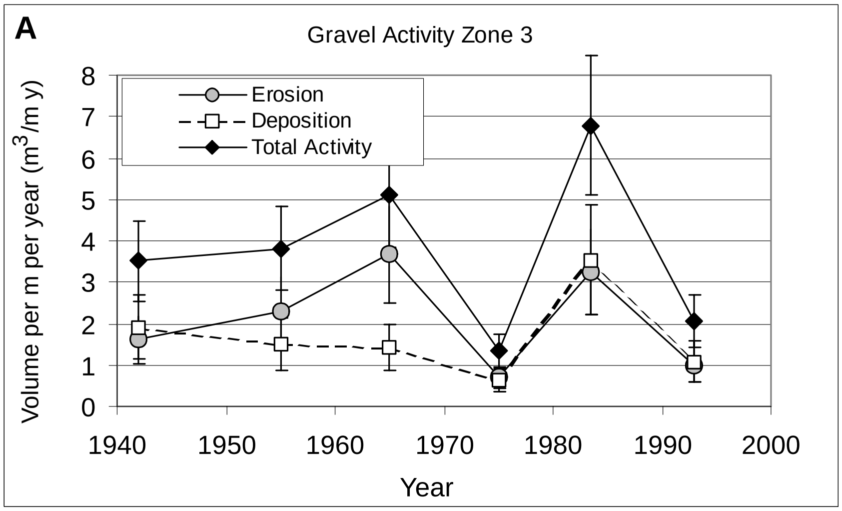


Figure 32: Graph showing normalized area of erosion and deposition in Zones 2 during six time intervals.



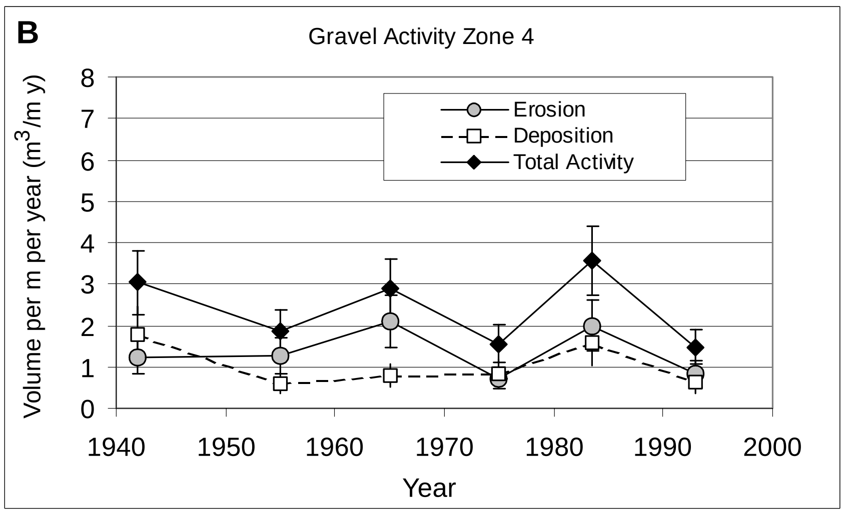


Figure 33 (A-B): Graphs showing gravel activity in Zones 3 and 4 through time.

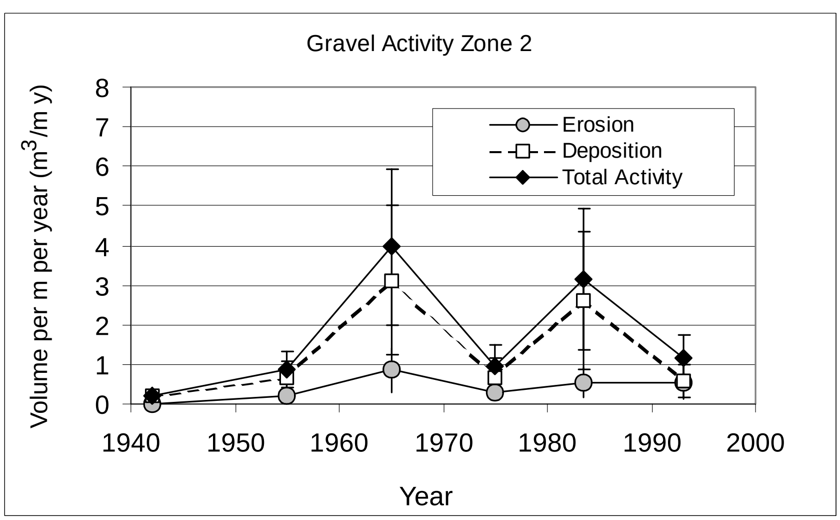


Figure 34: Graph showing gravel activity in Zone 2 through time.

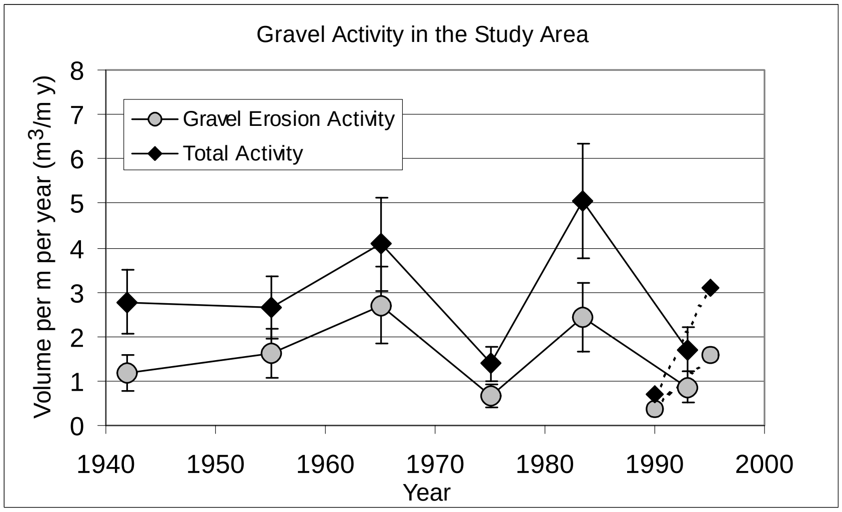


Figure 35: Graph showing composite total gravel activity and gravel erosion activity in Zones 2-4 through time.

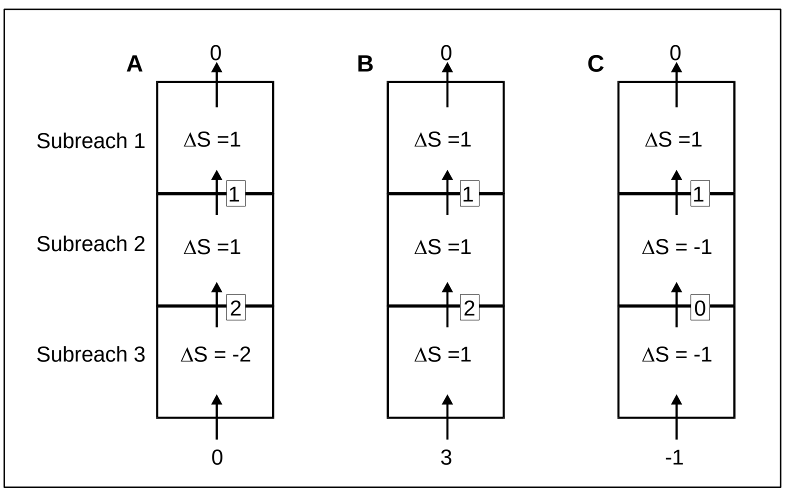


Figure 36: Diagram showing calculation for reach-scale gravel budgets using a zero-transport downstream boundary.

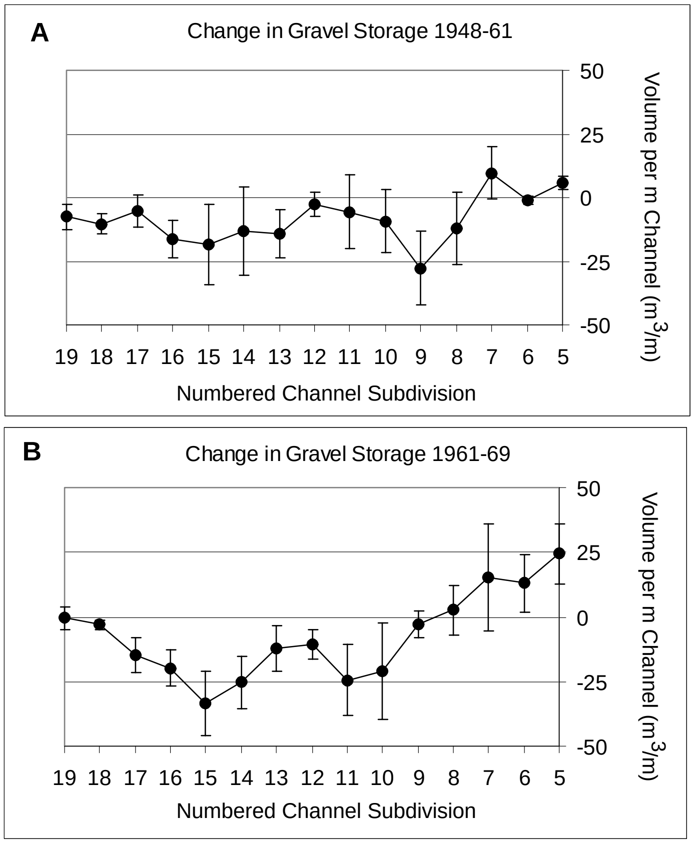


Figure 37 (A-B): Graphs showing longitudinal pattern of gravel storage changes between 1948 and 1961, and between 1961 and 1969.

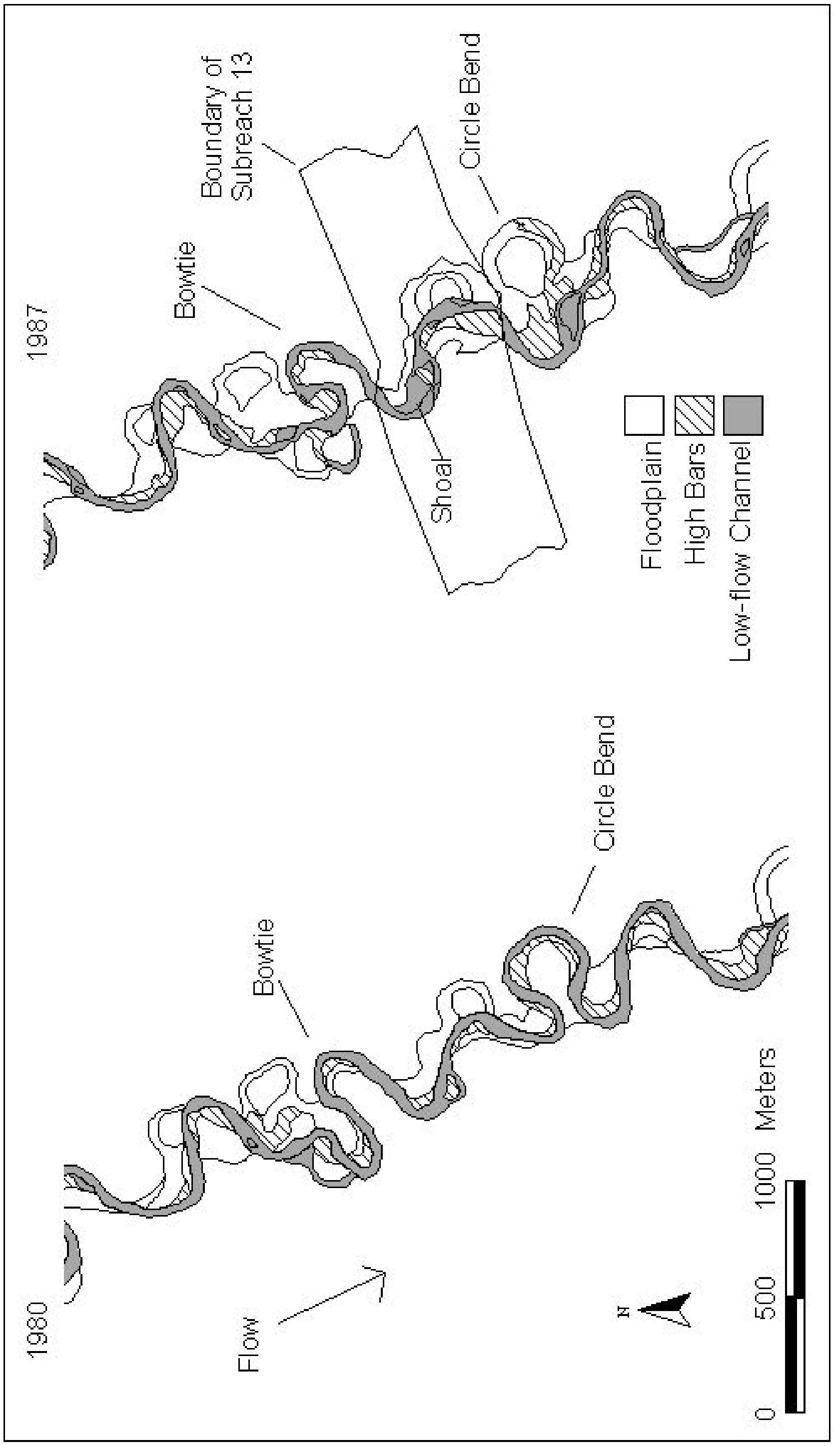


Figure 38: Maps showing the middle part of Zone 3 in 1980 and 1987.

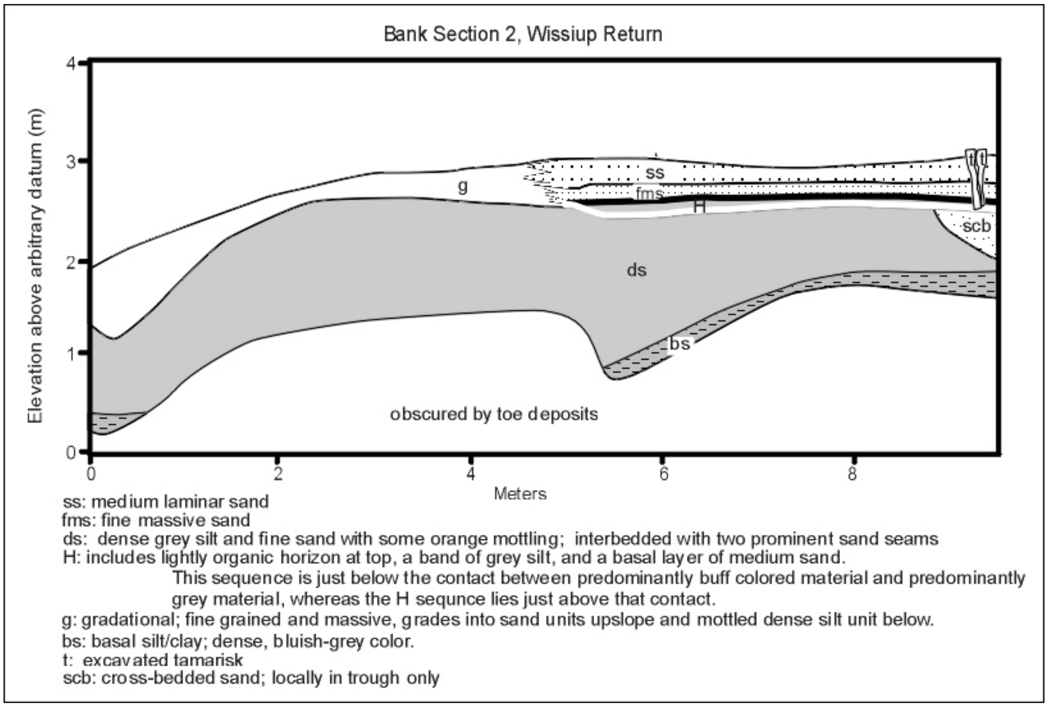


Figure 39: Diagram of cutbank stratigraphy at the edge of the pre-avulsion channel near Wissiup Return.

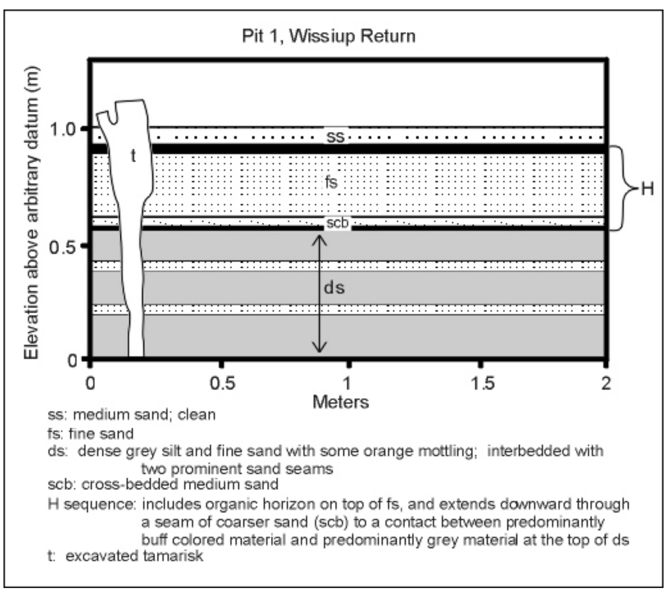


Figure 40: Diagram of stratigraphy in a pit in the tamarisk terrace near Wissiup Return.

DISCUSSION

Many attributes of the natural stream flow regime contribute to the diversity of aquatic and riparian habitats in channels and valleys (Poff et al. 1997). Thus, there are many aspects of the natural flow regime that must be considered in developing a comprehensive recommendation of the flows necessary and sufficient to maintain aquatic and riparian ecosystems. Alluvial valleys include the stream channel, the active floodplain, and higher terraces occasionally inundated by flood flows. Each of these areas plays a role in the ecological functions of the channel/valley system. We have pursued a strategy to quantifying flow needs to maintain a wide range of geomorphic and ecological habitats in conditions similar to their present states.

Determination of in-stream flow requirements necessitates determining the minimum discharges necessary to maintain some target physical condition of the channel and floodplain. River discharge is necessary to maintain a suite of ecologically important geomorphic environments:

1. The characteristics of riffles are maintained by frequent entrainment of gravel, which prevents long-term embedded conditions and accumulation of fines within the gravel framework;
2. Inundation of chute channels and the active floodplain is necessary to maintain channel complexity and to maintain connection between riparian and in-channel ecological processes;
3. High flows of sufficient magnitudes and durations to transport gravel and produce bank erosion are necessary to maintain bend extension and cutoff, exchange of sediment between the channel, floodplain, and terraces, and the overall characteristic of the lower Duchesne River as a dynamic and unstable channel/floodplain system;
4. Fine sediment delivered to the lower Duchesne River must be transported through the reach in order to prevent channel simplification caused by deposition in backwater habitats.

Fisheries studies presented in other chapters of this report have identified the entire gravel-bed portion of the lower Duchesne River as habitat used by Colorado pikeminnow. Our measurements demonstrate that the portion of the lower Duchesne River that is currently gravel-bedded can be subdivided into three zones of different characteristics, different histories, and different future trajectories of change. It is a challenge to interdisciplinary science to further identify the relative significance of these zones in terms of the habitat needs of endangered Colorado River fishes.

Gravel Mobilization

Riffles are maintained by the frequent mobilization of bed material gravels. Entrained gravels typically move during flood and come to rest at places that have lower shear stress than nearby areas. At lower discharges, these areas of temporary accumulation of gravel become riffles and are places of relatively high in-stream productivity.

Many riffles on the lower Duchesne River have been in similar locations for the past 70 years. At present, some of these riffles contain loose, easily moved cobbles and gravel. However, other sites contain gravels that are interlocked with one another and which include a large proportion of fine-grained sediments. These riffles are likely to be much less productive.

Our gravel mobility analyses shows that gravel entrainment on the bed of Duchesne River becomes widespread through riffles and runs at discharges of about 4,000 ft^3/s. Limited entrainment at some isolated locations may begin at discharges as low as about 2,500 ft^3/s, while entrainment at other locations may require more than 5,000 ft^3/s. These estimates were made at 15 riffle or run locations at three study sites, and are subject to the uncertainty inherent in modeling.

Inundation of the Floodplain and Other Adjacent Surfaces

Our study demonstrates that the alluvial surfaces adjacent to the Duchesne River are typically inundated by floods with long-term recurrence intervals of 2 to 2.6 years. Inundation of these surfaces therefore occurs frequently during wet periods, but is less frequent during dry periods. Connection between the channel and floodplain occurs by inundation of chute channels and by local overtopping of high bars. Local overtopping of high bars adjacent to the low-flow channel occurs at higher discharges of about 4,000 ft^3/s. Daily mean discharges of 4,000 ft^3/s have a recurrence interval of 1.3 years during wet periods and 3.6 years during dry periods. Terrace surfaces are not inundated except during rare floods. Flow into the main chutes and side channels occurs over a wide range of discharges averaging about 3,000 ft^3/s. This magnitude of daily mean discharge has a recurrence interval of about 2.4 years during dry periods and of about 1.1 years during wet periods. Higher floodplain surfaces are rarely inundated, requiring flood events with a recurrence interval of approximately 6 years. Inundation of the valley bottom downstream of the Oil Shack occurs in association with backwater flooding from the Green River. Its frequency is therefore controlled by the flow regime of the Green River.

The importance of floodplain inundation extends beyond maintenance of ecological connection between channel and floodplain. Inundation helps to retard invasion of riparian shrubs and trees onto the floodplain surface. Invasion of such shrubs and trees encourages vertical aggradation of fine sediment that ultimately leads to channel narrowing (Allred and Schmidt 1999) and increases the potential for channel adjustment.

Channel-forming Discharges

Our gravel entrainment and inundation analyses converge on a threshold discharge of about 4,000 ft^3/s for mobilizing a significant portion of the bed and inundating some overbank surfaces. This discharge level has a long-term recurrence interval of about 2.4 years and is exceeded about 1.6 percent of the time. The frequency of this threshold discharge is broadly consistent with its interpretation as the approximate bankfull or channel-forming discharge. However, the recurrence intervals for floods of this magnitude increased from 2.2 years for the period of record before completion of the Bonneville Unit of the CUP to about 3 years since Bonneville Unit diversions began. A consistent trend of channel narrowing since 1970 is in part the product of less frequent inundation of bar and floodplain surfaces by channel-forming flows. Prolonged periods with no flow exceeding the channel-forming threshold allow riparian vegetation to become established in the channel. Even large flood events, including the flood of record in 1983, have become less effective in restoring channel dimensions since the frequency of channel-forming flows declined after 1972. The recurrence interval of daily mean flows equal to or greater than 4,000 ft^3/s should therefore be kept at its long-term historical value of 2.4 years.

Comparison of the history of channel changes and activity rates with the historic hydrology demonstrates that the level of channel activity is low when flows above this critical discharge threshold are less frequent. The data indicate that an average annual channel-forming stream flow volume of at least 7,000 ft^3/s-days per year is needed to promote channel migration and maintain channel integrity. The total water volume necessary to attain 7,000 ft^3/s-days per year of channel-forming discharge depends on the magnitude of the floods contributing to the total, with less total water volume being required by larger floods of shorter duration. For practical purposes, it is necessary to cast this average annual channel-forming flow target in terms a few well-defined hydrographs that can be implemented in years when sufficient flow volumes are available. Below, we define two hydrographs designed to achieve a long-term average channel-forming flow volume of 7,000 ft^3/s-days per year. The smaller of the two hydrographs will produce 7,000 ft^3/s-days of channel-forming flow, and can be implemented on a relatively frequent basis. The larger of the two hydrographs maintains the long-term average channel-forming flow volume at the target level by taking advantage of relatively rare flow events. This approach recognizes that infrequent larger floods are 1) critical for maintaining physical habitat in the lower Duchesne River, and 2) are not significantly reduced by the existing diversion/storage infrastructure.

To develop a stream flow hydrograph that will produce 7,000 ft^3/s-days of channel-forming flow and can be implemented in moderately wet years, we analyzed the flow duration characteristic of eight moderately wet water years that averaged 7,008 ft^3/s-days of channel-forming flow. These eight years constitute all years from the gaged flow record for which the channel-forming flow volume was between 3,000 ft^3/s-days and 10,000 ft^3/s-days. During these years, the channel-forming threshold of 4,000 ft^3/s was exceeded an average of 10.7 days per year. The remainder of the hydrograph in excess of the threshold is well-described by the following magnitude/duration combinations: 4,400 ft^3/s exceeded 7 days per year, 4,700 ft^3/s exceeded 4 days, 5,100 ft^3/s exceeded 2 days, 5,600 ft^3/s exceeded 1 day per year (Figure 41). If implemented in all years in which the channel-forming threshold is exceeded, this hydrograph will have a recurrence interval of about 2.4 years, or approximately 40 percent of all years.

Because the hydrograph described above will occur in only 40 percent of the years, it will actually produce a long-term average channel-forming flow volume of only 2,800 ft^3/s-days. In some extremely wet years, however, much larger flow events will occur. During the six wettest years on record, about 55,000 ft^3/s-days of channel-forming flows were discharged on average. By definition, these years occur in about 1 year out of 10 and therefore contribute 5,500 ft^3/s-days per year to the long term average channel-forming flow volume. The simplest approach for specifying the hydrology of these extremely wet years is to simply regard them as years in which the 10-year peak event occurs, and all magnitude-duration combinations proposed for moderately wet years are also attained. The instantaneous 10-year peak flow on the lower Duchesne River is approximately 8,400 ft^3/s.

If 10 percent of the years are to be regarded as extremely wet, the percentage of moderately wet years must be adjusted to 30 percent (40 percent – 10 percent = 30 percent). The high flow hydrograph proposed above for moderately wet years will then contribute 2,100 ft^3/s-days per year to the long-term average. Summing the contributions of the wet and extremely wet years to the long-term average channel-forming flow yields a total of 7,600 ft^3/s-days per year.

Transport of Fine Sediment through the Lower Duchesne River

It is critical that fine sediment being supplied to the lower Duchesne River from all sources be transported through the reach. Otherwise excess sediment will be stored in the reach, affecting river morphology and the quality of the substrate. Reductions in stream flow in a river reach may be accompanied by reductions in the rate of suspended sediment transport through the reach. However, there may not be a corresponding decrease in the quantity of fine sediment supplied to the channel. We have demonstrated that local sediment sources are important components of the suspended sediment budget of the lower Duchesne River. The magnitude of these sources may remain the same or even increase, independent of changes in stream flow. Changes in the balance between the sediment supply and the sediment transport rate may cause the stream channel to reconfigure its geometry or plan form so that the imposed water and sediment loads can be transported through the system (Schumm 1969). For example, a 50-percent reduction in suspended sediment transport capacity caused by flow depletion on the upper Colorado River resulted in channel narrowing and loss of side channel habitat within a decade (Van Steeter and Pitlick 1998).

The estimated historical fine sediment flux at the USGS gaging station near Randlett can be maintained with less water than the historical stream flow if the stream flow is used in an efficient manner. Numerous hydrologic scenarios could be defined for transporting the necessary quantity of sediment. Because the channel-forming discharges specified above provide a large portion of the required fine sediment transport capacity, additional flows needed to transport the balance of the historical sediment load are best specified after the channel-forming targets have been determined. We therefore calculated fine sediment loads according to the hydrology defined by the proposed channel-forming flows, i.e., for extremely wet years, wet years, and all other years. For purposes of fine sediment transport analysis, it is useful to further subdivide the 60 percent of all years not classified as either extremely wet or wet into two groups. These two additional groups are defined as normal years (years with total annual flow between the 30^{th} and 60^{th} percentiles) and dry years (years with total annual flow less than the 30^{th} percentile).

For this analysis, we assume that the sediment rating relation computed for rising limb flows can be applied to all flows. This assumption is justified on the basis that suspended sediment concentrations are a function of sediment supply as well as a function of discharge. During periods of low stream flow and decreased sediment throughput, the supply of fine sediment within a reach will increase. As more fine material becomes available to the flow, the concentration of suspended sediment in the flow will increase (VanSickle and Beschta 1983; Rubin and Topping 2001). This supply-dependence of suspended sediment concentrations accounts for the differences observed in the rating relations between the rising and falling limb flows, and may reduce the need for changes in channel geometry to compensate for small reductions in discharge.

Applying our rising-limb sediment rating relations to representative annual hydrographs for each water year class yields sediment loads for each water year class, and use of a cumulative sediment transport curve allows graphical identification of the loads transported by specified portions of each hydrograph (Figure 42). The cumulative sediment transport curve for extremely wet years (total annual flow exceeding the 90^{th} percentile) is based on the assumption that the rise and fall of the hydrograph peaking with the 10-year flood will be similar to the historical hydrographs for flows greater than 3,000 m^3/s. Flow greater than 3,000 m^3/s in extremely wet years can transport an estimated cumulative sediment discharge equal to about 350 percent of the mean annual load (Figure 42).

During wet years (total annual flow between the 60^{th} and 90^{th} percentiles) application of the rising-limb rating relation to the hydrograph proposed for channel-forming flow greater than 4,000 m^3/s indicates that this hydrograph can transport about 83 percent of the mean annual load. Historically, an average of seven additional days on which flows were between 3,000 m^3/s and 4,000 m^3/s occurred during wet years. If a similar ramping rate between 3,000 m^3/s to 4,000 m^3/s is implemented, an additional 50 percent of the mean annual load will be transported. A total of 133 percent of the mean annual load can therefore be transported during wet years by implementing channel-forming flows, plus an additional 7 days on which discharges are greater than 3,000 m^3/s. Multiplying the loads transported during extremely wet years and wet year by the fraction of years in which these flows occur yields the total percentage of the mean annual load that can be transported in these years. The calculations indicate that 35 percent of the mean annual load can be transported by the channel-forming flow requirements in extremely wet years. In wet years, 40 percent of the mean annual load can be transported by flows consisting of the channel-forming flow targets, plus the additional duration target for flows greater than 3,000 m^3/s.

A water-sediment efficiency curve shows that sediment transport efficient in normal years (total annual flow between the 30^{th} and 60^{th} percentiles) begins to decline as cumulative discharge exceeds about 22 percent (Figure 43). Above this threshold, proportionally larger volumes of water are needed to increase sediment transport by a given increment. This cumulative discharge level corresponds with a discharge of 2,000 m^3/s. It would therefore be inefficient to reserve discharges of less than 2,000 m^3/s for fine sediment transport. Referring to Figure 42, all normal-year discharges greater than 2,000 m^3/s will transport about 61 percent of the mean annual sediment load, and all discharges greater than 3,100 m^3/s will transport about 35 percent of the mean annual sediment load. The normal-year interval between 2,000 m^3/s and 3,100 m^3/s will therefore transport about 26 percent of the mean annual load. Historically, a volume of water equal to 3,785 m^3/s-days above 2,000 m^3/s have been discharged in this flow band. This volume of water discharge above 2,000 m^3/s can be achieved by maintaining discharges greater than 2,500 m^3/s for 7.5 days. Because only 30 percent of the years are designated as normal years, this target will transport about 8 percent of the mean annual load overall.

Summing the percentages of the mean annual sediment load transported by the above flow regimes for extremely wet, wet, and normal years indicates that about 83 percent of the mean annual sediment load can be transported by the proposed targets. Portions of the annual hydrographs not considered above include the dry year runoff hydrograph, and the base-flow portions of the all years. By equating these stream flows to the discharges exceeded 50-percent of the time for all years, we estimate that at least 5 percent of the mean annual sediment load is transported by these remaining flows. This brings the total suspended sediment load transported by the proposed flow regime to within 90 percent of the historical mean annual load.

A Flow Regime for Channel and Habitat Maintenance

We propose a flow regime for maintenance of current habitat conditions in the lower Duchesne River incorporating three main components. These components are a minimum frequency with which the channel-forming flow threshold must occur, a minimum total volume of discharges in excess of the channel-forming threshold, and some moderate flows less than the channel-forming threshold to maintain adequate fine sediment transport capacity. The respective objectives of these components are to prevent the establishment of riparian vegetation within the existing channel, to ensure that gravel mobilization and processes of channel migration continue, and to balance the fine sediment budget for the reach.

These components can be implemented for each runoff season by classifying the year either as an exceptionally wet year (total annual flow exceeding the 90^{th} percentile), a wet year (total annual flow between the 60^{th} and 90^{th} percentiles), a normal year (total annual flow between the 30^{th} and 60^{th} percentiles), or a dry year (total annual flow less than the 30^{th} percentile), and applying the appropriate stream flow criteria (Table 23). By definition, 10 percent of all years will be exceptionally wet, 30 percent will be wet, 30 percent will be normal, and 30 percent will be dry. In terms of total annual flow volume, exceptionally wet years are those years in which the total annual discharge at the Randlett gaging station is greater than 765,000 acre-ft. The total annual discharge in wet years is between 435,000 acre-ft and 765,000 acre-ft. In normal years the total annual discharge is between 224,000 acre-ft and 435,000 acre-ft, and in dry years it is less than 224,000 acre-ft.

In 30 percent of the years (wet), daily mean discharges should exceed 4,000 ft^3/s for 10 days or more, and exceed 5,600 ft^3/s for at least one day. Intermediate discharge/duration combinations are given in Table 23. These discharges are designed to produce a total of 7,000 ft^3/s-days of channel-forming flow. In addition, daily mean discharges should exceed 3,000 ft^3/s for at least an additional 7 days (17 days total) to provide for fine sediment transport capacity. In the wettest 10 percent of the years (exceptionally wet), the proposed flow targets consist of an instantaneous peak discharge of 8,400, plus all lower flow magnitude-duration combinations suggested for wet years. In the 30 percent of the years classified as normal years, no channel-forming flows will occur. However, we propose that discharges in excess of 2,500 ft^3/s be maintained for a minimum of 7 days to provide for fine sediment transport capacity in these years. In the driest 30 percent of the years (dry years), no channel maintenance flows of any kind are proposed.

The Impacts of Future Withdrawals

The existence of wet and dry hydrologic periods in the Duchesne River basin over the past 50 years has been demonstrated in this report. Implementation of a flow protection policy involves recognition of these wet and dry cycles. The channel and floodplain of the Duchesne River is primarily maintained during the wet periods, and the channel likely accumulates sediment during dry periods. Increases in the duration of dry periods will allow riparian shrubs and trees to establish themselves more strongly and limit the ability of wet-period floods to restore channel dimensions and dynamic behavior. We have shown that even large flood events have failed to reverse channel narrowing since the recurrence interval for the channel-forming discharge of 4,000 ft^3/s increased from 2.2 years for the period before 1971 to about 3 years for the period after 1971.

Some analyses suggest that, at full development, water diversion projects will divert all but about 200,000 acre-ft of Duchesne River water (US Fish and Wildlife Service 1998). This would constitute an approximately 50 percent decrease in stream flow from current and historical values, and would make maintenance of the Duchesne River channel as it exists today impossible. Impacts would likely include the loss of substrate productivity due to the accumulation of fine sediments, a decrease in channel dimensions due to sedimentation and vegetative encroachment, and a decrease in topographic complexity due to greatly decreased rates of channel activity. These changes would likely result in loss of habitat for endangered native fishes.

Stream flow on the Duchesne River is already depleted relative to historical figures by operation of the Bonneville Unit of the Central Utah Project, and geomorphic responses to those depletions are already underway. Mean annual runoff has decreased from the historical mean by 5.5 percent since Bonneville Unit diversions began in water year 1972, and the recurrence interval of daily mean discharges of 4,000 ft^3/s has increased from 2.4 years for the full period of record to 3 years, as calculated from post-1971 data. Thus far, the impact of these withdrawals has been offset to some extent by the fact that wet years and large floods have occurred with similar frequencies and magnitudes both before and after 1972. The average annual volume of stream flow in excess of the channel-forming threshold for the post-project years is actually greater than the average for the full record.

To illustrate the effect of additional withdrawals, a scenario involving a additional flow depletions of 10 percent from all parts of the annual hydrograph is assumed. Such a decrease in stream flow would reduce the frequencies and magnitudes of channel-forming flows such that only 5,900 ft^3/s-days of stream flow in excess of 4,000 ft^3/s would occur on an average annual basis. This volume of channel maintenance flows may be insufficient to maintain the dynamic channel processes necessary for ecosystem maintenance. A 10 percent reduction in stream flow below historical levels would also cause a corresponding decrease in the percentage of the annual sediment load that could be moved through the system. Restoring the fine sediment transport capacity to the proposed levels would require dropping the threshold discharges at which spring runoff is available for fine sediment transport. Because of the non-linear relationship between discharge and suspended sediment concentration, threshold discharge levels would have to decrease by much more than 10 percent.

However, stream flow data comparing the flow regimes before and after construction of the Bonneville Unit show that depletions are not spread evenly over all four seasons or among all water years. The frequencies of some flow magnitudes are affected to a greater degree than others, and the amount of increase in the recurrence interval for a flood of a given magnitudes is highly sensitive to the precise seasonal and annual patterns of diversion. Operation of the Bonneville Unit may have caused an observed increase after 1972 in the recurrence interval for flows that inundate bar and floodplain surfaces and mobilize gravel. Careful management of diversion patterns may provide the means to maintain channel-forming discharges at target levels, while also ensuring adequate fine sediment transport, even with modest additional depletions of the system.

Table 23: Proposed flow regime for channel and habitat maintenance.

Hydrolgic Category	Percent of years	Flow and Duration Exceeded	Description of Anticipated Effects
Extremely Wet (> 645,000 acre-ft)	12%	7,000 at least 1day 6,800 at least 2 days 6,600 at least 3 days 6,400 at least 5 days 6,200 at least 8 days 6,000 at least 11 days 5,800 at least 14 days 5,600 at least 16 days 5,400 at least 18 days 5,200 at least 19 days 5,000 at least 20 days 4,800 at least 21 days 4,600 at least 22 days 4,400 at least 23 days 4,200 at least 24 days	These higher flows will promote channel migration, maintain off-channel topographic complexity, maintain channel dimensions, and rejuvenate riparian vegetation. Intense scouring of the channel bed will remove fine sediment from the gravel framework, and fine sediment will be flushed from the full range of low velocity habitats along the lower Duchesne River. These processes are necessary to maintain the current level of channel integrity and habitat diversity now present in the Duchesne River.
Wet (439,000 to 645,000 acre-ft)	28%	5,600 at least 1day 5,100 at least 2 days 4,700 at least 4 days 4,400 at least 7 days 4,000 at least 10 days 3,000 at least 17 days	Widespread bed entrainment will maintain riffle and pool topography, maintain channel dimensions, and contribute to channel migration. Regular flow events exceeding the bankfull stage are necessary to prevent the establishment of riparian vegetation within the bankfull channel. In addition, fine sediment will be flushed from gravel substrates and from many low velocity habitats adjacent to the main channel.
Normal (224,000 to 439,000 acre-ft)	30%	2,500 at least 7 days	These flows will transport fine sediment delivered to the lower Duchesne River in order to balance the sediment budget and prevent fine sediment accumulation in low velocity habitats.
Dry (< 224,000 acre-ft)	30%	No peak flow requirements	

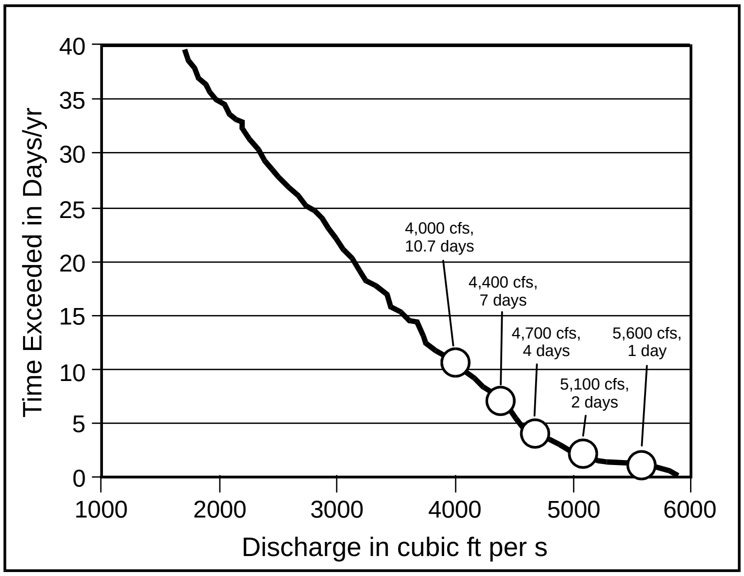


Figure 41: A portion of the flow duration curve for eight years with total annual channel-forming discharges between 3,000 ft^3/s-days and 10,000 ft^3/s-days. Magnitude/duration combinations defining the proposed channel-forming hydrograph for wet years are indicated.

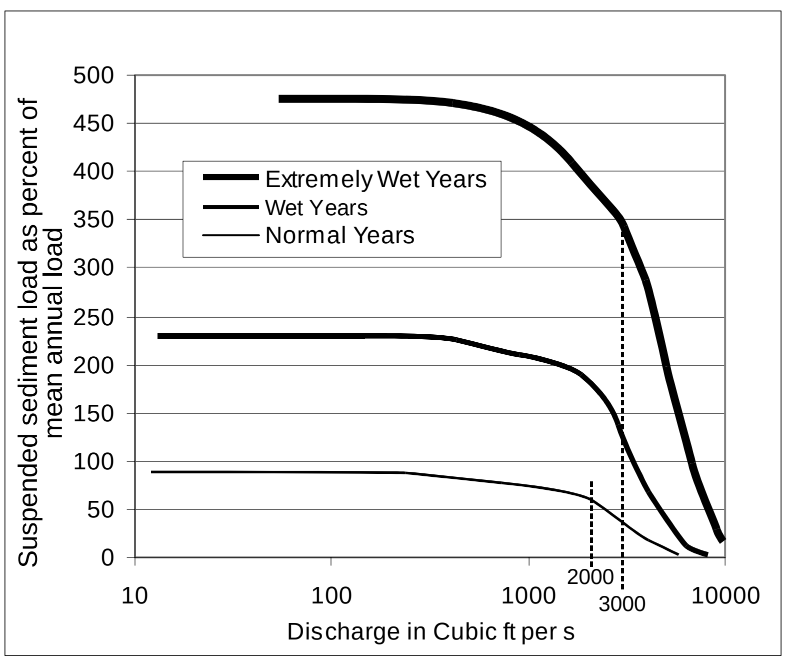


Figure 42: Cumulative sediment transport curves for extremely wet years (years with total annual flow exceeding the 90th percentile), wet years (years with total annual flow between the 60th and 90th percentiles), and normal years (years with total annual flow between the 30th and 60th percentiles).

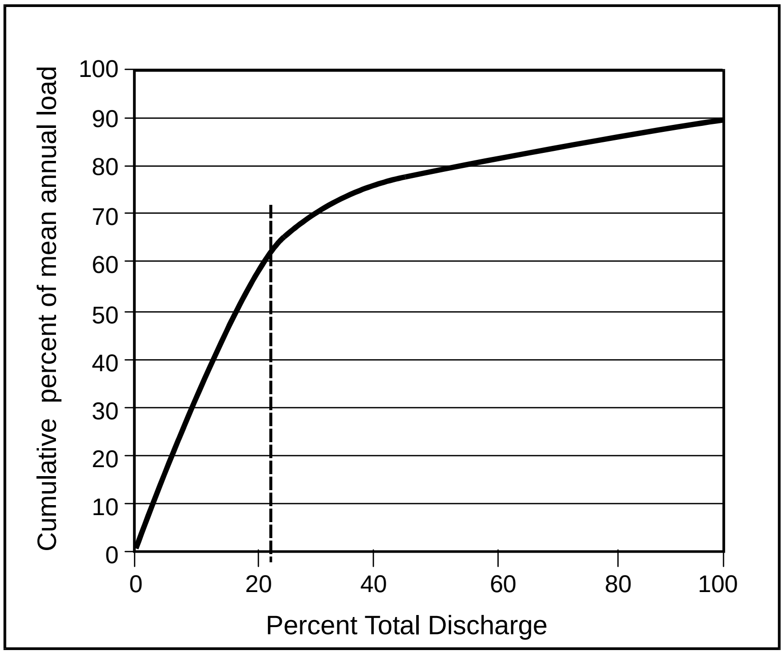


Figure 43: Water-sediment efficiency curve for normal years.

CONCLUSIONS

1. The lower Duchesne River consists of four distinct zones with differing morphologies and histories. Channel morphology and response to flow varies in time and between zones.
2. Channel-forming discharge on the lower Duchesne River is about 4,000 ft^3/s. Gravel mobilization and inundation of high bar surfaces occur at this discharge.
3. Little channel activity occurs during periods when the volume of stream flow in excess of the channel-forming discharge of 4,000 ft^3/s is less than 7,000 ft^3/s-days per year. Physical habitat is created and maintained during decades when the volume of stream flow in excess of 4,000 ft^3/s is greater than 7,000 ft^3/s-days per year. Calculation of flow volumes in excess of the channel-forming discharge is described on page 85.
4. The recurrence of daily mean discharges of 4,000 ft^3/s is about 2.2 years for the period from 1943 through 1971. The recurrence of this discharge has increased to 3 years since completion of the Bonneville Unit of the CUP in water year 1972.
5. The increase the recurrence period for daily mean discharges of 4,000 ft^3/s or greater since 1971 has contributed to a consistent trend of channel narrowing since 1969.
6. Fine sediment accumulation related to a 50-percent reduction in streamflow after the 1920s and an increase in the local sediment supply resulted in significant channel narrowing, the loss of side channel habitat, and large-scale avulsions on the lower Duchesne River.
7. The accumulation of fine sediment in the lower Duchesne River can be prevented by a flow regime that includes stream flow volumes in excess of 4,000 ft^3/s averaging at least 7,000 ft^3/s-days per year, plus an additional 7 days per year with discharges greater than 3,000 ft^3/s in wet years and 7 days per year with discharges greater than 2,500 ft^3/s in normal years.
8. Existing measurements of suspended sediment concentrations in the lower Duchesne River are inadequate for making well-constrained estimates of suspended sediment loads during high discharge periods. An extended sampling program to monitor suspended sediment concentrations in the lower Duchesne River during peak flow events should be undertaken.

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