Sediment supply versus local hydraulic controls on sediment transport and storage in a river with large sediment loads

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Abstract The Rio Grande in the Big Bend region of Texas, USA, and Chihuahua and Coahuila, Mexico, undergoes rapid geomorphic changes as a result of its large sediment supply and variable hydrology; thus, it is a useful natural laboratory to investigate the relative importance of flow strength and sediment supply in controlling alluvial channel change. We analyzed a suite of sediment transport and geomorphic data to determine the cumulative influence of different flood types on changing channel form. In this study, physically based analyses suggest that channel change in the Rio Grande is controlled by both changes in flow strength and sediment supply over different spatial and temporal scales. Channel narrowing is primarily caused by substantial deposition of sediment supplied to the Rio Grande during tributary-sourced flash floods. Tributary floods have large suspended-sediment concentrations, occur for short durations, and attenuate rapidly downstream in the Rio Grande, depositing much of their sediment in downstream reaches. Long-duration floods on the mainstem have the capacity to enlarge the Rio Grande, and these floods, released from upstream dams, can either erode or deposit sediment in the Rio Grande depending upon the antecedent in-channel sediment supply and the magnitude and duration of the flood. Geomorphic and sediment transport analyses show that the locations and rates of sand erosion and deposition during long-duration floods are most strongly controlled by spatial changes in flow strength, largely through changes in channel slope. However, spatial differences in the in-channel sediment supply regulate sediment evacuation or accumulation over time in long reaches (greater than a kilometer).

1. Introduction

Understanding how channels change in response to alterations in flow regime and sediment supply is fundamental to the application of fluvial geomorphology to many important societal questions, especially those concerning the impacts of human activities and climate change [Gilbert, 1877, 1917; Gilbert and Murphy, 1914; Exner, 1920, 1925; Mackin, 1948; Schumm, 1969; Petts, 1979, 1985; Williams and Wolman, 1984; Brandt, 2000; Singer and Dunne, 2001; Eaton and LaPointe, 2001; Topping et al., 2000a; Grant et al., 2003; Magilligan and Nislow, 2005; Schmidt and Wilcock, 2008; Singer, 2010; Belmont et al., 2011]. Predicting the magnitude and style of channel change in response to changes in stream flow and sediment supply is difficult, because geomorphic changes are dependent upon the spatial and temporal variability in the sediment supply, channel geometry, hydraulics, and the physical and/or biological controls on channel and floodplain stability and roughness. Furthermore, changes can be amplified or dampened by positive or negative feedbacks [Melton, 1958; King, 1970; Singer, 2010]. It is this complex nature of geomorphic evolution, framed within the context of climate change, water development, river management, and other anthropogenic alterations to river corridors, that necessitates the continued focus on this fundamental research problem.

Many studies have used sediment budgets to analyze changes in the sediment mass balance for rivers, and the associated mass balances are often used to understand and predict geomorphic change [Singer and Dunne, 2001; Erwin et al., 2012; Grams et al., 2013]. For example, in reaches that evacuate sediment because of a perturbation of the mass balance into sediment deficit, channel incision [Williams and Wolman, 1984; Lach and Wyzga, 2002; Grams et al., 2007], channel widening [Harvey and Watson, 1986], erosion of channel bars [Kondolf et al., 2002; Venditti et al., 2012], decreases in lateral migration rates, decreases in channel bed slope [Grams et al., 2007], and channel bed coarsening [Draut et al., 2011] may all occur. Conversely, in reaches that accumulate sediment due to perturbation of the mass balance into sediment surplus; bed
aggradation [Zahar et al., 2008; East et al., 2015; Warrick et al., 2015]; channel narrowing [Everitt, 1993]; increases in channel sinuosity, bed slope, and avulsion frequency [Leopold and Wolman, 1957; Bryant et al., 1995; Ashworth et al., 2004; East et al., 2015]; floodplain accretion [East et al., 2015; Dean et al., 2011]; and bed fining are all potential outcomes.

There are differences in the style of geomorphic change depending on whether sediment transport is dominated by bed load or suspended load, and depending on the size of sediment on the bed. In bed load-dominated rivers, bed coarsening may limit the degree of incision [Williams and Wolman, 1984; Brandt, 2000; Grams et al., 2007; Schmidt and Wilcock, 2008], whereas bed incision can proceed unabated if the bed material is fine [Brandt, 2000]. In conditions of sediment surplus, aggradation of the bed in bed load-dominated rivers may cause widening and the development of multiple threads if bank cohesion and stability are low [Brandt, 2000]. Alternatively, narrowing occurs if there is a large amount of fine sediment transported in suspension [Alfred and Schmidt, 1999; Grams and Schmidt, 2002; Dean et al., 2011]. In rivers dominated by suspended-load transport, sediment surplus can result in concurrent bed and floodplain aggradation, as well as narrowing [Everitt, 1993; Friedman et al., 2015]. Furthermore, silt and clay deposited from suspension may increase bank stability by increasing cohesion. Riparian vegetation can also increase bank stability [Pollen-Bankhead et al., 2009; Tal and Paola, 2007], trap sediment [Wilcox and Shafroth, 2013], promote channel narrowing [Grañ and Paola, 2001; Dean and Schmidt, 2011], and decrease the width-to-depth ratio of the channel [Manners et al., 2014].

Channel changes caused by sediment deficit or surplus conditions within a reach can be reduced or exacerbated by upstream changes in sediment storage within the channel network [Schumm, 1979; Graf, 1987]. For example, measurements of channel adjustment downstream from Elephant Butte Dam on the Rio Grande and Hoover Dam on the Colorado River demonstrated that large areas of upstream erosion were a significant source for downstream aggradation [Stevens, 1938; Borland and Miller, 1960]. Conversely, sediment accumulation within a reach may result in reduced sediment transfer to downstream reaches [Graf, 1987]. Thus, conditions of deficit or surplus may vary longitudinally over space and time, resulting in complex geomorphic responses that may propagate throughout a drainage network at different rates and magnitudes [Schumm, 1973; Schumm and Parker, 1973; Graf, 1987]. In field-based studies, it is often difficult to fully evaluate the spatial variations in channel geometry, flow strength, and/or sediment supply that cause these complex patterns of geomorphic change. This, in turn, makes it difficult to properly apply numerical, theoretical, and flume-based studies to the field scale. Comprehensive measurements of sediment transport at individual cross sections and longitudinally can help inform processes of geomorphic change, because spatial and temporal patterns of sediment deficit and/or supply directly influence sediment transport processes.

We use suspended-sediment transport measurements collected at high temporal resolution to construct a suspended-sediment budget and analyze temporal variability in suspended-sediment transport to track changes in sediment supply over a long reach of the Rio Grande in the Big Bend region of Texas, USA, and Chihuahua and Coahuila, Mexico. The sediment transport regime of the Rio Grande is dominated by suspended load and is in a state of sediment surplus, caused by upstream stream flow reductions [Schmidt et al., 2003; Dean and Schmidt, 2011]. Current management efforts, including the potential reoperation of upstream dams, aim to limit the rate and magnitude of channel narrowing associated with surplus conditions. Here we inform these efforts by developing a suspended-sediment budget and analyzing time-varying patterns in suspended-sediment concentration and grain size to link the geomorphic response of the Rio Grande to temporal changes in the sediment supply. We combine these data with analysis of the spatial variation in channel geometry to analyze longitudinal patterns of suspended-sediment transport and geomorphic response. This work demonstrates the utility of a comprehensive program of sediment transport and channel measurements in determining the relative influences that flow and sediment supply exert on the rate and magnitude of channel change in rivers perturbed into sediment surplus.

2. Hydrology and Geomorphology

The Rio Grande in the Big Bend region extends from the confluence with the Rio Conchos 490 km downstream to Amistad Reservoir (Figure 1) and is the international boundary between the United States and Mexico. In Mexico, the river is called the Rio Bravo. The Rio Grande in this region is predominantly single threaded and flows through wide alluvial valleys in structural basins and narrow canyons cut through
intervening ranges. Channel slope varies between ~0.0005 in the alluvial valleys and ~0.003 in the canyons. The channel bed is a heterogeneous mix of gravel, sand, silt, and clay [Dean et al., 2011]. Gravel is predominant at the mouths and downstream from most ephemeral tributaries [Dean and Schmidt, 2013]. A large amount of silt and clay and lesser amounts of sand are also supplied from these tributaries. The alluvial banks of the Rio Grande are primarily silt and clay with some sand lenses and sandy levees [Dean et al., 2011]. The banks and floodplain are densely vegetated with nonnative salt cedar (Tamarix spp.) and giant cane (Arundo donax), and with native willow (Salix exigua), seep willow (Baccharis spp.), and common reed (Phragmites australis) [Moring, 2002], among others.

Prior to the construction of Elephant Butte Dam on the Rio Grande in New Mexico in 1915 (i.e., “predam”), late spring high flows in the Rio Grande were caused by snowmelt in the southern Rocky Mountains. Large inflows from the Rio Conchos during the North American monsoon typically began at the time the snowmelt flood receded, and high flows in the Rio Grande extended into late summer and early fall [Schmidt et al., 2003; Dean and Schmidt, 2011]. High flows during the monsoon were caused by local thunderstorms in the Chihuahuan Desert and dissipating tropical storms in the Rio Conchos’ headwaters in the Sierra Madre Occidental [Dean and Schmidt, 2011, 2013]. The combination of high flows from snowmelt and the monsoon resulted in the predam Rio Grande having one of the longest annual floods in North America, lasting up to 6 months [Schmidt et al., 2003; Dean and Schmidt, 2011]. Meade et al. [1990] estimated that the Rio Grande had the sixth largest sediment discharge of any river in the U.S. and Canada, prior to large-scale human development. Dams and irrigation diversions in the upper Rio Grande completely eliminated the spring snowmelt flood by the 1920s [Dean and Schmidt, 2011]. Although many large dams and irrigation diversions were also constructed in the Rio Conchos watershed, the magnitude of flow reduction was less in the Rio Conchos. In 1967, Luis L. Leon Dam was completed on the lower Rio Conchos to allow full utilization of the Rio Conchos within Chihuahua. The present objective of reservoir management on the Rio Conchos is to release only the amount required by downstream Mexican irrigators, by international treaty, or to create storage capacity for flood control. Typical dam releases are very low, or zero, in drought years, and moderate-magnitude, long-duration dam release floods only occur in wet years for water resource management purposes. Any potential benefits of dam releases to the downstream ecosystem are unintentional.

The present hydrology of the Rio Grande in the Big Bend is extremely variable. Between the Rio Conchos confluence and Rio Grande Village (RGV) (Figure 1) base flows in winter and early spring may be less than 1 m$^3$/s. Groundwater inflow from the Edwards-Trinity aquifer increases base flow in the eastern part of Big Bend National Park near RGV. Short-duration, moderate-magnitude floods sometimes occur as early as May when the thunderstorm season begins. Flash floods from ephemeral tributaries enter the lower Rio Conchos, downstream from Luis L. Leon Dam, or the Rio Grande. Hereafter, we refer to these flash floods as “tributary-sourced” floods. These floods have peak discharges typically between 5 and 500 m$^3$/s and can inundate parts of the Rio Grande floodplain. The duration of tributary-sourced floods is rarely more than a...
day and may be only a few hours. In contrast, moderate-magnitude, long-duration, mainstream high flows
released from Luis L. Leon Dam are typically between 40 and 200 m$^3$/s and typically last more than 5 days.
These releases occurred every year between 2010 and 2014, which is an unusually high frequency for the
postdams era. To simplify, we refer to long-duration moderate-magnitude releases from Luis L. Dam as
“dam releases.”

Runoff from tropical storms in the Rio Conchos watershed can at times exceed available reservoir storage and
necessitate large releases from Luis L. Leon Dam with peak discharges $>$1000 m$^3$/s and durations of weeks to
months [Dean and Schmidt, 2011, 2013]. Flooding of this magnitude were common in the early 1900s but are
now infrequent. Based on long-term flow records, the most recent of these floods occurred in fall 2008
and had a peak discharge of 1490 m$^3$/s; a recurrence of 13–15 years for the entire gaging record since
1901, and 22 years since the completion of Luis L. Leon Dam [Dean and Schmidt, 2013]. Flooding of this
magnitude are referred to as “channel-reset” floods, because they substantially widen the channel and can
completely reconfigure the geomorphic organization of the river [Dean and Schmidt, 2013].

The geomorphic nature of the modern Rio Grande in the Big Bend is characterized by progressive channel
infilling over decadal timescales, and occasional channel widening during channel-reset floods [Dean and
Schmidt, 2011, 2013; Dean et al., 2011]. Channel infilling occurs through the accumulation of sand, silt, and
clay (collectively referred to as fine sediment) that causes bed aggradation, floodplain accretion, and the
development of inset floodplains. This sediment is dominantly carried in suspension. Channel-reset floods
rewidth the channel once every decade or two, but widening occurs to a lesser degree than the magnitude
of the intervening narrowing. Thus, the Rio Grande channel is being progressively “ratcheted” (in the sense of
Tal et al. [2004]) from a historically multithreaded planform [Dean and Schmidt, 2011] into a narrow single-
threaded condition with a simplified aquatic habitat whereby side channels, backwaters, embayments, and
other low-velocity parts of the channel are lost as the channel narrows. Between 1991 and 2008, the river
narrowed between 36 and 52% through the oblique and vertical accretion of as much as 3 m of fine sediment
[Dean and Schmidt, 2011; Dean et al., 2011]. The 2008 channel-reset flood rewidthed the channel by an aver-
age of 26 to 52%, and a new phase of narrowing is now occurring [Dean and Schmidt, 2013].

This study is focused on a 110 km reach between the U.S. Geological Survey (USGS) stream gages near
Castolon, TX (USGS gage number 08374550) and Rio Grande Village (RGV), TX (USGS gage number
08375300). This reach is predominantly alluvial and resides entirely within Big Bend National Park and is
hereafter referred to as “the study area” (Figure 1).

3. Physics-Based Context for Predicting Patterns in Suspended-Sediment Concentration and Geomorphic Change

Erosion and deposition are dependent upon the amount and size of sediment supplied to a reach, and the
hydraulics that transport that supply through the reach [e.g., Topping et al., 2000a, 2000b]. Sediment is sup-
plied from two sources: upstream and locally. The upstream sediment supply comes from the watershed,
whether from tributary-sourced floods or from mainstream floods released from Luis L. Leon Dam whose runoff
comes from the distant headwaters. In either case, the upstream sediment supply is the suspended-sediment
flux measured at a specific location. This flux integrates all of the upstream flow and sediment transport
conditions and is fully characterized by the concentration, grain size, and load of the suspended sediment.
In contrast, the local sediment supply is the fine sediment available for entrainment from the bed, banks,
and floodplain in the reach immediately upstream from each gage (up to several kilometers). The length
of this reach is determined by the scale over which the suspended sediment equilibrates with the bed-
sediment conditions over a range in flow conditions [Topping et al., 2007a]. This supply not only contributes
to the mass balance but also regulates the amount and grain size of sediment entrained and transported by
flood flows within the study area. Hereafter, we refer to the local sediment supply as the “in-channel” supply,
because most of the local supply is accessed by flows contained in the channel.

Below, we combine basic theoretical relationships with comprehensive field measurements of suspended-
sediment transport and geomorphic change to analyze spatial variability of suspended-sediment transport,
track temporal changes in sediment supply that come from upstream and local sources, and link these pro-
cesses to changes in channel morphology. We distinguish two upstream sediment sources—fine sediment
delivered to the Rio Grande by tributary flash floods and fine sediment delivered to the study area by
long-duration dam releases. We describe how stream flow and suspended-sediment transport during floods from these two upstream sources vary over time and space and how erosion and deposition are also affected by changes in the characteristics of the local sediment supply within the channel.

3.1. Predicting General Patterns of Erosion and Deposition

Large-scale predictions of channel response in relation to the divergence in the sediment flux can be made using the conservation of mass as described by Exner [1920, 1925]. Here we use a derivation and simplification similar to that proposed by Grams et al. [2013]. The general form of the Exner equation describes the relationship between geomorphic change (i.e., change in sediment storage) and sediment flux, expressed as follows:

$$\frac{\partial \eta}{\partial t} = -\frac{1}{(1 - \lambda_p)} \left( \frac{\partial V_s}{\partial t} + \nabla \cdot Q_s \right)$$

where \( \eta \) is the elevation of the bed, \( t \) is time, \( \lambda_p \) is bed-sediment porosity, \( V_s \) is the volume of sediment in suspension, and \( Q_s \) is sediment flux. Scaling analyses by Rubin and Hunter [1982] and Paola and Voller [2005] demonstrated that \( \partial V_s/\partial t \) has little effect on the evolution of the bed, because there is rarely enough suspended sediment within the water column to act as a sediment source for long periods of time; thus, this term can be neglected. If porosity of the bed is assumed constant, then aggradation or degradation of the channel bed is entirely controlled by the spatial divergence of the sediment flux [Grams et al., 2013]:

$$\frac{\partial \eta}{\partial t} \propto -\nabla \cdot Q_s$$

Since suspended-sediment transport dominates on the Rio Grande, sediment flux is defined here as the product of suspended-sediment concentration \( (C_s) \) and water discharge \( (Q) \):

$$Q_s = C_s Q$$

where \( Q \) is defined as the product of mean velocity \( (U) \), mean depth \( (h) \), and channel width \( (b) \):

$$Q = Uhb$$

Scaling analyses by D. J. Topping (Physics of flow, sediment transport, hydraulic geometry, and channel geomorphic adjustment during flash floods in an ephemeral river, the Paria River, Utah and Arizona: University of Washington, unpublished PhD dissertation, 406 pp., 1997, http://www.gcmrc.gov/library/reports/physical/Fine_Sed/Topping1997V1.pdf and http://www.gcmrc.gov/library/reports/physical/Fine_Sed/Topping1997V2.pdf) indicate that sediment transport theory developed for steady, uniform flow conditions is appropriate over the reach scale for all conditions except flood bores. Flood bores comprise a very small percentage of time, even in a flash-flood-dominated system like the Rio Grande and its tributaries, and therefore, we use steady, uniform flow theory to develop relations between suspended-sediment transport, flow, and the local, in-channel sediment supply.

Using the modified formulation of Rubin and Topping [2001, 2008] in Topping et al. [2010], suspended-sediment concentration \( (C_s) \) is dependent upon the reach-scale hydraulics, and the reach-scale amount and grain size distribution of sediment available for transport on the channel bed. Thus,

$$C_s \propto u^3 D_b^{-2.5} A_s$$

where \( u \) is the shear velocity (a measure of the flow strength), \( D_b \) is the reach-averaged median grain size of susceptible bed sediment, and \( A_s \) is the reach-averaged fraction of the bed covered by the sediment with median grain size \( D_b \). Taken together, \( D_b \) and \( A_s \) are measures of the in-channel sediment supply; as the in-channel sediment supply increases, \( D_b \) typically decreases and \( A_s \) typically increases to a maximum of 1.0. Unlike the nonlinear influence of \( D_b \) on \( C_s \), \( A_s \) exerts an approximately linear control on \( C_s \) when a uniform areal distribution of susceptible sediment exists on the bed under steady, uniform flow conditions [Grams and Wilcock, 2007].

The amount of sediment available for suspended-sediment transport in a reach is not directly determined by the total volume of susceptible sediment stored in the bed but, rather, is determined by \( D_b \) and \( A_s \) on the bed surface [Topping et al., 2007a, 2010]. At the reach scale, \( A_s \) and the volume of susceptible bed sediment should typically be positively correlated in alluvial rivers. Similarly, although \( D_b \) on the bed surface and the reach-averaged median grain size of the entire volume of the local, in-channel sediment supply may be
similar, these two attributes may not be well correlated if the bed surface is armored [Topping et al., 2010]. For simplicity, and because the Rio Grande in the study area is mostly alluvial, we assume that $A_s$ is positively correlated with the total volume of the local, in-channel sediment supply. In addition, we assume that the bed of the river is “well mixed” during large floods and that armoring is unimportant because there is significant scour and fill during Rio Grande floods, and multiple floods occurred during the period of study; thus, we assumed that $D_b$ is representative of the reach-averaged median grain size of the local supply.

Similar to (5), the median grain size of suspended sand ($D_s$) is dependent on local hydraulics and on $D_b$, but not on $A_s$ [Topping et al., 2010]. Thus,

$$D_s = u^{0.25}D_b^{0.75}$$

(6)

From (5) and (6) it is evident that changes in $A_s$ can be deduced for conditions of constant $u_*$ and $D_s$, and changing $C_s$, a logic that we employ to detect changing $A_s$. For constant $u_*$, constant $D_s$ in (6) implies constant $D_b$.

Equations (3)–(5) can be combined to reformulate a sediment-concentration-based approximation of the Exner equation, and because $u_* \propto U$ (von Karman, 1930), equation (2) then can be written as follows:

$$\frac{\Delta \eta}{\Delta t} = \frac{\Delta(u_*^{4.5}D_b^{-2.5}A_s h_b)}{\Delta X}$$

(7)

where

$$u_* = \sqrt{ghS}$$

(8)

and $X$ is the distance in the downstream direction, $g$ is the gravitational acceleration, and $S$ is the slope of the water surface for steady, uniform flow. Equation (7) illustrates that reach-scale erosion or deposition is controlled by spatial changes in the variables that serve as measures of flow strength ($u_*$), in-channel sediment supply ($D_b$ and $A_s$), and channel geometry ($h$ and $b$). Substitution of equation (8) into (7) further illustrates that erosion and deposition patterns are dependent on changes in flow strength that are driven by spatial changes in $h$ and $S$. Although spatial changes in flow strength typically exert stronger influence over erosional and depositional processes than do spatial changes in the local sediment supply or channel geometry, changes in the local sediment supply may dominate under certain conditions [Topping et al., 2000b; Grams et al., 2013].

Accurate prediction of geomorphic response requires that none of the variables within the parentheses on the right side of equation (7) be considered in isolation; however, evaluation of spatial changes in these individual variables can be used as a conceptual framework for understanding how spatial changes in flow strength, in-channel sediment supply, and channel geometry may affect channel changes of the Rio Grande. To illustrate how this framework can be used to make predictions, we consider three examples: (1) an upstream, tributary-sourced flood entering the Rio Grande when discharge in the Rio Grande is low; (2) spatial changes in flow strength under constant discharge and constant local, in-channel sediment supply; and (3) spatial changes in the in-channel sediment supply under constant discharge and flow strength.

First, during tributary-sourced floods, the discharge of the flood typically greatly exceeds the antecedent discharge of the Rio Grande such that the incremental increase in $Q$ of the flash flood upon entering the Rio Grande can be ignored. The slope of the Rio Grande is generally much less than that of its tributaries, and the width of the Rio Grande is generally equal to or wider than the width of its tributaries. Therefore, when a tributary-sourced flood enters the Rio Grande, there is a large spatial decrease in $S$, a large spatial increase in $b$, and therefore, a large spatial decrease in $h$. By equation (8), these spatial reductions in $S$ and $h$ result in a large spatial decrease in $u_*$. This combination of factors in equation (7)—large spatial decreases in $u_*$ and $h$—and an increase in $b$, lead to conditions favorable to local aggradation (+$\eta$) in the reach downstream from tributary confluences. Because the spatial decrease in $u_*$ is large in this example and is raised to the 4.5 power in equation (7), its effects dominate over the effects of the other, possibly offsetting, variables.

Second, during long-duration dam releases without tributary-sourced floods, $Q$ is typically constant for many days. By equations (4), (7), and (8), and because $u_* \propto U$, spatial changes in $u_*$ during long-duration dam releases will result from spatial changes in a combination of $S$, $h$, and $b$. For cases where the in-channel sediment supply ($D_b$ and $A_s$) is approximately constant longitudinally, we can expect erosion to occur...
wherever $h$ and/or $S$ increases and/or $b$ decreases, such as might occur at channel constrictions. By the same logic, we can expect deposition to occur in areas characterized by reductions in $S$ and/or $h$, and/or increases in $b$, such as at the transition from steep reaches to lower gradient meandering reaches and in areas upstream from hydraulic controls. In these cases, spatial changes in the flow strength driven only by longitudinal changes in $h$ and $S$ are responsible for determining the locations and the rates of deposition or erosion. Therefore, basic analyses of the general geomorphic setting should provide some insight about the likely locations of fine-sediment accumulation or evacuation during steady, long-duration dam releases.

Third, long-duration dam releases that follow an active thunderstorm season may result in a geomorphic response that is slightly modified from that in the previous example. In this example, the antecedent, in-channel sediment supply may not be spatially constant but may be fine (decrease in $D_b$) and increase in bed coverage (increase in $A_i$) in the downstream direction because of fine-sediment inputs by tributaries. In this case, even under conditions of constant flow strength or channel geometry, there may be a longitudinal increase in the suspended-sediment flux under constant $Q$. Thus, if the flow is sufficient to entrain material from the bed, erosion may occur because either a spatial decrease in $D_b$ or a spatial increase in $A_i$ may lead to a downstream increase in the sediment flux, even under constant $u_t$. Furthermore, erosion may occur even in cases where the flow strength decreases [Topping et al., 2000b], because even though the absolute value of the exponent on $u_t$ is greater than that on $D_b$, there are cases where longitudinal changes in the local, in-channel sediment supply can offset longitudinal changes in flow strength in controlling geomorphic response.

In light of these examples, we hypothesize that (a) tributary-sourced flash floods will primarily result in deposition in the Rio Grande; (b) longer-duration dam releases may be either depositional or erosional, depending on spatial changes in flow strength and channel geometry, and the spatial distribution of fine sediment within the channel at the time of the dam releases; and (c) reaches with a greater fine-grained sediment supply may have larger transport rates for similar flow strength.

### 3.2. Suspended- and Bed-Sediment Dynamics During Individual Floods

Suspended-sediment dynamics within individual floods may be more complex than the large-scale patterns described above. For example, during a tributary-sourced flash flood, the sediment supplied to the Rio Grande is much finer than the sediment typically composing the bed, and the newly supplied sediment travels downstream as an elongating sediment wave, with a component in suspension and on the bed. When the fine front of a sediment wave reaches a given location, the concentration of the finer grain sizes in suspension will be higher than that which can be supported by the grain size distribution of the bed, and there will be a mass transfer of the finest sizes from the bed into suspension, resulting in bed coarsening (i.e., winnowing). As the sediment wave passes, the concentration of the finer grain sizes of suspended sediment will decrease to be lower than that which can be supported by the grain size distribution of the bed, and there will be a mass transfer of the finest sizes from suspension to the床 and causes the bed to fine, wherein $D_b$ decreases and where $A_i$ may increase. As the sediment wave passes, the concentration of the finer grain sizes of suspended sediment will decrease to be lower than that which can be supported by the grain size distribution of the bed, and there will be a mass transfer of the finest sizes from the bed into suspension, resulting in bed coarsening (i.e., winnowing) (i.e., $D_b$ increases and $A_i$ may decrease) [Rubin et al., 1998].

As above, there are other physical processes (e.g., changes in the upstream sediment supply and bed form development) that can independently control the transport of some sizes of bed sediment [Colby, 1963; Guy, 1970; Dinehart, 1982; Topping et al., 2000a, 2000b] and often result in the poor correlation between the concentration of suspended sediment and the discharge of water [Gray and Simões, 2008]. In cases where the in-channel sediment supply becomes progressively depleted during a flood, the sediment in suspension and on the bed coarsens as suspended-sediment concentrations decrease [Rubin et al., 1998; Topping et al., 2000b]. This results in clockwise hysteresis in discharge-concentration space coupled to counterclockwise hysteresis in discharge-grain size space. In cases where the upstream sediment supply becomes enriched during a flood, such as during tributary-sourced flash floods, the peak in suspended-sediment concentration lags the kinematic discharge wave, and the magnitude of this lag increases as the flood wave travels downstream [Einstein, 1943; Heidel, 1956]. This lag develops because the celerity of a flood wave exceeds the mean velocity of the water [Lighthill and Whitham, 1955], and because finer grain sizes of suspended sediment travel faster than coarser grain sizes because they are carried higher in the water column, where flow velocities are generally greater [Rouse, 1937; McLean, 1992].
The parameter that characterizes the physics of this process is the Rouse number, $P$, which is the ratio between the settling velocity of a particle ($\omega_s$) and the upward forces on the grain ($\kappa u_*$) that keep it suspended:

$$P = \frac{\omega_s}{\kappa u_*}$$

where $\kappa$ is the von Kármán constant, typically taken to be 0.41 [Long et al., 1993]. Therefore, since the celerity of the floodwave exceeds the mean velocity of the water, and the finer grain sizes travel downstream faster than coarser grain sizes, a counterclockwise hysteresis loop in discharge-concentration space, coupled to clockwise hysteresis loop in discharge-grain size space may occur during short-duration floods.

Furthermore, during a short-duration flood, the flood discharge may recede before some proportion of sediment passes a downstream reach, resulting in upstream deposition of the coarser fractions of sediment. Given these patterns, we hypothesize that (a) considerable fractionation of the suspended sediment will occur, resulting in different spatial depositional patterns depending on the grain sizes of sediment in suspension. We also hypothesize that (b) there will be proportionally less deposition of silt and clay compared to sand during these short-duration floods.

4. Methods

4.1. Continuous Acoustic Suspended-Sediment Monitoring

Because stream flow and suspended-sediment concentration on the Rio Grande are often decoupled, stable relations between the discharge of water and suspended-sediment concentration cannot be assumed, and estimates of sediment transport using traditional sediment-rating relations will have unacceptably large biases and error [e.g., Glysson et al., 2001; Gray and Simões, 2008]. Calculation of accurate sediment loads in rivers where the transport of suspended sediment is decoupled from discharge requires measurements of suspended-sediment concentration made at high temporal resolution. We used a multifrequency acoustic method that utilizes 15 min measurements of acoustic attenuation and backscatter to monitor suspended-sediment transport [Topping et al., 2004, 2007b, 2015]. The details of this acoustic method are based on the method of Topping et al. [2015] and the theory of Thorne and Meral [2008] and Moore et al. [2013]. This method involves three steps: (a) acoustic attenuation is used to calculate suspended silt-and-clay concentration in the cross section; (b) acoustic backscatter is used to calculate the apparent suspended-sand concentration at each frequency, corrected for the backscatter produced by silt and clay; and, (c) the apparent suspended-sand concentration calculated at each frequency is used in combination with the form function of Thorne and Meral [2008] to calculate a two-frequency measure of the suspended-sand concentration and median grain size in the river cross section.

In November 2010, we installed two acoustic suspended-sediment gaging stations in the vicinity of existing Castolon and RGV stream gages (Figure 1). The sediment gages are officially named Rio Grande above Castolon, TX (USGS gage number 08374535) and Rio Grande above Rio Grande Village, TX (USGS gage number 08375295). Each sediment gage consists of a 1 MHz and 2 MHz side-looking acoustic Doppler profiler, paired with an automatic pump sampler. For simplicity, both the Castolon and RGV sediment gages and the nearby stream flow gages are referred to as Castolon and RGV in this paper.

The two-frequency acoustic data were calibrated using physical suspended-sediment samples. At high flow, standard depth-integrated samples were collected using a U.S. D-74 sampler deployed from a boat, and at lower flows, samples were collected with a U.S. DH-48 hand-held sampler while wading. Depth-integrated samples were collected using the Equal-Width-Increment (EWI) method at 10 vertical profiles across the channel [Edwards and Glysson, 1999].

When field crews were not available, suspended-sediment samples were collected automatically by the pump samplers. Samples collected by the pump samplers were calibrated to the cross section using paired EWI and pump measurements. Pump sampler calibrations were developed for silt and clay and for individual size classes of sand, as recommended by Edwards and Glysson [1999]. Calibrated-pump measurements included suspended-silt-and-clay concentration, suspended-sand concentration, and suspended-sand median grain size. Acoustic data were then calibrated to the EWI and calibrated-pump measurements and include concentrations of silt and clay, sand, and the median grain size of the suspended sand.
We calculated sediment loads using standard methods [Porterfield, 1972] using both the acoustic and physical suspended-sediment data. Calibrated acoustic data were combined with discharges measured at the nearby stream flow gages to calculate instantaneous loads of suspended silt and clay and suspended sand. Instantaneous loads were integrated over the hydrograph to calculate cumulative loads. Instantaneous median grain sizes were not calculated during periods where an acoustic Doppler profiler malfunctioned, because those calculations require measurements in two frequencies. An example of calibrated acoustic data is presented in Figure S1 in the supporting information. All EWI and calibrated-pump and acoustic data are available at http://www.gcmrc.gov/discharge_qw_sediment/ or http://cida.usgs.gov/gcmrc/discharge_qw_sediment/.

4.2. Suspended-Sediment Monitoring of Tributaries

Automatic pump samplers were installed on Terlingua and Tornillo Creeks, two large ephemeral tributaries (Figure 1b). Terlingua Creek enters the Rio Grande approximately 11 km upstream from Castolon, and Tornillo Creek joins the Rio Grande approximately 2 km upstream from RGV. Pump samples were collected at discrete time intervals, initially triggered by an increase in stage. All samples were analyzed for silt-and-clay concentration, sand concentration, and the suspended-sand grain size distribution. We assumed that the suspended sediment is well mixed in the cross sections during flash floods and that suspended sediment obtained from the pump samples was representative of the concentrations throughout the cross section. On Terlingua Creek, 15 min discharge data were measured at a stream gage operated by the International Boundary and Water Commission (gage number 08374300), and sediment concentrations were combined with these discharge data and integrated over all flows to calculate the cumulative silt-and-clay and sand loads delivered to the Rio Grande.

On Tornillo Creek, no stream flow measurements were made, because most flash floods occur at night, so a pressure transducer was installed and programmed to collect flow depth measurements every 15 min such that a near-continuous stage record was created. We then used high-resolution topographic data obtained from an aerial lidar (light detection and ranging) survey, and surveys of high-water marks to build a two-dimensional hydraulic model using the USGS international River Interface Cooperative model framework and its hydrodynamic model Flow and Sediment Transport with Morphological Evolution of Channel [McDonald et al., 2005]. We ran the model such that the root-mean-square error between modeled water surface elevations and the surveyed high-water marks was minimized [Griffiths et al., 2010]. We used output from our flow model to create a stage-discharge relation and combined this with our 15 min stage record to produce a 15 min record of discharge (see Text S1 for additional details). Sediment concentrations obtained from the pump samples were combined with the model hydrograph and integrated over time to determine the cumulative silt-and-clay and sand loads contributed to the Rio Grande.

4.3. Constraining Other Important Sediment Sources and Constructing the Sediment Mass Balance

Construction of sediment budgets based on the difference in flux at two gages provides insight into geomorphic change of the intervening reach and thus integrates all of the variables in equation (7):

\[ I - E = \Delta S \]  

(10)

where \( I \) is the influx of sediment, \( E \) is the efflux, and \( \Delta S \) is the change in sediment storage. To be meaningful, equation (10) must incorporate all influxes. However, in the Big Bend, many ungaged tributaries exist between Castolon and RGV that contribute sediment to the Rio Grande.

We analyzed the stream flow and sediment records at Castolon and RGV and identified spikes in silt-and-clay concentration at RGV that are not present at Castolon, which represent inputs from ungaged ephemeral tributaries. We modified equation (10) by adding an ungaged tributary \((I_{UT})\) term as well as a term to account for inputs by Tornillo Creek \((I_{TN})\):

\[ I + I_{UT} + I_{TN} - E = \Delta S \]  

(11)

To quantify the \( I_{UT} \) term, we analyzed the acoustic silt-and-clay concentration data and determined that pulses of silt and clay take an average of 24 h to travel between Castolon and RGV, and an average of 30 min to travel between Tornillo Creek and RGV. We shifted the instantaneous and cumulative silt-and-clay loads at Castolon and Tornillo Creek by these travel times such that the silt-and-clay loads at Castolon and Tornillo Creek were approximately coincident with the silt-and-clay loads at RGV. We then subtracted the time-shifted instantaneous silt-and-clay loads at Tornillo Creek from the instantaneous silt-and-clay loads...
at RGV and identified every event where the instantaneous silt-and-clay load at RGV was at least 500 kg/s greater than the time-shifted Castolon record. This method was used to identify the flash floods that originated in ungaged tributaries.

The threshold difference of 500 kg/s was used because it is approximately the lower limit of instantaneous silt-and-clay loads measured during Tornillo Creek flash floods, and because it is approximately the lowest value of the time-shifted difference in loads that exceeds the inherent variability of the method. For example, the traveltime of large silt-and-clay flux events between Castolon and RGV is not constant. Some events take slightly less than 24 h to transit the study area, and other events take slightly more time. This variation in traveltime results in oscillation by about zero for time-shifted differences in loads, with absolute values less than several hundred kg/s.

Estimation of the ungaged tributary silt-and-clay loads using this approach required accounting for the amount of silt and clay typically deposited upstream from RGV during ungaged tributary floods. During discrete flash floods that passed both Castolon and RGV, and during times of no activity on the ungaged tributaries, we calculated that between 27 and 41% of the cumulative silt-and-clay load measured at Castolon was deposited in the study area. Therefore, we estimated that approximately 33% of the silt and clay contributed by ungaged tributaries was also likely deposited upstream from RGV. We thus applied a correction factor of 1.5 to the silt-and-clay loads measured at RGV during each flood interpreted to originate in ungaged tributaries, to correct for the 33% of sediment that was likely deposited.

Flash floods in both Terlingua and Tornillo Creeks were assumed to be representative of flash floods in the ungaged tributaries, because they both drain a wide variety of the geologic settings in the Big Bend. Silt and clay loads in both of those watersheds exceed sand loads by a factor of approximately 12.5. Thus, once the ungaged tributary silt-and-clay loads were estimated, the sand loads of the ungaged tributaries were estimated by dividing the silt-and-clay loads by a factor of 12.5.

Using the above logic, calculations of ungaged tributary loads were defined for silt and clay

\[ 1.5 \left( \frac{I_{\text{RGV-SC}}}{I_{\text{TORN-SC}} - I_{\text{C-SC}}} \right) > 500 \text{ kg s}^{-1} \]  

and for sand

\[ I_{\text{UT-SC}}/12.5 = I_{\text{UT-SAND}} \]  

where \( I_{\text{RGV-SC}} \) is the instantaneous silt and clay load at RGV, \( I_{\text{TORN-SC}} \) is the time-shifted Tornillo Creek silt-and-clay load, \( I_{\text{C-SC}} \) is the time-shifted Castolon silt-and-clay load, \( I_{\text{UT-SC}} \) is the ungaged tributary silt-and-clay input, and \( I_{\text{UT-SAND}} \) is the ungaged tributary sand input.

There is large uncertainty in these estimates of ungaged tributary inputs of fine sediment, but we are unable to further constrain these estimates without measuring the many tributary inputs directly. We assumed that the uncertainty for inputs from ungaged tributaries was 50%. For a discussion of other potential sources of uncertainty, and how uncertainty was calculated for the suspended-sediment budget, see Text S2. All of the above calculations regarding traveltimes, silt-and-clay attenuation estimates, and the ratios of silt and clay to sand loads can be recalculated using the data served at http://www.gcmrc.gov/discharge_qw_sediment/. This website also includes an automated sediment-mass-balance computation tool that can be used to calculate the sediment mass balance for the study area for any time period of interest.

### 4.4. Back Calculation of Sediment Supply From Suspended-Sediment Data

We used the methods of Rubin and Topping (2001, 2008) to evaluate the relative influence that bed-sediment grain size and flow exert on suspended-sediment transport, through the calculation of their parameter \( \alpha \) using data obtained from physical suspended-sediment samples. \( \alpha \) is a nondimensional parameter that scales changes in bed-sediment grain size by changes in flow in order to determine which of those variables exert a stronger influence on suspended-sediment transport (Text S3). We calculate \( \alpha \) using the standard deviation of concentrations and median grain sizes of a series of physical suspended-sediment samples [Rubin and Topping, 2001, equations (8) and (9)] (equation (1) and Text S3), over periods of steady high flow. We also used the methods of Rubin and Topping (2001, 2008) to calculate a relative dimensionless measure \( \beta \) of the median grain size of the bed sediment \( D_50 \) (equation (7)) from the suspended-sand data. \( \beta \) includes the influence of both \( D_50 \) and \( A_j \); however, the behavior of \( \beta \) is dominated by \( D_50 \) [Topping et al., 2010] (Text S3). Values of \( \beta \) were calculated using equation (1) in Rubin and Topping (2008) for Castolon and RGV (equations (2) and (3) and Text S3).
Rubin and Topping [2001, 2008] derived $\alpha$ and $\beta$ using the Rouse mechanics-based theory of McLean [1992]. They then evaluated the accuracy of $\alpha$ to detect flow versus grain size regulated transport using the flume data sets of Guy et al. [1966] and evaluated the accuracy of $\beta$ backcalculations of $D_b$ using the flume data sets of Einstein and Chien [1953] and the river data set of Topping et al. [1999]. Rubin and Topping [2001, 2008] found that $\alpha$ accurately distinguishes flow-regulated from grain-size-regulated transport and found the $\beta$ backcalculations of $D_b$ to be in good to excellent agreement with the measured values of $D_b$ in these three data sets. Topping et al. [2010] conducted further tests of the accuracy of using $\beta$ to backcalculate $D_b$ and found that the agreement between backcalculations and measurements of $D_b$ were generally good during three floods on the Colorado River. These comparisons were conducted using $\beta$ calculated from suspended-sediment measurements made in the same river cross sections as the bed-sediment measurements.

There is good evidence that the $\beta$ backcalculations of $D_b$ may provide a more accurate measure of the reach-averaged $D_b$ upstream from a suspended-sediment measurement station than do a sparse set of bed-sediment measurements made at only one cross section. During higher discharges in the Rio Grande, the sediment in suspension likely equilibrates with the bed over spatial scales of many hundreds of meters. Therefore, $\beta$ provides a more representative relative reach-scale measure of $D_b$ than a limited number of direct samples of the bed because $\beta$ uses the physical suspension processes in the river to “sample” the bed in exactly the proportion that various bed-sediment environments interact with the flow [Topping et al., 2010]. Findings of Rubin et al. [2010] combined with Topping et al. [2010] confirm this process, whereby calculations of $\beta$ during the 2008 controlled flood experiment on the Colorado River in Grand Canyon [Topping et al., 2010, Figure 13A] showed the same spatial patterns as hundreds of measurements of bed-sediment grain size collected prior to the flood [Rubin et al., 2010, Figure 2G]. Therefore, although we have no direct observations of bed-sediment grain size and do not directly backcalculate $D_b$ in this study, we assume that calculations of $\beta$ provide an accurate relative measure of bed-sediment grain size changes.

Changes in $A_s$ that are distinct from changes in $D_b$ can only be inferred from suspended-sand data during cases where the flow strength is constant based on the logic associated with equations (5) and (6) in section 3.1. Using this logic, we use water discharge as a proxy for $U_*$, a justifiable approximation because $U_* = U$ and $U = Q/(h b)$. Channel geometry at Castolon and RGV are similar, and simulations from a one-dimensional hydraulic model employed in a previous study [Dean and Schmidt, 2013] show that $U_*$ is similar at both Castolon and RGV for similar discharges, and thus, this approximation is valid. If both the water discharge and median grain size of suspended sand are constant, but suspended-sand concentration decreases, the likely cause of this decrease in suspended-sand concentration is a decrease in $A_s$ (equations (5) and (6)). Therefore, we use changes in $\beta$ to deduce qualitative changes in $A_s$ during periods of relatively steady flow.

### 4.5. Longitudinal Trends in Suspended-Sediment Transport During a Steady Dam Release Flood

Between July and October 2013, a long-duration release from Luis L. Leon Dam occurred in three pulses. The purpose of this release was to create flood storage capacity within Luis L. Leon Reservoir. A 15 day pulse occurred in July and August and was approximately steady at 180 m$^3$/s. Subsequent pulses were released in September and October, and those pulses peaked at approximately 100 and 150 m$^3$/s, respectively.

We launched a 2 day river trip at Castolon on 2 August 2013 and conducted a quasi-Lagrangian longitudinal suspended-sediment sampling campaign during the steady part of this release wherein the measurements were made within a reference frame that approximated the velocity of the sampled water [Meade et al., 1985; Meade and Stevens, 1990; Topping et al., 2010]. During the sampling campaign, depth-integrated suspended-sediment samples were collected every 8 km using a fixed time interval between Castolon and RGV in an attempt to collect samples of the same packet of water. Sample collection times were based on the average surface water velocity, hence the term quasi-Lagrangian (hereafter referred to as “Lagrangian”). It was not feasible to collect complete EWI samples because of time constraints, and we limited our sampling to three depth-integrated samples along the channel centerline at each location. There were no observed tributary inflows between the gages during our sampling campaign.

We compared the measured suspended-sediment concentrations with local low-flow water surface slope, $\beta$, $b$, and $h$ at each sample location to investigate longitudinal patterns in concentration within the context of the general geomorphic and hydraulic setting. To estimate local slope at each sampling location, we used the average slope over a 1 km reach obtained from the 2012 lidar-generated digital elevation model. Each
1 km reach was centered on each sampling station. Channel widths ($b$) were extracted from active-channel boundaries delineated in a geographic information system using a combination of aerial photographs and the lidar data. Midchannel depth ($h$) was measured during the sampling campaign. $F$ tests were run on least squares linear regressions between suspended-sand concentration and $S$, $\beta$, $b$, and $h$.

### 4.6. Geomorphic Analyses of Channel Cross Sections

In 2009 and 2010, we established a channel monitoring program in three reaches (Castolon, Solis, and RGV) (Figures 1b and S2). The Castolon reach is approximately 2.5 km long and includes the Castolon gage. The Solis reach is 2.5 km long and is approximately 80 river km downstream from the Castolon reach. The RGV reach extends from the RGV gage 6 km upstream. We surveyed cross sections of the channel and floodplain using a Real-Time Kinematic GPS. The cross sections in the Castolon and RGV reaches were surveyed annually between 2009 and 2014, and the cross sections in the Solis reach were surveyed annually between 2010 and 2014. There are 19 cross sections in the Castolon reach, 16 cross sections in the Solis reach, and 33 cross sections in the RGV reach. Cross-section locations span the entire range of channel morphologies within each reach. We calculated changes in cross-section area to investigate rates of erosion and deposition relative to changes in the sediment mass balance. For a discussion of the sampling design, and uncertainties associated with channel cross-section measurements, see Text S4.

### 5. Results

Results from the suspended-sediment monitoring program indicate that at both Castolon and RGV, there may be up to ~3 orders of magnitude variation in silt and clay and sand concentration for any given discharge of water (Figure S3). To address the different processes that govern this large variability, we present five sets of analyses and link them to the hypotheses presented in section 3. In section 5.1, we show how spatial and temporal decreases in $u^*$ during tributary-sourced flash floods result in fine-sediment deposition within the Rio Grande, and how different grain size fractions become sorted downstream during these short-duration floods. We then analyze suspended-sediment dynamics and the sediment mass balance during long-duration dam releases in section 5.2 to describe how sediment transport varies at the two gages during similar $Q$, and therefore similar $u^*$, and how those differences are likely driven by temporal changes in the in-channel sediment supply. We then link those differences to the sediment mass balance. In section 5.3, we illustrate how spatial changes in flow strength, predominantly governed by changes in $S$ during steady flows, result in spatial variability in suspended-sand transport. We then analyze changes in $\beta$ over time to infer spatial and temporal changes in the in-channel sediment supply, and consequent effects on suspended-sediment concentration (section 5.4). Lastly, in section 5.5, we link changes in the sediment mass balance to on-the-ground measurements of geomorphic change.

These results indicate that temporal changes in the in-channel sediment supply arising from changes in $D_b$ and $A_s$ over long reaches (1-10s km) are responsible for temporal variation in suspended-sediment concentration at both sites. Additionally, during periods of steady flow over short time periods, spatial variability in suspended-sand transport is caused by spatial changes in $u^*$, largely governed by longitudinal variability in $S$.

#### 5.1. Rio Grande Discharge and Sediment Transport Characteristics During Tributary-Sourced Flash Floods

Discharge and suspended-sediment concentrations typically attenuate rapidly downstream when tributary-sourced flash floods enter the Rio Grande. Thus, there are large spatial decreases in $u^*$, and large quantities of sediment are deposited as described in example 1 in section 3.1, which supports our hypothesis (a). These types of floods therefore play an important role in the accumulation of fine sediment in the study area.

An extreme example of flood peak attenuation and fine-sediment deposition is illustrated during a flash flood on 2 June 2011. This flood arose in the lower Rio Conchos watershed, downstream from Luis L. Leon Dam. Discharge near the Rio Conchos confluence was approximately 200 m$^3$/s, and the flood wave attenuated to approximately 80 m$^3$/s at Castolon and to 20 m$^3$/s at RGV (Figure 2). The peak silt-and-clay concentration was an order of magnitude larger at Castolon than at RGV (Figure 2a). The peak sand concentration at Castolon was 260 mg/L (Figure 2b) and was less than 1 mg/L at RGV (Figure 2b). The dramatic attenuation
of discharge resulted in substantial decreases in concentration downstream, and more than 96% of the silt-and-clay load and 100% of the sand load was deposited between Castolon and RGV.

There were different hysteresis patterns between discharge and concentration for silt-and-clay and for sand. These differences are due to the different traveltimes of sand and of silt-and-clay, and different sources of supply for the two grain size fractions in different parts of the study area. Silt-and-clay concentrations were higher during the falling limb of the flood (Figures 2a and 2c), because the silt and clay lagged the discharge wave. Thus, silt-and-clay transport was primarily driven by the supply from far upstream. Clockwise discharge-sand concentration hysteresis (Figure 2d) and counterclockwise discharge-sand median grain size hysteresis (Figure 2e) occurred at Castolon. We speculate that the upstream supply of sand likely did not reach Castolon during this flood (section 3.2, hypothesis (b)), and the suspended sand transported past Castolon was locally derived from the bed a relatively short distance upstream from the gage. The decrease in sand concentration coupled with the increase in median grain size over the duration of the flood at Castolon indicates that progressive depletion of this local sand supply occurred.

Different sediment transport patterns were measured during a series of flash floods between 13 and 19 June 2013; these floods came from both the Rio Conchos and a number of tributaries (Figure 3). The differences indicate that suspended-sand transport during these floods was dependent upon different sediment sources.
at each gage. Sand concentration at both sites increased linearly with increasing discharge (Figure 3c); however, the suspended sand was finest at higher discharges at Castolon, and the suspended sand was the coarsest at higher discharges at RGV (Figure 3d). The finer grain sizes of sand at the highest concentration indicate that suspended-sand transport at Castolon was partially regulated by the upstream sediment supply. However, sand transport at RGV was mostly regulated by the local, in-channel sediment supply, and by hydraulics, where larger discharges always transported more sand, and the median grain sizes of sand coarsened with increasing discharge. Thus, even though the flood wave attenuated, suspended-sand concentrations were similar, or slightly larger at RGV, which indicates that the local in-channel supply near RGV was likely larger than at Castolon; this observation supports hypothesis (c) in section 3.1.

The uncertainty in the measurements is sufficiently large that we cannot determine if silt-and-clay accumulated or was evacuated from the study area. The mass balance for the 13–19 June 2013 flash floods indicates that 98,000 ± 172,000 t of silt and clay accumulated between the two gages (Figure 3e). Although the uncertainty bands are large, the majority of the uncertainty cloud is positive and suggests that this flood caused deposition of silt and clay between the two gages. The mass balance for sand was positive and demonstrates

Figure 3. Sediment dynamics at the Castolon and RGV gages during a flash flood from multiple source areas. (a) Discharge and concentration of suspended silt and clay. (b) Discharge and concentration of suspended sand. (c) Relation between discharge and suspended-sand concentration. (d) Relation between discharge and the median grain size of suspended sand. There were short-duration biases in acoustic sand concentration data; thus, only the physical measurements of sand concentration and grain size are shown. No biases were evident in the acoustic silt-and-clay concentration data, as shown by the physical measurement data (blue and red symbols) overlain on the acoustic data (lines). Error bars in Figures 3b–3d indicate 95% confidence intervals. (e) Change in storage for silt and clay, and sand between the RGV and Castolon. Green bands around the zero-bias lines in Figure 3e indicate the region of accumulating uncertainty arising from 10% possible persistent biases in measured Rio Grande silt and clay and sand loads, 20% possible persistent biases in measured Tornillo Creek silt and clay and sand loads, and 50% possible persistent biases in ungaged tributary silt and clay and sand loads. See Text S2 in the supporting information for a description of possible biases.
that sand was deposited between Castolon and RGV (Figure 3e). We examined other tributary-sourced flash floods and found that in nearly all cases, the universal downstream attenuation of discharge results in the deposition of fine sediment between Castolon and RGV. Additionally, there is proportionally less deposition of silt and clay, as supported by the Rouse-based hypothesis (b) in section 3.2.

5.2. Rio Grande Sediment Transport Characteristics During Long-Duration Dam Releases

During long-duration dam release floods, there may be substantial variability in suspended-sediment transport at Castolon and RGV, even though the magnitudes of the floods are essentially the same at each location. This variability appears to be linked to spatial differences and temporal changes in the in-channel sediment supply and is specifically reflected in the variability in suspended-sand transport. These findings support hypotheses (b) and (c) in section 3.1, whereby erosion and deposition are dependent upon the spatial distribution of fine sediment throughout the study area, and whereby a larger in-channel supply of fine sediment may result in larger transport rates for similar flow strength.

Dam releases in 2011 and 2012 were of similar magnitude and duration (Figures 4a–4d). The 2013 dam release consisted of three pulses, all of which were generally longer, and considerably larger, than the dam releases in 2011 and 2012 (Figures 4e and 4f). Tributary-sourced floods occurred during all of these releases for short periods, as indicated by the abrupt spikes in both discharge and sediment concentration (Figure 4).

There was little difference in measured silt-and-clay concentration between Castolon and RGV during each year’s dam releases (Figures 4a, 4c, and 4e), with the largest steady discharge silt-and-clay concentrations
occurring during the 2012 dam release flood. However, sand transport at Castolon and RGV varied greatly among the different dam releases. In 2011 and 2012, sand concentrations during the dam releases were approximately twice as large at Castolon as at RGV (Figures 4b and 4d, 5a and 5c). This pattern substantially changed during the 2013 dam release, whereby sand concentrations were considerably lower at Castolon.

Figure 5. Relations between discharge and sand concentration, and discharge and the median grain size of suspended sand during the (a, b) 2011, (c, d) 2012, and (e–h) 2013 dam releases. These data are from steady discharge parts of the floods when tributaries were not flooding; therefore, the effects of tributary flash floods have mostly been removed. There is an incomplete acoustic grain size data record; thus, grain size data are physical measurement data only. The gray point cloud is all of concentration and grain size data for steady discharge periods during all dam releases. Note that sand concentrations at the RGV sediment gage were much lower than at the Castolon sediment gage in 2011 and 2012 and that in 2013, sand concentrations were larger at the RGV sediment gage compared to the Castolon gage. These trends show the progressive downstream enrichment of sand over time between 2011 and 2013. Note that median grain sizes of the suspended sand were generally coarser at Castolon except during the 2012 dam release, and pulse 3 of the 2013 dam release.
than at RGV (Figures 4f, 5e, and 5g). In general, median grain sizes of suspended sand are coarser at Castolon than at RGV (Figures 5b and 5f). Flash flood activity can sometimes increase the finer sand supply upstream from Castolon, and median grain sizes of suspended sand were similar at the two sites during the 2012 dam release and during most of pulse 3 of the 2013 dam release (Figures 5d and 5h).

The very long duration of the 2013 dam release depleted the in-channel sand supply at Castolon. Clockwise hysteresis between discharge and sand concentration during pulse 2 in 2013 occurred at each gage, indicating the progressive depletion of sand in the reaches upstream from the gages (Figure 5e). Counterclockwise hysteresis between discharge and the median grain size of the suspended sand supports this interpretation; however, the hysteresis is subtle and is not discernible in Figure 5f. Therefore, sediment depletion was not great enough to substantially affect the median grain sizes in suspension. During pulse 3, sand concentrations at RGV were higher than during any of the other dam releases (Figure 5g). These high sand concentrations indicate that the in-channel sand supply at RGV had become enriched over time, and the relatively higher concentrations downstream at RGV compared to Castolon, combined with the clockwise hysteresis between discharge and suspended-sand concentration, indicate that sand was eroded from the study area.

Estimates of the mass balance in the study area indicate that long-duration dam releases erode silt and clay from the study area (Figure 6). During the steady state parts of all of the dam releases, the silt and clay budgets were negative (Figures 6a, 6c, and 6e); however, tributary-sourced flash floods offset much of the erosion that occurred during the steady state parts.
of the releases. The mass balances of silt and clay in 2012 and 2013 were indeterminate (Figures 6c and 6e) because many fine-sediment inputs from ungauged tributaries occurred, and there is large uncertainty in the estimates of the mass of sediment delivered from those sources.

In contrast to the sediment budgets for silt and clay, sand accumulated in the study area during the 2011 and 2012 dam releases but was eroded during the 2013 dam release (Figures 6b, 6d, and 6f). Accumulation occurred during the 2011 and 2012 dam releases, because higher sand concentrations occurred upstream at Castolon, and more sand entered the study area than was transported past RGV. The lower sand concentrations at RGV in 2011 and 2012, and sand accumulation within the study area, indicate that there was a local sand supply limitation upstream from RGV during those years. During the 2013 dam release, 100,000 ± 63,000 t of sediment was eroded (Figure 6f). Therefore, the higher concentrations downstream at RGV in 2013, and the net erosion of sand during the 2013 release indicate that the in-channel sand supply upstream from RGV had become enriched prior to the 2013 dam release. The sand enrichment within the study area was likely caused by the numerous flash floods that occurred between the 2012 and 2013 dam releases. The temporal changes of sand transport during periods of steady discharge, and therefore flow strength, thus provided important insight into the temporal changes in the local in-channel sediment supplies. These results show that the in-channel sediment supply increased near RGV over time, which was reflected in larger sediment transport rates for the same discharge (hypothesis (c), section 3.1).

5.3. Longitudinal Trends in Suspended-Sediment Transport

Downstream variations in suspended-sediment concentration during steady flows provide insight as to whether the mass balances calculated from the flux measurements at Castolon and RGV were uniformly distributed throughout the reach, or whether the mass balance was the net difference among shorter reaches where erosion or deposition occurred. Using an analysis of basic geomorphic attributes, combined with suspended-sediment transport data collected during the Lagrangian sampling campaign, we found that the locations and rates of erosion and deposition are controlled by spatial changes in $S$ during steady flow, because of the dominance of $u^*$ in equation (7). It is difficult to evaluate the effects of $h$ and $b$ on flow strength, because those variables were only measured at the sampling locations, and thus, we likely did not have a robust enough measure of $h$ and $b$ to fully characterize its influence on transport. These findings support hypothesis (b) in section 3.1 whereby spatial changes in flow strength result in spatial variability in suspended-sediment transport.

At the launch of the Lagrangian sampling campaign, the concentration of silt and clay at Castolon was 2150 mg/L; the concentration of silt and clay was 2800 mg/L at RGV a day and a half later when we completed the trip (Figure 7a). There was a slight decline in silt-and-clay concentration when we camped overnight, 56.5 km downstream from Castolon; thus, we sampled a different “packet” of water on the second day (Figure 7a). Aside from the overnight decline, suspended-silt-and-clay concentration progressively increased downstream, demonstrating that silt and clay was uniformly eroded from all parts of the study area during this 2 day period.

In contrast to silt and clay, there was no progressive longitudinal increase in the concentration of sand. Instead, there were alternating zones of increasing and decreasing sand concentrations that were caused by zones of sediment erosion and deposition, respectively (Figure 7a). In the upstream half of the study area, the zones of erosion and deposition were short—generally less than the distance between our sampling locations—and the transitions from one to another were abrupt. In the downstream half of the study area, there was a long zone of deposition, followed by a long zone of erosion. The zones of deposition occurred between 49 and 89 km downstream from Castolon, where there was a progressive decline in channel slope upstream from a tributary confluence; the zone of erosion was downstream from this tributary confluence and extended to the bottom of the study area (Figures 7a–7c).

These zones of erosion or deposition correlate to zones of steep and flat slope, respectively (Figures 7b–7d), as predicted by equations (7) and (8). The correlation is weak, however, because the linear regression between sand concentration and $S$ is unduly influenced by the two highest concentrations that exist for slopes less than 0.0005 (Figure 7d). Neglecting either one of these points results in a t test on the regression that yields a significant $p$ value ($< 0.05$). Equations (7) and (8) also suggest that zones of erosion and deposition could also be dependent upon $h$ and $b$; however, correlations were weak between those...
variables and suspended-sand concentration (Figure S4). Therefore, $h$ and $b$ were either not important drivers of suspended-sediment transport or their effects were not detectable at the scale of our sampling design.

Equation (7) also states that if flow strength is constant, spatial differences in the sediment supply, manifested by changes in either $D_b$ or $A_s$, should regulate the rate of erosion or deposition. Analyses of $\beta$ values show that the highest sand concentrations occurred in areas where the bed sediment was the coarsest (Figures 7e and 7f). This result is the opposite of the trend expected in equation (5). However, this is consistent with the above observation that the highest sand concentrations occurred in the regions of highest $S$, because areas of high $S$ generally have coarser beds. This suggests that spatial increases in $D_b$ were not large enough to offset spatial changes in $S$ during the sampling campaign. Calculations of $\alpha$, a parameter that scales the effects of grain size to flow, are much less than 1 (0.32 and 0.27 for narrow and wide grain size distributions [Rubin and Topping, 2001]), indicating that at the beginning of the release, longitudinal changes in hydraulics had a larger influence over suspended-sand concentration at our sampling locations than did longitudinal changes in bed-sand grain size. Therefore, during the initial part of the 2013 dam release, longitudinal changes in flow strength arising from changes in $S$ controlled the locations of erosion and deposition, and spatial changes in $D_b$ were not large enough to offset those changes in $S$. These findings are similar to those of Grams et al. [2013] which showed that local hydraulics exert a primary control on erosion and deposition in the Colorado River in Grand Canyon.
5.4. Temporal Variation in Sediment Supply

Results from the Lagrangian sampling campaign in the previous section show how spatial changes in sand transport in the Rio Grande are controlled by spatial changes in $u^*_s$ (arising from spatial changes in $S$) over a relatively short time period. In this section, we evaluate the importance of the in-channel sediment supply in regulating sand transport over longer time periods at Castolon and RGV. During steady flow, temporal changes in $\beta$ coupled with temporal changes in suspended-sand concentration can provide insights into temporal changes in the in-channel sediment supply, because these metrics can be used to infer changes in $D_b$ and $A_s$. Here we show that temporal variation in sediment supply strongly influenced sand transport at both gages, leading to periods where greater sand transport occurred at either Castolon or RGV for the same discharge. These differences in transport between the two gages determined whether net erosion or deposition occurred within the study area and supports hypothesis (c) in section 3.1.

$\beta$ values generally indicate that the sand on the channel bed was coarser near Castolon than near RGV. Tributary-sourced flash floods that occur upstream from Castolon cause the bed near Castolon to fine (Figure 8a). However, fining of the bed at Castolon is short lived, and the finer grain sizes are rapidly depleted, resulting in bed coarsening (increase in $\beta$) during flood recession. The sand on the channel bed at RGV, however, can either coarsen or fine during tributary-sourced flash floods, depending on the source area of the flood.

During the steady flow periods of the dam releases, analyses of $\beta$ at Castolon and RGV during similar discharge (and therefore similar $u^*_s$) highlight three different phenomena. First, larger $\beta$ values at Castolon (Figure 8b) coupled with larger suspended-sand concentrations during the 2011 dam release (Figure 5a) verify that there was likely an in-channel sand supply limitation downstream near RGV as described in section 5.2. Since $\beta$ was smaller, and therefore, bed-sand grain size was likely finer at RGV, the sediment supply limitation was likely manifested in a smaller relative area of the bed covered by sand (lower $A_s$) at RGV compared to at Castolon. This resulted in lower sand transport rates at RGV, and the accumulation of sand within the study area.

Second, $\beta$ values at Castolon and RGV during 2012 dam release were similar and were lower than $\beta$ values at Castolon during the 2011 dam release (Figure 8c), which indicates that the in-channel sand supply near Castolon had fined. Yet even though $\beta$ was similar at the two sites, and therefore bed-sand grain size was likely similar, sand transport rates were still considerably larger upstream at Castolon compared to RGV, which indicates that the in-channel sand supply limitation persisted near RGV (lower $A_s$).
Third, at the beginning of the 2013 release, $\beta$ values were similar at Castolon and RGV (Figure 8d); however, with each pulse, $\beta$ values progressively increased at Castolon indicating that the in-channel sand supply upstream from Castolon was progressively depleted. As described in section 5.2, suspended-sand concentrations were much greater downstream at RGV during the 2013 release (Figures 5e and 5g). The lower $\beta$ values, and the higher concentrations at RGV, with respect to Castolon, indicate that there was no longer an in-channel sand supply limitation upstream from RGV, and this was likely driven by an increase in $A_s$ between the 2012 and 2013 releases. This increase in the in-channel sand supply near RGV was likely a result of the many tributary-sourced flash floods that occurred during that period. Thus, the in-channel sand supply near RGV progressively increased between 2012 and 2013 (through a likely increase in $A_s$), shown by a general increase in suspended-sand concentration relative to Castolon. These increases in concentration at RGV occurred while $\beta$, and therefore $D_{br}$, remained relatively constant.

5.5. Measured Cross-Section Changes, 2009–2014

Resurveys of channel cross sections in the Castolon, Solis, and RGV reaches demonstrate that channel contraction, through bed aggradation and channel narrowing, was the dominant geomorphic response during the past 5 years (Figures 9 and 10). The highest rates of channel contraction occurred between 2009 and 2010, before sediment transport measurements began, and between 2012 and 2013 (Figure 9). Between 2009 and 2010, the average decrease in cross-section area was 4.2% and 1.3% in the Castolon and RGV reaches, respectively (Figure 9); the Solis reach was not surveyed in 2009. Between 2012 and 2013, the average decrease in cross-section area was 4.6%, 2.8%, and 3.2% in the Castolon, Solis, and RGV reaches, respectively. Some cross sections were reduced in area by ~15% during those years (Figure 9a). By the end of 2012, the three reaches

![Figure 9. Average (a) annual percent change in cross-section area and (b) cumulative percent change in cross-section area for the three study reaches between 2009 and 2014. Error bars represent minimum and maximum changes for each reach.](image)

![Figure 10. Examples of cross-section changes between 2009 and 2014. (a and b) Channel narrowing is shown. Bed scour caused by the 2013 dam release flood is also shown in Figure 10a.](image)
had lost an average of 11%, 3.9%, and 4.7% of cross-section area, respectively, when compared to the first 
surveys in 2009 or 2010. Examples of cross-section changes are shown in Figure 10.

Analyses of average cross-section change show that the greatest channel contraction occurred during 2012, 
which had the highest frequency of flash floods and had the largest positive changes in sediment storage. 
These data indicate that the largest geomorphic changes were linked to the frequency of tributary flooding 
and that the cumulative sediment budget correlates to on-the-ground measurements of change. Lastly, the 
only year when channel expansion occurred, or at least, no channel contraction occurred, was between 2013 
and 2014, which had the largest, longest-duration dam release. In the case of tributary flooding, channel 
contraction is linked to large, systematic reductions in $D_0$ (as inferred through changes in $\beta$, Figure 8a) and 
likely increases in $A_0$ (inferred through changes in concentration at constant discharge with no change in $\beta$). 
During large, long-duration dam releases, channel expansion and/or the maintenance of channel area is likely 
linked to large, systematic increases in $D_0$ and decreases in $A_0$.

6. Discussion

Geomorphologists have long sought to link geomorphic change to perturbations in the sediment mass 
balance, and numerous studies have employed versions of equation (10), as well as efforts to partition the 
budget such as in equation (11) [Trimble, 1983; Erwin et al., 2011; Grams et al., 2013]. Because changes in 
 sediment storage are typically small over short timescales, sediment budgets are usually calculated over 
timescales of a few years or decades in order to estimate channel change during the intervening period. In 
order to do so, sediment-rating curves, as well as other methods (e.g., cosmogenic isotope dating), are 
typically used to estimate annual sediment influx and efflux, because historically, it has not been possible 
to measure sediment transport in a temporally precise way.

The detailed flux measurements described in this study demonstrate that for rivers where suspended-
 sediment transport dominates the flux, sediment-rating curves are likely insufficient for flux and sediment 
 mass balance calculations. Here we continuously measured flux at the upstream and downstream ends of 
the study area, measured influx from two large tributaries, and estimated influx from other tributaries. The 
measurements described here showed that large variability in suspended-sediment transport occurred in 
the Rio Grande, with many different patterns of hysteresis. We demonstrate that the nature of the hysteresis 
differs for silt and clay and for sand, and we show that the hysteresis differs in time and space depending if 
suspended-sediment transport is controlled by local hydraulics, the upstream sediment supply, or the local 
sediment supply. Thus, we demonstrate that sediment influx and efflux changes with time as sediment 
accumulates or is evacuated from the channel.

Together with previous studies of the Rio Grande in the Big Bend, the findings described here provide a rich 
picture of a disequilibrium channel that progressively fills with fine sediment during decades of low and mod-
erate flows dominated by Chihuahuan Desert flash floods and occasionally evacuates fine sediment during 
very large channel-reset floods. Our measurements of fine-sediment flux describe the watershed and local 
controls on the rate of fine-sediment accumulation during the periods immediately following reset floods. 
We show that depositional and erosional processes change with time as the boundary conditions of the 
channel change. We also show that the characteristics of the flow regime—whether tributary-sourced floods 
or long-duration dam releases—affect the rate and characteristics of fine-sediment accumulation that in turn 
affect subsequent rates of suspended-sediment transport and channel change.

These findings demonstrate that the channel of the Rio Grande responds in complex ways to the changing 
flow and sediment regimes of the Rio Grande watershed and the changes in upstream reservoir operation. 
The complex response of channels to changing flow and sediment supply regimes has been extensively 
described elsewhere [Schumm, 1973; Schumm and Parker, 1973; Baker, 1977; Nanson, 1986; Graf, 1987; 
Kochel, 1988; Friedman et al., 1996], but without temporally and spatially precise measurements of flux that 
describe the physical processes that cause these responses. Schumm [1973] summarized how large-scale 
changes in flow or sediment supply can cause geomorphic perturbations that propagate up and down a drai-
nage network, resulting in spatially and temporally variable processes of erosion and deposition. However, 
the large spatial extent of these geomorphic interactions, and the potentially long timescales associated with 
geomorphic adjustments, often hinder the ability to investigate the details of those processes that occur at 
small spatial and temporal scales. The sediment transport measurements conducted with high temporal
resolution in this study provide the ability to investigate some the detailed sediment transport processes associated with the spatial and temporal complexities of large-scale geomorphic responses.

Most studies focusing on channel recovery after large floods show that the initial stages of recovery consist of sediment accumulation within the overcapacity channel. On alluvial rivers, with large suspended-sediment loads, sediment accumulation occurs through the oblique and vertical accretion of fine sediment along the channel margins resulting in the construction of “benches” that become stabilized by vegetation, and evolve into floodplains [Costa, 1974; Nanson, 1986; Friedman et al., 1996, 2005; Moody et al., 1999; Pizzuto et al., 2008; Dean et al., 2011]. The landforms created during the channel recovery process, and the rates associated with the development of those landforms are often the primary focus of the associated research. However, few studies discuss the spatial and temporal variability associated with those processes [Wilcox and Shafroth, 2013], and linkages between depositional processes and sediment supply and sediment transport at the field scale are even more rare [Pizzuto et al., 2008; Perignon et al., 2013; Griffin et al., 2014].

In this study, acoustic measurements of suspended-sediment transport, combined with physical sampling efforts provided the ability to quantify both the upstream sediment supply (manifested in the sediment flux), and temporal changes to the in-channel sediment supply during channel recovery processes on the Rio Grande. We found that the rate of sediment accumulation is strongly linked to the frequency of tributary-sourced flash floods. Coupled changes in suspended-sediment concentration and grain size provided the ability to infer temporal changes to the in-channel sediment supply driven by periods of high tributary flash flood frequency. We were specifically able to link patterns of hysteresis in the sediment transport record to both depositional and erosional processes and evaluated whether long-duration dam releases exacerbate or alleviate the sediment loading that occurs during those periods of high flash flood frequency. Lastly, during periods of steady discharge, we were able to determine that spatial variability in flow strength primarily controls spatial changes in transport at any given location, yet changes in sediment supply can strongly influence sediment transport rates over time.

We present a conceptual model that describes the suspended-sediment transport processes, and the relative time line of geomorphic change on the Rio Grande in the Big Bend (Figure 11). This conceptual model...
presents temporal patterns of geomorphic change and changes in sediment supply driven by tributary-sourced floods, and long-duration dam releases, starting with the postreset channel configuration, such that existed following the 2008 channel-reset flood [Dean and Schmidt, 2013]. This conceptual model builds on previous conceptual models [Schumm, 1973; Baker, 1977; Nanson, 1986] and additionally links attributes of sediment supply and transport to geomorphic changes.

Following channel reset, the channel is supply-limited with respect to fine sediment (Figure 11a). The supply limitation reflects the large magnitude of erosion that occurs during channel-reset floods, which widens the channel and evacuates large quantities of fine sediment. In this state, the bed is coarse (high $D_b$), the area of the bed covered by fine sediment ($A_s$) is small, and suspended-sediment concentrations ($C_s$) are relatively low. This channel configuration represents an “overcapacity” condition whereby the channel is sized to the large channel-reset flood [Lane, 1955; Wolman and Gerson, 1978; Miller, 1987]. Barring further flooding of that magnitude, the channel begins to accumulate fine sediment because flow and sediment transport regimes following channel resets are generally greatly reduced. Thus, following channel reset, both tributary-sourced floods, and long-duration dam release floods will result in the accumulation of fine sediment within a reach (given some sufficient upstream sediment supply), as occurred in 2011 and 2012 when both tributary-sourced floods and dam releases occurred (Figure 6). The accumulation of fine sediment will cause decreases in $D_b$, and increases in $A_s$ and $C_s$, for similar flow strength (Figures 8c and 11b). Thus, given the usual reductions in the flow and sediment transport regimes that occur after channel-reset floods, some fine-sediment accumulation is inevitable, as described in other studies of channel recovery after large floods [Schumm and Lichty, 1963; Burkham, 1972; Friedman et al., 1996; Moody et al., 1999].

From the condition presented in Figure 11b, two trajectories can occur. If only tributary-sourced floods occur, fine sediment will continue to accumulate within the channel, causing continued decreases in $D_b$ and increased increases in $A_s$ and $C_s$, resulting in further channel contraction (Figure 11c). In the absence of any long-duration, moderate-magnitude floods, this trajectory will continue and the channel will narrow over time toward the condition depicted in Figure 11e.

During this type of trajectory, vegetation often establishes on bare deposits because there is insufficient hydraulic disturbance to scour seedlings [Wilcox and Shafroth, 2013]. Vegetation will stabilize these new deposits, thereby adding to bank strength and hindering erosion that may occur in the absence of vegetation (Figure 11e) [Tai et al., 2004; Pollen-Bankhead et al., 2009; Manners et al., 2014]. This occurred on the Rio Grande during the 17 years of narrowing prior to the 2008 channel-reset flood [Dean and Schmidt, 2011]. Under this scenario, $D_b$ will be at its finest condition, and $A_s$ will approach 1.

The trajectory from Figures 11a to 11e can be characterized by positive feedback mechanisms as described by Dean and Schmidt [2011], who showed that channel contraction occurred at rapid rates between 1991 and 2008. In this circumstance, the loss of channel capacity associated with channel narrowing resulted in widespread vertical accretion of the floodplain. Even though the elevation of the floodplain grew vertically away from the channel bed, the frequency of floodplain inundation did not decrease because of the progressive loss of channel capacity, and the inability of the channel to convey progressively smaller flood flows over time. Based on equation (7), positive feedbacks can only occur unabated for so long. As $D_b$ approaches the finest condition, and $A_s$ approaches 1, $C_s$ should increase for any given flow strength, thereby counteracting the progressive infilling of the channel bed with fine sediment, and resulting in an increased propensity for erosion. However, $C_s$ may decrease if channel margin vegetation is of sufficient density to reduce flow velocities [Griffin et al., 2005] and therefore flow strength, since $u_*$ is proportional to $U$. In this case, erosion will only occur if (1) increases in flow strength associated with increases in slope (equation (7)) caused by the progressive infilling of the channel occurs or (2) there is a flood of sufficient magnitude (i.e., sufficient $u_*$) to erode the accumulated sediment. In the case of the 2008 channel-reset flood, it was case (2) that resulted in large-scale erosion.

Along an alternative trajectory, long-duration dam release floods occurring in conjunction with tributary-sourced floods will offset some of the fine-sediment accumulation that occurs. If the dam releases are of insufficient magnitude and/or duration to erode the accumulated sediment from the reach, the channel condition will continue to track along the trajectory from Figures 11b to 11c. If the dam release is of sufficient magnitude and/or duration to convey the available sediment supply, the channel geometry will be maintained and $D_b$, $A_s$, and $C_s$ will remain constant for any given $u_*$ (Figure 11d). This is likely only a hypothetical scenario because there is a wide range of spatial variation of $D_b$ and $A_s$ such that these variables will likely...
The erosion of the 2013 dam release likely caused in controlling the locations of erosion and deposition at any given time (equation (7)), a the Big Bend. This study shows how spatial changes in supply, and how changes in supply correspond to changes in sediment storage within the Rio Grande. Made with high temporal resolution provide the ability to identify spatial and temporal changes in sediment geomorphic change. In all, we show how comprehensive measurements of suspended-sediment transport sediment can be as important, if not more so, than spatial changes in hydraulics in controlling geomorphic change in long reaches over time. We additionally link patterns of sediment accumulation and channel contraction, as determined from the channel cross-section measurements, occurred in the Castolon reach in the upstream part of the study area. Contraction here was likely driven by the large sediment inputs by Terlingua Creek, 9 km upstream. Sediment loads contributed by Terlingua Creek account for approximately one third of all of the sediment that passes Castolon. Based on equation (7), the channel infilling near Castolon will result in decreases in Δs, increases in Aυ, and therefore, the more efficient transfer of sediment downstream, if not confounded by the vegetative effects discussed above. In this circumstance, the more efficient downstream transfer of sediment will result in increased sediment delivery to downstream reaches, with corresponding increases in channel contraction, and thus, progressive downstream infilling. Similarly, increases in the sediment supply downstream near RGV may also occur with erosion and bed coarsening upstream from Castolon, such as appears to have occurred during the 2013 dam release [Stevens, 1938; Borland and Miller, 1960]. Furthermore, flash flood activity on ungaged tributaries within the study area may also cause deposition and increases in the in-channel sediment supply near RGV, with no changes upstream near Castolon. This highlights the importance of comprehensive measurements of sediment transport at multiple locations, because spatial variability in supply is manifested in different patterns of sediment transport at different locations.

If a dam release is of great enough magnitude or duration, erosion of accumulated fine sediment can occur, and the condition in Figure 11b can track to the condition in Figure 11f. In this study, the trajectory of sediment supply and channel change followed a pattern of sediment accumulation and erosion whereby (1) the overcapacity channel accumulated fine sediment during both tributary flooding and dam releases (Figures 11a to 11b, e.g., 2011 and 2012), (2) tributary floods continued to fill the channel, increasing the in-channel sediment supply (Figures 11b to 11c, e.g., the 2012 thunderstorm season), at which point (3) the 2013 dam release was of sufficient magnitude and duration to erode some portion of the accumulated sediment (Figures 11c to 11f). The erosion of the 2013 dam release likely caused Δs to increase, and Aυ to decrease in the reach upstream from Castolon. Erosion in 2013 was sufficient enough such that there were some measurable increases in cross-section area in the three monitoring study reaches.

The trajectory from Figures 11a to 11f is thus defined by negative feedback mechanisms, whereby increases in the in-channel sediment supply result in increased erosion rates for any given flow strength (e.g., 2013 dam release). For this negative feedback scenario to persist, dam releases of sufficient magnitude and duration must occur in order to maintain channel geometry and convey the available supply. In this scenario, the increase in supply caused by tributary-sourced flash floods, combined with the erosional effects of dam releases, will result in a condition where the bed elevation, or channel cross-section area (i.e., the left side of equation (7)) will not progressively increase or decrease over time.

There may be many spatial complexities not included in this conceptual model. For instance, the greatest degree of channel contraction, as determined from the channel cross-section measurements, occurred in the Castolon reach in the upstream part of the study area. Contraction here was likely driven by the large sediment inputs by Terlingua Creek, 9 km upstream. Sediment loads contributed by Terlingua Creek account for approximately one third of all of the sediment that passes Castolon. Based on equation (7), the channel infilling near Castolon will result in decreases in Δs, increases in Aυ, and therefore, the more efficient transfer of sediment downstream, if not confounded by the vegetative effects discussed above. In this circumstance, the more efficient downstream transfer of sediment will result in increased sediment delivery to downstream reaches, with corresponding increases in channel contraction, and thus, progressive downstream infilling. Similarly, increases in the sediment supply downstream near RGV may also occur with erosion and bed coarsening upstream from Castolon, such as appears to have occurred during the 2013 dam release [Stevens, 1938; Borland and Miller, 1960]. Furthermore, flash flood activity on ungaged tributaries within the study area may also cause deposition and increases in the in-channel sediment supply near RGV, with no changes upstream near Castolon. This highlights the importance of comprehensive measurements of sediment transport at multiple locations, because spatial variability in supply is manifested in different patterns of sediment transport at different locations.

Framing our analyses within a simple derivation of the Exner equation (7) provided the ability to examine how various aspects of flow and sediment supply affect the geomorphic evolution of the Rio Grande in the Big Bend. This study shows how spatial changes in flow strength may be the most important variable in controlling the locations of erosion and deposition at any given time (equation (7)), a finding similar to Topping et al. [2000b] and Grams et al. [2013]. However, our study highlights how large spatial changes in sediment supply can be as important, if not more so, than spatial changes in hydraulics in controlling geomorphic change in long reaches over time. We additionally link patterns of sediment accumulation and evacuation, as determined from the suspended-sediment budget, to on-the-ground measurements of geomorphic change. In all, we show how comprehensive measurements of suspended-sediment transport made with high temporal resolution provide the ability to identify spatial and temporal changes in sediment supply, and how changes in supply correspond to changes in sediment storage within the Rio Grande.

7. Conclusions

This study illustrates how a variable flow regime coupled with a large sediment supply results in complex geomorphic changes over time and space on the Rio Grande in the Big Bend region. We continuously...
monitored suspended-sediment transport with high temporal resolution to understand how flow strength and sediment supply control the geomorphic evolution of the Rio Grande over different spatial and temporal scales. Our monitoring program indicates that spatial differences in the in-channel sediment supply control net erosion or deposition in long reaches (kilometers to tens of kilometers) over time. However, spatial changes in flow strength arising from changes in slope exert the strongest control on the locations and rates of erosion or deposition at any given time. These results are two of the many reasons that suspended-sediment transport rates are rarely stationary for any given discharge on the Rio Grande. In all, this study shows how comprehensive suspended-sediment monitoring with high temporal resolution at the field scale can be used to understand the geomorphic evolution of an alluvial river in sediment surplus.

Increases in the in-channel sediment supply primarily occur when short-duration, tributary-sourced, flash floods enter the mainstem Rio Grande. These floods quickly attenuate within the Rio Grande, and large proportions of the tributary-derived load are deposited in the mainstem river, causing fining of the channel bed and increases in the areal extent of fine sediment on the channel bed. In contrast, fine-sediment accumulation is small or fine sediment is eroded when moderate-magnitude, long-duration, upstream dam releases occur. The effectiveness of dam releases in exporting sediment is dependent not only upon the discharge and duration of the dam release but also upon the antecedent sediment supply within the channel sourced from tributary flash floods. Thus, dam releases occurring when the grain size of the bed is fine, and when a large area of the bed is covered by fine sediment will likely result in greater erosion throughout the study area. Therefore, the greatest erosional efficiency on the Rio Grande will be achieved following periods when there have been many tributary-sourced flash floods, and large spatial increases in the fine-sediment supply within the channel. These findings are consistent with on-the-ground measurements of geomorphic change. Erosional efficiency will be reduced if other factors, such as the growth of vegetation, stabilize in-channel deposits. This study highlights the different trajectories of geomorphic evolution that may occur under different flow and sediment conditions on the Rio Grande in the Big Bend region.

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