Sediment monitoring in Galveston Bay – Final Phase

Michael C. Slattery
Institute for Environmental Studies
Texas Christian University
Fort Worth, TX 76129

and

Jonathan D. Phillips
Copperhead Road Geosciences, LLC
720 Bullock Place
Lexington, KY 40508
(also University of Kentucky)

FINAL REPORT

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INTRODUCTION AND OVERVIEW

This report addresses the fifth year of a corporative research study that seeks to document the effects of the Lake Livingston dam on downstream sedimentation processes, in particular the delivery of sediment to the lower Trinity River and the Trinity Bay estuary and, ultimately, Galveston bay. Progress in the first two years addressed the dam effects question. In years three and four, the focus turned to identifying the major sediment sources for the Trinity River delta and Trinity Bay. In this final year we planned to resolve issues raised by results obtained during the first four years of the project with respect to the effects of various human and natural controls on sediment transport and storage in the lower Trinity River. The specific objectives of the project were to:

1. Continue sediment monitoring with increasing focus on major tributaries Long King and Menard Creeks
2. Conduct additional sediment fingerprinting via magnetic susceptibility
3. Assess geomorphic changes over the past 50 years in Long King and Menard Creek
4. Evaluate the effects of channel slope, flow, and water withdrawals from the Trinity River on sediment transport capacity in the Lower River
5. Examine the constraints imposed by geologic history and controls on geomorphic changes and sediment fluxes in the lower Trinity, particularly ancestral valley morphology and bedrock controls of channels
6. Determine the role of Holocene sea level change in controlling sediment delivery to Trinity Bay and the Trinity River Delta.

The report is presented in five parts. The first, **Fluvial sediment delivery and human impact in a large coastal plain river: The case of the Trinity River, Texas**, documents work performed under Task 1 above and is in manuscript form to be submitted to *Earth Surface Processes and Landforms*. The second part, **Downstream trends in discharge, slope, and stream power in a lower coastal plain river**, addresses Task 4 above and is a paper in press in the *Journal of Hydrology*. The third part, **Antecedent Alluvial Morphology and Sea Level Controls on Form-Process Transitions Zones in the Lower Trinity River, Texas**, presents results that relate to Tasks 5 and 6, and is a paper in press in *River Research and Applications*. The fourth part, **Channel adjustments of tributary streams within the lower Trinity River basin, Texas,** is Ph.D. work conducted by Zach Musselman (a graduate student of Dr. Phillips); a paper to be submitted to *Geomorphology* from this work is presented here. The final part, **Sediment fingerprinting in the lower Trinity River, Texas**, is a commentary on work conducted in Year 4 and the difficulty we faced in doing this work in Year 5.

Part 1

Fluvial sediment delivery and human impact in a large coastal plain river: The case of the Trinity River, Texas

Introduction

The impact of dams on land-to-ocean sediment flux has been widely documented (e.g., Graf, 1999; Vörösmarty et al., 2003; Walling and Fang, 2003). A recent analysis by Syvitski et al. (2005) suggests that humans have simultaneously increased the sediment transport by global rivers through soil erosion (by 2.3 ± 0.6 billion metric tons per year), yet reduced the flux of sediment reaching the world's coasts (by 1.4 ± 0.3 billion metric tons per year) because of retention within reservoirs. Over 100 billion metric tons of sediment is now sequestered in reservoirs constructed largely within the past 50 years (Syvitski et al., 2005).

However, in fluvial systems where upper and lower basins are decoupled, in the sense of limited upper-basin sediment being transported to the river mouth, upstream impacts on sediment production and transport such as dams may not be evident at the river mouth, no matter how significant their effects upstream. In such systems, the locus of deposition is frequently not the ocean, estuary, or delta, but floodplains in and upstream of the fluvial-estuarine transition zone. Sediment delivery to these "upstream mouths" may be a more accurate reflection of river sediment fluxes to the coastal and marine environment. This pattern of upper- and lower-basin decoupling has been documented in several rivers of the U.S. south Atlantic Coastal Plain (Phillips, 1991, 1992a,b, 1993, 1995; Slattery et al., 2002) and in drainage basins in the Great Lakes region and in Australia (Beach, 1994; Brizga and Finlayson, 1994; Olive et al., 1994; Fryirs and Brierly, 1999). If sediment delivery from the upper basin is indeed small compared to lower-basin sediment sources, then geomorphic changes in the lower river are likely to be linked to controls within the lower basin as opposed to changes in sediment delivery from the upper basin, including those associated with sediment trapping behind dams.

Our focus in this paper is on the lower reaches of the Trinity River, Texas, and the response of the river's sediment delivery system to the construction of a 2.2 billion m^3 reservoir 175 river kilometers from the coast. The purpose of the study is to document pre- and post-dam sediment transport within this meandering, alluvial reach of the river. Here we show that the effects of sediment retention behind the dam, even in a massive reservoir controlling ~ 95 percent of the drainage area, are unnoticeable in the lowermost river reach. In this system, a sediment storage bottleneck has created an essential decoupling, such that changes in sediment regimes in the upper basin are simply not reflected in the lower river reaches. Because sediment and freshwater fluxes to the coastal zone are typically measured or estimated based on gaging stations well upstream of the coast, and upriver from such sediment bottlenecks, our work shows that fluvial sediment delivery to the coast can in many cases be substantially overestimated.

Study site and methods

The 46,100 km^2 Trinity River drainage basin heads in north Texas and drains to the Trinity Bay, part of the Galveston Bay system on the Gulf of Mexico (Figure 1). The Lower Trinity River basin, defined here as the drainage area downstream of Lake Livingston, has a humid subtropical climate, and a generally thick, continuous soil cover. Soils on stable upland sites are mainly Ultisols and Alfisols. Drainage area at Livingston Dam, which was completed in 1968 to form Lake Livingston, is 42,950 km^2. The primary purpose of the lake is water supply for Houston; it has no flood control function.

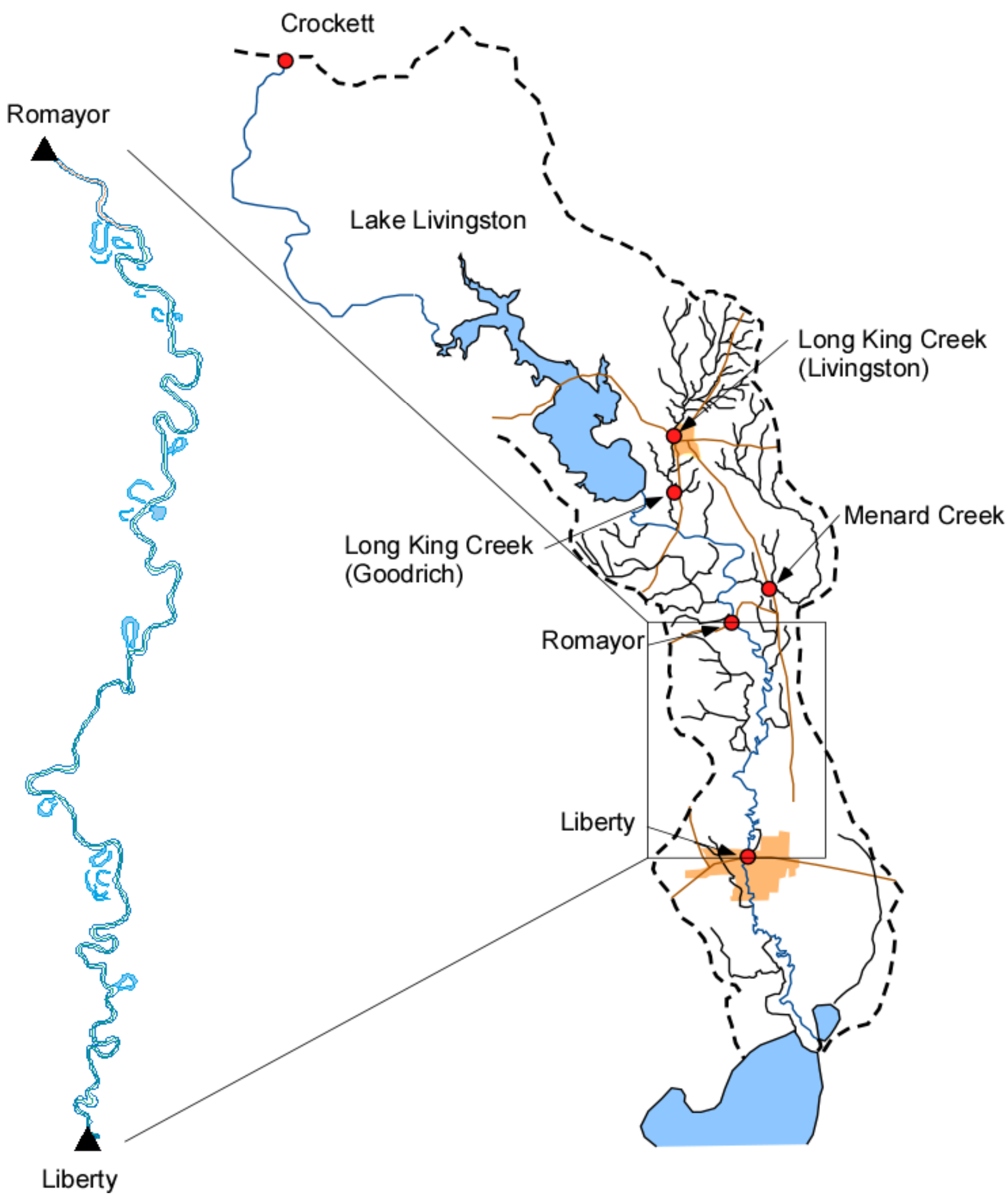

Figure 1. The lower Trinity River basin below Lake Livingston, Texas. USGS gaging station and sediment sampling sites are shown, along with a detailed planform view of the reach between Romayor and Liberty.

The Trinity River represents one of the rare cases where a significant suspended sediment record is available in the lowermost reaches of the river. The U.S. Geological Survey (USGS) and the Texas Water Development Board (TWDB) have collected suspended sediment samples at several sites from 1965 to 1989 (Figure 1). At the Romayor gaging station, 51 km downstream of the dam, the sediment record stretches back to 1936. We augmented this historic record through a field-sampling program from 2002-2006. Suspended sediment was sampled using a crane-mounted US D-74 depth-integrating sampler at Romayor as well as on the two major tributaries downstream of the dam, Long King Creek and Menard Creek (Figure 2). The
Long King Creek gaging site at Livingston has an upstream drainage area of 365 km^2, representing about 16% of the drainage area for the river downstream of the lake. We sampled directly at the USGS station on Highway 190, ~ 23 km from the mouth on the Trinity. This represents one of only two access points to the river from a bridge. Thus, sediment delivery to the Trinity is under-estimated at this site. We were able to sample at the second access point on Highway 1988 at Goodrich, but there is no long-term record of discharge here. We estimated discharge based on a cross-sectional survey of the channel, along with depth and surface velocity measurements during the storm events. On Menard Creek (which has a USGS gaging station but no historic sediment record), we sampled from the bridge on Highway 146, about 6 km from the Trinity confluence. The Menard Creek gaging station has an upstream drainage area of 394 km^2, representing about 17% of the drainage area for the river downstream of the lake.

Figure 2. Photograph showing the US D-74 depth-integrating sampler (taken on Long King Creek at Goodrich during the 17-18 November, 2004 storm).

In order to establish a longer-term, high resolution record of sediment transport in the Trinity, we installed a turbidity probe (YSI UPG-6000) at the Romayor site in April 2002. The probe was programmed to sample turbidity every six hours. The turbidity readings were then calibrated against depth-integrated samples taken over a range of flow conditions ($r^2 = 0.87$, $n = 26$) to give the suspended sediment-turbidity rating curve for the site. We found cross-sectional variability in suspended sediment concentrations to be less than 10 % at the gaging station giving us confidence in the turbidity record. We also sampled bed load at the sites using a Helley-Smith sampler.

Results

The historic sediment record for the Trinity River downstream of the dam at Romayor is shown in Figure 3. We chose two periods to compare pre- and post-dam sediment flux: 1959-1963 and 1977-1980. These time periods represent the most consecutive years in the record where the deviation of the annual flow duration curves were within 5 percent of each other. This gave us the ability to compare sediment transport under similar flow regimes before and after completion of the dam. The sediment rating curves for each time period are shown in Figure 4, along with our contemporary data derived from the turbidity record. The cumulative sediment record for Romayor is shown in Figure 5.

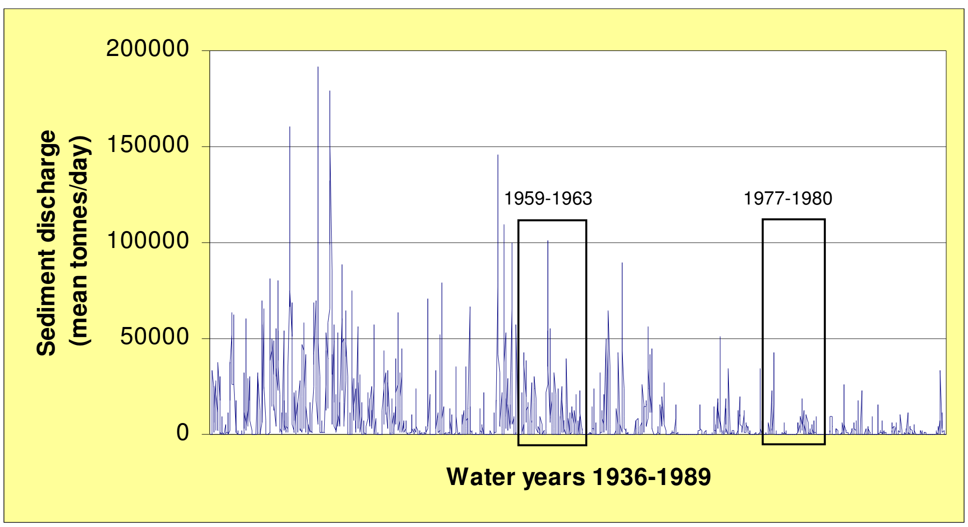

Figure 3. Historic sediment record at USGS station 08066500 (Romayor) from 1936-1989.

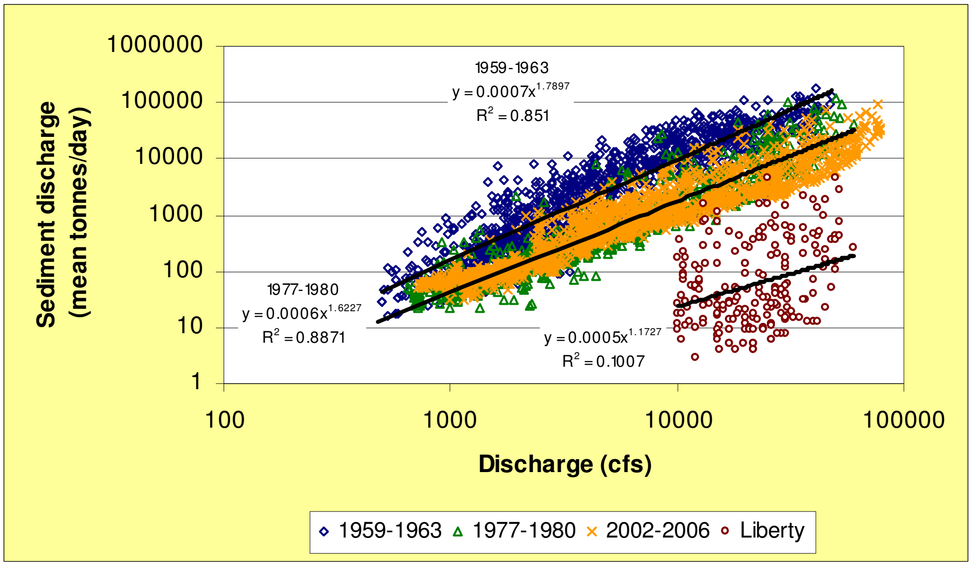

Figure 4. Sediment rating curves for USGS station 08066500 (Romayor) for periods 1959-1963, 1977-1980, and 2002-2006. Data for USGS station 08067000 (Liberty) for 1977-1980 are included.

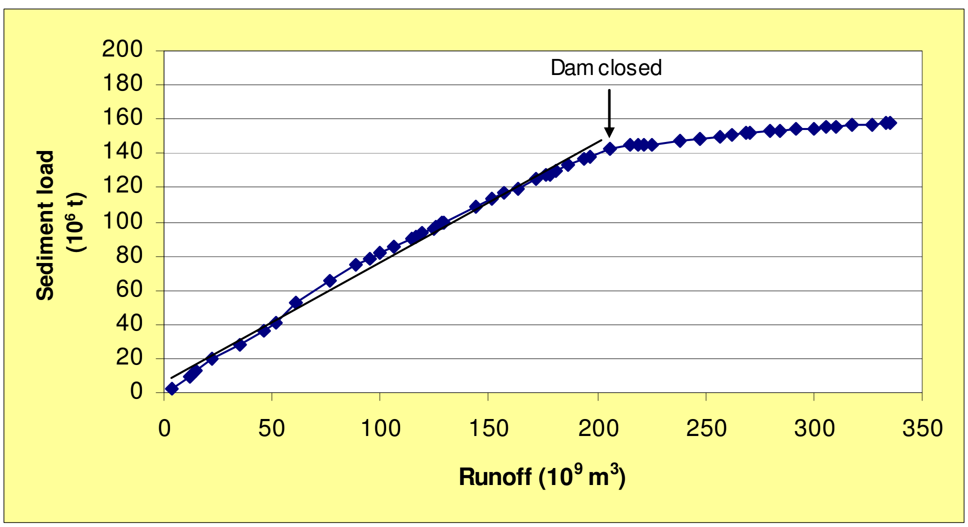

Figure 5. Cumulative sediment load as a function of cumulative runoff at USGS station 08066500 (Romayor) from 1936-1989.

The pre- and post-dam sediment rating curves (Figure 4) and the cumulative sediment record (Figure 5) for Romayor both show a clear decline in sediment transport following completion of Livingston Dam. Sediment loads have been reduced by between three to seven times following impoundment. The measured cumulative sediment load in 1989 was 65 percent less than the predicted load using a linear regression function fitted to the pre-dam data (Figure 5). Our contemporary data confirm the post-impoundment trend, plotting consistently within the 1977-1980 dataset (Figure 4). This suggests that sediment transport processes in the lower Trinity have remained unchanged during the

40 years following closure of the dam. However, sediment loads at Liberty, the lowermost station with any historic record, are consistently two orders of magnitude lower than those at Romayor. The very low sediment yields and concentrations at Liberty compared with those at Romayor suggest extensive alluvial storage between Romayor and Liberty, and that little sediment reaches the lower river at Liberty, with or without Lake Livingston.

The sediment rating curves for the two major tributaries on the lower Trinity, Long King Creek and Menard Creek, are shown in Figure 6, along with the post-dam record at Romayor. The flow duration curves for all three rivers are given in Figure 7. The suspended sediment data from Long King Creek and Menard Creek show that both tributaries transport more sediment to the Trinity than the Trinity transports itself *at equivalent discharges*. For example, at a discharge of 1000 cfs, sediment flux in the Trinity is ~ 45 tons/day, whilst in Menard Creek and Long King Creek, concurrent fluxes are ~ 244 tons/day and ~ 1,570 tons/day, respectively. Generally, across the range of flows, sediment discharge from Long King Creek is an order-of-magnitude greater than Menard Creek and two orders-of-magnitude greater than in the Trinity. Although the drainage areas upstream of both gaging stations on these tributaries are similar (Menard Creek = 394 km^2; Long King Creek = 365 km^2), the hydrologic regimes of the two rivers are significantly different (Figure 7). Long King Creek is a more dynamic and responsive river (i.e., hydrologically more "flashy") with steep rising limbs that generally crest 6 to 18 hours before Menard Creek under equivalent rainfall. Menard Creek's flow duration curve is flatter and low flows are elevated. This is consistent with the greater proportion of urban and agricultural land use in the Long King watershed, as opposed to the predominantly forested Menard watershed, much of which is within the Big Thicket National Preserve. Soils are deeper and more permeable in the Menard watershed, and runoff is slower and presumably dominated by subsurface pathways.

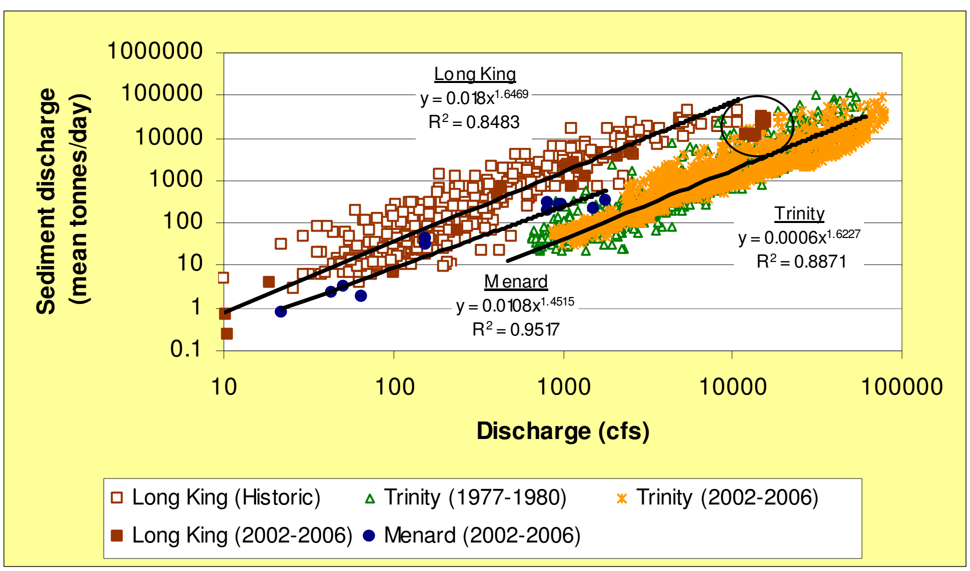

Figure 6. Sediment rating curves for USGS station 08066500 (Trinity at Romayor) for periods 1977-1980 and 2002-2006, USGS station 08066200 (Long King Creek) for 1973-1979 and 2002-2006, and USGS station 08066300 (Menard Creek) for 2002-2006.

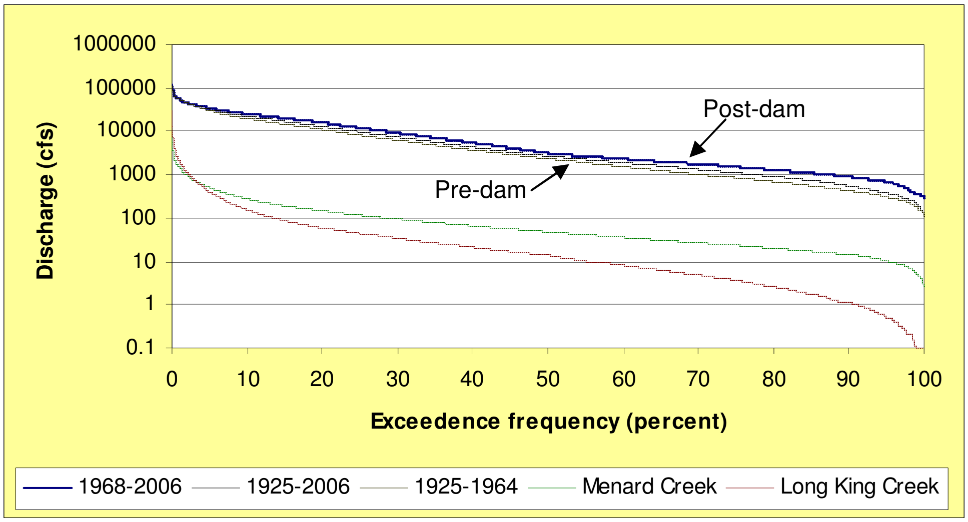

Figure 7. Flow durations curves for the Trinity River at Romayor (pre-dam 1925-1964; post-dam 1968-2006; complete record 1925-2006), Menard Creek (1971-2006) and Long King Creek (1971-2006).

The circled data points in Figure 6 warrant further discussion. These data represent samples taken on Long King Creek at extremely high flow during a storm on 17-18 November, 2004 (Figure 8 and photograph in Figure 2). The peak discharge recorded during this event was one of the highest on record, yet suspended sediment concentrations were lower than expected and under-predicted, according to the historic sediment rating curve. During this event, Long King Creek reached flood stage (19 feet; Q = 10,100 cfs) at 19:30 on 11/17, broke its banks shortly thereafter, and reached peak discharge 8 hours later at 15,700 cfs. The samples were taken right at peak discharge at 03:30. Because the channel had reached bankfull, we infer that sediment transport within Long King during this time was "transport-limited" and that these suspended concentrations represent an upper limit within this tributary basin. Using our data to extend the historic Long King rating curve, we were able to estimate a 24-hour storm-based sediment flux of ~ 74,200 tons. By comparison, sediment discharge at Romayor for the same 24-hour period was just 7,200 tons.

A substantial delta has developed at the mouth of Long King Creek (Figure 9). Tributary-mouth delta growth is not uncommon downstream of dams where flow in the main channel has been reduced, but this is not the case in the Trinity. While flow magnitudes have not been reduced, the lake has created asynchronicity in peak flows between Long King Creek and the Trinity. Peak sediment inputs from the creek are delivered before river discharge has increased enough to transport the coarser material. The delta is dominantly sandy, but characterized by alternating sand and clay layers. This suggests that at high flows backwater flooding, and perhaps recirculating eddies, occurs in the creek mouth, facilitating the fine-grained deposition. At normal and low water levels there is an obvious flow from the creek into the river and a turbidity plume.

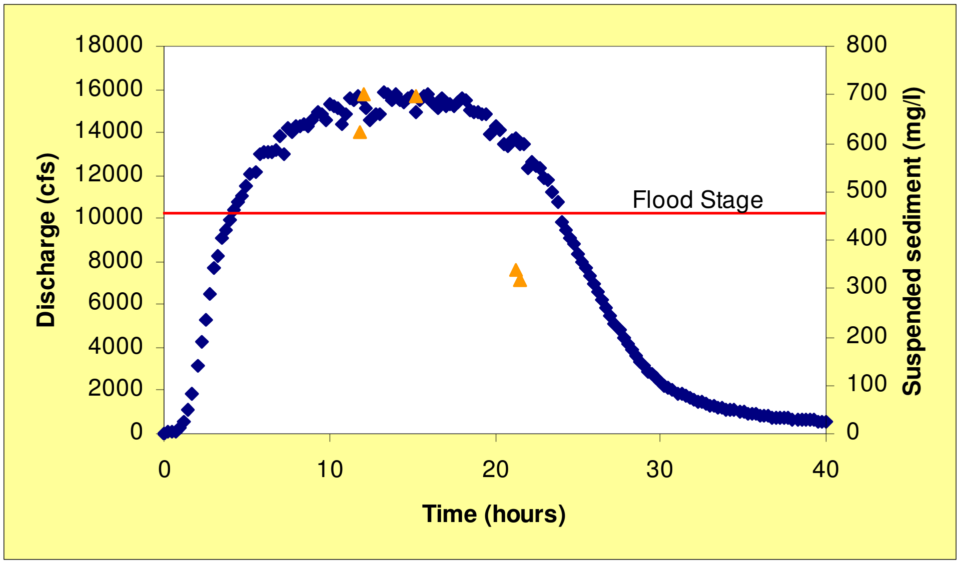

Figure 8. Discharge and suspended sediment (triangles) on Long King Creek for 17-18 November 2004 storm.

Figure 9. Coarse-grained delta at the mouth of Long King Creek. View looking downstream toward bridge at Goodrich.

Figure 10 shows the suspended sediment and bed load sediment rating curves for Romayor, Long King Creek, and Menard Creek for 2002-2006. Although there is ample evidence of bed load transport in the lower Trinity in the form of channel bars and floodplain sediment, the bed load data collected during this study suggest that this mode of sediment transport is much less important than suspended sediment transport. On both Long King Creek and Menard Creek, bed load is generally one to two orders-of-magnitude lower than suspended load. At Romayor, bed load is two to three orders-of-magnitude less than suspended load. However, we emphasize that bed load was much more difficult to sample than the suspended load, and we found substantial within-sample variability (see, for example, the data highlighted on Figure 10). There is considerable scatter in the bed load data across the range of discharges sampled, and none of the fitted regression curves were statistically significant. The only historic data we have for comparison are for Romayor where, on 12 occasions between 1972 and 1975 the USGS measured suspended and bed load on the same day. Bed load represented 1.4 percent to 21.4 percent of the total sediment load, with a mean of 9.7 percent. Thus, sediment transport estimates described in the revised sediment budget below, that are based on suspended measurements alone, were increased by 10%.

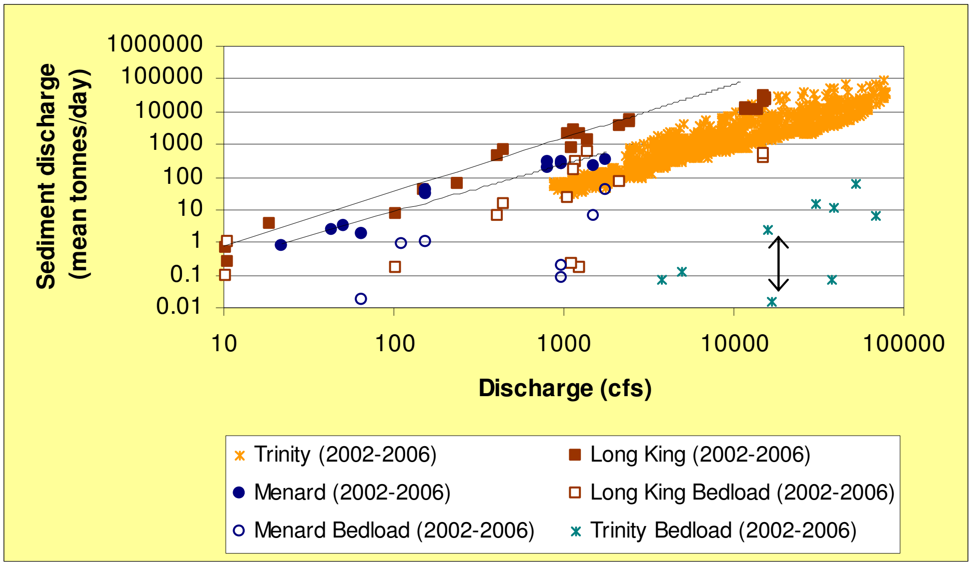

Figure 10. Suspended sediment (solid) and bed load sediment (open) rating curves for the Trinity at Romayor, Long King Creek, and Menard Creek for 2002-2006. Arrow indicates two samples collected within an hour of eachother.

The fluvial sediment budget for the Lower Trinity is shown in Figure 11. Originally published by Phillips et al. (2004), we include a revised version here because data from the 2002-2006 sediment monitoring program has allowed us to refine estimates of sediment delivery and transport within the lower Trinity.

Mean annual sediment yield at Romayor is 3.4 million tons/year, or about 1.7 million tons/year less than at Crockett upstream of the dam. Lake Livingston presumably accounts for much of the intervening storage. However, sediment yields at Romayor are almost 50 times those at Liberty, indicating that alluvial sediment storage is extensive in the lowermost reaches of the Trinity dwarfing sediment yield. In fact, there is more alluvial sediment is stored between Romayor and Liberty – that is, in the lower Coastal Plain portion of the river above tidal influences – than in Lake Livingston.

Mean annual sediment yield at the gaging station on Long King Creek at Livingston is ~ 170,500 tons/year, or 467 t/km^2 /year on a per area basis. If we extrapolate this sediment delivery rate to the entire Long King Creek basin (~ 508 km^2), then ~ 237,240 tons/year is delivered to the Trinity. This is considerably higher than sediment yield per unit area for any of the stations on the lower Trinity River. Mean annual sediment yield on Menard Creek is just 17,070 tons/year, or 43 t/km^2 /year. If we assume a conservative sediment loading of 50 t/km^2/year for the remainder of the drainage area at Romayor, downstream of the lake, then the total annual yield from all contributing sources along the 51 km reach between the dam and Romayor is 276,200 tons/year.

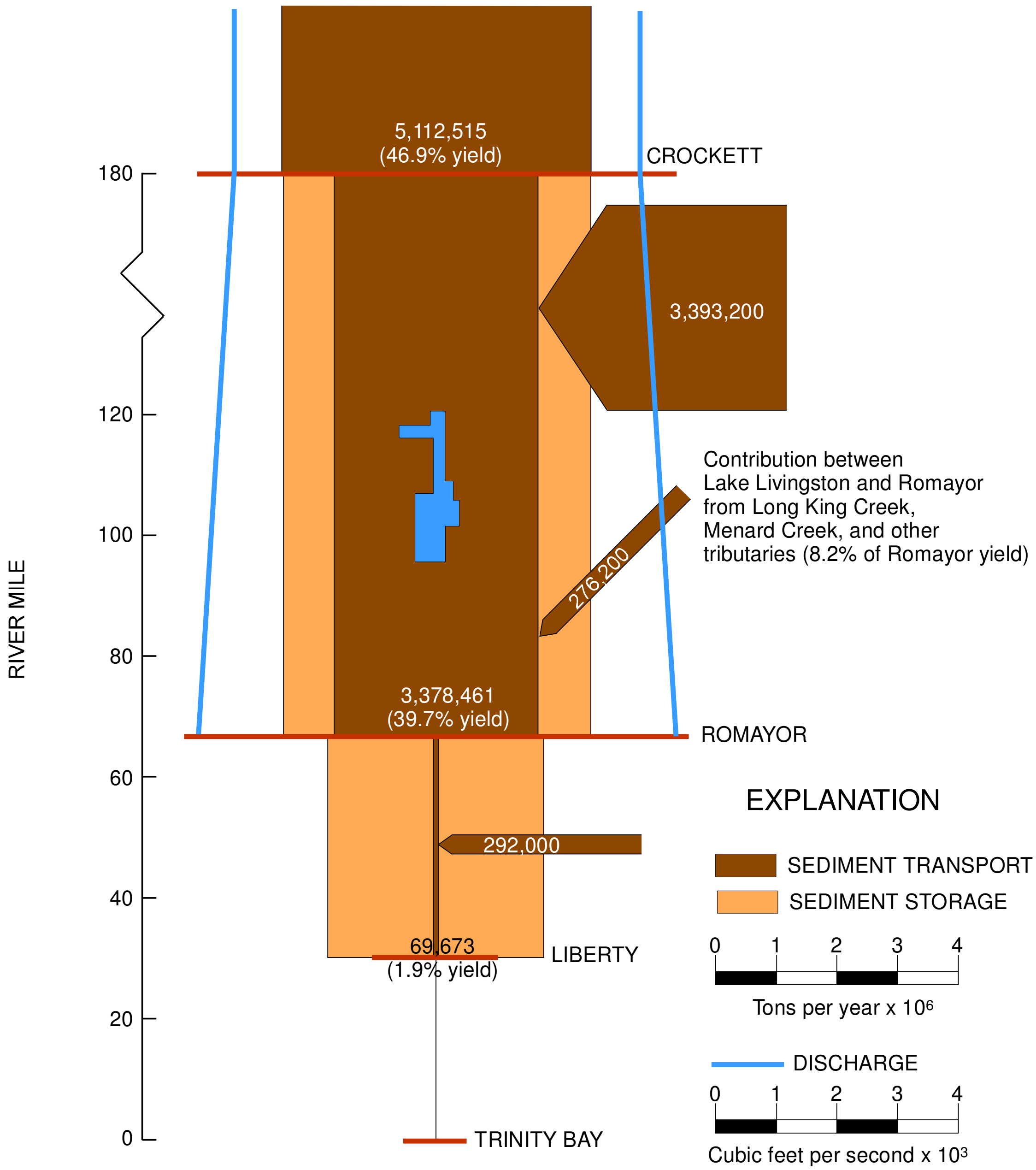

Figure 11. Fluvial sediment budget for the lower Trinity River.

Discussion and Implications

Because much of the upstream sediment load of the Trinity is captured in Lake Livingston, questions arise as to the source of sediments in the lower Trinity. The 276,200 tons/year sourced from the tributaries between the dam and Romayor represents only about eight percent of the sediment yield at the Romayor station. This implies that much of the sediment transported at Romayor comes from upstream of the dam – for example, is transported through the lake – or is derived from channel erosion downstream of the dam. We rule out the former hypothesis here based on observations of essentially clear water immediately downstream of the dam, even at high flows. Sediment concentrations in grab samples taken adjacent to the spillway never exceeded 100 mg/l. This 'hungry water' with unfilled sediment transport capacity has resulted in substantial lateral and vertical channel erosion downstream of the dam. We therefore conclude that the majority of the sediment transport at Romayor is derived from channel erosion between the dam and the gaging station along the 51 km reach. This interpretation is supported by results from an earlier study on channel change conducted on the Trinity below Lake Livingston which suggested contributions from channel erosion may exceed 50 percent (Wellmeyer et al., 2005). In this paper, the authors used historic aerial photographs from 1938 to 1995, digitized and imported into a GIS, to quantify long-term channel bank stability. Mean annual channel erosion was computed at 30.2 ha/year. Using the average channel depth of 7 m and a mean bulk density of 1.4 Mg/m^3 yielded a possible 2.96×10^6 Mg of sediment per year, which is equivalent to 87.6 percent of the annual sediment load measured at Romayor.

Sediment data from the Romayor station show a clear decline in sediment transport following completion of Livingston Dam. The reservoir is an effective trap for sediment, but the lower river is an even more effective sediment bottleneck, with a miniscule fraction of sediment produced within the basin delivered to the estuary. This situation has existed both pre- and post-dam. Fluvial sediment input to the gaging station at Romayor, about 51 km downstream of Lake Livingston and 126 km upstream of Trinity Bay, is about 3.4 million tons/year, with an additional 276,200 tons/year input from tributaries in the lower basin. Of this, only about 70,000 tons (or less) is transported to Trinity Bay, less than two percent of sediment delivered to the lowermost basin. This suggests that (i) sediment storage in the lower Trinity is greater than storage in Lake Livingston, and (ii) alluvial storage in the lower river is a bottleneck for sediment delivery to the coast, independently of the effects of upstream impoundment. The implication is that dams, however important they may be upstream, may have minimal impact on land-to-ocean sediment fluxes in such coastal plain systems.

The low sediment flux from the Trinity River system into Galveston Bay appears to be the result of the very low slopes, and correspondingly low stream power and transport capacity, within the reach. Unit stream power for normal flows (50 percent probability) was calculated at 2.508×10^{-4} for Romayor and 1.002×10^{-5} for Liberty (Phillips and Slattery, 2006). Increases in discharge between Romayor and Liberty are overwhelmed by the much reduced slope, so that stream power is at least an order of magnitude lower at Liberty than at Romayor. For higher flows, the difference is even more pronounced: stream power at Liberty is about two orders of magnitude lower than at Romayor. Beyond slope and stream power, accommodation space (a wider, lower floodplain) and greater frequency of overbank flow downstream are also import in promoting alluvial storage.

The upper and lower basins of the Trinity River are decoupled in the sense that very little upper-basin sediment is being transported to the river mouth. The implications of such alluvial buffering are quite profound. River sediment delivery to the coast is typically estimated based on monitoring stations that are significantly inland, above the storage bottlenecks, and not reflective of the low slope, low stream power reaches. In the Sabine, Neches, Trinity, Brazos, and Colorado Rivers on the Texas coastal plain, for example, the gaging stations used to measure or estimate sediment loading to the coast range from 54 to 98 km upstream of the river mouth. As illustrated by the case of the Trinity, sediment transport monitoring which does not represent the lower reaches of coastal plain alluvial rivers will result in overestimation of sediment flux to the sea in a contemporary sense, particularly for large alluvial rivers that discharge to passive margins.

While we are confident in our conclusions that relatively little sediment is delivered to the coast in many coastal plain rivers, and that low transport capacity plays a major role in this, there is clearly much to be done. Suspended sediment monitoring in lower coastal plain river sections is an obvious need, along with more studies of sediment sources, transport, and storage in lower river reaches. This is a major challenge. Sediment transport measurements and studies of fluvial sediment systems in coastal plain rivers demands dealing with large channels and large drainage basins. These are difficult logistical as well as conceptual tasks. And while small basins are generally more responsive to environmental change and generally easier to work with, the huge quantities of area, sediment, water, and other mass represented or transported and stored by coastal plain rivers demands that we engage them.

Conclusions

Livingston Dam has greatly reduced sediment input to the lower Trinity River. Clear, 'hungry' water immediately downstream of the dam with unfilled sediment transport capacity has resulted in channel incision and widening, a response that is limited to about 60 km downstream of the dam. At Romayor, sediment loads have recovered to *c.* 3.4 million tons/year, or about 40 percent of the sediment load entering the reservoir. The majority of the sediment in this reach is derived from channel scour and bank erosion, with tributaries accounting for less than 10 percent of the sediment delivered to the main channel. However, the reservoir has had essentially no effect in terms of sediment delivery to Trinity Bay, about 175 km downstream of the dam. The *c.* 125 km reach between Romayor and the Trinity delta is dominated by alluvial sediment storage. Yields at Liberty are only two percent of the load at Romayor, confirming that alluvial sediment storage in the lowermost reaches dwarfs sediment yield. Even after dam construction, sediment supply in the lower Trinity still exceeds transport capacity. Sediment storage is so extensive that the upper Trinity basin and the lowermost river reaches were essentially decoupled (in the sense that very little upper-basin sediment reached the lower river) even before the dam was constructed. This alluvial storage in the lower Trinity essentially buffers Trinity Bay from the effects of fluctuations in upstream fluvial sediment dynamics.

Acknowledgements

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Part 2

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Downstream trends in discharge, slope, and stream power in a lower coastal plain river

Jonathan D. Phillips [a,*], Michael C. Slattery [b]

[a] Tobacco Road Research Team, Department of Geography, University of Kentucky, Lexington, KY 40506-0027, USA
[b] Institute for Environmental Studies and Department of Geology, Texas Christian University, Fort Worth, TX 76129, USA

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Coastal plain river;
Trinity river

Summary Conceptual models of river-estuary interaction are typically based on a notion of systematic downstream change in the intensity of fluvial processes. Low slopes, backwater effects, and effects of antecedent topography and landforms may complicate downstream tends in water and sediment flux in coastal plain rivers. An analysis of the lower Trinity River, Texas shows no consistent downstream pattern of increases or decreases in the discharge, stream power, or water surface slope. Flows may decrease downstream due to coastal backwater effects in the lowermost reaches, and due to diversion of flow into valley-bottom depressions during high flows in both the fluvial and fluvial-estuarine transition reaches. In general, however, stream power and slope decrease in the lower reaches, consistent with earlier findings of limited fluvial sediment delivery to the coastal zone. Some tributaries may become distributaries at high but sub-bankfull flows, as backwater effects reverse flows into depressions associated with paleomeanders. The paleomeanders, and possibly the locations of these ''reversible'' channels, are related to antecedent topography associated with aggradation/degradation cycles over the past 100 Ka. Low-gradient coastal plain rivers may not function as simple conduits from land to sea. Further, the transition from fluvial to coastal dominance may be variable along the river, with the variability controlled not just by the relative magnitude of river and tidal or backwater forcing, but also by valley topography controlled in part by antecedent landforms.

Introduction

The conveyance of water and sediment from rivers to the sea is deceptively complex. As rivers approach the coast, particularly those crossing extensive coastal plains, fluvial dynamics change as channel and valley slopes decline and alluvial accomodation space increases. In the fluvial-estuarine transition zones backwater effects and lunar and wind tides influence water and sediment fluxes. Over longer timescales, these lower coastal plain reaches are also profoundly influenced by Quaternary and contemporary sea

* Corresponding author.
E-mail address: jdp@uky.edu (J.D. Phillips).

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level changes. A common conceptual model of hydrodynamics applied to, e.g., stratigraphic facies models, is based on the interplay of coastal/marine processes, which generally decrease in intensity inland, and fluvial processes, which decline in strength downstream (Cattaneo and Steel, 2003; Dalrymple et al., 1992). However, a number of studies in fluvial systems show that there may not be consistent downstream trends in factors such as stream power (Graf, 1983; Jain et al., 2006; Knighton, 1999; Lecce, 1997; Magilligan, 1992; Reinfields et al., 2004). The goal of this project is to examine downstream changes in stream power and the determinants of the latter, discharge and slope, in the lower Trinity River in the southeast Texas coastal plain. Process linkages between hydrology, geomorphology and ecology in coastal plain rivers remain largely undocumented (Hupp, 2000). This work seeks to help fill that gap, for the particularly problematic lower coastal plain.

River discharge is an important determinant of estuarine circulation, water chemistry, and flushing or residence time, and is thus critical with respect to water quality, estuarine ecology and fisheries (Longley, 1994; Powell et al., 2003). Fluvial discharge and sediment fluxes are typically measured a considerable distance upstream from the coast. Variations in discharge occurring downstream of these gaging stations will thus not be reflected in these records. Because these gaging stations are often upriver from lower coastal plain sediment bottlenecks in some cases fluvial sediment delivery to the coast has been substantially overestimated (Brizga and Finlayson, 1994; Olive et al., 1994; Phillips, 1993, 1997; Phillips et al., 2004; Phillips and Slattery, 2006).

In tidally dominated river estuaries there may be a relatively straightforward downstream progression from fluvial to tidal domination, reflected in landforms, sedimentary environments, and hydrodynamic zones (e.g. Renwick and Ashley, 1984), the latter obviously varying with river discharge and tidal cycles. The transition from fluvial to coastal dominance may be considerably more complicated and subtle in wave- and wind-dominated estuaries such as the Trinity Bay/Galveston Bay system considered in this study (Nichols, 1989; Phillips and Slattery, 2006; Wells and Kim, 1987).

In addition to the systematic changes in channel and valley morphology, slope, and the relative importance of fluvial vs. coastal processes, recent field experience on the lower Trinity River in southeast Texas suggested that the downstream changes in flow and sediment transport capacity might be even further complicated by factors such as inherited valley morphology, extensive water storage on floodplains, and low-water tributaries that might function as distributaries at high flow. This paper investigates the downstream trends in discharge, slope, and stream power in the lower Trinity River. The study area was selected in part due to past and ongoing geomorphological studies in the area, but the Trinity is advantageous for this study in having a number of gaging stations in the lower fluvial reaches and fluvial-estuarine transition zone. The specific environmental settings, land and water use and management, sea level histories, and other controls vary between rivers, but in a broad general sense the Trinity is not atypical of rivers on the US. Atlantic and Gulf Coastal Plains.

Background

Stream power

In humid-region perennial streams such as the Trinity River, channel, valley, and energy grade slopes typically decline, on average, as base level is approached, as illustrated by the typically concave-upward longitudinal profile. Discharge generally increases downstream, often as a step function reflecting tributary inputs. Cross-sectional stream power (power per unit channel length; terminology follows Rhoads, 1987) is a function of the product of slope (S) and discharge (Q):

$$\Omega = \gamma QS, \tag{1}$$

where γ is specific gravity.

Stream power does not necessarily increase systematically downstream because of the conflicting changes in discharge and slope, and local variations in width, depth, roughness, and other factors that may influence Q and S (Graf, 1983; Magilligan, 1992; Knighton, 1999). Nonlinear downstream changes in stream power were documented by Lecce (1997), who showed power peaking where drainage areas were 10–100 km^2 (in a 208 km^2 Wisconsin drainage basin) and decreasing rapidly downstream. The relative rates of change in discharge and slope determine the location of the Ω maximum, which in Knighton's (1999) model, applied to the Trent River, England, occurred at location intermediate between headwaters and lower reaches.

Downstream variations in stream power were assessed from digital elevation models (DEMs) in a small, steep Australian watershed by Reinfields et al. (2004), where longitudinal profiles were concave up, and channel gradients generally decreased downstream, with some localized variations. Channels with steep convex sections had locally steeper gradients in mid-profile, but still displayed lower gradients in the lower as compared to upstream reaches (Reinfields et al., 2004). Cross-sectional stream power exhibited no monotonic downstream trend, and in four study rivers was both higher, lower, and approximately the same in the lower as in the upper reaches. Specific stream power was uniformly lower in the downstream reaches, but also varied irregularly in the longitudinal direction. Jain et al. (2006) also used a DEM-based model for the upper Hunter River watershed, Australia, finding that stream power variations in headwaters were controlled mainly by discharge, while in the mid and lower reaches local variations in slope were the primary controls. Those results, and the theoretical models applied, showed irregular downstream trends in power, but with generally smoother, downward trends in the lowermost reaches (Jain et al., 2006).

In the lower Trinity River, Texas, Ω was found to be substantially reduced between upstream and downstream gaging stations at flood, bankfull, and near-bankfull flows (Phillips et al., 2005; Phillips and Slattery, 2006). This was attributed primarily to declines in slope (based on channel bed slope), though lower banks downstream and thus a tendency to reach bankfull at lower discharges also played a role.

In Magilligan's (1992) and Lecce's (1997) studies, discharge increased downstream, and in the other studies in perennial streams (Jain et al., 2006; Knighton, 1999;

Reinfields et al., 2004) discharge was assumed to increase downstream as a function of drainage area and/or total stream length. The latter is common and widely accepted, and Phillips et al. (2005) accordingly dismissed the apparent downstream decrease in bankfull flow in the lower Trinity as a function of bank height. Subsequent analysis of Trinity River flows during sub-bankfull events, however, suggested that a general downstream increase in discharge between gaging stations cannot necessarily be assumed.

While Lecce (1997) and Magilligan (1992) based their analyses on measured or modeled water surface slopes between stations, Phillips et al. (2005) used surveyed channel bed slope, and Jain et al. (2006), Knighton (1999), and Reinfields et al. (2004) assumed that downstream changes in energy grade slope reflect changes in channel slope. Magilligan (1988) showed that water surface slopes are a better approximation of energy grade slopes than either field-measured or map-derived channel bed slopes.

In some previous studies lithological control has been identified as a key determinant of factors such as valley width and valley slopes, which in turn help determine stream power (Graf, 1983; Lecce, 1997; Magilligan, 1992). Lithological controls are generally not thought to be strong, or even relevant, in coastal plain alluvial rivers such as the lower Trinity, however, where resistant, confining materials are rare. However, the Trinity (in common with other rivers of the region) has experienced a series of climate- and sea level-driven cycles of aggradation and degradation, such that inherited valley morphologies influence the contemporary river (Blum et al., 1995; Blum and Törnqvist, 2000; Morton et al., 1996; Rodriguez et al., 2005). While structural and lithological control in the usual sense is subtle at best in the lower Trinity River, antecedent topography may play a significant role in downstream variations in discharge, slope, and stream power.

Abrupt changes in the downstream trends of slope and stream power may represent critical transition points and foci of change (Reinfields et al., 2004). A critical transition zone has already been identified in the lower Trinity with respect to sediment transport and storage, and channel cross-sectional change (Phillips et al., 2004, 2005).

The mouth of the river

Generalizations about downstream changes in discharge and other hydrologic and hydraulic parameters are generally at least implicitly understood to apply to the portion of the river network which is both fluvially dominated (vs. influenced by coastal processes) and convergent. Convergent networks are dominated by net tributary inflows, while divergent networks are dominantly distributary, with net flow from the trunk stream into the tributary.

The seaward mouth of a river can be defined as the point at which a well-defined dominant channel can no longer be identified, at an open-water estuary or a delta apex. The mouth might also be defined as the point at which the dominant flow pattern becomes divergent or distributary rather than convergent. These points often do not coincide with the point at which channels are cut to below sea level, or with common upstream limits of backwater effects or salt wedges. Further, these ''mouths'' have been found to over channel distances of 50 to >100 km (Pierce and Nichols, 1986; Nichols et al., 1991; Phillips and Slattery, 2006).

In the Trinity, the transition to a distributary network occurs about 20 km upstream of the point at which the Trinity River enters Trinity Bay. Tidal influence is evident at the gaging station at Liberty, Texas, 85 km upstream, and the channel is cut to below sea level 110 km upstream. The lower coastal plain sediment storage bottleneck identified by Phillips et al. (2004) occurs about 130 km upstream of Trinity Bay. This suggests that downstream changes could be considerably more complex than a steady downstream increase in discharge and decrease in slope, followed by a gradual transition from fluvial to coastal dominance.

These issues are not only important for determination of sediment and water fluxes to the coast. The lower coastal plain reaches of rivers also typically contain large areas of ecologically and economically valuable wetlands such as bottomland hardwood forests, and both natural environments and anthropic features which are quite vulnerable to river floods, coastal storms, sea level change, subsidence, and other coastal plain dynamics.

Study area

The 46,100 km^2 Trinity River drainage basin, Texas, drains to the Trinity Bay, part of the Galveston Bay system on the Gulf of Mexico. The lower Trinity River basin (Fig. 1) has a humid subtropical climate and a generally thick, continuous soil and regolith cover. Most of the drainage area (95%) lies upstream of Livingston Dam, which was completed in 1968 to form Lake Livingston. The lake, a water supply reservoir for the city of Houston, has a conservation pool capacity of >2.2 billion m^3. The dam has no flood control function and Livingston is essentially a flow-through reservoir.

The contemporary and recent historical sediment budget, channel planform change, and changes in cross-sectional channel morphology between Lake Livingston and Trinity Bay have been analyzed elsewhere (Phillips et al., 2004, 2005; Wellmeyer et al., 2005). The alluvial morphology and stratigraphy of the lower Trinity (and the nearby and similar Sabine River) and the deposits and palaeochannels now submerged in Trinity and Galveston Bays and the Gulf of Mexico preserve evidence of climate, sea level, and upstream sediment delivery changes (Anderson et al., 1992; Thomas and Anderson, 1994; Blum et al., 1995; Anderson and Rodriguez, 2000; Rodriguez and Anderson, 2000; Rodriguez et al., 2001; Phillips, 2003; Phillips and Musselman, 2003). Therefore, contemporary modifications to flow and sediment regimes are superimposed on long-term changes controlled primarily by climate and sea level change.

The Trinity/Galveston Bay has a mean volume estimated at about 2.7 billion m^3 and drainage area of 85,470 km^2. About 54 percent of the drainage area, and of the freshwater inflow, is accounted for the Trinity River. Though Lake Livingston's capacity is more than 80% of that of Galveston Bay, analysis of pre- and post-dam discharge records at Romayor found no significant post-dam decrease in flow, and limited discharge change of any kind (Wellmeyer et al., 2005).

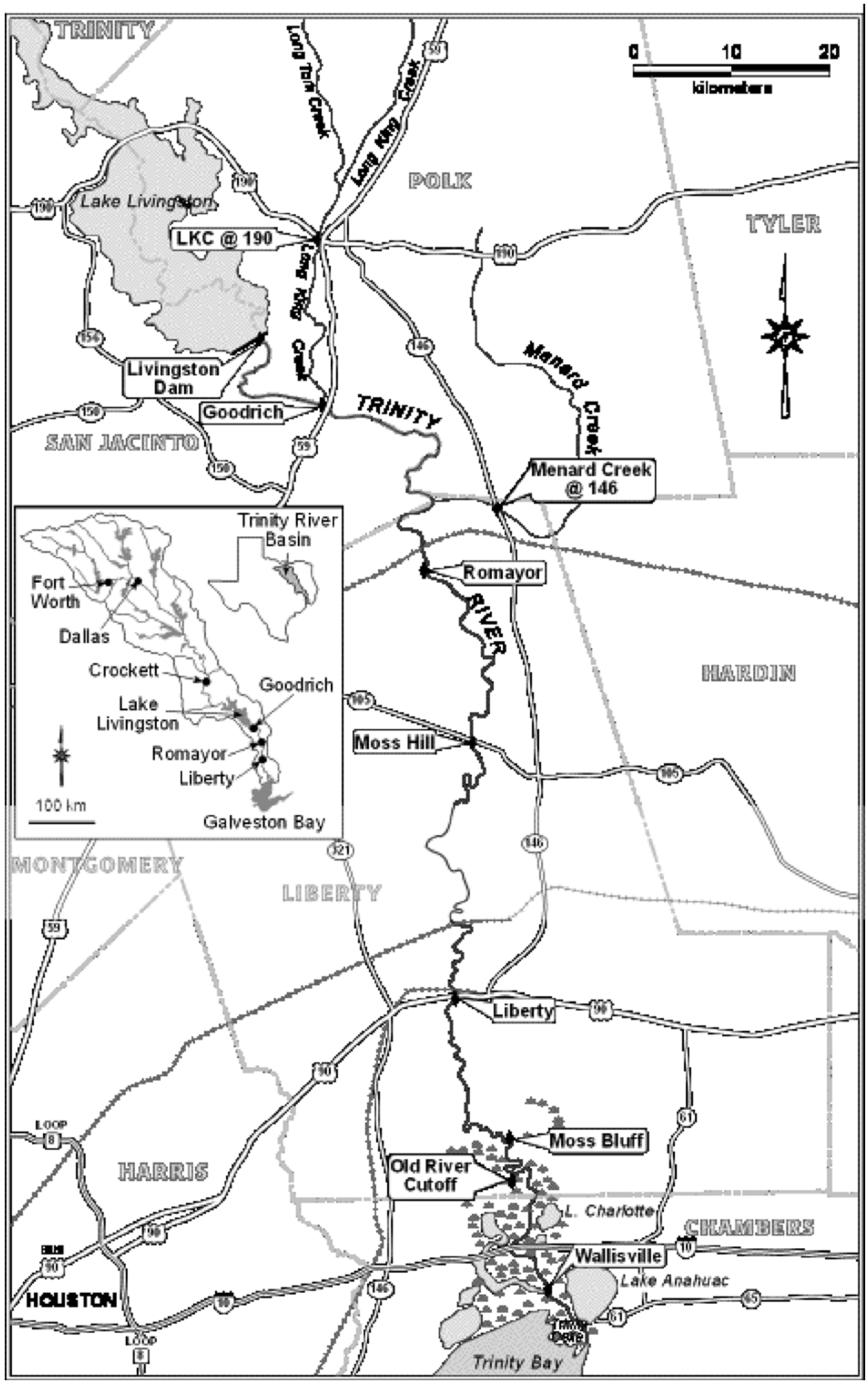

Figure 1 Study area, showing gaging stations and field sites.

Hydrodynamics of the Galveston Bay estuary have been studied in some detail (e.g. Powell et al., 2003), in part driven by concerns over potential effects of changes in freshwater inflow due to water diversions on salinity, water quality, and estuarine ecology. Work thus far has been focused almost entirely on the estuary, and driven chiefly by concern with fisheries production (GBFIG, 2003; Longley, 1994). The lower Trinity River has not been included in these studies, and is treated only as an input to Galveston Bay hydrology. Water diversions represent less than 10% of the mean discharge of the lower Trinity River, and a considerably lower proportion of high flows.

The details of sea-level history and coastal evolution in Texas are controversial (Blum et al., 2002), but most sources agree that Galveston Bay in its more-or-less modern position was formed about 4000 years ago. During lower

Quaternary sea level stands, the Trinity and Sabine Rivers converged on the continental shelf and cut an incised valley. From about 18,000 years BP to the present, the Trinity-Sabine incised valley has backfilled (Blum et al., 1995, 2002).

Methods

Data from nine gaging stations between Lake Livingston and Trinity Bay was used for this study. Five are Trinity River stations, and two (Long King and Menard Creeks) are stations on the two largest tributaries to the Trinity downstream of Livingston Dam. One station records changes in surface elevation and storage in Lake Livingston, and another (Old River cutoff) is a short distance from the river on a distributary channel in the Trinity River delta area. Station locations are shown in Fig. 1, and described in Table 1.

Discharge regime

For the three river stations with a sufficient period of record (Goodrich, Romayor, and Liberty), a number of reference flows were calculated using the standard formula

$$P = m/(n+1) \text{ or } T = (n+1)/m, \tag{2}$$

where m is the rank of the flow in the series and n is the total length of the series. Daily mean flows (reported in $ft^3\ s^{-1}$) were used to calculate P, the probability of exceedence, and T, the return period or recurrence interval. Reference flows include those associated with 50%, 10%, and 1% probability of exceedence by mean daily flows, and annual peaks with recurrence intervals of 1, 2, and 10 years. In addition, the mean annual discharge was determined from the entire available record of mean daily discharge. Finally, peaks were determined for the October, 1994 flood, which is the flood of record in the lower Trinity River, and a smaller flood in November, 2002.

The upstream—downstream trends in these reference flows was examined based on direct comparisons and differences between downstream and upstream stations (Liberty—Romayor; Romayor—Goodrich).

Hurricane Rita event

Hurricane Rita struck the southeast Texas coast and areas of adjacent Louisiana in late September, 2005. The eye of the storm and the most intense rainfall passed to the east of the Trinity River valley, but there was extensive precipitation in the lower Trinity Basin. Furthermore, wind-wave related damage to the earthen and stone Livingston Dam forced the Trinity River authority to make a rapid release to lower lake levels for damage assessments and repairs. The event therefore provided an opportunity to determine response to a dam release and precipitation confined chiefly to the lower basin, as opposed to being transmitted through the lake.

Based on the hydrograph responses to this event, water surface elevations and flow responses (discharge and/or stage) were determined for times corresponding with the start of the rising hydrograph limbs of Long King Creek and the Trinity at Goodrich and Romayor, the peak elevation and beginning of drawdown of the lake, the completion of the lake drawdown, and the flow peaks at Long King Creek, and Goodrich, Romayor, Liberty, and Moss Bluff. Long King Creek is taken as representative of the local, lower-basin runoff and tributary input, as opposed to releases from Lake Livingston.

Stage elevations at these times, coded as R1 through R9, were combined with gage datums to determine instantaneous water surface elevations. These were combined with distances between stations measured from 10-m resolution DEMs to determine water surface slopes.

The National Oceanic and Atmospheric Administration data buoy at Morgan's Point on upper Galveston Bay was used (via barometric pressure records) to pinpoint the arrival of the storm in the lower Trinity valley.

Table 1 Lower Trinity River (TR) gaging stations and year of establishment

Name	Location	Number	Measurements
Livingston Reservoir (1969)	177	0866190	H, storage
Long King Creek at Livingston (1963)	145[a]	0866200	H, Q
Menard Creek at Rye (1963)	130[a]	0866300	H, Q
Trinity River (TR) nr Goodrich (1965)	144	0866250	H, Q
TR at Romayor (1924)	126	0866500	H, Q
TR at Liberty (1940)	83	0867000	H, Q[b]
TR at Moss Bluff (1959)	32.5	0867100	H, Q[c]
Old River cutoff near Moss Bluff (2003)	30[a]	0867215	H, velocity
TR at Wallisville (2003)	6.5	0867252	H

Location refers to distance upstream from Trinity Bay, in kilometers. Number is the US. Geological Survey station number. Measurements of interest here include discharge (Q) and stage or gage height (H). All are operated by the US. Geological Survey except Livingston Reservoir (Trinity River Authority) and Old River and Wallisville (US Army Corp. of Engineers).

[a] Approximate distance from the bay of creek/river confluence.

[b] Discharge measurements discontinuous.

[c] Discharge estimated from stage by National Oceanic and Atmospheric Administration, West Gulf River Forecast Center (http://www.srh.noaa.gov/wgrfc/statlist.php?funct=obs&shefid=MFT2).

Valley topography

Topography of the lower Trinity Valley was analyzed based on 10-m resolution DEMs from the USGS National Elevation Dataset (NED) obtained via the USGS seamless data distribution center. The RiverTools software was used for visualizations of the topography, to construct elevation profiles and to evaluate topographically controlled flow directions. The flow analysis was based on the imposed gradient method of Garbrecht and Martz (1997). The algorithm used arbitrarily fills local pixel-scale depressions, so any broader depressions attracting flow were taken to be real rather than data artefacts. Digital orthophotquads (DOQQs) at 1- and 2.5-m resolutions, many taken during high water conditions associated with the 1994 flood, were also used to identify key geomorphic features. DOQQs and fieldwork confirmed that larger depressions in the DEM are present on the ground.

Field observations

The field area was visited in early November, 2005. No further overbank flows occurred between the Rita event and this fieldwork. Flow indicators (flood debris and deposits) were examined at seven locations between the mouth of Long King Creek and the Wallisville station. In addition, field surveys were conducted at the junction of Pickett's Bayou and the Trinity River, a short distance upstream of Moss Bluff. This stream connects the Trinity River with Old River, one of the delta distributaries. It was unclear from maps and photography the extent to which the bayou functions as a tributary or distributary.

Results

Discharge regime

Reference flows are shown in Table 2. Mean annual flows and some relatively frequently exceeded events show slightly higher values at Romayor, as expected, than at Goodrich 18 km upstream. However, for six of the nine reference events discharge is higher at Goodrich than downstream at Romayor. This is despite the fact that two major tributaries (Menard and Big Creeks) join the Trinity between the two stations.

By contrast, every reference discharge for Liberty except the 2002 flood is higher – often substantially so – than at either of the upstream stations. The gage datum at Liberty is 0.67 m below sea level, and the thalweg elevation when measured in early 2003 was −5 m (Phillips et al., 2005). The gage also often shows tidal influences. Tidal and backwater effects influence the stage/discharge relationship so that discharges are not estimated or published continuously. This may bias the published data toward river flow domination and thus inflate the mean annual flow and relatively frequent discharges (50% and 10% daily exceedence, and Q1). In the two specific high flow events, the peak for the 2002 flood was lower than for the upstream stations, and for the 1994 flood only eight percent greater.

Overall, the data in Table 2 indicate that there is not necessarily a consistent downstream increase in discharge, even within the always fluvially dominated Goodrich-Romayor reach.

Peak flow differences (downstream station minus upstream station) for the annual peaks are shown for Romayor–Goodrich and Liberty–Romayor for the period of overlapping records in Fig. 2. In most cases, peaks were apparently associated with the same flow event, as indicated by peaks occurring within 5 days or less of each other at adjacent stations. Negative values indicate that the peak flow for the downstream station was lower than for the upstream. In some cases, these could be associated with downstream flood wave attenuation for events dominated by releases from Lake Livingston. This could account for the increasing range of differences observed after 1968. However, negative differences are approximately equally common in the pre- and post-dam records.

Table 2 Reference flows for lower Trinity River gaging stations, in $m^3\ s^{-1}$

Reference flow	Goodrich	Romayor	Liberty
MAQ	231	246	509
50% exceedence	82	77	433
10% exceedence	677	640	1048
1% exceedence	1550	1541	822
Q1	2130	1970	2484
Q2	2400	2330	2835
Q10	3002	2925	3600
2002 flood	1872	2198	1602
1994 flood	3540	3455	3823

MAQ, mean annual discharge. Exceedence flows indicate the mean daily flow exceeded the indicated percentage of days. Q1, Q2, Q10 are peak flows with estimated recurrence intervals of 1, 2, and 10 years. The 2002 and 1994 floods are the maximum flow peaks.

Hurricane Rita discharge

The Morgan's Point station recorded its lowest pressure (983 mb) at 0900 on September 24, 2005 when the eye of Hurricane Rita passed closest to Trinity Bay. Clouds and rain bands preceded the eye of the storm. Though the most intense precipitation fell east of the Trinity River watershed, significant rainfall was recorded for September 23–24 at several locations in the region. The most at nearby stations was >170 mm at Beaumont, about 70 km east of the Trinity River. No meteorological stations within the lower Trinity basin directly recorded precipitation for this event, but 24-h precipitation estimates from the Lake Charles, Louisiana National Weather Service Radar indicated 25–100 mm in the lower Trinity basin.

Long King and Menard Creeks experienced steep rises in the hydrograph. Long King Creek showed an equally steep recession, whereas Menard Creek flow remained elevated for several days. This is consistent with the greater proportion of urban and agricultural land use in the Long King watershed, as opposed to the predominantly forested Menard watershed, much of which is within the Big Thicket National Preserve. The creeks began rising at about 0430 September 24 (Table 3), though the hydrograph had begun rising at

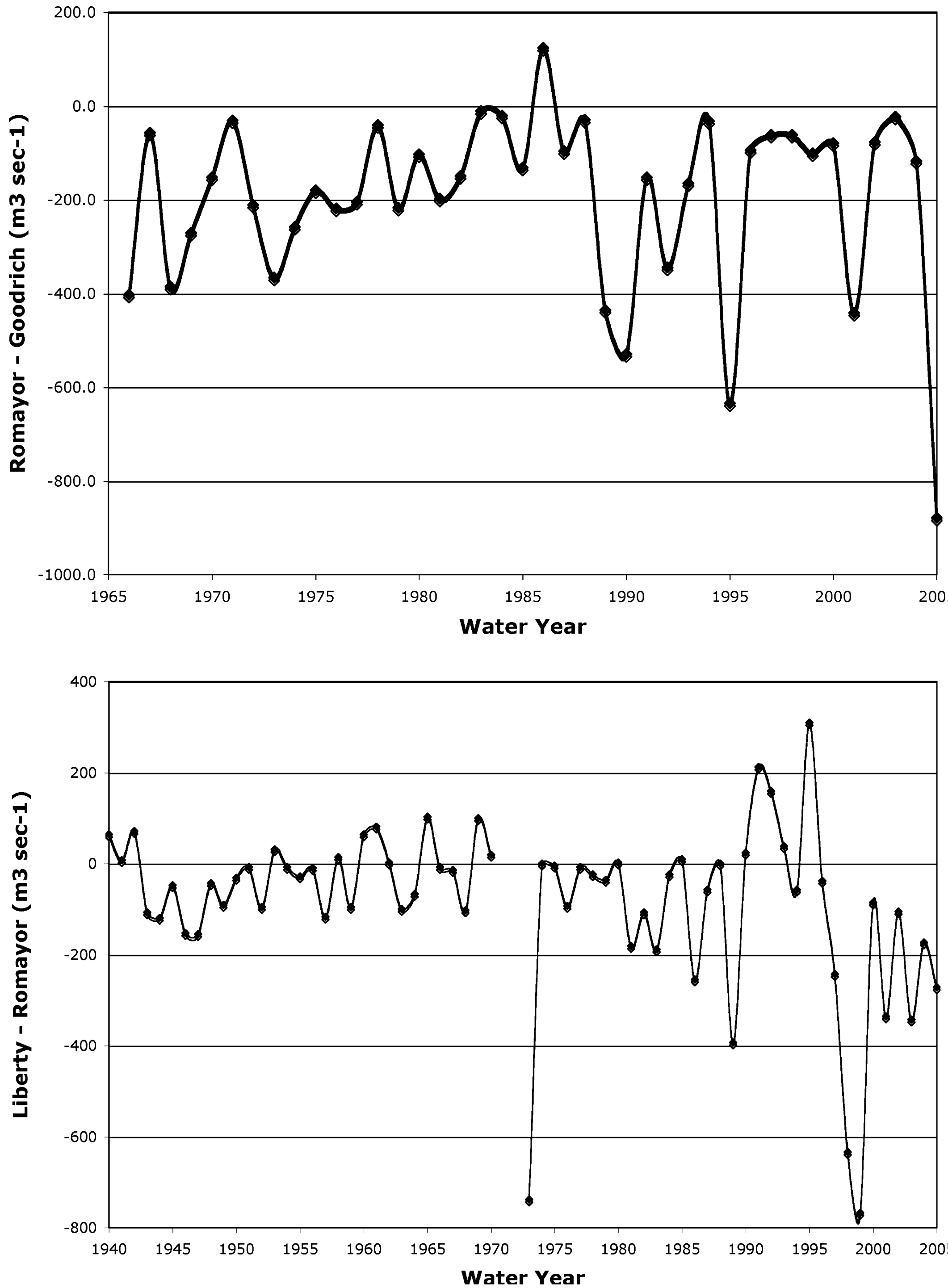

Figure 2 Annual peak discharge at the Romayor gaging station minus that at the Goodrich station 18 km upstream (top), and at the Liberty gaging station minus that at Romayor, 43 km upstream. Lines between points are included to facilitate visual comparisons and do not imply continuous change.

Romayor a bit earlier. Later the same morning the peak elevation of Lake Livingston occurred, and the drawdown began, along with the hydrograph rise at Goodrich. Long King Creek, Goodrich, and Liberty peaked on September 25, with Moss Bluff peaking early on September 27 (Table 3).

Storm runoff resulted in a roughly half-meter rise in the elevation of Lake Livingston. Wind-wave damage to Livingston Dam, however, required the Trinity River Authority to draw down the lake to inspect damage and begin repairs. Lake elevation peaked at 0800 on September 24, and was drawn down over the next three days, leveling off about 1 m below pre-storm water levels early on September 27 (Fig. 3).

The hydrograph responses of the river at Goodrich and Romayor (Fig. 4) show a rapid rise and recession similar to

Table 3 Key stages of the Hurricane Rita flow event, 2005

Code	Date & time		Significance
R1	9/24	0200	Start of hydrograph rise @ Romayor
R2	9/24	0430	Start of hydrograph rise @ Long King Creek
R3	9/24	0800	Peak elevation, Lake Livingston; start of drawdown
R4	9/24	0830	Start of hydrograph rise @ Goodrich
R5	9/25	0700	Peak @ Long King Creek
R6	9/25	1545	Peak @ Goodrich
R7	9/25	2330	Peak @ Romayor
R8	9/27	0200	Lake drawdown complete; Liberty near peak
R9	9/27	0315	Peak @ Moss Bluff

the Lake drawdown curve (Fig. 3), with the peak at Romayor occurring 7.75 h after Goodrich. In both cases, following recession the base flow remained only slightly elevated from the pre-storm flow.

By contrast, stations further downstream at Liberty and Moss Bluff (Fig. 5) showed a sustained rise in base flow. Note that while discharge at Liberty was partly estimated, the peak and recessional limb are based on published data. The Moss Bluff discharge, however, is entirely estimated. The West Gulf River Forecast Center of the National Oceanic and Atmospheric Administration sometimes estimates discharges for this station based on gage heights. An empirical equation based on such records for similar stages was used to estimate discharge for the Rita event. However, this station — even more so than Liberty — is affected by backwater effects and both lunar and wind tides, and there is thus some uncertainty in converting gage heights to discharge.

Water levels at Old River declined only slightly after the storm, as did those at the Wallisville station, where the tidal signal is clearly overprinted (Fig. 6).

Slope

Instantaneous water surface slopes for the nine key times during the Hurricane Rita event were determined by determining surface elevations based on gage heights and datums, and the channel distance between stations. Water surface profiles for the Rita event are shown in Fig. 7. At the highway 105 river crossing near Moss Hill flood debris and an interview with a local resident both suggested that the Rita water levels peaked just under the bridge. This implies a stage elevation of 16—17 m, consistent with the computed water surface slope between Romayor and Liberty.

The water surface slopes show that in all cases, slopes decrease substantially downstream of the Liberty station. Gradients from Goodrich to Romayor to Liberty are variable, but always positive and always >0.0011. Slopes between Liberty and Moss Bluff may be negligible or negative. From Moss Bluff to Wallisville water surface gradients are even more variable, ranging from +0.117 to −0.079, the steepest positive and negative slopes at any station. Negative slopes

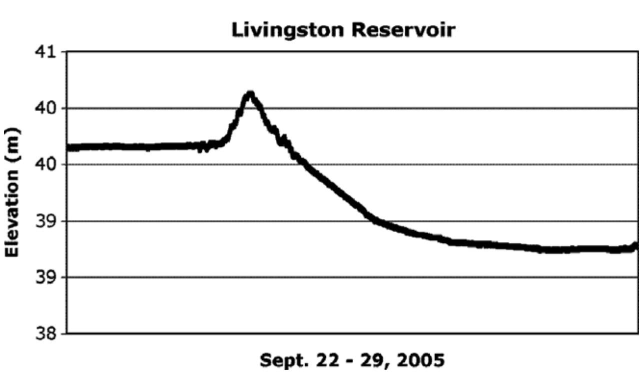

Figure 3 Water surface elevations for Lake Livingston for the week including Hurricane Rita, showing the rise in lake levels and subsequent drawdown via dam releases to assess and repair damages.

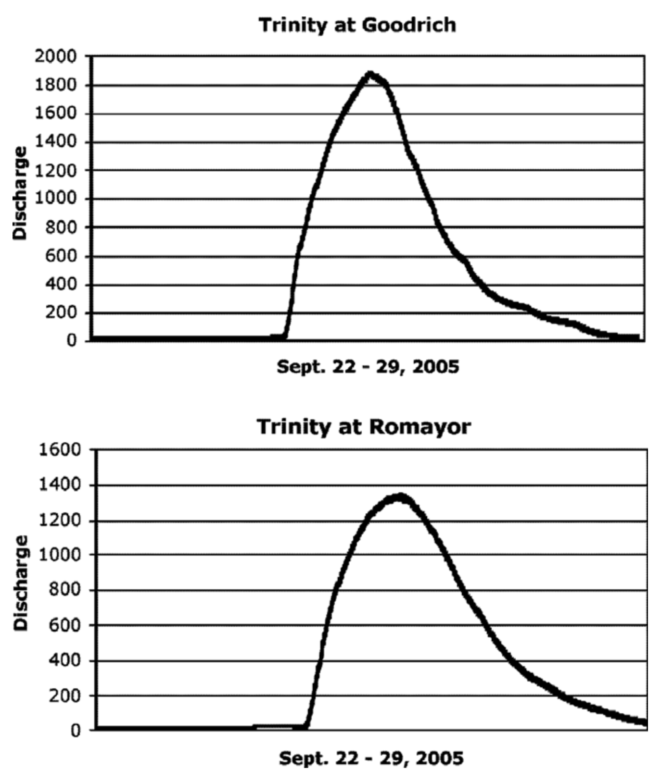

Figure 4 Hydrographs for the lower Trinity River at the Goodrich and Romayor gaging stations for the week including the Hurricane Rita flow event. Discharge was measured every 15 min.

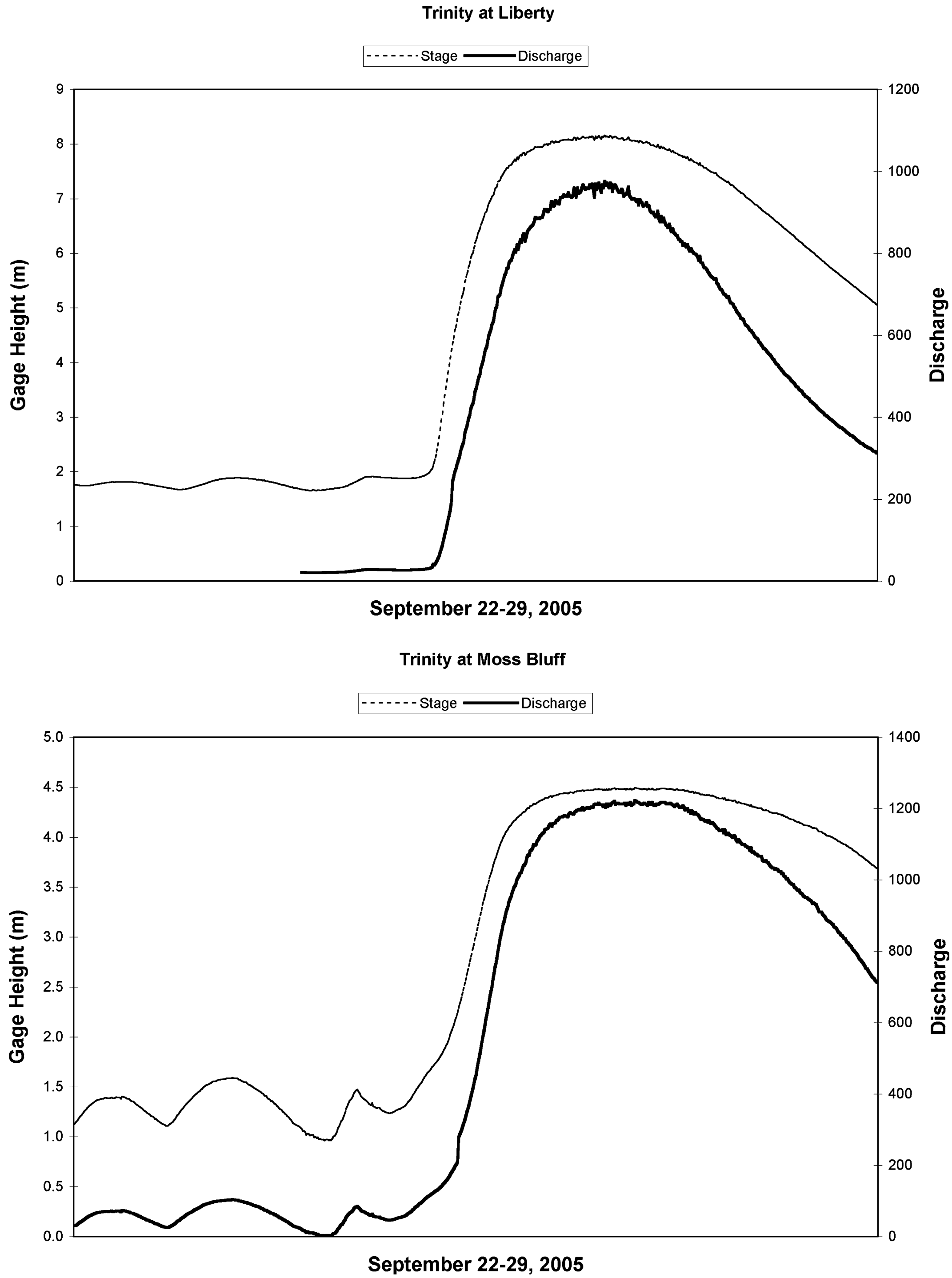

Figure 5 Stage (gage height) and discharge for the lower Trinity River at the Liberty and Moss Bluff gaging stations, with readings every 15 min. For the Liberty data, discharge was estimated by the author for the portion of the curve prior to the point indicated by the arrow. For Moss Bluff, discharge is entirely estimated by the author.

in the lower river can occur due to tidal effects and wind forcing.

Stream power

Cross-sectional stream power was estimated for a number of reference flows at the Goodrich, Romayor, and Liberty stations by Phillips and Slattery (2006) using channel thalweg slope as a surrogate for energy grade slope. As the previous section shows, water surface — and therefore energy grade — slopes may vary considerably between and within flow events. While water surface gradient is still only a surrogate for energy grade slope, and the distances between stations (18—50 km) are quite large, they allow a first-order assessment of the downstream variation of stream power during the Rita event.

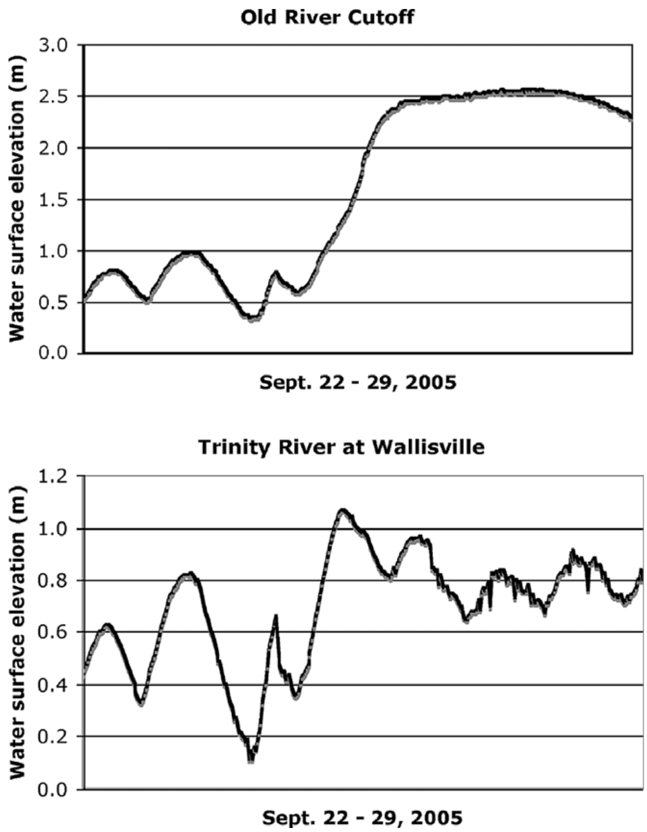

Figure 6 Water surface elevations for the week including the Hurricane Rita event at the Old River cutoff and Wallisville sites. Readings were taken every 15 min.

The data set allows calculation of either ''import'' or ''export'' stream power for each station (Fig. 8), using the instantaneous discharge and either the upstream or downstream water surface slope. During the first four samples of the Rita event (up to the start of the hydrograph rise at Goodrich), power is low at all cross sections. As the river peaks at Goodrich and Romayor, stream power increases substantially, and is much higher than at the downstream stations. As Liberty and Moss Bluff peak, the flood wave from the lake drawdown has passed the upstream stations, where stream power is now less than the downstream points (Fig. 9).

The stream power trends for the Rita event are consistent with previous work indicating a sediment storage bottleneck downstream of Romayor, whereby power is insufficient to transport the imposed sediment load, reduced though it is by trapping in Lake Livingston (Phillips et al., 2004).

Floodplain, tributary, and distributary morphology

The topography and geometry of the floodplain and tributaries were examined between the Goodrich and Romayor stations to investigate possible causes for the reduction in flow that sometimes occurs between the stations (Table 2). This was also noted in the Rita event, as the peak discharge at Goodrich was 39% higher than at Romayor.

Mussel Shoals Creek, which joins the Trinity downstream of the Goodrich station (Fig. 10), does so at an angle which is more characteristic of a distributary than a tributary channel. These are sometimes termed barbed tributaries, but to some geomorphologists the latter term implies

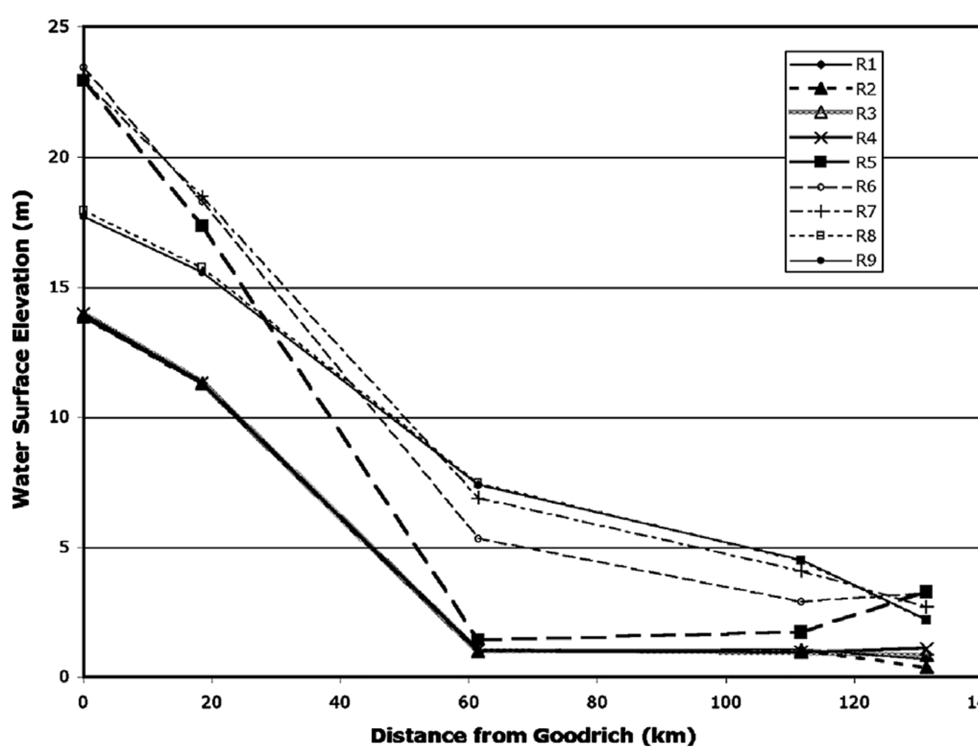

Figure 7 Water surface profiles from Goodrich to Wallisville for nine sample times during the Hurricane Rita flow event. Note the negative slopes for some measurements between Moss Bluff and Wallisville in the lowermost segment.

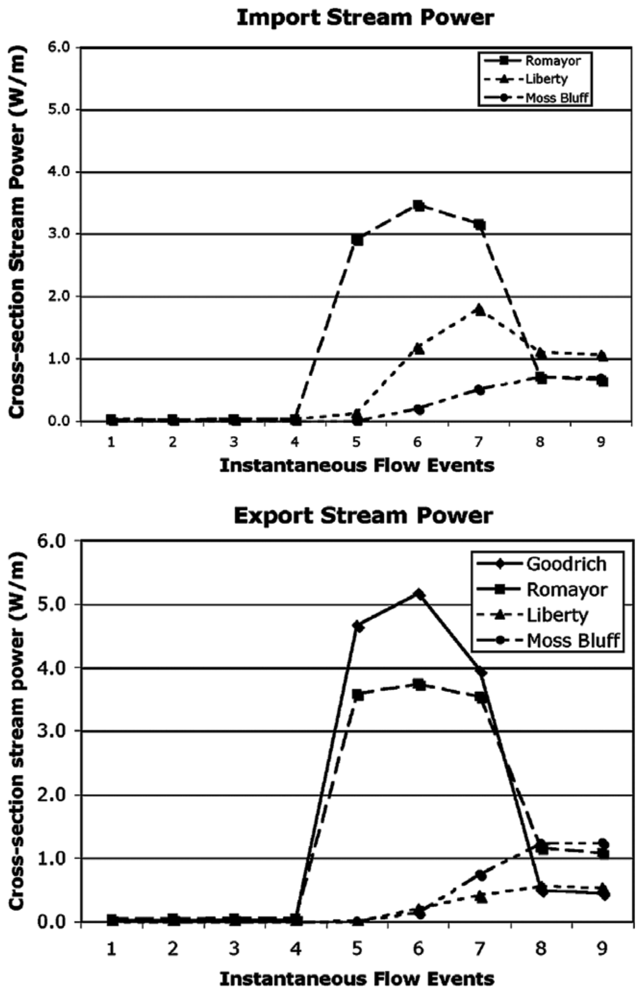

Figure 8 Stream power at Trinity River gaging stations for nine Hurricane Rita instantaneous flows, based on discharge and upstream (import) or downstream (export) slope.

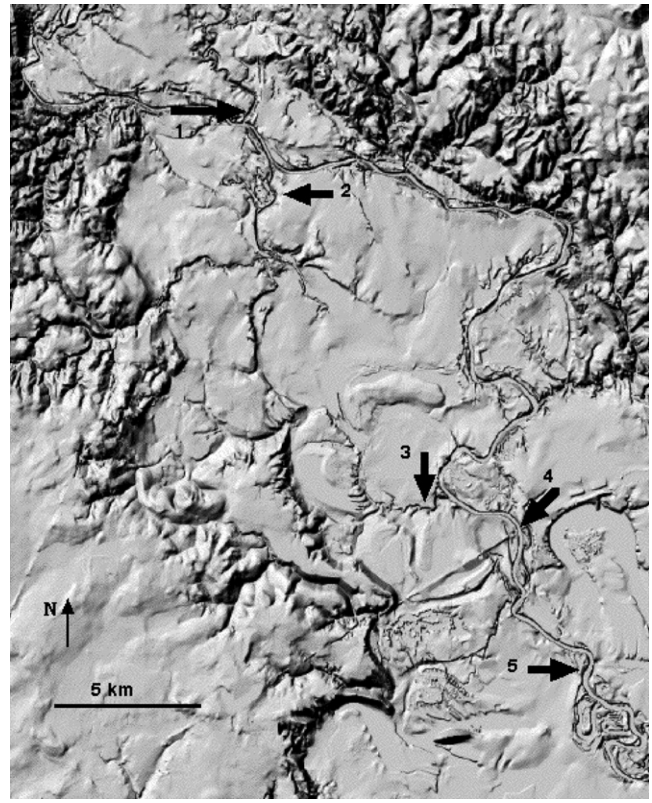

Figure 10 Shaded relief map (50× vertical exagerration) of the lower Trinity River valley in the vicinity of the Goodrich and Romayor gaging stations. Numbered arrows identify (1) Goodrich gage site; (2) Mussel Shoals Creek; (3) Big Creek at the southern end of Grama Grass Bottom; (4) Romayor gage site; and (5) approximate location of the morphological transition zone. The Romayor gage is located at 30°25′30″N and 94°51′02″W. Big Creek and lower Grama Grass bottom begin backflooding from the Trinity River as stages at Romayor rise from about 15 to 19 m amsl. Mussel Shoals Creek begins backflooding from the river as Goodrich stages rise from approximates 21 to 23 m amsl.

stream capture, which is not the case here. Analysis of topographic gradients from the DEM indicate that portions of the channel drain away from the river, toward Grama Grass Bottom. Simulated flooding of the DEM to uniform depths indicates that (assuming water surface elevations at the confluence are approximately the same as at the Goodrich gage), that Mussel Shoals Creek begins backflooding from the river as water elevations at Goodrich rise from 21 to 23 m. This is below bankfull stage in this vicinity. During the Rita event stages at Goodrich reached this level late on September 24. The longitudinal profile of the creek channel suggests that backwater flooding to about 22 m could direct flow upstream.

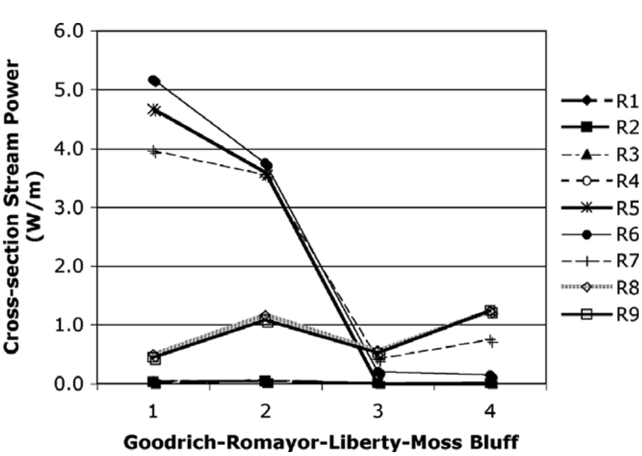

Figure 9 Stream power (export) at four stations for nine Hurricane Rita instantaneous flows.

Big Creek, the largest tributary of the lower Trinity on the west side of the valley, joins the river at the expected acute angle, and flows through the southern end of Grama Grass bottom. The mouth of Big Creek, observed in the field at low flows, was not discharging water into the Trinity (though there was signifigant flow at several cross-sections of the upper reaches of the creek). A DEM analysis similar to that above indicates that Big Creek and lower Grama Grass bottom begin backflooding when stage elevations at Romayor rise from about 15 to 19 m. Again, this is well below bankfull levels. The Trinity at Romayor reached this stage late on September 24.

Thus, as river stages rise, Mussel Shoals and Big Creeks do not merely backflood, but become distributaries rather than tributaries of the Trinity, delivering water to the depressional areas of Grama Grass bottom, thus reducing the proportion of flow passing the Goodrich gage which is recorded at Romayor.

A short distance downstream of Moss Bluff, the Trinity clearly transitions to a dominantly divergent, distributary network at the confluence of Old River cutoff. Pickett's Bayou, which joins the Trinity upstream of Moss Bluff, connects the river with Old River in a marshy area of the Trinity River delta. It is not clear from maps the extent to which the bayou is a tributary of the Trinity or Old River. In the field, the confluence of Pickett's Bayou and the Trinity River has no single dominant mouth (or inlet). Rather, at least five subchannels dissect the river bank. Field surveys indicate the beds (Fig. 11) are 3.5–4 m above the river channel. Bayou channel slopes and flow indicators show the dominant flow pattern is clearly away from the river. Thus it appears that Pickett's Bayou serves as a tributary of Old River during low and normal water flows, draining a portion of the delta and adjacent terrraces. During high flows, however, the bayou becomes a distributary of the Trinity River. The elevation of the bayou channels at the river bank is approximately the same as that of the top of the point bar opposite the confluence. At this site, the distributary function comes into play at approximately bankfull flow. Shaded relief and surface images derived from the DEM (Fig. 12)

Figure 11 (top) One of several channels at the confluence of Pickett's Bayou and the Trinity River. The elevation of the tributary channel is well above normal Trinity Water levels, but slopes away from the river bank, so that the bayou functions as a distributary during high flows. (bottom) Two of at least five channel inlets at the confluence of Pickett's Bayou and the Trinity River.

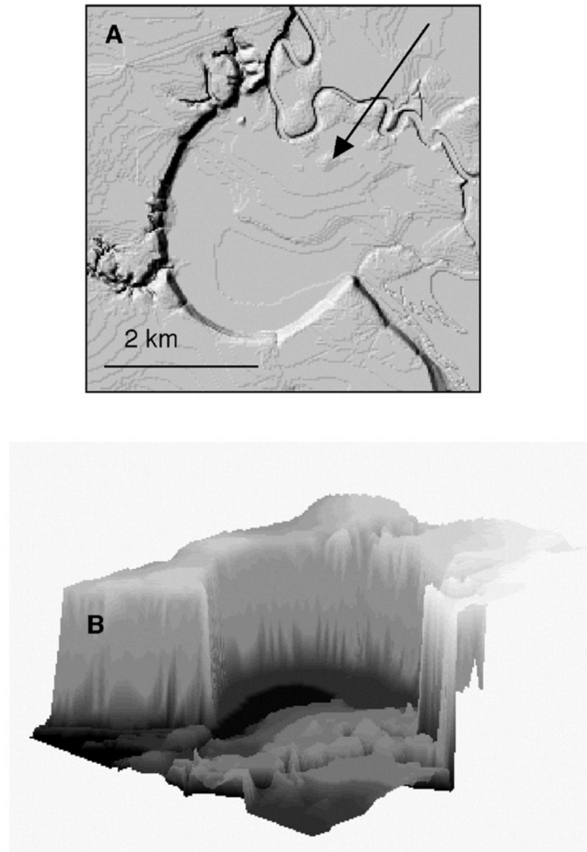

Figure 12 Trinity River Floodplain topography in the vicinity of Pickett's Bayou near Moss Bluff, Texas. A shaded relief map (A) shows the Trinity River, and the large palaeomeander defining the western valley wall. The arrow indicates the direction of view for the shaded surface model (B), which illustrates the topographic controls which tend to direct flow away from, rather than toward, the Trinity River channel.

show that topographic gradients lead generally away from the river toward the southeast.

Both Grama Grass bottom and the depression shown in Fig. 12 are palaeomeanders of the ancestral Trinity River. The role of this inherited valley morphology in determining modern flow patterns will be addressed in the discussion.

Discussion

Despite being a humid perennial stream with no significant transmission losses, and despite numerous tributary inputs, discharge as recorded at gaging stations does not necessarily increase downstream in the lower Trinity River. In the clearly fluvially dominated reach from Goodrich to Romayor, while mean annual discharge is slightly higher at the downstream station, discharges associated with six reference flows (1%, 10%, and 50% exceedence probability, and recurrence intervals of 1, 2, and 10 years) are actually lower at the downstream station. Peaks associated with a moderate 2002 flood were higher at Romayor, but the peak for the 1994 flood of record was higher at Goodrich. Peak flows in the 2005 Hurricane Rita event also showed an apparent decline in flow between Goodrich and Romayor. Annual peak flows are often higher at the Romayor station.

The apparent cause of the discrepancy is backflooding and flow reversal in two tributaries, Mussel Shoals and Big Creeks. At higher than average but less than bankfull flows these creeks are backflooded by the river, and local topographic gradients lead to Grama Grass bottom, a depression within the river valley. This flow diversion may reduce discharge recorded at Romayor. If the magnitude or duration of high water is sufficient to fill the depressions, however, no peak flow reduction downstream of Goodrich is likely to occur.

Gaging stations further downstream are influenced by tidal and coastal backwater effects. Mean and reference flows at Liberty are substantially higher than at the upstream stations, but the discharge data are biased towards fluvially dominated events. Event peaks at Liberty may be lower than at the upstream stations, as shown by the 2002 flood and the Rita event.

Water surface slopes decline systematically from Goodrich to Romayor to Liberty, but further downstream slopes may be negative due to tidal and backwater effects. Slopes in the lowermost reach from Moss Bluff to Wallisville are the most variable, including the steepest positive and negative water surface slopes, reflecting the downstream translation of the Lake Livingston dam release and the backwater flooding effects of the storm.

Downstream of Moss Bluff the Trinity River is clearly dominantly divergent and distributary at all times, discounting periods of backwater effects and upstream flows. At least one upstream confluence is also distributary at high flows. Pickett's Bayou diverts water from the river at flow stages slightly less than bankfull. The bayou thus serves as a tributary of local runoff to Old River most of the time, but as part of the Trinity distributary network at high flows.

The depressional areas of both Grama Grass bottom and the Pickett's Bayou area are associated with palaeomeanders. The Trinity River is flanked by a modern floodplain and flights of several Pleistocene terraces. The oldest and highest are termed the Beaumont terrace, correlative with the Prairie surface in Louisiana. The modern lower Trinity River valley is cut into the Beaumont surface. Dates for the Prairie-Beaumont terrace in Louisiana and Texas range from 33 to 195 Ka, with a date from Winnie, Texas (the closest site to the Trinity) of 102.3 ± 8.3 Ka (Otvos, 2005). Blum et al. (1995) date the incision into the Beaumont terraces at about 100 Ka, broadly consistent with Thomas and Anderson's (1994) date of about 110 Ka, and within the range of Beaumont dates indicated by Otvos' (2005) synthesis (74–116 Ka).

Below the Beaumont surface, and often merging into the modern floodplain, are a series of up to three alluvial terraces, traditionally referred to as Deweyville, though they are not now generally believed to be part of a single terrace system (Blum et al., 1995; Morton et al., 1996). The paleomeanders in the lower Trinity Valley, often expressed as swampy depressions or meander scrolls, occur on the Deweyville surfaces, with radii of curvature and amplitudes suggesting significantly larger paleodischarges than at present (Alford and Holmes, 1985; Blum et al., 1995). These are generally cut laterally into Beaumont sediments. Between incision into the Beaumont and the current Holocene sea level rise, the Trinity underwent several entrenchment/aggradation cycles (Blum et al., 1995; Morton et al., 1996; Thomas and Anderson, 1994).

While the antecedent topography associated with incision into the Beaumount surface, and the Deweyville terraces and paleomeanders, does not constitute geological control in the traditional sense, it does apparently exert important influences on the modern river. Rodriguez et al. (2005) emphasized the importance of alluvial terrace inundation in creating flooding surfaces during transgression of the Galveston/Trinity Bay/Trinity delta area. Phillips et al. (2005) related the morphological and process transition zone in the river to the upstream limits of the effects of Holocene sea level rise. This study suggests that the location and gradient of tributaries and distributaries is strongly influenced by the antecedent landforms, and that water and other mass fluxes may be diverted from the river channel at high river flows.

Conclusions

There are no systematic downstream patterns of increases or decreases in the discharge, stream power, or water surface slope of the lower Trinity River. Discharge in the river channel may decrease downstream due to coastal backwater effects in the lowermost reaches, and due to diversion of low into valley-bottom depressions during high flows in both the fluvial and fluvial-estuarine transition reaches. In general, however, stream power and slope decrease in the lower reaches, consistent with earlier findings of limited fluvial sediment delivery to the coastal zone.

Some river tributaries may become distributaries at high but sub-bankfull flows, as backwater effects reverse flows into depressions associated with paleomeanders. The paleomeanders, and possibly the locations of these ''reversible'' channels, are related to antecedent topography associated with aggradation/degradation cycles over the past 100 Ka or so.

Results reinforce the notion that coastal plain rivers may not function as simple conduits from land to sea, and that the transition from fluvial to coastal dominance may be variable along the river, with the variability controlled not just by the relative magnitude of river and tidal or backwater forcing, but also by valley topography controlled in part by antecedent landforms.

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Part 3

Antecedent Alluvial Morphology and Sea Level Controls on Form-Process Transitions Zones in the Lower Trinity River, Texas

Introduction

Alluvial coastal plain rivers are not considered to greatly affected by geological controls, at least in the form of the structural and lithological variations that are typically important in fluvial geomorphology in other settings. Alluvial river valleys are in some senses self-formed, and alluvial coastal plain rivers are often found in passive-margin, tectonically stable areas, with unconsolidated substrates whose resistance does not greatly constrain fluvial erosion. Despite this, geological controls expressed in the form of persistent landforms and antecedent topography may be quite important in alluvial coastal plain rivers. The purpose of this study is to examine the potential effects of antecedent alluvial morphology and the history of sea-level driven changes on the lower Trinity River, Texas in determining the location of a critical "hinge point" or form/process boundary. Note that the term "antecedent" is used here in its basic meaning of pre-existing or initial conditions relative to the modern river, and is not necessarily meant to invoke the Davisian concept of antecedent streams (those which maintain their general course in the face of crustal warping).

Several recent studies have shown that antecedent topography and inherited geologic structures are critical in controlling the Holocene evolution of passive-margin coastal plain coastlines (Riggs et al. 1995; Dillenburg et al., 2000; McNinch, 2004; Harris et al., 2005). Inherited landscape controls in fluvial systems are well known, particularly with respect to valley size and shape, and have been shown to be important in some alluvial rivers without significant bedrock constraints (e.g. Fryirs, 2002; Wasklewicz et al., 2004). Bishop and Cowell (1997) showed the interactions of fluvial system and coastline development on an embayed coast in terms of lithological and morphological controls on coastal development. Studies of sedimentary deposits in Trinity Bay, Texas, the Galveston estuary as a whole, and the continental shelf offshore, showed that the incised river valleys of the ancestral Trinity and Sabine Rivers provided critical controls over subsequent evolution (Thomas and Anderson 1994; Rodriguez and Anderson, 2000; Rodriguez et al., 2001; 2005). Along-strike variability in antecedent topography associated with variable inner-shelf slope gradients was linked to variable Holocene coastal retreat rates in east Texas by Rodgriguez et al. (2004), who reaffirmed the general principle that the geologic setting in general, and antecedent topography in particular, plays a key role in controlling coastal and coastal plain evolution.

Rivers draining to the coast--particularly those that cross broad coastal plains--often exhibit a transition zone from complete fluvial domination upstream to domination by coastal processes in the estuary, with a systematic but complex and variable combination of fluvial and coastal influences in between. From the geomorphological perspective, this is reflected in variations in landforms, channel and valley morphologies, and sediment transport/storage regimes. The lower end of such transition zones has received considerable attention from stratigraphers due to the need to interpret sedimentary sequences representing alluvial, deltaic, estuarine, and marine facies. However, the fluvial-to-estuarine transition zone may extend well upstream of estuaries and deltas. In the coastal plain portion of the Trinity River, Texas, a critical transition zone has been identified, reflected in sediment dynamics, valley morphology, and channel response to upstream disturbances (Phillips et al., 2004; 2005). The purpose of this paper is to further explore the earlier suggestion that this transition corresponds with the upstream limit of the effects of Holocene sea level rise (Phillips et al., 2005), and to determine whether antecedent morphology associated with Quaternary marine terraces and/or alluvial valleys plays a role in determining the location of this transition.

Fluvial to Coastal Transitions

The morphological mouth of a river can be readily defined in most cases on the basis of where a well-defined channel or set of channels discharges into an estuary or other open water body. However, the transition from fluvial to coastal domination could also be defined on the basis of transition from a convergent to a distributary flow network, geochemical criteria such as salinity, hydrographic criteria such as upstream limits of tidal influence, topographic criteria such as the point at which the channel is cut to below sea level or channel width/depth ratios, and geomorphic/sedimentological criteria such as coastal plain or deltaic sediment bottlenecks, or loci of deposition. Any of these is likely to fluctuate at various time scales according to river flows, tides, sea level change, and other factors. The location of these variously defined fluvial/coastal boundaries can be tens of kilometers, and often more than 100 km, upstream of the morphological mouth (Giese et al., 1979; Renwick and Ashley 1984; Nichols et al., 1991; Phillips and Slattery, 2006).

Studies of the downstream geomorphic effects of Lake Livingston and Livingston Dam, 176 km above Trinity Bay, showed that channel changes attributable to the dam extend about 55-60 km downstream of the dam, and that this corresponds with obvious changes in valley morphology, reflected in a wider valley, lower elevations, and frequent oxbows downstream (Phillips et al., 2005). Sediment starvation effects of the dam are also not evident downstream of this point, where pronounced increases in sediment storage and decreases in sediment transport occur (Phillips et al., 2004). We refer to this reach as the critical zone, to distinguish it from the more general concept of the fluvial-estuarine transition zone. The higher sinuosity downstream of the critical zone led to the suggestion that the transition corresponds to the upstream limit of Holocene sea level rise effects on the Trinity River (Phillips et al., 2005).

A process and morphological transition might simply reflect the current location of upstream encroachment, proceeding more-or-less gradually through time. In coastal plain rivers of North Carolina, Phillips (1992) found that the boundary between fluvially-dominated alluvium and locally-derived autocthonous predominantly organic-rich alluvium corresponded with a subtle scarp marking a Pleistocene paleoshoreline. This suggests that the upstream propagation of the effects of base level rise may be stalled for periods of time at significant "steps" associated with inherited landforms. Rodriguez et al. (2005) have shown that alluvial terrace inundation is an important autocyclic mechanism in the formation of sedimentary sequences of the Galveston Bay estuary, Texas (of which the Trinity River is the major tributary). Essentially, even if sea level rise and sediment supply are constant, transgression of an alluvial terrace surface represents a significant threshold, resulting in the formation of a flooding surface and geologically rapid reorganization characterized by a sudden increase in accomodation space and an upstream shift in coastal facies (Rodriguez et al., 2005). This leads to the hypothesis that the geomorphological critical zone on the lower Trinity is controlled by antecedent topography and forms associated with Trinity River alluvial terraces, and/or with coastal plain terraces and paleoshorelines.

Determining the controls over critical process/form transition zones in coastal plain rivers has a number of practical implications. Such changes in process dominance may help predict downstream limits of the effects of upstream perturbations such as dams or channel modifications, as is the case in the Trinity. Further, such geomorphic boundaries often correspond with critical ecotones. Finally, if antecedent alluvial and coastal plain morphology controls the location of upstream effects of sea level change, then mapping of these features can be extremely useful in assessment and prediction of the hydrological and ecological, as well as the geomorphological, effects of sea level change. Form/process transition zones associated with the inherited topography may be important "hinge points" not only for geomorphology, but for water, land, wetland, and biological resource management in the river corridor.

Study area and background

Trinity River and Galveston Bay

The 46,100 km^2 Trinity River drainage basin flows to the Trinity Bay, part of the Galveston Bay system on the Gulf of Mexico (Figures 1, 2). The lower river is defined here as the portion downstream of Lake Livingston. The climate is humid subtropical.

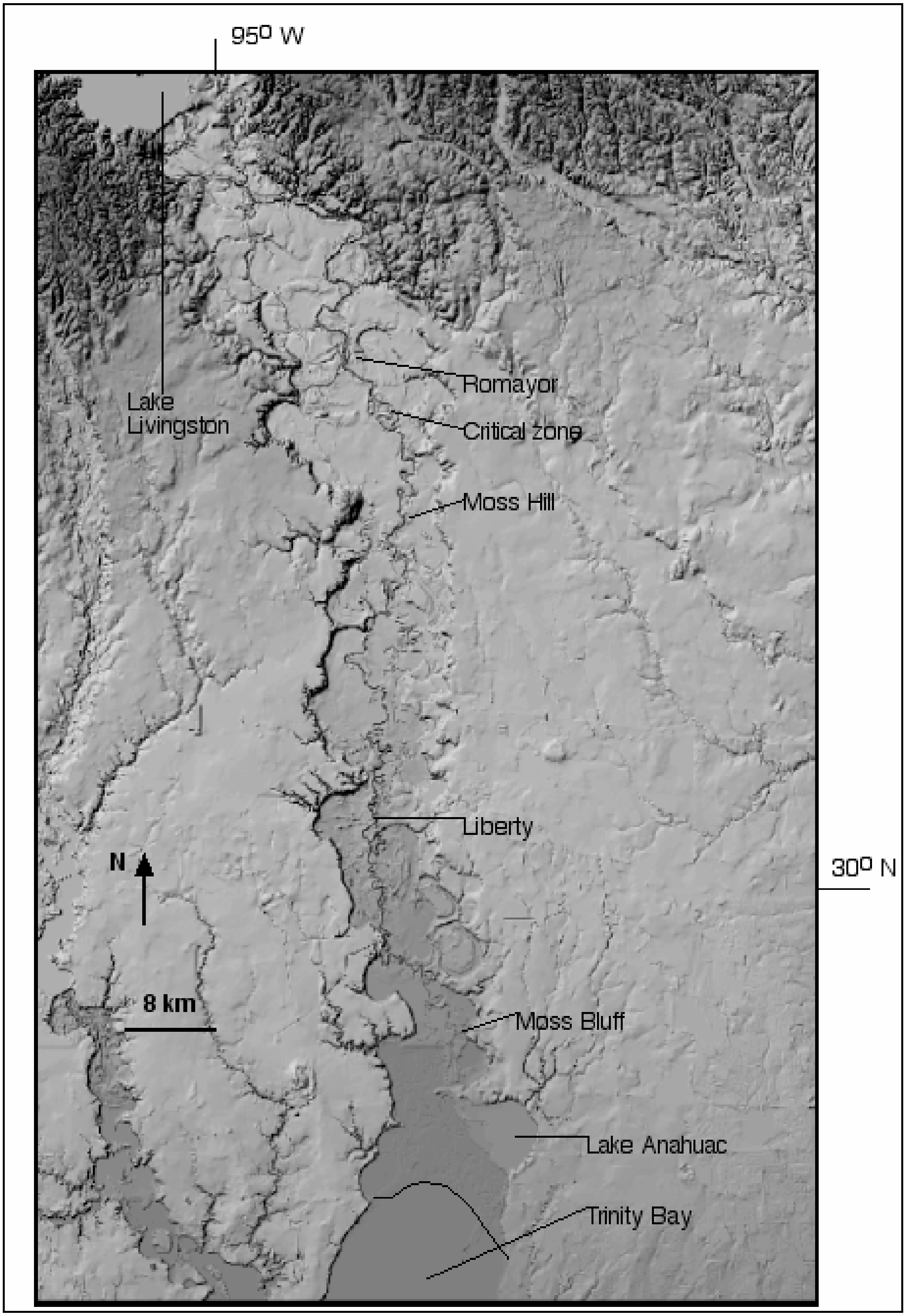

Figure 1. Shaded relief map (50X vertical exagerration) of the lower Trinity River Valley.

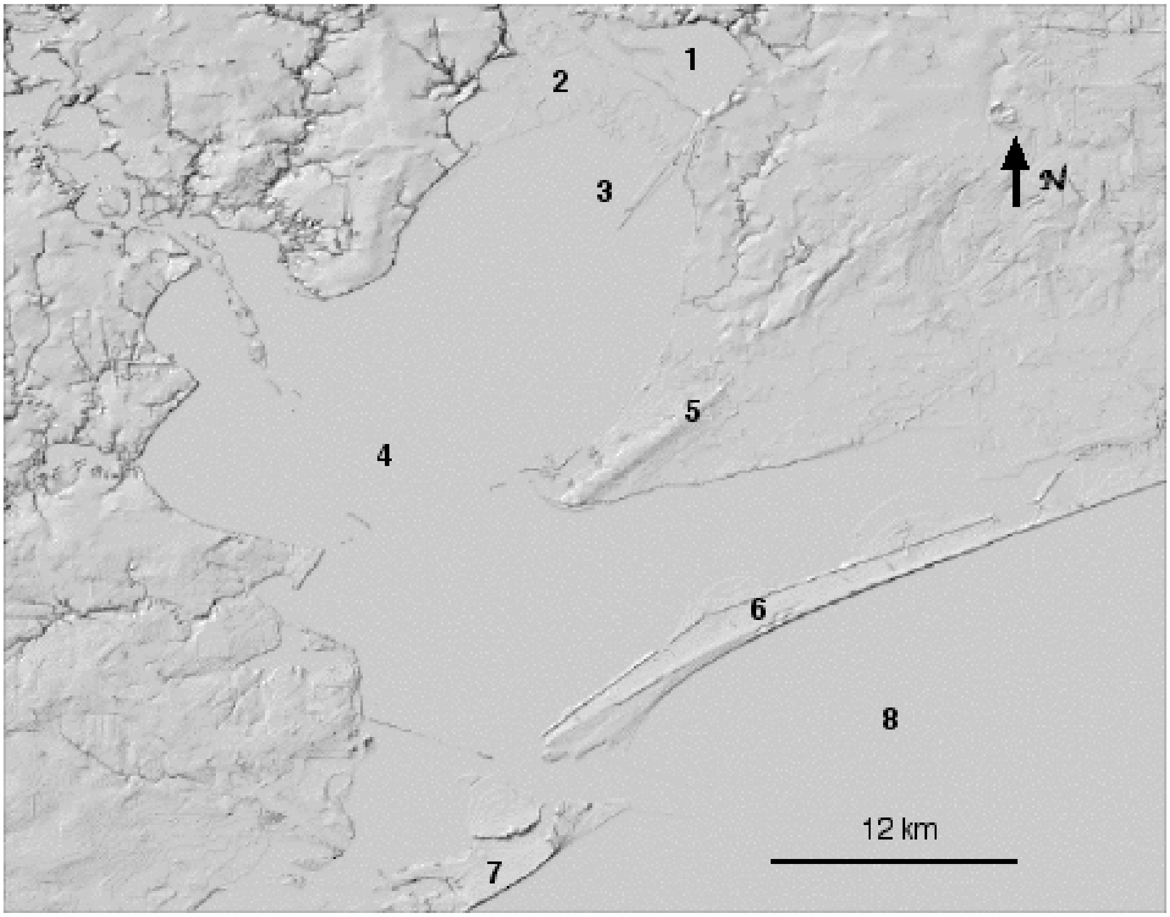

Figure 2. Shaded relief map (50X vertical exaggeration) of the Galveston Bay area, Texas. Numbered locations are: (1) Lake Anahuac; (2) Trinity River delta marshes; (3) Trinity Bay and Trinity River navigation channel; (4) mid Galveston Bay; (5) Pleistocene beach ridge associated with the Ingleside barrier paleoshoreline; (6) Bolivar Peninsula; (7) Galveston Island; (8) Gulf of Mexico.

The lower Trinity floodplain contains numerous oxbow lakes, meander scars, and other evidence of Holocene and historical channel change, confined within an incised valley. Evidence of Pleistocene channel migration is preserved on alluvial terraces. The river channel has extensive evidence of bank erosion and point bar accretion. The lower river is therefore an actively migrating channel and has been throughout the Quaternary.

Galveston Bay, which includes Trinity Bay, is located in southeast Texas, adjacent to the Houston-Galveston metropolitan area. The estuarine surface area is about 1,554 km^2, and mean volume is about 2.7 billion m^3 (GBFIG, 2003). The bay is a lagoon-type estuary, separated from the Gulf of Mexico by Galveston Island and the Bolivar Peninsula. The bay's drainage area is 85,470 km^2, of which about 54 percent is the Trinity River.

Details of sea-level and Quaternary coastal evolution in Texas are disputed (c.f. Blum et al., 2002; Otvos, 2005), but most sources agree that Galveston Bay was formed in roughly its current location about 4000 years ago. During lower sea level stands in the Quaternary, the Trinity and Sabine Rivers cut incised valleys that converged on the present continental shelf. For the past 18,000 years the offshore Trinity-Sabine incised valley has backfilled (Blum et al., 1995; 2002).

Nichols (1989) reported short-term sea level rise rates of 5.5 mm yr^{-1} for Galveston Bay, from tidal gage records, and long-term rates of 1.4 mm yr^{-1}, from stratigraphic evidence. Coastal submergence, which accounts for both water level and land surface elevation changes, is substantially higher due to subsidence associated with water and hydrocarbon withdrawal, as well as autocom-paction of sediments. Recent tide gage records and interferometry show submergence rates averaging about 7.6 mm yr^{-1} (Stork and Sneed, 2002). White et al. (2002) combined an estimated eustatic sea level rise of 2.2 mm yr^{-1} with mean subsidence of 8.1 mm yr^{-1} at four lower Trinity valley benchmarks to arrive at an estimate of 10.3 mm yr^{-1} coastal submergence.

The Trinity River changes from a typical convergent drainage network to a divergent, distributary network downstream of Moss Bluff, Texas, 19.5 km upstream of Trinity Bay. The longitudinal profile indicates that the channel bed elevation is cut to sea level 110 km upstream of the bay, and tidal influence is evident at the gaging station at Liberty, 85 km upstream. The upstream limit of the sediment storage bottleneck, just downstream of the scour zone below the dam (e.g., the critical zone), is about 125 to 130 km from Trinity Bay.

Alluvial Terraces

The Trinity River is flanked by a modern floodplain and flights of several Pleistocene Terraces. The oldest and highest are termed the Beaumont terrace, correlative with the Prairie surface in Louisiana. Shafer (1966) presumed that the Beaumont terraces in the lower Trinity River were deposited during the last interglacial of the Wisconsin period, 30 to 60 Ka. Dates for the Prairie-Beaumont terrace in Louisiana and Texas compiled by Otvos (2005) range from 33 to 195 Ka, with a date from Winnie, Texas (the closest site to the Trinity) of 102.3 $\pm$ 8.3 Ka. Otvos' (2005) analysis places the deposition of the Beaumont terraces in Texas, which are 50 to 100 km wide from the coast, at 74 to 116 Ka--broadly consistent with Blum et al. (1995) and Thomas and Anderson (1994).

Between the Beaumont surface and often merging into the modern floodplain are a series of up to three alluvial terraces. These are usually referred to as Deweyville, though they are not now generally believed to be part of a single terrace system (Blum et al., 1995; Morton et al., 1996). In most locations two (Shafer, 1966), "at least two" (Blum and Price, 1998), or three (Blum et al., 1995; Morton et al., 1996; Rodriguez et al., 2005) separate "Deweyville" surfaces are recognized. The lowermost Deweyville surfaces are only slightly higher than the modern floodplain, and in some cases are buried by the latter, with natural levees of the modern floodplain higher than backswamps of the lower Deweyville

(Alford and Holmes, 1985; Anderson et al., 2005; Blum et al., 1995). Aerial photographs show obvious palaeomeanders in the Trinity Valley, expressed as swampy depressions or meander scrolls. These occur on the Deweyville surfaces, with radii of curvature and amplitudes suggesting significantly larger palaeodischarges than at present (Alford and Holmes, 1985; Blum et al., 1995).

In the lowermost Trinity, Shafer (1966) dates the Deweyville terraces at 5-7 Ka; and Alford and Holmes (1985), in the nearby Sabine River, at 4-9 Ka. In the Colorado River, Texas, Blum and Price (1998) place the deposition of the Eagle Lake Alloformation, youngest of the group, from 20 to 14 Ka, followed by incision from 14-12 Ka, and then Holocene valley fill. The three Deweyville surfaces are designated (youngest to oldest) the Fredonia, Sandjack, and Merryville alloformations by the Louisiana Geological Survey (Heinrich et al., 2002).

In the Sabine River, Otvos' (2005: 102) chronology indicates entrenchment from about 100 to 50 Ka, and aggradation, producing two terraces, from 40 to 20 Ka. These were followed, based on optically stimulated luminescence dating, by entrenchment from 20 to 18 Ka and aggradation from 18 to 2 Ka (Otvos, 2005: 102). The Sabine and Trinity systems were connected during lower sea level stands on what is now the continental shelf, and Thomas et al. (1994) date the oldest incision of the Trinity-Sabine system at about 110 Ka. Blum et al. (1995) estimate the incision associated with the Beaumont terraces at about 100 ka, associated with marine oxygen isotope stage 5 (115 to 75 Ka). Multiple episodes of lateral channel migration, degradation, and aggradation occurred within those incised valleys during isotope stages 4, 3, and 2 glacials as channels graded to shorelines further out on the current continental shelf (Blum et al., 1995; Morton et al., 1996).

In the Colorado River, Texas, deposition of the youngest Deweyville alloformation from 20-14 Ka was followed by bedrock valley incision 14-12 Ka, with Holocene valley filling since (Blum and Price, 1998). Waters and Nordt (1995) found that the lower Brazos River, Texas was a competent meandering stream from 18 to 8.5 Ka, leaving thick coarse lateral accretion deposits (such as those associated with Deweyville terraces) as it migrated across the floodplain. The transition to an underfit stream incised into those deposits and dominated by vertical accretion is dated to 8.5 Ka, with avulsions in narrow and unstable meander belts occurring on several occasions since (Waters and Nordt, 1995).

Morton et al.'s (1996) analysis implies Trinity River incision sometime after about 13 Ka, with aggradation triggered by sea level rise and progressive onlap and burial of Deweyville surfaces sometime during isotope stage 1, from about 10 Ka. This is consistent with analyses of offshore and estuarine sediments, which indicate that Galveston Bay began forming initially by flooding of incised valleys about 8 Ka, with subsequent, apparently rapid inundation of valleys creating the approximate modern version of Galveston Bay about 4 Ka (Anderson et al., 1992). Rodriguez et al. (2005) identified flooding surfaces in Galveston Bay from decreases in sedimentation rates and changes from delta plain to central estuarine basin facies in cores. Formation of these surfaces dates to 8.2 and 7.7 Ka, at depths matching the elevations of relatively flat alluvial terraces.

The Critical Zone

The critical zone is a boundary between different channel responses, channel and valley morphologies, and sediment transport and storage regimes. Studies of channel morphological responses of the Trinity River to Livingston Dam, built in 1968, were reported by Phillips et al. (2005). Seven cross-sections from just downstream of the dam to Romayor, about 52 km downstream, showed morphological evidence of channel scour and/or widening in response to the dam. Resurveyed bridge cross-sections at three sites also showed the scour effects. From the Romayor site, exposed bedrock in the channel, indicating recent scour, can be observed in the channel a short distance downstream of Romayor. However, no such evidence is visible at a cross-section examined in the field for this study, 8 km downstream of Romayor. At 10 cross-sections between Romayor and Trinity Bay, including two resurveyed bridge crossings, no morphological response to the dam was observed (Phillips et al., 2005). No evidence of incision was noted, and lateral channel change was associated with point bar-cutbank pairs on migrating meanders.

Analysis of suspended sediment transport data from gaging stations at Romayor, about 8 km upstream of the critical zone, and Liberty, about 45 km downstream, show pronounced differences in sediment transport regimes (Phillips et al., 2004). Mean annual sediment yield at Romayor is nearly 3.4 million t yr^{-1}, with a specific yield of 76 t km^{-2} yr^{-1}. At Liberty, by contrast, the numbers are less than 69,000 t yr^{-1} and 1.6 t km-2 yr^{-1}. Additionally, while the Romayor station shows a clear reduction in sediment transport following closure of the Livingston Dam, there is no evidence of any change at Liberty (Phillips et al., 2004). Downstream of Liberty low stream power and ample accomodation space creates a sediment storage bottleneck such that little upstream sediment was reaching the lower reaches of the river even before the dam was constructed. Phillips et al. (2004) pinpointed the transition in sediment storage regimes at what is called the critical zone in this paper, just downstream of a Deweyville palaeomeander scar, at a point where floodplain elevation generally decreases, width increases, and numerous modern oxbow lakes appear (Figure 1).

The reaches up- and downstream of the critical zone also differ significantly in sinuousity, slope, and stream power. Cross-sectional stream power at any given reference flow is 4.5 to 33 times greater at Romayor compared to Liberty, despite the higher discharges downstream, and unit stream power is 20 to 100 times higher upstream of the critical zone (Phillips and Slattery, 2006). The difference is mainly attributable to slope, as channel bed slopes are 25 times steeper upstream of Romayor.

Methods and data sources

The critical zone is about 7-8 km downstream of the highway 787 crossing of the Trinity River near Romayor. In addition to cross-sections analyzed in previous work, the identified transition zone was visited in the field. The site was evaluated for evidence of scour in the form of exposed bedrock in the channel, and bank erosion other than in cut banks opposite point bars.

Geologic maps at the 1:250,000 scale are available (Houston and Beaumont sheets; Barnes, 1982; 1992). These maps distinguish between modern alluvium, the late Pleistocene Deweyville formation (including high-level deposits), and older coastal plain formations, including the Beaumont and Lissie formations (Pleistocene).

As resolution of the geologic mapping is relatively coarse, the soil survey of Liberty County, Texas (Griffith, 1996) was also used. Soils are mapped at a 1:24,000 scale. Soil series in the river valleys are associated with geologic formations and landscape surfaces (Aronow, 1996). Using the Liberty survey and the Official Series Descriptions (OSD) database of the U.S. Department of Agriculture (Soil Survey Staff, 2005), eight soil series were identified as occurring on the contemporary active floodplain. Soils mapped in Liberty County were considered to occur (not necessarily exclusively) on Pleistocene alluvial terraces if they were identified as occurring on Pleistocene alluvial or fluvial terraces, on fluvial terraces and adjacent uplands, or on alluvial sediments on uplands. Six series met this criteria. A 1:24,000 scale map was then produced of the transition zone area and adjacent up and downstream portions of the valley, aggregating the mapped soils into Pleistocene terrace and Holocene alluvial soils.

Digital orthophotography quadrangle images (flown in 1994) were obtained from the Texas Natural Resources Information Service for the entire lower Trinity valley. These were used for identifying oxbow lakes and other general morphological and land use features. Their primary use, however, was in identifying and mapping the large paleochannels which occur on many of the Deweyville surfaces.

Digital elevation data at 30-m resolution was also obtained through the U.S. Geological Survey and analysed using the LandSerf and RiverTools programs. Shaded relief models were visually evaluated for topographic evidence of palaeoshorelines, terraces, and other relevant features. Cross-valley and down-valley profiles were examined to identify terrace surfaces, and the digital elevation models (DEM) were "flooded" by raising water levels to arbitrary data to identify locations of potentially rapid response to rising base levels.

Results

Soils and Geology

The available geologic maps (Barnes, 1982; 1992) distinguish between two levels of Deweyville Terrace, shown in Fig. 3. No obvious transition is evident in the vicinity of the critical zone. Note also that the Trinity Valley, incised into the Beaumont formation, is, if anything, wider upstream than downstream of the critical zone.

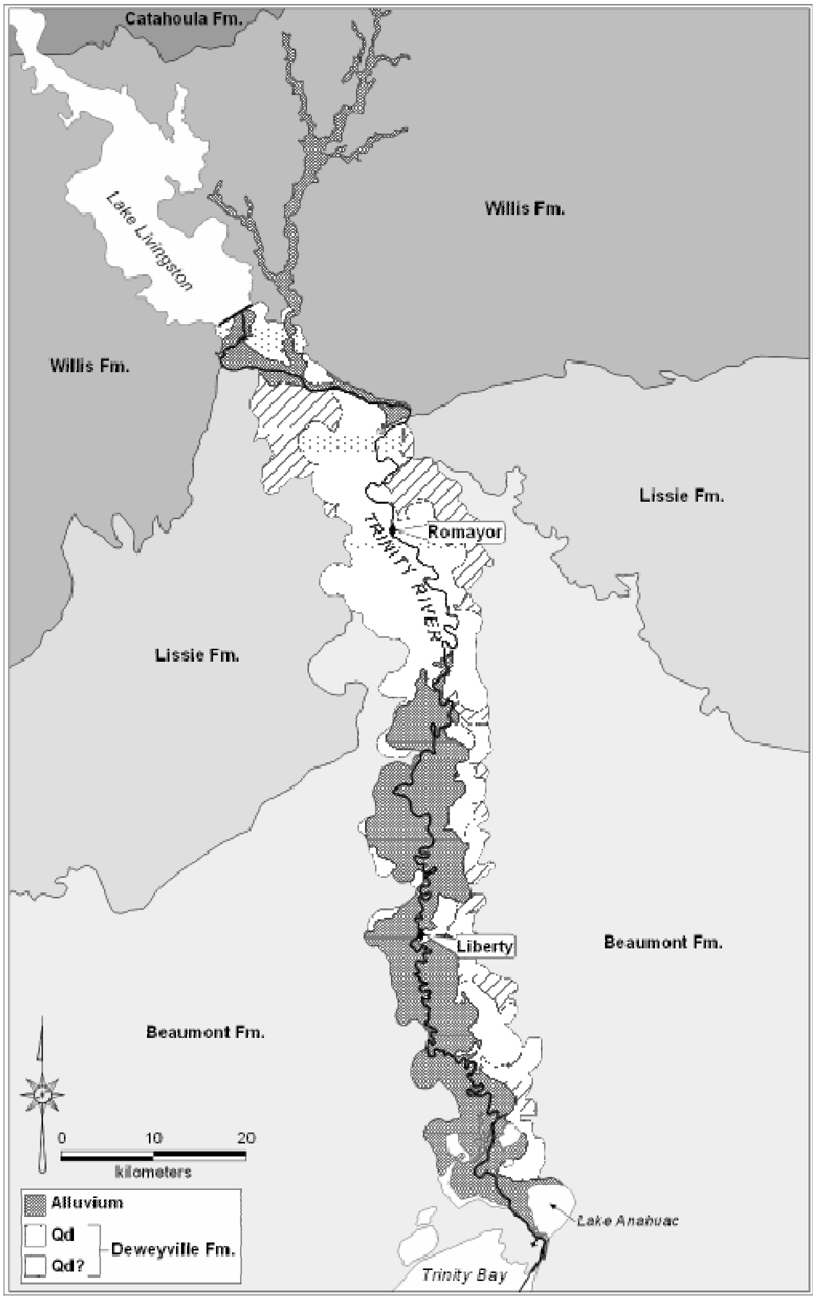

Figure 3. Generalized geology of the Trinity River Valley. The Willis, Lissie, and Beaumont formations are early, middle, and late Pleistocene, respectively. The late Pleistocene Deweyville formations are discussed in the text.

The alluvial soils mapped in the lower Trinity River valley upstream of the delta are shown in Table 1. In general, soils on the Beaumont and older alluvial and marine terrace surfaces are identified as uplands, flatwoods, and coastal prairie (Griffith, 1996). Soils identified as occurring on stream terraces are chiefly within the incised valley of the Trinity River or larger tributaries. Floodplain soils are mainly Entisols with minimal pedogenic development, with the exception of Owentown series, which exhibits enough cambic B horizon development to be classified as an Inceptisol, and the Kaman series, a Vertisol whose diagnostic properties are dominated by high amounts of smectitic clays. Terrace soils are more strongly developed, and are mainly Alfisols with argillic horizons. The exception is the Alaga series, which is widely mapped in sandy marine, coastal, and alluvial deposits throughout the U.S. Atlantic and Gulf coastal plain.

Table I. Alluvial Soils of the Lower Trinity River Valley. Series names are given, along with the U.S. Soil Taxonomy, and the description of their landscape position in the Liberty County, Texas soil survey).

Modern Alluvial Floodplain

Estes: fine, montmorillonitic, acid, thermic, Aeric Fluvaquents
- floodplains

Fausse: very-fine, montmorillonitic, nonacid, thermic, Typic Fluvaquents
- low backswamps and on remnants of oxbows along floodplains

Hatliff: coarse-loamy, siliceous, nonacid, thermic Aquic Udifluvents
- floodplains

Kaman: fine, montmorillonitc, thermic, Typic Pelluderts
- floodplains

Mantachie: fine-loamy, siliceious, acid, thermic Aeric Fluvaquents
- floodplains

Owentown: coarse-loamy, siliceous, thermic, Fluventic Dystrochrepts
- floodplains

Pluck: fine-loamy, siliceous, nonacid, thermic, Typic Fluvaquents
- floodplains

Voss: Mixed, thermic, Aquic Udipsamments
- floodplains

Alluvial Terrraces

Alaga: thermic, coated Typic Quartzipsamments

•ridges of stream terraces along the floodplain

Aris: fine, montmorillonitic, thermic, Typic Glossaqualfs
•broad flats along drainageways; depressions and remnant drainageways

Bienville: sandy, siliceous, thermic, Psammentic Paleudalfs
•stream terraces

Kenefick: fine-loamy, siliceous, thermic, Ultic Hapludalfs
•stream terraces

Landman: loamy, siliceous, thermic, Grossarenic Paleudalfs
•uplands and stream terraces

Spurger: fine, mixed, thermic, Albaquultic Hapludalfs
•low ridges of stream terraces along the floodplains

The Fausse series is formed in recently-deposited sediments, but these are generally in depressions associated with Deweyville paleomeanders rather than modern oxbows. The latter, if filled, are most commonly mapped as Mantachie. The Voss soil series is associated with active or stabilized sandy point bars. Lower-elevation and backswamp areas of the modern floodplain are typically occupied by the Kaman series, while natural levees and other slightly higher areas are generally Owentown. Some of these soils, particularly the Kaman, are also found in depressions on alluvial terraces. The Fausse, Kaman, Mantachie, Owentown, and Voss dominate the floodplain soil geography in the vicinity of the critical zone. In the same vicinity Spurger soils generally demarcate relict meander scars of Pleistocence terraces within the incised valley.

The soil geography of the critical zone area allows discrimination into modern floodplain and (Deweyville) terrace surfaces, as shown in Figure 4. This mapping derived from the soil survey maps reflects the modern alluvium inset within the terrace deposits, and shows a pronounced widening of the modern floodplain at the critical zone.

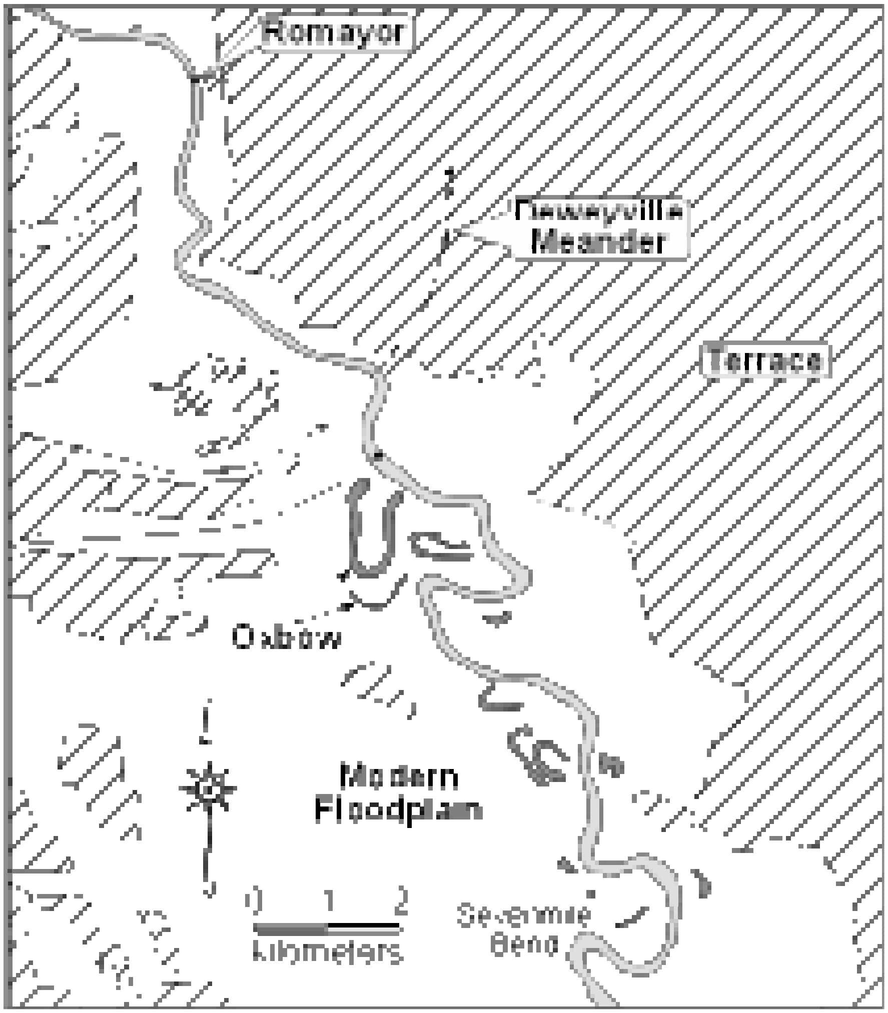

Figure 4. Pleistocene terrraces and the modern river, floodplain, and oxbows in the vicinity of the critical zone.

Both the detailed (1:24,000) and general (1:316,800) soil maps were examined for evidence of more-or-less linear trends in the distribution of sandy soils on uplands whose trend would bisect the river valley, as evidence of possible paleoshorelines and beach ridges. None were found.

Topography

The general relief shown in the location map (Fig. 1) shows the valley incised into the Beaumont surface, with a variety of topographic levels within the valley. Three elevation transects oriented down the axis of the lower Trinity valley (Fig. 5) show the general trends in elevation. The profiles (Fig. 6) show low points at river and tributary channel crossings. High areas represent remnants of Deweyville terrace surfaces, an interpretation confirmed in the field at several locations by the presence of rounded fluvial gravels common in the Deweyville deposits. Singular high points, appearing as spikes in the profiles, are associated with causeways for roadways and railways.

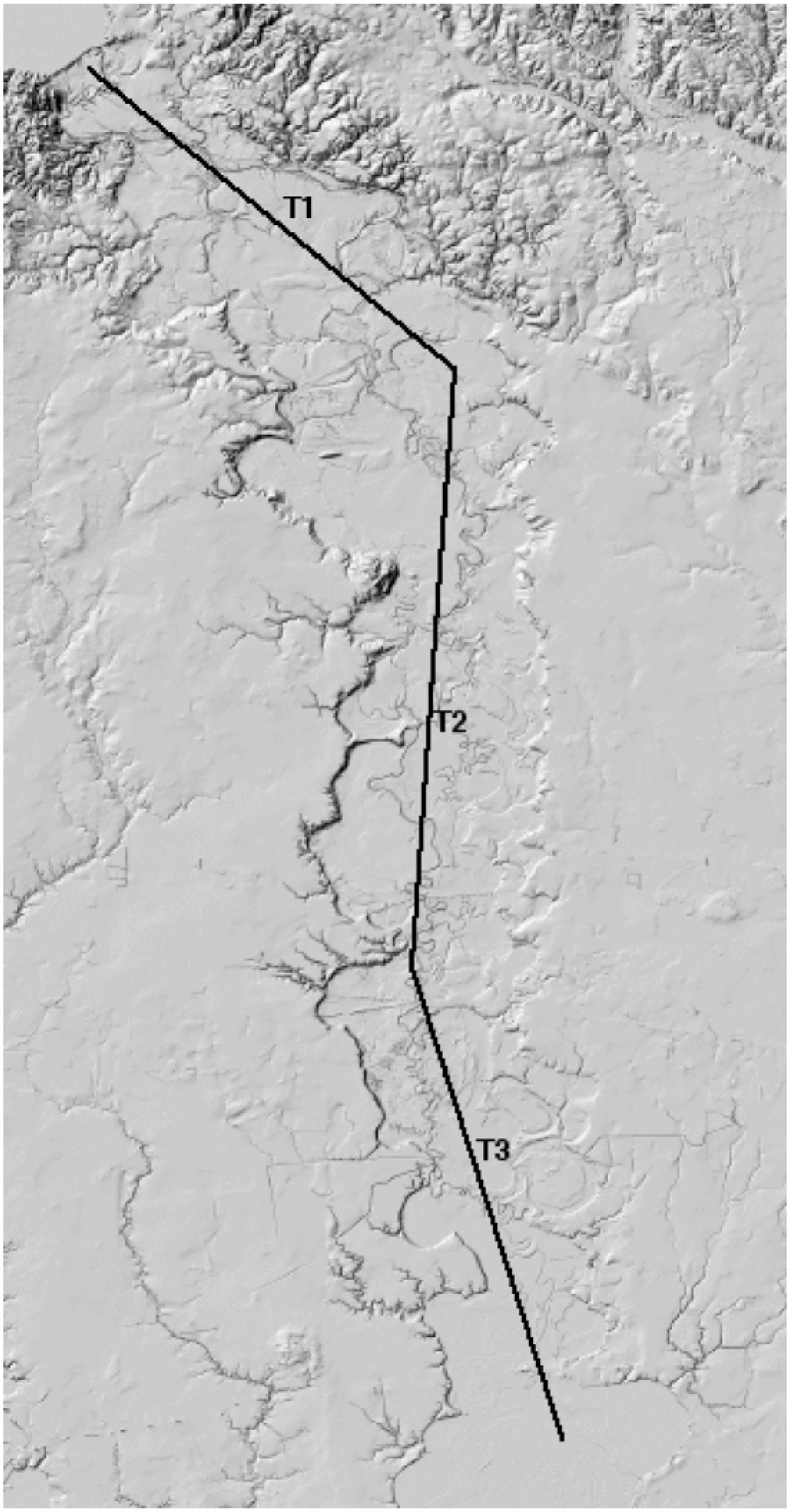

Figure 5. Location of down-valley topographic transects.

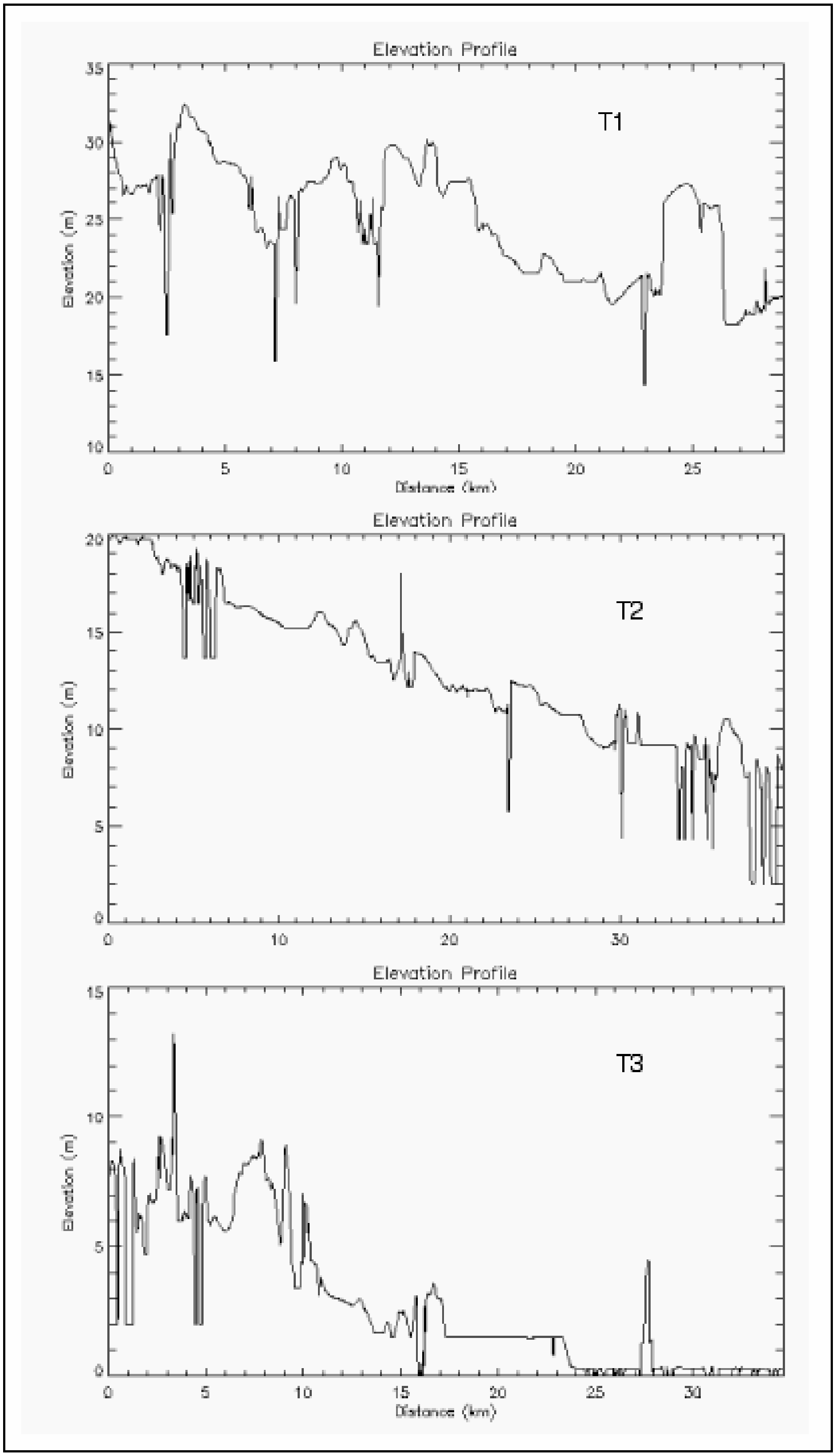

Figure 6. Downvalley elevation transects (see figure 5 for locations). Note differences in vertical scale.

In profile T1 (fig. 6) there are four discernable surfaces. While most of the lower Trinity valley is incised into the Beaumont formation, in the area downstream of Lake Livingston the Lissie formation comprises the valley walls, as shown in the geologic map (Figure 3), and showing up as the dissected terrain in the shaded relief (Fig. 1). The highest surface (L1), sloping from about 33 m elevation to about 27 m, is interpreted as the Beaumont terrace. A lowermost surface (L4), into which the channels are cut, is about 27 m in elevation at the upper end of the profile, to about 18 m at the lower. Channel margin levee deposits are evident adjacent to channels on this surface, which includes the modern floodplain and portions of the lowermost Deweyville terrace. The latter often merges subtly with the modern floodplain sediments. The L2 and L3 levels represent the upper and middle Deweyville terrace levels. This interpretation was confirmed in the field at a number of locations by the presence of rounded fluvial gravels common in Deweyville deposits. The large-magnitude paleomeanders are cut to L4, and appear to be incised into L3 and cut laterally in some cases into L2.

Profile T2 (fig. 6) shows continuations of the L3 and L4 surfaces. Remants of the L2 surface are present in this reach of the valley near the valley sides. The L3, L4 surfaces also project into profile T3, with two additional surfaces appearing in the lowermost part of the transect, associated with the upper Trinity River delta (from about 13 to 24 km) and the delta marshes and tidal flats (from 24 km).

A number of cross-valley profiles (Figs. 7 and 8) show the steep valley walls cut into the Lissie and Beaumont formations (Fig. 3). The L1 – L4 surfaces are evident to varying extents in these profiles.

Paleomeanders

The large-magnitude paleomeanders are observed on topographic maps, digital elevation models, and are evident in soil surveys. They tend to be particularly evident on color aerial photography, however, as the distinctive shape, size, and soil, hydrologic, and vegetation patterns are readily apparent. Paleomeanders in the vicinity of the critical zone were mapped from the digital orthophotoquads, and checked against topographic and soil map information.

Figure 9 shows the contemporary river, paleomeanders, and modern oxbows from Livingston Dam through the critical zone While paleomeanders exist both up and downstream of the critical zone (Lake Anahuac, at the head of Trinity Bay, is a flooded paleomeander), the only case where one of the paleomeanders is observed to transect the river is at the critical zone. Several meander scars with sizes consistent with modern river exist on the Beaumont surface in the upper portions of the study area. Beginning where the valley is cut into the Beaumont formation, the larger paleomeanders appear. However the critical zone is the only location between the lake and the bay where one of these meanders is bisected by the modern river.

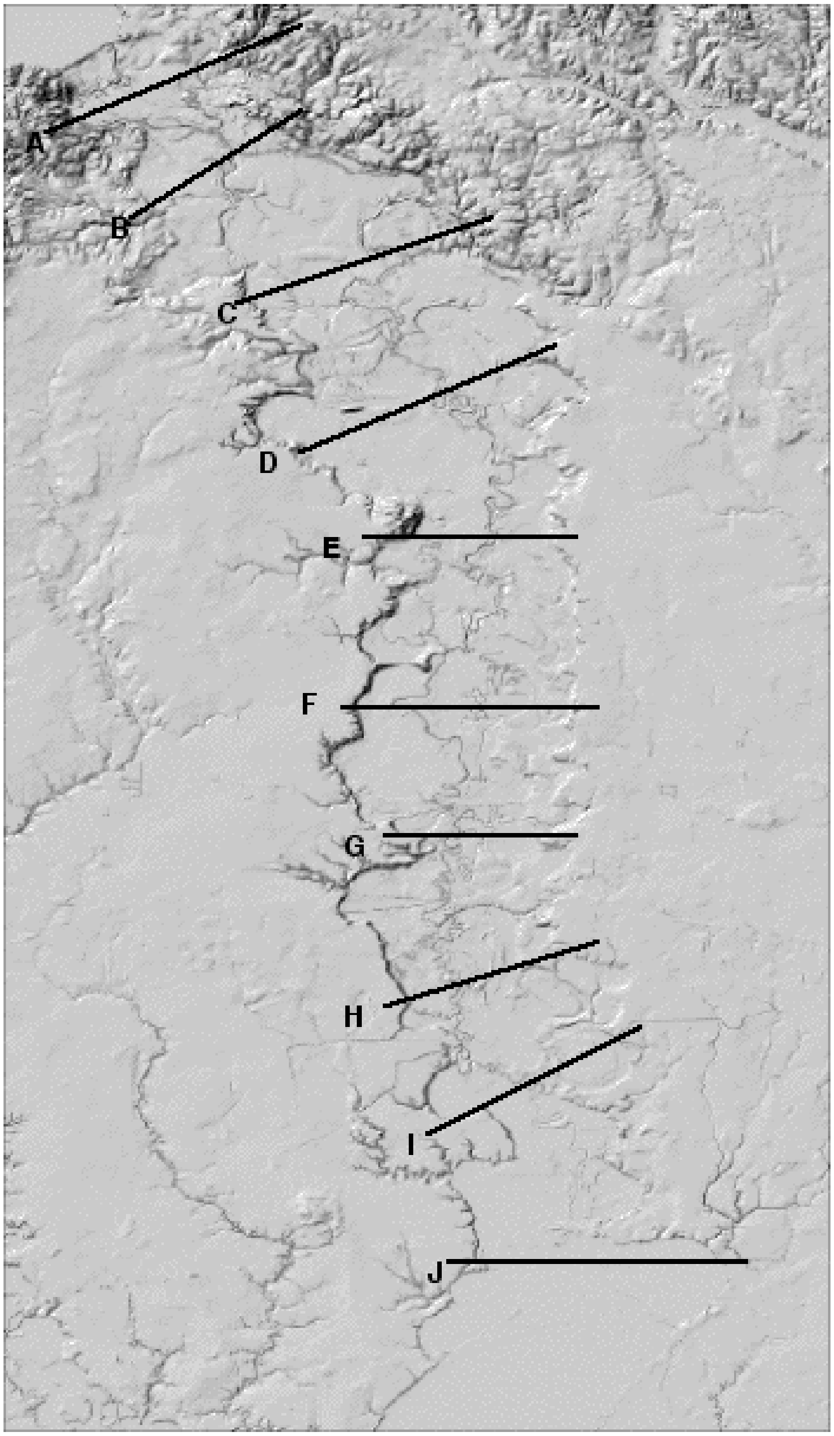

Figure 7. Location of cross-valley transects.

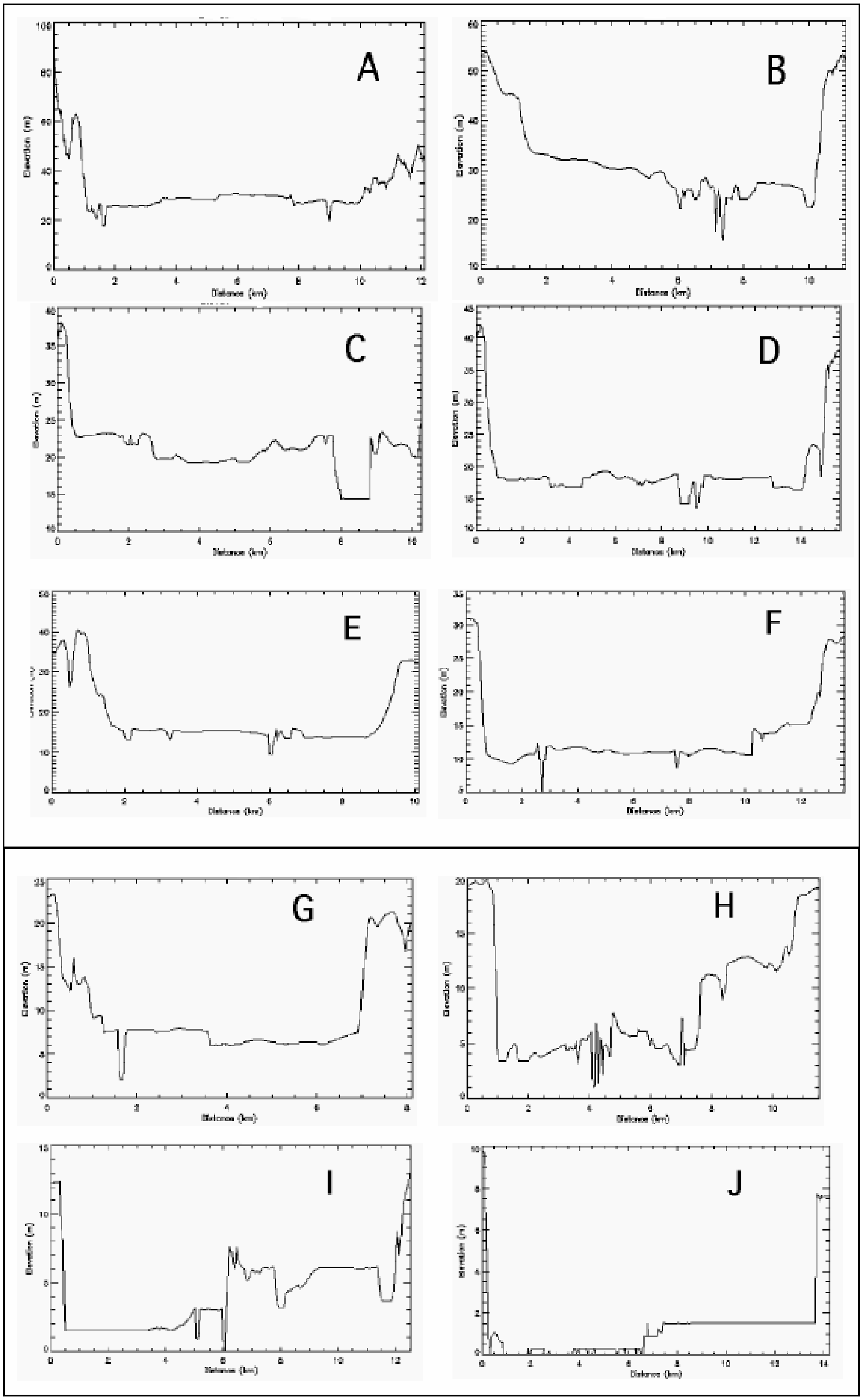

Figure 8. Cross-valley transects A-J (see Fig. 7 for location). Note differences in vertical scale.

Elevation profiles (Fig. 8) show that the paleomeanders are consistently at levels above that of the modern river channel. Paleomeanders upstream of Romayor (Fig. 9) have channel elevations at $\geq$19 m, and are about 5 m above the level of the river channel nearby. The gooseneck paleomeanders downstream of Romayor, by contrast, are 1 to 2 m lower in elevation, and 3 to 4 m above the modern river channel. This, along with the geometry of the paleochannels, suggests at least two different meander systems preserved in the alluvial terraces.

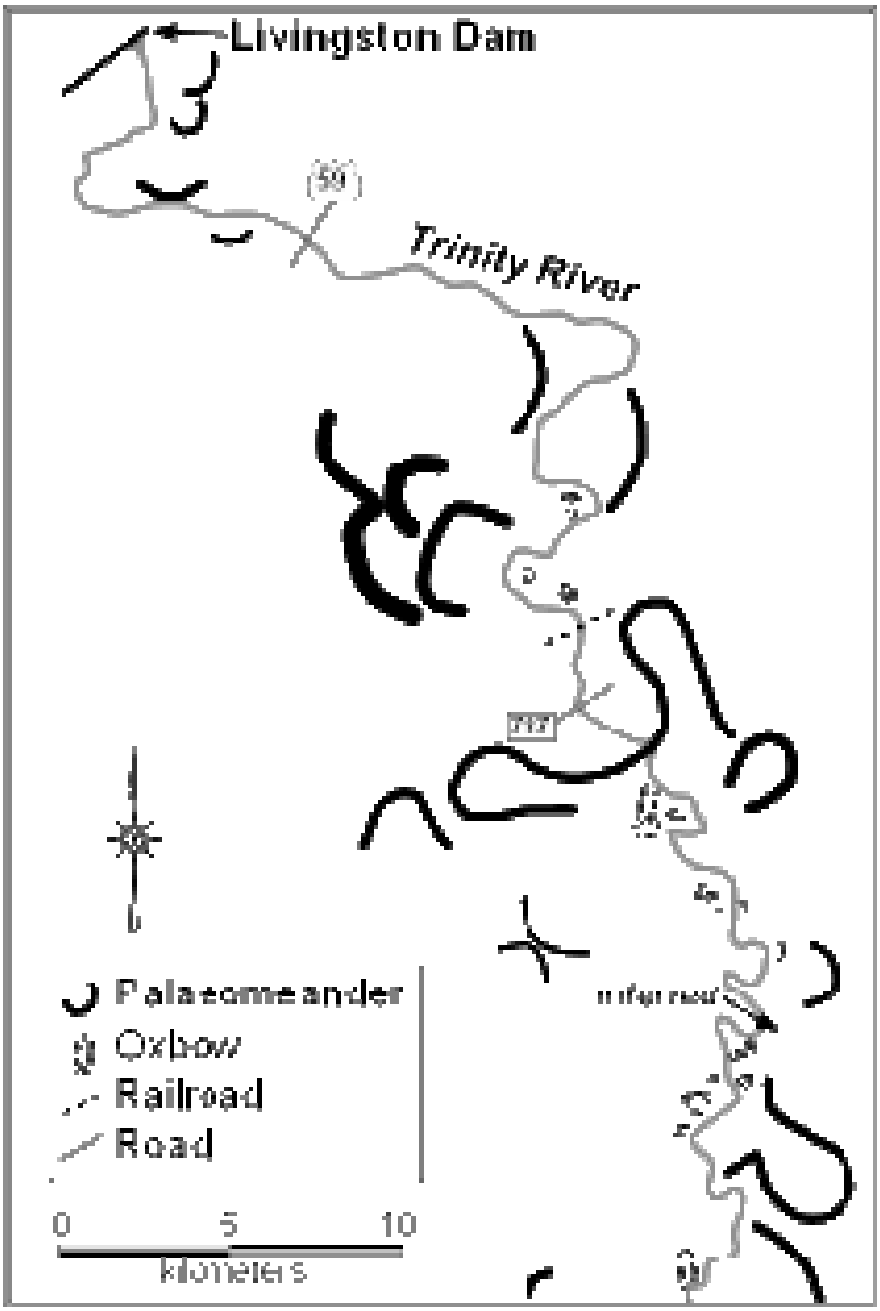

Figure 9. Large-magnitude Pleistocene palaeomeanders, mapped from digital orthophotographs, from Lake Livingston to downstream of the critical zone.

Discussion and interpretations

For about 50 km downstream of Lake Livingston, scour since the construction of Livingston Dam has cut down to more resistant pre-Quaternary clays and sandstones. Other than this, however, there are no significant resistance constraints in the lower Trinity, as the entire valley is inset into alluvial, coastal, and marine unconsolidated sediments. The inherited morphology from previous episodes of lateral migration, aggradation, and degradation, however, provides significant influences and controls on the modern river, however.

Morton et al. (1996) found that coastal plain rivers of southeast Texas tend to be entrenched at three levels, consistent with the results of this study, with the youngest terrace controlling gradients, patterns, and locations of modern channels.

Proceeding upstream from Trinity Bay, up to a short distance upstream of Moss Bluff the Deweyville surfaces are onlapped by modern floodplain and delta sediments. This is also marked by a transition from a convergent to a divergent, distributary channel network. For rivers in the region generally, Morton et al. (1996) found that the onlap position also marked a transition in lateral migration style. Downstream, channels are sinuous but relatively stable in position, with lateral accretion accomplished by fine-grained overbank sedimentation on meander beds. Upstream of the onlap position there is more active channel migration with lateral accretion on sandy point bars (Morton et al., 1996). Results in the Trinity are consistent with that trend.

The critical zone marks a second transition or hinge point, and results of this study indicate an association with drowning of alluvial terrace surfaces. Both the lowest/youngest and intermediate Deweyville surfaces are evident throughout the lower Trinity, with the oldest/highest present mainly on the valley edges downstream of the critical zone, and more prevalent upstream.

Sinuousity (channel length divided by valley length) from the Livingston Dam to the critical zone is 1.13, and increases to 1.61 from there to Liberty (Phillips et al., 2005). This decrease in sinuousity in the upstream direction—consistent with channel response to a rise in base level—also corresponds to the section of the valley where the L2 (upper Deweyville surface) becomes more prominent in mid-valley, and to the only location where a Deweyville meander (cut down to L4, incised into L3, and cut laterally into L2) can be clearly traced across the modern channel.

The location of the critical zone is therefore interpreted to be associated with the encroachment of sea-level-driven effects on channel and valley morphology. The transitional zone coincides with the encroachment of the aggrading Holocene river onto L2, the upper Deweyville allformation. Rodriguez et al. (2005) recognize relatively rapid changes associated with flooding of terrace surfaces in the lowermost onlap reaches. These are presumably associated with comparably rapid upstream translation of geomorphic effects. A relatively sudden (in the geological sense) upstream transgression of the critical zone is likely when the upper Deweyville surface is breached.

The critical zone is not associated in any obvious way with any evident paleoshorelines or escarpments, but Pleistocene beach ridges and dune fields, other than the Ingleside Barrier in the vicinity of Galveston Bay (see fig. 2) have have not been identified in this area, making identification of paleoshorelines difficult.

The two hinge points in the lower Trinity mark transitions in river channel and valley morphology and in dominant processes which are further associated with variations in aquatic and riparian ecosystems, channel change, and fluvial response to human modifications. The results of this study studies of fluvial/alluvial landscape evolution over Quaternary and Holocene time scales may be highly relevant to river resource management, particularly in the context of defining critical transition zones or points, and in predicting limits to upstream or downstream propagation of disturbances.

Conclusions

The critical zone of the lower Trinity River marks an important transition in river channel and valley forms, dominant processes, and resulting geomorphological, hydrological, and ecological characteristics. The location of this zone is not a circumstantial result of a given rate of up- or downstream propagation of effects. Rather, the location marks the contemporary upstream extent of the effects of Holocene sea level rise on the Trinity River. This in turn coincides with the point at which the Pleistocene upper Deweyville alluvial terrace surface is encountered. A more rapid rate of change and relatively sudden upstream disciplacement of this zone is likely when the upper Deweyville surface is flooded.

Antecedent fluvial and alluvial topography inherited from previous aggradation, degradation, and lateral migration episodes is likely to be an important control over modern fluvial forms and processes in other alluvial coastal plain rivers as well. Identification and mapping of such features may be extremely useful in pinpointing critical transition zones for water resource managers.

Acknowledgements

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Part 4

Channel adjustments of tributary streams within the lower Trinity River basin, Texas

Introduction

The understanding of the impacts and effects dams may have on a fluvial system has been significant and recognized for more than 50 years (Petts and Gurnell, 2005). World-wide dam building accelerated rapidly in the 1950s and peaked in the late 1960s (Beaumont, 1978; Graf, 2005; Petts and Gurnell, 2005). The effects of large dams became more obvious in the 1970s and 1980s as geomorphological research slowly and steadily increased. Research today concerning dams has progressed to the stage of theory-building (Graf, 2005).

According to Petts and Gurnell (2005), research on the downstream effects of dams often focuses on three themes: 1) channel dynamics, 2) the role of riparian vegetation, and 3) channel change causing ecological change. Within most studies, the downstream effect on tributaries gets little attention beyond the confluence with the impounded stream.

Confluences and the reach within a trunk stream that a confluence occurs have been recognized as areas that are sensitive to system perturbations and may illustrate the direction of change within that system (Andrews, 1986; Church, 1995; Harvey, 2002; Benda et al., 2004). At confluences the impounded stream's channel response controls the effect to the tributary. A tributary stream's base level is the channel bottom of the trunk stream. Any change in hydraulic geometry (widening, narrowing, aggrading or degrading), or planform change in the trunk stream will effect the base level of a tributary. Mainstem channel change that has altered base level in tributaries has caused the upstream migration (in tributaries) of knickpoints, entrenchment, bankfull width increases and channel caving (Germanoski and Ritter, 1988; Kesel and Yodis, 1992).

The effects of tributary base level lowering are often considered a coupling effect-a change in the mainstem translated to the tributary. Coupling effects between trunk and tributary streams may be a key control of the geomorphic function of a river system (Brierley and Fryirs, 1999). Upstream coupling caused by channel adjustment in the mainstem (Galay, 1983) can often cause a desynchronization of flood hydrographs in the tributary and trunk stream system (Willis and Griggs, 2003). Desynchronized hydrographs can produce a situation in which the tributary peaks before the trunk and subsequently transports exuberant amounts of sediment to the confluence. When the tributary reaches its mouth, with higher stream power than the trunk stream, coarser particles are deposited and over successive events a delta may form (Topping et al., 2000; Willis and Griggs, 2003).

The spatial and temporal extent of a perturbation throughout a fluvial system will depend upon three factors; distance decay, response propagation rates, and landscape sensitivity. The distance decay and response propagation rates of dam effects within tributary streams will be controlled by coupling processes from the impounded-trunk stream, as well as other influences on channel morphology. Distance decay describes the spatial extent of a system perturbation as changes associated with that particular disruption become less and less detectable with greater distance from the source. A system is likely to become more insensitive to a perturbation with increased resistance due to dampening affects attributed to distance away from the source of a disturbance, or reach a point at which overriding system components overwhelm any influence of the perturbation. Change in 'local' (e.g. a mainstem stream to a tributary) base level is a geomorphic control that has been shown to dampen out with distance (Leopold and Bull, 1979). Therefore the subsequent responses within a tributary system to the effects of downcutting within a trunk stream (causing a drop in base level in a tributary) would become less detectable further from the perturbation (e.g. the confluence).

Working in the Brazos River system, Nordt (2004) showed that tributary streams, rather than mainstem streams, are more sensitive to climate change and fluctuating sediment supply; suggesting that low order streams and associated trunk streams may not respond similarly to geomorphic change. Similarly, as dams cause changes in sediment supply and flow conditions and associated geomorphic changes to downstream reaches, tributary systems should not be expected to respond in the same way as a trunk stream.

The response of tributary systems downstream of an impoundment will vary depending upon many contingencies. The relative sizes of the tributary/trunk stream system, distance from and time since the perturbation, and upstream/downstream coupling processes will affect morphological and flow conditions within the tributary streams. The response of tributary systems below an impoundment may be more strongly influenced by varying local conditions and historical contingencies such as land use/cover, vegetation and geology.

Regional setting

Glacial-eustatic cycles have played a particularly influential role in sea-level effects on Texas coastal plain rivers. While many of these rivers are likely still responding to eustatic cycles, it appears that the reaction of coastal plain rivers to natural processes is outpaced by anthropogenic alterations to the landscape (such as impoundments and fluid withdrawal) (Morton and Purcell, 2001).

Numerous studies have documented the coastal plain evolution of rivers within Texas through the Holocene (Blum and Price, 1998; Rodriguez et al., 1998; Anderson and Rodriguez, 2000; Rodriguez and Anderson, 2000; Rodriguez et al., 2001; Rodriguez et al., 2000ab), with contemporary studies focusing on sedimentation rates (Longley et al., 1994; White et al., 2002), fluvial-coastal systems (Giardino et al., 1995), and sediment transport/residence time (Hudson and Mossa, 1997; Phillips, 2001a; Phillips and Marion, 2001; Yeager et al., 2002; Phillips, 2003a). Previous work on the modern Trinity River system includes sedimentological studies that focused on wetlands in the fluvial-deltaic area (Morton and Paine, 1990; White and Calnan, 1991; Solis et al., 1994; Rodriquez and Anderson, 2000; White et al., 2002).

Within the lower Trinity basin (Figure 1), studies focusing on dam related affects have shown a notable geomorphic impact for at least 60 km downstream of Lake Livingston. Between this reach and Trinity Bay an apparent sediment "bottleneck" exists, seemingly buffering the delta/estuary system from upstream sediment regime changes (Phillips, 2003b; Phillips et al., 2004; Phillips and Slattery, 2006). This fluvial-estuary transition zone has been reworked numerous times through the Holocene (Anderson and Rodriquez, 2000) and has migrated the "mouth" of the river as much as 200 km in the upstream-downstream direction (Thomas and Anderson, 1994; Phillips et al., 2004; Phillips and Slattery, 2006).

The entire Trinity River drainage basin has an area of 46 100 km^2, with headwaters in north-central Texas (Figure 1). The four forks that combine to form the Trinity River flow through the Dallas-Fort Worth metroplex, through the piney woods of east Texas, finally draining to Trinity Bay. The seventh largest estuary in the United States (Pulich and White, 1991), the Galveston Bay system includes the Trinity Bay, which is the only natural bay-head delta (Trinity River) in Texas that has prograded in geologically recent times (White and Tremblay, 1995).

Closed 28 September 1968, Livingston Dam is a flow-through reservoir which functions primarily as a water supply for the city of Houston, Texas. The conservation pool capacity of the lake is greater than 2.2 billion m^3, with a capacity/inflow ratio of 0.316, based on the conservation pool capacity and an extrapolation of mean annual flow per unit drainage area for the Crockett gauging station (Phillips and Musselman, 2003). Located approximately 175 km above the Trinity Bay, Lake Livingston impounds 95 percent (42 950 km^2) of the Trinity River's drainage area. The upper Trinity basin (above Lake Livingston reservoir) has a total of twenty-nine dams, concentrated around the Dallas-Fort Worth metroplex, which are managed for flood remediation (Wellmeyer et al., 2005).

Livingston Dam has been shown to have a minimal impact on the downstream flow regime of the Trinity River (Wellmeyer et al., 2005). While no changes in high flow conditions exist following impoundment, low flows have been shown to be slightly elevated. The post-dam period however, is characterized by considerably higher amounts of precipitation and might be masking the complete impact of flow regulation (Wellmeyer et al., 2005).

Cross-sectional morphological changes were investigated in the lower basin by Phillips et al. (2005). While high and moderate flows were not altered by the dam (Wellmeyer et al., 2005), sediment transport was greatly affected. Livingston Dam has a trap efficiency of 81 percent, based on the curve of Brune (1953). The principal sources of evidence used to determine channel adjustments included resurveys of channel cross-sections at highway bridge crossings, and field indicators of geomorphic change. The channel response, which is limited to about 60 km downstream of the dam, is characterized by incision, widening, coarsening of channel sediment and a decrease in channel slope (Phillips et al., 2005).

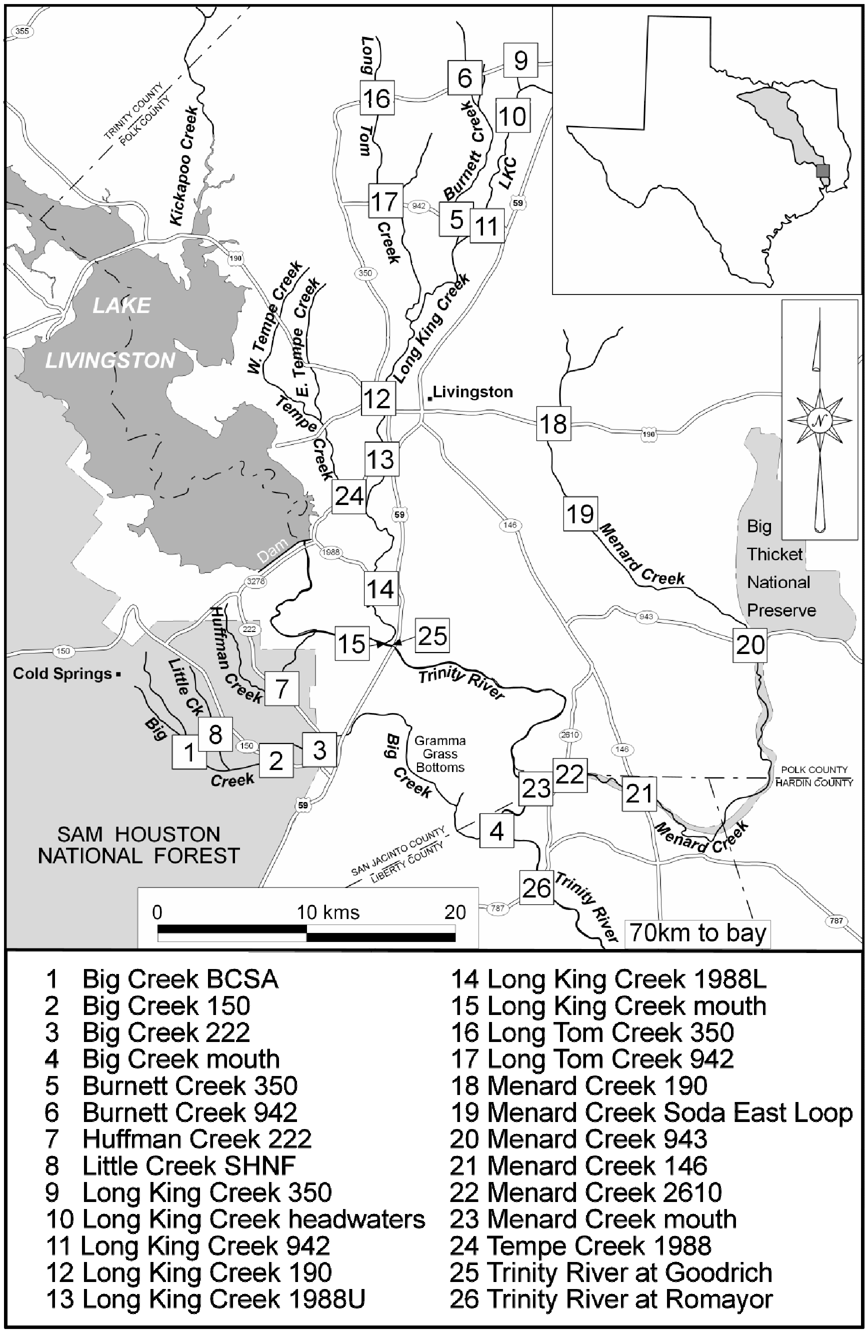

Figure 1. The Trinity River Basin, SE Texas, showing the study area immediately below Lake Livingston. USGS gauging stations are located at tributary sites 12 and 21 and Trinity sites 25 and 26.

Materials and methods

The purpose of this study is to describe and explain river channel cross-sectional change in tributary streams within the affected 60 km reach of the Trinity River, Texas, downstream of Livingston Dam (Figure 1). All four first-order tributaries investigated have their confluence with the Trinity within the 60 km reach affected by the dam.

The general pattern of system effects addressed in this study is the consistency of response (and direction) with respect to geomorphic change such as channel change (in width, depth, slope and roughness).

The typical cross-sectional and reach variability within a fluvial system may lead to variations in the quantitative rates and extent of changes. But, the more fundamental issue of modes of adjustment, defined here as qualitative combinations of increases, decreases, and negligible changes in hydraulic variables, allows for a more qualitative description of change in channel geometry and hydraulics, especially where a lack of baseline of data makes quantitative measurements of change impossible (Phillips et al., 2005).

The response of a river to a dam can be directly measured only if monitoring of the river occurred prior to dam construction; this is the case for the Trinity River system. The Livingston dam was constructed during a time when active USGS gauging stations were located above and below the impounded reach, as well as on two first-order tributaries in the lower basin. Figure 2 shows the mean annual discharge record for both tributary stations. Daily stage-discharge data was obtained from USGS gauging station records and used to construct flow duration curves (Figure 3). The Long King Creek and Menard Creek gauging stations do not have lengthy pre-dam records; they have been in operation since 1963 and 1965 respectively.

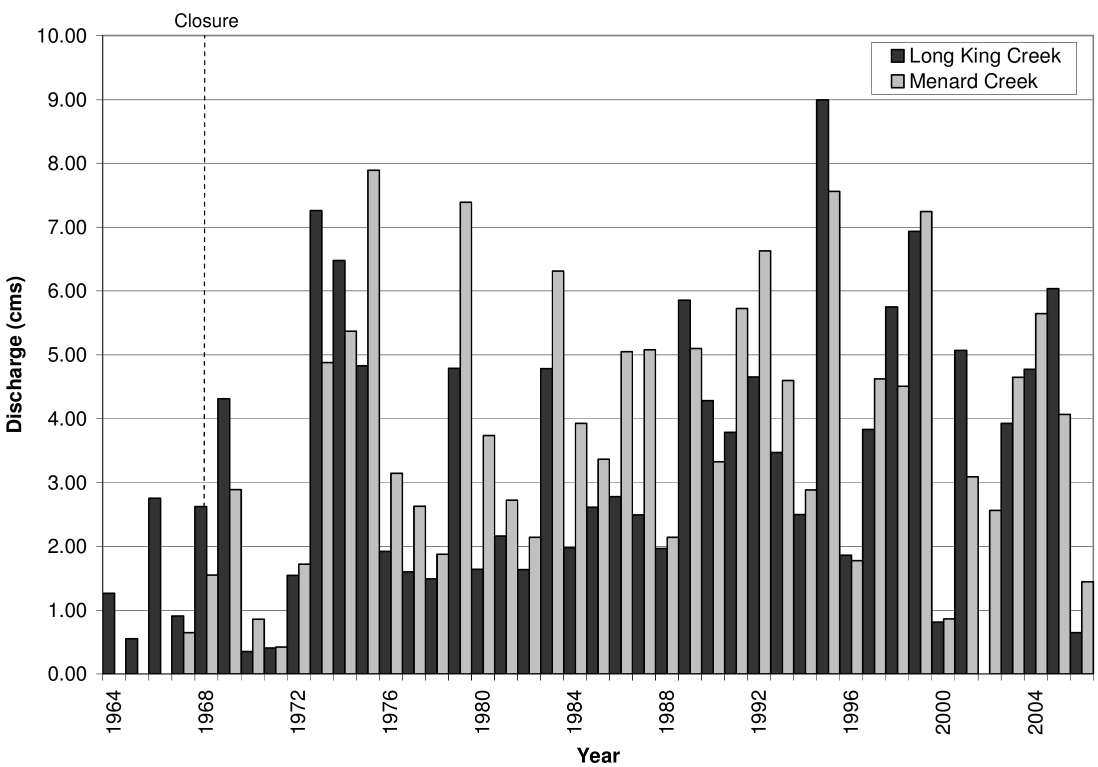

Figure 2. Mean annual discharge at the Long King Creek and Menard Creek gauging stations, 1964-2006.

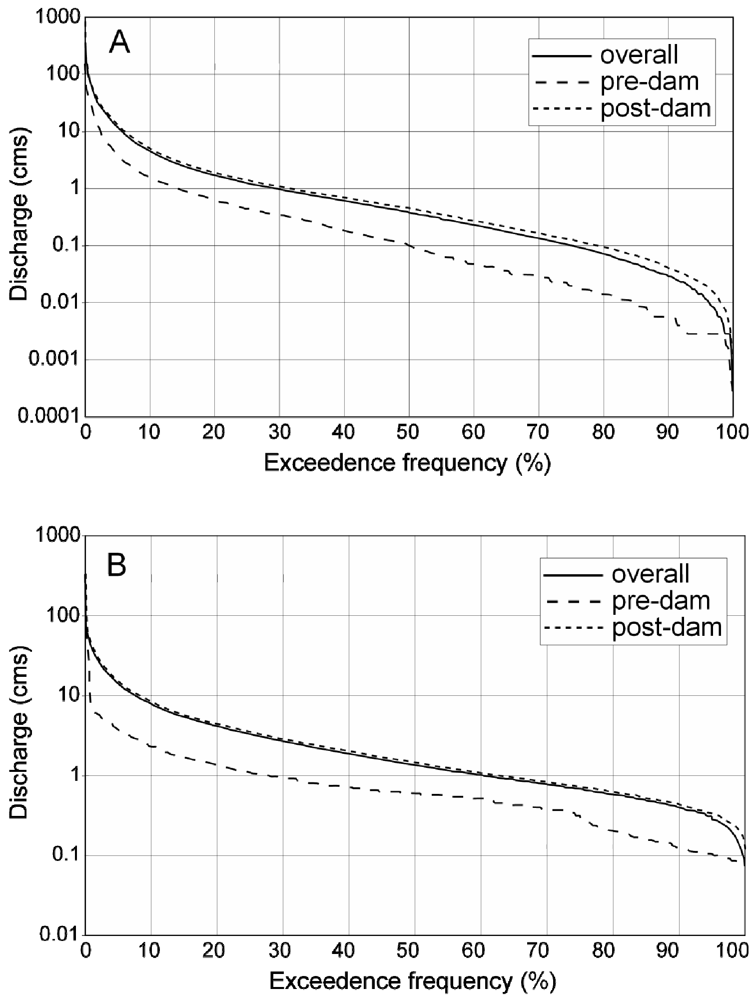

Figure 3. Flow duration curves at the Long King Creek (A) and Menard Creek (B) gauging stations. The curves show little change in flow regime following impoundment.

Channel adjustments within the tributaries was investigated through resurveying sites of historical surveys of channel cross-sections at highway bridge crossings, aerial photos, and field indicators of erosion, sedimentation and channel change.

Bridge Cross-sections

A series of cross-sections over successive years may provide a great deal of information about changing geomorphic and hydrologic characteristics of a stream (Kesel and Yodis, 1992; Yodis and Kesel, 1993). Phillips et al., (2005) used channel cross-sections surveyed from bridge crossings over the Trinity River to show a dynamic channel with multiple modes of adjustment. Using a similar technique, channel changes in width, depth and cross-sectional area were measured at 13 sites along seven tributaries.

Channel cross-sections for thirteen bridge crossings were obtained from the Texas Department of Transportation (TXDOT). Figure 4 shows five of the thirteen crossings. These five are shown because they illustrate the most dynamic crossings of the thirteen sites. The obtained data varied for each site, but all the crossings included at least 3 channel surveys that occurred between the years 1996 and 2002. Ten of the cross-sections were resurveyed in July 2003 using the same methods as employed by the bridge engineers, a weighted drop line. The three cross-sections that were not resurveyed in 2003 were judged in the field to have very little channel activity, with stable channel banks and floor and no detectable geomorphic change. The data from the cross-sections were then compared to determine changes in width, mean depth, max depth, width/depth ratio and cross-sectional area. The measurements were all relative to a banktop-to-banktop datum determined for each survey site.

Others have successfully used this technique of at-a-station hydraulic geometry change to document channel change through time (Phillips et al., 2005). Similarly, Kesel and Yodis (1992) and Yodis and Kesel (1993) have used historical channel surveys at bridge sites to show the impact of human modifications to two coastal plain rivers within southwestern Mississippi, USA. Typically bridge crossings would not necessarily be considered representative of a stream's behavior for numerous reasons, including a tendency to choose: 1) locally narrow reaches; 2) stable channels; and 3) stable floodplains when constructing bridges; and 4) due to the nature of bridge anatomy, scour tends to occur around pilings and bridge supports (Phillips et al., 2005). Considerations should also be given to the amount of disturbance that may occur during bridge construction and how this may affect subsequent creek surveys. Nevertheless, the bridge crossings do represent the only historical records of cross-sectional change.

While only one tributary bridge was constructed before impoundment of Lake Livingston, the remaining sites offer valuable insight into the recent morphological changes of the Trinity River tributaries. Changes in banktop-to-banktop width, mean depth and width/depth ratio also reveal the direction or mode of change in the tributaries.

Aerial photos

Historical aerial photos and satellite imagery are a valuable resource when describing planform change. In the lower Trinity basin, aerial photos were available from the United States Department of Agriculture, and digital orthographic quarter quadrangles from the Texas Natural Resources Information System.

Historic air photos, as well as more recent photos and imagery, may be used to study planform change through time. Other studies have conducted similar studies using a GIS approach (Downward et al., 1994; Winterbottom and Gilvear 2000; Simon et al., 2002; Wellmeyer et al., 2005). In this study, qualitative changes at the mouth of Long King Creek were determined by mapping changes from a series of aerial photos (Figure 5). Only one upstream location along LKC was investigated through aerial photo

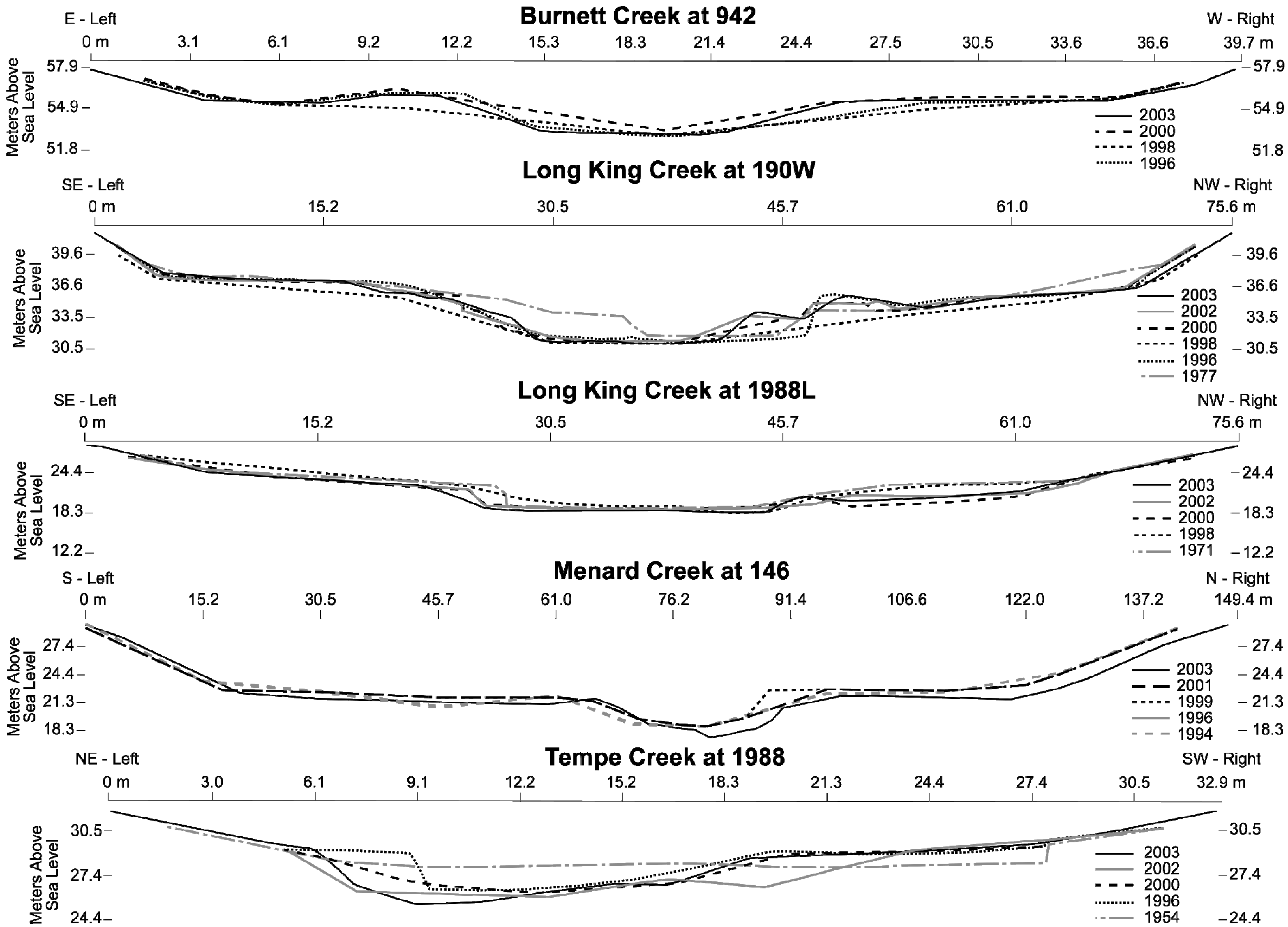

Fig. Bridge. Sample cross sections over the five most dynamic bridge crossings.

coverage (Figure 6). At this site a cutoff had created an oxbow sometime between 1968 and 1982. These were the only two sites selected for description for two reasons. First, the mouths of the tributaries are dynamic locations of change, and hold the greatest potential for dam influenced effects. Second, as a result of the imagery resolution, other sites within the study area are not discernible. Four individual years of photographic coverage were available: 1958, 1968, 1982, and 1995.

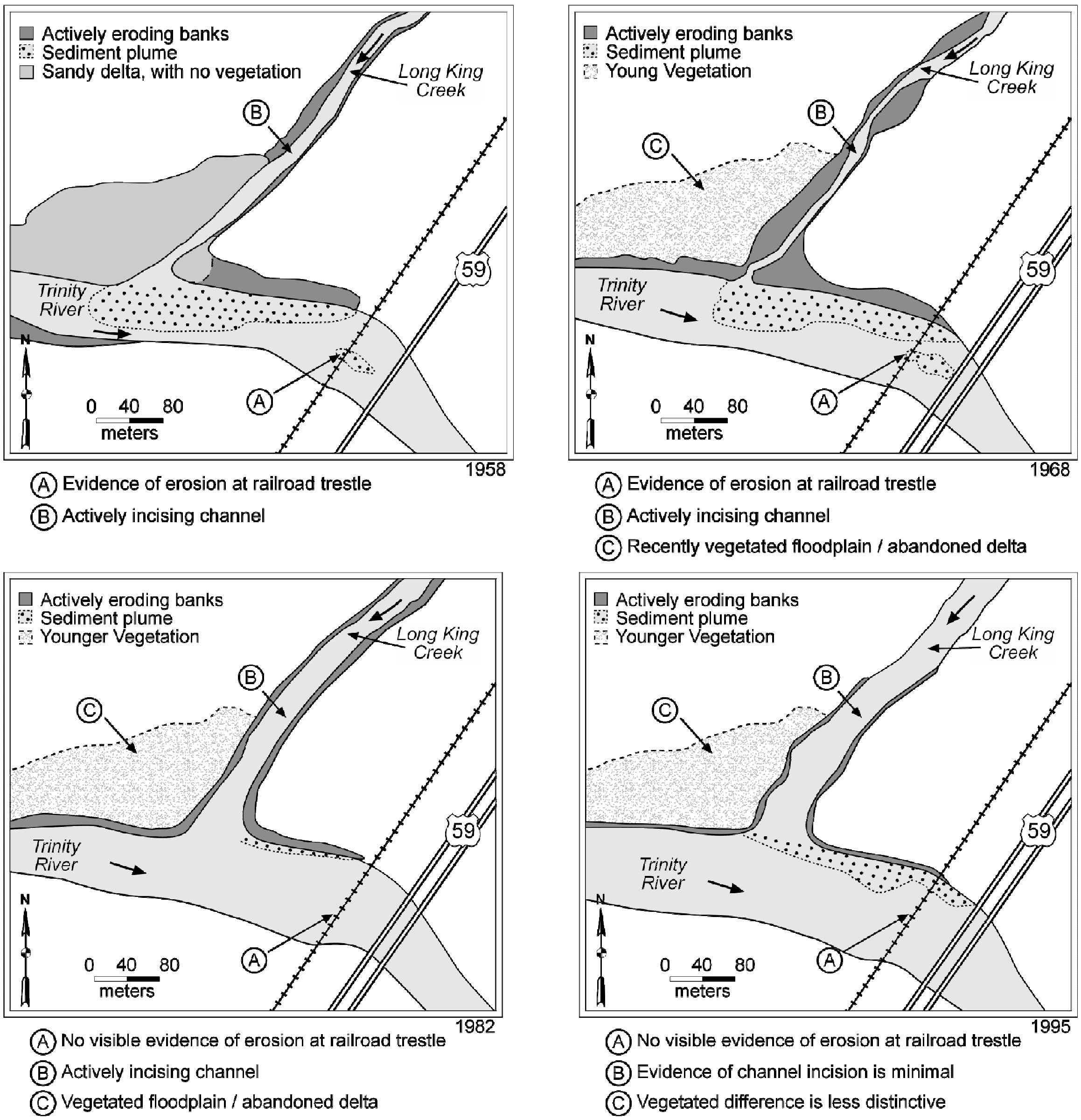

Figure 5. Channel and floodplain changes at the mouth of Long King Creek.

Figure 6. Long King Creek cutoff about 3.85km upstream of confluence with the Trinity River. This is the only notable location of upstream planform change that is resolvable in aerial photographs.

Geomorphic indicators of change

At the thirteen bridge sites, and an additional ten sites (23 total), field evidence of geomorphic changes was assessed. Indicators of geomorphic change include channel and bank morphology, vegetation, changes to cultural features such as bridges, dendrogeomorphic evidence (such as exposure or burial of tree roots) and comparisons of observations made during the study period (2001-2005) with earlier maps and aerial photographs (Table 1).

Field indicators of geomorphic change that were used to interpret bank erosion and channel widening included: fresh or active erosional scarps, cut banks, bank failures, woody debris in or near channels, exposed tree roots and root crowns, and tilted trees. Decreases in channel width were considered to occur if there was evidence of accretion or infilling on both banks.

Field indicators that were used to interpret channel incision included: tilted trees on banks and on floodplains, evidence of scour around anthropogenic features, knickpoints, exposure of a resistant clay layer within channel beds, relic channel shelves, bank scarps, and vegetation lines.

Numerous indicators of both channel widening and incision were observed at many of the field sites. Evidence of accretion or aggradation occurred at eight sites. Accretion or aggradation was considered to occur at sites where cultural features (e.g. east Texas creeks are unfortunately a popular dumping ground for old appliances and trash in general) were partially buried in sediment or floodplain sedimentation was evident. This is based on the idea that when dumped, the item thrown into the stream was flush with the channel floor. No sites revealed evidence of lateral migration; field evidence of erosion that was dominantly occurring on one bank while the adjacent bank showed evidence of accretion or infilling.

Table 1. Field evidence of channel responses to geomorphic change

Stream	Site	Geomorphic indicators of change
Big Ck.	BCSA	Undercut banks, trees fallen across channel, bedforms present
	150	Erosion scarps on banks, floodplain accretion
	222	Undercut banks, trees fallen across channel, bedforms present
Burnett Ck.	350	Trees bending into channel, roots exposed on channel banks, bank erosion, buried cultural feature
	942	Undercut banks in pools, roots exposed in banks, tilted trees into channel
Huffman Ck.	222	Slight undercutting of channel banks, stable banks, bedforms present, tilted trees
Little Ck.	SHNF	Slight undercutting of channel banks, stable banks, bedforms present, partially buried cultural feature
Long King Ck.	350	Bedforms present, exposed tree roots on banks, undercut trees on banks, tilted trees into channel
	headwaters	Bank erosion, knickpoints present, roots exposed in banks, tributaries not graded
	942	Bank erosion, exposed tree roots on banks
	190	Bank erosion, knickpoint present, floodplain accretion
	1988U	Bedforms present, stable banks, buried cultural feature
	1988L	Point bar accretion, bank stabilization with vegetation encroachment
	mouth	Bank stabilization with vegetation encroachment, floodplain accretion, delta formation, vegetation line
Long Tom Ck.	350	Undercut banks, exposed tree roots on banks, tilted trees into channel, buried cultural feature
	942	Undercut banks, tree roots exposed in channel banks, bedforms present, undercut bridge abutment on left bank

Menard Ck.	190	Undercut banks, few exposed tree roots
	Soda E. Loop	Bank erosion, exposed tree roots on banks, partially buried cultural feature, trees in channel
	943	Stable banks
	146	Stable banks, exposed bridge abutments, exposed tree roots in channel
	2610	Trees fallen across channel, roots exposed, erosion scarps on banks, bedforms present
	mouth	Floodplain accretion, delta formation and river sandbar migration and breaching, bank erosion
Mud Ck.	942	NA
Tempe Ck.	1988	Bank erosion, exposed tree roots on banks, sand bar mobility, trees in channel

NA: Not available (In November 2005, the stream had been rerouted and no flow was present.)

Results

Bridge cross-sections

Although a single bridge cross-section may not be a good indicator of channel changes taking place in a steam, a series of cross-sections taken over a period of years may provide significant information on the dynamic changes of a channel's geomorphic and hydrologic characteristics. At ten of the bridge sites, no attempt has been made to confine the flow, of the other three, two have had minor morphological influence while one is significantly engineered to influence stream morphology.

The temporal scale of data available for each individual bridge cross-section varies from five to 49 years. Numerous observations can be made from the bridge cross-sections (Figure Bridge). The data collected includes cross-sectional area, banktop-to-banktop width, max depth, mean depth, and the width/depth ratio (Table 2).

While each individual cross-section may reflect effects of recent scour and fill events, some observations may be deduced. The average rate of thalweg elevation change for all thirteen streams is -1.89 cm/yr (Table 3). A negative rate indicates that these thirteen locations have been, on average, degrading over time. The average rate in the change of cross-sectional area for all thirteen streams is 0.88 m^2/yr (Table 3), indicating a loss of alluvium from storage. The average rate in the change of the banktop-to-banktop width at all thirteen locations is 0.44 m/yr (Table 3). The positive rate indicates that on average these streams, at these locations, are becoming narrower.

Table 2. Channel dimensions at bridge cross-sections. 2003 surveys by the author; earlier surveys from the Texas Department of Transportation.

Stream	Site	Date	Cross-sectional area* (m^2)	Width (m)	Maximum depth (m)	Mean depth (m)	Width/maximum depth
Big Ck.	150	1980	14.5	22.42	1.19	0.65	18.86
		2003	11.2	12.35	1.52	0.89	8.10
	222	1971	30.3	19.99	2.26	1.46	8.86
		2003	32.5	22.45	2.19	1.42	10.23
Burnett	942	1996	24.1	15.63	2.50	1.50	6.25
Ck.		2003	23.6	14.53	2.53	1.58	5.74
Huffman	222	1998	12.6	14.84	1.61	0.85	9.22
Ck.		2003	11.3	12.85	1.51	0.87	8.54
Long King	942	1996	1.8	9.46	0.27	0.19	34.48
Ck.		2002	2.7	10.76	0.49	0.25	22.06
	190	1977	29.3	18.01	2.56	1.52	7.03
		2003	80.9	28.36	4.63	2.66	6.12
	1988U	1998	144.2	54.32	4.38	2.61	12.41
		2003	75.9	25.18	3.47	2.23	7.25
	1988L	1971	65.3	25.33	3.41	2.51	7.43
		2003	117.9	41.75	4.28	2.73	9.75
Menard	943	Original	5.99	6.86	1.34	0.82	5.12
Ck.		2000	9.36	10.47	1.46	0.87	7.16
	146	1994	99.4	63.49	3.59	1.55	17.69
		2003	100.1	72.96	4.51	1.47	16.18
	2610	1996	89.4	38.18	4.22	2.37	9.04
		2002	84.5	33.63	3.64	2.51	9.25
Mud Ck.	942	1996	13.1	9.38	2.07	1.37	4.52
		2003	17.5	11.36	2.62	1.48	4.33
Tempe Ck.	1988	1954	32.6	23.85	1.63	1.33	14.60
		2003	25.5	12.69	3.11	1.98	4.08

Channel slopes were also calculated using the bridge cross-sectional data for LKC, MC and Big Creek. Differences in channel thalweg elevations between upstream and downstream bridge crossings was divided by the distance between the two sites. Over four year spans in LKC (1998-2002) and MC (1996-2000), and a five year span in Big Creek (1998-2003) channel slopes in all three streams decreased. LKC slopes changed from 0.0009913 to 0.0009739; MC slopes from 0.0008934 to 0.0008892; and Big Creek slopes from 0.003289 to 0.003105.

Tempe Creek at FM 1988 is the only site with data available from the pre-dam era. At this location degradation has occurred since 1954. The thalweg elevation has fallen each year the site was surveyed (averaging 5 cm/yr over a 49 year span and 13 cm/yr over the last 7 years). Tempe Creek appears to have adjusted drastically morphologically since the first survey. The 1954 stream survey appears to show a rather engineered and unnatural stream cross-section (Figure 4). This rather wide and flat channel may explain the drastic degradation in thalweg elevation.

Table 3. Morphological data from bridge cross-sections.

Stream	Site	Distance from Dam (km)	Years	X-section change (m2)	Banktop-to-banktop width (m)	Thalweg change (cm)
Big Creek	150	72.7	23	3.37	10.07	27.4
	222	68.5	32	-2.25	-2.46	6.1
Burnett Creek	942	56.9	7	0.44	1.10	9.1
Huffman Creek	222	18.1	5	1.24	1.98	-7.6
Long King Creek	942	58.0	6	-0.90	-1.30	12.2
	190	39.7	26	-51.60	-10.35	-73.2
	1988U	35.3	5	68.31	29.14	-7
	1988L	22.7	32	-52.54	-16.42	-66.4
Menard Creek	943	73.8	?	-3.36	-3.61	-6.1
	146	49.2	9	-0.68	-9.47	-118.6
	2610	41.3	6	4.88	4.55	39
Mud Creek	942	58.2	7	-4.34	-1.99	-54.9
Tempe Creek	1988	32.7	49	7.04	11.16	-248.1
			Average rates per year	0.88	0.44	-1.89

Surface water measurements from the U.S. Geological Survey show little change in width/mean depth ratios at LKC 190 since measurements began in 1962, at MC 146 an increase in width/mean depth ratios occurred after 1980 (Figure 7). This increase in MC suggests an increased frequency of overbank flow and a possible hydraulic geometry response to damming of the trunk stream.

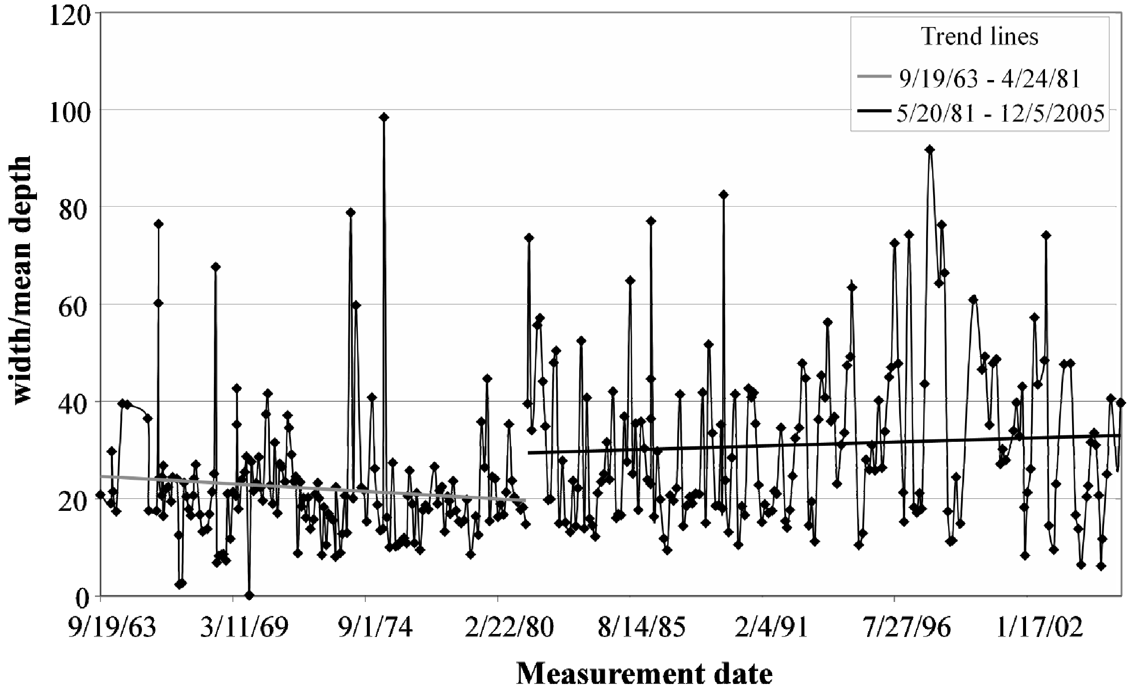

Figure 7. Width/depth ratios at Menard Creek 146, based on U.S. Geological Survey field measurements.

Aerial photos

<u>LKC mouth</u>

Figure 5 shows maps produced from the four years of photos at the mouth of LKC. The 1958 photo suggests that this location is very geomorphologiclly active; a sediment plume is exiting LKC and entering the Trinity, aggradation is causing a delta to build, and the channel banks appear to be eroding (Figure 5). The right bank of LKC's delta appears to have little or no vegetation. LKC's channel also appears to be actively incising at this time. Evidence of this incision includes a small incised gully at the upstream, right bank edge of the delta and a vertical bank cut into the alluvium near the left bank end of the delta.

In 1968 (Figure 5) the right bank delta area appears to be vegetated and aggraded. Aggradation on the right bank delta area (since 1958) situates the surface elevation closer to the elevation of the floodplain to the north. A break in slope is much less noticeable as vegetation has colonized this side of the delta. LKC's channel also appears to continue to incise. The vertical bank cut into the alluvium near the left bank end of the delta is still visible. A sediment plume is also visible entering the Trinity.

In 1982 (Figure 5) the vegetation on the right bank of LKC's delta has firmly established itself. The water levels in this photo are higher in both the Trinity and LKC then in the two previous photos. The high discharges are likely covering the delta. The water flowing in LKC appears lighter in color than the water flowing in the Trinity, likely resulting from higher concentrations of suspended sediment in LKC; a minor sediment plume is entering the Trinity.

Similarly high flow conditions are observed in the 1995 photo (Figure 5); the higher stage in both LKC and the Trinity is likely masking the delta. The younger vegetation on the right bank of the delta is more difficult to distinguish from the older vegetation, and a small sediment plume is entering the Trinity. On 18 October, 1994, the Trinity River peaked at 3398 m^3/s, the flood of record. LKC peaked one day prior at 852 m^3/s, also the flood of record. The geomorphic changes caused by this single event within the lower Trinity system were quite significant and are likely masked by the high water in the image.

<u>LKC upstream</u>

Figure 6 shows a segment of LKC about 3.85 km (apex of the meander) upstream from the confluence with the Trinity. At this location a cutoff formed sometime between 1968 and 1982. This cutoff and subsequent oxbow formation shortened the stream by 0.5 km, and is the only significant (upstream) planform change observed from multiple image comparisons. The foreshortening of LKC at this location would have caused an increase in energy within the system by increasing the slope within this reach. The discharge in LKC when the images were captured was relatively low. Although the 1982 photo has lower resolution, in both photos point bars and alluvium covered banks are visible.

Geomorphic indicators of change

Geomorphic indicators of change may be used to assess channel change, and may imply the direction in which change is or has occurred. Field mapping and observations of indicators of change within the tributaries suggest that at most of the sites the tributaries are geomorphically active, either widening and/or degrading. Field evidence of channel responses at the field sites is summarized in Table 1. A tight gray clay acts as a local 'bedrock' in some portions of the lower Trinity basin. Alluvium covering this gray clay, often occurring as various bedforms, was considered active and mobilizable (Table 1).

<u>Qualitative change</u>

At most of the sites geomorphic activity is evident (Table 4). No single mode of adjustment appears to be the dominant response (13 different modes of adjustment) to a system perturbation. Although, channel incision and widening appear to be the dominant action occurring in the tributaries. This sort of activity would be expected in these streams, regardless of system perturbations, as they evolve and dissect the landscape. The most dynamic location of geomorphic change occurs at the confluence with the Trinity River. Only two of these sites (the mouths of LKC and MC) were observed in the field during this study. There appears to be no fundamental differences in the reactions of the tributaries regardless of distance from the dam, or east versus west sides of the basin.

Table 4.
Qualitative changes in width, depth, slope and roughness estimated from geomorphic indicators of change, aerial photos and bridge cross-sections.

Stream	Site	Width	Depth	Slope	Roughness
Big Ck.	BCSA	+	+	-	+
	150	-	+	-	+
	222	+	-	-	+
Burnett Ck.	350	+	+	+	+
	942	+	+	+	+
Huffman Ck.	222	+	+	+	-
Little Ck.	SHNF	+	+	-	+
Long King Ck.	350	+	+	+	+
	headwaters	+	+	+	-
	942	+	0	+	0
	190	+	+	+	+
	1988U	-	-	-	-
	1988L	+	+	-	+
	mouth	+	+	-	+
Long Tom Ck.	350	+	+	+	+
	942	+	+	+	+
Menard Ck.	190	+	+	0	0
	Soda E. Loop	0	+	0	+
	943	+	0	0	0
	146	+	+	+	+
	2610	-	-	-	+
	mouth	0	+	-	+
Mud Ck.	942	ND	ND	ND	ND
Tempe Ck.	1988	-	+	+	+

(+) = increase; (-) = decrease; 0 = no significant change; ND = no data

Synthesis and summary

While Livingston Dam has caused a disruption in the sediment system of the lower Trinity River (Phillips and Musselman, 2003; Phillips et al., 2004; Phillips et al., 2005), it may have had a negligible effect on the tributaries.

U.S. Geological Survey records show an increase in discharge and a possible change in flow regime in the post-dam era on both gauged tributaries (LKC and MC). As Wellmeyer et al. (2004) noted for the lower Trinity, this increase might be attributable to increased precipitation in east Texas in the post-dam period. During the pre-dam period (while data was being collected for LKC and MC) east Texas was experiencing a mild to moderate drought (Riggio et al., 1987).

Surveys at thirteen bridge crossings showed a tendency toward degradation may exist in the tributaries. On average the creeks' thalweg elevations dropped 1.89 cm/yr. Further evidence of degradation in the creek channels included geomorphic indicators of change. 19 of the 25 sites (76 percent) had geomorphic indicators that suggested degradation had occurred within relatively recent times (within the past 50 years). At the remaining six sites buried cultural features (such as a refrigerator, engine block, toilet and bridge features) suggested that aggradation had occurred in recent times.

Both incision and aggradation suggest an active sediment system with erosion and transportation constantly removing and adding alluvium.

An increase in slope of the LKC system may have resulted from an increased energy regime within the Trinity system. Between 1968 and 1982, an oxbow lake was created 3.85 km upstream from the confluence with the Trinity. This is the only resolvable (from aerial photos) planform change within the tributary systems.

Cross-sectional area change was measured from thirteen bridge surveys. The average rate at which cross-sectional area changed was 0.88 m^2/yr. This rate of change suggests that over time there has been an increase in channel area (e.g. alluvium in storage is being removed from within the channel). At 17 sites (68 percent), field evidence suggests that width has increased within relatively recent times (50 years). The average rate at which banktop-to-banktop width has changed (-0.439 m/yr) suggests channel narrowing in the tributary creeks.

LKC and MC confluences with the Trinity River are areas sensitive to system change. Using aerial photography that spans 37 years, geomorphic change at LKC's mouth was shown to be dramatic. At its confluence with the Trinity, LKC has continued to build a substantial delta system through cycles of degradation and accretion. Vegetation on the delta has been stripped by numerous storm events over the years, but has continued to recolonize and encroach upon the channel. The channel itself has shifted laterally across the delta during the four years of field work in this study.

At its confluence with the Trinity, MC has been less active than LKC. MC has experienced a drop in base level as the Trinity degraded and shifted laterally after impoundment. At the mouth, MC has built a delta system which interacts complexly with the Trinity. Gullying on the channel banks and channel erosion provide sediment to the system.

Discussion

The tributaries in the lower Trinity basin are dynamic systems. In the lower Trinity River system, geomorphic characteristics are largely dominated by Holocene sea level change and the response to extreme events (e.g. flood of record in 1994), so that dam effects are relatively localized. The geomorphic indicators of change within the tributary streams suggest highly active systems changing often in response to varying flow conditions. This makes system responses to the Livingston impoundment difficult to distinguish from other fluvial adjustments. The response to dam-induced Trinity River downcutting has apparently not progressed very far up the tributaries. The responses to imposed change caused by the impoundment of the Trinity River are concentrated at the mouths, and may not be detectable beyond the confluences. The mouths of the three largest first-order tributaries are behaving quite differently in response to trunk stream adjustments.

While the Trinity River's downstream adjustments to Lake Livingston has caused adjustments within the tributary streams, contingency, nonlinearity and other complex responses make it difficult to identify any consistent response.

The response of a fluvial system to a point-centered perturbation such as a dam could be expected to start at the location of the disturbance and propagate downstream. The response in the lower Trinity system has been observed for about 60 km downstream of the dam. Further downstream channel incision and/or widening and slope decreases are not evident (Phillips et al., 2005). The question raised is whether the upstream sites are unaffected by the dam, or whether the response has not propagated these distances in the 35 years since the disruption.

Spatial and temporal propagation

The spatial and temporal propagation of a disturbance through a system is contingent upon local factors and the magnitude, rate, and duration of the change. Disturbance migration rates through streams similar to those in this study have been shown to vary with basin size; so that rates of migration on mainstem channels are an order of magnitude greater than in their tributary basins (Yodis and Kesel, 1993). If the order of magnitude relationship holds true the disturbance within the Trinity tributaries would not have propagated very far upstream. For example, dam effects propagated 60 km downstream in about 35 years. This suggests a mean propagation rate of 1.7 km/yr. This exceeds rates found by Galay (1983), who reported downstream propagation of stream bed degradation after impoundment in sand bed rivers at rates of 0.72, 0.93, and 0.66 km/yr. Using the estimated rate from the lower Trinity (1.7 km/yr), the disturbance would have reached LKC's confluence in 9.5 years. Based on a change in width/depth ratios after 1982, Phillips et al. (2005) estimated that downcutting at this site would have occurred by the early 1980s. Assuming 25.5 years ago the disturbance reached LKC's confluence, and a propagation rate of 0.17 km/yr (e.g., an order of magnitude less than the mainstem), the disturbance would have propagated upstream in LKC no more than ~4.5 km since impoundment. This implies that there has not been enough time since impoundment for the disturbance to propagate to any of the upstream sites along LKC (LKC 1988L is the closest upstream field site to the mouth at 6.6 km). Even the nearest upstream tributary site to the dam, Huffman Creek at 222 would not be affected at these rates.

While these calculations are admittedly crude, they do serve to indicate that the lack of evidence of propagation of effects at upstream sites is consistent with general findings in other studies where distance decay (Germanoski and Ritter, 1988) and landscape sensitivity (Yodis and Kesel, 1993) influence the propagation of a disturbance within a fluvial system.

Regardless of propagation rates, Leopold and Bull (1979) argue that a change in base level will have little upstream effect. They concluded that base level changes affect the vertical part of the longitudinal profile only locally, whereas upstream hydrologic controls determine the more regional profile. Experimental studies have also shown that changes in base level may produce localized effects, and further upstream responses are limited with distance (Koss et al., 1994).

While upstream sites on all the tributaries have geomorphic indicators of change suggesting active behavior within the system, none of the sites revealed evidence that suggests an increase in activity caused by a recent perturbation. Evidence at numerous sites suggested that these tributaries do react to extreme events such as the 1994 flood.

Conclusions

While the geomorphological effects of dams on downstream hydrology, sediment discharge and ecosystems have been extensively studied, these studies rarely consider the effects of the impoundment on the downstream tributaries. In many of these studies tributaries are noted for contributing significant inputs of energy (flow) and mass (sediment) to the mainstem system. Other than the inputs to the impounded trunk stream, however, few studies have considered tributary effects beyond the confluence with the mainstem. Although confluences are critical locations which may amplify local disturbances within a fluvial system (Benda et al., 2004), upstream coupling within a tributary may force system changes beyond its confluence.

Flow data in the lower Trinity basin rule out modifications in the discharge regime as a significant cause of change. First, the two gauged tributaries (Long King Creek and Menard Creek) along with the three stations on the Trinity River (Goodrich, Romayor and Liberty) do not show any indication of post-dam alterations in flow. On the mainstem, slightly elevated flows in the post-dam period have been attributed to higher-than-average precipitation during this corresponding period (Wellmeyer et al., 2004). Precipitation records at Liberty, TX (in the Trinity basin but below the study area), show a general increase in the amount of precipitation over the past century. During the time period in which discharge data are available for Long King Creek and Menard Creek, pre-dam annual mean precipitation amounts are 20 cm less than during the post-dam period and 10 cm less than during the entire span of pre-dam precipitation record. This suggests that the short pre-dam discharge data for both tributaries may not accurately represent the historical flow trend.

Deltas at the mouths of LKC and MC may suggest a change in sediment dynamics, but are likely attributable to the slight changes in hydrographs. Even though no general change in flow regime is associated with the dam, flood waves are slowed as they pass through Lake Livingston. Thus, tributary flows are out of phase with the Trinity River. Subsequently, the tributaries peak sooner. When the tributaries are carrying their maximum sediment loads to the mainstem, the Trinity has not yet reached its maximum transport potential, and deposition occurs. While changes in the characteristics of the LKC delta have occurred, a delta existed prior to 1968 and dam emplacement. As Trinity flows increase, stream power increases, transporting portions of the recently deposited alluvium. While the Trinity flow increases, tributary flows are decreasing, creating backwater flooding. Evidence of backwater deposits occurs on the delta surfaces at the mouths of LKC and MC.

The variable reactions at the mouths of the two largest tributaries may be interpreted using a nonlinear-complex response approach. Petts's (1982) complex response downstream of an impoundment was characterized by alternating phases of erosion and deposition. These responses or phases are controlled within a reach by the character of the processes (the interaction between trunk and tributary) and channel morphology before impoundment. The response differences at the mouth locations of Long King Creek and Menard Creek indicate they may be influenced most strongly by local conditions. The LKC basin resides in a different land use setting (more populated and developed) than the MC basin (more rural and including a nature preserve). These local conditions and the reaction of the trunk stream within the reach of the confluence are most likely the controlling factors of the nonlinear-complex tributary responses.

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Part 5

Sediment fingerprinting in the lower Trinity River, Texas

In Year 4 of this project we analysed 138 potential catchment source samples from within the Trinity River system using mineral magnetics. The samples showed considerable variability in their magnetic properties. While the data suggested that quantitative sediment contribution estimates from the various sub-catchments of the Trinity River system may be possible, the large scale of the system and the inherent spatial variability of the potential source materials at the sub-catchment level suggested that such as approach would be a significant undertaking. Characterising the multitude of potential sources would require extensive and rigorous field sampling.

In his consultancy report in Year 4, Dr. John Walden suggested that a more pragmatic approach would involve a methodology whereby the sediments in transport up-stream of a major tributary are used as end-members to unmix the composition of the sediments in transport down-stream of the tributary. While this approach is more likely to be effective than one based upon characterising potential sources from within the catchment, in a system as large as the Trinity River, it would still require significant resources to fully implement. Ideally, this should involve two sub-catchments of different magnetic characteristics – one with high concentration of magnetic minerals and another with low/intermediate concentrations. In each case, this should involve a suspended sediment sampling program both upstream and downstream of the sub-catchment tributary, with a view to characterising the compositions of the sediments. Such an evaluation program could establish the feasibility of a sediment source mixing model based on environmental magnetic properties for ascribing the contribution to the main channel sediment transport system made by the respective sub-catchments.

Although we intended to conduct such a study, as per Task 2, we decided not to pursue fingerprinting in Year 5. The problem was essentially methodological: there were no access points on the Trinity River either upstream or downstream of where the two major tributaries (i.e., Long King Creek and Menard Creek) joined the main channel. It was also impossible to sample suspended sediment within the tributaries close to where they joined the Trinity. Because the procedure concentrates upon characterising the suspended sediment both up-stream and down-stream of major tributaries, the lower Trinity is simply not the ideal location to evaluate this approach. In addition, the sediment budget work gave us a good estimation of the contributions from these two sub-catchments.