FINAL REPORT

YEAR 4 OF LOWER

TRINITY PROJECT
DOWNSTREAM TRENDS IN DISCHARGE, SLOPE, AND STREAM POWER IN THE LOWER TRINITY RIVER, TEXAS

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FINAL REPORT – YEAR 4 OF LOWER TRINITY PROJECT

Downstream Trends in Discharge, Slope, and Stream Power in the Lower Trinity River, Texas

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Overview

This report is divided into two parts. The first examines downstream trends in fluvial processes in the Lower Trinity. This work is currently in press in the *Journal of Hydrology*, and is presented here in its entirety. The second part of the report presents work on accretion rates in the delta. All sediment monitoring work for 2005-2006 will be included in the final project report due March 2007.

PART 1: DOWNSTREAM CHANGES IN FLUVIAL PROCESSES

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 accommodation 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 level changes.

The locus of sediment deposition in low-gradient coastal plain rivers may be well upstream of deltas, estuaries, or even fluvial-estuarine transition zones (Phillips and Slattery 2006). Sediment and freshwater fluxes to the coast are typically measured or estimated based on gaging stations well upstream of the coast, and upriver from the lower coastal plain sediment bottlenecks (Phillips and Slattery 2006). Thus, 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 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 fuction as distributaries at high flow. The purpose of this paper is to investigate 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 U.S. 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) is a function of the product of slope (S) and discharge (Q),

$$\Omega = \gamma Q S \qquad (1)$$

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) showed this to be the case for arroyo systems in Utah, and Magilligan (1992) demonstrated irregular downstream trends in stream power for reference floods in a humid perennial stream in Wisconsin. In Magilligan's (1992) study lithological controls on valley width and channel slope played a predominant role in the downstream variations in flood power. Nonlinear downstream changes in stream power in a Wisconsin watershed were documented by Lecce (1997), who showed power peaking where drainage areas were 10 to 100 km^2 (in a 208 km^2 drainage basin) and decreasing rapidly downstream. Lithological controls were also found to be important in Lecce's (1997) study. Knighton (1999) considered downstream variation in stream power, based on a standard downstream increase in discharge as a function of drainage area, and slope determined by a concave-up longitudinal profile described by an exponential function. 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 Reinfelds 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 (Reinfelds 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; Reinfelds et al. 2004) discharge was assumed to increased 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 Reinfelds 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 Tornqvist 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 (Reinfelds 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 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.

In the James River, Virginia, Nichols et al. (1991) subdivided the lowermost river into the bay mouth at Chesapeake Bay, the estuary funnel, and the meander zone. The upper end of the estuary funnel was defined on the basis of typical salinity patterns, width-depth ratio, and sedimentology. The upstream limit of the meander zone coincides with the inland limit of tidal influence (Nichols et al. 1991; figure 13). In the Tar-Pamlico River, North Carolina, the open-water mouth, transition to a distributary network, location of common salt wedge penetration (50 percent probability in a given year), transition from sand- to mud-bed channel, upstream limit of tidal/backwater influence, and point at which the bed is cut to below sea level occur at six different locations covering 50 km of channel distance (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 in east-central 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. The bay is a lagoon-type estuary, separated from the Gulf of Mexico by Galveston Island and Bolivar Peninsula. Using data from the National Estuarine Inventory, Nichols (1989) calculated the ratio between bay volume and mean annual freshwater inflow as 0.2, with a mean water residence time of 40 days. Nichols (1989) also calculated a flow ratio, based on mean freshwater inflow during half a tidal cycle divided by the tidal prism (volume of water exchanged over a tidal cycle) as 0.183. The bay's drainage area is 85,470 km^2, of which about 54 percent is the Trinity River, which provides a similar proportion of the mean annual freshwater inflow. Though the Lake Livingston's capacity is more than 80 percent that of Galveston Bay, analysis of pre- and post-dam discharge records at Romayor found no significant post-dam decrease in flow, and limited change of any kind (Wellmeyer et al. 2005).

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).

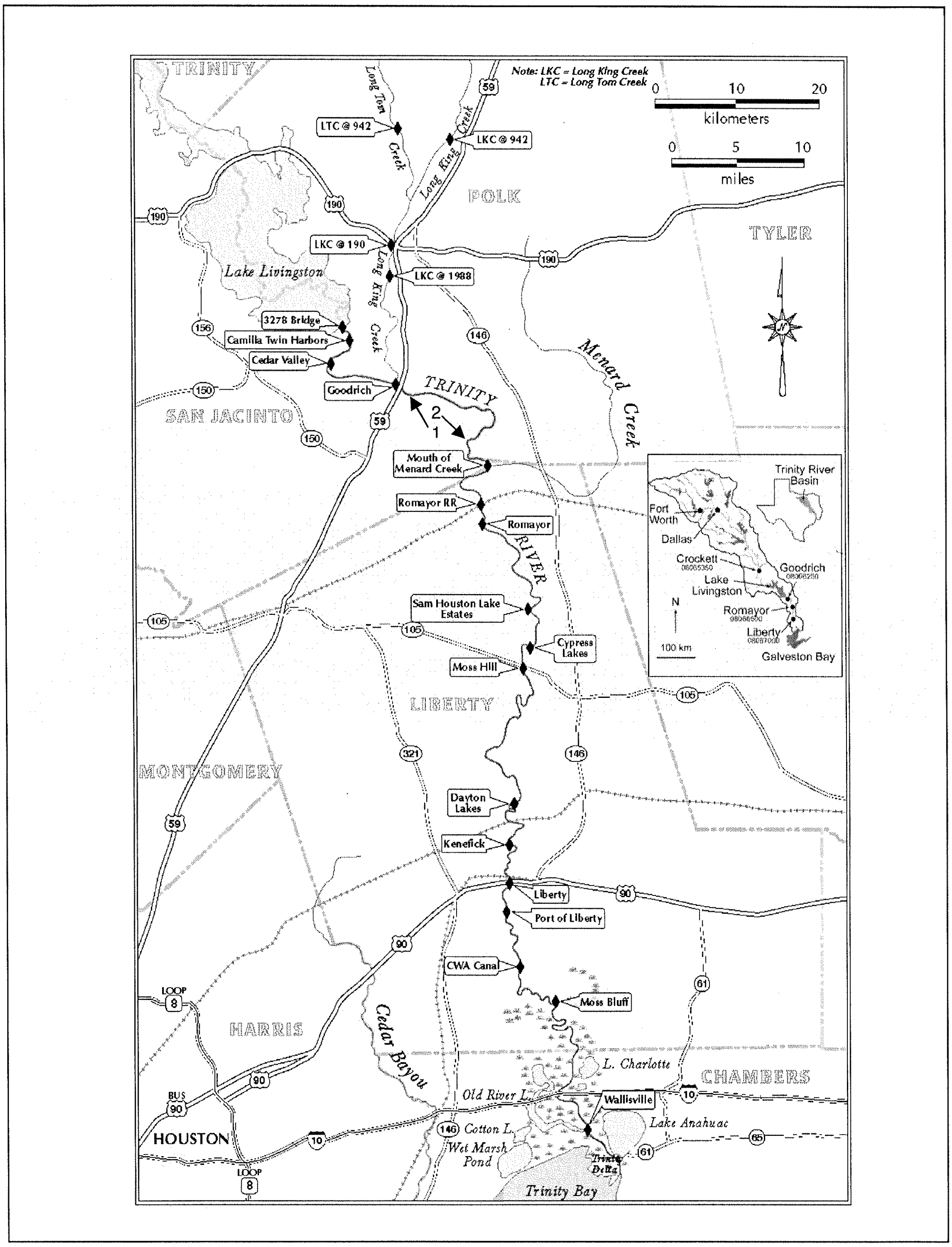


Figure 1. Study area, showing gaging stations and field sites. Arrow labeled 1 is Mussel Shoals; arrow labeled 2 is Big Creek at the south end of Grama Grass Bottom.

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 Figure 1, and described in Table 1.

Table 1. Lower Trinity River (TR) gaging stations. Location refers to distance upstream from Trinity Bay, in kilometers. Number is the U.S. Geological Survey station number. Measurements of interest here include discharge (Q) and stage or gage height (H).

Name	Location	Number	Measurements
Livingston Reservoir	177	0866190	H, storage
Long King Creek at Livingston	145[1]	0866200	H, Q
Menard Creek at Rye	130[1]	0866300	H, Q
Trinity River (TR) near Goodrich	144	0866250	H, Q
TR at Romayor	126	0866500	H, Q
TR at Liberty	83	0867000	H, Q[2]
TR at Moss Bluff	32.5	0867100	H, Q[3]
Old River cutoff near Moss Bluff	30[1]	0867215	H, velocity
TR at Wallisville	6.5	0867252	H

[1]Approximate distance from the bay of creek/river confluence.
[2]Discharge measurements discontinuous
[3]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)

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

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

Where m is the rank of the flow in the series, and n is the total length of the series. P is the probability of exceedence, and T is the return period or recurrence interval. Reference flows include those associated with 50, 10, and 1 percent 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 were examined based on direct comparisons and differences between downstream and upstream stations (Liberty-Romayor; Romayor-Goodrich). Discharge is reported in $ft^3\ sec^{-1}$; these values were converted to $m^3\ sec^{-1}$.

Table 2. Lower Trinity River gaging stations, with station names as referred to in text. Refer to table 1 for official designations. Operating agency refers to the operator of the station; the U.S. Geological Survey (USGS) distributes data and maintains records for all stations.

Name	Operating agency	Established
Lake Livingston	Trinity River Authority	1969
Long King	USGS	1963
Menard	USGS	1965
Goodrich	USGS	1965
Romayor	USGS	1924
Liberty	USGS	1940
Moss Bluff	USGS	1959
Old River	US Army Corps of Engineers	2003
Wallisville	US Army Corps of Engineers	2003

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 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 digital elevation models 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.

Valley Topography

Topography of the lower Trinity Valley was analyzed based on 10-m resolution digital elevation models 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, and 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 artifacts. Digital orthophotoquads (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 3. 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 percent 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 3 indicate that there is not necessarily a consistent downstream increase in discharge, even within the always fluvially-dominated Goodrich-Romayor reach.

Table 3. Reference flows for lower Trinity River gaging stations, in m^3sec^{-1}. 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.

Reference Flow	Goodrich	Romayor	Liberty
MAQ	231	246	509
50% exceedence	82	77	433
10% exceedence	677	640	1048
1% exceedence	1550	1541	1822
Q1	2130	1970	2484
Q2	2400	2330	2835
Q10	3002	2925	3600
2002 flood	1872	2198	1602
1994 flood	3540	3455	3823

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 Figure 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.

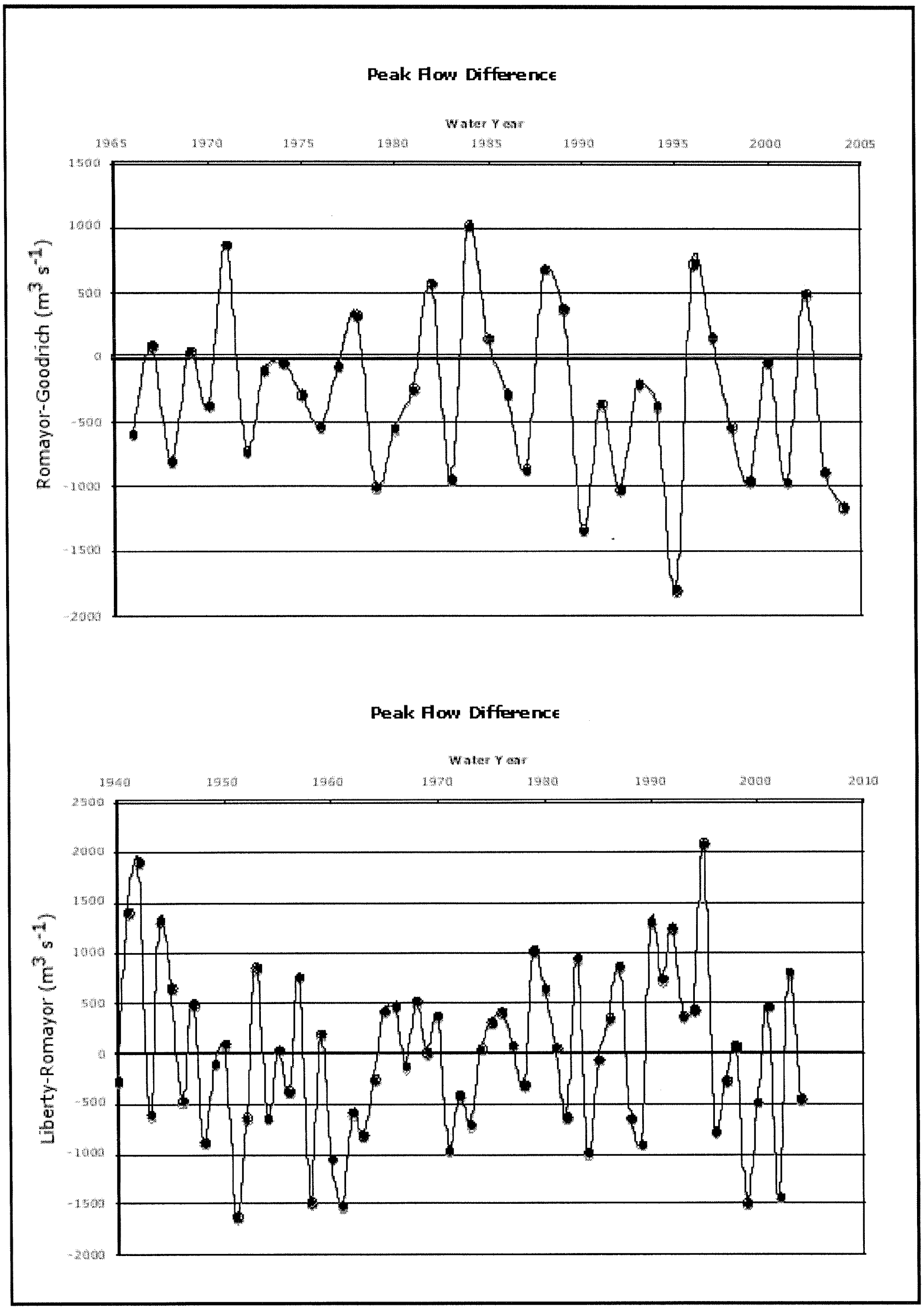


Figure 2. Annual peak discharge at the Romayor gaging station minus that at the Goodrich station 18 km upstream (top), and at the Liberty gaging station minus that at Romayor, 43 km upstream.

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-hour precipitation estimates from the Lake Charles, Louisiana National Weather Service Radar indicated 25 to 100 mm in the lower Trinity basin.

The response of two tributaries is shown in Figure 3. Both experienced steep rises in the hydrograph. Long King Creek experienced an equally steep recession, whereas Menard Creek flow remained elevated for several days. This is consistent with the more highly developed Long King watershed, which has considerably more urbanized area, as well as pasture and other agricultural land uses. The Menard watershed, much of which is within the Big Thicket National Preserve, is predominantly forested. The creeks began rising at about 0430 September 24 (Table 4), though the hydrograph had begun rising at 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 4).

Table 4. 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

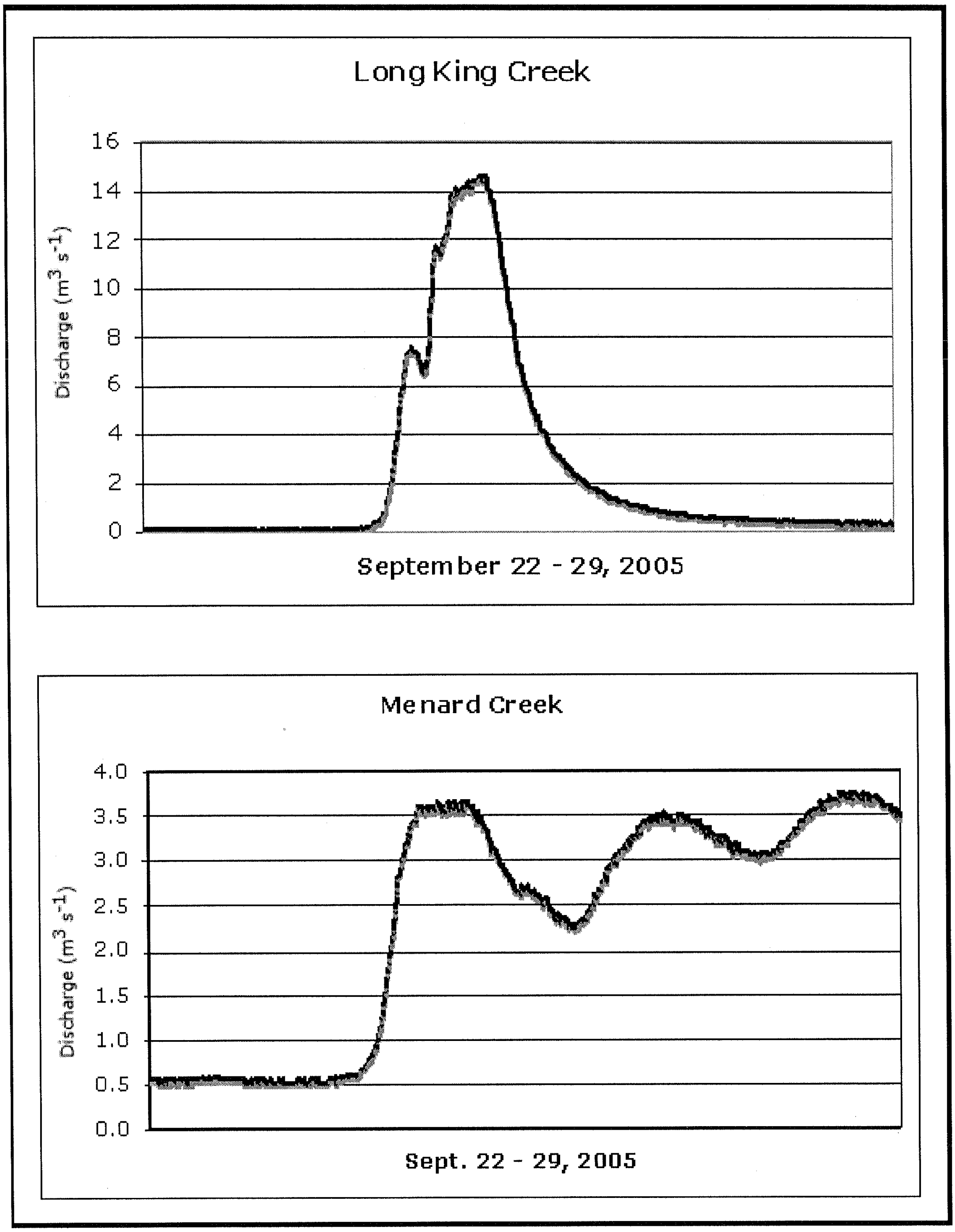


Figure 3. Hydrographs for two lower Trinity River tributaries for the week including the Hurricane Rita flow event. Discharge was measured every 15 minutes.

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 (Figure 4).

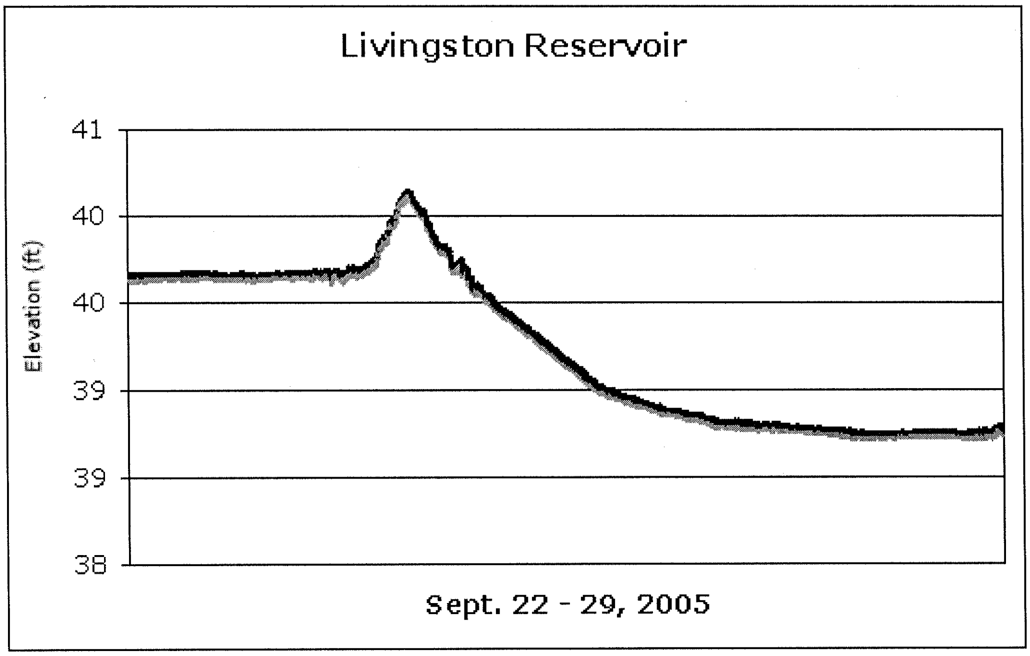
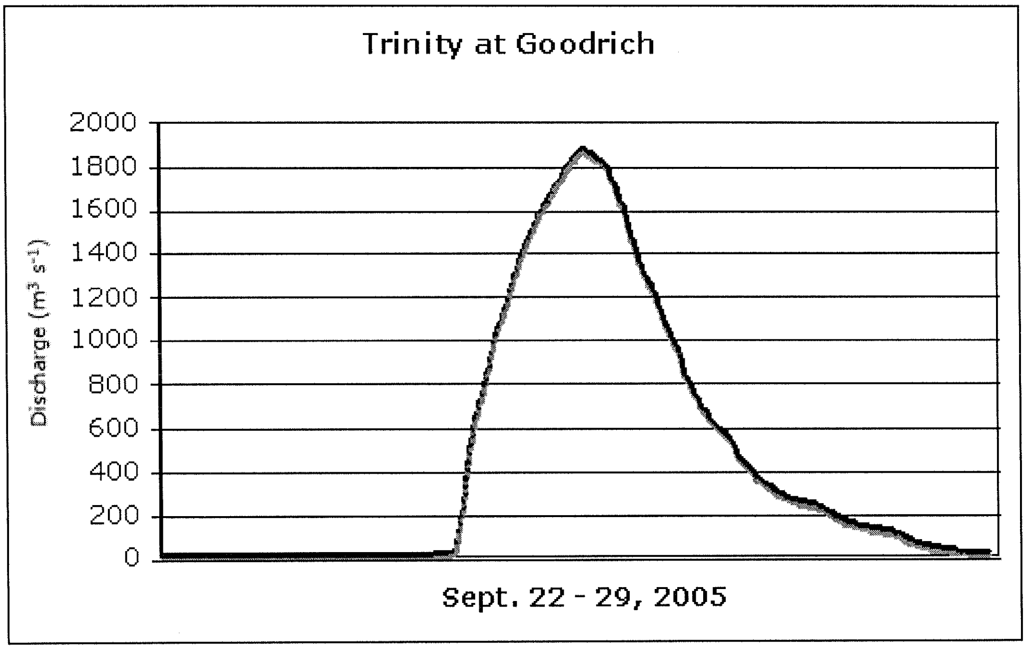


Figure 4. Water surface elevations for Lake Livingston for the week including Hurricane Rita, showing the rise in lake levels and subsequent drawdown via dam releases to assess and repair damages.

The hydrograph responses of the river at Goodrich and Romayor (Fig. 5) show a rapid rise and recession similar to the Long King hydrograph (Fig. 3) and the Lake drawdown curve (Fig. 4), with the peak at Romayor occurring 7.75 hours 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 (Figure 6) showed a sustained rise in base flow. Note that while discharge at Liberty was partly estimated by the author, 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 (Figure 7).



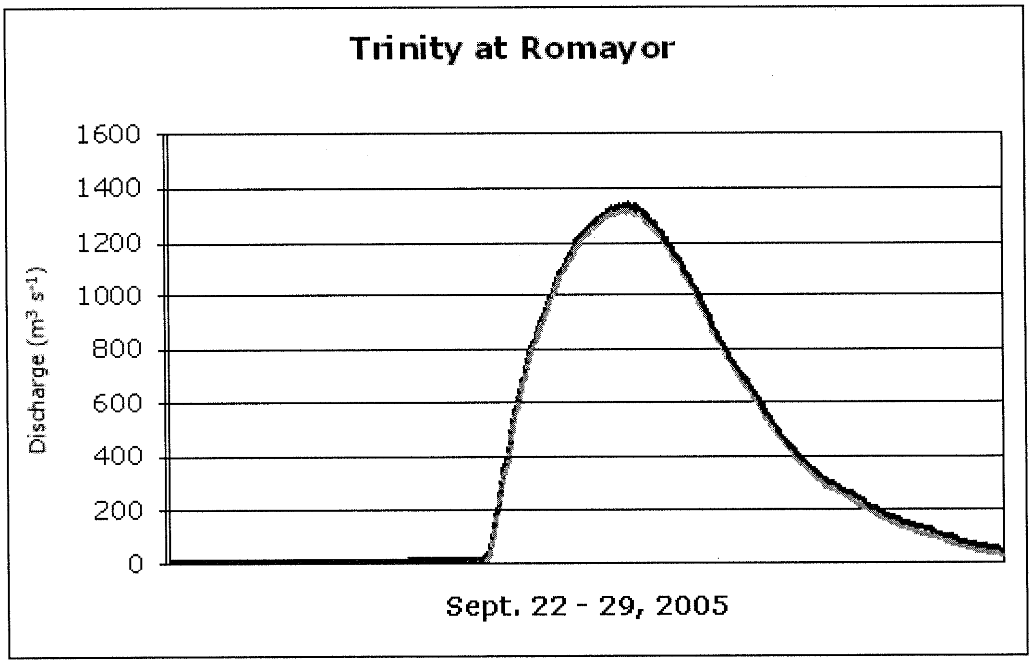


Figure 5. Hydrographs for the lower Trinity River at the Goodrich and Romayor gaging stations for the week including the Hurricane Rita flow event. Discharge was measured every 15 minutes.

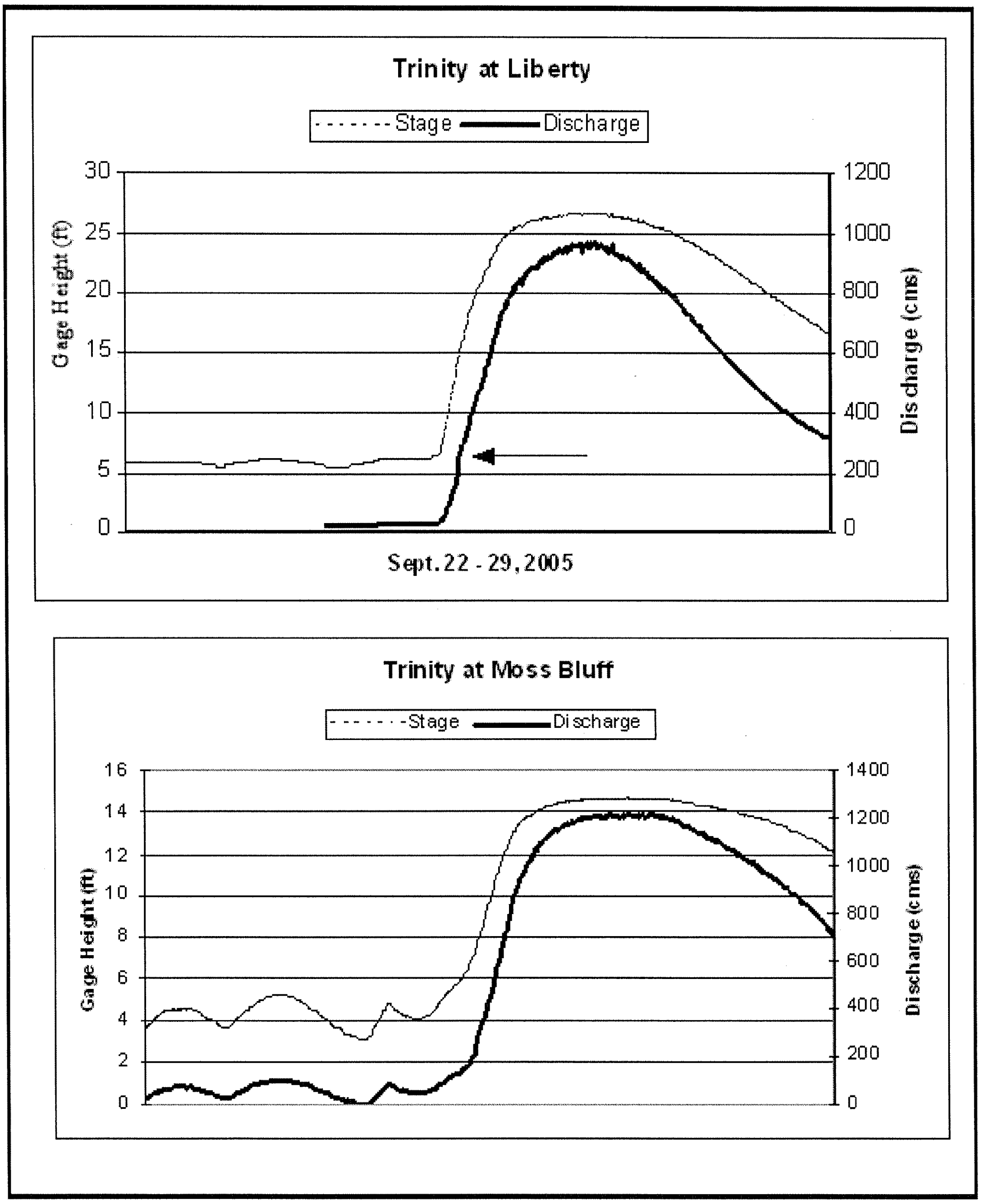


Figure 6. Stage (gage height) and discharge for the lower Trinity River at the Liberty and Moss Bluff gaging stations, with readings every 15 minutes. For the Liberty data, discharge was estimated by the author for the portion of the curve prior to the point indicated by the arrow. For Moss Bluff, discharge is entirely estimated by the author.

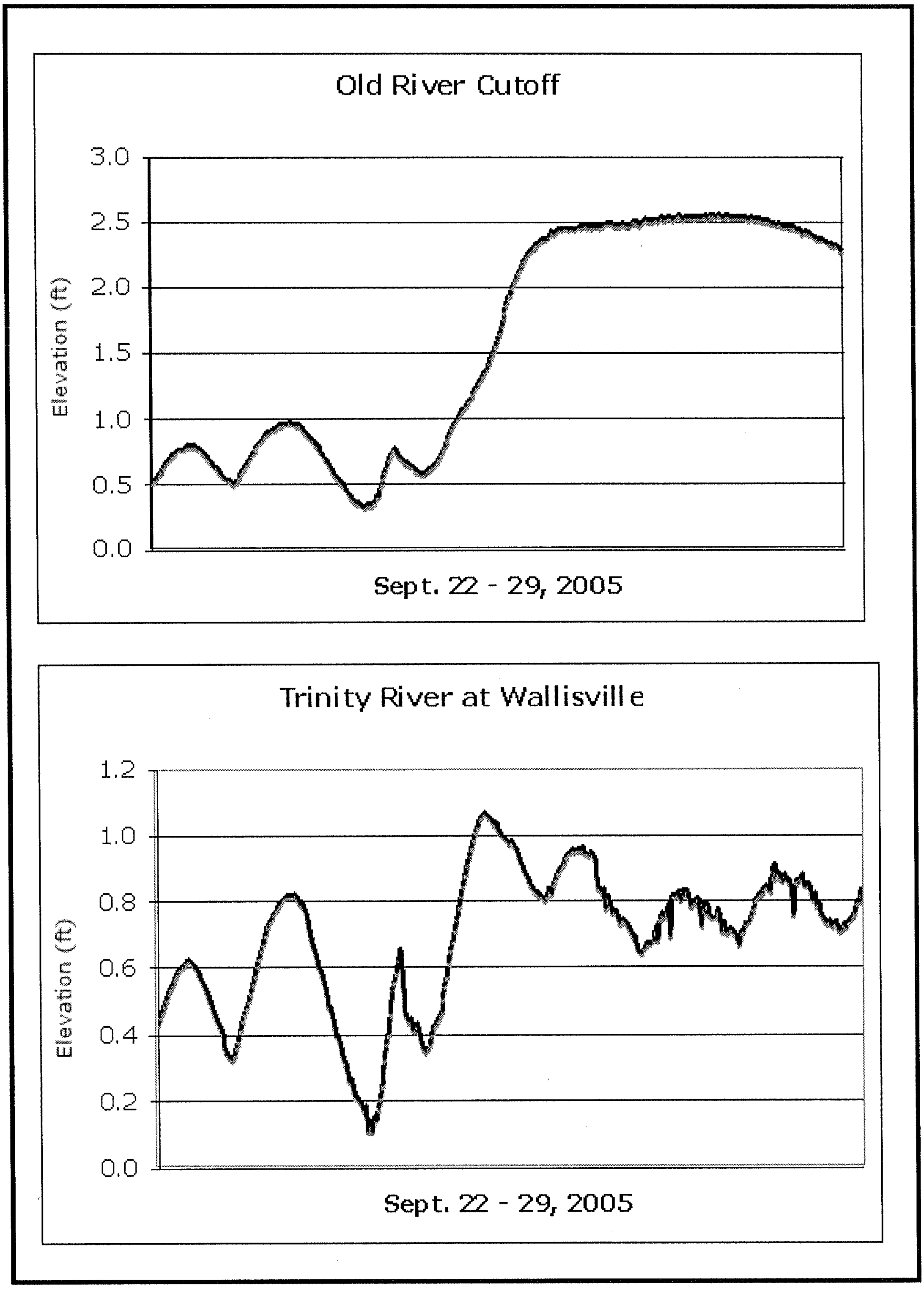


Figure 7. Water surface elevations for the week including the Hurricane Rita event at the Old River cutoff and Wallisville sites. Readings were taken every 15 minutes.

Slope

Channel bed elevations were measured in 2003 at five bridge cross-sections, which include the Goodrich, Romayor, and Liberty sites (Phillips et al. 2005). The thalweg slopes from near Livingston Dam to Goodrich, Goodrich to Romayor, and Romayor to Liberty are shown in Table 5. Water surface slopes from a July 12, 2004 event presented by Phillips and Slattery (2006) are also shown. Additionally, 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. In all cases Table 5 shows the slope from the next upstream station (e.g., the Romayor slope represents the Goodrich to Romayor gradient, etc.). Water surface profiles for the Rita event are shown in Fig. 8. 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 ot 16 to 17 m, consistent with the computed water surface slope between Romayor and Liberty.

Table 5. Slopes at lower Trinity River gaging stations. WS 1 is based on water surface slopes at midnight, 12 July 2004 (after Phillips and Slattery 2006). The other water surface slopes are during the Rita flow event as described in Table 4.

	Goodrich	Romayor	Liberty	Moss Bluff	Wallisville
Thalweg	.0003834	.0002508	.0000100		
WS 1		.0001784	.0002053	.0000371	
WS R1		.0001400	.0002382	-.0000016	.0.020971
WS R2		.0001387	.0002391	-.0000005	.0327170
WS R3		.0001427	.0002397	-.0000007	.0063139
WS R4		.0001432	.0002395	-.0000010	-.0006507
WS R5		.0003007	.0003704	-.0000056	-.0793389
WS R6		.0002794	.0003010	.0000482	-.0160179
WS R7		.0002414	.0002698	.0000555	.0714494
WS R8		.0001178	.0001930	.0000591	.1174534
WS R9		.0001158	.0001903	.0000576	.1160273

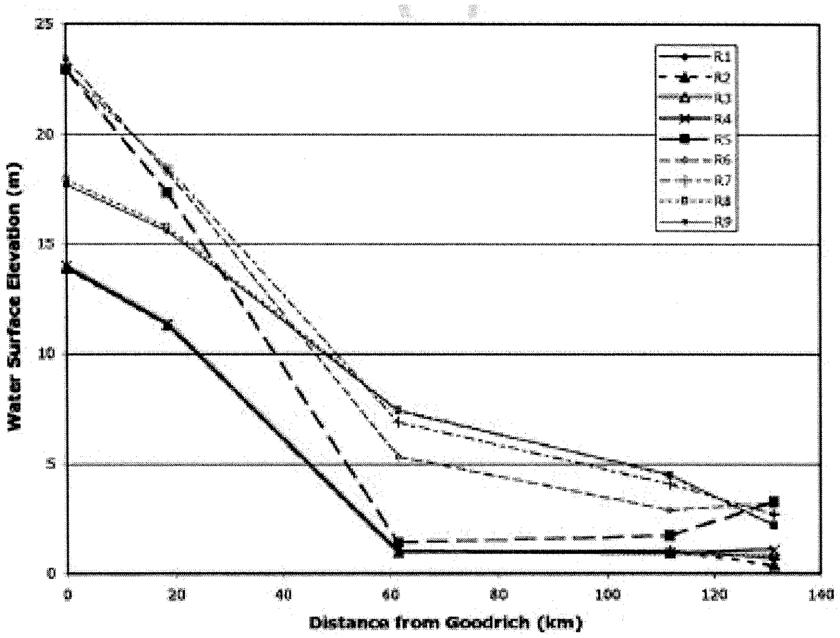


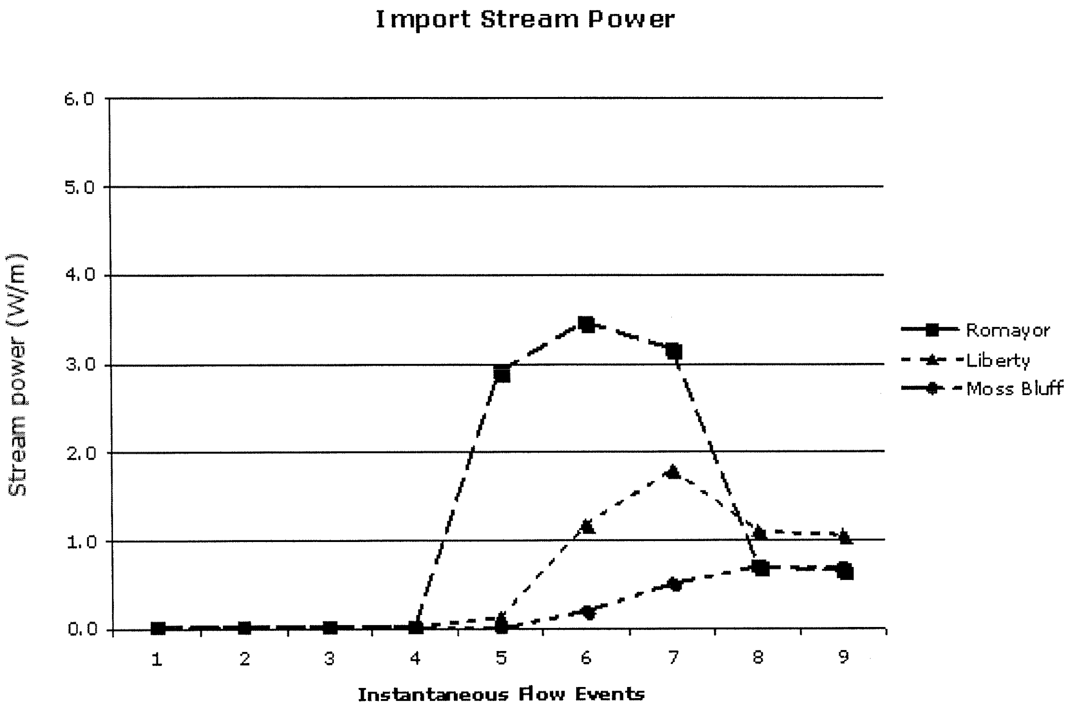
Figure 8. Water surface profiles from Goodrich to Wallisville for nine sample times during the Hurricane Rita flow event.

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 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 water surface gradient 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 to 50 km) are quite large, they allow a first-order assessment of the downstream variation of stream power during the Rita event.

The data set allows calculation of either "import" or "export" stream power for each station (Fig. 9), 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. 10).



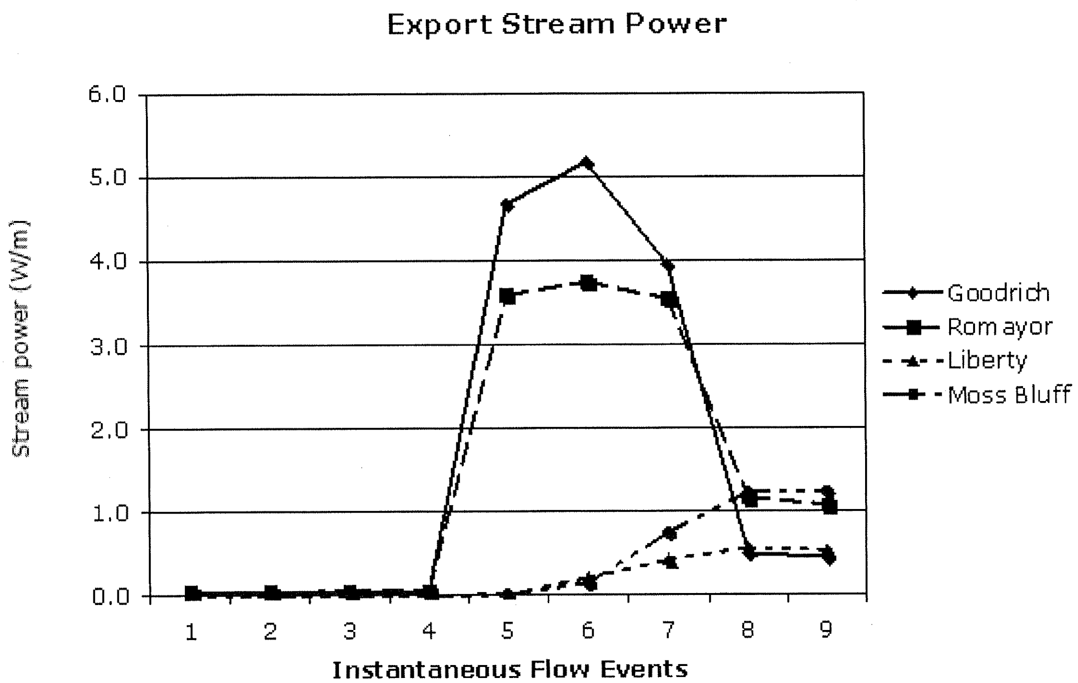


Figure 9. Stream power at Trinity River gaging stations for nine Hurricane Rita instantaneous flows, based on discharge and upstream (import) or downstream (export) slope.

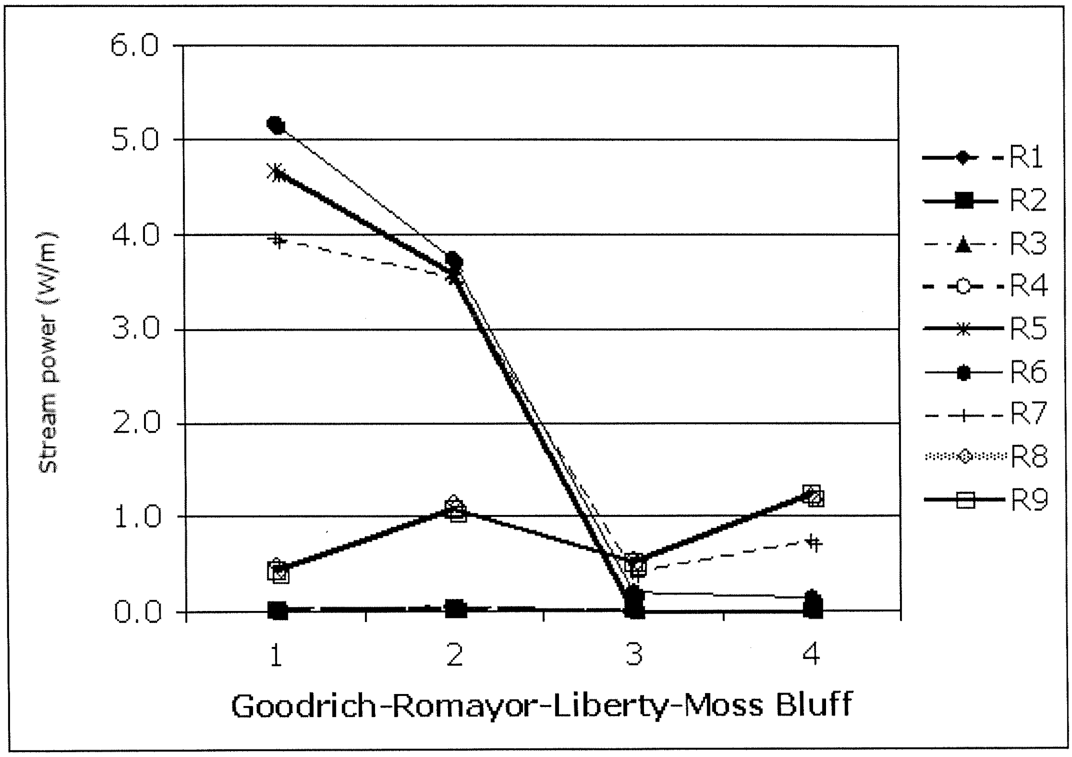


Figure 10. Stream power (export) at four stations for nine Hurricane Rita instantaneous flows.

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 3). This was also noted in the Rita event, as the peak discharge at Goodrich was 39 percent higher than at Romayor.

Mussel Shoals Creek, which joins the Trinity downstream of the Goodrich station (Fig. 11), does so at an angle which is more characteristic of a distributary than a tributary channel. Analysis of topographic gradients from the digital elevation model 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

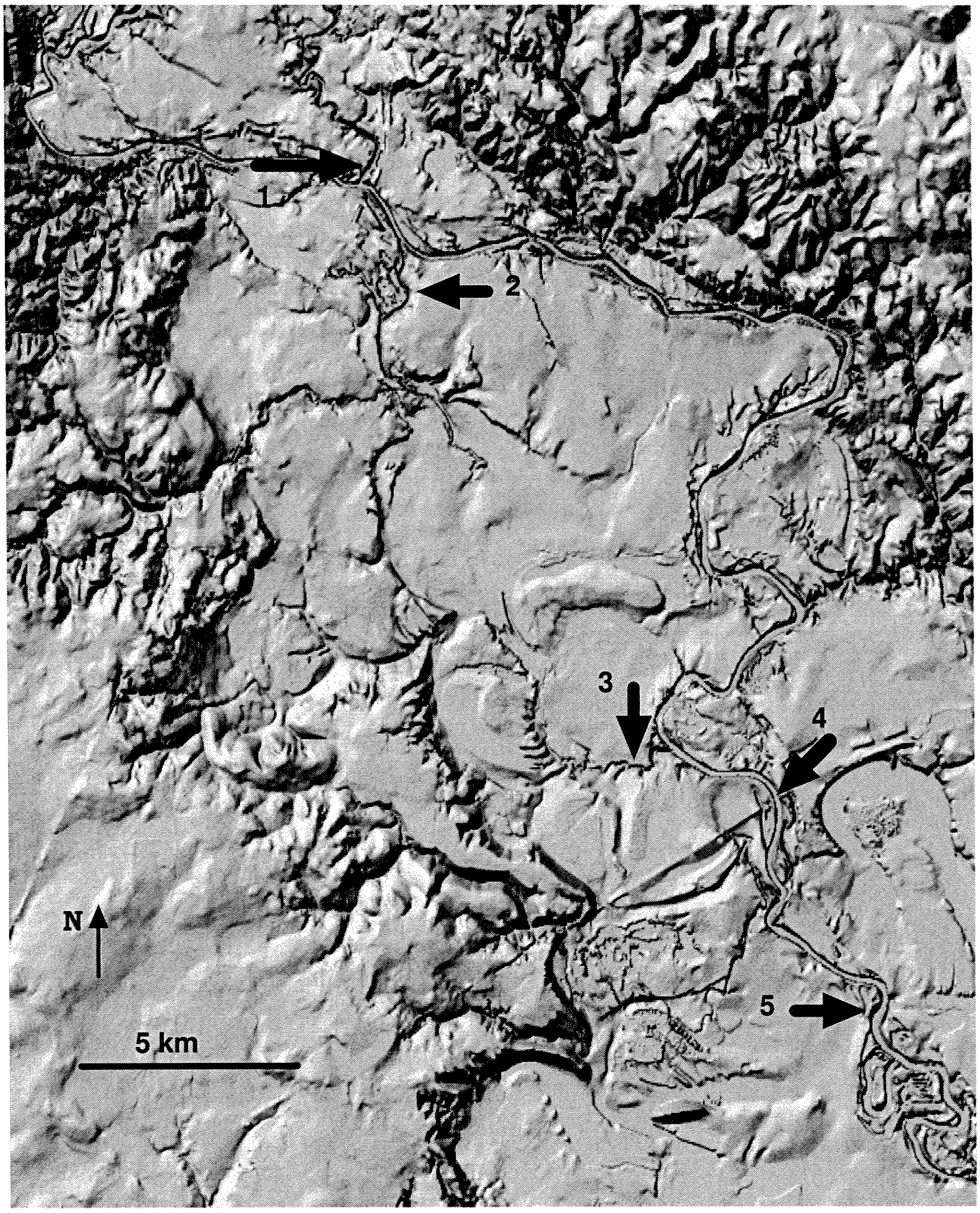


Figure 11. Shaded relief map (50X vertical exagerration) of the lower Trinity River valley in the vicinity of the Goodrich and Romayor gaging stations. Numbered arrows identify (1) Goodrich gage site; (2) Mussel Shoals Creek; (3) Big Creek at the southern end of Grama Grass Bottom; (4) Romayor gage site; and (5) approximate location of the morphological transition zone. The Romayor gage is located at 30°25′30″ N; 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 (23 to 30 ft gage heights). Mussel Shoals Creek begins backflooding from the river as Goodrich stages rise from approximates 21 to 23 m amsl (29 to 35 ft gage heights).

water elevations at Goodrich rise from 21 to 23 m (recorded gage heights of 29 to 35 feet). This is below bankfull stage in this vicinity. During the Rita event stages at Goodrich reached this level late on September 24. Figure 12 shows the longitudinal profile of the creek channel, derived from the DEM, showing that backwater flooding to about 22 m could direct flow upstream.

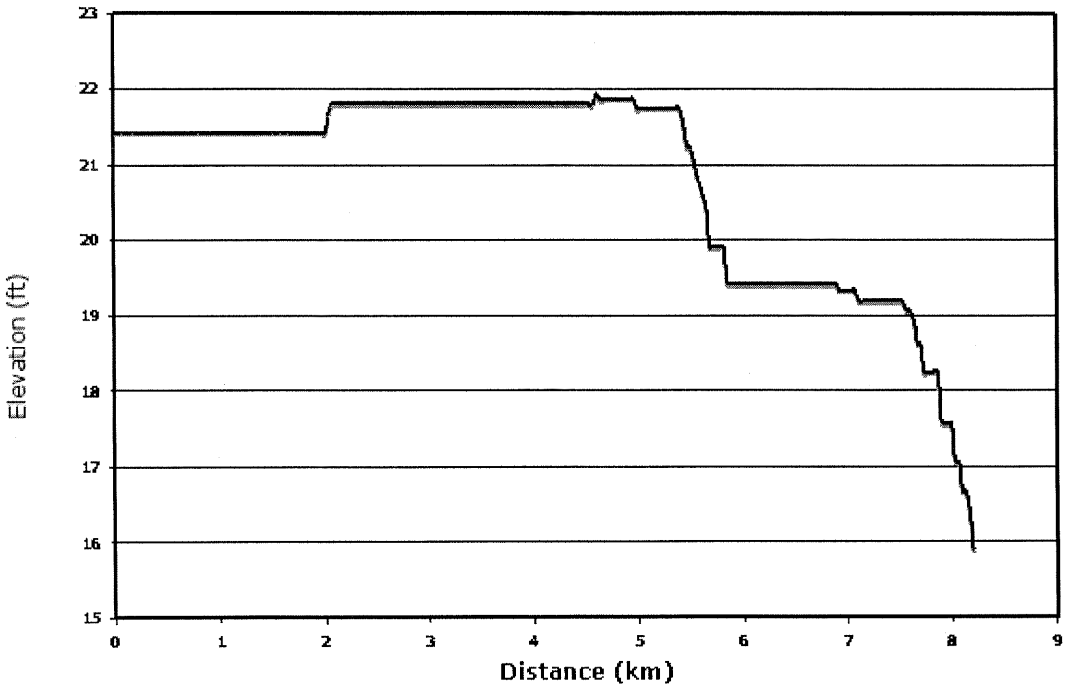


Figure 12. DEM-derived longitudinal profile of Mussel Shoals Creek.

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 significant 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 (gage heights of 23 to 30 feet). 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. 13) are 3.5 to 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. 14) show that topographic gradients lead generally away from the river toward the southeast.

Both Grama Grass bottom and the depression shown in Fig. 14 are palaeomeanders of the ancestral Trinity River, often called Deweyville meanders. The role of this inherited valley morphology in determining modern flow patterns will be addressed in the discussion.

Figure 13. 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.

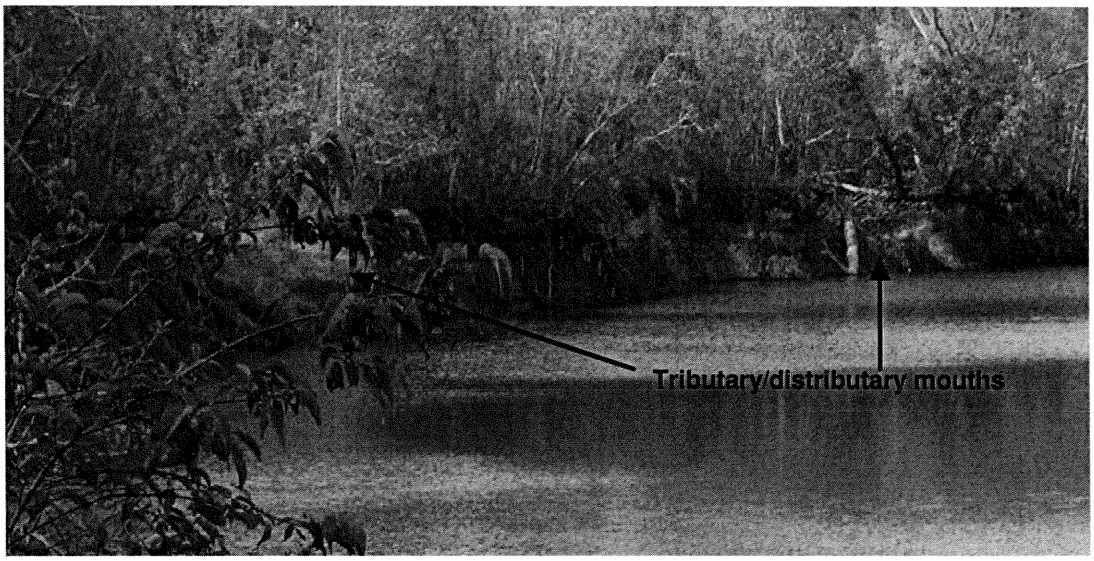


Figure 13 continued: Two of at least five channel inlets at the confluence of Pickett's Bayou and the Trinity River.

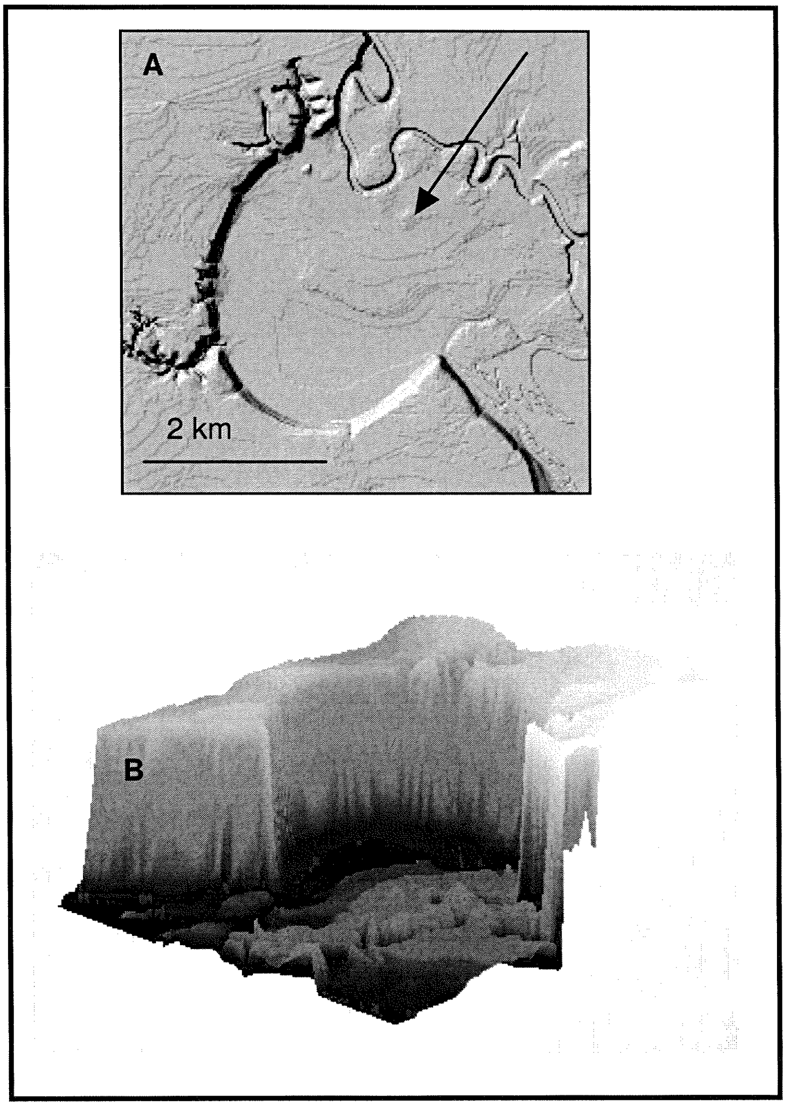


Figure 14. Trinity River Floodplain topography in the vicinity of Pickett's Bayou near Moss Bluff, Texas. A shaded relief map (A) shows the Trinity River, and the large palaeomeander defining the western valley wall. The arrow indicates the direction of view for the shaded surface model (B), which illustrates the topographic controls which tend to direct flow away from, rather than toward, the Trinity River channel.

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 percent 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 into the bottom 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 (Figure 15). 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 $\pm$ 8.3 Ka Otvos (2005). Blum et al. (1995) date the incision into the Beaumont terrades at about 100 ka, broadly consistent with Anderson et al.'s (1994) date of about 110 ka, and within the range of Beaumont dates indicated by Otvos' (2005) synthesis (74 to 116 Ka).

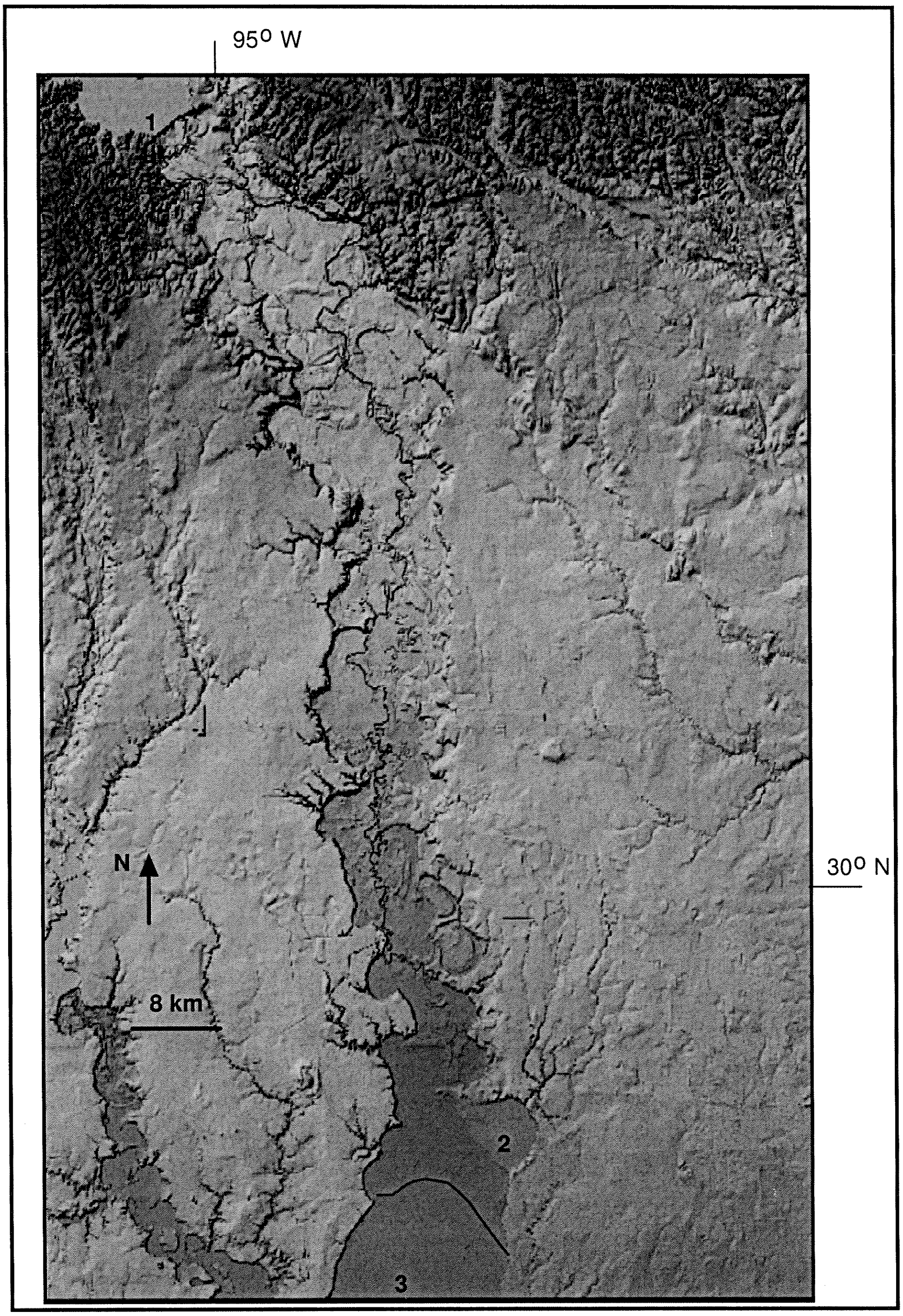


Figure 15. Shaded relief map of the lower Trinity River Valley (50X vertical exaggeration). The approximate Trinity Bay shoreline is drawn on the figure. Numbered sites are (1) lower Lake Livingston and Livingston Dam; (2) Lake Anahuac in the Trinity delta area; and (3) Trinity Bay.

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 et al. 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. Anderson 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. 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 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.

REFERENCES

Alford, J.J., Holmes, J.C. 1985. Meander scars as evidence of major climate change in southwest Louisiana. Annals of the Association of American Geographers 75, 395-403.

Anderson, J.B. Rodriguez, A.B., 2000. Contrasting styles of sediment delivery to the east Texas shelf and slope during the last glacial-eustatic cycle: implications for shelf-upper slope reservoir formation. Gulf Coast Association of Geological Societies Transactions 50, 343-347.

Anderson, J.B., Thomas, M.A., Siringan, F.P., Smyth, W.C. , 1992. Quaternary evolution of the east Texas coast and continental shelf. In C.H. Fletcher III and J.F. Wehmiller (Editors), Quaternary Coastlines of the United States: Marine and Lacustrine Systems, SEPM (Society for Sedimentary Geology), Tulsa, OK, pp. 253-263.

Blum, M. D., Carter, A. E., Zayac, T., Goble, R. J., 2002. Middle Holocene Sea-Level and Evolution of the Gulf of Mexico Coast (USA). Journal of Coastal Research. Special Issue 36, 65-80.

Blum, M.D., Morton, R.A., Durbin, J.M., 1995. "Deweyville" terraces and deposits of the Texas Gulf coastal plain. Gulf Coast Association of Geological Societies Transactions 45, 53-60.

Blum, M.D., Törnqvist, T.E., 2000. Fluvial responses to climate and sea-level change: a review and look forward. Sedimentology 47, 2-48.

Brizga, S.O., Finlayson, B.L., 1994. Interactions between upland catchment and lowland rivers: An applied Australian case study. Geomorphology, 9, 189-201.

Garbrecht, J., Martz, L.W., 1997. The assignment of drainage direction over flat surfaces in raster digital elevation models. Journal of Hydrology 193, 204-213.

Graf, W.L., 1983. Downstream changes in stream power in the Henry Mountains, Utah. Annals of the Association of American Geographers 73, 373-387.

Jain, V., Preston, N., Fryirs, K., Brierly, G. 2006. Comparative assessment of three approaches for deriving stream power plots along long profiles in the upper Hunter River catchment, New South Wales, Australia. Geomorphology, in press.

Knighton, A.D., 1999. Downstream variation in stream power. Geomorphology 29, 293-306.

Lecce, S.A., 1997. Nonlinear downstream changes in stream power on Wisconsin's Blue River. Annals of the Association of American Geographers 87, 471-486.

Magilligan, F.J., 1988. Variations in slope components during large magnitude floods, Wisconsin. Annals of the Association of American Geographers 78, 520-533.

Magilligan, F.J., 1992. Thresholds and the spatial variability of flood power during extreme floods. Geomorphology 5, 373-390.

Morton, R.A., Blum, M.D., White, W.A., 1996. Valley fills of incised coastal plain rivers, southeastern Texas. Transactions of the Gulf Coast Assocation of Geological Societies 46, 321-329.

Nichols, M.M. 1989. Sediment accumulation rates and relative sea-level rise in lagoons. Marine Geology 88, 201-219.

Nichols, M.M., Johnson, G.H., Peebles, P.C., 1991. Modern sediments and facies model for a microtidal coastal plain estuary, the James estuary, Virginia. Journal of Sedimentary Petrology 61, 883-899.

Olive, L.J., Olley, J.M., Murray, A.S.,Wallbrink, P.J., 1994. Spatial variability in suspended sediment transport in the Murrumbidgee River, New South Wales, Australia. In: Variability in Stream Erosion and Sediment Transport, International Association of Hydrological Sciences publication 224, Wallingford, U.K., pp. 241-249.

Otvos, E.G., 2005. Numerical chronology of Pleistocene coastal plain and valley development; extensive aggradation during glacial low sea-levels. Quaternary International 135, 91-113.

Phillips, J.D., 1993. Pre- and post-colonial sediment sources and storage in the lower Neuse River basin, North Carolina. Physical Geography 14, 272-284.

Phillips, J.D., 1997. Human agency, Holocene sea level, and floodplain accretion in coastal plain rivers. Journal of Coastal Research 13, 854-866.

Phillips, J.D., 2003. Alluvial storage and the long term stability of sediment yields. Basin Research 15, 153-163.

Phillips, J.D., Musselman, Z.A., 2003. The effect of dams on fluvial sediment delivery to the texas coast. Proceedings of Coastal Sediments 2003, American Society of Civil Engineers, New York, pp. 1-14.

Phillips, J.D., Slattery, M.C. 2006.. Sediment storage, sea level, and sediment delivery to the ocean by coastal plain rivers. Progress in Physical Geography, in press.

Phillips, J.D., Slattery, M.C., Musselman, Z.A. 2004. Dam-to-delta sediment inputs and storage in the lower Trinity River, Texas. Geomorphology 62, 17-34.

Phillips, J.D., Slattery, M.C., Musselman, Z.A. 2005. Channel adjustments of the lower Trinity River, Texas, downstream of Livingston Dam. Earth Surface Processes and Landforms 30, 1419-1439.

Reinfields, I., Cohen, T., Batten, P., Brierly, G.J., 2004. Assessment of downstream trends in channel gradient, total, and specific stream power: a GIS approach. Geomorphology 60, 403-416.

Rodriguez, A.B., Anderson, J.B., 2000. Mapping bay-head deltas within incised valleys as an aid for predicting the occurrence of barrier shoreline sands: an example from the Trinity/Sabine incised valley. Gulf Coast Association of Geological Societies Transactions 50, 755-758.

Rodriguez, A.B., Fassell, M.L., Anderson, J.B., 2001. Variations in shoreface progradation and ravinement along the Texas coast, Gulf of Mexico. Sedimentology 48, 837-853.

Rodriguez, A.B., Anderson, J.B., Simms, A.R., 2005. Terrace inundation as an autocyclic mechanism for parasequence formation: Galveston estuary, Texas, U.S.A. Journal of Sedimentary Research 75, 608-620.

Thomas, M.A., Anderson, J.B., 1994. Sea-level controls on the facies architecture of the Trinity/Sabine incised-valley system, Texas continental schelf. In: R.W. Dalrymple, R. Boyd, B.Z. Zaitline (Editors), Incised-Valley Systems: Origin and Sedimentary Sequences. SEPM (Society for Sedimentary Geology), Tulsa, OK, pp.63-82.

Wellmeyer, J.L., Slattery, M.C., Phillips, J.D. 2005. Quantifying downstream impacts of impoundment on flow regime and channel planform, lower Trinity River, Texas. Geomorphology 69, 1-13.

PART 2: TRINITY RIVER DELTA ACCRETION RATES

INTRODUCTION

Sedimentation and accretion rates in deltas, marshes, and alluvial floodplains may occur over long time periods when direct measurement is impossible, and in any case are typically discontinuous and slow between major sedimentation events. The common—and often only—recourse is to measure accumulation in terms of thickness or mass and divide by the time period over which the deposition is known or estimated to have occurred. This results in what might be termed a "virtual rate." Sedimentation rates estimated in this way not only include averages of periods of faster or slower sedimentation, but also represent the net of accumulation minus removal (by erosion, oxidation of organic matter, or other processes). Thus, one should be aware that an accretion rate of (say) 5 mm yr^{-1} over 50 years does not necessarily imply steady accumulation at that rate; only that over a 50 year period episodic depositional episodes of various magnitude (perhaps coupled with removal episodes and processes) resulted in a net accretion of 250 mm over that time period. These virtual rates (in this and other studies) are perhaps better conceptualized as net accumulation amounts standardized for comparative purposes to account for the amount of time elapsed.

Accretion rates are most often reported in dimensions of LT^{-1}. This convention will be used here, with rates reported in units of mm yr^{-1}.

Long Term Accretion

White et al. (2002) reported a mean depth of Holocene mud in the Trinity River delta area of 8.1 m. A stratigraphic cross-section across the delta at Wallisville presented by Morton et al. (1996) shows thicknesses of Holocene sediment ranging from approximately 2 to 11 m. The Chambers and Liberty County soil surveys (Crout, 1976; Griffith, 1996) both show Trinity River delta soils deeper than 1.5 to 2 m. Soil survey descriptions usually extend only to a depth of 5 to 7 feet, but on both surveys the soils mapped in the delta are indicated as forming in much thicker recent sediments. Similarly, McEwen (1963) found deltaic sediments to be generally thicker than 3 m, but did not sample the entire thickness. A core on the distal edge of the modern bayhead delta presented by Rodgriguez et al. (2005) extended to 13.5 m in Holocene deposits. Based on this, the characteristic thickness of 8.1 m from White et al. (2002) is used here.

A date of 8 Ka is used as the time of initiation of delta sediment accumulation, based on Rodriguez et al.'s (2005) study of the Galveston estuary and the ancestral incised valley of the Trinity River. That study indicates a terrace inundation event at 7.7 to 8.2 Ka, supported by a basal radiocarbon date of 7.738 $\pm$ 60 Ka (conventional; 8.495 $\pm$ 50 calibrated) in a core in the modern bayhead delta.

The characteristic thickness of the Holocene delta deposits represents a mean accretion rate of about 1 mm yr^{-1}. If local thicknesses are assumed to range from about 2 to 15 m, local mean accretion rates would range from ~0.25 to ~1.9 mm yr^{-1}.

Sedimentation rates are not spatially or temporally uniform, and both the size and location of the delta surface has changed over the course of its Holocene evolution. Nonetheless, the amount of sediment necessary to cover a delta with the modern surface area of 126.1 km^2 (excluding open water) was calculated to obtain an order-of-magnitude, ballpark estimate of the amount of sediment necessary to construct the delta given mean accretion rates up to 3 mm yr^{-1} (Fig. 16). Each mm of sediment thickness over the surface area requires 126,100 m^3 of sediment. While newly-deposited sediment typically has bulk densities of ~ 1 t m^{-3}, the typical bulk densities of floodplain and marsh soils in the delta area is about 1.4 t m^{-3}, due to autocompaction and settling. Thus the cubic meters necessary to cover the delta to a particular thickness was multiplied by 1.4 to obtain the tones of sediment required. Figure 16 also shows the tonnes required to cover 25, 50, and 75 percent of the delta surface.

Despite numerous uncertainties and assumptions, Figure 16 suffices to assess the feasibility that fluvial sediment inputs approximating contemporary inputs (about 70,000 t yr^{-1}) could have constructed the modern delta. The results suggest that river sediment inputs less than 100,000 t yr^{-1} could maintain accretion rates of 1 mm yr^{-1} for only half or less of the contemporary delta surface area. This suggests at least two possibilities:

(1) River sediment loads were higher earlier in the Holocene.
(2) Other, non-fluvial sediment sources are significant contributors to the delta sediment budget.

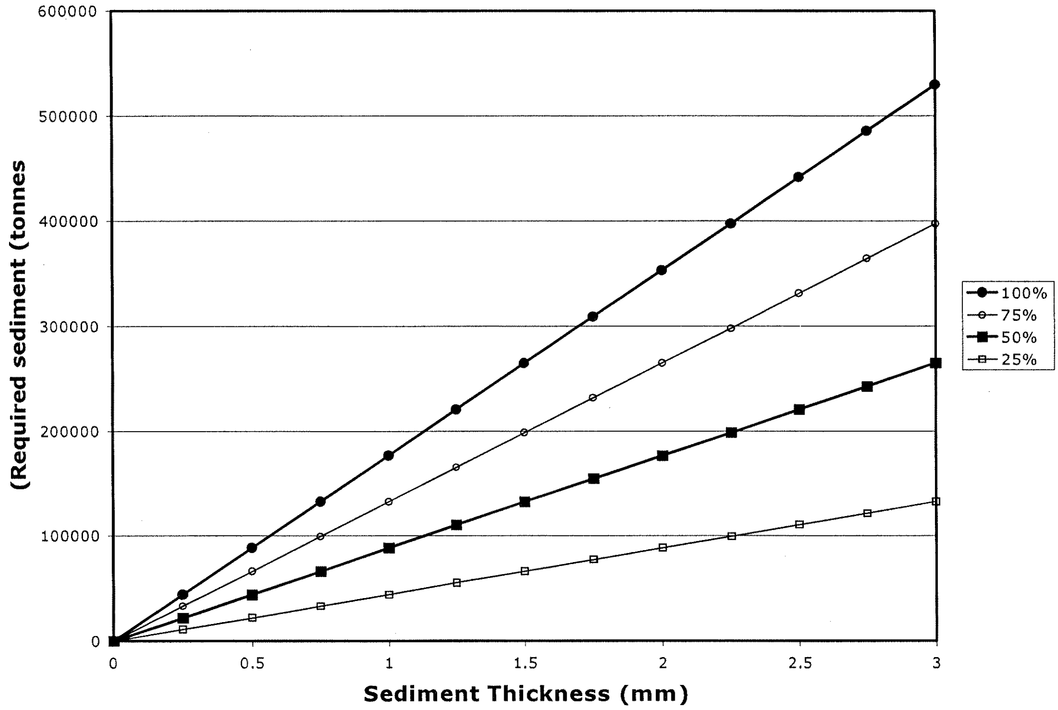


Figure 16. Tonnes of sediment necessary to cover the modern Trinity River delta to specified thicknesses, corresponding to accretion rates up to 3 mm yr^{-1}, assuming 25, 50, 75, and 100 percent of the surface area is accreting.

CONTEMPORARY DELTA SEDIMENTATION RATES

Marsh Sites

White et al. (2002) used ^{210}Pb methods to determine accretion rates at 12 marsh sites in the Trinity River delta (locations in Fig. 17). This method applies to the past 100 years, and the results show rates ranging from 1.6 to 13.05 mm yr^{-1}, with a mean of 5.1 (Table 6). A similar mean rate (5.4 mm yr^{-1}) was reported for the Trinity delta marshes by White and Calnan (1990). An additional core analyzed by Rodriguez et al. (2005) shows a rate of 2.42 mm yr^{-1} over about 720 years.

Rodgriguez et al. (2005) present calibrated radiocarbon dates for a ~` 13.5 m core in the Trinity bayhead delta. Dividing the depth of dated features by their calibrated age yields an estimate of mean accretion rates. The overall rate suggested by the date of the deepest, oldest sample indicates a rate of 1.58 mm yr^{-1} over about 8,500 yr. The nine individual depth increments in the core, however, indicate mean rates of 0.52 to 5.68 mm yr^{-1} over periods ranging from 175 to 2,580 yr.

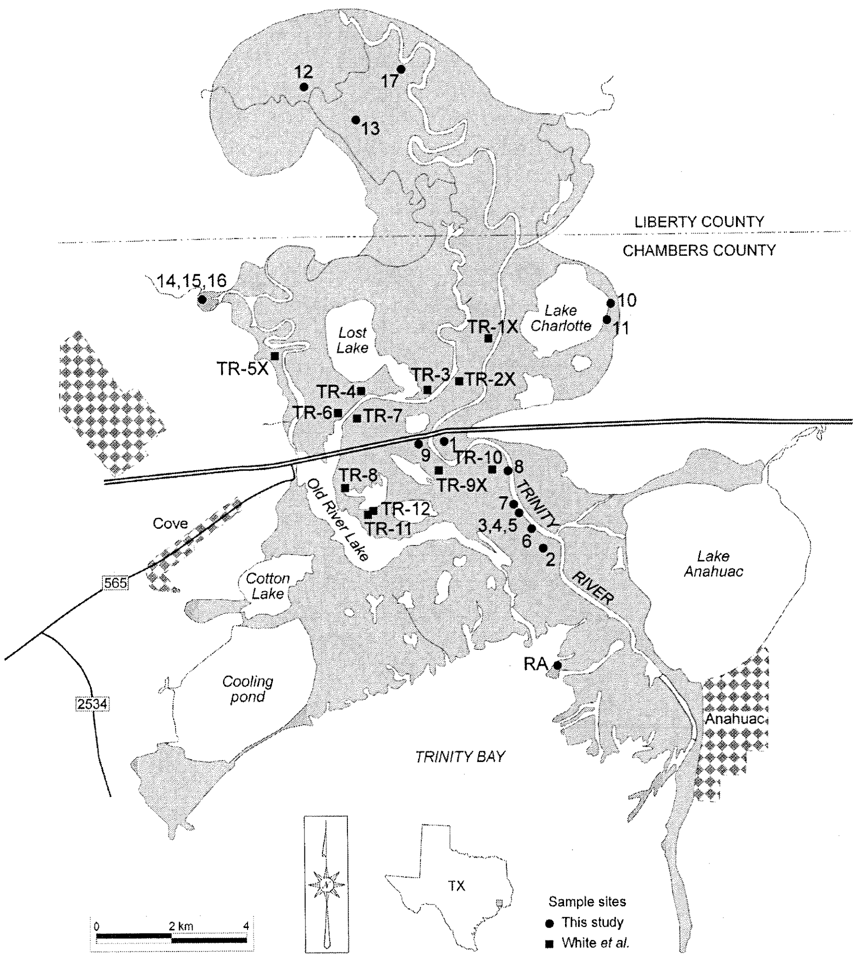


Figure 17. Sample sites in the Trinity River delta area.

Table 6. Accretion rate at Trinity delta marsh sites, using ^{210}Pb methods (from White et al. 2002), with the margin of error indicated. Locations are shown in Figure 18.

Site	Accretion rate (mm yr^{-1})	$\pm$
TR-1X	3.90	0.002
TR-2X	2.92	0.003
TR-3	5.57	0.003
TR-4	9.49	0.014
TR-5X	4.45	0.008
TR-6	3.03	0.030
TR-7	7.19	0.001
TR-8	4.45	0.019
TR-9X	2.81	0.003
TR-10	3.19	0.001
TR-11	1.58	0.006
TR-12	13.05	0.005

Swamp Forest Sites

In this study, recent accretion rates were determined using dendrochronological methods. In essence, this involves measuring the burial of tree roots and basal flares by sediment, and determining the (maximum) date of burial via counts of tree rings. These methods are reviewed and described by Hupp and Osterkamp (1996) and Strunk (1998). They have previously been used in the lower Trinity River (Phillips et al. 2004) and elsewhere in southeast Texas (Phillips 2001). In addition to the necessity of trees, forested sites were used to supplement the marsh sites used in other studies. Results are shown in Table 7 and locations in Figure 18. Rates range from negligible to nearly 17 mm yr^{-1}. Note that because the method depends on significant burial, very low rates of accretion (particularly for young trees) are difficult to detect. The information used to calculate the rates is shown in in Table 8.

The major discrepancy between the ^{210}Pb marsh data and the dendrochronology swamp data is that the latter covers shorter periods of time and is more likely to reflect recent depositional events (or the lack thereof). The century timescale of the marsh dates ensures that at least some significant depositional events will be recorded. Further, the marsh sites are in similar topographic settings, while the forest sites reflect some microtopographic variability associated with ridge-and-swale topography on the floodplain.

Table 7. Accretion rates at Trinity delta forest sites using dendrochronology methods (this study).

Site	Accretion rate (mm yr^{-1})	
1	~0	
2	<1	
3	0 to 3.3	
4	1.2 to 6.5	
5	3.3	
6	~0	Mean rate = 3.3
7	~0	
8	~0	
9	~0	
10	~0	
11	~0	
12	~0	
13	~0	
14	0 to 16.7	
15	16.7	
16	10.0	
17	>8	

Table 8. Description of sites reported in Table 7.

Location	*Trees*	*Burial*	*Age*	*Comments*
1 Wallisville rookery area, left bank, east side of I-10 bridge	Bald cypress	None	N/A	No evidence of recent sedimentation. 10 cm of black organic much over gray clay
2 Wallisville birding trail, 29.8111 N, 94.7405 W	None; sediment deposition on rock riprap assessed	Veneer	5	Adjacent to Phragmites marsh
3 Riverside access area; 29.8255 N, 94.7410 W		None to		13 trees examined; five buried.
4 (same as 3)	Bald cypress	20 to 110	17	
5 (same as 3)		20	6	
6 Right bank, Wallisville area; 29.8242 N 94.7412 W				
7 Riverside access area; 29.8304 N 94.7412 W		None	N/A	2 cm of mud over previous year's litter layer on channel shelf

Location	Trees	Burial	Age	Comments
8 Right bank, Wallisville area; 29.8320 N 94.7428 W		None	N/A	
9 Picnic area adjacent to frontage road, west side of I-10 bridge	Live oaks	None	N/A	
10 east side Lake Charlotte, Cedar Hill Park, 29.8737 N 94.7152 W	Bald cypress	None; some evidence of reworking of depositional veneer	N/A	Sandy, podzolized material apparently from adjacent terrace
Location	*Trees*	*Burial*	*Age*	*Comments*
11 east side Lake Charlotte, 29.8746 N 94.7171 W	Bald cypress	None; some evidence of reworking of depositional veneer	N/A	Sandy, podzolized material apparently from adjacent terrace
12 Trinity NWR Champion Lake; 29.9211 N 94.7997 W	Mixed bottomland hardwoods	None	N/A	No recent sediment deposits
13 Trinity NWR, Pickett's Bayou 29.8746 N 94.7171 W	Mixed bottomland hardwoods	None	N/A	No recent deposits
14 Lost River near CR 1409 29.8739 N 94.8739 W	Green ash	None to		13 trees examined. Four buried.
15 (same as 14)	Red maple	50	3	
16 (same as 14)	Green ash	80	8	
17 Pickett's Bayou at Trinity R. 29.9350 N 94.7790 W	Mixed hardwoods	400 to 600	Ukn	Equipment failure precluded ring count; trees 10 to 20 cm diameter at breast height

Delta Sedimentation Rates: Comparisons and Context

Overall, the data indicate some parts of the delta receiving minimal sedimentation-- too small in magnitude of too slow in rate to be detected via burial of tree root crowns and basal flares. Other parts of the system are accreting at rates in excess of 15 mm yr^{-1}. These rates are consistent with the mean rates recorded for accretion of bayhead deltas in the Gulf Coast region (Table 9), which generally range from about 3 to 8 mm yr^{-1}.

Table 9. Published measurements and estimates of bayhead deltaic sedimentation rates in the Gulf Coast region over periods of 100 years or less. .

Location	Range (mm yr^{-1})	Mean	Source
Trinity River, TX	0 - 16.7	3.3	this study
Trinity River, TX	1.6 - 13.0	5.1	White et al. 2002
Trinity River, TX		5.4	White & Calnan 1990
Lavaca-Navidad, TX	0.9 - 10.3	3.3	White et al. 2002
Nueces, TX	0.7 - 7.6	2.6	White et al. 2002
Colorado River, TX		7.5	White & Calnan 1990
Tensaw River, AL		7.6	Aust et al. 1991

The accretion rates in the delta may also be compared to rates measured using dendrochronology on alluvial floodplains in the lower Trinity River valley between Livingston Dam and the delta. Phillips et al. (2004) reported mean rates of 18 to 40 mm yr^{-1} at three sites. These were supplemented in this study by three additional measurements at the Trinity River National Wildlife refuge area near Hardin, which gave similar results (Table 10). In general, these rates exceed those in the delta area. The lower Trinity River floodplain accretion rates in turn are consistent with measurements in the Gulf Coastal Plain region (Table 11).

Table 10. Dendrogeomorphic estimates of recent floodplain accretion rates at lower Trinity River sites upstream of the delta (first three sites from Phillips et al. 2004; Trinity River National Wildlife Refuge near Hardin, TX from this study)

Site	Number of measurements	Age range (years)	Accretion rate (mm yr^{-1}) Mean	Min	Max
Goodrich	10	1 - 27	18.5	0	41.0
Moss Hill	6	1 - 16	45.4 18.5[1]	3.6	180.0 41.2[1]
Liberty	3	2 - 21	39.9	28.1	56.7
Trinity River NWR	3	6 - 38	43.3	13.1	61.7

[1]Mean and maximum excluding one measurement indicating 180 mm of deposition in one year.

Table. 11. Published measurements and estimates of floodplain accretion rates, Gulf Coastal Plain region.

Location	Range (mm yr^{-1})	Mean	Source
Trinity River, TX	13.1 - 61.7	43.3	This study
Trinity River, TX	0 -180.0	30.4	Phillips et al. 2004
Loco Bayou, TX	11.2 - 61.0	25.0	Phillips 2001
Loco Bayou, TX[1]	0.1 - 3.1	3.0	Yeager et al. 2005
Lake Houston, TX[2]		21.9	Van Metre et al. 2002
Raccouri Old R., LA	34.0 - 58.0	45.7	Rowland et al. 2005
Black Swamp, AR	0.1 - 6.0		Hupp & Morris 1990
Cache River, AR	0 - 26.4		Kleiss 1996

[1]Estimated from data reported in units of g cm^{-2}
[2]Lake bottom sediment accumulation

DISCUSSION

Data from this study show that the most rapidly accreting delta sites are near the river channel near the delta apex, along Lost River, a tributary draining adjacent terraces and uplands, and some near-channel sites in the Wallisville vicinity. Sites further from the Trinity River channel (and a few near the channel) had very low or negligible sedimentation rates. Data from White et al. (2002) are less variable overall, but the sites with the highest and lowest rates are the two closest together.

Nichols (1989) reported sea level rise in the Galveston Bay area of 5.5 mm yr^{-1} for the short term, from tidal gage records, and 1.4 mm yr^{-1} for Holocene time scales. These are similar to the mean Trinity delta marsh accretion rates of 5.1 and 5.4 mm yr^{-1} reported by White et al. (2002) and White and Calnan (1991), and to the estimated mean long-term delta accretion rate estimated in this report. Coastal submergence (accounting for sea level change and land surface elevation changes) is substantially higher, averaging about 7.6 mm yr^{-1} in the Galveston Bay region, from recent tidal gage records and interferometry measurements (Stork and Sneed 2002). White et al. (2002) combine 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} relative sea level rise (coastal submergence).

Subsidence is associated with natural autocompaction of sediments, and can also be influenced by tectonic or isostatic subsidence. The latter are not factors in the contemporary or Holocene Trinity delta area. However, hydrocarbon production in southeast Texas has been shown to induce subsidence (and in some areas fault reactivation) (Morton et al. 2001). Further, in the Galveston Bay area groundwater withdrawal contributes to accelerated subsidence (Stork and Sneed 2002).

Mean sedimentation rates in marshes, and at most specific sites, is inadequate to keep pace with coastal submergence, accounting for the wetland loss and marsh shoreline erosion occurring in the lower delta. At the sites of the most active accretion, rates are greater than coastal submergence, however, suggesting increasing fragmentation of deltaic wetlands as coastal submergence proceeds.

Contemporary fluvial sediment delivery to the lower Trinity River is inadequate to account for the current delta; much less so to keep pace with sea level rise. As mentioned earlier, this implies either higher sediment loads in the past or additional, non-fluvial sediment sources. While paleodischarges (and thus, potentially, sediment loads) were higher earlier in the Quaternary, as indicated by meander scars in Pleistocence alluvial terraces, there is no evidence of a different discharge regime within the last 5 to 10 Ka. Historical, anthropic land use changes have likely raised erosion rates within the Trinity River basin, but the lower Trinity is a sediment transport bottleneck, and increases or decreases in erosion upstream of Romayor have little impact on the delta (Phillips et al. 2004; Phillips and Slattery 2006).

Other sediment sources include aeolian, slopewash from adjacent terrace uplands, coastal-derived material resuspended from Trinity Bay and delta channels and lakes, and sediment imported from local tributaries such as Lost River. Autocthonous organic matter is also a significant component of marsh and backwater swamp sediments. Aeolian inputs are assumed to be minor. Slopewash from adjacent terraces is locally significant (as at two sites on east side of Lake Charlotte) but cannot account for a significant portion of overall delta accumulation. Characteristics of sediment collected in the delta, including magnetic signatures, angularity, and iron oxide coatings suggests that most are reworked—that is, they have been stored for a significant time within the fluvial or deltaic system as opposed to material recently eroded from uplands and transported downstream (Slattery and Phillips 2005).

While many streams within the delta region, such as Old River and Pickett's Bayou, are distributaries whose drainage area is wholly or chiefly within the delta, other tributaries, however, such as Lost River and Turtle Bayou, originate in uplands and may contribute significant sediment, as indicated by floodplain sedimentation sites along Lost River reported here.

CONCLUSIONS

Recent sedimentation rates in the Trinity River delta average about 3.3 mm yr^{-1} on forested alluvial floodplains within the delta, and 5.1 mm yr^{-1} in lower-lying marsh areas. These rates are higher than apparent Holocene mean accretion rates, but are considerably lower than recent rates of coastal submergence in the region.

Contemporary sediment delivery by the Trinity River is not adequate to account for the Holocene growth of the delta, and there is no independent evidence of higher sediment loads earlier in the Holocene. Trinity River inputs appear to be supplemented by reworking and resuspension of local Trinity Bay and delta sediments, autochthonous organic matter, sediment delivery from tributaries in the delta area, and (locally) slopewash from adjacent terraces.

REFERENCES

Aust, W.M., Lea, R., Gregory, J.D. 1991. Removal of floodwater sediments by a clearcut tupelo-cypress wetland. Water Resources Bulletin 27: 111-116.

Crout, J.D. 1976. Soil Survey of Chambers County, Texas. Washington: USDA Soil Conservation Service.

Griffith, K.L. 1996. Soil Survey of Liberty County, Texas. Washington: USDA Soil Conservation Service.

Hupp, C.R., Morris, E.E. 1990. A dendrogeomorphic approach to measurement of sedimentation in a forested wetland, Black Swamp, Arkansas. Wetlands 10: 107-124.

Hupp, C.R., Osterkamp, W.R. 1996. Riparian vegetation and fluvial geomorphology. Geomorphology 14: 277-295.

Kleiss, B.A. 1996. Sediment retention in a bottomland hardwood wetland in eastern Arkansas. Wetlands 16: 321-333.

McEwen, M.C. 1963. Sedimentary Facies of the Trinity River Delta, Texas. PhD dissertation, Rice University, Houston, 99 p

Morton, R.A., Blum, M.D., White, W.A., 1996. Valley fills of incised coastal plain rivers, southeastern Texas. Transactions of the Gulf Coast Assocation of Geological Societies 46: 321-329.

Morton, R.A., Purcell, N.A., Peterson, R.L., 2001. Shallow Stratigraphic Evidence of Subsidence and Faulting Induced by Hydrocarbon Production in Coastal Southeast Texas. U.S. Geological Survey Open File Report 01-274

Nichols, M.M. 1989. Sediment accumulation rates and relative sea-level rise in lagoons. Marine Geology 88: 201-219.

Phillips, J.D., 2001. Sedimentation in bottomland hardwoods downstream of an east Texas dam. Environmental Geology 40, 860-868.

Phillips, J.D., Slattery, M.C. 2006.. Sediment storage, sea level, and sediment delivery to the ocean by coastal plain rivers. *Progress in Physical Geography* in press.

Phillips, J.D., Slattery, M.C., Musselman, Z.A. 2004. Dam-to-delta sediment inputs and storage in the lower Trinity River, Texas. Geomorphology 62: 17-34.

Rodriguez, A.B., Anderson, J.B., Simms, A.R., 2005. Terrace inundation as an autocyclic mechanism for parasequence formation: Galveston estuary, Texas, U.S.A. Journal of Sedimentary Research 75: 608-620.

Rowland, J.C., Lepper, K. Dietrich, W.E., Wilson, C.J., Sheldon, R. 2005. Tie channel sedimentation rates, oxbow formation age and channel migration rate from optically stimulated luminescence (OSL) analysis of floodplain deposits. Earth Surface Processes and Landforms 30: 1161-1179.

Slattery, M.C., Phillips, J.D. 2005. Relative Importance of Fluvial and Nonfluvial Sediment Sources to Galveston Bay. Report to the Texas Water Development Board, Austin.

Stork, S.V., Sneed, M. 2002. Houston-Galveston area from space--A new tool for mapping land subsidence. U.S. Geological Survey Fact Sheet 110-002, URL: http://water.usgs.gov/pubs/fs/fs-110-02/

Strunk, H. 1998. Dating of geomorphological processes using dendrogeomorphological methods. Catena 31: 137-151.

Van Metre, P.C., Sneck-Fahrer, D.A. 2002. Water Quality Trends in Suburban Houston, Texas, 1954-1997, as Indicated by Sediment Cores from Lake Houston. U.S. Geological Survey Fact Sheet 040-02, 6 p.

White, W.A., Calnan, T.C., 1991. Submergence of vegetated wetlands in fluvial-deltaic area, Texas Gulf coast. In: Coastal Depositional Systems of the Gulf of Mexico. 12th Annual Research Conference, Society of Economic Paleontologists and Mineralogists, Gulf Coast Section, 278-279.

White, W.A., Morton, R.A., Holmes, C.W. 2002. A comparison of factors controlling sedimentation rates and wetland loss in fluvial-deltaic systems, Texas Gulf coast. Geomorphology 44, 47-66.

Yeager, K.M., Santschi, P.H., Phillips, J.D., Herbert,B.E. 2005. Suspended sediment sources and tributary effects in the lower reaches of a coastal plain stream as indicated by radionuclides, Loco Bayou, Texas. Environmental Geology 47: 382-395.