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Sediment storage, sea level, and sediment delivery to the ocean by coastal plain rivers

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Abstract: Coastal and marine sedimentary archives are sometimes used as indicators of changes in continental sediment production and fluvial sediment transport, but rivers crossing coastal plains may not be efficient conveyors of sediment to the coast. Where this is the case, changes in continental sediment dynamics are not evident at the river mouth. Stream power is typically low and accommodation space high in coastal plain river reaches, resulting in extensive alluvial storage upstream of estuaries and correspondingly low sediment loads at the river mouth. In some cases there is a net loss of sediment in lower coastal plain reaches, so that sediment input from upstream exceeds yield at the river mouth. The lowermost sediment sampling stations on many rivers are too far upstream of the coast to represent lower coastal plain sediment fluxes, and thus tend to overestimate sediment yields. Sediment which does reach the river mouth is often trapped in estuaries and deltas. Assessment of sediment flux from coastal plain rivers is also confounded by the deceptively simple question of the location of the mouth of the river. On low-gradient coastal plains and shelves, the location of the river mouth may have varied by hundreds of kilometers due to sea-level change. The mouth may also differ substantially according to whether it is defined based on channel morphology, network morphology, hydrographic or hydrochemical criteria, elevation of the channel relative to sea level, or the locus of deposition. Further, while direct continent-to-ocean flux may be very low at current sea-level stands, sediment stored in estuaries and lower coastal plain alluvium (including deltas) may eventually become part of the marine sedimentary package. The role of accommodation space in coastal plain alluvial sediment storage has been emphasized in previous work, but low transport capacity controlled largely by slope is also a crucial factor, as we illustrate with examples from Texas.

Key words: alluvial storage, coastal plain rivers, coastal sediments, river mouth, sediment delivery, sedimentary archives.

I Introduction

Coastal and marine basins are the ultimate sink for most riverborne sediment, and rivers are the major source of ocean sediment. Coastal and marine sedimentary archives thus represent a potential record of changes

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in river sediment transport and continental sediment production driven by human impacts, climate change, tectonics, or other environmental changes. This assumes that significant changes in erosion and sediment flux in drainage basins is recorded in sediment delivery to the coast. However, the connection between changes in sediment dynamics within a fluvial system and sediment loads at the river mouth is not always strong or direct. The small mountainous drainage basins that provide (relative to their area) a disproportionately large proportion of river sediment yield to the sea (Milliman and Syvitski, 1992; Ludwig and Probst, 1998) may indeed exhibit relatively direct connections between drainage basin sediment dynamics and coastal/marine sediment inputs. In California, for instance, dam construction in small mountainous rivers resulted in a quick and noticeable reduction in sediment supplies to the coast (Willis and Griggs, 2003). However, in some rivers sediment output or accumulation in sedimentary basins has been remarkably consistent despite changes in climate, sea level, vegetation, and human impacts (Summerfield and Hulton, 1994; Gunnell, 1998; Métivier and Gaudemar, 1999; Dearing and Jones, 2003; Phillips, 2003a). In some cases this may be because an overriding control such as relief or tectonic forcing overwhelms variations in climate and other factors, but several authors emphasize the role of alluvial buffering (Métivier and Gaudemar, 1999; Dearing and Jones, 2003; Phillips, 2003a). Alluvial buffering implies that sediment storage and timelags may make the output from some drainage basins slowly or relatively unresponsive to environmental change. This is because alluvial storage during periods of high sediment supply and remobilization when supply is low relative to transport capacity minimize variations in sediment output.

Geomorphologists have long recognized (eg, Meade, 1982) that coastal plain rivers may deliver only a fraction of their sediment load to the coast, and that sediment gaging stations (generally well upstream of the coast) typically overestimate the latter. Despite this, sediment gaging records continue to be used to assess factors such as the potential effects of dams and human landscape modifications on sediment delivery to the sea (Pont *et al.*, 2002; Vorosmarty *et al.*, 2003; Walling and Fang, 2003). Likewise, coastal and marine sedimentary archives are interpreted for evidence of environmental change within drainage basins (eg, Sandweiss, 2003; Vaalgamaa and Korhola, 2003).

The purpose of this paper is to review the factors contributing to low sediment output in coastal plain rivers and complicating efforts to determine sediment yield to the coast, with particular emphasis on the deceptively simple problem of defining the mouth or outlet of the river. We also seek to explore the role of transport capacity (in addition to accommodation space) in alluvial storage and buffering, with particular emphasis on channel slope. While some new data is introduced, this work is based largely on a review of the literature, and on a reinterpretation of data from our own previous and ongoing work.

Milliman and Syvitski (1992) point out that the sediment contributions of large rivers to the ocean is often overestimated because most such rivers discharge to passive margins and marginal seas, with much sediment sequestered in subsiding deltas. Dearing and Jones (2003) emphasize the lack of responsiveness of large drainage basins to environmental changes, such that continental margin sedimentary archives may be poor indicators of such change. In some cases storage bottlenecks and timelags create an essential decoupling, such that changes in sediment regimes in the upper basin are simply not reflected in the lower river reaches (Brizga and Finlayson, 1994; Olive *et al.*, 1994; Phillips, 1995; 2003a; 2003b; Fryirs and Brierly, 1999; 2001; Shi *et al.*, 2003; Phillips *et al.*, 2004).

This implies, at least for large rivers on passive margins, that recent trends in river sediment loads discussed by Walling and Fang (2003), however important they may be upstream, may have minimal impact on land-to-ocean sediment fluxes. Vörösmarty *et al.* (2003) document extensive sediment retention by dams and the potential major global impact on sediment delivery to the sea, agreeing with Walling and Fang (2003) that dams are the single most important contemporary impact on river sediment transport, and echoing Graf (1999), who pointed out that hydrologic impacts of dams far exceed those of even the most extreme predictions of climate change. If upper and lower basins are decoupled, however, dam impacts may not be significant at the river mouth, however significant the effects upstream. In southeast Texas, for instance, even where large reservoirs controlling 75–95% of the drainage area retain massive amounts of sediment behind dams, no dam-related changes in alluvial sedimentation are noticeable in the lowermost river reaches (Phillips, 2003b; Phillips *et al.*, 2004).

Though large alluvial rivers transport a disproportionately small amount of sediment per unit drainage area compared to steeper, smaller rivers draining directly to the coast, an overwhelming total amount of the global fluvial sediment transport is by large alluvial rivers, many of which cross extensive coastal plains. Understanding sediment delivery in coastal plain rivers is therefore a prerequisite to understanding fluvial sediment inputs to marine environments, the fate of sediments in alluvial rivers, and the extent to which coastal and marine sedimentary archives reflect upstream changes in sediment production and transport. We focus here on relatively large rivers with headwaters above the coastal plain that cross extensive coastal plains, though the generalizations also apply to streams confined to coastal plain environments.

Some coastal plain rivers do have high sediment loads and delivery ratios (the Amazon, for example), and the mouths of some rivers may respond relatively rapidly to upstream changes. The Atchafalaya River, Louisiana, and the Colorado River, Texas, for instance, have begun building new deltas in historic or recent times, while sediment supply to the Nile delta was severely reduced by the Aswan high dam and by irrigation and drainage canals (Stanley, 1996). The focus here, however, is on coastal plain rivers in which sediment yield at the mouth may be very low, due to our impression that the extensive alluvial sediment storage in the lower reaches of coastal plain rivers has been underappreciated, and that the delivery of sediment to the coast is frequently overestimated. We do not claim all coastal plain rivers have low rates of sediment delivery to the ocean, or that delivery rates are constant in even the broadest sense. Our focus is simply to address the issue of why so many coastal plain rivers transport so little sediment to the coast.

II Sediment storage, sinks, and transport

Three factors seem to be critical in explaining low sediment delivery ratios in coastal plain rivers: estuarine, deltaic, and coastal wetland sediment trapping; alluvial storage upstream of estuaries; and low sediment transport capacity.

1 Estuaries as sediment sinks

Estuaries, deltas, and coastal wetlands may be effective sediment traps. In the US Atlantic drainage, Meade (1982) considers estuaries and marshlands to be the ultimate sink for river sediments, at least on a millennial timescale. He estimates that certainly less than 10%, and probably less than 5%, of river sediment reaches the continental shelf or deep ocean. Zhang *et al.* (1990) found that estuarine sinks lead to a similarly small portion of sediment from the Huanghe River reaching the sea. Milliman and Syvitski (1992) note that fluxes to oceans from large rivers (nearly all of which discharge onto passive margins or marginal seas) are overestimated by data from gaging stations due to sediments sequestered in subsiding deltas.

Radionuclide tracers show that the upper Chesapeake Bay (USA) within and above the turbidity maximum retains nearly all the fluvial sediment delivered to it (Donoghue *et al.*, 1989), and Colman *et al.* (1992) concur that the vast majority of fluvial sediment from the Susquehanna River is stored in the upper bay. Colman *et al.* (1992) speculate that this may move down-bay as the upper bay fills, and generally frame their discussion of Holocene sedimentation in the Chesapeake in the context of the geologically ephemeral, rapidly changing nature of coastal plain estuaries, a theme that will be addressed below. Though fluvial inputs are small compared to shore erosion, Marcus and Kearney (1991) found that sediment from the South River, Maryland, is predominantly trapped in river mouth marshes and subtidal storage within a kilometer of the river mouth.

A sediment budget for the tidal fluvial/estuarine transition zone of the Raritan River, New Jersey, shows that the lower river is a sediment sink for fluvial inputs (Renwick and Ashley, 1984). The Albemarle-Pamlico estuarine system (North Carolina), is an efficient trap for all particulates (Harned *et al.*, 1995; McMahon and Woodside, 1997). Though fluvial inputs to the Neuse River estuary (a tributary to the Albemarle-Pamlico) are very low inputs due to extensive storage of coastal plain alluvium (Phillips, 1993; 1997), the Neuse is an efficient trap for those particulates that are delivered (Benninger and Wells, 1993).

In Texas, while Morton (1994) believes that barrier islands are predominantly made of reworked delta sands, he notes that where rivers discharge into estuaries they do not directly contribute to barrier island sediment budgets. This is broadly consistent with Anderson *et al.* (1992), who found that southeast Texas barriers are composed of offshore sands, largely derived from drowned deltaic and fluvial deposits of the Trinity and Sabine Rivers.

Thus, in some cases coastal plain rivers may contribute little sediment to the ocean because of trapping in estuarine environments. This raises the question of the extent to which upstream changes are reflected in estuarine sedimentation.

2 Alluvial sediment storage in coastal plain rivers

Multiple studies have shown that some rivers deliver relatively little sediment to the coastal zone, such that sediment delivery to upper estuaries may be severely limited. Sediment delivery ratios of total basin erosion to sediment yield at the outlet of less than 10% on an average annual basis has been shown for a number of coastal plain rivers, or upper- and lower-basin decoupling. These include rivers of China (Zhang *et al.*, 1990; Shi *et al.*, 2003), Australia (Brizga and Finlayson, 1994; Olive *et al.*, 1994; Fryirs and Brierly, 1999; 2001) the south Atlantic US coastal plain (Phillips, 1991; 1992a; 1992b; 1993; 1997; Slattery *et al.*, 2002), and Texas Gulf coastal plain (Phillips, 2003b; Phillips *et al.*, 2004). The majority of the imbalance between sediment production and sediment yields at the basin mouth is accounted for by source-to-sink timelags, and extensive colluvial and alluvial storage. The well-known relationship between sediment delivery ratios and drainage area, whereby the delivery ratio tends to get smaller as contributing areas increase (Walling, 1983; Dearing and Jones, 2003), reflects this phenomenon.

Studies of sediment provenance in infilling estuaries with extensive inland drainage basins sometimes show that fluvial inputs are small compared to coastal and marine sediment sources. This is also indicative of extensive alluvial storage in the rivers feeding the estuary.

In tributary estuaries of the Chesapeake Bay system, for example, shoreline erosion seems to be a more important sediment source than rivers. Rapid sedimentation in upper and central portions of tributary estuaries has often been attributed to accelerated post-European-settlement soil erosion, but data from Maryland suggest that coastal shoreline erosion is the dominant sediment input along the entire length of many tributary estuaries (Marcus and Kearney, 1991). Coastal contributions to the South River estuary sediment budget are 4–12 times higher than fluvial, and fluvial inputs have little direct impact on sediment accumulation in the majority of the estuary (Marcus and Kearney, 1991). This is consistent with Yarbro *et al.* (1983), whose sediment budget for the Choptank River estuary showed that shoreline erosion in the estuary contributes seven times more sediment than fluvial inputs. Even in the main Chesapeake estuary, the drowned valley of the Susquehanna River, Skrabal's (1991) analysis of clay minerals shows a dominantly marine sediment source, including landward sites previously thought to be dominated by river inputs. Colman *et al.* (1992) indicate that both the continental shelf and Susquehanna River are important sediment sources for the bay, but that over the Holocene marine sources are probably dominant. The Chesapeake and other Atlantic drainage estuaries of the USA were found by Meade (1969) to have landward transport of bottom sediments, and infilling from marine sources, up to the limit of upstream bottom flow.

The main sediment source for the inner Severn River estuary (UK) is probably the estuary itself, not the river (Hewlett and Birnie, 1996). In the Savannah River estuary (USA), isotope tracers show that 65% of inorganic suspended sediments and the upper five cm of benthic sediments are of marine origin (Mullholland and Olsen, 1992). The percentage is even higher in lower, higher-salinity portions of the estuary – up to 99% for salinities greater than 14 ppt – and Mullholland and Olsen (1992) emphasize the importance of the Savannah estuary (the drowned valley of a large river system), and other estuaries on submergent coastlines, as sinks for particles from the coastal ocean. Chemistry of sediments in the Neuse River estuary indicates that little modern river sediment is stored there, and that marine sediment sources and landward transport are prominent (Benninger and Wells, 1993). In van Dieman Gulf, northern Australia, Woodroffe *et al.* (1993) found that, despite a drainage basin of $>10{,}000\,km^2$, fluvial sediment sources were inadequate to account for observed infill and coastal progradation, and concluded that the major sediment source is seaward.

The point is not to suggest that dominantly coastal and marine sediment sources are the norm for all estuaries; there are examples of estuaries with dominantly fluvial sources, and of estuaries with dominantly fluvial sources at the upstream and marine at the downstream ends. Rather, the aim is to show that, even in estuaries with large fluvial drainage basins, much of the fluvial sediment may be stored upstream of the estuary.

3 Sediment transport capacity

While the specific predictive equations for fluvial sediment transport continue to be controversial and uncertain, it is generally agreed that (assuming a transport-limited system), sediment transport is proportional to stream power, which provides a good index of transport capacity.

Cross-sectional stream power is given by:

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

where γ is the specific weight of water, Q is discharge, and S the slope. This represents the total transport capacity of the river at a given cross-section as a rate of energy expenditure. The stream power per unit weight of water is:

$$P_u = \gamma Q\, S/\gamma w\, d = VS \qquad (2)$$

where A is cross-sectional area, V is mean velocity, and Q = AV.

While discharge generally increases downstream (with exceptions to be discussed below), large decreases in slope may occur in the lower reaches of coastal plain rivers. This occurs due not only to regional slope controls (ie, crossing a low-gradient, low-relief plain), but also to the fact that lower river reaches may be cut to below sea level, and may be subject to backwater effects from lunar or wind tides.

Coastal plain rivers are often perceived as sluggish, and velocities may indeed be quite low in lowermost reaches. This could result in very low unit stream power and consequently low transport capacity relative to the quantity of incoming water and sediment. Where backwater effects occur, discharge can also be reduced. These effects may extend well upstream of the head of the estuary. In the Trinity River, Texas, for example, the channel thalweg is cut to below sea level as far as 110 km upstream of Trinity Bay (Phillips *et al.*, 2005). In the James River, Virginia, the tidally influenced zone includes the entire coastal plain portion of the River (Nichols *et al.*, 1991).

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 Texas Water Development Board collected suspended sediment samples at several sites from 1965 to 1989. Sediment loads at the Romayor station, the lowermost station with suspended sediment records from the US Geological Survey, were two orders of magnitude higher than those at Liberty, downstream where the channel is cut to below sea level (Phillips *et al.*, 2004). This generalization applies to total sediment loads, sediment concentrations, and sediment yields (load per unit area). Kim's (1990) two years of measurements of sediment flux from the Neuse River, North Carolina, into its upper estuary are about two-thirds lower than typical yields on coastal plain tributaries in the region (Phillips, 1995).

The issue of stream power and transport capacity will be explored further in the case study.

4 Gaging station locations

River sediment contributions to estuaries and the sea are generally based on suspended (and occasionally bedload) sediment records at the lowermost gaging station at which sediment transport is regularly monitored. These are often a considerable distance upstream of the river mouth. In rivers of the Texas coastal plain, for example (Sabine, Neches, Trinity, Brazos, and Colorado), 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, records from these stations will result in overestimates of sediment loads delivered to the coast.

Milliman and Syvitski (1992), in a compilation of sediment loads from rivers around the world, include entries for both the Tar and Pamlico Rivers, North Carolina. These are actually the same river; the name changes near the head of the estuary. The listing for the Tar River is apparently based on sediment records from the station at Tarboro, North Carolina, which shows a mean annual sediment load of 1.1×10^6 tonnes, and a yield of $2.0\,\mathrm{t\,km^{-2}\,yr^{-1}}$. The listing for the Pamlico River is apparently based on records from a station upstream at Louisburg, North Carolina, which is actually representative only of the upper, Piedmont portion of the watershed. This gives an annual load of 2.1×10^6 and a yield of $19\,\mathrm{t\,km^{-2}\,yr^{-1}}$. Beyond the name confusion (understandable in a data compilation and insignificant with respect to the message of the paper), the mean annual loads show not only the expected decline in yields with basin area, but also a decline in total sediment loads, indicating an alluvial storage bottleneck between Louisburg and Tarboro. A similar phenomenon is reported for the nearby Neuse River by Simmons (1988).

The storage and load reductions in the Louisburg-Tarboro reach are likely minor compared to those further downstream, as the Tarboro station is about 78 km upstream of the estuary. Extensive storage in this area, and upper-lower basin decoupling with respect to sediment, has been demonstrated by Phillips (1992a). The streamflow gaging station 48 km downstream of Tarboro at Greenville, North Carolina, shows clear evidence of backwater effects and predominantly wind-driven water level and flow changes transmitted upstream from the estuary. The gage datum is more than a meter below sea level. Stations downstream at Grimesland and Washington, just above and below, respectively, the head of the estuary, show frequent negative (upstream) flows. The amount of sediment passing Greenville, Grimesland, and Washington in the downstream direction would be expected to decrease substantially compared to the upstream stations, a conclusion borne out by pedologic and sedimentological evidence (Phillips, 1992a).

Another illustration is given by the Yellow River, China. Xu (2003) indicates that since the 1970s a decline in sediment yield to the sea can be linked to precipitation and land-use change. Xu's conclusions with respect to sediment yield at the Lijin gaging station, which is used to reflect river input to the delta and coast, are sound. But the work of Shi *et al.* (2003; see also Zhang *et al.*, 1990) on the sediment budget of the Yellow River delta shows that loads at Lijin, 53 km upstream of the delta apex, overrepresent delivery to the coast. Sediment inputs to the delta are estimated by Shi *et al.* (2003) based on accounting for variable but considerable alluvial accumulation in the intervening valley – that is, delivery to the delta is less than what passes Lijin.

III The mouth of the river

In the discussion above, the mouth of the river has been defined simply as the point at which a well-defined dominant channel ceases to exist at an open-water estuary or delta apex. In the context of land-to-sea sediment fluxes it is worth considering this question more carefully. As discussed below, there are several possible criteria for defining the river mouth. As the location of these may vary considerably along the valley, the definition of the mouth could conceivably result in large variations in estimates of sediment export from a fluvial system. This will be discussed in general terms below, and illustrated via case studies.

Nichols *et al.* (1991) indirectly addressed this issue in their study of the James River, Virginia. The estuary, a tributary to the Chesapeake Bay, is subdivided into the bay mouth, the estuary funnel, and the meander zone. The mouth of the river could conceivably be defined as the bay mouth, the upper end of the estuary funnel (defined on the basis of salinity, width-depth ratio, and sedimentological criteria), or the upper limit of the meander zone (upper limit of tidal influence). Additional mouths could be defined within the meander zone, based on the high-discharge tidal limit and channel width/depth ratio (see Nichols *et al.*, 1991: Figure 13).

The simplest and most intuitive definition of the river mouth is based on channel morphology, as used above, and is signified by the end of a distinct, well-defined channel or the beginning of an open water body. In many coastal plain rivers, however, this point may be cut well below sea level, strongly influenced by lunar and wind tides, and may be characterized by dominantly coastal sediments, hydrodynamics, and ecological characteristics.

Alternatively, the mouth could be defined on the basis of network rather than channel morphology, as the point at which the channel network changes from dominantly convergent to divergent (distributary). This is most obviously the case at the apex of a delta, but may in some cases occur upstream or in the absence of what would conventionally be considered a delta.

Hydrographic and hydrologic characteristics could also be used to define the river mouth. Three logical candidates would be the point at which the channel bed is cut below sea level, the upstream limit of tidal influence (including wind-driven tides) or coastal ponding, or the upstream limit of saltwater penetration (usually specified as some threshold salinity, typically 0.5 ppt). The tidal and saltwater criteria must, of course, be referenced to flow conditions.

Criteria related to sedimentology or sediment transport/deposition dynamics might also be employed to define the river mouth. The turbidity maximum or flocculation zone, which also varies with flow conditions, is commonly referenced by estuarine ecologists and sedimentologists. Changes in dominant bed sediment characteristics (for example from sand- to mud-dominated) might also be employed, or changes in the dominant sediment source as indicated by sedimentological, chemical, mineral, or alluvial pedological criteria (eg, Nichols *et al.*, 1991; Phillips, 1992a; 1992b; Cattaneo and Steel, 2003). A pronounced increase in sediment storage or decrease in transport might also be argued to mark a mouth of the river, though this may be difficult to pinpoint without detailed fieldwork.

These different definitions of the river mouth, summarized in Table 1, may vary significantly in their locations along the channel, and thus the amount of river sediment delivered to the mouth could be quite different depending on the conceptual framework adopted. Using the head of Trinity Bay as a reference point, the upstream distance of the river mouth based on various criteria for the Trinity River, Texas, is shown in Table 2. Estimated sediment loads at each point

Table 1 Potential criteria for defining the mouth of a river draining to the coast

Criterion	Mouth
1. Channel morphology	End of well-defined channel; transition to open water
2. Network morphology	Transition from dominantly convergent to divergent network
3. Channel bed elevation	Thalweg cut below sea level
4. Tidal influence	Upstream limit of tidal influence or coastal back-water effects at reference flow condition(s)
5. Salinity	Upstream limit of threshold salinity at reference flow condition(s)
6. Turbidity	Turbidity maximum or flocculation zone at reference flow condition(s)
7. Sedimentology	Transition in source or nature of sediments (for example, sand to mud bed)
8. Sediment transport/storage	Transition in sediment transport and storage dynamics (transport bottleneck)

Table 2 Location of the mouth of the Trinity River, based on various criteria (see Table 1). Based on maps and field observations. Estimated sediment input based on data in Phillips *et al.* (2004)

Criterion	Estimated distance upstream from Trinity Bay (km)	Estimated mean annual river sediment input (tonnes × 1000)[1]
1. Channel morphology	0	$\ll$70
2. Network morphology	19.5	$<$70
3. Channel bed elevation	110	75 to 100
4. Tidal influence	85 to 110	65 to 75
5. Salinity	7 to 85	$<$70
6. Turbidity	2 to 10	$\ll$70
7. Sedimentology	20 to 30	$<$70
8. Sediment transport/storage	130	3400

[1]Mean annual sediment load at Liberty = 69,673 t yr^{-1}; mean annual sediment load at Romayor = 3,378,461 t yr^{-1}.

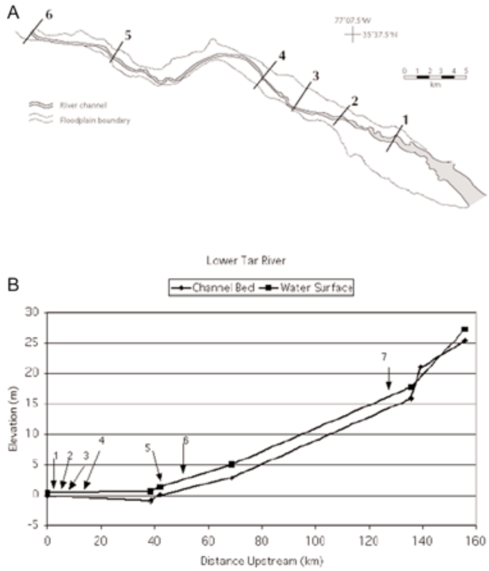

Figure 1 (Above) The lower Tar River and upper Pamlico River Estuary, between Greenville and Washington, North Carolina. The estimated location of the river mouth based on six different criteria is shown by the numbered lines. These are: (1) channel morphology (pronounced widening at the mouth of Tranter's Creek); (2) network morphology (beginning of distributary subchannels in the flooplain); (3) salinity (Grimesland gaging station, where there is a 50% annual probability of salt wedge penetration); (4) sediments (transition from dominantly upland/mineral to dominantly autocthonous organic alluvial soils); (5) gaging station at Greenville; approximate upstream limit of regular tidal influence; (6) approximate point at which channel bed is cut below sea level. The longitudal profile of both the channel bed and water surface is also shown (below), with the six locations above indicated above; plus (7) the point just downstream of the Tar River Falls where alluvial sediment storage increases dramatically

are inferred from Phillips *et al*. (2004). Table 2 suggests that using different definitions of mouth on the Trinity could result in assessments of sediment export from the basin varying by up to two orders of magnitude. Table 2 indicates that a mouth defined on the basis of sediment flux and storage may be of the most fundamental importance, but could be a somewhat self-fulfilling outcome given that this 'mouth' was defined on the basis of an identified transition point in sediment transport and storage dynamics. However, this transition zone coincides with a pronounced change in sinuosity that we have interpreted as representing the upstream limit of effects of Holocene sea-level rise (Phillips *et al.*, 2005). Possible locations of the mouth based on some of the criteria are shown for the Tar-Pamlico River, North Carolina, in Figure 1.

However defined, the river mouth is a dynamic, geologically ephemeral feature. On a low-gradient coastal plain and shelf, relatively small changes in sea level may result in large longitudinal shifts in the river mouth, resulting in a spatial sequence of geomorphic, hydrologic, and sedimentary environments that (depending on preservation) may be recorded in vertical sedimentary sequences across a broad area of contemporary coastal plains and shelves (Nichols *et al.*, 1991; Anderson *et al.*, 1992; Dalrymple *et al.*, 1992; Congxian, 1993; Nichol *et al.*, 1994; Thomas and Anderson, 1994; Cattaneo and Steel, 2003).

This dynamism and ephemerality point to the perhaps obvious but important point that generalizations about sediment fluxes in coastal plain rivers are contingent on a number of factors, in particular sea level and the rate and direction of sea-level change.

IV Case study: southeast Texas

The rivers of the southeast Texas coast (Figure 2) in general, and the Trinity River (already used as an illustration above) in particular, are used to illustrate some the points raised above. The characteristics of the 46,100 km^2

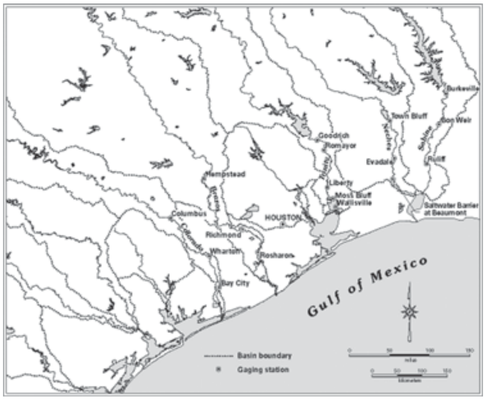

Figure 2 Rivers of the southeast Texas coastal plain, and gaging stations referred to in the text

Trinity drainage basin are described elsewhere (Phillips *et al.*, 2004; 2005).

1 Stream power in the lower Trinity River

a Methods: Stream power in the lower Trinity River, in the coastal plain downstream of Lake Livingston, was examined based on streamflow gaging data collected by the US Geological Survey at three stations – Liberty, Romayor, and Goodrich, about 83, 126, and 144 km, respectively, upstream from Trinity Bay. Mean daily flows from station inception through the 2003 water year were examined, comprising 63, 79, and 38 years of record, respectively.

Probabilities and return periods (recurrence interval, RI) were calculated for the entire record using the relation $p = 1/RI = m/(n + 1)$ where m is the rank of the event and n is the number of days in the record, and flows associated with 50, 10, and 1% probabilities were determined. These represent mean daily instantaneous discharges having a 50, 10, or 1% chance of being exceeded on a given day. Flows associated with recurrence intervals of 1, 2, and 10 years were also determined. Finally, peak daily flows were determined for each station for the October 1994 flood of record, and a moderate November 2002 flood.

Cross-sectional stream power can be computed directly from discharge, but unit stream power requires velocity, which is not routinely measured. Thus surface water measurements (generally 6–12 field measurements per year over the period of record) were used to develop velocity-discharge relationships. These data were experimentally fitted with linear, power, exponential, logarithmic, and second- and third-order polynomial functions to obtain the best fit and highest predictive power as determined by the R^2 value. For Liberty, a bimodal relationship made curve fitting the entire data set difficult (Figure 3). A linear trend was fitted by hand to the high-flow limb of the curve most relevant to the reference discharges.

From the velocity-discharge relationships, the velocity associated with each reference event was calculated. Cross-sectional and unit stream powers were calculated using thalweg slopes based on field surveys of five bridge cross-sections. These are 0.0002834 for Goodrich, 0.0002508 for Romayor, and 0.00001002 for Liberty.

b Results: The reference discharges generally decrease downstream from Goodrich to Romayor, and increase from there to Liberty (Table 3). The best-fit equations for velocity were as follows:

Goodrich: $V = 2 \times 10^{-10} Q^3 - 7 \times 10^{-7} Q^2 + 0.0014\,Q + 0.3428 \quad R^2 = 0.87$

Romayor: $V = -2 \times 10^{-7} Q^2 + 0.0011\,Q + 0.2601 \quad R^2 = 0.83$

Liberty: $V = 0.35 + 0.00013352\,Q$

The velocities for a given reference event (Table 3) are similar or identical at the two upper stations, and are substantially lower at Liberty. Because channel thalweg slopes were used in computation, variation in stream power at a station is entirely proportional to variations in discharge and velocity. Stream power is slightly lower at Romayor than Goodrich, but there are major reductions at Liberty. Increases

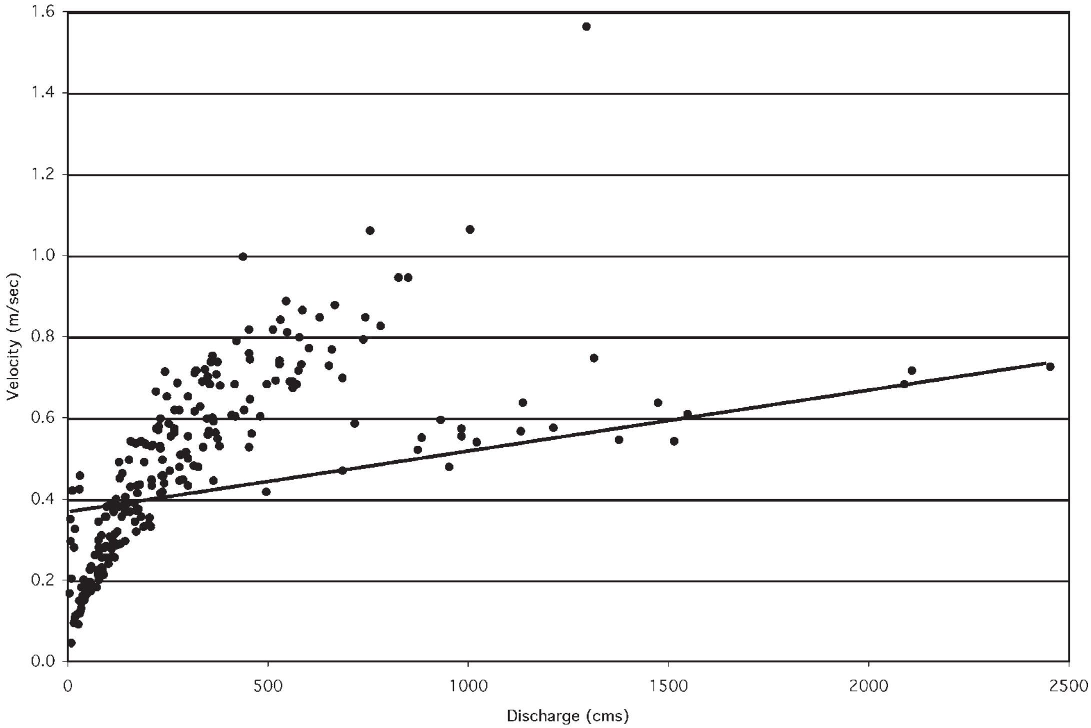

Figure 3 Velocity-discharge relationships for the gaging station on the Trinity River at Liberty, Texas. Velocities for the reference discharges were estimated based on the equation for the trend line shown

in discharge downstream are overwhelmed by the much-reduced slope, so that cross-sectional stream power is an order of magnitude lower at Liberty than at Romayor. The difference is even more pronounced with respect to unit stream power, where Liberty is about two orders of magnitude lower. The downstream/upstream ratios of stream power, discharge, and slope are shown in Table 4, which illustrates the strong influence of slope.

The sediment transport bottleneck in the lower Trinity River can be attributed largely to the fact that much of the lower river is cut to below sea level, to the very low slopes, and to the correspondingly low stream power and transport capacity. The downstream trend for normal flows (50% probability) and the 1994 flood peak are shown in Figure 4. This points to the critical role of slope, which is addressed further in the next section.

2 Slope

The extensive storage and low transport in the lower Trinity River is largely attributable to very low slopes in the lower river. But this conclusion is based partly on an assumption that energy grade slopes reflect the channel thalweg slopes. It also raises the question of whether similar trends in slope are apparent in other coastal plain rivers.

To investigate slope trends, water surface slopes were calculated from water surface elevations for coastal plain gaging stations on the Sabine, Neches, Trinity, Brazos, and Colorado Rivers, Texas. An arbitrary date and time was chosen by selecting 00:00 hours on the closest date immediately preceding the analysis which met two criteria: (1) higher than average but not rare discharge; and (2) water levels not influenced by flood waves or major dam releases. This was

Table 3 Discharge (Q, $m^3 sec^{-1}$), mean velocity ($m sec^{-1}$), cross-sectional and unit stream power (CX power, Unit power; $W m^{-2}$) for six reference discharges and two flood events at three Trinity River cross-sections. Reference discharges are based on exceedence frequencies and return periods. For example, 10% Q indicates mean daily discharge with an exceedence probability of 10% and Q2 represents a flow with a mean return period of two years

	Q	V	CX power	Unit power
Goodrich				
50% Q	82	0.45	0.23	0.00012831
10% Q	677	0.88	1.88	0.00024939
1% Q	1550	1.48	4.30	0.00041943
2002 flood	1872	1.82	5.20	0.00051652
Q1	2130	1.65	5.92	0.00046761
Q2	2400	1.74	6.67	0.00049312
Q10	3002	1.77	8.34	0.00050162
1994 flood	3540	1.67	9.83	0.00047328
Romayor				
50% Q	77	0.34	0.19	0.00008527
10% Q	640	0.88	1.57	0.00022070
1% Q	1541	1.48	3.79	0.00037118
2002 flood	2198	1.71	5.40	0.00042887
Q1	1970	1.65	4.84	0.00041382
Q2	2330	1.74	5.73	0.00043639
Q10	2925	1.77	7.19	0.00044392
1994 flood	3455	1.67	8.49	0.00041884
Liberty				
50% Q	433	0.41	0.04	0.00000411
10% Q	1048	0.49	0.10	0.00000491
1% Q	1822	0.59	0.18	0.00000591
2002 flood	1602	0.56	0.16	0.00000561
Q1	2484	0.68	0.24	0.00000681
Q2	2835	0.73	0.28	0.00000731
Q10	3600	0.83	0.35	0.00000832
1994 flood	3823	0.86	0.38	0.00000862

12 July 2004 (9 July for the Sabine, where data for 12 July were not available at all stations). Gage height recorded by the US Geological Survey at this time was added to the gage datum to determine water surface elevation. Distance between gaging stations was measured from Texas Department of Transportation maps at a scale of 1:166,667. Water surface elevation at the river mouth, defined by channel morphology, was assumed to be zero.

Water surface profiles are shown in Figure 5, and Table 5 shows water surface slopes for the lowermost reach (mouth to the next station upstream) and mean slopes for the entire study sections (uppermost coastal plain station to mouth). Figure 5 and Table 5 show that slopes in the lowermost reaches are very low. In combination with tidal and other backwater effects, this suggests low stream power and a high probability of sediment deposition upstream of the open-water

Table 4 Ratios of cross-sectional and unit stream power, discharge, and channel slope between successive Trinity River stations for various reference discharges, with the value at the downstream station divided by that at the next upstream station (R/G = Romayor/Goodrich; L/R = Liberty/Romayor)

	CX power		Unit power		Discharge		Slope	
	R/G	L/R	R/G	L/R	R/G	L/R	R/G	L/R
50% Q	0.83	0.22	0.66	0.05	0.94	5.63	0.88	0.04
10% Q	0.84	0.07	0.88	0.02	0.95	1.64	0.88	0.04
2002 flood	0.88	0.05	0.88	0.02	0.99	1.18	0.88	0.04
1% Q	1.04	0.03	0.83	0.01	1.17	0.73	0.88	0.04
Q1	0.82	0.05	0.88	0.02	0.92	1.26	0.88	0.04
Q2	0.86	0.05	0.88	0.02	0.97	1.22	0.88	0.04
Q10	0.86	0.05	0.88	0.02	0.97	1.23	0.88	0.04
1994 flood	0.86	0.04	0.88	0.02	0.98	1.11	0.88	0.04

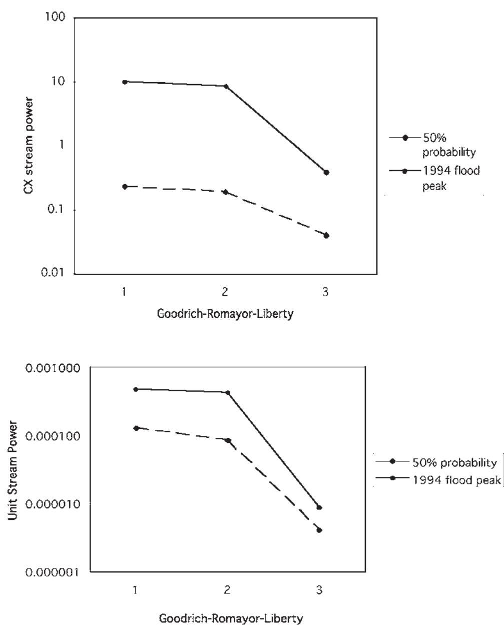

Figure 4 Downstream trends in cross-sectional and unit stream power for normal flows (50% probability) and the flood of record in 1994 (note logarithmic axis)

estuary. Concave river longitudinal profiles are common, and these tend to have low slopes in the lower reaches. However, Figure 5 shows that, within those low-slope reaches, very low slopes may be encountered in the lower reaches of coastal plain rivers.

The mean coastal plain slopes of the Brazos and Colorado Rivers are substantially higher than those of the Sabine, Neches, and Trinity, which also have substantially smaller drainage areas. The Colorado and Brazos have built substantial recent deltas, and contribute sediment to the littoral zone, while the Sabine, Neches, and Trinity do not (Morton *et al.*, 2004). The Brazos and Colorado have essentially filled the ravinement valleys cut during lower sea-level stands. This filling is generally attributed to higher sediment loads (Anderson *et al.*, 1992; Thomas and Anderson, 1994; Blum and Price, 1998). As there is no particular reason to believe that the Brazos and Colorado experienced greater denudation relative to their size during the Quaternary, however, it can be speculated that steeper regional slope in the coastal plain portion of the river has allowed the Brazos and Colorado to deliver a higher proportion of their loads to the Gulf of Mexico.

The Trinity River has the steepest slope in the reach just above the mouth, with much-reduced slope upstream, which may explain the sediment storage patterns identified by Phillips *et al.* (2004), which show an alluvial bottleneck between Romayor and Liberty.

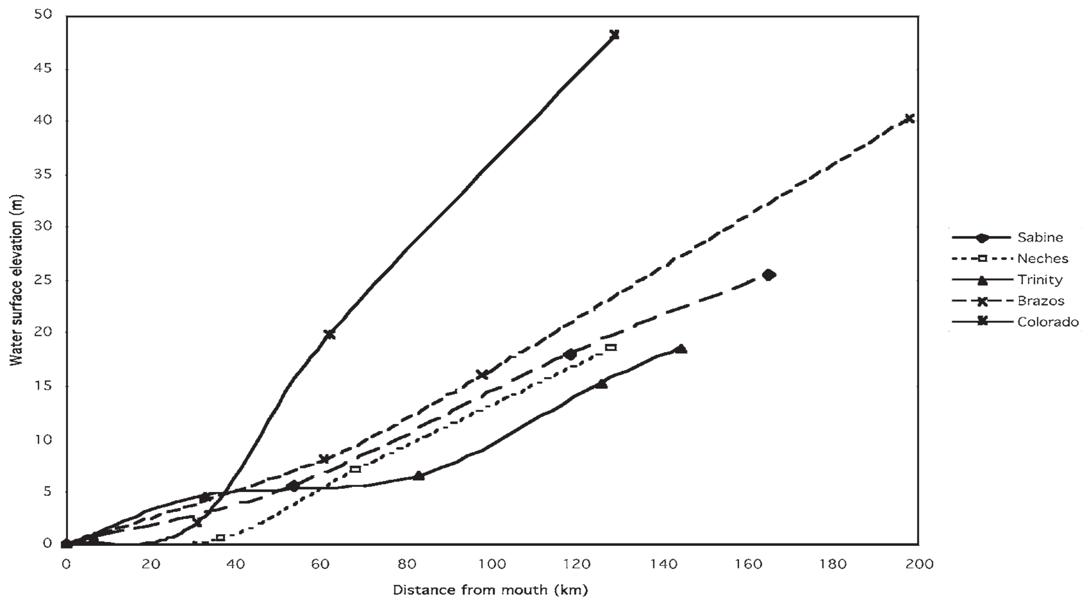

Figure 5 Water surface profiles for the lower reaches of rivers of southeast Texas, for flow conditions at 00:00 hours 12 July 2004 (00:00 9 July for the Sabine)

Table 5 Water surface slopes of east Texas coastal plain rivers for the lowermost reach, and mean slope for the entire coastal plain reach, for flow conditions at midnight, 12 July (9 July for the Sabine) 2004

River	Slope of lowermost reach ($\times 10^{-4}$)	Mean slope ($\times 10^{-4}$)
Sabine	1.026	1.544
Neches	0.167	1.451
Trinity	1.346	1.289
Brazos	1.336	2.039
Colorado	0.678	3.746

An important implication is that sediment transport monitoring which does not represent the low slope, low stream power, lower reaches of coastal plain rivers will result in overestimation of sediment delivery to the coast.

V Discussion and conclusions

Coastal and marine sedimentary archives may not reflect changes in continental erosion and fluvial sediment dynamics, because coastal plain rivers – at least during rising or high stillstands of sea level – may not transport much of their sediment load to estuaries or offshore basins. Sediment budgets of coastal plain rivers indicate extensive alluvial sediment storage upstream of estuaries, and low sediment delivery ratios. Estuarine sediment provenance studies confirm this, by showing dominantly coastal and marine sediment sources even in drowned river valley estuaries with substantial fluvial inflows. The lower coastal plain reaches of alluvial river valleys are in some cases effective sediment bottlenecks which buffer coastal systems from effects of upstream changes in sediment production and transport.

Lower coastal plain alluvial valleys may be drowned by rising sea level and ravined at lower sea levels. At current sea levels, the locus of fluvial deposition is not necessarily the ocean, estuary, or delta, but floodplains in and upstream of the fluvial-estuarine transition zone. This raises the question of what constitutes delivery to sediment to the outlet of the basin, and how to define the mouth of a river. There are at least eight different morphological, hydrographic, hydrologic, or sedimentological criteria that might be used to define the river mouth; in general these are significantly upstream of open-water estuaries or delta apices. Sediment delivery to these 'upstream mouths' may be a more accurate reflection of river sediment fluxes to the coastal and marine environment. Different definitions of the mouth could result in assessments of fluvial sediment export which differ by orders of magnitude.

Current assessments of river sediment supply to the coast (as opposed to the lower coastal plain) are too high. Sediment delivery to the coast is generally estimated from measurements at monitoring stations that are upstream of the storage bottlenecks. These sites do not represent the low slope, low stream power, lower coastal plain river reaches. Changes in sediment loads at these stations may have negligible impacts on coastal sediment systems.

Alluvial (as well as estuarine and deltaic) sediment storage during high or rising base levels is generally attributed to large accommodation space relative to sediment supply (Blum and Törnqvist, 2000; Cattaneo and Steel, 2003). This is indeed important, as lower elevations and wider valleys typical of coastal plain reaches provides ample opportunity for deposition. Overbank flooding is often more common in lower reaches, providing more opportunity for floodplain deposition. The role of fluvial sediment transport capacity, however, has been underappreciated. Slope-controlled reductions in transport capacity are widely recognized where rivers emerge onto coastal plains. In some cases slope-controlled declines in stream power occur in the lower coastal plain reaches – in the Trinity River example, stream power decreases by an order of magnitude or more in this zone. This reduction in transport capacity becomes even more pronounced as tidal and coastal backwater effects increase downstream.

The relative lack of attention to slope and transport capacity may be due in part to the assumption that the chief response of rivers to base level change is cut-and-fill. In the case of rising sea level, for instance, channels cut to below sea level might be expected to quickly aggrade. This is clearly not the case for the Trinity and Tar Rivers discussed here, where channel beds are below sea level well upstream of their estuaries – other coastal plain rivers are likely similar in this regard. Cutting and filling is only one way of adjusting slope, and several field and experimental studies in coastal plain rivers indicate that changes in channel planform (meandering) may be the primary response (Alford and Holmes, 1985; Autin, 1992; Schumm, 1992; Koss *et al.*, 1994; Leigh and Feeney, 1995). Thus Holocene sea-level rise could induce an increase in sinuosity, which would have the effect of reducing slope independently of any change in bed elevation. The depositional locus might then occur in the zone of increased meandering, within a low-transport-capacity reach where the bed elevation approaches and eventually falls below sea level. This appears to be the case in the Trinity River, where the transition zone between a scour-dominated reach downstream of Livingston Dam and the lower coastal plain alluvial storage bottleneck coincides with a pronounced increase in sinuosity (Phillips *et al.*, 2004; 2005). This suggests that such high-meandering lower coastal plain river reaches may be the most fruitful places to search for sedimentary archives of upstream changes in fluvial sediment delivery.

While the phenomenon of very low sediment delivery to the coast in many coastal plain rivers is a clear conclusion of this study, there is obviously much to do. The paucity of sediment monitoring in the lower coastal plain is a clear shortcoming, and not an easy one to redress. Measurements in coastal plain rivers demand dealing with large channels and large drainage basins; difficult practical as well as conceptual tasks. Most studies of sediment sources, fates, and provenance are focused either on drainage basins upstream of the lowermost gaging station or on estuaries. There is, therefore, a major gap in between, and we need to learn more about sediment sources, transport, and storage in lower river reaches and fluvial-estuarine transition zones. Such studies, particularly when conducted in the context of environmental change, should be cognizant of the likelihood that factors such as sea level, climate, tectonics, human impacts, and other factors are simultaneously and synergistically driving the system. Additionally, the boundary conditions and degrees of freedom for both fluvial and coastal response are conditioned by inherited geological factors.

Small drainage basins are, other things being equal, easier to work with. As Dearing and Jones (2003) point out, small basins are generally more responsive to environmental change. In this sense, any search for sensitive, manageable sedimentary archives of environmental change should avoid large alluvial rivers crossing coastal plains. In the context of comprehending continental and global-scale changes, however, the enormous amounts of land area, sediment, water, and other mass associated with or transported and stored by coastal plain rivers demands that we take them on.

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