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Geomorphic and sedimentary response of rivers to tectonic deformation: a brief review and critique of a tool for recognizing subtle epeirogenic deformation in modern and ancient settings

John Holbrook [a,*], S.A. Schumm [b]

[a] *Southeast Missouri State University, Department of Geosciences, 1 University Plaza 6500, Cape Girardeau, MO 63701, USA*
[b] *Colorado State University, Department of Earth Resources, Fort Collins, CO 80521, USA*

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Abstract

Rivers are extremely sensitive to subtle changes in their grade caused by tectonic tilting. As such, recognition of tectonic tilting effects on rivers, and their resultant sediments, can be a useful tool for identifying the often cryptic warping associated with incipient and smaller-scale epeirogenic deformation in both modern and ancient settings. Tectonic warping may result in either longitudinal (parallel to floodplain orientation) or lateral (normal to floodplain orientation) tilting of alluvial river profiles. Alluvial rivers may respond to deformation of longitudinal profile by: (1) deflection around zones of uplift and into zones of subsidence, (2) aggradation in backtilted and degradation in foretilted reaches, (3) compensation of slope alteration by shifts in channel pattern, (4) increase in frequency of overbank flooding for foretilted and decrease for backtilted reaches, and (5) increased bedload grain size in foretilted reaches and decreased bedload grain size in backtilted reaches. Lateral tilting causes down-tilt avulsion of streams where tilt rates are high, and steady down-tilt migration (combing) where tilt rates are lower. Each of the above effects may have profound impacts on lithofacies geometry and distribution that may potentially be preserved in the rock record. Fluvial sedimentary evidence for past tilting is traditionally based on the assumption that depositional features reminiscent of modern fluvial tectonic effects are evidence for past tectonic effects where it is closely associated with historically active structures, or where non-tectonic causes cannot be invoked; however, caution must be exercised when using these effects as criteria for past or current tectonic warping, as these effects may be caused by non-tectonic factors. These non-tectonic causes must be eliminated before tectonic interpretations are made.

Keywords: rivers; tectonics; epeirogeny; geomorphology; sedimentology; neotectonics

1. Introduction

The gentle and broad warpings of crustal interiors indicative of epeirogeny may have many causes (salt tectonics, Collins et al., 1981; transpression/transtension, Russ, 1982; intraplate stress, Holbrook, 1996; loading, Jordan, 1981; etc.), but the end result in most cases will be the same, tilting of the contemporary topographic surface. Rivers are particularly sensitive to such tilting because of the gradient changes imposed. This is especially true of low-gradient rivers, the type characteristic of continental in-

* Corresponding author. E-mail: jholbrook@semovm.semo.edu

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teriors. In fact, Melton (1959) estimates that 25–75% of all streams in non-glaciated areas are tectonically influenced or controlled.

Large and mature epeirogenic features, such as the U.S. Colorado and Tibetan plateaus, may represent easily identified warpings of continental interiors. Smaller and/or incipient epeirogenic warpings are by their very nature subtle, can often be difficult to identify in modern settings, and are a true challenge to recognize in the ancient. The sensitivity of rivers and streams to epeirogenic tilting can thus prove to be an asset when seeking these gentle warpings in both modern and ancient settings. This paper offers a brief synopsis and evaluation of the effects of subtle tilting on modern fluvial systems, and the means and likelihood for preservation of these effects in the sedimentary record. Much of the seminal work on this subject has addressed subtle warpings that lack the regional extent necessary to meet the purest definition of epeirogeny (e.g., Russ, 1982). These studies of more localized warping are discussed here in addition to those of more regional extent because they are a proxy for incipient and smaller-scale epeirogeny, they offer scale models for larger-scale epeirogeny, and they are the main locations where study has been sufficiently detailed to gain a meaningful understanding of the more general effects of tilting on rivers.

The information here is meant to serve as an initial guide for workers attempting to identify both modern and ancient epeirogenic deformation from the evidence provided by rivers and/or fluvial deposits. The concentration here is on individual alluvial rivers, which includes those rivers that flow through their own sediments and are not bedrock confined. Considerable tectonic information, however, can also be gained by examining drainage patterns for entire river/tributary networks (Howard, 1967; Muehlberger, 1979; Cox, 1994; Keller and Pinter, 1996). More comprehensive and detailed discussion of tectonic effects on rivers is presented in Schumm et al. (in press).

Epeirogenic tilting may be considered in end-member terms of either longitudinal (parallel to floodplain orientation) or lateral (normal to floodplain orientation) tilting. Both conditions are addressed in this paper for both modern and ancient settings. Sedimentary evidence for past tilting traditionally is based on assuming that ancient deposits bearing features similar to those produced by modern fluvial tectonic effects are evidence for past tectonic effects where closely associated with historically active structures, or where non-tectonic causes cannot be easily invoked. We adhere to this assumption.

2. River response to longitudinal tilting

Where rivers encounter zones of active subsidence or uplift, their normal longitudinal profile will be deformed. The river will either traverse, or will be deflected by, the deformed zone. The following sections detail the case of river deflection, followed by four possible responses to traversing of a deformed zone. As a general caution, streams may also yield these responses owing to changes in grade caused by locally increased sedimentation or erosion (e.g., the Jordan River; Schumm, 1977). Stratigraphic examination at suspected tectonic anomalies, however, should reveal if there is abnormal accumulation or removal of sedimentary units beneath the anomaly, and these observations should be made before a tectonic explanation is accepted for either ancient or modern settings.

2.1. Deflection

Geomorphologists observed long ago that rivers will tend to be deflected by surficial warping (Goodrich, 1898; Zernitz, 1932; Howard, 1967). This is mostly because this is one of the most readily observable of all the possible effects of deformation on rivers. Deflection of the river around an uplift or into a zone of subsidence will be manifest as an abrupt shift in the river course coincident with the deformed zone. Streams naturally tend to gravitate toward the subsided zone, if the zone is proximal to the river and there are no topographic barriers between the river and zone. In turn, a river will tend to cross a zone of uplift if the rate of stream incision is substantially greater than the uplift rate, or if the orientation of the uplift is such that the uplift is not easily avoided. This is especially true if the river course is well established at the site of uplift before deformation begins (i.e., a superposed river).

Low-gradient rivers are especially sensitive to epeirogenic movements, such that even very subtle deformation can alter the course of major rivers. For instance, Miranda and Boa Hora (1986) describe several Tertiary uplifts in the upper Amazon Basin that generate only a few tens of meters of relief over several kilometers of distance, yet these uplifts are sufficient to alter the course of the Amazon River and its tributaries.

Such diversions will appear in the rock record as a preferential concentration of channel belts in down-warped areas (Bridge and Leeder, 1979; Kvale and Vondra, 1993; Hazel, 1994; Autin et al., 1995; May et al., 1995), and/or deflection of paleocurrents toward tectonic lows (Schwartz, 1982; DeCelles, 1986; Greb and Chesnut, 1996). For example, Pliocene–Recent fluvial deposits of the Sorbas Basin of southeast Spain have paleocurrent trends that show deflection around sites of modern uplift and into flanking structural lows (Mather, 1993). Syndepositional elevation of these uplifts is further evidenced by the presence of intraformational angular unconformities within these deposits above the structural highs (Mather, 1993). The Lower Cretaceous Antlers Formation of central Texas reveals a preferential increase in sand thickness (as much as 2×) in pre-Cretaceous structural lows (i.e., the Kingston and Sherman synclines). This suggests Cretaceous reactivation and sagging of these features, causing preferential diversion of sand-depositing river channels into these coevally deforming tectonic lows (Hobday et al., 1981). Tectonic lows also may be characterized by localized valley incision as indicated by lows in scoured unconformities, and/or anomalously thick valley-fill strata (Weimer, 1984; Holbrook, 1992) (Fig. 1).

Where rivers are in close proximity to bedrock, deflections may follow fracture patterns without involvement of significant warping (Howard, 1967; Droste and Keller, 1989). Likewise, rivers may be deflected around locally resistant materials (Fisk, 1944). Sharp deflections do not necessarily reflect tectonic deformation where bedrock control is a factor, or highly resistant materials are present in the alluvium. Likewise, rivers will avulse in the absence of tectonic stimuli, owing to climatically or eustatically driven aggradation of the channel belt (Allen, 1978; Blakey and Gubitosa, 1984; Shanley and McCabe, 1991). Channel deflections must closely co-

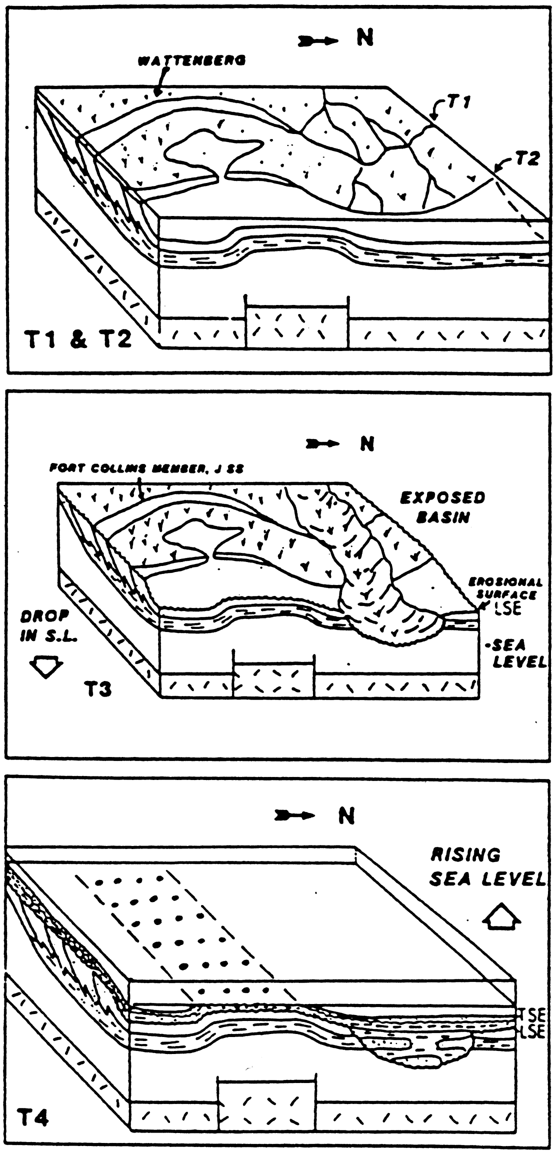
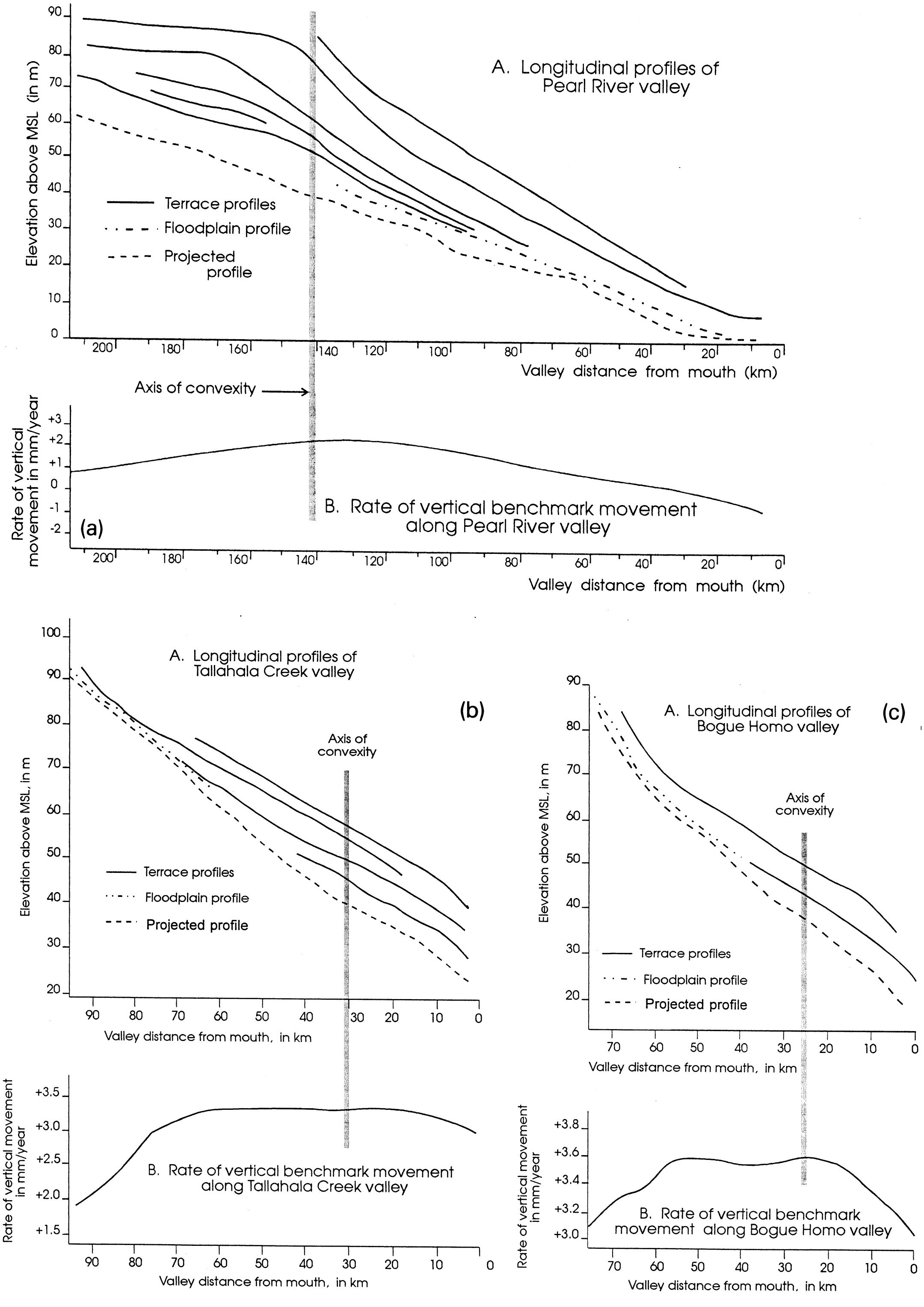


Fig. 1. Development of tectonically controlled valley fill. During time *T1* and *T2* (sea level highstand), the area above the down-dropped tectonic block accommodates deltaic deposition, while the uplifted block is draped by lagoonal deposits. During time *T3* (sea level drop), valley incision is preferentially concentrated in the down-dropped block, and the lagoonal deposits on the up-lifted block experience erosion and/or soil development. During time *T4* (sea level rise), the valley is aggraded, and both valley and lagoonal deposits are transgressed (Weimer, 1984).


A. Longitudinal profiles of
Pearl River valley
Elevation above MSL (in m)
Terrace profiles
Floodplain profile
Projected
profile
Valley distance from mouth (km)
Axis of convexity
Rate of vertical
movement in mm/year
B. Rate of vertical benchmark movement
along Pearl River valley
(a)
Valley distance from mouth (km)
A. Longitudinal profiles of
Tallahala Creek valley
(b)
Axis of
convexity
Elevation above MSL, in m
Terrace profiles
Floodplain profile
Projected profile
Valley distance from mouth, in km
Rate of vertical movement
in mm/year
B. Rate of vertical benchmark movement
along Tallahala Creek valley
Valley distance from mouth, in km
A. Longitudinal profiles of
Bogue Homo valley
(c)
Axis of
convexity
Elevation above MSL, in m
Terrace profiles
Floodplain profile
Projected profile
Valley distance from mouth, in km
Rate of vertical movement
in mm/year
B. Rate of vertical benchmark
movement along Bogue Homo valley
Valley distance from mouth, in km

incide with structures, before such deflections are considered tectonic instead of coincidental.

2.2. Longitudinal profile adjustment

As deformation affecting the ground surface also deforms the topography beneath the river course, the most fundamental effect crossing a site of deformation is warping of the channel profile relative to the regional average valley gradient. For example, up-warping may cause convexity of terraces, valley floor, water surface, and/or stream thalweg, mimicking the shape and location of bedrock structure (Fig. 2). In turn, local subsidence may result in concavity of these same features. Burnett (1982), Burnett and Schumm (1983), Ouchi (1985), Jorgensen (1990), Marple and Talwani (1993), Fischer (1994), and Schumm et al. (1994) all describe examples of both stream and valley profile warping.

An illustrative example of river profile deformation is described by Burnett (1982) and Burnett and Schumm (1983) over the Wiggins uplift of southern Mississippi. The Wiggins uplift spans the width of the state of Mississippi, and currently experiences as much as 4 mm/year uplift. Wiggins uplift is traversed by the Pearl River, Tallahala Creek, and Bogue Homo Creek southeast of Jackson, Mississippi. Each of these streams have convex stream terraces coincident with the shape and position of Wiggins uplift (Fig. 2). In each case, this reflects the cumulative deformation of abandoned stream floodplains after their formation.

Plots of channel thalweg or water-surface elevation against valley distance (projected channel profiles) reveal deformation of the current valley profile. Projected channel profiles of these streams show less dramatic convexity, relative to average regional gradient, in Tallahala and Bogue Homo creeks than their respective terrace profiles, and no convexity in the Pearl River at all (Fig. 2). Apparently, the greater discharge, and thus erosive power of the Pearl River has enabled quick response to uplift, allowing it to reestablish grade in the face of profile deformation. Rapid down-cutting is revealed by the lack of a well developed floodplain and the current 12 m of incision below the lowest terrace near the uplift axis. Tallahala and Bogue Homo creeks lack comparable stream power, and thus are convex (Burnett, 1982; Burnett and Schumm, 1983). This illustrates that projected channel profile is only a lasting and sensitive indicator of channel profile deformation for rivers with low stream power. Any such concavity will be subject to rapid beveling for larger rivers, and will thus be more temporary. Even for the largest rivers, however, beveling of convexities will not be immediate, as the Mississippi River still bears convexity over the Lake County uplift that is apparently related to deformation during the 1811–1812 New Madrid earthquakes (Russ, 1982).

Fig. 2. Longitudinal profiles of terraces, the active floodplain, and the modern channel for the Pearl River (A), Tallahala Creek (B), and Bogue Homo Creek (C) of southern Mississippi where it crosses the Wiggins uplift (Burnett, 1982).

One means of restoring a projected channel profile to a consistent grade after profile deformation is by aggradation or degradation (Maizels, 1979; Burnett, 1982; Ouchi, 1983, 1985; Jorgensen, 1990; Marple and Talwani, 1993). In the idealized case, a river restores a steady grade by aggradation upstream and downstream of an uplift (Fig. 3A). This is currently observable where the Rio Grande River crosses a broad dome over the Socorro magma body north of Socorro, New Mexico. Here, the Rio Grande River is undergoing aggradation in the region of lowered slope upstream from the domal axis. It is incising the area of steepened gradients over the dome, and it is aggrading where slopes become lower again

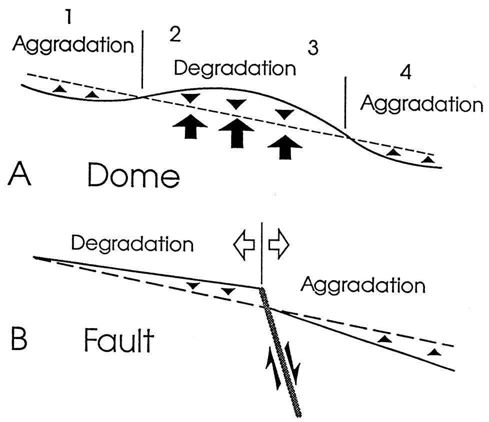


Fig. 3. Generalized aggradational and degradational response of a stream as it crosses a dome (A) and a fault (B).

downstream and adjacent to the dome (Ouchi, 1985). Similarly, rivers will tend to incise in regions of steepened slope entering a subsided zone, and aggrade in the low-gradient reaches over the axis of subsidence (Fig. 3B) (e.g., Rose Creek Narrows of Humbolt Creek, Nevada; Jorgensen, 1990).

Loss of stream power and aggradation is an anticipated response for streams encountering lowered slopes on the approach to the axis of an impeding uplift. Because of the loss of load accompanying aggradation, and the increased slopes, such streams can be expected to have increased erosive power and degrade as they cross the uplift axis and proceed across the downstream flank. If streams fail to reclaim sufficient load by degradation of the uplift, however, they will not aggrade downstream of uplifts and will continue to erode. Similarly, areas of subsidence will not tend to experience aggradation if the streams are supply-limited upon entering the subsided zone.

Owing to their ephemeral nature, primary preservation of stream profile anomalies over deformation is unlikely. Additionally, reconstruction of individual channel profiles from amalgamated fluvial deposits presents a challenge. Aggradation and degradation related to profile correction, however, may appear in the stratigraphic record as a product of secondary profile preservation. Degradation over zones of uplift or subsidence may be manifest in the stratigraphic record as thinning and/or erosional intraformational angular unconformities in fluvial strata (Riba, 1976; Miall, 1978; Anadón et al., 1986). Several cases for uplift have been made based on development of intraformational angular unconformities and thinning trends in fluvial strata (Schwartz, 1982; DeCelles, 1986; Meyers et al., 1992; Holbrook and Wright Dunbar, 1992; Pivnik and Johnson, 1995; Greb and Chesnut, 1996). Schwartz (1982), for example, argued that positive Early Cretaceous reactivation of the Boulder Batholith within the Western Interior foreland basin of southwestern Montana influenced deposition of the fluvial Kootenai and Blackleaf basin-fill formations. Uplift on the Boulder Batholith was apparently sufficient to generate radial paleocurrent patterns in fluvial sandstone flanking the uplift, thinning of fluvial strata above the uplift, and increase in the lithic component of flanking sandstone bodies. This is all presumably caused by erosion over the uplift where an intraformational angular unconformity was generated.

Caution must be exercised when using aggradation and degradation as indicators of syndepositional deformation of channel profiles. Namely, fluctuations in discharge as well as sediment load may cause aggradation and degradation in a stream (Love, 1960; Blum and Valastro, 1994; Bettis and Autin, 1997). Such changes should be manifest as local and abrupt alterations in channel-element size, geometry, and/or fill over the aggraded and degraded reach that are not explained by local gradient change. These possible causes must be eliminated if aggradation or degradation are to be attributed to deformation in modern or ancient conditions. Likewise, spotting such anomalies requires good age control in sediments that are inherently difficult to date. This is often compounded by the tendency of differential gravitational compaction to cause apparent local thickness anomalies within thick units.

Care must also be taken when observing channel profile adjustments in the modern settings, as causes other than tectonic warping may impose channel profile anomalies. In particular, a nickpoint may migrate up a stream in response to stream grade adjustment after base-level lowering or deformation; however, as nickpoints are mobile, they will not typically consistently coincide directly with sites of deformation. The abrupt local steepening of grade associated with a nickpoint, however, will be distinct from a concavity or convexity in most cases. Also uplift may expose resistant bedrock in the uplift core beneath otherwise alluvial rivers, increasing the amount of profile convexity by inhibiting efforts of streams to erosively reestablish grade. Examination for resistant bedrock in uplifts should therefore be made before profile convexities are attributed entirely to warping. Though a good initial indicator of deformation, such complexities mean that profile adjustment can rarely be used to make a case for uplift or subsidence without other corroborating evidence.

2.3. Channel pattern adjustment

Channel pattern may be altered directly or indirectly by increase or decrease in slope imposed by an impeding zone of uplift or subsidence (Fig. 4). Sufficiently decreased slopes may cause rivers to

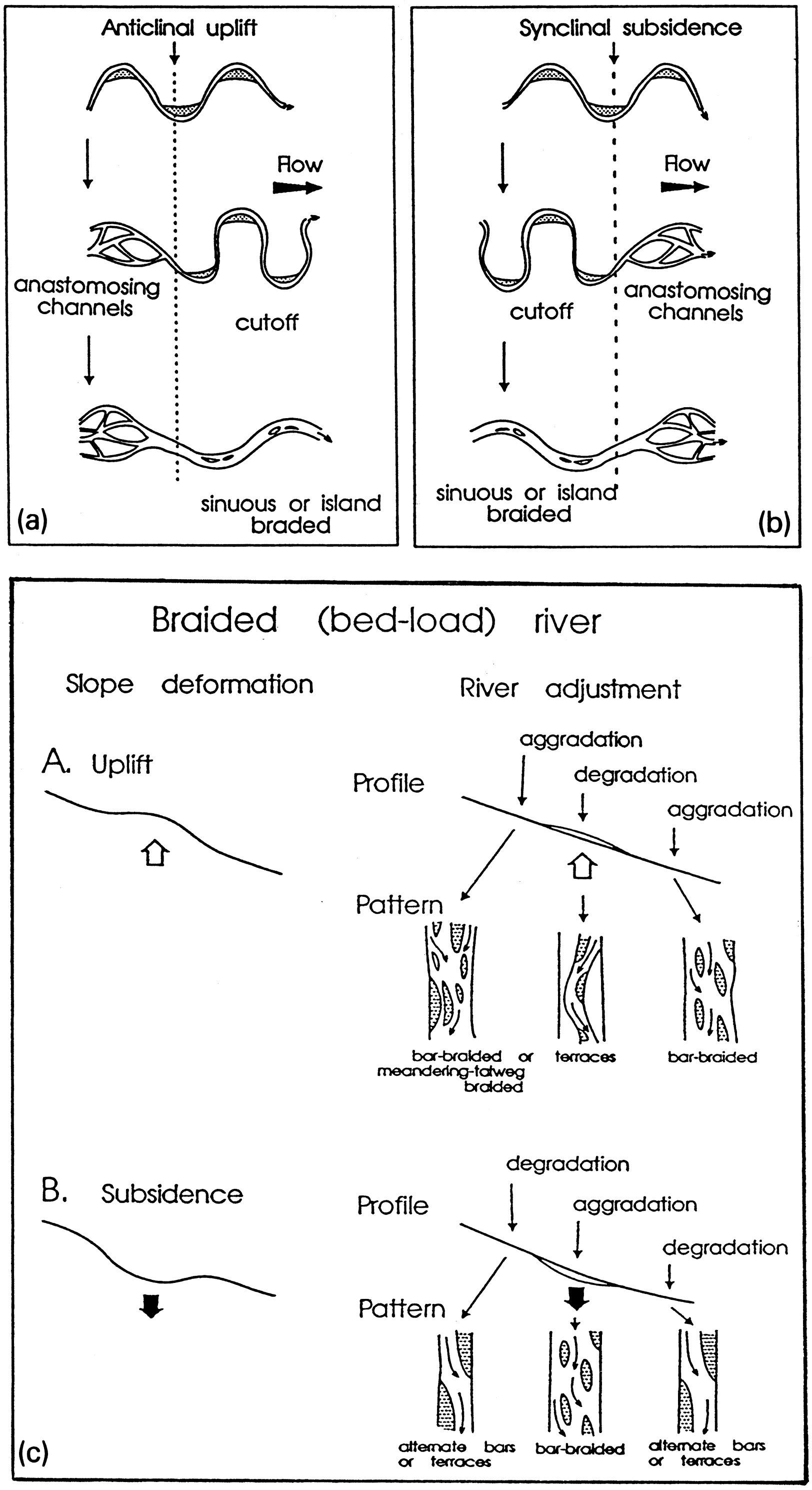


Fig. 4. Some typical pattern responses where meandering or straight streams cross a zone of uplift (A) or a zone of subsidence (B), as well as some responses of braided streams to similar conditions (C). (Ouchi, 1983).

transform along the path of braided to meandering to straight or anastomosing patterns, and vice versa (Fig. 4a,b). More typically, rivers undergo minor variations within pattern without undergoing complete shifts in pattern type (Fig. 4C). The most commonly observed of these intrapattern adjustments is for a meandering channel to increase its sinuosity in response to increased slope, or decrease its sinuosity as an adjustment to decreased slope (Welch, 1973; Adams, 1980; Vanicek and Nagy, 1980; Russ, 1982; Burnett, 1982; Ouchi, 1985; Jorgensen, 1990; Schumm and Galay, 1994; Schumm et al., 1994; Boyd and Schumm, 1995). A classic example occurs where the Mississippi River crosses the Lake County uplift, a 10 m high topographic bulge caused by active deformation in the New Madrid seismic zone of southeastern Missouri. Here, the low-gradient ($\sim$0.0001) Mississippi River reduces its sinuosity on the up-dip flank of the uplift where gradients are lowered. On the down-dip flank where gradients are increased, the sinuosity increases (Russ, 1982; Schumm et al., 1994).

Pattern alteration in braided streams is less obvious, and less studied. One notable example is from the gravel-bed Jefferson River, Montana (Jorgensen, 1990). Here, the river is more sinuous upstream and downstream of an uplift axis, because sediment chokes the system, and the increased sinuosity more effectively distributes the sediment laterally. The stream is actually straighter on the downstream flank of the uplift where erosion cleans the channel of excess sediment. This suggests that the relationship between sinuosity and slope seen in meandering streams may be more complex for gravel-bed braided streams.

Flume studies may also provide insight into potential tectonic reaction of braided streams. These studies reveal that braided streams undergoing the type of aggradation commonly observed in reduced-slope reaches show an increase in braid index (total channel length/valley length), number of braid bars, and number of lingoid dunes, whereas degradation caused decreased number of braid bars and decreased channel width (Germanoski and Schumm, 1993) (Fig. 4C).

Though intrapattern variations are more commonly cited, complete shifts in pattern type also occur. In some cases, deformation has resulted in such extreme gradient increase, that meandering streams have altered locally to braided patterns (Twidale, 1966; Burnett, 1982). Likewise, where slopes on straight or meandering streams have been lowered substantially, some rivers have adopted an anastomosing pattern, especially where aggradation rates are also high (Burnett, 1982; Ouchi, 1985; Marple and Talwani, 1993) (Fig. 4A,B).

Recognition of ancient pattern alteration in response to channel profile deformation requires detailed reconstruction of channel patterns of contemporaneous fluvial deposits above structures and their environs. Numerous lithofacies models proposed for distinction of stream pattern type (e.g., braided, meandering, straight, anastomosing) from fluvial strata are discussed at length in Miall (1996). Probably a more useful technique, where exposure permits, is architectural-element analysis (Miall, 1985, 1996). Architectural-element analysis focuses on identifying individual depositional components of fluvial deposits, and is thus more effective for recognition of individual bars, channels, levees, floodplains, etc. Ratios and characteristics of these components can be used to reconstruct not only the general river pattern, but can also yield insights into intrapattern features such as number and type of bars, sinuosity, and floodplain extent. As most modern channel perturbations over structures do not involve complete shifts in pattern type, such fine-scale reconstructions are often required.

A few examples of tectonically induced longitudinal pattern perturbation are cited from the stratigraphic record (Peterson, 1984; DeCelles, 1986; Srivastava et al., 1994; Holbrook and White, 1998). Holbrook and White (1998) recount one such example from Lower Cretaceous rocks of the Mesa Rica Sandstone in northeastern New Mexico. These strata appear to have been deposited by very low sinuosity single-channel streams on the up-stream flank of the Sierra Grande basement uplift and in areas well down paleodip from the uplift. Evidence for this is provided by dominance of sand-rich/active channel-fill elements, low width/depth ratio (10–20) of channel-fill elements, an almost complete lack of lateral-accretion elements, and lack of multi-lateral channel scours and multiple bar forms within major channel fills. In contrast, correlative Mesa Rica deposits on the downstream flank of the Paleozoic

Sierra Grande structure have approximately equal proportions of channel-fill and lateral-accretion elements, and a predominance of muddy/abandoned channel fills over sandy/active channel fills. This is evidence for higher sinuosity for streams in this reach. Increased sinuosity is very local and coincides exactly with the downstream flank of the Sierra Grande structure, inferring that the cause of this sinuosity anomaly is tectonic rather than climatic and/or sediment-supply induced. Early Cretaceous uplift on the Sierra Grande basement structure is thus argued. Apparent sinuosity variation within Mesa Rica Sandstone is closely analogous to the situation described previously in this section whereby the Mississippi River is forced to increase sinuosity in response to steepened slopes on the downstream flank of the rising Lake County uplift (Russ, 1982; Schumm et al., 1994).

Channel patterns may also alter owing to changes in flood peaks, mean discharge, and type of sediment load (Schumm, 1972, 1977; Autin et al., 1991). This means that any attribution of channel-pattern shifts to channel profile deformation requires elimination of these variables. Discharge and sediment load ratios are measurable quantities in modern streams, thus examination for longitudinal variation in these variables is in order before tectonic influence is inferred. In ancient systems, variations in these variables will be recorded indirectly and approximately by variations in channel-scour size (discharge), and average grain size of coarse and fine fractions (sediment load type).

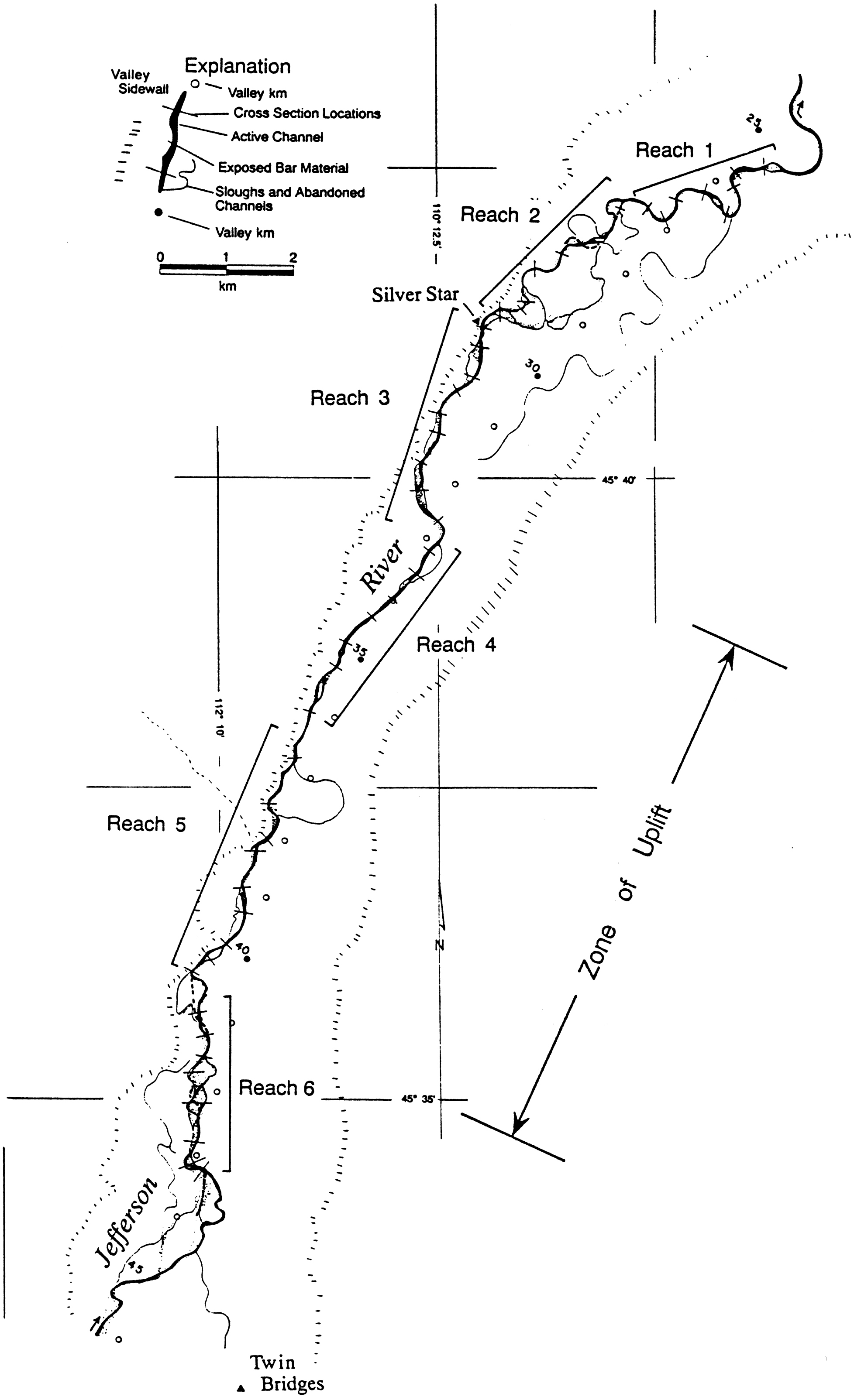
Where incision into bedrock or other resistant units (e.g., filled ox bows) impede channel migration, anomalously high sinuosities may result upstream through compression of downstream-migrating meanders (Gardner, 1975; Yeromenko and Ivanov, 1977; Jin and Schumm, 1987; Schumm et al., 1994). This results in increased sinuosity upstream from bedrock incision, and decreased sinuosity downstream. Such sinuosity perturbations are not necessarily related to profile deformation. In some cases, however, localized uplift may be the reason for appearance of bedrock in a channel (e.g., sinuosity variations of the Mississippi River as it encounters Tertiary clay uplifted by the Monroe uplift; Schumm et al., 1994). Evidence for incision of resistant materials should be examined in both modern and ancient settings where interpretations of tectonic influence on pattern are inferred.

2.4. Cross-section adjustment

A complicated collection of shifts in hydraulic variables may occur as streams cross zones of uplift or subsidence that impact stream cross-sectional shape. In general, decreased slopes result in lower bankfull discharge, decreased stream power, higher width/depth ratios, and higher flood frequency, and vice versa. Jorgensen (1990) describes such a case where the gravel-bedded Jefferson River crosses an active uplift in Montana (Fig. 5). Where the river first encounters the uplift, slopes are lowered. This causes the Jefferson River to dump much of its bedload, forming bars that the flow must pass around. Sediment storage thus widens the channel. The river encounters steeper slopes as it crosses the uplift that increase stream power and promote removal of stored sediment from the channel bottom. This means that channel width, width/depth ratio, and sediment storage decrease, and channel capacity (bankfull discharge) increases as the river passes from the reach just above to the reach just over the axis of the uplift (Fig. 5).

Though channel shape is often affected by deformation, just how that shape is expressed varies with the river. In contrast to the Jefferson River example, the Nile River widens over steepened zones (Schumm and Galay, 1994), thus increasing bed friction. Likewise, increased slopes may prompt channels to incise deeply into underlying strata (Burnett, 1982; Ouchi, 1985; Jorgensen, 1990). Overall channel shape is thus not a consistent indicator of deformation, and will tend to reflect a complex response to conditions individual to that stream.

The most consistent effect on rivers by profile deformation is reduced incision and shallower depths where lowered slopes are encountered. Reduced gradient and incision typically translates to loss of bankfull channel depth, which tends to promote more frequent floods (Ouchi, 1985; Jorgensen, 1990). In some cases, flooding results in permanent swamps (Ouchi, 1985; Marple and Talwani, 1993) or lakes (Doornkamp and Temple, 1966; Rasanen et al., 1987; Dumont, 1992, 1993), where deformation has effectively dammed river courses. Where rivers encounter increased gradient, increased stream power


Explanation
Valley Sidewall
Valley km
Cross Section Locations
Active Channel
Exposed Bar Material
Sloughs and Abandoned Channels
Valley km
0 1 2
km
Reach 1
Reach 2
Silver Star
Reach 3
River
Reach 4
Reach 5
Reach 6
Jefferson
Zone of Uplift
Twin Bridges
45° 40'
45° 35'
112° 10'
110° 12.5'
N
25
30
35
40
45

(a)

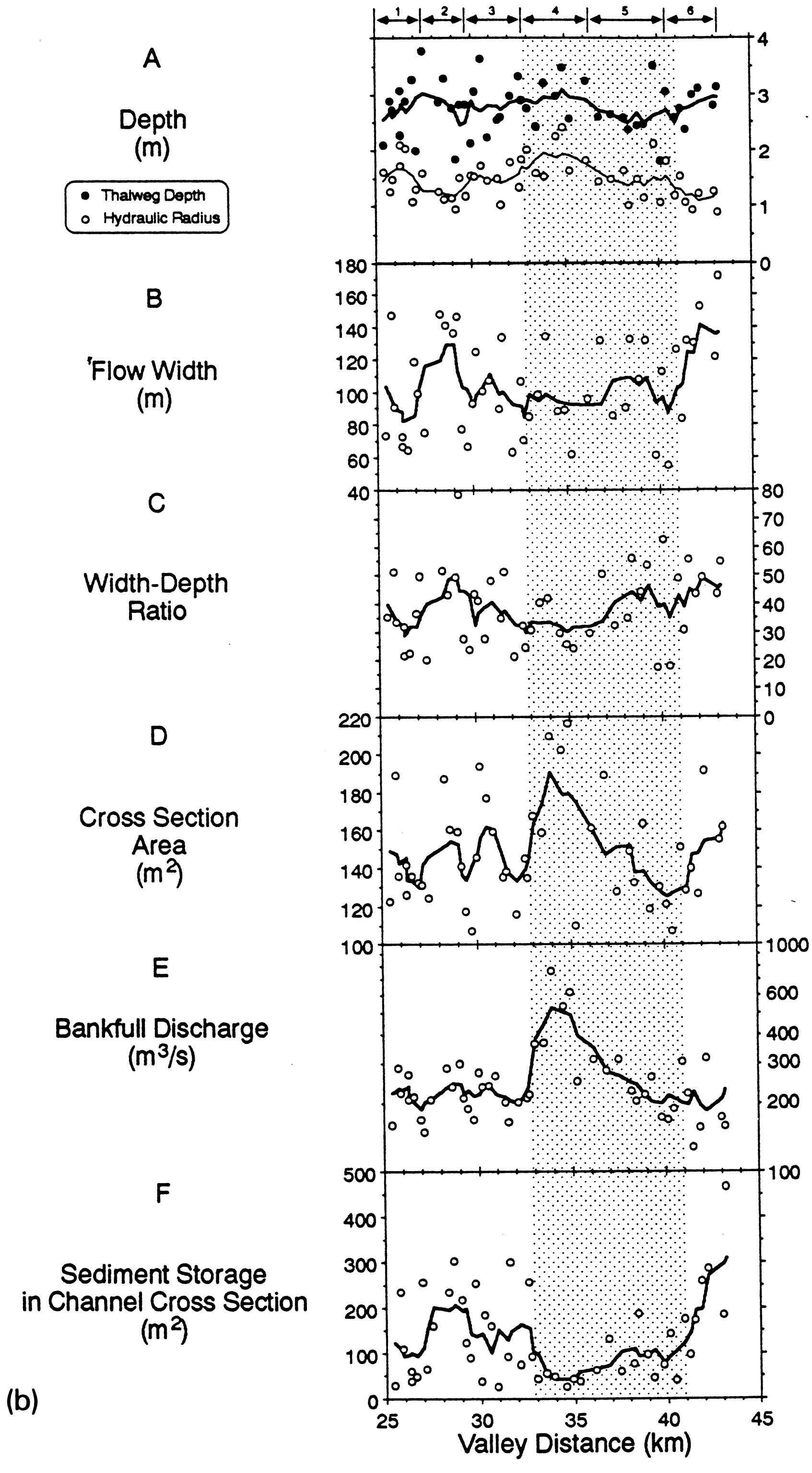


Fig. 5. Hydraulic responses of the Jefferson River, east flank of Tobacco Root Mountains, southwestern Montana, as it crosses a zone of uplift (Jorgensen, 1990).

and incision results in deeper effective channels and greater bankfull discharge, so overbank flooding is less apparent (Burnett, 1982; Ouchi, 1985; Jorgensen, 1990).

Width/depth ratios can be attained by direct observation of channel-element geometry in fluvial strata. Unfortunately, comparison of width/depth ratios above and adjacent to suspected sites of paleodeformation are no more indicative than such determinations in the modern. More indicative evidence of increased flood frequency, swamp development, or lake impoundment in sites of lowered slope, however, will often be preserved as increased overbank-fine element preservation, swamp deposits, or lake deposits, respectively. For instance, Guccione and Van Arsdale (1994) cite lake deposits associated with the St Francois River just upstream from the seismically active Blytheville Arch in the New Madrid Seismic zone of southeast Missouri. Even if damming is not sufficient to develop lake deposits, evidence of increased flooding should be apparent. Increased swamp development on the upstream side of uplifts, or within subsided zones should prompt increased preservation of organic matter in most cases. This may be preserved as carbonaceous overbank fines, coal/lignite deposits, and/or hydric paleosols. Increased preservation of overbank fines with or without organic preservation has also been cited as an indication of increased relative subsidence (e.g., Bridge and Leeder, 1979); however, preservation of clay is closely linked with supply of clay, making this indicator a reflection of climatic influences in the source area as well (see Garrels and Christ, 1965).

Channel width, depth, width/depth ratio, bankfull discharge and sediment storage are all highly dependent on discharge, total sediment load, and sediment load type (Schumm, 1977). Anomalies in these variables alone, make for flimsy evidence for tectonic warping. The environments and lithofacies related to the resultant changes in flood frequency, however, can prove to be useful evidence for tectonics in both modern and ancient environments. Erosively resistant strata, however, may also promote water impoundment. This may be related to tectonic up-warping, or be unrelated to tectonics entirely. Inspection for bedrock incision should thus be made. If this bedrock is undeformed, a tectonic interpretation may not be in order.

2.5. Grain-size change

Variations in stream power in response to increased or decreased slopes in deformed zones has a direct effect on grain size of stream bedload. In general, tectonically increased slopes will be characterized by increased bedload grain size, and vice versa (Ouchi, 1985; Jorgensen, 1990). The Guadalupe River in Texas for instance experiences steeper slopes as it crosses the Sam Fordyce fault zone and enters a zone of subsidence. The river has a dramatically increased bedload grain size where it crosses the fault zone and encounters increased stream power (Ouchi, 1985).

Use of grain-size variation as an indicator of deformation can prove to be complicated. First, grain size is not only a reflection of stream power, but also a reflection of sediment availability. For instance, the Humbolt River, Nevada, aggrades as it crosses a zone of subsidence. The river deposits most of its coarse load there (Jorgensen, 1990), so deposits will be finer downstream regardless of the slope encountered. In another example, the Bogue Homo and Tallahala creeks of southern Mississippi do have coarser bedload where their gradients increase across the Wiggins uplift. The coarser grain size, however, is largely due to the fact that the uplift exposes the gravel-rich Citronelle Formation to stream erosion. Increased grain size here is then more a reflection of bedrock supply than stream power, and thus grain-size increase is an indirect effect of uplift (Schumm et al., 1994).

It is also the case, that anything causing locally increased slope can cause increased stream power and increased grain size. Streams crossing a nickpoint, or a ledge of resistant bedrock may thus produce similar results. Because of their complicated nature, bedload grain-size variations should be used with extreme caution when interpreting present or past deformation.

3. River response to lateral tilting

Two styles of lateral channel migration typically result as a response to lateral tilting (Fig. 6). These are sudden avulsion of a stream toward the lower down-tilt part of the floodplain, or ‘combing’ (sensu

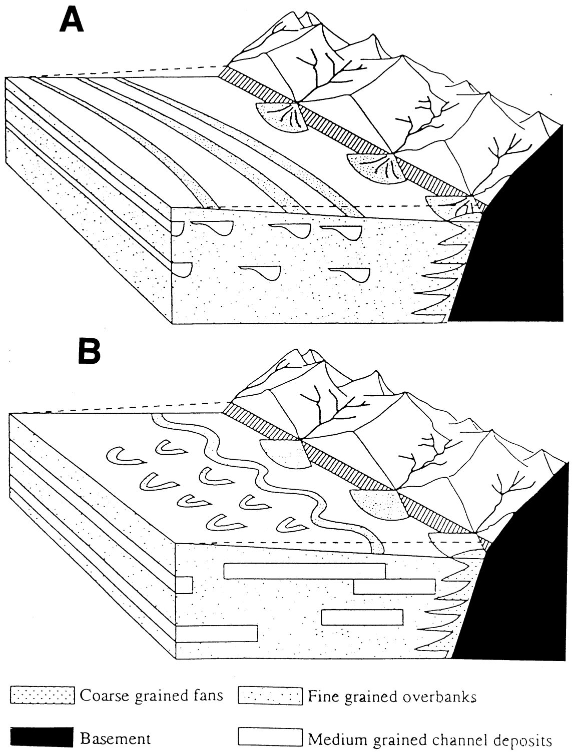


Fig. 6. Generalized channel-belt deposits resulting from down-tilt avulsion (A) and down-tilt combing (B) (Peakall, 1995).

Todd and Went, 1991), which refers to slow migration by preferential downslope erosion and/or meander cutoff on one side of the river (Alexander and Leeder, 1987; Leeder and Gawthorpe, 1987). In general, avulsion tends to produce isolated sand ribbons and immature channel belts (Fig. 6), whereas combing results in wide channel belts (Alexander and Leeder, 1987; Peakall, 1995). If rivers comb, instead of avulse, tilting must generate gradients sufficient to induce preferential lateral migration but small enough not to cause immediate avulsion. The limited number of quantified examples suggests that areas of lateral migration by avulsion have tilt rates of $\geq 7.5 \times 10^{-3}$ radians ka^{-1}, but examples of lateral migration by combing have tilt rates 2–3 orders of magnitude smaller (Peakall, 1995).

Other causes for unidirectional lateral migration of channels, besides tilting, have been proposed. For instance, some fan channels migrate unidirectionally without tilting, apparently because aggradation of former channel courses and levees act as barriers to channel movement, progressively deflecting channels in one direction in some cases (Wells and Dorr, 1987). Where bedrock-controlled, rivers might migrate down shallow-dipping resistant substrate in areas that were tilted prior to deposition, a process called monoclinal shifting (Gilbert, 1877). As well, rivers may be ‘pushed’ away from a basin margin by fan growth or generally high sediment supply on one side of the drainage (Blair, 1987). Bank aspect, freeze–thaw cycles (Lawler, 1986), and the influence of prevailing winds (Fairchild, 1932) also have been suggested as causes of preferential river migration, but these controls have not as yet gained widespread acceptance.

For down-tilt lateral migration of a channel to occur, the down-tilt side must also be the site of minimum elevation. In most tilted basins, this will be the case. In instances where sediment supply from the down-tilted side is greater than the subsidence rate, however, the basin may be overfilled on the down-tilt side, making the locus of minimum topography slightly offset from the locus of maximum subsidence. This and the factors above should be considered before the evidence described below is used to support lateral epeirogenic tilting.

3.1. Avulsion

Cross-valley tilting will commonly force rivers to avulse toward the lower/down-tilted side of the floodplain (Bridge and Leeder, 1979; Alexander and Leeder, 1987; Dumont and Garcia, 1991; Dumont and Hanagarth, 1993). Evidence that a river has avulsed toward the down-tilted side of its floodplain exists as an asymmetric position of a river in its valley and/or progressive unidirectional abandonment of channels in the down-tilt direction (Fig. 7) as evidenced by Osage ‘underfit’ type streams on the up-tilted side (see Dury, 1970). Dumont and Hanagarth (1993), for instance, cite five successive Holocene stages of down-tilt shifting for the Beni River in the active asymmetric Beni Basin of Peru. Avulsion history is based on current underfit streams in the basin which have similar channel width and meander wavelength as the modern Beni River, and become younger toward the current river position.

Bridge and Leeder (1979), Alexander and Leeder (1987) and Gawthorpe and Colella (1990) modeled

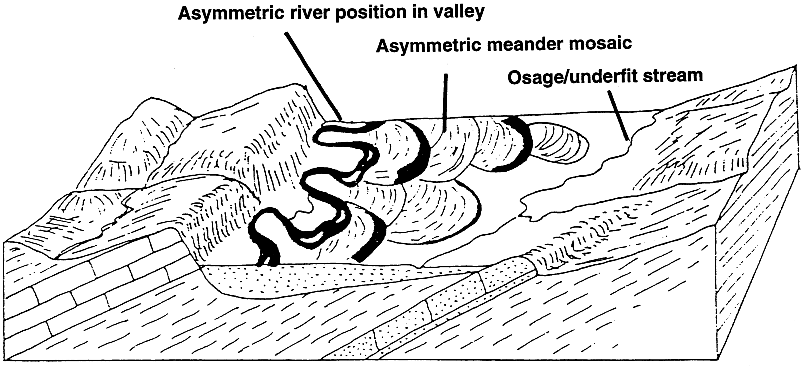


Fig. 7. Some features typical of floodplains in valleys undergoing lateral tilting. Tilting commonly results from epeirogenic warping of the craton, asymmetric rifting, tectonic loading by thrusts, and tilting of sags in piggy back basins on thrusts (Schumm et al., in press).

effects of basin tilting on alluvial architecture and noted that preferred channel avulsion toward the fault in a tilted basin will cause channel belts to be clustered on the down-tilt side of the basin. Such modeling has shown that floodplain tilting does not affect cross-section-averaged values of channel-belt deposition, but it does reduce the effective floodplain width. This causes channel-belt proportion and interconnectedness to increase locally on the down-tilt side, and vice versa (Alexander and Leeder, 1987; Leeder and Alexander, 1987; Bridge and Mackey, 1993). Alexander and Leeder (1987) further note that soils are exposed longer with little overbank deposition on the up-tilted side of tilted floodplains, resulting in more mature soils and paleosols in these areas. Channel-belt deposits will also tend to be wider in tilted basins if combing was a significant component of channel migration (Leeder and Alexander, 1987).

Mack and James (1993) tested these models by comparing syntectonic Plio–Pleistocene braided fluvial deposits from asymmetrically (Palomas and Mesilla basins) and symmetrically (Hatch-Rincon and Corralitos basins) subsiding sub-basins of the Rio Grande rift system, southern New Mexico. They determined that, compared to symmetrical basins, fluvial deposits in asymmetrical/tilted basins have the following: (1) a narrower effective floodplain (where effective refers to areas of floodplain prone to channel occupation); (2) a higher percentage of multistory channel sandstone bodies; (3) a higher ratio of channel to floodplain deposits near the basin axis; (4) fewer paleosols and paleosols with a generally lower degree of maturity near the basin axis; and (5) a lower proportion of sets of planar cross-beds.

3.2. Combing

Rivers may comb laterally in the down-tilt direction where tilt rates are not excessive. Evidence for combing of modern streams exists as asymmetric mosaics of recently formed meander loops which are dominantly concave to the axis of maximum subsidence (Mike, 1975; Leeder and Alexander, 1987; Alexander et al., 1994), as evidenced by oxbows and large meander scrolls (Fisk, 1944) (Fig. 7).

Leeder and Alexander (1987) studied the Madison River of Montana which currently flows through an actively tilting basin. They noted that preferential meander-belt shift in the down-tilt direction tended to destroy all cutoff loops generated on the downslope side of the meander belt, leaving only those meander belts formed on the up-tilt side. The resulting meander loop mosaic is thus dominated by loops oriented concave to the locus of maximum subsidence. There are three important implications from this observation that apply to channel-belt architecture in fluvial sediments of syndepositionally tilted basins: (1) lateral-accretion surfaces will tend to dip preferentially in the up-tilt direction; (2) meander-belt sand bodies without preferential dip or lateral-accretion surfaces can be assumed to have evolved randomly without the influence of imposed tectonic slope; and (3) under aggrading conditions, the base of the sand body produced by combing will climb up section in the down-tilt direction (Leeder and Alexander, 1987). Further examination of the Madison River by Alexander et al. (1994) revealed that the combing process involved various stages of incision and aggradation that complicated the picture beyond the original simplistic model, but the central premises of the original work held.

A few examples of unidirectionally oriented lateral-accretion surfaces have been observed in the rock record, and are explained as a result of coeval lateral tilting using the assertions of Leeder and Alexander (1987) (Todd and Went, 1991; Woolfe, 1992; Hazel, 1994). Though the principles are most easily applied to meandering streams (Woolfe, 1992; Hazel, 1994), they may be applied to braided systems as well with a few modifications. For instance Todd and Went (1991) note that lateral-accretion surfaces in braided stream deposits may dip either asympathic (dipping in direction opposite of channel combing) or sympathic. Sympathic accretion sets form as lateral bars migrate preferentially in the down-tilt direction, whereas asympathic accretion sets represent channel-to-bank filling of slough channels developed on lateral bars as these bars migrate down-tilt and detach from the up-tilt bank (Todd and Went, 1991).

Nanson (1980) observed an exception to the Madison River example from the tilted floodplain of the Beatton River, Canada. Here, meander loops were interpreted to have been cut off preferentially on the down-tilt side of the river channel because the amplitude of these loops was greater than the amplitude of the up-tilt bends. This appears to be a special case, however, where the Beatton River is confined to a narrow valley and unable to comb freely (Peakall, 1995).

Paleocurrent orientations have also been used to argue for tilting and lateral migration of fluvial channels, and are best applied where combing is a factor (Nakayama, 1994, 1996; Martins-Neto, 1994). Martins-Neto (1994) observed that Proterozoic fluvial strata from axial rift-fill deposits of the São João de Chapada Formation, Brazil, shift from east to southeast upward in the succession. He used this to argue for progressive tilting and lateral migration of channels southward (perpendicular to the basin axis) over the course of fluvial deposition.

4. Discussion

Rivers have multiple responses to epeirogenic warping (Figs. 6 and 8), and some are closely interrelated. The lowered stream power imposed by lowered gradients, for instance, is the main cause for aggradation, which in turn promotes increased overbank flood frequency, and vice versa (Fig. 8a). These responses thus tend to occur in association (e.g., Tallahala Creek, Burnett, 1982). Bedload grain-size changes should also be closely tied to stream power and slope; however, this response is typically complicated by issues of source stratum (Fig. 8a). The pattern on the other hand may be altered in lieu of, or in concert with, aggradation/degradation as a means of compensating for slope-altered stream power (Fig. 8b). Deformation of channel profile generally reflects lack of adjustment of the channel to altered slopes, and as such reflects a temporary condition. By the same token, profile deformation may be one of the best indicators of recent warping; but only if local bedrock control, local shifts in sediment supply, and nickpoint incision can be eliminated as causes. Where rivers lack the stream power, or are improperly oriented, to adjust to uplift by any of these means, rivers will tend to be deflected (Fig. 8a). Though other mechanisms for progressive and unidirectional lateral migration of rivers than tilting have been proposed, this remains a reasonably reliable indicator of epeirogenic warping.

Each of the conditions described above, except possibly profile deformation, has the potential to be preserved in the stratigraphic record. The difficulty in most cases is that similar effects may be generated by non-tectonic mechanisms. Where depositional features analogous to those produced at modern sites of tectonic deformation appear in the rock record, tectonic interpretation is based on reasonable elimination of other possible causes. Tectonic interpretations are strengthened greatly by close association of tectonically interpreted depositional features with historically active structures.

The tectonic effects discussed in this paper should be viewed as a collection of tectonic indicators. Like most indicators used in the geosciences, these have caveats and alternative explanations that must be addressed before application of the indicator can be considered valid. Likewise, few geologic interpreta-

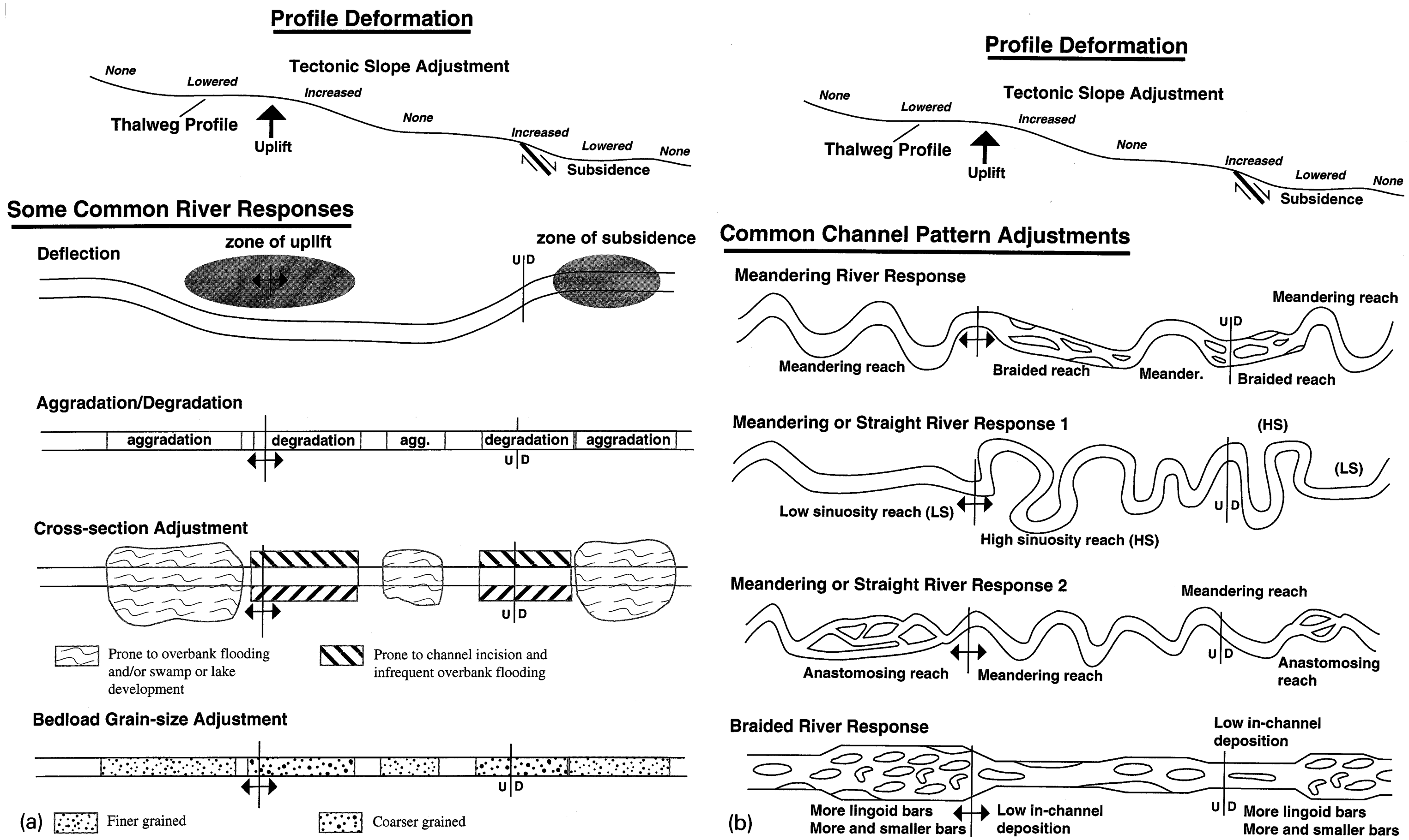


Fig. 8. Summary of common channel responses to longitudinal profile deformation. (a) Deflection and variations related directly to shifts in stream power. (b) Some common variations of channel pattern.

tions can be substantiated based on the presence of one indicator. Interpretors of both modern and ancient systems should investigate for as many of the listed tectonic indicators as possible before making a case for a tectonic perturbation.

The inherent difficulty in structurally recognizing broad gentle warping, and the effort and expense of non-targeted geodetic and leveling studies, makes fluvial interpretation an especially attractive and powerful tool for locating sites of modern warping. Such studies also provide a cost-effective precursor to geophysical investigations of structures, allowing investigators a way to optimize expensive data collection procedures. Where ancient epeirogeny has not left obvious structures, fluvial sediments may offer the only clues to timing, location and magnitude of deformation. The controls on fluvial lithofacies distribution by ancient structures discussed in this paper are also common place. Several authors (reviewed in Miall, 1996 and Schumm et al., in press) have drawn attention to the importance of these controls in development of petroleum and coal exploration/production strategies. Various authors have also pointed to the enigma of rapidly deforming, yet low-amplitude epeirogenic deformation over many modern structures (Adams, 1980; Schweig and Ellis, 1994). Probably one of the most overlooked uses of fluvial tectonic studies is using fluvial sediments for assessing the mesoscale (10^3–10^4 years) structural behavior for mid-continent deformation.

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