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The formation and maintenance of single-thread tie channels entering floodplain lakes: observations from three diverse river systems

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Abstract. Tie channels connect rivers to floodplain lakes on many lowland rivers and thereby play a central role in floodplain sedimentology and ecology, yet they are generally unrecognized and little studied. Here we report the results of field studies focused on tie channel origin and morphodynamics in three contrasting systems: the Middle Fly River, Papua New Guinea, the Lower Mississippi River, and Birch Creek in Alaska. Across these river systems, tie channels vary by an order of magnitude in size, but exhibit the same characteristic morphology and appear to develop and evolve by a similar set of processes. In all three systems, the channels are characterized by a narrow, leveed, single-thread morphology with maximum width approximately one tenth the width of the mainstem river. The channels typically have a V shaped cross-section, unlike most fluvial channels. These channels develop as lakes become isolated from the river by sedimentation. Narrowing of the connection between river and lake causes a sediment-laden jet to develop. Levees develop along the margins of the jet leading to channel emergence and eventual levee aggradation to the height of the mainstem levees. Bi-directional flow in these channels is common. Outflows from the lake scour sediment and prevent channel blockage. We propose that channel geometry and size are then controlled by a dynamic balance between channel narrow-

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ing by suspended sediment deposition and incision and widening by mass failure of banks during outflows. Tie channels are laterally stable and may convey flow for hundreds to a few thousand of years.

1. Introduction

On many lowland rivers, sediment-laden flows exit the mainstem river and enter floodplain lakes where they deposit their sediment load and create single-thread leveed channels (Figure 1). Through continued deposition, these channels grow, prograde and convey sediment-laden flow farther into lakes. During rising river stages, flow is directed into lakes. As river stages fall, however, lakes drain back to the river and flow in these channels reverses direction. Through tie channels, many river systems maintain both temporally and laterally stable connections between the river and off-river water bodies (ORWB) (Figure 1). A review of maps, aerial photographs, and satellite images reveals that such channels are a widespread feature of lowland floodplains around the globe (Figure 2).

The term "tie channel" originates from *Blake and Ollier* [1971] who used it to describe the numerous channels connecting the Fly River in Papua New Guinea to ORWB that experience regular reversals in flow between the river and floodplain. Published studies of floodplain systems suggest that tie channels play an important role in the hydrology, sedimentation and geomorphic development of these fluvial systems. For example, a detailed study of the hydrology and depositional mechanics of the Middle Fly River floodplain, Papua New Guinea [e.g. *Day et al.*, 2008] quantified that floodplain channels convey about 20% of the flow and sediment from the river onto the floodplain. Though not focused on tie channels, prior studies on the importance of connectivity between rivers and ORWB indicate connectivity influences the distribution of water and sediment onto the floodplain and into ORWB [*Smith and Perezarlucea*, 1994; *Mertes et al.*, 1996; *Dunne et al.*, 1998] and serves critical biological and geochemical functions [*Ward et al.*, 2002] such as: allowing the transfer of carbon from floodplains to rivers [*Lewis et al.*, 2000; *Thoms*, 2003]; providing access for aquatic species to areas of refugia [*Swales et al.*, 1999] and rearing habitats [*Pringle et al.*, 2000]; and influencing the biological and chemical characteristics of ORWB [*Knowlton and Jones*, 1997; *Miranda*, 2005].

Although of broad significance, we have found only a few papers outside those by our research group which offer some description of these channels. Prior work on the Middle Fly River documented the role tie channels play in the routing of water and dispersal of sediment across the floodplain [e.g. *Day et al.*, 2008, 2009]. Using Optically Stimulated Luminescence (OSL) dating of sediment, *Rowland et al.* [2005] determined tie channel extension and levee accretion rates on Fly River tie channels and compared these rates to rates for tie channels both on the Lower Mississippi River and a small river in Alaska (Birch Creek). The same Lower Mississippi River tie channel was also the focus of a more detailed sedimentological and morphological study reported by *Rowland and Dietrich* [2006].

Here we compare the data initially reported in *Rowland and Dietrich* [2006] for the Lower Mississippi River with previously unreported data for both the Middle Fly River to address the following two questions: 1) is there a characteristic morphology of tie channels, and 2) what processes lead to their formation, persistence and growth? These three field studies encompass tie channels which vary widely in size and setting providing a broad perspective on this class of floodplain channel. For each field site we quantified the channel morphology, rates of channel development, and sediment grain-size of deposits, and documented processes that appear to control channel evolution and form. Despite the widely varying sizes, bed material and hydrology, the tie channels were found to have

a similar cross-sectional morphology and to be a common scale relative to the mainstem source channel. In all cases the channel morphology appears to result from a dynamic balance between levee sedimentation, channel scour and mass failure of the banks. The similarity of form and process leads us to propose a conceptual model of tie channel development. While important in their own right, we also suggest that our findings about tie channels have application to single-thread deltaic channels in general.

2. Field Sites and Observations

2.1. Middle Fly River Setting and Data Collection Summary

The Middle Fly River which lies along the western border of Papua New Guinea drains 18,400 km^2 and has a mean annual discharge of 2,244 m^3/s [*Dietrich et al.*, 1999](Figure 2). The sand-bedded Middle Fly River flows along a sinuous 450 km reach between the junction of the Fly and the Ok Tedi Rivers (D'Albertis Junction) and the confluence of the Fly and the larger Strickland River (Everill Junction). We examined tie channels along the entire length of the Middle Fly River. The lowland channel increases in drainage area from 12,000 km^2 to 18,400 km^2 as its slope decreases downstream from 6.6x10^{-5} to 2x10^{-5} and its bed material decreases in size from 0.3 mm to about 0.1 mm [*Dietrich et al.*, 1999]. The Middle Fly lies in a wet-tropical climatic zone strongly influenced by ENSO cycle climatic events. Annual precipitation exceeds 10 m in the headwaters and drops to 2 m on the lower portion of the Middle Fly, with drought conditions commonly occurring during El Nino events [*Dietrich et al.*, 1999]. The Fly River is not dammed and its floodplain is unmodified by agriculture or flood control measures, but mining (and to a much lesser degree, logging) have greatly increased the sediment load in the Fly River system. Beginning in 1985 the Ok Tedi mine, located on the tributary of that name, began riverine disposal of tailings. Prior to the mine, the suspended sediment concentration of the Middle Fly averaged 100 mg/L and the average annual sediment discharge was about 5×10^6 t/yr [*Dietrich et al.*, 1999]. Post-mine suspended sediment concentrations in the 1990s ranged on the order of 400 to 500 mg/L while the total sediment load to the Middle Fly River increased by about four to five times [*Day et al.*, 2008]. Prior to the onset of mining activities, 60% of the ORWB along the Middle Fly River were connected to the river via tie channels [*Day et al.*, 2008]. Since mining and by 2005, at least one major tie channel in the lower reaches of the Middle Fly River as become plugged with sediment [*Day et al.*, 2009], the state of other tie channels since 2005 is not presently known to the authors.

An extensive field program conducted along the Middle Fly River from 1990 to 1998 [e.g. *Day et al.*, 2008] included a detailed study of tie channel morphology and hydrology. Over the course of the program researchers surveyed up to 17 cross-section locations along 8 tie channels and recorded long profiles of 6 channels using a combination of differential global positioning system (DPGS) and stadia rod and hand level surveys. The 1990 to 1998 research program established a network of water-level recorders tied to a common datum across the river system and along tie channels. Using water-surface slopes derived from the water-level recorders and cross-sectional surveys, *Day et al.* [2008] calculated velocities and estimated discharges along several tie channels. Based on the accumulation of mine-derived copper in sediments, *Day et al.* [2008] determined deposition rates across the Middle Fly floodplain and in oxbow lakes, and concluded that up to 40% of the total load of the river is deposited on the floodplain, with one-half of that occurring as overbank deposits adjacent to tie and tributary channels. Our additional surveys conducted in 2003

included measurements of tie channels widths and lengths with a hand-held GPS and laser range finder, and collection of sediment samples for grain-size and OSL analysis. We used a Coulter LS130 Laser Diffraction Analyzer with a size range sensitivity of 0.4 to 900 microns to determine grain-size distributions. *Rowland et al.* [2005] provide a detailed description of the OSL sampling and analysis methods.

Along the Middle Fly River, tie channels link the mainstem to both oxbow lakes (47) and blocked valley lakes (29). Oxbow lakes with tie channels appear to be predominantly formed by neck rather chute cutoff processes. On the Fly River, blocked valley lakes formed when tributaries became dammed in response to sea level rise induced aggradation of the mainstem [*Blake and Ollier*, 1971; *Dietrich et al.*, 1999](Figure 1f) and are similar to floodplain lakes described by *Vernon* [1942], *Baker* [1978], and *Kuenzi et al.* [1979]. These blocked valley lakes tend to have larger surface areas but shallower depths than oxbow lakes (3 to 5 m versus 10 m) and can dry out during periods of drought. We differentiate Fly River oxbow lakes into two types: open and closed. In closed lakes, tie channels provide the only inlet or outlet for channelized flow to the lake. Open lakes have additional channels that allow for the exchange of water with the floodplain, other ORWBs and/or the river.

2.2. Middle Fly River Observations

Sequential aerial photographs, satellite imagery, and direct field observations of two oxbow lake cutoffs on the Fly River (one of which is depicted in Figure 3) indicate that tie channels form contemporaneously with meander cutoff and lake formation. Rather than eroding into fully established plugs, tie channels appear to develop as flow into and out of the lakes prevents plug sedimentation along one of the oxbow margins. As sedimentation progressively seals off the oxbow limb an increasing percentage of water between river and lake is routed through the channel until it is the only conduit for flow into and out of the lake below flood stage. The channels may form in either the up- or downstream meander limb or occasionally in both. In other, less common instances, a channel may develop when the mainstem river migrates back into the existing oxbow lake creating a direct connection between the river and the open waters of the lake (Figure 1d). *Caldwell and Fitzgerald* [1995] found that flow-reversing outlet delta channels in the New England region of the USA were consistently connected to the outside of river bends and oriented in the downstream direction (forming an acute angle with the mainstem bank). On the Middle Fly River, this configuration is the most common but not universal. The connection angle of the channels may influence the diversion of sediment from the mainstem into the channels. The acute angle likely restricts the divergence of bed load material into the channel, as the coarse sediment would have to reverse transport direction and head in an upstream direction to enter the mouth of the channel. On the Middle Fly River, the first tie channel to plug with sediment following increased sediment load due to mine tailing discharge was one that connected in line with the river, pointed in a downstream direction.

Rainfall events commonly occur over only a fraction of the Fly River basin leading to asynchronous river and floodplain hydrological responses to storm events. Frequent storm events in the headwaters result in flood pulses that move down the system and pump water out through the network of channels onto the floodplain and into lakes. Rain on the seasonally dry mud-rich floodplain ponds, leads to flooding and outflow to the mainstem. *Day et al.* [2008] noted that even during a flooded state, however, rain on the

plain can induce a pressure rise that drives water off the floodplain back to the mainstem via the tie channels. During falling river stage water drains to the mainstem through tie channels, at times with considerable velocities [*Day et al.*, 2008]. Through the system of water-level recorders *Day et al.* [2008] documented water-surface slopes along tie channels up to $8.6\text{x}10^{-4}$ and $6\text{x}10^{-4}$ during flow from the lakes to the river (outflow) and from river into lakes (inflow), respectively. Both slopes are greater than the mainstem slope by a factor of ten and correspond to estimated velocities up to 1.7 m/s, well in excess of mainstem average bankfull velocities of 1 m/s [*Dietrich et al.*, 1999].

Figures 1c,e,f present typical views of tie channels entering lakes along the Middle Fly River. These single-threaded channels snake into the lakes, abutting one margin of the lake near the junction with the river before invariably traversing the lake, sometimes multiple times, commonly at near right angles (Figure 1c). The narrow zone of deposition bordering the channels and the absence of scroll bars or other evidence of lateral migration suggest that these channels are laterally stable. The sinuosity of tie channels appears to be due to two effects: turning of the lakeward propagating channel against the resistant lake margins and/or bathymetric obstacles, and temporary plugging of the channel outlet into the lake followed by a diversion to one side or the other of the plug. In 2003, we observed in the field, a new channel outlet developing through a levee breach located upstream of a blocked outlet.

On the Middle Fly River, the maximum levee-crest elevation of tie channels is equal to the height of the mainstem river levees, and occurs at the junction of the channel and the mainstem (Figure 4). Levee heights steadily decrease with distance into the lake. Long profile surveys of levee crests along two Fly River channels gave gradients of

$5\text{x}10^{-4}$, similar to water-surface slopes recorded along one of these channels, and an order magnitude greater than the mainstem water-surface slope.

In general, tie channel beds are several meters above the adjacent mainstem river bed. The longitudinal profiles of the channel bed and adjacent banks are generally parallel, except where the profiles join the mainstem and where they terminate in the lake in a mouth bar (Figure 4). Near the river end of the channel, the bed slopes steeply toward the river but typically does not grade to the mainstem bed. The bed of the channel presented in Figure 4, for example, lies 2 m above the bed of the Fly. Further toward the lake the bed slope reverses direction and runs roughly parallel to the adjacent levees leaving a distinct elevated inflection (or "sill") in the bed profile (Figure 4). The relative distance of this sill from the mainstem into the lake ranged from 0.2 to 0.6 and averaged 0.4 the total length of the channel for the six channels surveyed. The survey depicted in Figure 4 extends into the lake and also records an abrupt rise in the bed elevation where the channel enters the lake and shoals into a mouth bar. All three channel surveys which extend this far into the respective lakes show a similar shoaling.

In cross section, tie channels are narrow and V-shaped with no clear break in slope between the channel banks and bed (Figure 5). The channels have a mean width to depth ratio of 5 (the mainstem Middle Fly has an typical width to depth ratio of ~ 20) and bank slopes that vary from 10 to 57 degrees with a mean of 27 degrees (Table 1). Steps and local irregularities along the channel banks (Figure 5) result from slump blocks and bank failures. Dense vegetation characteristically obscures the banks and helps to anchor failure blocks along the channel walls (Figure 6).

We observe different along channel trends in width between tie channels connected to closed versus open oxbow lakes (Figure 7). Tie channels associated with closed oxbow lakes narrow with distance from the river (Figure 7a). Whereas channels connected to open oxbow lakes (where tributaries also enter) show no systematic trend in width (Figure 7c). Tie channels connected to blocked valley lakes also show a decrease in width with distance from the mainstem river, but the trend is not as strong (lower r^2 values) as that observed for closed oxbow lakes. Mean channel depth tends to decrease with the narrowing width, hence mean channel-bank slope tends to be spatially constant (Figure 8). In Figure 8 we lacked sufficient data to distinguish trends between tie channels connected to open and closed oxbow lakes.

Comparison of grain-size data for tie channel deposits to the suspended sediment carried by the mainstem, at a reach about 145 km down the Middle Fly River from D'Albertis Junction, shows that the coarsest layers in the channel levees are composed of the coarse tail of the suspended load carried by the Fly (Figure 9). Near-surface samples of the levee crests, along three channels (26 total samples) were 10-20% clay, 60-70% silt and 10-30% fine sand. Hence the levees, while finer than the Fly River bed appear deficient in fine silt relative to that found in suspension. At one tie channel, water-level monitoring data obtained by G. Day (unpublished) permitted calculation of the magnitude and duration of the total boundary shear stress. These data show the boundary shear stress is sufficiently high that even the coarsest fraction of the Fly River suspended load would be carried in suspension along the tie channel for significant fraction of time during both inflow and outflow. This would explain the size distribution of the levee deposits (derived from suspension).

Sedimentation by tie channels leads to extension of the channels into previously open lake waters and progressively fills the lakes. Using OSL dating, *Rowland et al.* [2005] concluded that channel extension rates on selected Fly River tie channels increased from a long term (measured over 350 to 900 years) rate of 2 to 7 m/year to 40 m/year since the 1980s. Both the magnitude and timing of the extension rate increase appears consistent with the estimated 4- to 5- fold increase in sediment load due to riverine disposal of mine tailing into the Fly River system. The long term rate of extension and surveys of tie channel topography allow us to estimate that tie channel extension typically adds 6,000 tons of sediment per year per lake. Extrapolated to all lakes connected by tie channels, this rate is equivalent to a net lake deposition of 2% of the pre-mine suspended sediment load of the Fly. *Day et al.* [2008] estimate that the 900 km of floodplain channels (tie and tributary channels) disperse about 20% of the post-mine river suspended sediment load into all Middle Fly floodplain settings.

2.3. Lower Mississippi River Setting and Data Collection Summary

Draining over 3.2 million km^2 of North America, the Lower Mississippi River has a mean annual discharge of 18,400 m^3/s [*Mossa*, 1996]. In 2002 and 2005, field investigations were conducted on a single tie channel connecting the Raccourci Old River (ROR) oxbow (30° 54.7'N, 91° 36.62'W) to the Mississippi River (Figure 1a). The ROR oxbow lake lies approximately 65 km upstream of Baton Rouge, Louisiana just below Red River landing and 27 km downstream of the Old River Control Structure. The Old River Control Structure currently diverts approximately 30% of the Mississippi River flow into the Atchafalaya system. The water-surface slope of the Mississippi River just upstream of the ROR site, at Natchez, Mississippi, is 5.2x10^{-5} [*Biedenharn et al.*, 2000].

In the vicinity of the ROR, the bed of the Lower Mississippi River is dominated by fine sand and suspended sediment concentrations average 420 mg/L (measured between 1950 and 1991) [*Mossa*, 1996]. Approximately half of the suspended load (45%) is clay ($< 4\mu$m) sized [*Catalyst Old River Hydroelectric Limited Partnership*, 1999]. Since the 1850s the concentration of suspended sediment carried by the Lower Mississippi River has decrease between 50 [*Horowitz et al.*, 2001] and 70% [*Kesel et al.*, 1992]. Over the 20th century, the timing of suspended sediment decreases corresponded to soil conservation measures across the Mississippi River watershed in the 1930s, and dam construction on major tributaries, bank stabilization, and diversion of flow at the Old River Control Structure since the 1950s [*Keown et al.*, 1986; *Kesel et al.*, 1992; *Mossa*, 1996]. Our review of maps and aerial photographs of the Lower Mississippi River from Baton Rouge, Louisiana to Memphis, Tennessee found that 24 lakes (65% of total) at one time connected to the Mississippi River via tie channels. Today only half the lakes remain connected to the river and dredging, agriculture and/or flood control measures impact most of the connecting channels.

Unlike much of the Lower Mississippi River, the ROR oxbow lake is not isolated from the river by flood control levees. The lake formed by 1851 following an 1848 man-made cutoff aimed at shortening the river for navigation [*Gagliano and Howard*, 1984]. Since cutoff, the original flood control levees (along the margins of pre-1848 river course) have been maintained and the lake and its associated floodplain still receive periodic floodwaters. On the tie channel itself, the only apparent human alteration of the channel is a low-head dam installed in 1965 to facilitate boating and recreation during summer months by maintaining minimum lake levels. The impact of the dam on channel morphology appears limited, though it may have influenced the rate at which the channel advances lakeward

[*Rowland and Dietrich*, 2006]. In Alaska, beaver dams across tie channels commonly control lake levels in a similar manner [*Brown and Fleener*, 2001].

Our field studies at the ROR channel in 2002 included five cross-sectional surveys using a stadia rod, level, tape and laser range finder and collection of sediment cores for OSL dating. In 2005, we returned to the site and surveyed the bathymetry of the channel outlet into the lake using a hand-held GPS and depth finder and collected bed samples of the channel and lake using a ponar-type grab sampler. The cores and grab samples were analyzed for grain-size distribution. OSL dating provided deposition rates and an independent check on channel extension rates determined from hydrographic surveys and aerial photographs [*Rowland et al.*, 2005].

Though the field data on the ROR are more limited than for the Middle Fly River, extensive data resources exist for the Lower Mississippi River that document the evolution of the channel, and the hydrology and sediment load of the river. Topographic documentation of the ROR channel began in 1883 and since then regular hydrographic surveys conducted by the Mississippi River Commission and the United States Army Corps of Engineers (USACE) have recorded its development. Aerial photographic records of the site date to the 1940s, and satellite imagery supplements this data set over the last two decades. In 2003, the Louisiana Federal Emergency Management Agency (FEMA) Project collected high-resolution topographic data of the site using airborne laser swath mapping (ALSM) also commonly referred to as LiDAR [*Watershed Concepts*, 2005]. ALSM mapping provides detailed data on levee morphology and along-channel trends in levee height and channel widths. Stage records of the Mississippi River, obtained 8 km upstream of the

ROR at the Red River Landing gage, date to 1851. Extensive sediment transport data for the river have been collected at the Tarbert Landing station located 14 km upstream.

Prior research on the Lower Mississippi River also provide limited documentation on tie channels, locally known as "batture channels". *Russell* [1939] and *Fisk* [1947] both note the occurrence of prograding deltas in oxbow lakes but make no observations regarding their origin or role in lake processes. *Gagliano and Howard* [1984] provided data regarding the occurrence, rate of development and some morphological characteristics of these channels and introduced a four-stage conceptual model for oxbow lake evolution that included a phase in which batture channels play a role in lake sedimentation. In his exhaustive look of Lower Mississippi River, *Saucier* [1994] also noted the role of batture channels in introducing silts and clays into oxbows. Finally, *Shepherd* [2001] and *Guccione et al.* [2002] provide limited stratigraphic data from cores penetrating filled batture channels at Reelfoot Lake, Tennessee, USA.

2.4. Lower Mississippi River Observations

With a maximum width of 100 m and length of 9 km, the ROR is approximately four times larger than the typical Fly River tie channel, but shows distinct similarities in morphology, sedimentology and evolution. In planform (Figure 1a), the ROR channel is a single thread, sinuous, narrow (relative to its depth) channel with pronounced levees. Historical hydrographic surveys and aerial photographs record the channel extension into the lake and indicate that bends in the channel course remain fixed over time with no measurable lateral movement of the channel centerline [*Rowland and Dietrich*, 2006]. If we assume that the processes governing channel development have remained largely unchanged since channel formation, the prograding development of the channel and its lateral stability allow us to view changes in channel morphology along the channel in a temporal context. We can substitute space for time and assume that the youngest portion of the channel near the outlet into the lake represents the early stages of channel development and the oldest sections near the river represent the most evolved morphologies.

The long profile, cross-section surveys and along-channel width measurements of the ROR channel are presented in Figures 10, 11, and 12, respectively. In long profile, the ROR channel shows the same general characteristics as the Fly River channel (Figure 4): decreasing levee-crest elevations with distance from the river and a shoaling of the bed leading up to a mouth bar. Similar to the Fly River channel, the ROR channel also shows a steepening of the levee slope over the last tenth of the channel length. Levee crests on opposites sides of the ROR channel may vary in elevation with up to 2.5 m difference observed at some bends. At these locations, the width of the levees flanking the channel tend to be wider on the side of the channel with higher levee crests.

There are some important differences with the Fly River observations. Along the first 2 km (~20% of the length) of the ROR levee crests, the elevation is essentially constant, unlike the progressive drop in elevation of the Fly tie channels. Furthermore the bed slopes toward the mainstem river for nearly 7 km or 80% of the channel length. Hence, the inflection or sill is much further along the channel. Whether the dam was located at this point to take advantage of this inflection, or if dam has subsequently influenced the channel bed slope is not presently known.

In cross section the ROR channel shows the same narrow, V-shaped geometry as observed on the Fly River (Figure 11). Width to depth ratios range from 3.6 to 7.8 and average 5.8 (Table 1). A comparison of cross-sections along the ROR channels reveals

that the channel narrows and becomes less deep with distance from the river (Figure 12). Levee flanks bordering the channel also decrease in lateral extent and steepen with distance from the river. When nondimensionalized by the local river width, the rate of channel narrowing on the ROR channel is approximately the same as that observed in Fly River closed oxbow lakes and shows approximately the same relative maximum width of ten percent of the mainstem width (Figure 12). Despite the scatter in width measurements, the width of the ROR channel appears to remain relatively constant over the first two tenths of the channel length, the same interval over which we observed relative constant levee-crest elevations.

Mass failures of the channel banks are common along the length of the ROR. For much of the channel, these failures are rotational, evidenced by back-rotated blocks (Figure 13). The failures grow in size with the increasing size of the channel and individual failures near the river end of the channel extend for hundreds of meters along the channel and propagate up to 10 m into the levee crests. Fresh scarps show vertical offsets ranging from of a couple meters on large failures to tens of centimeters on newer channel banks near the outlet to the lake. The presence of back–rotated tree stumps exposed in the bed of the channel suggest multiple reactivations of failure blocks over time. Sediment drapes over failure blocks suggests that sediment deposition smooths the topographic irregularities from the cross-sectional profile and rebuilds the slumped banks.

Due to the size of the channel banks, and the large-scale slumps, the internal stratigraphy on the ROR levees is well exposed along the banks of the channel. The banks consist of horizontal to sub-horizontal alternating layers of fine to medium sand and layers of mud with organic debris. The layers are massive in character and range in thickness from a few centimeters up to tens of centimeters. No primary sedimentary structures such as ripple marks and cross-bedding were observed although faint planar laminations are visible. The size distribution of the sandy (coarse) layers is similar to that reported for the Lower Mississippi River suspended load but is much finer than the river bed material (Figure 14). The muddy bank layers are about 60% silt and equal parts of fine sand and clay (D_{50} of 18 μm). The mouthbar deposits consist of well-sorted sand with a D_{50} of 300 μm and a size distribution similar to that of the river-bed surface. Lakeward of the mouth bar, pro-delta sediments consist of clayey silt which decrease logarithmically in size for 1 km, after which point the surface layer remains relatively uniform in grain-size (D_{50}=10μm, 24% clay, 72%silt) [*Rowland and Dietrich*, 2006]. Sediment collected from the ROR channel 1700 m upstream of the mouthbar has a grain-size distribution similar to lake bed sediments [*Rowland and Dietrich*, 2006]. In the channel bed, however, these clayey silts have undergone sufficient consolidation to render them nearly impenetrable to grab sampler collection. The similarity in grain-size and elevation of the channel bed relative to the mouthbar/prodelta crest elevation (Figure 10) suggest that the ROR tie channel has incised into its own prodelta deposits as it has prograded into the lake.

Rowland et al. [2005] used OSL dating, historical hydrographic surveys aerial photographs to study changes in the rate that the ROR channel has advanced into the oxbow lake since its formation in 1851. From the 1850s to the 1930s the channel advanced at 84 m/yr, decreased to 53 m/yr from the 1930s to the 1950s and slowed to 18 m/yr since the 1980s. *Rowland and Dietrich* [2006] explored the possible influence of changes in Mississippi River sediment load, loss of flow volume due to lake infilling, changes in river hydrology and construction of the low-head dam on the rates of the ROR extension.

They concluded that changes in river sediment load was the dominant cause for extension rate changes. As with the Fly River tie channels, both the timing and magnitude of the extension rate changes corresponds well to changes in the sediment load of the river.

Using historical hydrographic surveys and digital elevation models, *Rowland and Dietrich* [2006] estimated the total volume of sediment introduced to the ROR oxbow by the tie channel, and calculated a sedimentation rate of $4.5\text{x}10^5$ tons/yr. By extrapolating this rate to other Lower Mississippi River oxbow lakes with evidence of tie channels (prior to large-scale floodplain modifications) *Rowland and Dietrich* [2006] estimate as much as 11 million tons/yr or 4% of the total pre-1963 load [*Keown et al.*, 1986] of the river may have been deposited into floodplain lakes by these channels.

2.5. Birch Creek Setting and Data Collection

The Birch Creek, a tributary to the Yukon River and the smallest of the three river systems studied here, lies in the subpolar continental climatic zone near the Arctic Circle [*Kostohrys and Sterin*, 1996]. In the summer of 2002, we investigated tie channels along a 30 km lowland section of the Birch Creek marked by discontinuous permafrost located approximately 10 km northeast of Circle, Alaska (65° 53.78' N, 144° 18.07' W) (Figure 1d). A drainage area of 6,500 km^2 supplies a mean annual discharge of 28 m^3/s to this section of the Birch Creek [*Kostohrys and Sterin*, 1996]. Due to strong seasonality, maximum discharges of up to 550 m^3/s occur during spring and summer rain events, while periods of no flow occur during winter ice over. Characterized by high sinuosity (river length/valley length: 2.6) and a low gradient ($2.6\text{x}10^{-4}$), this section of the Birch Creek appears to lie in the gravel-sand transition with mixed sand/gravel bed and point bars. Aside from placer mining operations in the upper reaches of the Birch Creek, no human modifications (dams, diversions, flood control or channel engineering) impact the river system. The placer mining contributes to elevated water turbidity with mean total suspended solids (TSS) of 151 mg/L [*Vohden*, 1999].

Data collection involved surveying channel long profiles and cross-sectional morphology using a stadia rod, meter tapes and levels, at two lakes. The lakes included a newly forming channel in a meander cutoff which occurred prior to 1980 and a well-developed prograding channel in an oxbow lake. At these sites, we also collected a series of sediment samples for grain-size analysis, and at the well developed channel we sampled for OSL dating.

2.6. Birch Creek Observations

Figures 15 and 16 present the long profile and cross-sectional surveys for the well-developed Birch Creek oxbow channel. This channel appears to have formed when the Birch Creek migrated into an existing oxbow lake (Figure 1 d). Similar to the ROR tie channel, the Birch Creek channel–levee crests have a relatively similar height for approximately 100 m (20% of the total length) then decrease in elevation with distance from the creek. The channel bed slopes toward the river for the first 100 m then drops in elevation at about the same rate as the levee crests until it rises and merges with the levees at the mouth bar. At 135 m from the river, the channel is narrow with steep banks (28°) and a width to depth ratio of 3.5 (Figure 16). As the channel enters the lake (cross sections at 450 and 495 m from the river) levees become less pronounced and the width to depth ratio increases as the bed shoals into the mouthbar. Along the well developed sections of the channel, the maximum width of 10 m is approximately ten percent of the average local width of the Birch Creek. At distances 135 m to 400 m from the river, the channel

narrows to ~3 m and has a width to depth ratio just under 2. A summary of the width to depth ratios and bank slopes for both surveyed Birch Creek tie channels is presented in Table 1.

The Birch Creek channel banks are heavily vegetated and experience widespread mass failures (Figure 17). The bulk composition of sediments in the Birch Creek tie channel banks contain less that 5% clay and equal parts fine sand and silt with abundant unaltered organic matter. During sediment coring, we encountered frozen materials at depths greater than 75 cm below the levee crest. In this system, with its low clay content, a combination of vegetation and permafrost conditions appears to provide bank cohesion. Sediments deposited in the mouthbar were the coarsest (D_{50} = 200 μm) observed anywhere along the tie channel, in either the channel bed or banks, but were finer than the gravel bed of the mainstem. Based on OSL dating of sediments cored from the Birch Creek tie channel levees, this 500 m long channel is approximately 300 years old and has extended lakeward at a rate of 1.2 m/yr [*Rowland et al.*, 2005].

3. Discussion

The tie channels at all three sites display a common morphology: 1) V-shaped channel cross sections with low width to depth ratios; 2) a width about ten percent the mainstem width at its origin; 3) levees that slope toward the lake and diminish in elevation toward the lake at more than 10 times the mainstem river slope; 4) bed-surface slope which reverses direction from riverward to lakeward; and 5) a sinuous planform that records a directional shift in tie channel extension rather than arising from lateral migration. At all three sites channel scour, bank slumping and suspended sediment deposition appear to sustain channel cross-sectional form. Alternating sand and silt-clay deposits make up the levees while the channel bed at the outlet into the lake tends to be primarily sand. Tie channel extension rate is strongly linked to sediment supply in the mainstem river.

3.1. Conceptual Model

To explain these features and link them to observed and inferred processes we propose a conceptual model for the formation and evolution of tie channels entering oxbow lakes (Figure 18). Shortly after meander cutoff (Figure 18a) limbs of the oxbow tend to plug with coarse sediments. The processes by which abandoned channel segments become sealed on a range of river systems has been extensively discussed in the literature and for oxbow lakes the sealing processes is commonly attributed to the deposition of bed load sediments across the meander cutoff limbs [see *Fisk*, 1947; *Allen*, 1965; *Russell*, 1967; *Reineck and Singh*, 1980; *Gagliano and Howard*, 1984; *Shields and Abt*, 1989; *Saucier*, 1994; *Hooke*, 1995; *Piegay et al.*, 2000; *Bridge*, 2003]. On the Lower Mississippi River, *Gagliano and Howard* [1984] estimated that complete plugging of the ends of cutoff meanders takes between 2 and 10 years, while *Hooke* [1995] observed that sealing of old channels takes between 1 and 7 years along the Rivers Bolin and Dane in the United Kingdom. Based on aerial photographs and satellite imagery of cutoffs on the Middle Fly River and the Birch Creek, the time for lake isolation on these river systems ranges from less than a decade (Middle Fly River) to greater than 20 years (Birch Creek).

Flow that initiates the tie channel starts in a shallow depression across one of these newly forming plugs (Figure 3) and progressively conveys more of the flow into and out of the lake as the plug aggrades and the oxbow limb seals. Based on our observations of newly forming channels on both the Middle Fly River and the Birch Creek, this new

channel forms coincident with limb sealing and does not develop by erosion into the plug of a fully sealed meander.

Confinement of inflow by the developing channel creates a sediment-laden jet. Sedimentation along the margins of the jet builds subaqueous levees constraining and reducing jet spreading [*Rowland*, 2007]. The focused flow is then able to incise the pro-delta deposits and to extend deposition farther into the lake (Figure 18b). Continued deposition onto the subaqueous levees, facilitated by varying lake levels, leads to the emergence of the levees above a minimum lake level (set by the sill elevation) and rapid colonization by vegetation (not depicted in Figure 18).

With each cycle of water inflow, rising lake levels progressively inundate the levees, leading to a shift in depositional mechanics away from unconfined spreading of a jet to channelized conditions. If the water stage in the channel rises faster than the surrounding lake waters, flow may overtop the levee crests and spread sediment laterally down the flanks of the levees. When lake level fully inundates the levees, continued inflow to the lake will be constrained by the still waters surrounding the channel. *Adams et al.* [2004] proposed that advective sedimentation would occur under the former conditions and that diffusive mechanisms would dominate under inundated conditions. Diffusive sedimentation would be driven by turbulent shearing between flowing water in the channel and the stagnant waters of the lake. The sharp velocity contrast between the channel and lake waters will result in lateral shear zones similar to those discussed by *Leighly* [1934], modeled experimentally by *Sellin* [1964] and numerically by *James* [1985], and described along the levees of the mainstem Middle Fly River by *Day et al.* [2008]. Such diffusive mechanics are likely very similar to the shear driven turbulence occurring across the jet that develops as flow discharges into the lake at the channel terminus. Along the developing levees, however, deposition would be restricted to narrow zones centered on the levee crests (Figure 18b cross sections C-C' and D-D').

Along the ROR, evidence of advective levee growth takes the form of crevasse splays where water spills from the channel through topographic lows incising short channels into the crest and depositing lobes of sediment along the levee flanks. The large disparities in levee-crest elevations from one side of the channel to the other and the associated disparity in the size of the levees, however, suggest another possible mechanism for levee growth. Once flooded by rising lake waters, sediment-laden inflow may rise above the height of the bounding levee crests leading to shearing and diffusive deposition. At bends, however, this elevated flow may detach from the underlying channelized flow and enter the surrounding lake waters. This mode of deposition would conceptually be similar to "flow-stripping" thought to occur in submarine channels systems [e.g. *Piper and Normark*, 1983].

Regardless of the mechanism, during inflow, the coarser fractions of the sediment load deposits onto levees forming the discrete sand and silt layers observed in channel banks. Once the lake and river water levels reach the same elevation, and inflow ceases, mud and organics will deposit on submerged levees. The thickness of the mud layers will increase with duration of inundation and with suspended-sediment concentrations of the inflowing waters. Lower elevation regions, such as the levee flanks and portions of the channel furthest from the river, will experience longer periods of inundation.

During outflow from the lake, high boundary shear stresses arising from steep hydraulic gradients scour the channel bed and help maintain the channels against plugging with sediment deposited during the waning stages of inflow (Figure 18c). Bed erosion is greatest

near the river end of the channel where high levee crests confine the outflow and steep bed gradients facilitate sediment flushing. Farther into the lake, outflow is less confined, due to lower levee crests, and water enters the channel radially over the submerged channel (Figure 18c). As lakes drain, the hydraulic gradient, depth of the outflow and, therefore, the boundary shear stress of the outflow decrease. The location of the bed sill may represent the maximum lakeward propagation of significant bed erosion. This propagation distance is set by the duration of outflows with shear stresses at or above the critical stress needed to erode the channel bed.

Bank failures, which control channel widening, may occur during or immediately following outflow events when rapidly exposed, saturated banks, experience high pore pressures and undercutting by bed erosion. In addition to undercutting, channel banks may fail due to deposition on the levee crests resulting in oversteepening and/or loading the banks beyond a threshold condition. During subsequent inflow events, overwidened sections of the channel experience deposition by sediment drapes leading to a dynamic process of widening by failure and narrowing by localized deposition. Similar cyclic widening by bank failures and narrowing by sediment draping has been reported along the River Dee in the United Kingdom [*Gurnell*, 1997a, b].

The lateral stability of tie channels likely arises in part from the cohesive nature of the banks. This cohesion arises from three major sources. First, similar to other prograding deltas that incise into fine grained, cohesive prodeltas [e.g. *Ikeda*, 1989], the incision of tie channels into their prodeltas creates resistant banks that help "fix" the channel location and hinder lateral migration. Second and third, the presence of discrete layers of cohesive mud in the channel banks and abundant vegetation further stabilize the sediments deposited in levees overlying the prodelta platform. In the case of the Birch Creek, substantial bank stability appears to be derived from frozen banks. Finally, on the ROR we observed that bank failures on one side of the channel were commonly matched by failures on the opposite bank. It appears that when one bank fails it may trigger failure of the opposite bank by deflecting flow around the failure block up against the opposing bank, undercutting it. In this manner, the channel widens about its centerline and does not continually fail on one bank and accrete on the opposite which would promote migration.

Tie channels may also lengthen if the mainstem river migrates away from the site of the original meander cutoff (Figure 18c). Eventually, channel sedimentation fills the lake to a point where the volume of water flowing into and out of the lake is insufficient to maintain a channel. If not erased by lateral migration of the main channel across the floodplain, the remaining lake will become filled with sediment and organic detritus carried by floodwaters and by runoff from the surrounding floodplain (Figure 18d). Abandoned channels on the Middle Fly River, distinguishable from the surrounding floodplain by the trees lining the former levees, do not appear to extend beyond the apex of the former oxbow lakes.

Gagliano and Howard [1984], *Saucier* [1994] and *Shepherd* [2001] presented models detailing the evolution of oxbow lakes along the Lower Mississippi River. All three models included stages of lake progression from active river meander to a terrestrial stage following complete lake sedimentation. In these models the batture channels (tie channels) play a role in the partial infilling of lakes with sediment during the "lacustrine" stage. None of these models, however, provides details or mechanics of tie channel formation or evolution.

Individual aspects of the development and morphology of tie channels, as detailed here, have similarities with aspects of other types of channels known to connect mainstem

rivers with portions of lowland floodplain systems. These types of channels include, but are not limited to: side and back channels, side and dead arms, crevasse and splay [in the sense of *Smith*, 1996], slough, or just simply floodplain channel [in the sense of *Mertes et al.*, 1996; *Dunne et al.*, 1998]. Unnamed bidriectional channels connecting distributary channels to thermokarstic lakes across the Mackenzie River Delta [*Marsh et al.*, 1999; *Hill et al.*, 2001] also have functional similarities to tie channels, though morphologic data on these channels is not available to make a detailed comparison with our field sites. Of all types of floodplain channels, however, tie channels morphologically appear to have the most similarities to the narrow, leveed, Stage III-type splay channels of *Smith* [1996] occurring in Windy Lake area along the Lower Saskatchewan River, Canada [*Smith and Perezarlucea*, 1994]. Tie channels, however, differ from these splay channels in a number of important respects. Tie channels 1) rarely bifurcate; 2) convey flow bi-directionally; 3) form coincident with lake formation rather than breaching the existing mainstem levees; 4) maintain a stable connection to the mainstem for hundreds to thousands of years; and 5) appear to extract only the suspended fraction of the mainstem river's sediment load. We note that the morphologies we observe are not universal for bidirectional floodplain channels. For example, the Pitt River connecting Pitt Lake with the Fraser River in British Columbia, Canada is a bidirectional channel but has a width to depth ratio of 50 and is approximately the same width as the Fraser River [*Ashley*, 1980]. These differences from our observations, may in part arise from the fact that flow into the Pitt from the Fraser is driven by ocean tides and that large quantities of bedload sediment are transported from the Fraser into the Pitt.

3.2. Controls on Channel Geometry

That scale-independent, self-similar channel geometries should arise from jet sedimentation is perhaps not surprising given the fundamental self-preserving hydrodynamic structure of jets [e.g. *Albertson et al.*, 1950; *Bates*, 1953; *Abramovich*, 1963; *Schlichting*, 1968]. Jet self-similarity, however, is scaled by the geometry of the jet orifice (in natural systems the channel geometry at its terminus); this geometry sets the dimensions of the jet, not the reverse. To understand the observed self-similar geometry of tie channels, therefore, it is necessary to understand what controls the geometry of the channel upstream of the channel terminus where jet-like conditions do not prevail.

The cross-sectional geometry of tie channels bear a strong similarity to cosine channel geometry derived from threshold channel theory [for review see *ASCE Task Committee on Hydraulics, B.M. and Modeling of River Width Adjustment*, 1998]. Tractive force models posit that equilibrium channels evolve to a condition where the boundary shear stresses result in a condition of impending motion for grains on the channel perimeter [*ASCE Task Committee on Hydraulics, B.M. and Modeling of River Width Adjustment*, 1998]. Both the apparent dominance of mass failures as a channel width control and shear stresses in excess of that needed for suspension in tie channels, however, suggest a tractive force theory is not applicable to these systems.

Similar to tie channels, tidal inlets experience bi-directional flow and fill and drain closed bodies of water. Tidal inlets have been observed to display a relationship between equilibrium inlet cross-sectional area and the size of the tidal prism [e.g. *O'Brien*, 1969; *Jarrett*, 1976; *Escoffier*, 1977]. The tidal prism is the volume of water exchanged with a coastal body of water through the inlet during the ebb and flood of the tide. An analogous volume for tie channels would be the lake area times the stage range experienced by the

channel. To calculate an equivalent prism for tie channels, we used the height between the channel sill and maximum levee height to estimate stage range, and the measnred surface areas of lakes. Variation in these calculated volumes, however, show no clear relationship with channel cross-sectional area for channels on the Fly River, the Birch Creek or the ROR.

The ubiquitous presence of mass failure blocks along all tie channels studied here intuitively suggests a linkage between the characteristic channel geometries observed and controls on mass failure processes. We explore this possible linkage using a simple planar failure approach for predicting the maximum stable bank height [*Lohnes and Handy*, 1968]:

$$H_c = \frac{4C\sin\theta\cos\phi}{\gamma(1-\cos(\theta-\phi))}, \tag{1}$$

where C is cohesion, g is gravity, θ is the bank slope, ϕ is the friction angle, and γ is the unit weight of the bank materials. Equation 1 predicts a gentler bank slope for increasing bank heights if cohesion and friction angle are held constant. The observed changes in bank slope and channel depth observed along the Fly River channels (Figure 8) suggest that a simple bank stability model (e.g. equation 1) provides a first order explanation of observed tie channel geometries. Additionally, for a V-shaped channel where $\tan\theta$ equals the maximum depth divided by half the channel width, the width to depth ratio of the channel may be expressed as $2/\tan\theta$. Given this formulation, observed tie channel-bank slope values of 15 to 45 degrees correspond to width to depth ratio of 8 to 2, respectively.

Bank stability constraints may be the first order control on channel geometry and the along channel trends in geometry. These constraints alone do not explain, however, the central tendency for tie channel inlet width to be about ten percent of the mainstem channel width. This scaling may be set by the maximum depth of the tie channel, which will in turn be constrained by the stage range of the mainstem river.

The maximum tie channel levee elevation is that of the mainstem river bank and the minimum bed–elevation is set by the low–flow stage of the mainstem, with this low–flow stage serving as the effective base–level of the tie channel. At low stages, flow velocities through tie channels are greatly diminished and back water deposition from the mainstem into the tie channel may be favored; the lower the mean low stage of the mainstem, the lower the tie channel can cut before these back water effects take effect. On the Lower Mississippi River, the mean range in stage of 11.5 m (measured between 1851 and 2004 at Tarbert Landing - data source US Army Corps of Engineers) is close to the 12.5 m maximum depth measured along the ROR tie channel. On the Middle Fly River the average stage range observed in the lower half of the Middle Fly River is also similar to the mean maximum–depth of the tie channels. We lack stage data for the Birch Creek to make a similar comparison on that river system.

If river stage range controls tie channel depth and hence channel width (via slope stability, i.e. $width = 2H_c/\tan\theta$), why does mainstem river width provide a functional length scale for normalization? The answer appears to relate to a rough scaling between river width, depth and stage range in these lowland systems. If tie channel width (W_{tc}) equals $2H_c/\tan\theta$ and the mainstem river width (W_m) equals $H\beta$, where H is the mainstem mean bankfull depth and β the mainstem width to depth ratio, then:

$$\frac{W_{tc}}{W_m} = \frac{2H_c}{\beta H \tan\theta}. \tag{2}$$

The Middle Fly River has an average width to depth ratio of 20 and the Lower Mississippi River in the vicinity of the ROR has a width to depth ratio of 60. For the Middle Fly River an H_c of 7 m results in a relative tie channel to mainstem width of 10 percent for a θ of about 28 degrees, a value close of the mean observed bank slope of 26 degrees. On the ROR, to match the observed nine percent tie channel to mainstem width ratio requires a θ of about 15 degrees, slightly less than but close to the 16 degrees observed near the ROR tie channel confluence with the Lower Mississippi River.

While these proposed relationships offer a promising first–cut explanation for the observed tie channel scaling and geometry, refinement and testing of interrelationships between river stage, river width and the dynamics of tie channel geometry would greatly benefit from additional data from multiple river systems. Additionally, while we do observe (in aerial photographs and satellite imagery) very similar channel planforms and width scaling between tie channels and their mainstem rivers for many of the river systems highlighted in Figure 2, to those channels discussed here, greater insights into tie channel morphodynamics may be gained by studying exceptional tie channels, such as channels that show bifurcations, carry bedload or exhibit some degree of lateral mobility. In the absence of complete numerical or physical models of tie channels, exploring the range of these systems in nature offers the only opportunity to isolate and test the processes that control the morphodynamics of these channels.

4. Conclusions

Field studies of tie channels on the Middle Fly River, the Lower Mississippi River and the Birch Creek (Alaska) indicate that channels in each of these systems display morphologically similar forms in both planform and cross section. The common morphology of a narrow, leveed, V-shaped channel, that scales in maximum width to about 10 percent of the mainstem width arises from a common set of processes that appear to operate in largely a scale-independent manner. Channels initially form as flow into and out of cutoff meanders become restricted by sediment sealing the cutoff from the river. The restricted flow creates a sediment-laden jet which leads to subaqueous levee formation along the jet margins. As the levees grow, become emergent and are colonized by vegetation, the depositional mechanics governing levee formation transitions from those of a partially confined jet to those of channelized flow depositing sediment under overbank conditions. In both depositional conditions, shearing between still, sediment-poor lake waters and inflowing sediment-rich waters help to drive the transfer of sediment to the flow margins. Along the developing channels, overbank processes such as crevasse-splays and detachment of superelevated flow at bends appear to also result deposition down the levees flanks.

Widespread bank failures along the length of channels in all three river systems suggest that the shape and size of the channels reflect a dynamic relationship between channel scour, widening due to slumping, and narrowing due to bank sedimentation from suspended sediment deposition. The material properties of the channel banks (effective strength) and the stage variation of the mainstem river both influence geometry. The mainstem river stage sets both the maximum height to which channel levees may aggrade and the base level to which channel beds may grade. These two extremes control the maximum channel depth, material strength (friction angle and cohesion) then sets channel-bank slope and width. In these lowland systems, river stage appears to roughly scale with mean river bankfull depth, which in turn scales with river width, giving rise to the observed ten percent scaling between maximum tie channel width and river width.

As tie channel levee elevations decrease with distance from the mainstem river the depth of the channels decrease and there is a corresponding increase in channel slopes and narrowing of the channels.

Based on our estimates from both the Middle Fly and the Lower Mississippi River, sedimentation from tie channels diverts a small but significant (2 to 4%) fraction of the total river load into oxbow lakes. Given the widespread geographic distribution and long term stability of tie channels they likely play critical but largely overlooked roles in the ecological health of those lowland river systems that have them. In form and evolution, tie channels represent simple deltaic channels. The morphodynamics of these channels reported here, therefore, provides general insight about the form and evolution of all channels formed by sediment-laden flows entering still water.

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Figure 1. Tie channels. An * marks the channel junction with the river and # indicate the channel terminus in the lake. **a)** The Raccourci Old River channel located approximately 65 km north of Baton Rouge, Louisiana, USA. The Mississippi River flows from the left to the right across the upper portion of the image. Approximately half the oxbow lake is shown. Image is a false color aerial photograph, 2004 U.S. Geological Survey Digital Ortho Quarter Quadrangles (DOQQs) source: http://atlas.lsu.edu/. **b)** Enlargement of sediment-laden jet entering the lake at the Raccourci Old River channel. **c)** Fly River, Papua New Guinea (PNG), (1998 aerial photograph courtesy of Ok Tedi Mining LTD). **d)** Channel on the Birch Creek, a small tributary to the Yukon River in Alaska, USA, (June, 2002). **e)** Fly River blocked valley lake known as Bai Lagoon. (1992 aerial photograph courtesy of Ok Tedi Mining LTD). **f)** Channel on the Herbert River in PNG which connects the Strickland River to Lake Murray. Each limb of the lake is approximately 200 m wide (2003).

Figure 2. Map showing the distribution of river systems with known tie channels. Rivers identified from maps, aerial photographs, satellite imagery and published descriptions of floodplains. The inventory is considered illustrative rather than exhaustive. Three enlargements show locations of study sites for this work.

Figure 3. Meander bend cut off and development of a tie channel. Series of images of a bend located at river mile 380 (90 km downstream of D'Albertis Junction) on the Fly River, Papua New Guinea. **a)** 1988 Landsat 5 image showing meander bend prior to cut off in 2000. Flow in channel goes from left to right. **b)** 2000 Landsat 7 image of the bend shortly after meander neck breach and cut off. Sediment-laden water enters the lake through the downstream meander limb resulting in channel formation seen in **c)** the 2004

Landsat image. **d)** 2003 photo taken from the river looking into the lake with the nascent channel marked. Images a and b obtained from http://www.landsat.org/ortho/index.htm. Image c supplied courtesy of Geoff Pickup and Ok Tedi Mining LTD.

Figure 4. Long profile surveys of the bed (dashed line with triangles) and levees (solid lines) of a Fly River tie channel. The channel connects to an open oxbow lake. Elevations are reported relative to mean sea level.

Figure 5. Two cross-sectional surveys of the Pangua tie channel (8 river miles above Everill Junction) located 127 and 910 m from the channel confluence with the Fly River. The abrupt steps in the bank slopes represent down-dropped slump blocks.

Figure 6. Photograph of a Fly River channel highlighting the role of vegetation in this system. The levee crests lie several meters beyond the edge of the vegetation on either side. The total channel width at this location is approximately 15 m.

Figure 7. Plots of normalized channel width (W^*) versus normalized channel length (D^*) plotted by lake type. Width is normalized by the average local river width and the channel length is normalized by the total surveyed length of channel. **a)** Closed oxbow lakes (21 measurements from 4 channels). Best fit linear regression W^*=0.1-0.062D^*, r^2=0.88, slope $p<0.001$. **b)** Blocked Valley lakes (10 measurements from 4 channels). Best fit linear regression W^*=0.141-0.060D^*, $r^2 = 0.45$, slope p=0.03 **c)** Open oxbow lakes (32 measurements from 4 channels). Best fit linear regression W^*= 0.104-0.009D^*, r^2=0.018, $p<0.4$.

Figure 8. Channel-bank slopes (a) and depth (b) measured at 16 cross sections along 9 Fly River tie channels. Bank slopes (shown normalized by total channel length) increase

with distance from the river (Slope = 6.32x+24.6, r^2=0.14) and depths decrease (Depth = -2.25x+6.5, r^2=0.36).

Figure 9. Grain-size distributions from the levees of a tie channel and the suspended and bed material loads of the Middle Fly River at a location ~145 km downstream of D'Albertis Junction. Levee distribution is the average of 11 individual samples collected at 250 m intervals along the channel, the river suspended sediment is the flow weighted average at a cross section (Day unpublished data) and the bed sample is a single grab sample from the middle of the river.

Figure 10. Long profile surveys of the bed (dashed line with triangles) and levees (dashed and solid lines) of the ROR tie channel. Elevations are reported relative to mean sea level.

Figure 11. Two cross-sectional surveys of the ROR channel located 1 and 7.8 km from the channel confluence with the Lower Mississippi River. The abrupt step in the left bank of the 1 km survey is a down-dropped failure block. Cross sections are a composite of field surveys of between the levee crests and LiDAR data on the levee flanks.

Figure 12. Plots of normalized channel width (W^*) versus normalized channel length (D^*) for the ROR channel. Width is normalized by the average local river width and the channel length is normalized by the total channel length. Best fit linear regression W^*=0.093-0.051D^*, r^2=0.67, slope $p<0.001$. Channel width measured from peak levee-crest elevations on opposite channel banks derived from LiDAR based DEM.

Figure 13. Section of the ROR channel 1 km from the junction with the Mississippi River. The channel is approximately 95 m wide at this location. The vertical scar in the foreground and the back rotated tree stumps in the channel mark failure blocks.

Figure 14. Grain-size distribution for the ROR channel levees (coarse layers) and mouthbar compared to distributions of the Lower Mississippi River suspended and bed load fractions. Levee distribution represents the average of 10 individual samples and the mouthbar 2 samples. Mississippi distributions measured at Tarbert Landing averaged from data collected between 1974 and 1991 at river stages >9 m [*Catalyst Old River Hydroelectric Limited Partnership*, 1999]. Size distributions of the river suspended load undetermined for particle sizes < 3.9 μm.

Figure 15. Long profile surveys of the bed (dashed line with triangles) and levees (solid line) of the Birch Creek tie channel. Elevations are reported relative to an arbitrary datum.

Figure 16. Three cross-sectional surveys from the 500 m long Birch Creek channel shown in Figures 1d and 15. The surveys are located 135, 450, and 495 m from the channel confluence with the creek. Elevations are relative to lake elevation at the time of the survey.

Figure 17. Section of the Birch Creek tie channel approximately 200 m from the channel outlet into the lake. Vegetated slump blocks are visible on both sides of the channel along the bottom left side of photograph.

Figure 18. Conceptual model of channel evolution in a closed oxbow lake. **a)** Formation of the channel occurs coincident with the sealing of the cutoff meander bend and creation of an oxbow lake. **b)** Continued deposition at the river junction leads to the complete sealing of the oxbow lake from the river (A-A') and creation of a sediment laden jet of water. At the terminus (D-D') of the channel, subaqueous levees form where coarse sediment rapidly settles out of suspension along the margins of the jet and a small mouth

bar forms along the channel centerline. **c**) during outflow from the lake, sediment-poor water spills into the channel over inundated levees (C-C'). In the channel, outflow scours the channel bed, removes sediment deposited during the waning stages of inflow, undercuts the banks and triggers bank failures. **d**) Once the channel ceases to function, the lake slowly fills with mud and organic detritus delivered by floods and runoff from the surrounding floodplain.

Figure 1.

Figure 2.

Figure 3.

Figure 4.

Figure 5.

Figure 6.

Figure 7.

Figure 8.

Figure 9.

Figure 10.

Figure 11.

Figure 12.

Figure 13.

Figure 14.

Figure 15.

Figure 16.

Figure 17.

Figure 18.

Table 1. Mean Morphological Properties

River System	W/D ± Std (n)	Bank Slope (deg) ± Std (n)	Levee Slope† ± Std (n)	Sill Location‡
Fly River	5.3±1.6(16)	27±8(60)	$5\text{x}10^{-4}$	0.4±0.2(6)
ROR	5.8±1.8(5)	25±8(10)	$5\text{x}10^{-4}$	0.8
Birch Creek	8.7±8.8(11)§	23±11(4)	$5\text{x}10^{-3}$	0.2

W/D = width to depth ratio

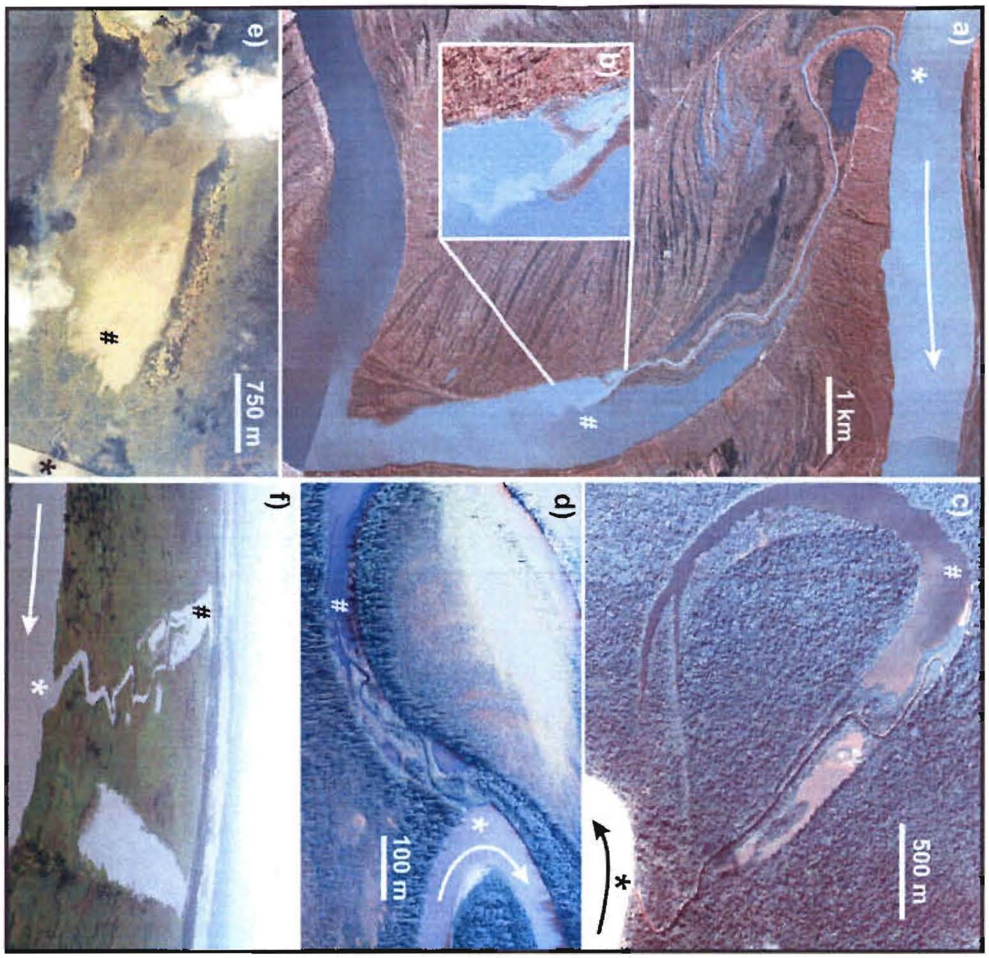
std = standard deviation

n = number of measurements

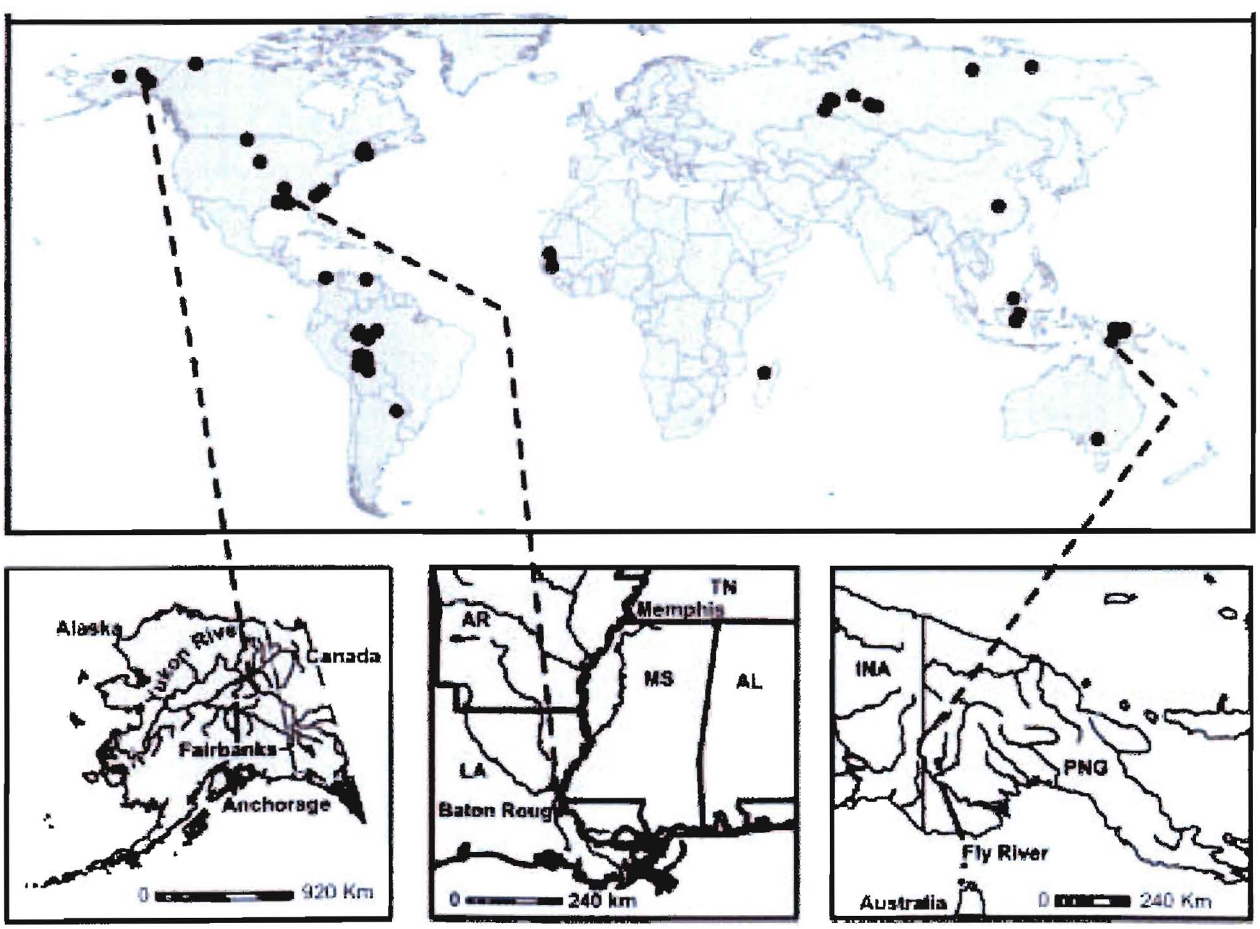
† slope of the levee crest along majority of tie channel length

‡ sill distance from river/total tie channel length

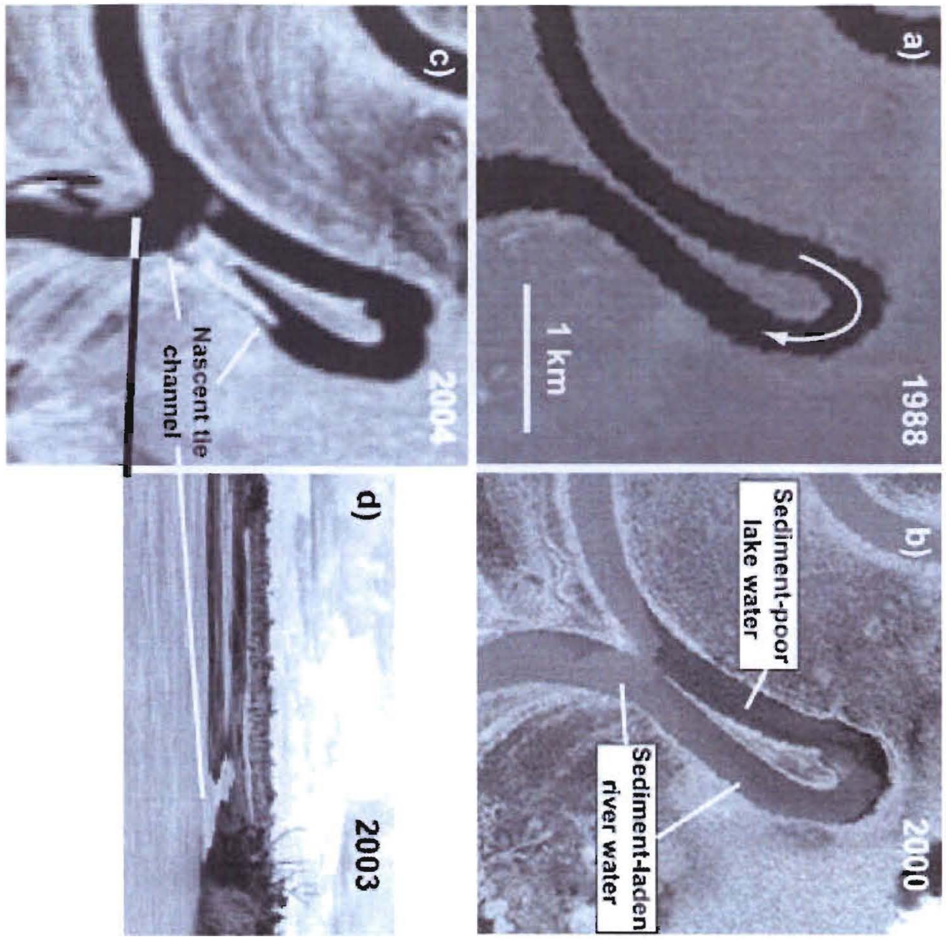
§ includes measurements on newly forming tie channel and measurements at the distal most end of the well developed tie channel



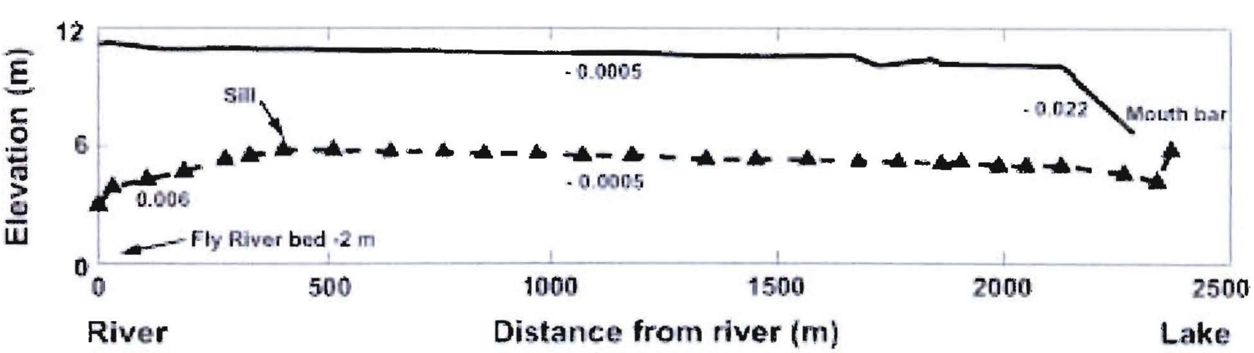
a)
b)
c)
d)
e)
f)
1 km
500 m
100 m
750 m



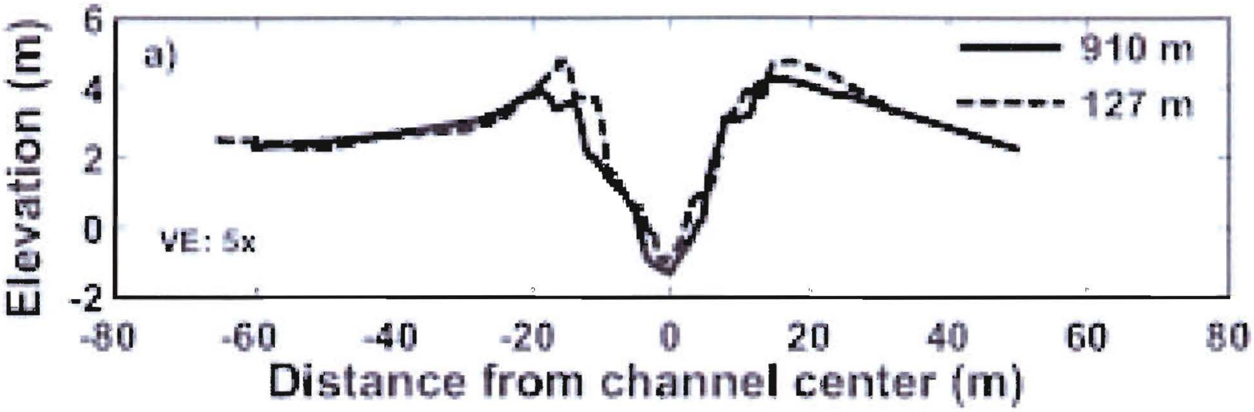
Alaska
Yukon River
Canada
Fairbanks
Anchorage
0 920 Km
AR
TN
Memphis
MS
AL
LA
Baton Roug
0 240 km
INA
PNG
Fly River
Australia
0 240 Km

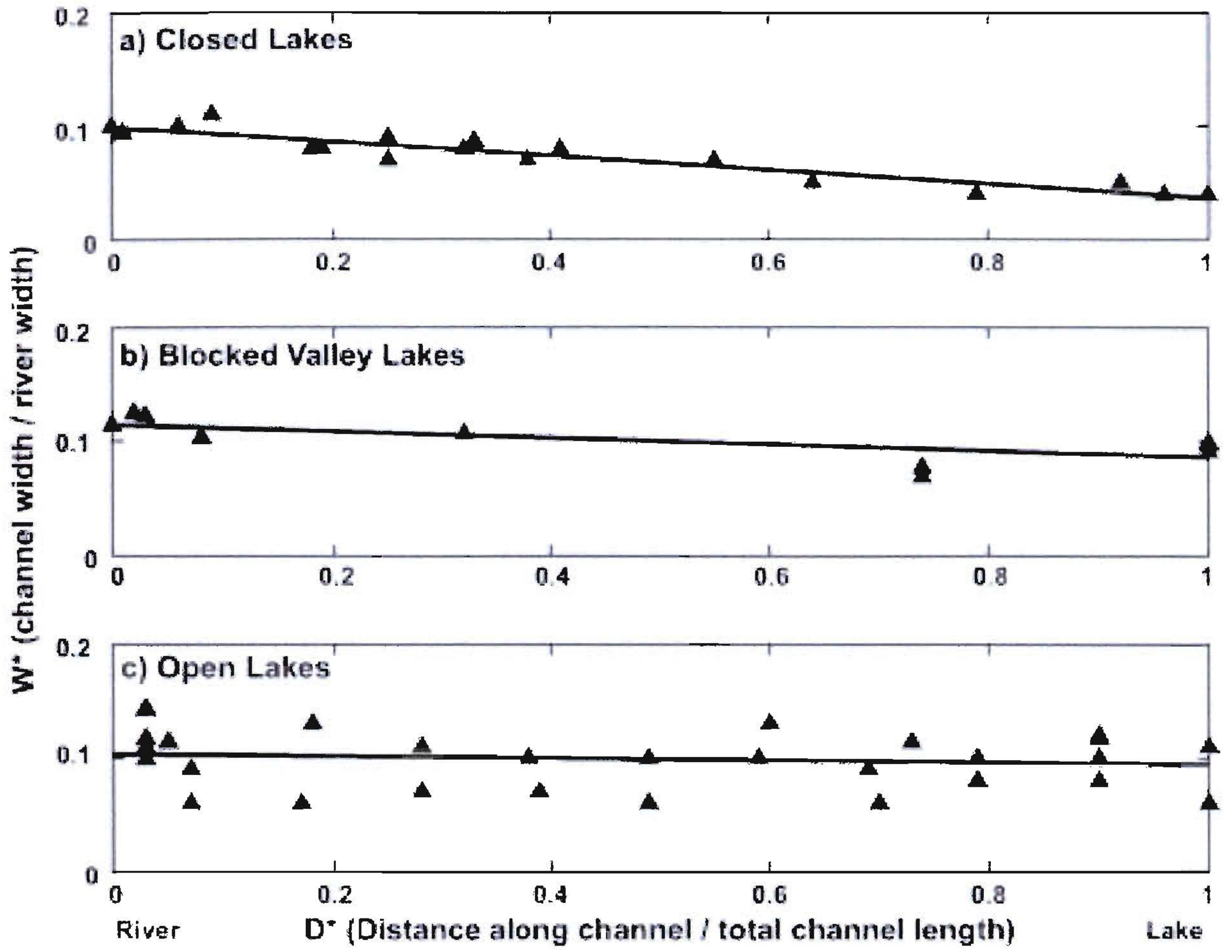


a)
1988
1 km
b)
2000
Sediment-poor lake water
Sediment-laden river water
c)
2004
Nascent tie channel
d)
2003

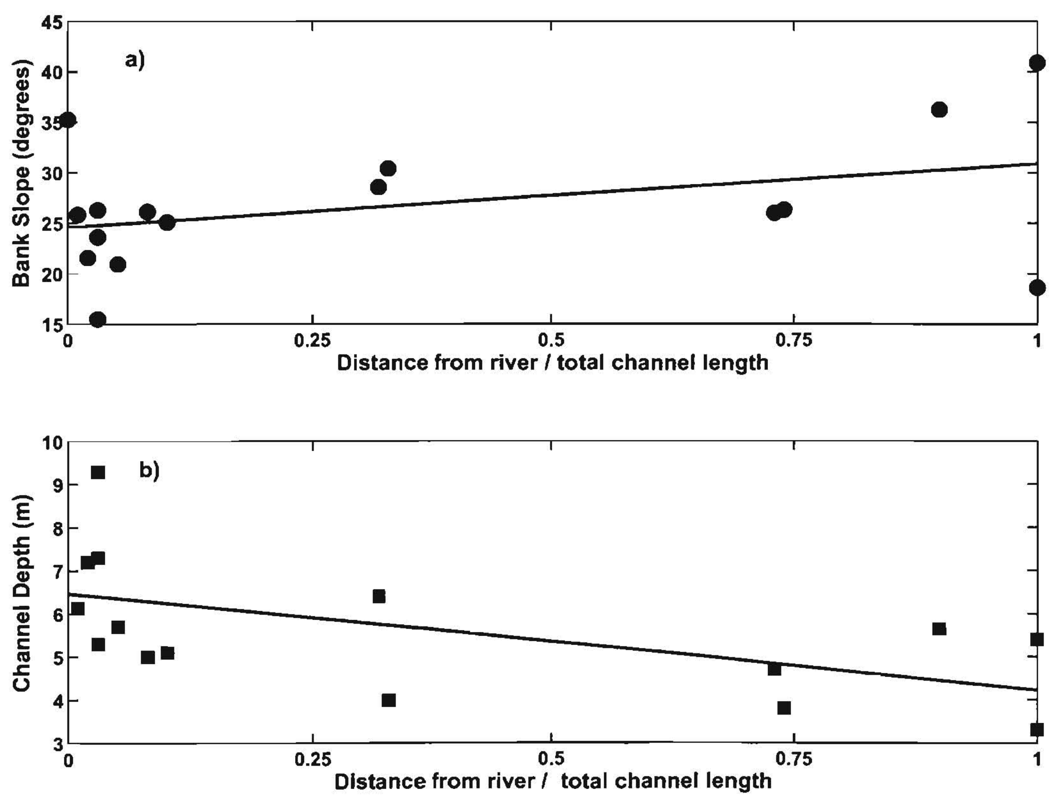


Elevation (m)
12
6
0
Sill
- 0.0005
- 0.022
Mouth bar
0.006
- 0.0005
Fly River bed -2 m
0
500
1000
1500
2000
2500
River
Distance from river (m)
Lake

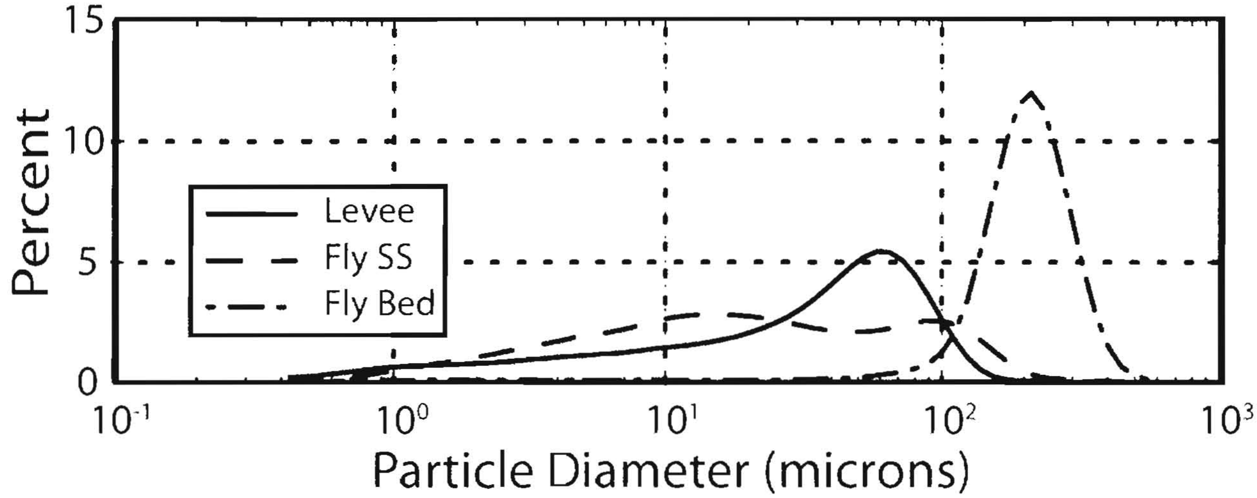

a)
910 m
127 m
VE: 5x
Elevation (m)
Distance from channel center (m)
6
4
2
0
-2
-80
-60
-40
-20
0
20
40
60
80

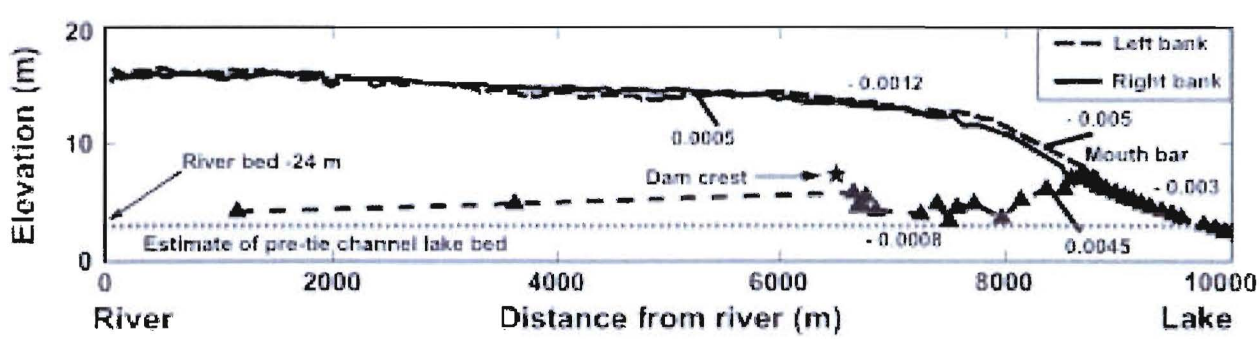


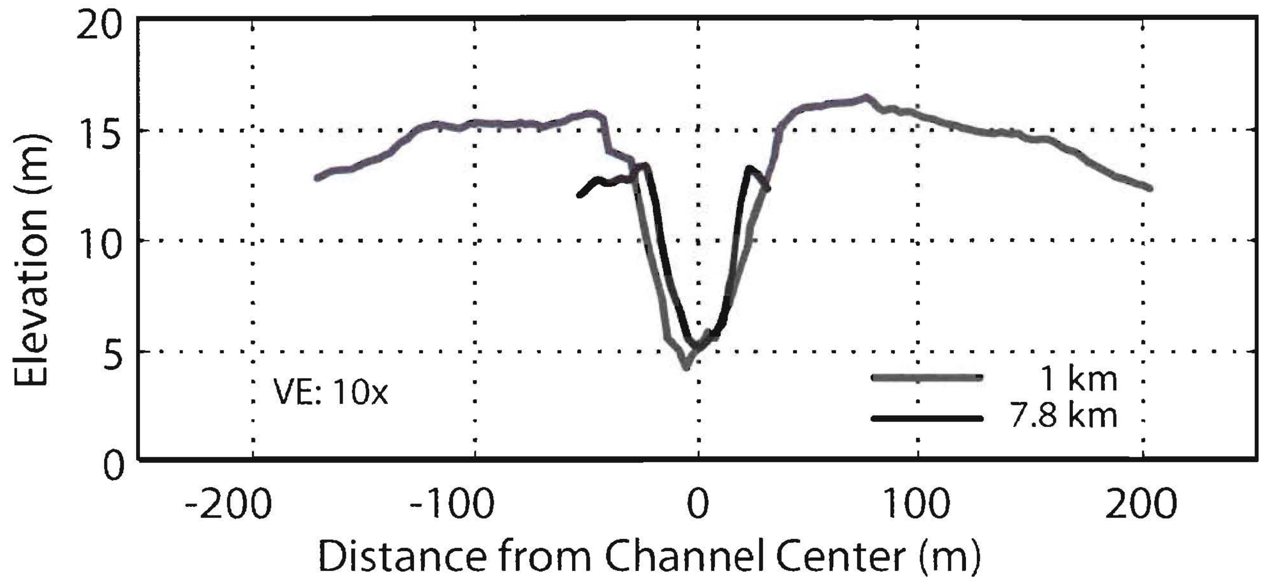
a) Closed Lakes
b) Blocked Valley Lakes
c) Open Lakes
W* (channel width / river width)
0.2
0.1
0
0
0.2
0.4
0.6
0.8
1
River
D* (Distance along channel / total channel length)
Lake



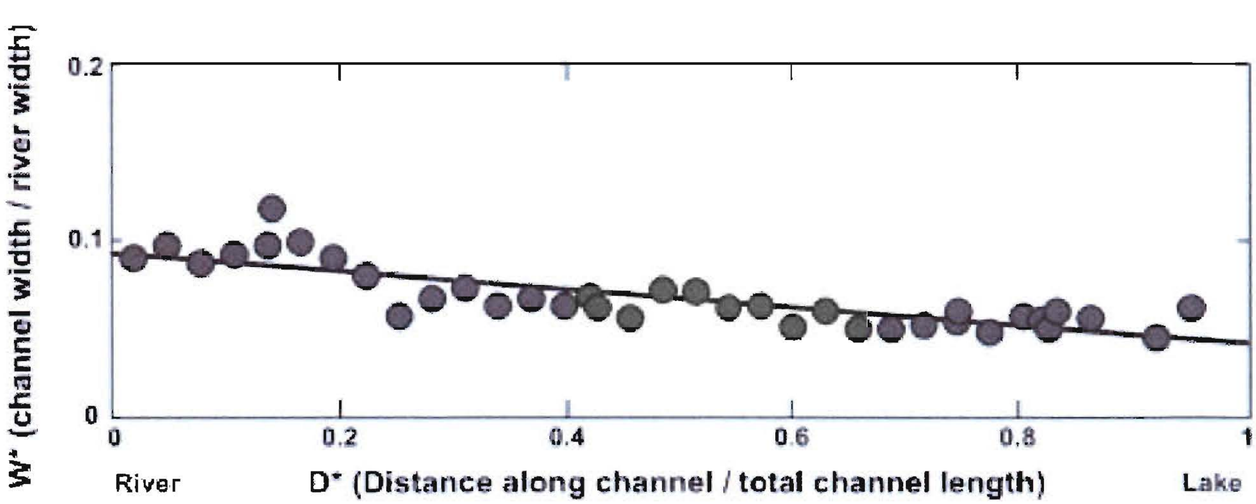
a)
Bank Slope (degrees)
45
40
35
30
25
20
15
0
0.25
0.5
0.75
1
Distance from river / total channel length
b)
Channel Depth (m)
10
9
8
7
6
5
4
3
0
0.25
0.5
0.75
1
Distance from river / total channel length


Percent
Particle Diameter (microns)
Levee
Fly SS
Fly Bed
15
10
5
0
10^{-1}
10^{0}
10^{1}
10^{2}
10^{3}

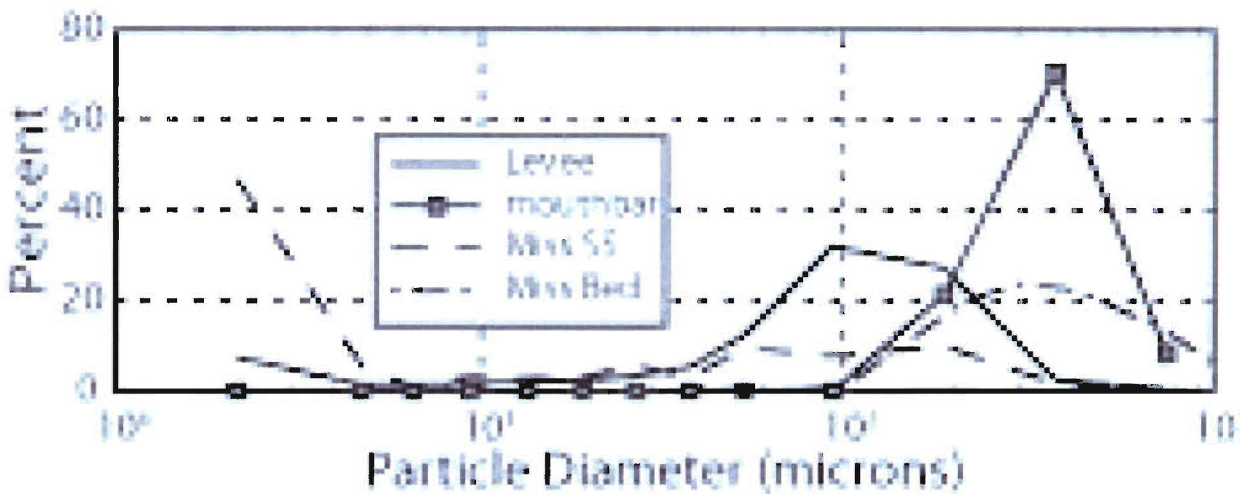

Elevation (m)
Left bank
Right bank
- 0.0012
0.0005
- 0.005
Mouth bar
- 0.003
River bed -24 m
Dam crest
Estimate of pre-tie channel lake bed
- 0.0008
0.0045
20
10
0
0
2000
4000
6000
8000
10000
River
Distance from river (m)
Lake



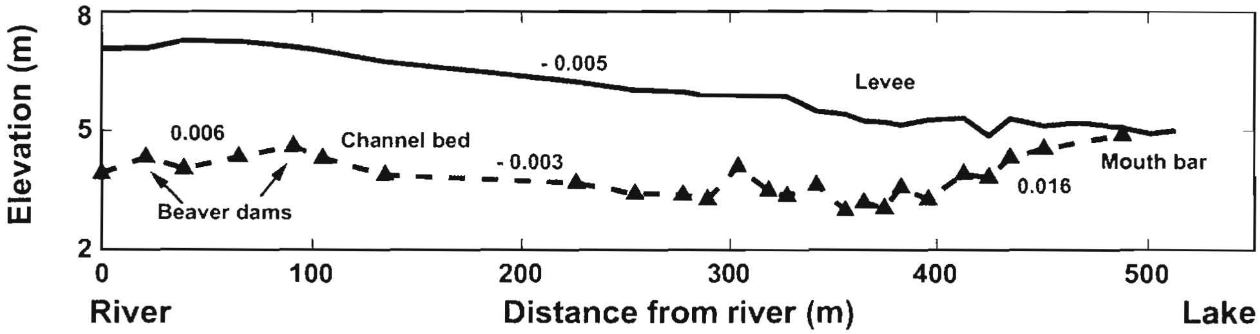
Elevation (m)
Distance from Channel Center (m)
VE: 10x
1 km
7.8 km



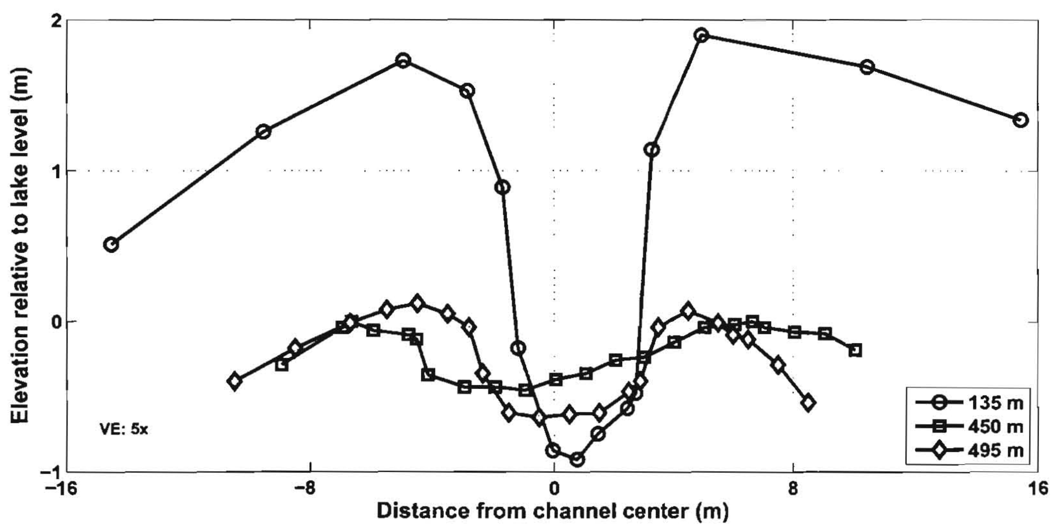
W* (channel width / river width)
D* (Distance along channel / total channel length)
River
Lake



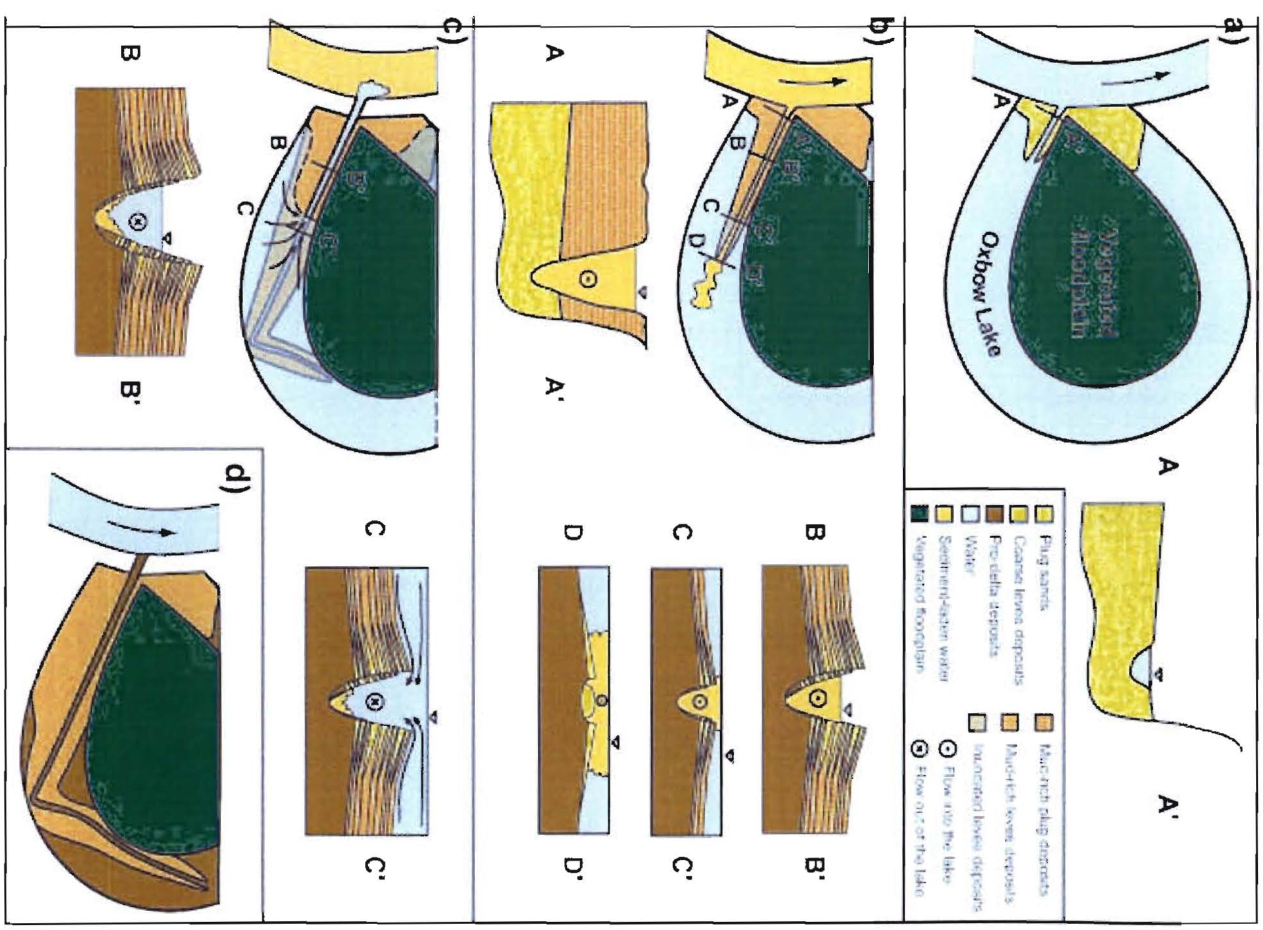
Percent
80
60
40
20
0
Levee
mouthbar
Miss SS
Miss Bed
10⁰
10¹
10¹
10²
Particle Diameter (microns)



Elevation (m)
8
5
2
- 0.005
Levee
0.006
Channel bed
- 0.003
Beaver dams
Mouth bar
0.016
0
100
200
300
400
500
River
Distance from river (m)
Lake



Elevation relative to lake level (m)
2
1
0
−1
−16
−8
0
8
16
Distance from channel center (m)
VE: 5x
135 m
450 m
495 m



a)
Oxbow Lake
Vegetated floodplain
A
A'
b)
c)
d)
B
B'
C
C'
D
D'
Plug sands
Coarse levee deposits
Pro-delta deposits
Water
Sediment-laden water
Vegetated floodplain
Mud-rich plug deposits
Mud-rich levee deposits
Incipient levee deposits
Flow into the lake
Flow out of the lake