Sediment load and floodplain deposition rates: Comparison of the Fly and Strickland rivers, Papua New Guinea

Kathleen M. Swanson,1 Elizabeth Watson,2 Rolf Aalto,3 J. Wesley Lauer,4 Marie T. Bera,5 Andrew Marshall,6 Mark P. Taylor,2 Simon C. Apte,7 and William E. Dietrich8

Received 6 July 2006; revised 13 June 2007; accepted 4 September 2007; published 29 March 2008.

Rates of aggradation and infilling of accommodation space along lowland channels in response to postglacial sea level rise should depend on sediment supply. The Strickland and Fly rivers join at just 6 m above sea level and have experienced the same Holocene sea level rise. Historically, the Strickland has carried about 7 times the sediment load and 1.4 times the water discharge as the Fly. Therefore we hypothesize that the lowland Strickland floodplain should be more developed and consequently should presently be capturing proportionately less sediment than the floodplain of the lowland Fly River. We use mine-derived elevated Pb and Ag concentrations in 111 shallow (<1 m) floodplain cores collected in 2003 to determine deposition rates across the lower Strickland floodplain. Sediment deposition rates decrease across the floodplain with distance from the channel bank, and the average rate of deposition is 1.4 cm/a over the first 1 km. Overbank deposition along the lowland sandbedded Strickland results in ~13% loss of the total load, ~0.05%/km of channel length of the main stem. Deposition rates over the first 1 km from the channel bank on the Strickland are about 10 times those on the Fly (for estimated natural sediment loads); however, the proportional loss per channel length on the Strickland is less than that on the Fly (0.09%/km of main stem channel length) because of an extensive network of tributary and tie channels that convey sediment to the floodplain on the Fly. Furthermore, the lateral migration of the Strickland channel is ~5 times that on the Fly, such that most overbank deposits on the Strickland are returned to the channel, causing the net loss of sediment to the floodplain to be small. We conclude that the Strickland River, which has a much higher overbank deposition rate than the Fly River as a function of distance from channel bank, nonetheless has significantly less net accumulation than the Fly because (1) a large proportion of the sediment load is conveyed up tributaries and tie channels on the Fly, and (2) lateral migration (which presumably results from the higher load) on the Strickland sweeps sediment back into the channel. Hence our field observations support our initial hypothesis, though the primary reason for this is because of active lateral migration rather than low overbank deposition rates.


1. Introduction

All large lowland rivers entering the sea are in various states of response to postglacial maximum sea level rise. The sediment accommodation space created by low sea stand incision and then rapid sea level rise tends to cause initially high net loss of sediment to the floodplain, but as the river rebuilds its slope and infills adjacent lowlands, the loss rate should decline [Muto and Swenson, 2005]. The timescale over which a river shifts from significant net loss of sediment to primarily downstream sediment transfer should depend on the relative sediment supply and on rates of local subsidence and sea level, or base level, rise [Paola, 2000]. This timescale is of fundamental significance to the offshore record because there is a time delay of sediment delivery to the marine environment. As the accommodation...
space infills, the depositional signal propagates through the fluvial system and could persist for more than 10,000 years [e.g., Pitman, 1978; Angevine, 1989]. River morphology may also still be evolving in response to sea level change [e.g., Aslan and Autin, 1999; Tornqvist, 1993; Latrubesse and Franzinelli, 2005].

The Fly River in Papua New Guinea provides an accessible fluvial system where the influence of sediment supply on river morphodynamics and net sediment storage response to rising base level may be studied. The middle Fly and lower Strickland rivers join at Everill Junction, 400 km upstream of the delta front (all distances are streamwise unless otherwise noted, Figure 1). The Strickland drains twice the area and carries ~7 times the natural sediment load of the Fly River where they join. In this paper, we hypothesize that morphological differences between the rivers are the result of sediment supply-driven differences in response to sea level rise, and that the lower Strickland River, which is presumably more fully adjusted to modern sea level, should lose a much smaller proportion of its annual load to the floodplain than does the middle Fly.

Here, we briefly describe the morphologic differences between the Fly and Strickland channels and adjacent floodplains and then quantify a sediment deposition rate for 10 years on the Strickland (1993–2003) using mine-derived tracers. These deposition rates are similar to those reported in a companion paper in which 

\[ {^{210}}Pb \] measurements provide timing of depositional events [Aalto et al., 2008]. Our rates differ significantly from those reported by Day et al. [2008] for the middle Fly. Unexpectedly, we find that the average deposition rate for the first kilometer across the floodplain from the channel bank is much higher on the Strickland. High lateral migration rates (presumably due to the high sediment load) on the lower Strickland cause a relatively rapid return of this sediment to the channel, resulting in much lower net sediment deposition on the floodplain.

2. Study Site

As shown in Figure 1, the Strickland and Fly rivers drain the Southern Fold Mountains of Papua New Guinea, an area of rapid uplift [Pickup, 1984, Hill et al., 2002]. They transport gravel down to broad lowlands, where the rivers transition to sand-dominated beds and a meandering channel planform across floodplains (<10 km wide) composed of elevated scroll bar topography bordered by backswamps. Where the rivers join at Everill Junction (bankfull elevation ~6 m above sea level), the median diameter of the Strickland River bed material is ~0.2 mm [Bera et al., 2005] and ~0.1 mm on the Fly [Dietrich et al., 1999]. The lower Fly here is tidally influenced and flows to its delta, which terminates 400 km downstream. The Strickland drains 36,000 km², as compared to 18,400 km² for the Fly above Everill Junction. A greater proportion of the larger Strickland drainage is in the steep headwaters, and consequently, the sediment load is much greater than that delivered to the Fly. The natural load of the Fly is estimated to be just 10 Mt/a, compared to 70–80 Mt/a on the Strickland [Dietrich et al., 1999]. Mean annual flow just above Everill Junction is 2244 m³/s on the Fly and 3100 m³/s on the Strickland [Dietrich et al., 1999]. This difference is smaller than that of the drainage area because the headwaters of the Fly are exceptionally wet, often exceeding 10 m/a of rainfall. The middle Fly channel length is ~420 km with another ~70 km of river channel up the Ok Tedi to the gravel-sand transition. On the Strickland, the gravel-sand transition lies ~250 km upstream of Everill Junction. Hence the potential sediment trap in the sandbed reach is nearly twice as long on the Fly as on the Strickland. Surprisingly, despite the Fly system crossing a major region of tectonic activity, there is little evidence of significant uplift or subsidence in the reaches studied. On the Fly and Strickland, pre-Holocene terraces (with bright red sediment) occur at relatively low heights above the current floodplain. For more details about the geology, hydrology and geomorphology of the Fly system, see Dietrich et al. [1999].

The lower Strickland River is much steeper than the middle Fly River. Figure 2 shows estimates of channel longitudinal profiles for both systems from Shuttle Radar Topographic Mission (SRTM), provided by NASA. The middle Fly profile, which has also been surveyed using differential Global Positioning System (GPS), is concave up, dropping from 6.6 × 10⁻⁵ upstream to 1 × 10⁻⁵ at Everill Junction. The profile of the Strickland is much steeper, with the slope of the lower sandbedded reach averaging 1 × 10⁻⁴. Importantly, there are well-defined bars, commonly island bars, in the bends on the Strickland, whereas on the middle Fly there were no bars for the lower 170 km. On average, the Strickland shifts 5 m/a laterally.
Junction, is 1 reach of the Strickland, to 250-km upstream of Everill from 90-m grid SRTM data. The mean slope of the sandbed 1995. These data agree well with the 5-km averages of with a differential GPS survey on the Fly performed in Figure 2.

F01S03
SWANSON ET AL.: SEDIMENT LOAD AND FLOODPLAIN DEPOSITION

Figure 2. Longitudinal profiles of the Fly and Strickland rivers derived from SRTM data. The Fly data are compared with a differential GPS survey on the Fly performed in 1995. These data agree well with the 5-km averages of elevation along the Fly, derived from 90-m grid SRTM data. The Strickland data are also 5-km averages of elevation from 90-m grid SRTM data. The mean slope of the sandbed reach of the Strickland, to 250-km upstream of Everill Junction, is $1 \times 10^{-4}$.

[1] One of the more striking features of the Strickland is the Holocene scroll bar complex that forms an elevated ridge ~6 km wide surrounded by swamps across the floodplain. This feature is evident in elevation SRTM data shown in Figure 3, though it is possible that the ridge is accentuated in this data because of the tendency of trees to grow on the elevated, better drained scroll bar complex. A detailed view of the floodplain ridge and scroll bar complex is shown in Figure 4. During 1998 and 2000 floods, flow spilled from the Strickland through an oxbow lake toward the north and traveled west along a tributary channel, the Mamboi River, into Lake Murray (Figures 1 and 4). The Mamboi River has remained an intermittent connection since that time [Porgera Joint Venture, 2003]. The resulting extensive area of cutting and deposition associated with this process may foreshadow an eventual avulsion.

3. Determination of Floodplain Sediment Deposition Rate Using Anthropogenic Tracers

[2] In 1991, tailings and rock waste from the Porgera gold mine located in the headwaters of the Strickland began entering the system, introducing sediment with a variety of elevated metal concentrations [Apte, 2001]. Silver and lead, the chosen tracers for this study, are ideal because of their postdepositional stability [Apte, 1995, 1996; Watson, 2006] and their presence at significantly elevated concentrations in mine-derived sediments discharged to the river system [Apte, 1995]. Particulate Pb and Ag strongly covary both in suspended and floodplain sediments [Watson, 2006]; hence measuring both elements serves as a data quality check. On the Strickland, Pb and Ag concentrations are only weakly influenced by the relative amount of organic matter in the sample, and grain size has a minor influence [Watson, 2006]. The mine-derived tailings are mostly silt and clay (~63 μm), and the strongest Pb-Ag relationship was found in sizes <63 μm. Only this size fraction was analyzed as a tracer in this study. There may be a tendency for samples rich in clay to be characterized by higher tracer concentrations, but our data cannot presently address this relationship of grain size and concentration. Suspended-sediment monitoring shows that near the mine, the suspended load is dominated by the mine waste (suspended load decreases with increasing discharge), but this effect ends with increasing discharge and drainage area downstream (Figure 5). A downstream decrease in trace metal concentrations occurs because of dilution by uncontaminated sediments from tributaries and to sediment exchange with the bed and banks. Various estimates have been made of the mine-derived component of the sediment load down the Strickland. Using tracer techniques, Apte [2001] reports a value of 10–11% as of 2000 at SG 4 (Figure 1). A. Markham (personal communication, 2006) calculated a sediment budget for the river system, and concluded that the mine-derived component at SG 4 progressively increased from 1991 to 1998, and since then has averaged ~15% of the suspended-sediment load. Further exchange and net deposition would make the mine-derived component even less by the time the Strickland joins the Fly. This small contribution is in sharp contrast to the middle Fly, where sediment load has been elevated by ~4.6 times the natural load because of mining discharges [Day et al., 2008].

[7] The period of significant mine operation is fairly short (10 years at the time of sample collection in 2003) and tailings discharge has varied with time. The mass of tailings discharged to the river increased progressively from 1992 to 1998 (Figure 6a). The concentration of Pb in the suspended sediment just 8 km downstream from the mine (SG1, Figure 1) increased during this time frame as well, but a very large spike occurred (due to lack of natural sediment dilution) during the 1997–1998 drought to flood cycle associated with a strong El Niño event (Figure 6b). Figure 6c shows the mean daily flow at a gauging station (SG3) ~220 km above our upstreammost floodplain sampling location. Though located in the upper catchment, these data show the severe drought of 1997 and suggest the

3.1. Sampling Strategy and Analysis

Two sampling campaigns were conducted on the Strickland floodplain to detect the spatial extent of mine-derived tracer accumulation across the Strickland floodplain. In 1997 (4 years after the commencement of significant discharge of mine tailings to the river, Figure 6a), five transects across the floodplain from near the gravel-sand transition (~410 km below the mine) to ~15 km upstream of Everill Junction (and 670 km below the mine) were sampled [Day et al., 1998]. These transects, numbered 1–5, are shown in Figure 7. Pits were dug at distances of 0, 50, 100, 250, 500 and 1000 km from each bank, and a 1-cm slice was taken at the surface and at 29–30 cm below the surface. The floodplain was dry at the time and access was entirely via helicopter. For consistency in this survey,
relatively straight sections were selected for each transect. Suspended load and bed material samples were also collected, although river stage was particularly low when the samples were taken.

[11] In 2003, the same transects were sampled in addition to 5 transects across actively shifting bends (Figure 7). On the basis of work investigating the Beni River in Bolivia [Aalto et al., 2003], it was anticipated that deposition rates would be higher if there was significant flow across the bends and out of the outer banks, particularly on the downvalley side of an actively migrating bend. In this report, data from three transects in bends, numbered transects 10, 13 and 15, are reported. Seven drop cores were also collected from the deltaic front of a tie channel entering an oxbow and extending along a transect to its distal end, 3 km into the lake. Cores were collected along the shore of this oxbow as well, numbered transect 12 (Figure 7). Sampling was severely constrained by standing water remaining on the floodplain after a flood event that preceded our arrival and by a temporarily low river stage. Our access boat could only travel ~100 km upstream of Everill Junction. Sampling above this point was accomplished using small boats, canoe, or helicopter. On the floodplain it was difficult to travel farther than 500 m from any bank without encountering deep standing water. Where possible, we sampled the transects at 5, 50, 100, 250, 350 and 500 m from each bank. Helicopter access was provided for a few days and distal cores (~1000 m) were collected on transects 1–5 for comparison to the 1997 study. To document whether Strickland sediment is currently spilling into Lake Murray via the Mamboi River (Figure 7), we used small boats to collect 27 cores along banks of the Mamboi and in Lake Murray.

[12] At sample locations other than those sampled via helicopter, three separate cores were collected. A push core with a 2.5-cm-diameter, 1-m-long Polyethylene liner was used to collect most of our cores. Penetration was rarely the full 100-cm length, and the longer of the two cores was selected for $^{210}$Pb analysis. Finally, a 5-cm-diameter hand PVC core liner was forced into the surface to ~10 cm depth and 1-cm slices were subsectioned from the cores and bagged in the field for shallow Pb and Ag analysis. At the helicopter sites, which were usually inundated, a single core was collected using a 5-cm gravity corer dropped from the

Figure 5. Suspended-sediment concentrations measured at increasing distances downstream from Porgera mine at stream gauges shown in Figure 1. Note that suspended-sediment concentration decreases with increasing discharge near the mine, SG1, indicating dilution, but increases with increasing flow farther downstream, SG4. This indicates that the mine loading does not have a large effect on the total load transported by the river, especially during large flows.

Figure 6. (a) Daily tailings discharge into the Strickland by Porgera mining (annual load in 1991 was similar to 1993, but daily discharge values were not available). (b) Particulate lead concentration at SG1, just downstream of the mine. (c) Daily flow records for SG3, 165 km from the mine and 470 km upstream from Everill Junction shown in Figure 1 (data provided by Porgera Joint Venture).
cores for $^{210}$Pb analysis, and 97 hand cores were collected. Transect 12, and curved transects 10, 13, and 15. Straight transects 1–4, moderately curved transect 5, oxbow push cores for Pb and Ag analysis.

A hovering helicopter and retrieved with a rope. A total of 144 push cores for Pb and Ag analysis $\sim 1$ m in length, 197 push cores for $^{210}$Pb analysis, and 97 hand cores were collected.

In order to characterize the bank and floodplain topography, each of the transects was surveyed (where access was not impeded by standing water) with a self-leveling transit. The coordinates for each core location were also recorded. Channel depth measurements were made with a sonar scanner.

3.2. Sediment Analysis

Chemical analyses were conducted at the Center for Environmental Contaminants Research of CSIRO Land and Water, Sydney, Australia, where well-developed analytical procedures were already in place. Cores were initially sectioned into 1-cm samples at 0–1, 1–2 and 29–30 cm depths. Then additional subsamples were analyzed to increase the resolution of data, so as to document vertical variation in deposition rate. In total, over 800 core sections were analyzed.

As noted above, only the fraction of the sediment that is $<63 \mu m$ was analyzed for this study in order to better isolate the mine-derived signal. The $<63-\mu m$ sediment fractions were isolated by wet sieving, dried and subjected to microwave-assisted digestion in a mixture of concentrated HCl and HNO$_3$. Metal concentrations in the resulting digest solutions were then determined using a combination of inductively coupled argon plasma emission spectrometry (ICP-AES), inductively coupled plasma mass spectrometry (ICP-MS) and graphite furnace atomic absorption spectrometry (GFAAS). Further details can be found in the work by Day et al. [1998] and Watson [2006]. For quality assurance purposes, one standard reference sample (PACS-2, National Research Council, Canada) and one blank were included in each analytical batch.

The organic carbon (OC) content was estimated for each sample through loss on ignition. Though OC% in each sample was generally small (\~{}9%). The concentration of metals measured by the ICP-AES and GFAAS was corrected to discount the average mass of organic material in the sample so as not to underestimate trace metal concentrations.

3.3. Determination of Deposition Rates

We used two different approaches to calculate sediment deposition rates from the Pb and Ag core profiles. In both approaches we assumed that by the time the mine-derived sediment reaches the Strickland floodplain system, it is well mixed with the natural load and has the same physical characteristics and behavior as natural sediment when it disperses across the floodplain.

The first method (inventory approach) assumes that the concentration of the tracers transported to the spot of deposition is a temporal and spatial constant. An integrated concentration (i.e., inventory) of the tracer is then measured through the core, and a deposition rate is calculated from the total amount of mine-derived sediment (for details, see Day et al. [2008]). Deviations at depth from the incoming tracer concentration are assumed to be due to mixing of mine-derived sediment with background floodplain sediment (i.e., through bioturbation and physical mixing processes), which may cause dilution of the signal and deeper penetration of mine-derived tracers than would occur by burial alone. The effective depth of sedimentation, $L_e$, is calculated by determining the proportion of each core interval that is composed of sediment with a set incoming trace metal concentration, $\varepsilon_i$, and a typical bulk density of contaminated surface sediment, $\rho_b$. This calculation is done for each interval depth ($i$) between successive subsamples and summed to get the equivalent depth of sedimentation with elevated mine-derived particles as shown in equation (1):

$$L_e = \sum_{i=1}^{n} \frac{z_i \cdot \rho_b (\varepsilon_e - \varepsilon_b)}{\rho_b (\varepsilon_e - \varepsilon_b)}$$

Here $\rho_b$ is the bulk density of uncontaminated sediments, $\varepsilon_d$, is the average, or measured, metal concentration for the interval depth $z_i$, and $\varepsilon_b$ is the background metal concentration. To apply (1), estimates must be made of incoming metal concentrations, $\varepsilon_i$, which as indicated by subscript, $i$, could vary with time and therefore depth. We also need to estimate the background concentration, $\varepsilon_b$, and directly measure the other properties. Measured trace metal concentrations above the calculated incoming concentrations are assumed to be unmixed and the entire interval length is included in the equivalent depth of sedimentation.

The other method used here (the depth-of-occurrence approach) assumes that the deposition rate is high enough and the depositional period short enough that bioturbation is ineffective, and that vertical variation in mine-derived Pb and Ag is due to variations in the concentration of tracer being delivered. In this second case we assume that the first occurrence of any elevated concentration of Pb or Ag relative to background defines the arrival of mine-derived sediment. This depth of occurrence directly translates to a rate for the time since the commencement of mining. Because the metal concentration of each core is defined by a series of discrete subsamples that may not capture the exact depth at which this occurs, the depth of first occur-
rence is linearly interpolated on the basis of the respective depths and concentrations of adjacent subsamples. As in the inventory case, a background concentration has to be assigned.

[20] Both methods also include a density correction. Reported sediment deposition rates are for sediments with a bulk density of 1.5 g/cm³ [Aalto et al., 2008]. Once the sediment deposition rate was calculated for each core, we then explored the spatial pattern of deposition to determine how to obtain the total flux of overbank sediment.

3.4. Derivation of Background and Mine-Contaminated Sediment Concentrations

[21] Data from suspended material and floodplain sediment analyses were used to determine both background and incoming concentrations. First, the upper limit of trace metal concentrations was determined for suspended-sediment samples from Strickland River tributaries not influenced by mining (32 µg/g Pb, 0.32 µg/g Ag, as reported by Apte [2001]). To quantify the background metal concentrations, we reviewed 111 cores to determine the depth of first occurrence for concentrations greater than any measured in the tributaries. A histogram of Pb and Ag concentrations in the samples taken below this first occurrence was then plotted (198 samples in total, Figure 8). The Pb values were approximately normally distributed with a median value of 16 µg/g, and 95% of the subsamples contained 25 µg/g Pb or less. The 80th percentile value for Pb was 22 µg/g. The Ag values were skewed toward the smaller concentrations with a median value of 0.07 µg/g, a 95th percentile value of 0.16 µg/g, and a distinct break at the 80th percentile of 0.1 µg/g. On the basis of this analysis, we used 25 µg/g Pb and 0.16 µg/g Ag with the understanding that 22 µg/g Pb and 0.1 µg/g Ag represent more conservative alternatives. The deposition rate was determined for both methods described below using both the 95th percentile and 80th percentile background values. Lead and silver concentrations were closely correlated in floodplain sediments (r = 0.85, n = 706). For the sake of brevity, we will primarily discuss the results for the analysis of Pb with a background value of 25 µg/g.

[22] The concentrations of the first two subsamples from each core, 0–1 and 1–2 cm respectively, where there was a distinct variation above the background values for both Pb and Ag were plotted to determine the incoming trace metal concentrations (n = 109). These data were normally distributed and had mean values of 49 µg/g Pb and 0.5 µg/g Ag, respectively (Figure 9). These values are in agreement with the measured concentrations of what appeared to be very young deposits in the field and with concentrations measured in suspension during the falling limb of the 2003 flood (60 µg/g during a discharge of about 3100 m³/s).

4. Deposition Patterns on the Strickland River Floodplain

4.1. Deposition of Mine-Derived Sediment on the Floodplain

[23] Figure 10 shows examples of cross-stream variation in surface and 30 cm depth Pb concentrations at transect 5 for the 1997 and 2003 survey, 4 and 10 years, respectively, after significant volumes of mine-derived sediment entered the Strickland (Figure 6a). These data clearly reveal the overbank deposition of mine-derived sediment, with initial deposition closer to the channel, followed by a widening of the active depositional area and significant elevation of Pb concentrations in the surface sediments in 2003. As shown in Figure 6a, tailings loading reached full mine operation level in 1998. So, the large change is in keeping with this increased loading, and with the spike in Pb concentrations that occurred during 1997 and for a period thereafter. Note that all measured concentrations at 29–30 cm were below background in 1997, but were elevated above background near the channel on both banks by 2003.

4.2. Profile Analysis

[24] Core profiles generally showed a consistent trend of Pb and Ag accumulation in surficial sediments. In this section, transect 5 is discussed in detail in order to illustrate some key observations. Figure 11 shows the cross-sectional topography and corresponding Pb concentration profiles for 11 cores taken across the floodplain. As seen in Figure 12
Figure 10. Comparison of Pb concentration in surface and 29–30 cm samples taken in 1997 and 2003 at transect 5. Note the overall increase in concentrations at the surface and at depth as well as the spread of contamination across the floodplain over time. The distances plotted are to the nearest location on the main channel.

Figure 11. Transect 5 topography and individual core concentration profiles. (a) Field-surveyed topography (distal elevations are estimated) with core locations shown. The distances shown are along the transect, but the distal cores are actually closer to upstream reaches of the channel. (b) Pb concentration profiles for each of the 11 cores along transect 5. The distal cores, L1000 and R1000, are approximately 500 and 800 m from the upstream left and right banks, respectively. The number at the bottom of each subplot is the interpolated depth at which the Pb concentration would equal the background value of 25 µg/g.
shift may simply reflect this change. The profiles, when sampled intensively, unexpectedly reveal considerable vertical variation, and there are distinct subsurface concentration maxima. The right bank 150-m core has a spike (153 μg/g Pb, the highest recorded in any floodplain core) at 35–36 cm below the surface. On the left bank, the spike is at 3–4 cm below the surface (at 5, 150, and 250 m from the bank), and the measured concentrations are smaller. We suggest the vertical variation in Pb and the spike record the temporally variable and stage-dependent metal concentration in the suspended load (see Watson [2006] for more discussion). Although on an annual basis the mine load is ~15% of the total load at SG 4 (360 km from the mine), during periods of low flow and periods of drought, the proportion of the sediment that is derived from the mine can increase considerably. Conversely, during floods, the large introduction of natural sediments dilutes the concentrations of mine-derived tracers to low values. The spike in concentration may record the first flush of sediment after the 1997 drought [Watson, 2006], with the higher proportion of mine-derived sediment in this material causing a significant increase in metal concentrations in the floodplain sediment. Continued overbank flow, and subsequent flood events with increased natural sediment loads would then reduce the Pb and Ag concentrations. This may be the case in three cores where the concentrations in the first 1 cm approach background levels (5 m left, 50 m and 150 m right).

The cores reveal deposition patterns across the floodplain that are consistent with the position of the transect in a bend and with local topographic effects. In Figure 11, we have labeled each profile with the depth where we infer mine-related sedimentation began, given a background concentration of 25 μg/g. The deposition rate is then calculated by correcting for density variation and dividing by 10 years. On the right cut-bank side of the bend, overbank deposition rates were 3.5–6 cm/a out to 150 m. Further coring across the floodplain was not possible because of deep standing water and time constraints (access to distal cores was entirely by helicopter). We calculated a deposition rate of ~1.0 cm/a for a core collected by helicopter >1 km from the right bank, where it is crossed by the transect. This site, however, was ~800 m from the nearest bank upstream. On the left bank side 5 and 50 m from the bank, the cores are contaminated through their entire depth (60 and 73 cm, respectively). These deposits probably record rapid vertical accretion on the advancing point bar. Hence the strong vertical variation probably records stage-dependent Pb concentration in the suspended load. Farther across the floodplain, the deposition rate drops to 1.2–1.3 cm/a until the 1000-m core is reached, where the net deposition rate increases to 1.8 cm/a. We note that this site is closer to the outer bank of an upstream bend (~500 m from the bank) and suggest that the increase in sediment accumulation rate records the influence of this upstream bend. This finding caused us to correct all distances relative to the nearest bank rather than along a particular transect.

[27] To estimate the sediment deposition rate from the inventory method at transect 5 and subsequent transects, we apply equation (1) assuming the background values for Pb of 25 μg/g and Ag of 0.16 μg/g in the higher case, and 22 and 0.1 μg/g in the lower case. Incoming concentrations at flood stages have not been monitored, and as suggested by the analysis of transect 5, the concentrations will be highly variable. We did not attempt to estimate the temporal record of ξ, given the likelihood that incoming concentrations are highly variable with both seasonal and stage effects, which would be extremely difficult to measure and predict. This method provides a lower estimate for the average deposition rate, about half that indicated by the depth-of-occurrence method, but mirrors the spatial patterns. For transect 5, the calculated deposition rates for 25 μg/g Pb background concentration on the right cut bank has a range of 1.4–3.1 cm/a in the near-channel cores with the distal core recording 0.4 cm/a. The corresponding rates on the left bank are 3.5 and 5 cm/a at 5 and 50 m from the channel, respectively, dropping below 0.7 cm/a at 250 m from the channel and increasing to 1.0 cm/a in the distal core.

4.3. Calculated Deposition Rates

[28] Table 1 summarizes the sediment deposition rate for each core site across all transects sampled, and Figure 13a shows the deposition rate for each core as a function of distance from the nearest channel bank. It should be noted that vertically adjacent core subsamples from which the linear interpolation of the depth of occurrence is calculated are more than a 10 cm apart for 21 of the cores sampled. Of the 111 cores analyzed in this study, 28 were contaminated throughout their depth (hence we have only a minimum deposition rate) and 9 showed no evidence for deposition of mine-derived sediment. For the remaining 74 cores, the depth of first mine sediment occurrence was linearly interpolated between two values separated by an average core length of 10.4 cm with a standard deviation of 10.4 cm. Through this 10-cm depth interval, Pb concentration gradients averaged 5.5 μg/g-cm with a standard deviation of 7.1 μg/g-cm. With the linear interpolation procedure, the estimated depth of first mine-derived sediment would decrease with increasing metal concentration in the upper layers. Thus our deposition rates depend to some degree on...
the incoming metal concentrations, which vary spatially and temporally across the floodplain.

[29] As shown in Table 1, the average deposition rate for each side of the channel, calculated as the numeric integral of sediment deposition rate as a function of distance from the channel divided by transect width, varies from 0.04 cm/a (on the right bank of transect 4) to 4.85 cm/a (right bank of transect 10). The average deposition rate per channel distance decreases from nearly 5 cm/a near the bank to 0.35 cm/a 1450–2000 m from the channel. Figure 13a separates cores from straight sections, with relatively slowly shifting banks (transects 1–4) from cores in bends (transects 5, 10, 13, and 15), and from a core collected on the banks of an oxbow (transect 12).

[30] Several patterns emerge. Deposition rate generally decreases with distance from the bank, but local values are highly variable. An exponential function is commonly applied to floodplain sedimentation data [e.g., Tornqvist and Bridge, 2002; Walling and He, 1998; Day et al., 2008]. Figure 13b shows that the data generally follow an exponential decline, but near the bank the rate of decline is systematically higher than predicted by the function. Several of the near-bank cores may record the effects of lateral shifting and rapid vertical accretion on newly emergent bars. Farther out, overbank deposition was the dominant process. This change in process may explain the shift in deposition rate with distance, which is apparent in the data.

[31] On a given transect, deposition generally differed between the left and right banks. Although picked for their generally straight reaches, none of the five transects established in 1997 were in perfectly straight channels. Transects 1, 2, 4 and 5 show curvature and in each case the bank and the adjacent floodplain for a few 100 m was higher on the outer bank. Transect 3 is centered in a relatively straight reach between two bends in which the channel width progressively increases downstream. In all cases, the sediment deposition rate was higher on the bank with the higher floodplain elevation, suggesting that the higher rate was associated with stronger overbank flows because of bank curvature.

[32] As expected, the more strongly curved transects (5, 10, 13 and 15) had higher overbank deposition rates (over the comparatively short distances from the channel that were measured). The reach with the least floodplain sediment deposition was transect 4. Comparison of 2000 and 1972 satellite imagery reveals this section to be the most stable of all the transects, showing no visible shifting. Transect 12 samples were collected on the banks of an oxbow rather than normal to the bank of the active channel. To compute deposition as a function of distance we measured the distance to the nearest bank of the Strickland main stem. Sediment deposition was low along the entire transect, ranging from 0.1–0.4 cm/a.

4.4. Floodplain Sediment Deposition Rate

[33] There are several approaches that could be used to estimate total floodplain deposition of sediment (M, t/a). One would be take the average rate (in cm/a), Da of all cores, multiply that times the active area (A) of deposition on both sides of the channel, and apply an appropriate average bulk density (ρs), i.e.,

\[ M = \rho_s D_a A \]  

Our samples, however, are strongly biased toward areas near the bank (due to access difficulties). Instead, we estimated a distance-dependent deposition function for the entire data set and integrated that function across the floodplain. Sparsely sampled regions as defined by distance from channel were grouped with adjacent samples to reduce the bias toward individual samples, as demonstrated in Table 1. Figure 13a shows that deposition rates are similar for straight and curved reaches very close to the channel, because of high deposition rates associated with relatively minor channel shifting. Sediment deposition rates remain high farther from the

Table 1. Summary of Sediment Accumulation Rate Calculated for Each Core Using the Depth-of-Occurrence Method for a Background Concentration of 25 μg/g Pb

<table>
<thead>
<tr>
<th>Transect</th>
<th>Distance From Channel, m</th>
<th>475–625</th>
<th>800–1025</th>
<th>1000–1450</th>
<th>1875–2000</th>
<th>Average</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 left</td>
<td>3.1 3.1 2.5 1.9 1.2 0.9 1.53</td>
<td>0.4 1.5 0.96</td>
<td>0.7 1.4 1.61</td>
<td>0.2 0.3 0.9 0.71</td>
<td>0.0 0.37</td>
<td>3.25</td>
</tr>
<tr>
<td>2 left</td>
<td>7.8 1.9 2.6 1.6 0.8 1.9 2.3</td>
<td>0.7 1.4 1.61</td>
<td>0.2 0.3 0.9 0.71</td>
<td>0.0 0.37</td>
<td>3.25</td>
<td></td>
</tr>
<tr>
<td>3 left</td>
<td>6.8 0.6 0.5 0.7 0.6 1.5 2.0</td>
<td>2.0 2.2 3.25</td>
<td>0.0 0.1 0.2 0.3</td>
<td>0.1 0.2 0.3</td>
<td>3.25</td>
<td></td>
</tr>
<tr>
<td>3 right</td>
<td>6.8 0.6 0.5 0.7 0.6 1.5 2.0</td>
<td>2.0 2.2 3.25</td>
<td>0.0 0.1 0.2 0.3</td>
<td>0.1 0.2 0.3</td>
<td>3.25</td>
<td></td>
</tr>
<tr>
<td>3 left 10</td>
<td>3.3 5.5 5.9 7.0 3.5 2.1 0.4</td>
<td>2.1 2.9 4.85</td>
<td>0.7 0.8 1.8</td>
<td>1.2 1.3</td>
<td>3.25</td>
<td></td>
</tr>
<tr>
<td>3 right 10</td>
<td>3.3 5.5 5.9 7.0 3.5 2.1 0.4</td>
<td>2.1 2.9 4.85</td>
<td>0.7 0.8 1.8</td>
<td>1.2 1.3</td>
<td>3.25</td>
<td></td>
</tr>
<tr>
<td>11 left</td>
<td>2.7 2.3 4.8 1.0 1.0 1.0 1.0</td>
<td>0.2 0.1 0.2 0.1 0.35 0.35</td>
<td>2.7 4.46 0.91 0.91 0.35 0.35</td>
<td>2.7 4.46 0.91 0.91 0.35 0.35</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Average</td>
<td>2.87 4.46 3.07 2.87 1.45 1.65 1.29 1.88 0.56 1.15 0.95 2.87 4.46 0.91 0.91 0.35 0.35</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*Numerically integrated average deposition for each transect is shown; units are cm/a. Data were combined with adjacent samples where two or fewer cores recorded the deposition rate for a single distance from the channel.*
channel banks on reaches with sharp channel curvature than on straight reaches. Inspection of the maps of the Strickland (Figure 7) suggests that the proportion of bank length in bends and straight sections is roughly equal, so we will treat the deposition data as a single data set. An exponential fit to data binned by distance from the channel (Table 2) systematically underestimates the high rates of deposition near the channel bank (Figure 13b). Rather than impose a function on these data, we numerically integrated the data and divided by the active deposition width to get a mean deposition rate (cm/a). We then assigned this average rate to $D_a$ in (2), with the associated channel area.

[34] The distances over which equation (2) can be applied were limited to the region where the pattern of deposition was well defined. There are only 4 data points beyond 1 km from the channel bank, and two of these points show very low values (0 and 0.1 cm/a). Hence the deposition rate is not well constrained farther out, though the rate there is distinctly lower than near bank values. Furthermore, because of channel sinuosity, projection of deposition rate beyond 1 to 1.5 km for the width, $w$, is problematic because there is significant overlap of bank-normal projections. On the Strickland, the actual area of the floodplain bordering 1 km either side of the channel is 455 km$^2$, and for 1.5 km out it is 597 km$^2$ (compared to the estimated total floodplain area of 1165 km$^2$, with channel surface area deleted from the total). If we estimate the average deposition beyond 1 km and apply it to the total floodplain area, we will ultimately underestimate the total loss to deposition, because the area increases by only 30% when the active depositional width increases by 50%. For these reasons, we estimate the deposition on the basis of a 1 km width. The use of 1 km is more conservative and makes the width identical to that used on the Fly River, to which we wish to make comparisons [Day et al., 2008]. We also use the GIS determined area of 455 km$^2$ for the Strickland floodplain.

[35] Our upstreammost sample location (267 km from Everill Junction), transect 5, is in a gravel bedded reach that progressively steepens upslope. On the basis of satellite imagery, floodplain borders the channel for another 49 km upstream, but in this reach the channel shifts more slowly [Aalto et al., 2008], and the discharge is flashier than in the lower Strickland. It seems possible that rates and patterns of overbank deposition may change significantly up the canyon. In order to compare with the sandbedded Fly rates, it is perhaps most appropriate to use the 267 km length of sandbedded channel below transect 5 when calculating $A$ in equation (2). However, for consistency with the deposition rates presented by Aalto et al. [2008] who perform their analysis for the ~318 km of sinuous single thread lowland Strickland, we also report results for a 318-km lowland channel length. Aalto et al. [2008] calculate channel area, $A$, as channel length times the 1-km active width of deposition on either side of the channel ($A = 636$ km$^2$), and acknowledge that this leads to an ~10% overestimate of the total area because of overlapping bank-normal projections. For consistency with their results, we use $A = 636$ km$^2$ when calculating the floodplain losses for the longer channel length.

[36] Table 2 summarizes the 1-km average sediment deposition rate (corrected to a bulk density of 1.5 g/cm$^3$), $D_a$ in equation (2), for both the inventory and depth-of-occurrence methods employing different assumptions about background and incoming Pb and Ag levels. We also report the total annual mass deposited overbank (M in equation 2) and its proportion of the total sediment load (estimated to be 70 Mt/a). The inventory method yields smaller values because this method assumes that values between background and estimated incoming concentration represent
dilution by bioturbation with uncontaminated sediments. These rates define the lowest probable deposition rate. The highest rate was associated with the lowest estimated background Pb and Ag concentrations using the depth-of-occurrence method. This sediment deposition rate is probably too high, as deep cores, clearly not contaminated, at transect 12 were counted as contaminated throughout their entire length because the uniformly low Pb concentrations were slightly above the assigned background concentration.

[37] Our best estimate, based on the depth-of-occurrence method, is that the mean sediment deposition rate across the floodplain to 1 km from the channel bank is 1.4 cm/a or 9.2 Mt/y for the sandbedded river length of 267 km or 13 Mt/y for 318 km of active lowland channel. These deposition rates equal 13 and 19% of the estimated mean annual load of 70 Mt/a. These rates normalized by channel length are 0.05%/km and 0.06%/km, respectively.

[38] Two other depositional processes are not accounted for in this analysis. One oxbow connected via a tie channel was sampled and high rates of sediment accumulation were measured there. All core samples to a depth of 40 cm (bottom of core) were contaminated out to a distance of 1200 Mt/a annual load. Although we have not quantified the Holocene sediment discharge to the lower Strickland, there is no evidence of major stream capture at this time, or of large climatic shifts. Hence the elevated load on the Strickland, derived from headwater channels cutting deep canyons into rapidly uplifting lands, has most likely occurred throughout this period. Radiocarbon dates in cores up to 15 m deep obtained along the middle Fly and lower Strickland demonstrate that aggradation kept pace with sea level rise, even during early postglacial rapid rise [Chappell and Dietrich, 2003]. Lauer et al. [2008] infer through modeling that the sediment load on the Strickland was much higher than the Fly over this period. The Fly, on the other hand, did experience an early Holocene pulse of sediment due to a 7 km3 landslide in its headwaters [Blong, 1991]. Much of that sediment remains in the headwaters, but initial erosion of the deposit may have contributed to the Fly River keeping pace with sea level rise. On the other hand, we see evidence of a 7 km width of the deposit. Our best estimate, based on the depth-of-occurrence method, is that the mean sediment deposition rate across the floodplain to 1 km from the channel bank is 1.4 cm/a or 9.2 Mt/y for the sandbedded river length of 267 km or 13 Mt/y for 318 km of active lowland channel. These deposition rates equal 13 and 19% of the estimated mean annual load of 70 Mt/a. These rates normalized by channel length are 0.05%/km and 0.06%/km, respectively.

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[40] Differences in morphology associated with sustained differences in sediment load on rivers responding to sea level rise are seen on other rivers, notably the Rio Negro and the Solimoes-Amazon River. Dunne et al. [1998] describe the morphologic diversity and dynamics of the Amazon, emphasizing that sediment exchanges between the floodplain and the channel exceed the 1200 Mt/a annual load. There are extended reaches with large bars, rapid lateral channel shifting, and extensive floodplain deposition. In contrast, the Rio Negro, which drains 27% of the

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**Table 2. Summary of Results for Each Method Using both the 80% and 95% Values for Background Concentration**

<table>
<thead>
<tr>
<th>Method</th>
<th>Distance From River Bank, m</th>
<th>1-km Average</th>
<th>2-km Average</th>
<th>Mass Lost, Mt</th>
<th>Mass Lost, %</th>
</tr>
</thead>
<tbody>
<tr>
<td>25 µg/g Pb</td>
<td>2.9</td>
<td>4.5</td>
<td>3.1</td>
<td>2.9</td>
<td>2.1</td>
</tr>
<tr>
<td>22 µg/g Pb</td>
<td>2.9</td>
<td>4.5</td>
<td>3.4</td>
<td>3.0</td>
<td>2.3</td>
</tr>
<tr>
<td>0.16 µg/g Ag</td>
<td>2.7</td>
<td>2.5</td>
<td>3.2</td>
<td>3.0</td>
<td>2.3</td>
</tr>
<tr>
<td>0.1 µg/g Ag</td>
<td>3.3</td>
<td>3.8</td>
<td>3.3</td>
<td>3.4</td>
<td>2.7</td>
</tr>
</tbody>
</table>

Note: Rates are recorded in cm/a. The average rate is determined by numeric integration to 1 km divided by width. The proportion of the load lost is calculated for 1 km on either side of the channel for the entire 455-km² floodplain (for the 267-km sand-bedded channel reach) relative to the discharge of 70 Mt/a load.
Amazon basin where they join, discharges only 8 Mt/a to the Amazon [Latrubesse and Franzinelli, 2005]. The lower 300 km of the Rio Negro consists of a system of islands, lakes and channels dominated by backwater effects of the Amazon. Latrubesse and Franzinelli [2005] argue that the Rio Negro progressively responded to Quaternary climate change and sea level rise, but that the lower reach of the river did not have sufficient sediment load to keep pace with postglacial sea level rise on the Amazon. Consequently, the main stem Amazon response to sea level is driving the morphodynamics of the lower Negro.

[42] Comparable rates of floodplain deposition have been documented on other large lowland rivers. On the Brahmaputra-Jamuna River, Allison et al. [1998] report deposition rates as a function of distance from channel banks (like our Figures 13 and 14) and fit a power law to their data. The function defined by Allison et al. [1998] matches with our observations and those of Aalto et al. [2008] (Figure 14). Allison et al. [1998] also note, as we have found, that an exponential function represents the data poorly. For their study reach, they report 8% of the sediment load was being deposited on the floodplain over 110 km, or a value of 0.07%/km of channel reach, a rate similar to our measurements. On the tidal reach of the Amazon below Obidos, Dunne et al. [1998] conclude that there is a loss of 350 Mt/a relative to a total load of 1200 Mt/a as the river travels 450 km toward the sea. This gives a deposition rate of 0.065%/km of channel length. On the 833 km reach between Sao Paulo de Olivenca and Obidos, which is not tidally influenced, floodplain deposition rate is ~0.1% of total load/km of channel length. On the Mississippi, Kesel et al. [1992] report a 39% loss of sediment to the floodplain over 1674 km, or 0.02%/km of channel length. Much higher rates of loss per channel length (well above 0.1%/km of channel length) are found on the relatively small rivers studied by Nicholas et al. [2006] in the United Kingdom. Overall, the values we find for the Strickland are similar to those reported elsewhere for most large rivers, although they are lower than those reported for the Beni River, Bolivia [Aalto et al., 2003].

[43] Comparison of depositional rates on the Strickland and Fly reveal some surprising results. The deposition rate for the first 1 km of floodplain from the bank is nearly 10 times greater on the Strickland (1.4 cm/a) than that documented for the natural load on the Fly (0.1 to 0.2 cm/a, Day et al. [2008]). Figure 14 shows the measured rates on the Fly compared to those of the Strickland, but these rates are elevated because of mine waste loading on the Fly. Day et al. [2008] propose that these rates are elevated in proportion to the increased load by the mine; hence the natural load rates are estimated to be ~4.6 times lower. Given that the Strickland carries ~7 times the natural load of the Fly annually, it makes sense that annual accumulation rates would be greater. This may seem to contradict our stated hypothesis, but other differences become important. We note that the Strickland deposits 13–19% of its total sediment load overbank through its lower reaches, whereas, on the middle Fly, 40% of the annual load is sequestered. This difference is partially a result of the longer reach on the Fly, but correcting for length, the sediment deposition rate as a proportion of the load is 0.05–0.06%/km per channel length on the Strickland, and 0.09%/km channel length on the Fly. On the Fly, however, half of the sediment deposition occurs when sediment-laden flows are pushed up tributary and tie channels and spill onto to the floodplain at a considerable distance from the main stem. Including these additional channels, the effective channel length on the middle Fly is 1325 km (as compared to just the main stem of 420 km). Normalizing by this greater length gives a sediment loss rate of 0.03%/km of channel length on the Fly. Thus, per main stem channel length, the lower load on the Fly leads to lower floodplain deposition rates, but the extensive network of tributaries and tie channels into which the Fly...
pumps sediment causes it to deposit a much greater proportion of its sediment load on the floodplain than does the Strickland.

[44] Although the percentage sediment load deposition per unit main stem channel length is less on the Strickland, the absolute deposition rate in Mt/y is much higher than the natural rate on the Fly. However, this does not mean that the Strickland is still rapidly infilling the accommodation space (in the sense of Blum and Tornqvist [2000]) associated with postglacial sea level rise. The average migration rate of 5 m/a on the Strickland [Aalto et al., 2008] implies that the channel can easily migrate across the 1-km wide depositional width in just 200 years, sweeping most of the deposited sediment back into the channel. Furthermore, the exchange rate (in Mt/a) with the floodplain due to lateral migration is much higher than the overbank deposition rate. A lateral migration rate of 5 m per year, times a channel distance of 318 km and an estimated average bank height of 13 m yields ~30 Mt/a (for a density of 1.5 g/cm³). In contrast, on the Fly, the migration rate is 1–2 m/a in the upper two thirds of the middle Fly and zero in the lower third of the middle Fly. Dietrich et al. [1999] estimate that bank migration causes it to deposit a much greater fraction of its sediment load on the floodplain than does the Strickland.

[45] This analysis points to two important issues: (1) topographic controls on a river’s response to sea level rise and (2) the long-term stratigraphic record versus the short-term deposition rates. Figures 3 and 4 show that the Strickland has built a scroll bar ridge (or meander belt) across a wider valley trough. Potential accommodation space on the Strickland River is still plentiful in backswamp areas bordering this belt and in Lake Murray. Former meander belts, similar to those that populate the Holocene alluvial surfaces of other large systems such as the Mississippi River [e.g., Aslan and Autin, 1999] and Colorado River [e.g., Blum and Tornqvist, 2000], however, are absent. The recent flood breakout down the Mamboi River (Figure 7) into Lake Murray defines a potential avulsion path, which, should the Strickland shift there, would cause a significant channel displacement and lead to large areas of sediment accumulation. We note that the current outflow channel of Lake Murray, the Herbert River, seems much too wide for its drainage area, and is bordered by oxbows despite the current absence of bars or lateral migration of the channel. One possibility is that the Herbert River was an earlier path of the Strickland, and the path down the Mamboi was exploited much earlier in the Holocene.

[46] Blum and Tornqvist [2000] suggest that on the Colorado system complete filling of the accommodation space would favor avulsion. Slingerland and Smith [2004] emphasize the tendency for aggrading floodplain systems to avulse. On the Strickland, even with more infilling of backswamps, the valley width is still relatively narrow and may inhibit avulsion but for the path down the Mamboi. Our measurements of floodplain deposition and lateral migration also indicate that despite the extensive backswamp areas bordering the channel, little net sediment accumulation appears to be occurring in these distal locations. Hence, well before complete “filling” of accommodation space, the rate of floodplain sediment accumulation appears to have dropped to low values. Three factors appear to contribute to this state. First, as just reviewed, there is limited space for avulsion and potential accommodation areas are not reached. Second, the rapid drop off in sediment deposition on the floodplain with distance from the channel means limited sediment delivery to more distal portions of the floodplain. Third, vertical accumulation rate may be relatively low. This rate will tend to keep pace with the combined effects of subsidence, sea level rise and delta extension [Aalto et al., 2008]. There may also be some effect of hydrostatic warping due to sea level rise [Chappell and Dietrich, 2003]. We see little evidence for either subsidence or uplift on the lower Strickland. The late Holocene slow sea level rise and delta growth, then, requires little accumulation along the Strickland to keep pace.

[47] Another important issue is the disparity on the Strickland between the relatively rapid short-term floodplain deposition rates and the apparent slow accumulation rates over the long term. An inverse relationship between deposition rate and period of observation was noted by Sadler [1981] and often referred to as the “Sadler effect”. Bridge [2003] notes specifically that average floodplain deposition rates tend to decrease with increasing time interval of record. Mechanisms for this disparity, found throughout sedimentary systems, have been debated [e.g., Sommerfield, 2006]. Here it appears that a net balance between short-term (event and decadal) overbank deposition and progressive lateral migration could lead to a small net accumulation over the long timescales. In this case, the inverse relationship between deposition rate and observation time emerges from the changing balance of processes over time.

6. Conclusions

[48] The Fly River system provides us with a natural experiment in which two large rivers with greatly different sediment loads respond to the same postglacial sea level rise. Our field investigations and modeling reported here and in companion papers (Aalto et al. [2008], Day et al. [2008], and Lauer et al. [2008] document modern floodplain deposition rates and place constraints on longer-term accumulation rates. Together these data demonstrate strong differences between the Strickland and Fly rivers that arise because of sediment load.

[49] The Strickland discharges ~7 times the natural load of the Fly. Close to where the two rivers join, the Strickland is nearly 10 times steeper, carries bed material whose median size is twice that of the Fly, and has a lateral migration rate of 5 m/a as compared to nearly zero (in the past 50 years) on the Fly. Furthermore, the Strickland has wide shallow point bars that commonly are divided into partially vegetated islands, whereas the lower 170 km of the Fly upstream from the junction with the Strickland is essentially bar-free. The Fly discharges sediment into tributaries and tie channels leading to net accumulation in the floodplain. Slow lateral channel shifting also means that...
overbank deposition is contributing to net accumulation, although natural rates of deposition are slow (0.1–0.2 cm/y to 1 km from the bank) on the Fly. In contrast, the Strickland deposits sediment overbank at 1.4 cm/a averaged over 1 km from the channel bank, but high lateral migration rates sweep sediment back into the river, leading to relatively low net sediment accumulation rates. Hence the river with the higher load has a higher floodplain deposition rate, but a faster lateral migration rate, leading to much less of its floodplain overbank deposits contributing to the total net storage and infilling of accommodation space. This dynamic balance supports our initial hypothesis that the Strickland has infilled more of its accommodation space and is correspondingly losing less of its sediment load to floodplain deposition. The current low rates of accumulation within the scroll bar complex are set by the combined effects of sea level rise and delta advance (both of which have been slow in the late Holocene). The relatively narrow valley in which the Strickland flows may have prevented multiple Holocene meander belts from forming as has been found elsewhere.

[50] Deposition rate decreases rapidly with distance from channel bank, but the pattern differs between the Strickland and the Fly. Both the mine tracer occurrence measurements reported here and the 210Pb data reported by Aalto et al. [2008] show an abrupt decrease in floodplain deposition rate with distance from the channel bank that is not well represented by an exponential function (often used to characterize near bank deposition [e.g., Tornqvist and Bridge, 2002]). Deposition extends beyond 1 km, and can occur out to 2 km from the channel. Rates of overbank deposition were greater farther from the channel on bends than on straight sections. Interestingly, the average deposition rate as a function of distance from bank on the Strickland is very similar to that reported on the Brahmaputra-Jamuna River [Allison et al., 1998] and Beni River [Aalto et al., 2003]. In contrast, the Fly River overbank deposition rates are well defined by an exponential, and deposition essentially ceases 1 km from the bank. These documented rates, and their similarities and differences between river systems, are not yet explained quantitatively.

[51] We have argued from inference and modeling that the long-term (Holocene) sediment load on the Strickland has probably been persistently higher than the Fly because of the greater drainage area in the rapidly uplifting and eroding headwater mountains. It is difficult, however, to reconstruct this history. Knowledge of the depth and rate of sediment accumulation over the Holocene along the Strickland would help provide some important constraints. Seismic surveys and dating of sediments obtained from deep cores across the floodplain would be invaluable.

[52] Taken together, these observations argue that the time evolution (and resulting morphodynamics and stratigraphic record) of a large lowland river in response to sea level rise will depend strongly on upstream sediment load to the river. That load, however, is not necessarily uniformly dispersed into available floodplain accommodation space. Low sediment load leads to slow development, and even after 20,000 years since the onset of sea level rise and more than 6000 years of nearly constant sea level, the Fly River continues to lose a large proportion of its sediment load to its floodplain. In contrast, the Strickland has more fully responded, and transfers the majority of its sediment load through its lowland valley, even though substantial areas of accommodation space remain unfilled.

[53] Acknowledgments. Primary support for this project came from an NSF Margins (Source to Sink) grant (NSF EAR-0203577). Development and refinement of the 210Pb dating technique was supported in part by NSF EAR-0310339 and EAR-0403722. Some support also came from the National Center for Earth-surface Dynamics. The Western Venturer, which was essential to the program, was provided by Ok Tedi Mining Limited (with a special thanks to Jim Veness). Helicopter and food support in the field and support for laboratory analysis were provided by Porgera Joint Venture (Tim Omundsen, Charlie Ross, and Jim McNamara). Valuable hydrologic monitoring data were also provided by Porgera Joint Venture (through the help of Andrew Markham (Hydrobiology, Australia)). Lihir Mining (Geoff Day) gave valuable logistical assistance in Port Moresby and provided some equipment. Cliff Reife, John Sanders, and Ian Hargreaves were invaluable to the field campaign. Stuart Simpson and David Bishop (CSIRO) are thanked for their contributions to the chemical analyses undertaken as part of this project. The authors are grateful for the reviews and comments provided by Michael Blum, Gary Parker, and Charles Nittouer.

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