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The Fly River, Papua New Guinea: Inferences about River Dynamics, Floodplain Sedimentation and Fate of Sediment

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ABSTRACT

The Fly River drains the Southern Fold Mountains of Papua New Guinea where it descends down steep valleys bordered by peaks of up to 4000 m and then crosses a low-relief, nearly flat plain to the Gulf of Papua. Only about 30% of the 75 000 km^2 basin is in the rapidly eroding uplands. The basin receives up to $10\,m\,a^{-1}$ rainfall in the uplands and still exceeds $2\,m\,a^{-1}$ in the lowlands. Here we focus on the 450 km long Middle Fly River and its floodplain. Along this reach, the slope decreases by a factor of more than three (from 6.6×10^{-5} to 2×10^{-5}), the median grain size declines by a comparable amount (to about 0.1 mm) and duration of flooding greatly increases, causing the rainforest vegetation of the upper reach floodplain to give way to a swamp grass-dominated lower reach. Our analysis of observations on the Fly sheds some light on three questions: (1) what controls the rate of channel migration? (2) what controls the rate of floodplain sediment deposition? (3) what controls downstream fining in a large sand-bedded river? Previous studies have interpreted the well preserved floodplain features of former channel positions (scroll bar complexes, filled oxbows and oxbow lakes) as indicating a rapidly migrating, unstable river which frequently reworks its floodplain. Simple assumptions regarding frequency of loop cutoff, comparison of available topographic maps, and recent high resolution aerial photographs all indicate that, instead, the Fly River migrates slowly, typically less than 0.0045 channel width per year in the forested floodplain reach, and appears to be virtually stagnant in the swamp grass reach. The well preserved floodplain features in the roughly 200 km long swamp grass reach suggest that channel migration rates here were higher in the recent past. One mechanism which might allow a meander to slow or cease movement is for the river slope to decrease, causing the boundary shear stress on the bank to drop below a critical value. Such a change may also be responsible for the unusually large-wavelength bends in the swamp grass reach which have long, nearly straight limbs. Slope reduction may have been brought about by tectonics, hydro-isostatic effects, eustatic effects or backwater influence from evolution of a large downstream tributary (the Strickland River). Rather than indicating highly active channel migration, the well preserved floodplain features instead give evidence of low rates

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of overbank deposition, probably less than 1 mm a^{-1} in the forested reach and of the order of 0.1 mm a^{-1} in the swamp grass reach during the late Holocene. Floodplain deposition occurs by advection of sediment with overbank flows, by lateral diffusion from the sediment-rich Fly main channel, and by transport up tributaries or small channels connecting the river with off-river water bodies (tie channels). Low rates of sedimentation in the swamp grass reach (despite long periods of inundation) result from: (1) low concentration of sediment in the overbank flows; (2) dense swamp grass vegetation which may damp turbulence of advecting overbank flows and reduce diffusion; and (3) inhibition of river overbank flows across the floodplain due to pre-existing standing water resulting from local rainfall and low drainage rates. Prevention of tie channel plugging with sediment may depend on rather rare hydrological events, perhaps linked to drought-producing El Niño events which cause large pressure head differences between the river and the off-river water bodies and enable the strong currents in the tie channels to scour. The observed downstream fining of the Fly can result from only about a 15% reduction in the total sand load which may occur by preferential net deposition of the coarser sand in the channel bed and surrounding floodplain.

INTRODUCTION

Large river systems may show significant lags in response to changes in controls due to long travel distances. Church and Slaymaker (1989), for example, have argued that rivers draining western Canada are still passing the increased sediment load arising from glaciation which terminated over 10 000 years ago. Fluctuating sea level has forced waves of incision and aggradation on rivers, and the current morphology of large lowland rivers may strongly reflect responses that are still evolving to the current sea level high stand (e.g. Ikeda, 1989). An analysis of any large lowland river draining to the sea, then, must face the issue of sorting out the evolutionary relicts of the river's response to past sediment load, water discharge and sea level rise from the current dynamic linkages between morphology and current conditions. Importantly, such rivers may be responding simultaneously to large changes created in the Holocene and to recent effects associated with land use, and these effects may force opposing or reinforcing morphologic adjustments on the river.

The Fly River in Papua New Guinea (Figures 14.1 and 14.2) is currently the subject of intense investigations directed towards documenting and predicting the geomorphic and ecological effects of sediment waste from a large copper mine that has been operating since 1985 in the headwaters of the river (e.g. Eagle et al., 1986; Higgins et al., 1987; Pernetta, 1988; Higgins, 1990; Smith and Hortle, 1991; Wolanski and Eagle, 1991; Day et al., 1993; Smith and Bakowa, 1994). Of particular concern with regard to the geomorphology have been the effects of increased sediment load on the long-term planform stability of the river, the rate of channel aggradation, and the rate of floodplain sedimentation. A monitoring programme has been established to assess hypotheses developed from earlier and ongoing studies and to guide ongoing theoretical development directed towards predicting long-term effects of the mine discharges. In this paper we re-examine previously collected field data, report new findings and, through some simple estimations, reach conclusions that in some cases differ significantly from those of previous studies. Here we will focus on the pre-mine behaviour of the system which had been negligibly influenced by human activity. The river geomorphology has been strongly influenced by Holocene sea level rise, therefore distinguishing historical relicts from ongoing tendencies will be an important issue in our analysis. We specifically

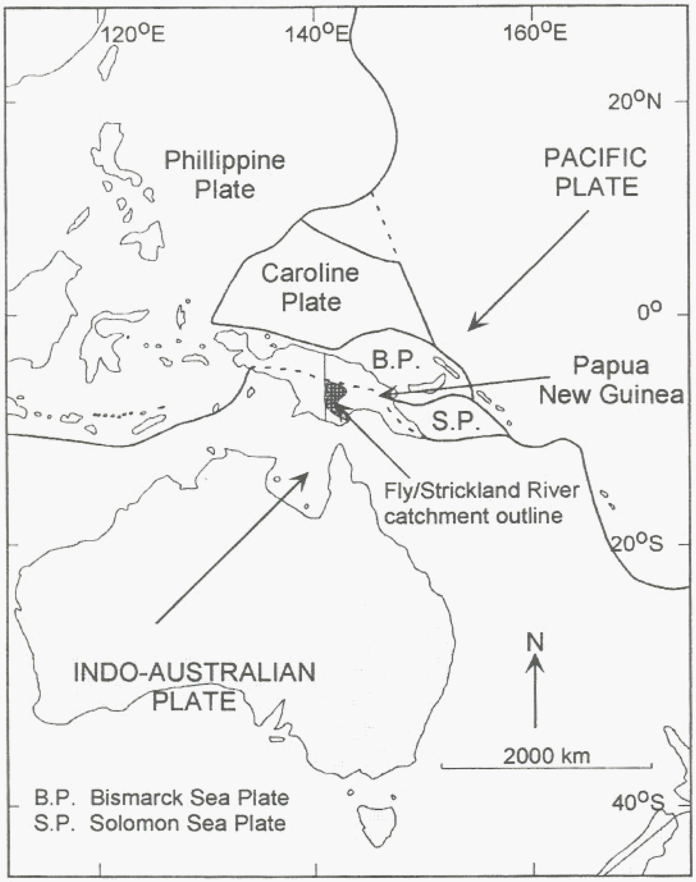

FIGURE 14.1 Plate tectonic framework of the Fly River catchment in Papua New Guinea showing relative plate motion (modified from Hill, 1991)

examine three general questions: (1) what controls the rate of channel lateral migration? (2) what controls the rate of deposition of sediment on the floodplain? and (3) what controls downstream fining of bed material in lowland sandy rivers?

THE FLY RIVER SYSTEM

The overall character of the Fly River system owes much to its tectonic setting. According to Pigram et al. (1989), middle Oligocene collision of the northern carbonate shelf margin of the Australian craton with a subduction system led to the formation of a foreland basin. This caused a broad flexure of the margin; as the northerly migration of the plate pushed into a tropical climate, an extensive carbonate platform formed. By the late Pliocene and Quaternary, sediment shed from the mountains overwhelmed the foredeep and clastic deposition, and rapidly spread southward burying the carbonate platform. Continued tectonic activity has caused previously buried Quaternary sediments

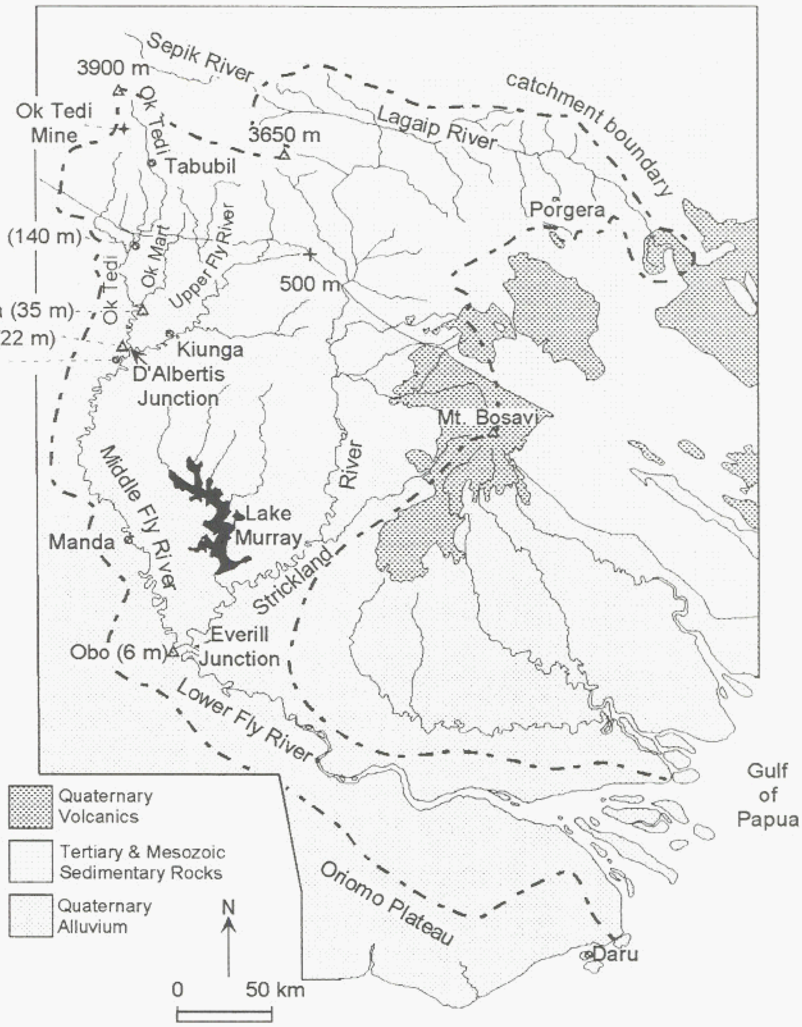

FIGURE 14.2 Fly River catchment showing very generalized geology (modified from Bureau of Mineral Resources, Geology and Geophysics, Canberra, ACT, 1976). Numbers in parentheses next to site names are elevations of the bank. (Note that "Ok" means "river" in the local dialect)

to be uplifted and, consequently, dissected, such that while the topography across the former carbonate platform is subdued, it is locally highly dissected (Loffler, 1977). Based on exploratory drilling, about 1000 m of Quaternary sediments mantle the carbonate platform at Ningerum (on the Ok Tedi, Figure 14.2) progressively thinning to about 280 m at Lake Murray (Taylor, 1979). This suggests that the deposition rate has been of the order of 1 mm a^{-1} near the mountains, decreasing to about 0.1 mm a^{-1} towards the Gulf of Papua.

This tectonic setting leads to: (1) high sediment loads in the headwaters; (2) steep, massively unstable topography which can shed large pulses of sediment to rivers; (3)

mechanically weak sediment which quickly breaks downs upon transport in the rivers (i.e. Parker, 1991); (4) orographically induced high rainfall rates; (5) a rapid decline in river slope and an overall very low gradient in the lowland bordering the mountains; (6) possible tectonic influence on river orientation (Blake and Ollier, 1969, 1971; Blake, 1971); and (7) limited net storage of sediment on land upstream of the Gulf of Papua.

The Fly River system consists of three major tributaries, the Ok Tedi, Fly and Strickland, all of which originate in the steep, rapidly uplifting subduction zone complexes and marine sediments of the Southern Fold Mountains where peaks reach up to 4000 m in elevation (Figure 14.2). These tributaries quickly descend through narrow canyons, crossing the uplifted Quaternary sediments at about 500 m, and join in the lowland area enroute to the delta. Where the Ok Tedi meets the Fly, the elevation of the water surface at bankfull is about 18 m (the local bank height for Kuambit reported in Figure 14.2 is on a terrace), and the river is 846 km from the delta mouth. At this junction (known as D'Albertis Junction), the drainage area of the Ok Tedi is 4350 km^2 and that of the Fly is 7600 km^2. The Fly then follows a 412 km sinuous path to its junction (Everill Junction) with the Strickland River. Here the Strickland drains 36 740 km^2 while the Fly drains only 18 400 km^2; this notwithstanding, the main stem below the junction is called the Fly River. Downstream of Everill Junction the Fly River gains an additional 19 860 km^2 before discharging into the Gulf of Papua. Of the 75 000 km^2 of land that drains via the Fly to the Gulf, only about 30% occupies the steep headwaters area.

Rainfall depends strongly on elevation, with annual rainfall in excess of 10 m occurring at the mine site on the Ok Tedi (elevation about 1500 to 2000 m), and decreasing from 8 m at Tabubil (600 m) to about 2 m at Obo and Daru near sea level (Markham, 1995; Harris et al., 1993). A distinct seasonality occurs in the southern half of the catchment with June to August tending to be drier. Despite heavy rainfall, droughts still strike, particularly in the lowlands, causing the Fly to fall and drain its normally swampy floodplain.

River flows increase downstream with increasing drainage area and become less flashy such that along the Middle Fly (the reach of the Fly between D'Albertis and Everill Junctions), flood stage may remain approximately constant for many months during which time the floodplain will be under several metres of water. We have observed continuous standing water in the swamp reach of the Fly floodplain for more than 18 months, and according to Higgins (1990) much of that water may originate from local rainfall and runoff. Average annual flow on the Fly below D'Albertis Junction is $1930\,m^3\,s^{-1}$ (Higgins, 1990), about $2244\,m^3\,s^{-1}$ at Obo just above Everill Junction and about $3110\,m^3\,s^{-1}$ on the Strickland. Hence, the average annual flow at the delta is less than $7000\,m^3\,s^{-1}$ (not the $15\,000\,m^3\,s^{-1}$ proposed by Blake and Ollier, 1971). Tidal influence on stage extends upstream of Everill Junction, and perhaps, during low river stage, as far as Manda, 570 km upstream from the Gulf.

Before the start of mining on the Ok Tedi, the sediment discharge from the Ok Tedi and Upper Fly was about $5 \times 10^6\,t\,a^{-1}$ each (based on early sediment monitoring records), but was subject to large temporal variations due to occasional massive landsliding (Pickup et al., 1981). Little sediment appeared to be added between D'Albertis Junction and Everill Junction, despite the drainage area increasing from 11 950 to 18 400 km^2 along this reach, because this low-lying portion of the catchment has gentle topography, is densely vegetated, and the tributaries are mostly blocked and empty into lakes as they enter the floodplain area. Instead, net sediment loss may have dominated at this time. According to various estimates and direct monitoring, the sediment load from the Strickland is about 70 to 80 $\times 10^6$ t a^{-1}, and while there is no doubt that the Strickland sediment yield was much higher than the Middle Fly before mining began, the uncertainty on this estimate is large. Perhaps less than 30% of this load is coarser than silt. The natural sediment discharge from the Fly above Everill Junction is relatively high compared to other rivers of comparable drainage area, but low compared to other rivers with comparable runoff (from data reported by Milliman and Syvitski, 1992). By implication, this means the sediment concentration in the Fly must have been fairly low, even though it is chronically turbid. On the Fly just below D'Albertis Junction the mean sediment concentration was about 100 mg l^{-1}, whereas on the Strickland upstream of Everill Junction it was about 770 mg l^{-1} (average of 53 depth-integrated samples collected during 1991 to 1996 by personnel working for Ok Tedi Mining Ltd). Figure 14.3 shows the downstream variation in runoff and sediment yield throughout the Ok Tedi–Fly system. On a per unit area basis, sediment yield and runoff decrease, illustrating how the uplands area is the dominant source for both water and sediment.

While bed material size on the Ok Tedi–Fly River declines downstream with decreasing slope, it does not do so uniformly (figure 2 in Pickup, 1984; Parker, 1991). Pickup proposed that five bed material zones, bounded by sharp transitions, make up the downstream variation in bed sediment characteristics. They are: (1) source (boulders and gravel, poorly sorted, close to sediment supply); (2) armoured (gravel over sand-rich

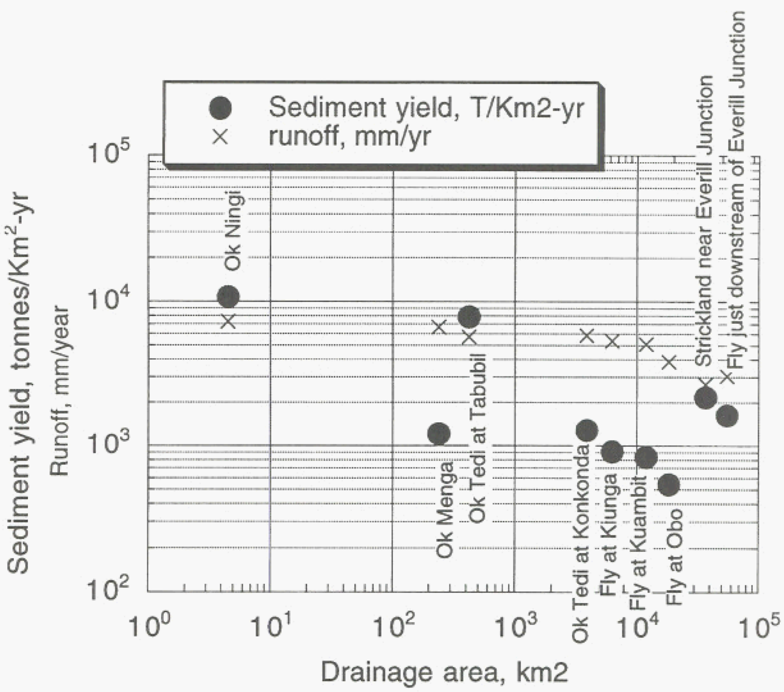

FIGURE 14.3 Downstream variation in sediment yield and runoff along the Ok Tedi and Fly River. Data are from Pickup et al. (1981) and OTML hydrology reports

substrate); (3) gravel–sand transition (bimodal); (4) sand (mean size about 0.2 to 0.3 mm); and (5) backwater (mean size less than 0.16 mm, but fluctuates with stage). Although he did not specifically locate these zones, they roughly correspond to: (1) the Ok Tedi above Ningerum (see Figure 14.2 for location); (2) Ningerum to above the Ok Mart Junction (at Konkonda); (3) a short reach just upstream of the Ok Mart; (4) Ok Mart to roughly half-way to Everill Junction; and (5) the remainder of the Fly to Everill Junction. Pickup suggested that slope, sediment supply and geomorphic history enforce these sharp bed material breaks. The backwater zone was proposed to be due to Strickland water level at its confluence with the Fly River. Based on a simple mass balance, Pickup estimated that net deposition of all sand at the distal end of the zone would cause the sand zone to reach the junction with the Strickland between 1400 years and 22 000 years if the bed level of the Strickland were to remain constant.

Blake and Ollier (1971), in a pioneering examination of the Fly River system, recognized several important attributes of the floodplain of the Fly. They distinguished five distinct floodplain environments: scroll complexes; back swamps (similar to Allen's (1965) floodbasins); alluvial plains of minor tributaries; blocked valley swamp; and lakes (Figure 14.4). Blake and Ollier proposed that the sediment-laden Fly and Strickland Rivers aggraded during Holocene sea level rise, blocking all small tributaries draining from the lowlands, and creating blocked valley lakes (including Lake Murray) and swamps. The scroll complexes were inferred to be aggraded relative to surrounding plains (Blake and Ollier, 1971, p.5) giving rise to the back swamp areas. They identified numerous geomorphic features on aerial photographs and introduced the term "tie channel" to describe the small channels that almost invariably connect the main stem Fly River to cutoff and blocked valley lakes and through which the water may flow in either direction. They conclude that the well formed, numerous scroll bar complexes on the Fly indicate that "the alluvial plains are highly unstable and that lateral migration and accretion dominate over vertical accretion". Loffler (1977), Pickup et al. (1979) and Pickup and Warner (1984, p.39) agree with this inference of active, rapid channel migration on the Fly. We will re-evaluate this interpretation below.

Blake (1971) reported a radiocarbon age date of 27 000 years BP in sediments found near Kiunga (Fly River above D'Albertis Junction) which he believes are overlain by a regionally extensive alluvial unit referred to as the Lake Murray beds. Blake and Ollier (1969) further proposed that the land mass south of the Fly River (the Oriomo Plateau, Figure 14.2) was uplifted 15 m after deposition of the Lake Murray beds and that this uplift contributed to forcing an ancestral Fly River (which they assumed headed southwestwards to the sea) to head eastwards to the Gulf of Papua. Subsequently Loffler (1977, p.20, 98) questioned whether the regional stratigraphy was sufficiently understood that a single date can be widely applied. He argued that the fans from the volcano Mount Bosavi (Figure 14.2) overlie the broad alluvial plain proposed by Blake to have been dissected in the last 27 000 years, and yet the Bosavi volcano construction must extend well into the Pleistocene. Finally he argued that the rate of erosion required in 27 000 years to generate the fine-scale ridge and valley topography that borders the Fly–Strickland north of the Oriomo Plateau would greatly exceed that expected from a low-lying alluvial plain. Instead, he proposed that the Fly River entrenchment and dissection of the surrounding hills occurred repeatedly during the Pleistocene as sea level fluctuated with the glacial cycles. Because other rivers turn from a southerly direction to an easterly

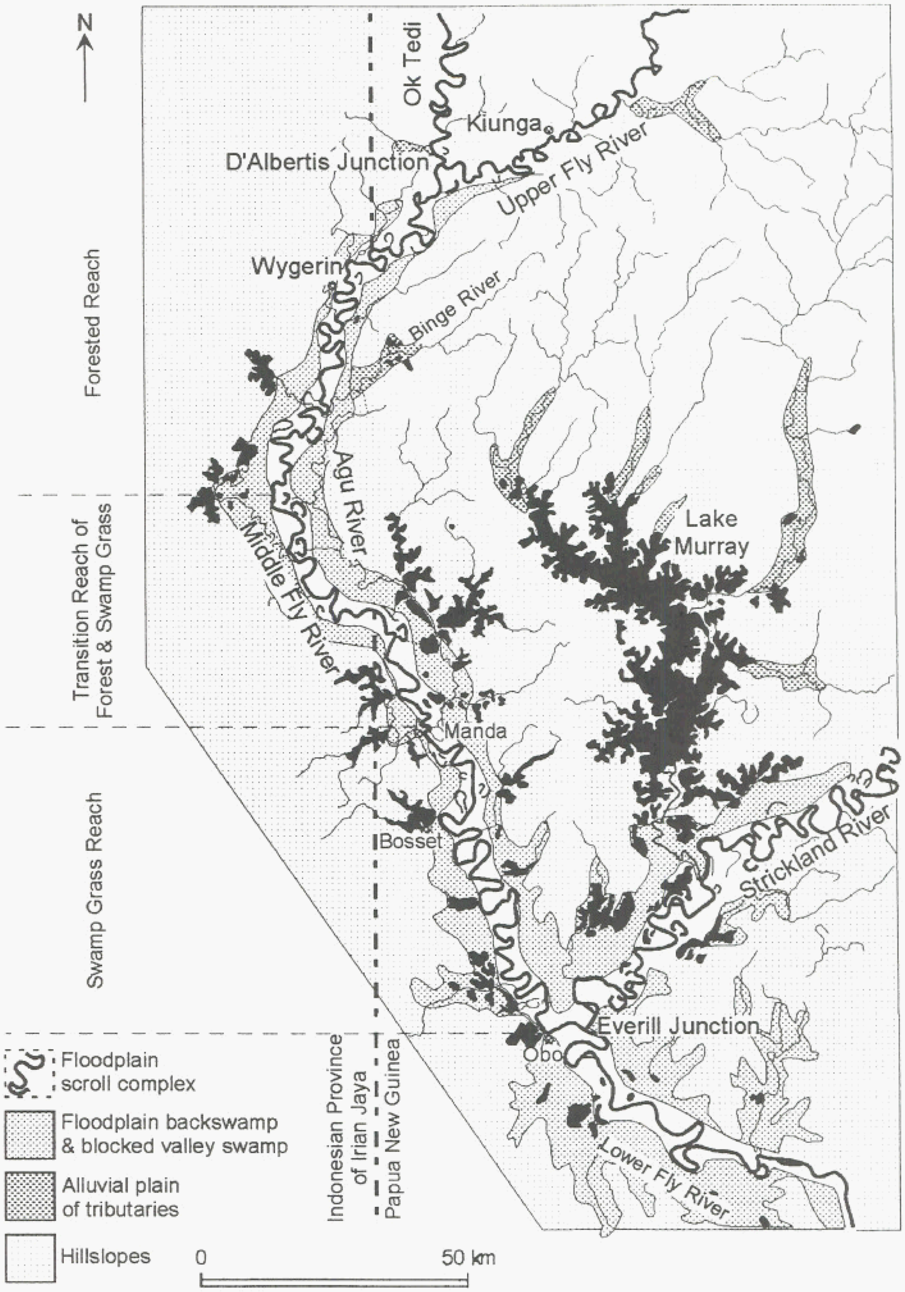

FIGURE 14.4 Generalized depositional environments of the Fly River and lower Strickland area (modified from Blake and Ollier, 1971)

direction to the sea in this area, he also reasoned that the Fly River was not turned by recent uplift of the Oriomo Plateau, but instead has simply followed the "predominant slope of the depositional surface" of the alluvial plains (Loffler, 1977, p.90). Harris et al. (1996) reported that deep incised valleys, up to 120 m deep, extend across the continental shelf into the Gulf of Papua, indicating that the Fly River has flowed eastwards since at least the last glacial maximum.

Pickup et al. (1979) used the Blake (1971) observation and the presence of weathered alluvium in the bed of the Fly River near Kiunga to infer that the Fly River cut down approximately 10 m between 27 000 and 17 000 BP and then aggraded to its present level by 5000 BP. They concluded that the river aggraded at about 1 mm a^{-1} during sea level rise (a rate similar to the longer time scale deposition recorded in the Quaternary sediments). Taylor (1979) also relied on the Blake (1971) observation as well as data that indicate cessation of volcanism at Mount Bosavi about 30 000 to 50 000 BP to estimate an average aggradation rate in the Middle Fly River area of about 1 mm a^{-1}. Dietrich (1988), using these data and other inferences, reasoned that the aggradation rate decreased from about 1 mm a^{-1} to about 0.1 mm a^{-1} down the Middle Fly. Hettler and Lehmann (1995) reported a similar conclusion.

Harris et al. (1996) reviewed current understanding of Late Quaternary sea level changes in this area, pointing out that over much of the last 100 000 years eustatic sea level has been in the range of 40 to 70 m below its present position. They also suggested that the Holocene transgression may have experienced brief periods of rapid rise and that current eustatic sea level was reached about 6500 years ago. Isostatic adjustments due to rising water levels (i.e. Chappell et al., 1982) and sediment loading on the shelf probably have occurred, but their magnitudes are not yet established.

In the following, we examine more closely the spatial and temporal patterns of channel migration, floodplain deposition and grain size variation along the Middle Fly observed by earlier workers.

RIVER PLANFORM DYNAMICS OF THE MIDDLE FLY RIVER

Along the Middle Fly River from D'Albertis to Everill Junction, the river has a sinuosity of 2.1, and within the distinct meander belt there is abundant evidence of past channel position, including complex scroll patterns and filled cutoffs and cutoff lakes connected to the main channel via tie channels (Figures 14.5 and 14.6). The floodplain and meander belt progressively widen downstream from each being about 4 km at Kiunga to 14 and 8 km, respectively, by Everill Junction (Pickup et al., 1979). Channel width, on the other hand, is about 200 m at Kiunga, increases to an average width of about 350 m within 90 km downstream of D'Albertis, and then narrows downstream thereafter to about 250 m before reaching Everill Junction (Pickup et al., 1979). Vegetation is an uninterrupted rainforest from D'Albertis Junction through the reach bordering the Indonesian province of Irian Jaya. This gradually gives way to a swamp grass reach where the Agu River joins the Fly (near Manda) and continues as such until after the Strickland enters the Fly (Figure 14.4).

Throughout the Middle Fly reach, the meanders are not entirely free; occasionally, along the outer banks of bends, higher banks composed of intensely weathered, bright red alluvium are exposed which clearly offer greater resistance to channel migration. Pickup

FIGURE 14.5 Scroll bars and oxbows of the swamp grass reach of the lower Middle Fly. Not all scroll bars shown. Based on aerial photographs take in 1982. Flow direction is from top to bottom of figure. Arrow is located in the first bend downstream from where the Agu River joins the Fly (at a bend apex on the Fly)

et al. (1979) proposed these banks to be composed of Pleistocene sediment, which seems very likely. There is also a distinct change in the meander morphology along the river. In the forested upper Middle Fly, meanders have typical shapes and wavelength-to-width ratios comparable to those of most rivers (about 11), but in the swamp grass reach this ratio grows to 20 and the bends tend to consist of long straight limbs attached to short, highly curved bend apices reminiscent of meanders in muskeg (Figures 14.4–14.6).

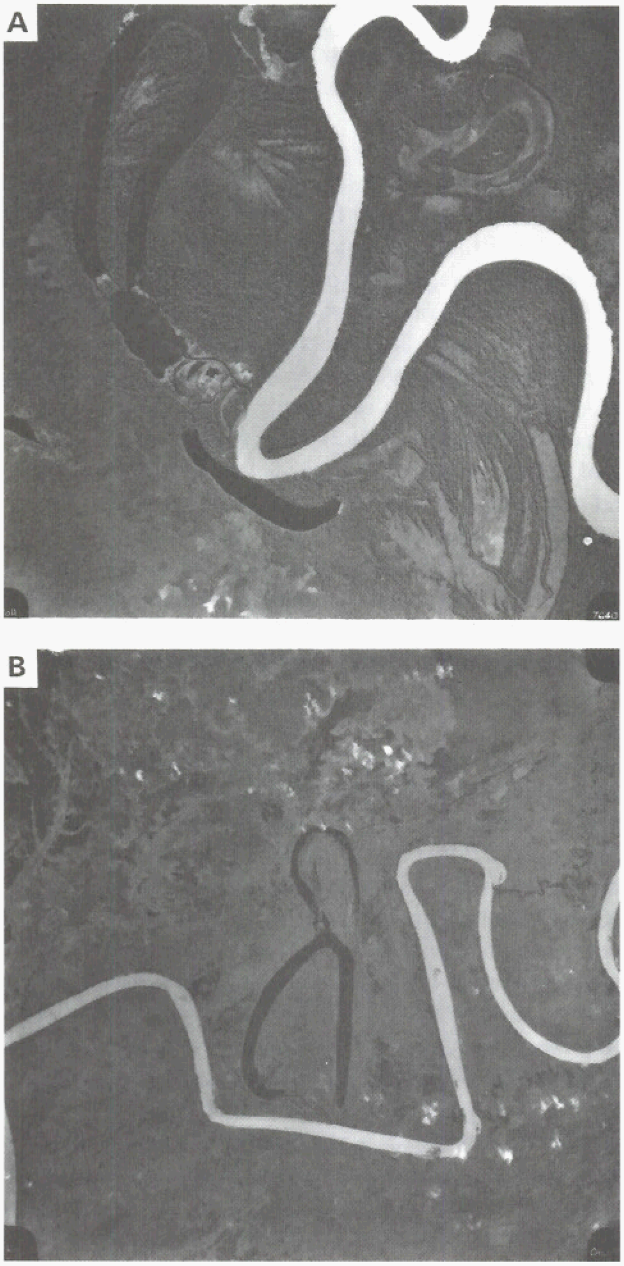

FIGURE 14.6 Aerial photographs of the Fly in the forested reach (A) and swamp grass reach (B). In (A), which is downstream of the Binge River (Figure 14.4), oxbows in various stages of infilling are visible and scroll bars are recorded by long curved rows of trees. A distinct tie channel joins the Fly with the oxbow lake. In (B), which is just upstream of the Obo gauging station (at bottom of Figure 14.5), a double oxbow with typical bends lies in contrast to the dogleg pattern of the current Fly River. A tie channel connects the oxbows to the Fly and scroll bars dominate the fine-scale topography within the meander belt

In order to determine migration rates and assess the inference that the river "meanders vigorously", three different measurements were made. First, a crude assumption, but one for which there is evidence on other rivers (e.g. Fisk, 1944), is that the sinuosity of the river and resulting slope are relatively constant over time. In this case, the number of meander bends for a given reach would stay approximately constant, with reaches that experienced cutoff quickly recovering their sinuosity and other bends increasing in size. Hence, if a cutoff occurs, the loop will be replaced, and if we know the total number of bends and the rate of cutoff, we can calculate the mean migration rate.

On the Middle Fly, according to Pickup et al. (1979, p.104), three loops were cut off since 1899, suggesting a cutoff rate of three per 90 years or one every 30 years. Between D'Albertis and Everill Junctions there are 111 individual bends (over 400 km). If these cut off at the rate of one in 30 years and are immediately replaced with new loops such that this number stays constant, then on average an individual loop forms and is then cut off in 111×30 years or 3330 years. Cutoff loops typically have an amplitude of about 1.5 km at the time of cutoff. Hence if the typical loop grows to 1.5 km in 3330 years, its maximum lateral migration rate at the bend apex is on average about $0.5\ \mathrm{m\,a^{-1}}$. Studies of meander loop evolution typically show that they tend to first grow and migrate downstream and then enlarge laterally, slowing down as the loops become large (e.g. Hooke (1984), Larson (1995) and as implied in the Nanson and Hickin (1986) relationship between bank erosion rate and radius of curvature to width ratio). The average migration rate of $0.5\ \mathrm{m\,a^{-1}}$, then, is probably low compared to the early stages of actively migrating loops, and gives a minimum average rate of about 0.002 channel widths per year.

The second method of estimating channel migration rates consisted of comparing topographic maps. Two high quality maps were available: 1:100 000 topographic maps produced by the Royal Australian Survey Corp in 1969 and based on aerial photography from 1963 to 1966, and 1:50 000 navigational charts produced by the Snowy Mountains Engineering Corporation (SMEC) in 1981 and modified from the Australian Survey Corp maps using 1980/81 Landsat Imagery and ship radar observations during surveying. Although of relatively high quality, these data have unknown positional inaccuracies due to the lack of extensive ground control. The strength of our findings reported below relies on the comparison among the three methods. Older, less reliable maps included: 1:250 000 US Army Map Service maps from aerial photography and field observations made in 1953–1963, and 1:500 000 US Army Map Service dated 1942 and compiled from many sources including the Netherlands New Guinea and Papua maps at 1:250 000 and US Hydrographic charts. We also found a 1:500 000 1943 map captured from the Japanese after World War Two that was modified from an Australian 1:1 000 000 map printed in 1941. This map was particularly important for examining the swamp grass reach. Aerial photographs dating from the 1930s apparently exist but we have not yet been able to gain access to them.

Channel migration analysis was divided into three distinct reaches: Kiunga to D'Albertis Junction; D'Albertis Junction to the upstream end of the swamp reach at the southern end of the Indonesian border; and from the upstream end of the swamp reach to Everill Junction. Of the 18 bends between Kiunga and D'Albertis only 30% had measurable displacement. Average shifting in these active bends over an approximate 20 year period was typically about 1 to 1.5% of channel width (or about 2 to $4\ \mathrm{m\,a^{-1}}$).

Comparisons using the older Japanese maps were less certain because of clear mapping errors; nonetheless, these maps also support the interpretation that migration is occurring at a measurable pace in only a fraction of the bends and the rate is less than 2% of channel width per year in these bends. Hence, average channel migration rate overall in this reach is estimated to be less than $0.02 \times 0.30 = 0.006$ channel width per year and may be as low as 0.003 channel width per year.

Between D'Albertis Junction and the end of the Indonesian border upstream of Manda, US Army maps were compared to the SMEC maps to determine migration rates. Of the 62 bends along this international border, about 30% showed evidence of measurable displacement, with most of these bends in the southern part of the reach. Average shift rate where it occurred was about 1.5% of channel width per year (or about 4 to $5\,\mathrm{m\,a^{-1}}$). This pre-mine estimate for channel migration rate is about $0.015 \times 0.30 = 0.0045$ channel width per year.

In the swamp grass reach, we found the Japanese maps to be reasonably accurate and only minor differences were found between these maps made in the early 1940s with those published in 1981. Comparison of the Australian Royal Survey Corp maps with the SMEC navigation charts showed only one bend other than a short reach affected by a cutoff to have any measurable displacement. Nearly the entire reach has shown no perceptible migration for over 50 years.

The third method of analysis became possible when aerial photographs were taken in 1992 and compared to 1982 aerial photographs obtained by BHP Engineering. This work was accomplished by Barr Engineering Co. (1995) by digitizing channel centrelines and analysing channel shift using a program called MEANDER, originally described by MacDonald et al. (1991). Barr Engineering divided the distance from Kiunga to D'Albertis Junction into three reaches, and found the average shift normal to the downstream direction similar in each, averaging 0.0036 channel width per year. Downstream of D'Albertis, they divided the 200 km distance to the start of the swamp reach into eight separate analyses of migration. Figure 14.7 summarizes their findings, showing that migration rate was found to be highest just downstream of D'Albertis Junction and declined to values of about 0.0047 channel width per year at the downstream end. Shortly below their lowest reach, the swamp reach begins and the migration rates drop to very low values. Barr Engineering interpreted the elevated migration rate downstream of D'Albertis Junction as resulting from bar growth due to introduction of mine waste to the Ok Tedi, which began in earnest in 1985. Whereas this is probably correct, this pattern of accelerated bank migration downstream of a tributary that introduces significant bedload can occur naturally. This is precisely what happens downstream of Everill Junction, where significant channel migration has been previously noted (SMEC, 1981, p.52, figure 17). In contrast, in the swamp reach upstream of Everill Junction, bars are normally absent and channel migration rates are presently nearly zero. Hence, the pattern of downstream decreasing migration rates below D'Albertis probably existed before, but has been exaggerated by the introduction of mine-related sediment to the system.

Comparison of these three estimates of migration rates gives remarkably similar results (especially given the assumptions employed and inaccuracies of the first two methods). Cutoff rates give a minimum estimate of 0.002 channel width per year. Comparison of topographic maps up to 1981 indicates migration rates between 0.003 and 0.006 for the

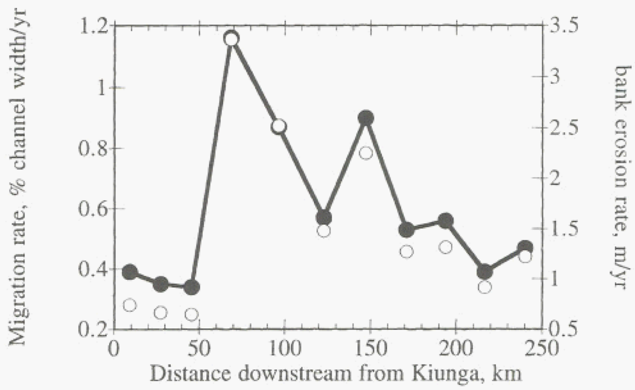

FIGURE 14.7 Channel migration rate downstream of Kiunga shown as metres per year (open circles) and percentage of channel width per year (solid circles) (data from Barr Engineering Co., 1995)

reach upstream of D'Albertis, 0.0045 below D'Albertis to the swamp reach, and no migration in the swamp reach. Repeated recent aerial photographs give values of 0.0036 channel width per year above Kiunga, and an accelerated rate (due to mine-related sediment) below D'Albertis Junction which declines to 0.0047 channel width per year in the unaffected lower reach above the swamp reach.

Despite the well preserved evidence of lateral migration of the Fly, the documented migration rates do not reveal "intensive meandering activity" that could cause "rapid floodplain destruction" (Pickup and Warner, 1984, p.39) as many have suggested. The migration rates are low, or, in the case of the swamp reach, essentially zero. An analysis of migration rates of 18 rivers in western Canada by Nanson and Hickin (1986) gave a median migration rate of about 0.015 channel width per year. A compilation by MacDonald et al. (1991) of 16 streams in Minnesota, USA, gave values of migration per unit channel width per year of 0.0044 to 0.066, with an average value of 0.0187. Most of the rivers in the two studies were narrower rivers conveying smaller discharges. Two reaches on the Mississippi River, however, which had widths of 56 and 67 m, had migration rates of 0.005. Larsen (1995) reviewed migration rates of a number of rivers (besides the MacDonald et al. sites) including several that had widths greater than 100 m. None of the rivers greater than 100 m wide had migration rates less than 0.013 channel width per year, a value much higher than the mean pre-mining rates on the Fly River. Empirically, then, the migration rates for the Fly fall within the low range of values reported for other rivers.

If the strong preservation of scroll bars and old channel paths in the floodplain cannot be explained by rapid or vigorous lateral migrations, then their preservation instead suggests that the rate of overbank deposition is sufficiently low that floodplain features are not obliterated in spite of the slow channel migration. On the Fly above the swamp grass reach, if the average lateral migration is 0.004 channel width per year, the mean channel width is 240 m and the meander belt width is 6 km, then to traverse half that distance at this rate (assuming that alternate bends migrate in opposite directions) would take about 3000 years, roughly the time estimated for a loop to form and cut off. The traverse time would be shorter if the peak migration rate during loop expansion is higher, as it probably is. If cutoff loops do represent 3000 years of floodplain development, then historical rates of floodplain overbank deposition must be less than 1 mm a^{-1} as 3 m would probably eradicate much of the scroll bar relief so prominent on the floodplains.

This issue is particularly pertinent to the swamp reach upstream of Everill Junction where very little bank migration was detected for the past 50 years, yet scroll bar features, cutoff loops and old channel paths etch the floodplain (Figure 14.5). This seeming contradiction suggests that the rate of channel migration has reduced significantly.

What would cause lateral migration in a large lowland river to nearly cease? Howard (1992) reasoned that four factors constrain bank erosion rates: (1) rate of deposition on the point bar; (2) ability of the stream to remove the bedload component from eroded bank deposits; (3) ability of the stream to entrain cohesive bank slump deposits; and (4) rate of weathering of cohesive bank materials. It appears that constraint 1 would at least partially explain the accelerated bank migration rate below D'Albertis Junction. In the swamp grass reach there are no well developed point bars, but the channel is relatively narrow, having a width/depth ratio of 15 or less, hence local flow velocities are not significantly diminished due to a lack of point bar confinement. Constraint 4 is unlikely to undergo the apparently relative rapid transition suggested by the preservation of floodplain features. So this points to a decreased ability of the river to entrain (constraint 3) and/or transport (constraint 2) in-place or fallen bank material.

Howard (1992) reviewed the basis for the widely used bank erosion law that assumes the erosion rate (E) is proportional to the near-bank velocity perturbation (u_b), an assumption originally proposed by Ikeda et al. (1981), i.e

$$E = Ku_b \tag{14.1}$$

$$u = U(1 + u_b) \tag{14.2}$$

in which K is a proportionality constant, u is the local vertically averaged velocity and U is the average velocity for the channel cross-section. While field observations support this assumption, Howard also pointed out that Odegaard (1989) and Hasegawa (1989) argue that depth perturbation near the outer bank may be a better or at least an equally important predictor of bank erosion rates. Furthermore, Howard reasoned that there is likely to be a critical near-bank shear stress below which bank erosion ceases, hence the simple velocity perturbation assumption may be incomplete. Alternatively, then, one might formulate Equation (14.1) using a critical bank velocity, u_c:

$$E = K(u_b - u_c) \tag{14.3}$$

Such a model would provide a direct mechanism for slowing or even halting channel migration through either increases in bank resistance or decrease in the perturbation velocity. The most likely case is reduction in perturbation velocity due to reduction in mean slope. If a river was actively migrating and experienced a slope decline (due to local base level rise, for example), then mean velocity might decline and with that the perturbation velocity. If the slope declined progressively, the perturbation velocity would remain in excess of critical for progressively smaller lengths of channel until the entire bend ceased to migrate. Such a decline could lead to a morphological change as well, from the typical bends of actively migrating rivers, to the dogleg bends with large wavelength-to-width ratio of the present swamp reach.

On the Fly River, average slope decreases from about 6.6×10^{-5} near D'Albertis Junction to about 2×10^{-5} in the swamp reach (based on SMEC (1981) data). Correspondingly, the bankfull boundary shear stress declines from about 80 to $30\,\mathrm{dyn\,cm^{-2}}$. In addition, for a given discharge (or for bankfull discharge), average velocity declines from Kuambit to Obo (Figures 14.2 and 14.8). It is possible, then, that earlier in the Holocene, the slope was steeper in the lower Middle Fly and then reduced to its current value. Slope reduction could have resulted from sea level rise, build-up of the Strickland and backwater development, or tilting due to tectonics. We briefly consider each of these controls.

After deglaciation at the end of the Pleistocene, sea level quickly rose to close to its current level by about 6500 years ago (as reviewed by Harris et al., 1996). We infer that the change in slope along the Fly is more recent than 6500 years ago because the floodplain features are well preserved (some preliminary dating of oxbow lakes appears to support this inference). Hence, sea level rise seems an unlikely mechanism. Build-up of the Strickland and development of a backwater up the lower Middle Fly was proposed by Pickup and Warner (1984) to explain the swampy conditions; they did not, however, comment on swamp reach migration rates. At Everill Junction, the Strickland drains twice the drainage area of the Fly and probably carries 10 times the sediment load. The Strickland floodplain near Lake Murray is frequently flooded, but forest lines the channel and spreads out into the surrounding floodplain, suggesting that it is better drained than the lower Middle Fly. A build-up of sediment in the Strickland would presumably occur progressively, and the backwater development could just be the result of ongoing

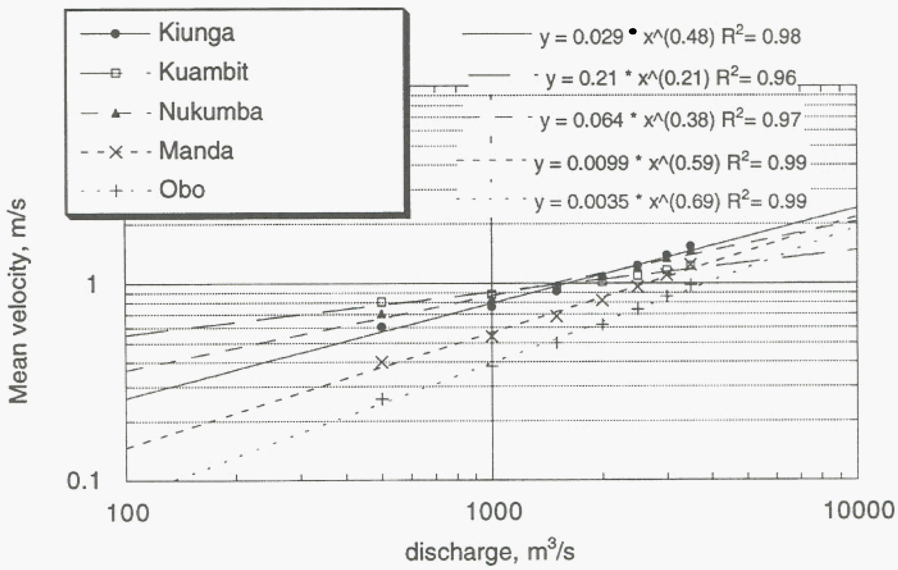

FIGURE 14.8 Variation in mean velocity with discharge at gauging stations along the Fly. Nukumba is about 5 km downstream of Kuambit and Manda is at the head to the swamp grass reach just upstream of where the Agu River joins the Fly

response of the Strickland to Holocene sea level rise. The bank elevation of the Fly falls only about 3 m over the roughly 152 km swamp reach, hence relatively minor aggradation of the Strickland could greatly influence the average slope.

One indication of backwater effects on the swamp grass reach is the stage dependent slope that can be calculated from hydraulic geometry relationships at gauging stations. Using Manning's relationship:

$$u = d^{0.67}S^{0.5}/n \tag{14.4}$$

solving for the slope, S, and replacing the mean velocity, u, and average depth, d, with the hydraulic geometry relationships (e.g. Leopold et al., 1964):

$$u = kQ^m \quad \text{and } d = cQ^f$$

gives

$$S = (kQ^m)^2 n^2/(cQ^f)^{1.33}$$

or

$$S = k^2 c^{-1.33} Q^{(2m-1.33f)} n^2 \tag{14.5}$$

For the Obo gauging station, Equation 14.5 is

$$S = 3.8 \times 10^{-7} Q^{1.35} n^2$$

showing that the slope is strongly discharge-dependent. A Manning's n of 0.03 gives reasonable estimates (ignoring stage dependency of resistance) of slope in this reach increasing from 3.8×10^{-6} to 1.7×10^{-5} as discharge increases from 1000 to 3000 $m^3 s^{-1}$. This strong stage dependency may reflect the varying influence of the Strickland on the Fly, whereby the effect is strongest at low flow, causing the swamp grass reach to have an exceptionally small slope. This analysis is consistent with field observations made during low flow in which we saw a dramatic decrease in suspended sediment concentration as the flow entered the swamp reach. While there is strong evidence that the Strickland creates a backwater on the Fly, it does not necessarily follow that this backwater is responsible for the apparent change in migration character of the swamp reach.

Given the location of the Fly River on the leading edge of the Australian plate, it is also not unreasonable to suggest that recent tectonics may have tilted or lowered the lower Middle Fly, causing it to become swampy and reducing its slope to the point where bank migration has nearly ceased. On the *Geology of Papua New Guinea* map (Bureau of Mineral Resources, Geology and Geophysics, Canberra, ACT, 1976), a fault is depicted as starting just upstream of the swamp reach and running parallel to the axis of the Fly valley, passing just downstream of Everill Junction and curving eastward following the Fly valley to where the river begins to widen rapidly as it enters the delta. How this fault, if it is correctly shown, affects the Fly River is not clear, but it does suggest that tectonics could affect this reach.

By whatever mechanism, if the slope has been reduced at the transition from the forested to the swamp grass reach, there should be a tendency for aggradation. We have recognized no obvious evidence for this. Furthermore, preliminary results of high resolution surveying of bank and floodplain heights using a global positioning system fail to show a significant slope break at this transition: a nearly constant gradient extends from nearly 100 km upstream of the transition to Obo at the lower end of this reach. This is consistent with the lack of a distinct zone of aggradation. The lack of aggradation at the transition reach raises the possibility that the higher channel migration in the past in the current swamp reach was due to historically higher sediment loads. According to Blong (1991), about 8800 BP a 7 km^3 landslide swamped the Ok Tedi near Tabubil. However, the majority of this pulse of sediment had probably passed through the system, however, well before the bank migration slowed in the current swamp reach.

At this time, although we can make a strong case that the pace of current channel migration is too slow to develop the complex floodplain features of the swamp reach, we cannot yet assign causality.

SEDIMENTATION RATES AND PROCESSES

Floodplain deposition

Geological evidence cited above suggests that long-term rates of floodplain deposition decline downstream along the Middle Fly from about $1\,mm\,a^{-1}$ to $0.1\,mm\,a^{-1}$. These low rates are compatible with preservation of scroll bar features formed by the slowly migrating Fly River. Futhermore, preliminary analysis of radiocarbon dating in the swamp reach supports the inference that sedimentation rates are on the order of $0.1\,mm\,a^{-1}$ (Hettler and Lehmann, 1995). Higgins (1990) estimated that 3% of the mine-derived sediment currently carried by the Fly River will be deposited on the floodplain. If this deposition rate is representative of the fate of the entire sediment load, then we can estimate a floodplain deposition rate. Mean sediment load of the river due to mine waste addition is about 50×10^6 tonnes annually and if 3% of that is spread out over the roughly $3300\,km^2$ floodplain then for a bulk density of $1\,t\,m^{-3}$ the deposition rate is $0.5\,mm\,a^{-1}$; for a bulk density of $1.7\,t\,m^{-3}$, it is $0.3\,mm\,a^{-1}$. Sampling of floodplain copper concentrations confirm these estimates and show that the aggradation rate is higher in the forested reach than the swamp reach (Day et al., 1993). These estimates are for the high sediment load associated with the mine which is roughly five times natural load.

Inferences regarding preservation of floodplain features suggest that the swamp grass reach, despite being chronically in flood, actually has the lowest floodplain deposition rate of the Middle Fly. Even before mining began, the sediment discharge of the Fly River above Everill Junction was higher than average for a basin of its size (Milliman and Syvitski, 1992, figure 2), and now it is very high. Why is the sedimentation rate of the floodplain not higher? What controls the rate of floodplain sedimentation?

Floodplain deposition is controlled by the frequency and duration of river flow onto the floodplain and the concentration and size of sediment carried by the river. Three distinct processes transport sediment from the Fly to the surrounding plain: advection by overbank flow; diffusion during flood stage; and transport up tie channels and tributaries. Advection by overbank flows depends on the relative heights of water in the river and on the floodplain. On the Fly, the strongest advective transport probably occurs when the water level across the floodplain is low and the river experiences a large flow generated from storms in the upper catchment. The probability of this occurring is much greater in the forested upper Middle Fly than in the poorly drained, chronically flooded swamp grass reach.

Once the swamp reach becomes flooded, there is probably little pressure gradient between the river and the flooded plain. Water is lost by evapotranspiration from the plain, but can be completely replenished by rainfall on the floodplain itself. In fact, during the flooded state in the swamp reach, it is common to see black, sediment-free floodplain waters bleeding into the river after a few hours of intense local rainfall has elevated the floodplain waters relative to that of the river. Figure 14.9A shows the river just at the transition from flooding onto and draining of the floodplain. On the far bank (flow is from left to right), light-coloured sediment-rich water has spilled behind the tree-lined levee a distance of about one-half the channel width across flooded swamp grass. On the right bank, an irregular boundary between light-coloured river water and black floodplain waters reveals the Fly River partly spilling sediment-rich water onto the plain and partly draining sediment-free floodplain waters.

Higgins (1990, p. 406) estimates that on average for only one month per year does "mainstream flow rather than local runoff contribute significantly to water stored in the floodplain". Hence, despite being flooded for periods in excess of 18 months, the swamp grass reach is not receiving a large advective transport of sediment. This must be a significant contributor to the low overbank deposition rates in this reach. It may also explain why the forested, better drained reach upstream has higher overbank deposition rates; frequency and duration of overbank flows onto the floodplain in this reach are probably greater than in the swamp reach.

A simple calculation emphasizes this point. Let us assume that all the water on the floodplain originates from the channel and has a concentration, C, the same as that in the channel. We can ask how deep, h, must this water be to provide enough sediment to deposit a specified thickness, d for a given bulk density of sediment, ρ_s:

$$C \times (1/\rho_s) \times h \times (\text{unit area of the floodplain}) = d \times (\text{unit area})$$

$$h = d \times \rho_s/(C) \tag{14.6}$$

If $C = 100\,\text{mg}\,\text{l}^{-1}$, $\rho_s = 1\,\text{g}\,\text{cm}^{-3}$, and $d = 0.1\,\text{mm}\,\text{a}^{-1}$, then $h = 1$ m. Clearly, then, if the Fly River were to spread across the entire floodplain just 1 m deep carrying sediment at the same mean concentration as in the channel, it could cause $0.1\,\text{mm}\,\text{a}^{-1}$ of sediment deposition. Certainly the concentration near the surface will be less than the mean for the flow as a whole and the bulk density will increase well above 1.0 with deep burial; both of these effects require deepening of the standing water needed to cause the required deposition. However, the floodplain waters are typically much deeper in the swamp grass reach, often being 3 to 5 m deep, yet the deposition rate is nonetheless probably much closer to $0.1\,\text{mm}\,\text{a}^{-1}$. For the swamp grass reach to have such a low deposition rate it seems reasonable to conclude that rainwater contributes significantly to flooding here and that this inhibits overbank sediment-rich flows.

Dietrich and Parker (1992), in an unpublished report, describe the findings of the first three years of an intensive sampling programme being conducted by G. Day to monitor the spread of mine-derived sediment across the Fly River floodplain. These preliminary data gave floodplain deposition rates identical to that estimated by Higgins (1990) of 3%

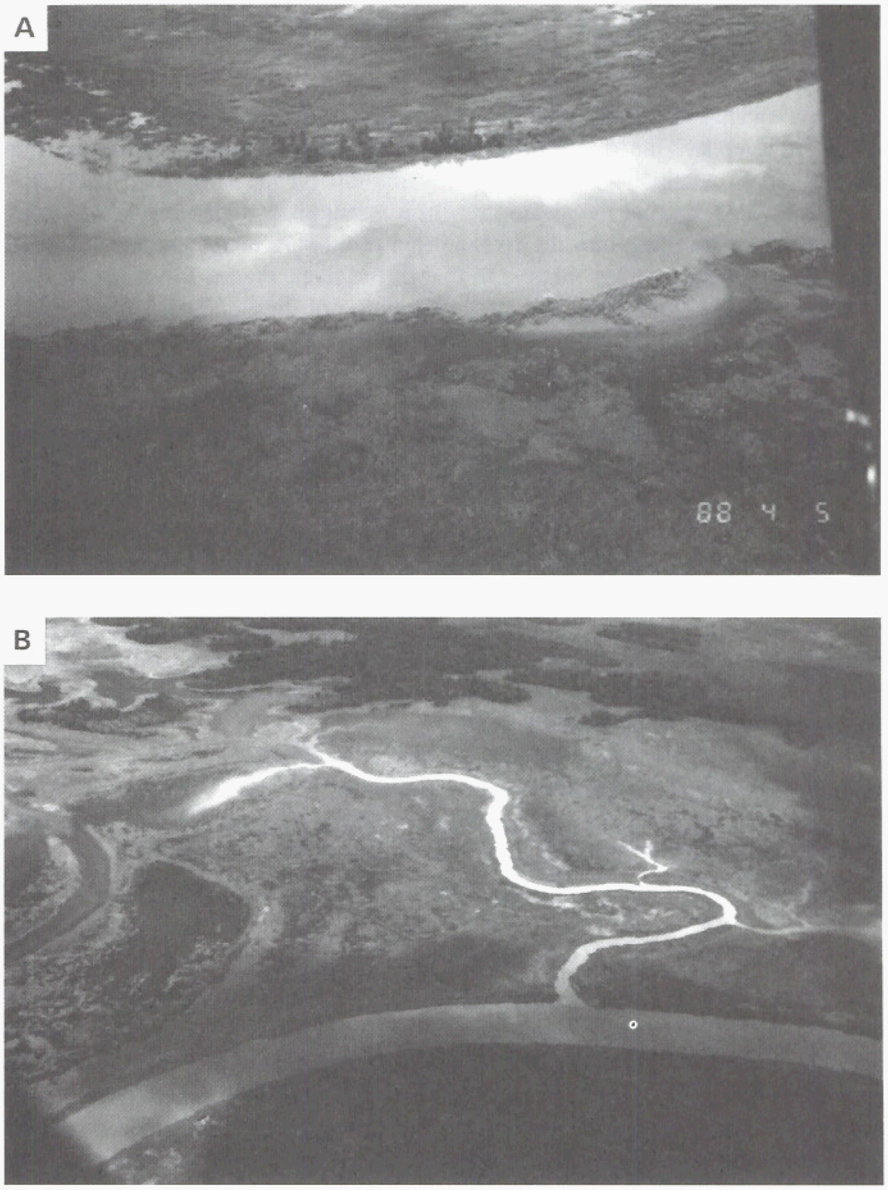

FIGURE 14.9 Photographs showing the penetration of light-coloured, sediment-rich Fly River waters onto the already flooded sediment-poor floodplain. (A) View of the Fly in the swamp reach taken in 1988 (flow from left to right). River-derived waters have reached a distance of about one-half the channel width onto the floodplain from the left (top of picture) bank. Along the right bank the boundary between dark floodplain-derived waters and light-coloured river waters is irregular, with limited overbank deposition in some areas and spilling of floodplain waters to the river elsewhere. The river is about 250 m wide. (B) View of the Agu River about 10 km upstream from the Fly junction. Here the Fly River water is travelling up the Agu and is spilling via a network of distributary tie channels into lakes on the adjacent swampy floodplain. The Agu is roughly 70 m wide

of the total load, and indicated that the deposition rate was roughly 1 to 2 mm a^{-1} in the forested reach and less than 0.5 mm a^{-1} in the swamp grass reach. They pointed out that the chronically high water level of the swamp grass reach would prevent the development of pressure gradients from the main channel to the floodplain and thus inhibit floodplain deposition there.

Recent analysis of remote sensing data in several large river systems around the world by Mertes (1997) suggests that the inhibition of sediment-laden overbank flows across the floodplain due to pre-existing water on the floodplain is a widespread phenomenon. Based on our experience on the Fly, this tendency for a given floodplain system may vary widely depending on antecedent conditions on the floodplain. Following a drought in which the swamp grass reach of the Fly completely drains, if rainfall occurs primarily in the uplands then flooding in the lowlands may arise from overbank advection rather than from local precipitation. It may be these rare events that contribute the most important sediment load to the floodplain. An intensive field monitoring programme is now underway on the Fly to document controls on flooding.

The second process of sediment delivery to the floodplain, diffusive transport, occurs because of concentration gradients between the river source and the floodplain sink (e.g. Pizzuto, 1987). It will be strongly influenced by grain size, as coarser particles, once out of the turbulence of the channel flow, will tend to settle out (according to grain size). Presumably advective transport dominates over diffusive transport in the forested reach. Diffusive transport may take on greater significance in the swamp grass reach if advective transport is inhibited. The dense swamp grass may reduce large-scale turbulence and thereby retard diffusive and advective transport. At this time we cannot quantify the relative importance of these processes.

The third process, tie channel and up-tributary flow, follows discrete pathways into the floodplain environment and can cause significant sedimentation across it. Sediment-laden flow up tie channels and tributaries occurs more frequently than does overbank flow across the floodplain. Flow up tie channels may also be an important pathway for rising river flow to flood the surrounding plains before overbank flow develops. We have seen Fly River water travel 40 km up the Agu River (Figure 14.4), and along the way spill out of the Agu (Figure 14.9B), depositing sediment in bordering valleys and plains. Tie channels commonly have distinct deltas oriented away from the Fly, where they enter adjoining oxbows or blocked valley lakes (as noted by Blake and Ollier, 1971), recording the transport of the Fly River sediment often many kilometres across the floodplain. On a tie channel which connects two oxbows to the Fly just upstream of Everill Junction (Figure 14.6B), Markham and Day (1994) monitored flow direction and water level, and found that in this tidally influenced reach, flow could alternate direction in the tie channel many times a day.

The high sedimentation generated by transport through the tie channels raises the question as to why the channels do not quickly plug with sediment. We have witnessed two distinct stage conditions that maintain the channels. The river reaches a low stage during rapid drawdown of the Fly after a period of low rainfall in the uplands or during chronic drawdown during El Niño droughts. The forested reach of the river has been observed to drop over 4 m in less than 48 h following reduced rain in the mountains. This creates a strong pressure gradient from the off-river water bodies towards the mainstem along the tie channels. During one such event we observed flow cascading down the tie channel from an oxbow lake, cutting a trough into the underlying sediments. We have also observed very strong flow up the tie channels during a period when the river stage rose more quickly than rainwater flooded the adjacent plains. This suggests that maintenance of the tie channels depends on particular, relatively rare hydrological events.

A very crude estimate of oxbow infilling, largely due to tie channel sediment contribution, can be made by assuming that production rate of oxbows is balanced by infilling rate such that there is no net increase through the late Holocene of the number of oxbow lakes. The rate of oxbow production may be about one in 30 years (as discussed above) and there are about 29 oxbows between D'Albertis and Everill Junction (about 25 have tie channels). To keep pace with production an individual oxbow would have to fill in 870 years (29 multiplied by 30). For an initially 10 m deep lagoon, this implies a sedimentation rate of $1\,\mathrm{cm\,a^{-1}}$. We can estimate the number of times per year flow must enter the oxbow to generate this deposition through a simple mass balance. If deposition is assumed to occur by settling of sediment at concentration C at N times per year, then when the depth of the lake is H with a fluid density of ρ_w, the deposited sediment will have a grain density of ρ_s and a porosity of p, and deposition rate D is simply:

$$D = NHC(\rho_w/\rho_s)(1/1-p) \tag{14.7}$$

If H equals 1000 cm, $\rho_w = 1.0\,\mathrm{g\,cm^{-3}}$, $\rho_s = 2.65\,\mathrm{g\,cm^{-3}}$, $p = 0.35$, $C = 100$ ppm, and $D = 1\,\mathrm{cm\,a^{-1}}$, then N is 17 times per year. Clearly, as the oxbow shallows with deposition, the frequency of flow from the Fly into the lake would have to increase to maintain this deposition rate. Hence it is likely that deposition rates decline through time; to average $1\,\mathrm{cm\,a^{-1}}$, early stage deposition rate would have to be higher.

Alternatively, if we assume instead that the oldest oxbows (which tend to be completely filled) were formed shortly after eustatic sea level stabilized (about 6500 years BP), then the average aggradation rate is $1.5\,\mathrm{mm\,a^{-1}}$. The actual sedimentation rate probably falls between these two estimates of 1 to $10\,\mathrm{mm\,a^{-1}}$, and is much higher than the overbank deposition rate on the floodplain.

One other distribution system deserves special attention. Along the eastern side of the Fly River floodplain between the Binge River and the upstream end of the swamp reach, the Agu River drains southward parallel to the general trend of the Fly (Figure 14.4). It drains at least $1800\,\mathrm{km^2}$ of lowland landscape and passes through a seasonal lake system before it joins the Fly at the head of the swamp reach just above Manda. Various topographic maps indicate that the Agu is connected to the Fly via tie channels at a minimum of five locations (Figure 14.4). Recent field observations during an extended period of flooding confirm that significant flow is draining from the Fly via these channels into the Agu. During high stage, the Agu is also connected to the Binge River upslope, which means that nearly all of the eastern side of the Fly River floodplain could at times drain via the Agu River. As previously mentioned, under certain conditions, flow on the Fly will also travel up the Agu from the tributary junction near Manda. The channels which convey flow from the Fly River to the Agu also convey sediment, hence the Agu system must play a major role in the lateral distribution of sediment, in this case taking sediment beyond what might otherwise be considered the edge of the floodplain.

Downstream fining

The bed material of the pre-mine Fly River systematically fines downstream from Kiunga to Everill Junction, where it coarsens for a short distance with additional sediment from the Strickland, and then fines out into the delta (Figures 14.10, 14.11 and 14.12). This downstream fining raises the possibility that significant deposition is required along the Fly system, which would seem to be at odds with the inferences gained from preservation of floodplain features and the slow meander migration rates. Furthermore, the data do not reveal a sharp grain size reduction at the start of the swamp reach (about 240 km from Kiunga), despite the apparent backwater effects that should reduce sediment transport capacity here.

We examine this problem in two ways. First, we analyse downstream changes in potential mode of sediment transport, then we perform a simple mass balance analysis. The downstream decline in slope of the Fly, if not accompanied by some combination of

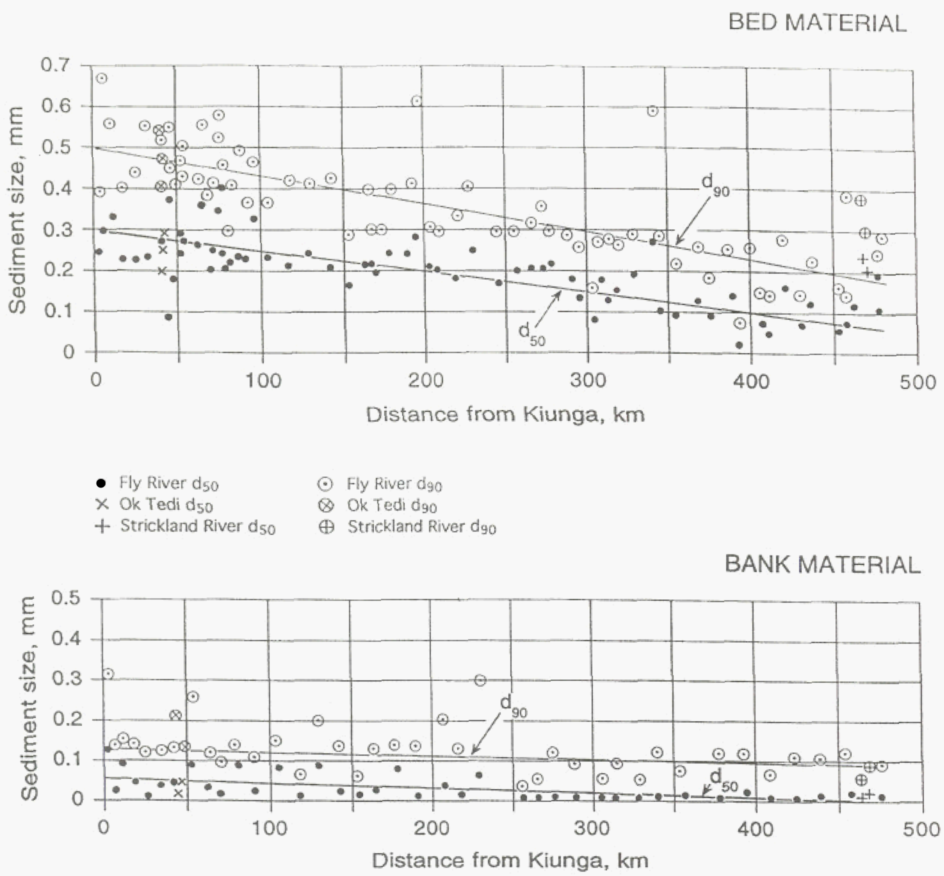

FIGURE 14.10 Downstream change in bed material and bank material (top of bank adjacent to channel). Symbols d_{50} and d_{90} refer to the median grain size and the size for which 90% of the bed is finer, respectively (slightly modified from Pickup et al., 1979)

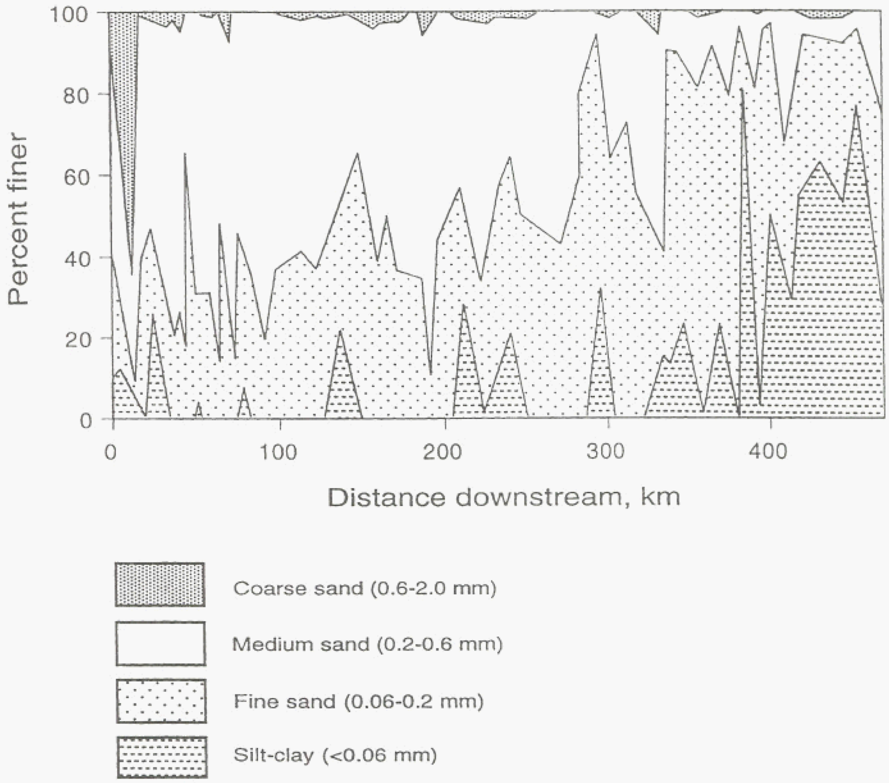

FIGURE 14.11 Bed material grain size variation from Kiunga to Everill Junction (modified from Pickup et al., 1979).

increasing depth, decreasing grain size or declining form drag, should lead to diminished sediment transport capacity of bed material load.

In Figure 14.12, we compare the downstream varying bankfull shear velocity with the downstream decline in bed material size. In order to estimate the shear velocity, the total boundary shear stress was estimated from slope values reported by SMEC (1981) and mean depths determined from cross-sections reported by SMEC and Pickup et al. (1979). Due to the influence of channel sinuosity, bars and finer scale bedforms (dunes and ripples), this total boundary shear stress was reduced by a factor of three (based on experience elsewhere; Dietrich et al., 1984) to estimate the average boundary shear stress responsible for sediment transport. We assumed that for suspension, the local shear velocity must equal or exceed the particle settling velocity (e.g. Raudkivi, 1990); we used the empirical curves of Dietrich (1982) to convert grain size to settling velocity. The vertical axis labelled "maximum grain size suspended" was scaled to correspond to the opposing values of shear velocity using an empirical relationship derived from Dietrich's curves. Hence, it is possible to read on the left axis the shear velocity, and on the right axis the corresponding grain size that can be suspended by that shear velocity.

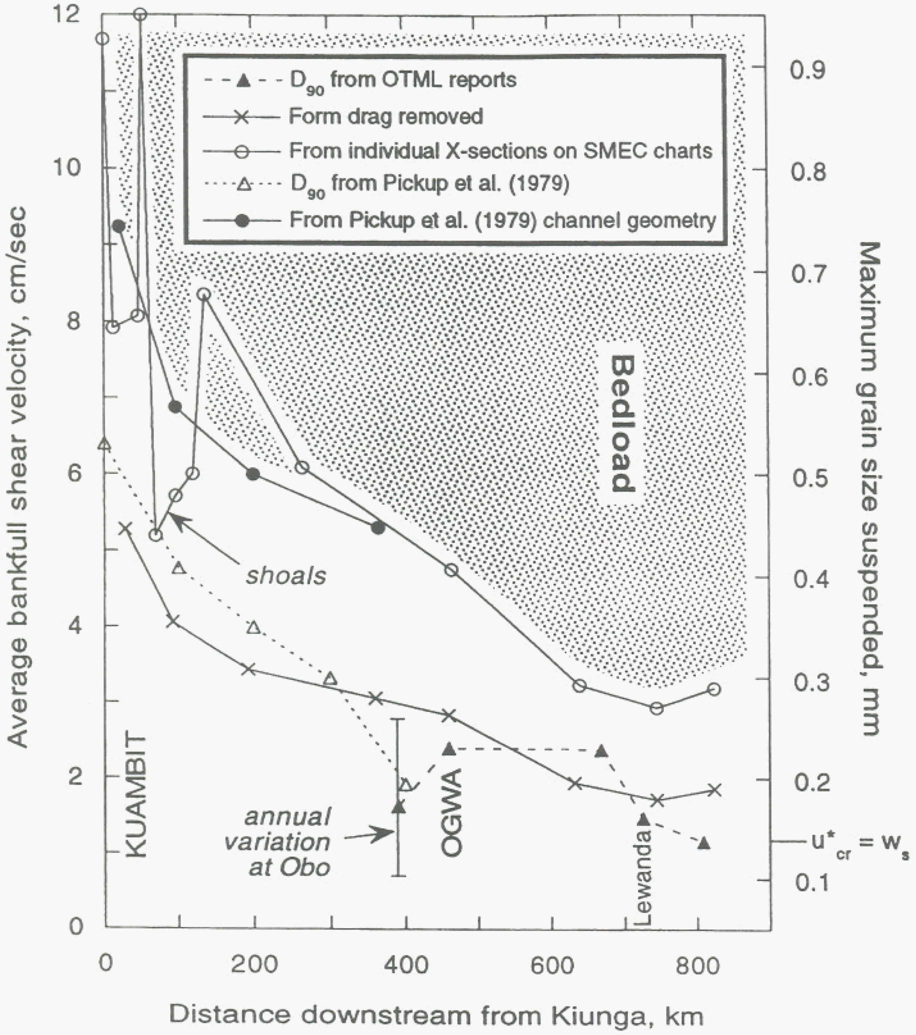

FIGURE 14.12 Downstream variation in average bankfull shear velocity (defined as the $(gdS)^{0.5}$), form drag corrected shear velocity, median and d_{90} of bed material, and maximum grain size that can be suspended by the calculated shear velocity. Note that we use the criteria that suspension occurs when the shear velocity equals or exceeds the settling velocity. If the settling velocity exceeds the shear velocity it will travel primarily as bedload and this area is shown as shaded for the shear velocity values not corrected for form drag. The point where the critical shear velocity for initial motion is equal to settling velocity of the particle is shown on the left vertical axis. Annual variation at Obo is shown because there is large temporal variance of bed material grain size, apparently due to periodic backwater effects of the Strickland.

In Figure 14.12 the downstream varying shear velocity is compared with the downstream varying grain size for which 90% of the particles are finer. The two curves closely follow each other. If the form drag correction of the total boundary shear stress is correct, then this plot also indicates that as the shear velocity declines so does the grain size, and the shear velocity is close to values capable of suspending the entire bed. Apparently, virtually all bed material can be carried in suspension through the Fly despite the large downstream decrease in shear velocity. While Figure 14.12 shows that downstream fining compensates downstream diminished shear velocity, it does not indicate how this fining occurs and whether it requires significant net deposition of sediment.

In a sand-bedded river, downstream grain size reduction due to breakdown of coarser sand during transport is unlikely. Some breakdown undoubtedly occurs during weathering and soil formation when sediment is temporarily stored in the floodplain. If the average bank erosion for the 250 km of active Middle Fly is just $1\,\mathrm{m\,a^{-1}}$ and that contributes a bank of material 10 m high, with a bulk density of $1.7\,\mathrm{t\,m^{-3}}$, then this introduces 4.3×10^6 tonnes into the river annually, a value equivalent to about half of the annual total load below D'Albertis Junction. Residence of this sediment in the floodplain probably is on average several thousand years, and in this environment weathering does clearly (as seen in bank exposures) alter sediments. However, weathering is inhibited by ground saturation and this probably significantly reduces weathering of deep sediments. One clear indication, however, that it is not only weathering that causes downstream fining, is the fact that the downstream fining has persisted (albeit somewhat muted) despite an increase in sediment load by over a factor of five times due to mine waste dumping.

The bed could fine downstream if a significant portion of the coarser sand is lost to net deposition in the bed and floodplain contributing to aggradation. Alternatively, a greater portion of the fine sand carried in suspension upstream could reside on the bed and travel as bedload or weak suspended load, causing dilution of the sand.

Simple mass balance assumptions can be used to test which of these two mechanisms could explain downstream fining. If grain size reduction occurs without net deposition of sediment due to dilution of coarser sediment with the fine sediment coming out of suspension, then there must be an increase of bedload transport in proportion to the addition of fines. The amount of total sand load coarser than a specified size that travels as bedload at section i is:

$$f_i p_i a_i T_i$$

where T_i = total sediment load at i, a_i = proportion of total load that is sand at i, p_i = proportion of sand load that travels as bedload at i and f_i = proportion of bed material coarser than a specified size i.

In this equilibrium case, if we compare grain sizes at two sections, 1 and 2, then

$$T_1 = T_2; \quad a_1 = a_2; \quad a_1 T_1 = a_2 T_2$$

and

$$f_1 p_1 a_1 = f_2 p_2 a_2$$

hence

$$p_2 = p_1 f_1 / f_2 \tag{14.8}$$

Note also that $p_1 a_1 T_1$ = bedload discharge at section 1, $(p_1 a_1 T_1)/T_1 = p_1 a_1$ = proportion of total load that is bedload at section 1, and $p_2 a_1$ = proportion of total load that is bedload at section 2.

According to Figure 14.11, at Wygerin 60% of the bed is coarser than 0.2 mm, whereas at Bosset Lagoon only about 20% of the bed material is coarser. This is just above where significant silt and clay were found in the bed. The proportion of the total load that travels as bedload just downstream of D'Albertis Junction (at Kuambit) has been estimated from both measurement and theoretical analysis to be less than 10% (Ok Tedi Mining Ltd, 1987, p.25); here we will use 5 to 10%. Limited data at Kuambit suggest that about 20% of the suspended load is sand, hence $a_{\Gamma} = 0.95(0.2) + 0.05(1)$ to $0.9(0.2) + 0.1(1)$ or 0.24 to 0.28. This also implies that the proportion of the sand load that travels as bedload, p_1, is 0.21 to 0.36. In Equation 14.8 with $f_1 = 0.6$ and $f_2 = 0.2$, $p_2 = 3p_1$, that is, the proportion of sand load that travels as bedload would increase three times between Wygerin and Bosset. Consequently p_2 would equal 0.63 to 1.0 by Bosset; that is most of the sand would be travelling as bedload by Bosset and the proportion of the total load travelling as bedload would be 0.15 to 0.28.

These simple calculations argue that fining downstream is probably *not* caused by downstream addition of fine sand residing as bed material and travelling as bedload. Note, too, that there is probably little sediment contributed from the few tributaries between D'Albertis and Everill Junctions. No field evidence suggests that bedload transport is more significant at Bosset; in fact, the reduction in the size of bars in this lower reach may indicate an opposite trend. A large increase in downstream bedload transport also seems unlikely under the observed condition of downstream declining boundary shear stress. The relatively large proportion of the total load required to travel as bedload here is also very unlikely. Furthermore, Pickup et al. (1979) found that top-of-channel bank deposits adjacent to the channel still contained up to 30% sand in this reach.

If decrease in grain size is due to net deposition of the coarser fraction from the bedload only, then:

$$a_1T_1 \neq a_2T_2$$

but

$$f_1p_1a_1T_1 - f_2p_2a_2T_2 = a_1T_1 - a_2T_2 \tag{14.9}$$

i.e., the amount of total sand coarser than a specified size in the bedload at section 1 minus the amount at section 2 downstream equals the difference in the total sand load. Rearranging Equation 14.9, assuming $p_1 = p_2$, and using the values from the previous examples gives:

$$a_2T_2 = a_1T_1(1 - f_1p_1)/(1 - f_2p_2)$$

$$= 0.91a_1T_1 \text{ to } 0.85a_1T_1$$

This surprising result states that the dramatic decrease in medium sand content in the bed material from Wygerin to Bosset and the corresponding decrease in median size from about 0.23 mm to 0.13 mm could occur if less than 15% of the total sand load is lost to deposition. The reason this number is small is the relatively small proportion of the total load that is bedload and the assumption that the medium sand travels only as bedload. This latter assumption is well supported by the expected mode of transport based on shear stress values (Figure 14.12). Also, some medium sand is found as overbank deposits. However, as long as the percentage of the suspended load coarser than 0.2 mm and the total suspended load discharge remain constant between Wygerin and Bosset, the results are the same.

The selective deposition of coarser sand that leads to downstream fining probably occurs by several mechanisms. Slow aggradation of the bed of the river can account for some. For example, if the sediment load before the mine below D'Albertis Junction was 10×10^6 tonnes annually of which 28% was sand, and of this total sand 15% were deposited in the bed between Wygerin and Bosset (about 220 km), then assuming a bulk density of 1.9 (upon deeper burial), the annual aggradation of this reach for a 200 m wide channel bed would be $(10 \times 10^6 \times 0.28 \times 0.15)/(1.9 \times 220\,000 \times 200)$ or about $5\,\mathrm{mm\,a^{-1}}$, a rate much greater than inferred to be the long-term floodplain deposition. This suggests that significant portions of the medium sand are also deposited on levees and in off-river water bodies.

DISCUSSION AND CONCLUSION

Our analysis suggests that some inferences that have previously been made about the Fly were correct, and others, despite seemingly clear evidence, were not. Despite appearances, the Fly River is currently not an unstable river rapidly crossing its floodplain. On the other hand, estimates of long-term floodplain deposition rates are supported by further analyses presented here. These estimates and other observations, however, seem to contradict a proposal by Harris et al. (1996) regarding sediment storage in the Fly River floodplain. In essence, they report detailed analyses of delta and offshore sedimentation patterns for the Holocene and conclude that the large sediment input from the uplands areas cannot be accounted for in deltaic and near-shore deposition. To solve this missing mass problem, they propose that in the Holocene 30 m of aggradation took place in the Fly and Strickland floodplains. This deposition, they suggest, may have occurred in as little as 3000 years, implying an aggradation rate of $1\,\mathrm{cm\,a^{-1}}$. While Early to Mid-Holocene sedimentation rate was probably much higher than present, our estimates and those of previous workers do *not* support this higher sedimentation rate. Furthermore, Pickup et al. (1979) report encountering weathered (probably Pleistocene) sediments in the bed of the Fly River near Kiunga, implying limited incision during glaciation.

It seems likely that the swampy conditions with reduced channel migration in the roughly 200 km reach of the lower Middle Fly River formed relatively recently, perhaps due to slope reduction caused by backwater of the Strickland or possibly tectonic effects. A key to understanding what happened is to find out when slope reduction occurred, and an effort is underway to date channel migration history throughout the Middle Fly.

Qualitative observations suggest that sediment is delivered to the floodplain by at least three processes: overbank advection, diffusion, and transport via tie channels and tributaries. The most dramatic case of tributary transport is the drainage of the Middle Fly into the parallel-draining Agu River and the upriver transport of sediment from the downstream tributary junction. Very few data are available on rates of sedimentation caused by these various processes on large lowland floodplains. We can infer that rates of sedimentation will be controlled by near-surface river flow sediment concentration and grain size, frequency of overbank flow and frequency of flow up tie channels and tributaries. These latter frequencies appear to be tied to timing of upland versus lowland rainfall, which dictates how wet the floodplain already is when the flow on the Fly rises.

Recent analysis of remote sensing data in several large river systems around the world by Mertes (1997) suggests that the inhibition of sediment-laden overbank flows across the floodplain due to pre-existing water on the floodplain is a widespread phenomenon. Observations on the Fly suggest there will be annual variation in the relative timing of upland versus lowland rainfall and there will be longer time variations, for example El Niño drought events, that play a significant role in determining when important depositional events occur across the floodplain. A programme is now underway to try to monitor the floodplain hydrology directly so as to understand the controls on timing and magnitude of flooding.

Downstream fining of bed material on large sandy rivers has received little attention compared to that given to the phenomenon on gravel-bedded rivers (e.g. Paola et al., 1992). Nonetheless, dramatic fining can occur. Leopold et al. (1964) report data on the Mississippi River collected before 1935 when the river was not converted to a network of dams and lakes. These data show the mean grain diameter decreasing in size at a constant rate over 1500 km from about 0.65 mm just below New Madrid, Missouri, to less than 0.2 mm below New Orleans, Louisiana. This fining results from systematic loss of coarser sediment (including a small amount of gravel in the upstream reaches). Based on our analysis of the Fly, rather than reason that this reflects a large net loss of sediment and a significant decline in the river's transport capacity, we suggest that such fining can occur with a relatively minor amount of net aggradation of the river bed or net discharge to the floodplain environment. Alternatively, there could have been a dilution effect caused by addition of significant amounts of finer sediment from the large tributaries which, unlike the Fly, line the Mississippi.

Through reanalysis of existing data and addition of some new data on the Fly River, we have been able to reject some inferences, confirm or at least support others, and offer a few new interpretations. Field and modelling research is now underway that should allow us to investigate much more quantitatively the issues raised here, no doubt introducing new ones.

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