One-dimensional numerical modeling of the long-term morphodynamic evolution of a tidally-dominated estuary: The Lower Fly River (Papua New Guinea)

Alberto Canestrelli a,c,*, Stefano Lanzoni b, Sergio Fagherazzi a

a Department of Earth Sciences, Boston University, Boston, MA 02143, USA
b Departamento IGEA, University of Padua, Italy
c Department of Geosciences, Penn State University, State College, PA 16803, USA

A R T I C L E   I N F O

Article history:
Received 21 December 2012
Received in revised form 20 May 2013
Accepted 27 June 2013
Available online xxx

Keywords:
Long-term morphodynamics
Tidally-dominated estuary
Fly River
PRICE-C scheme
River-tidal transition

A B S T R A C T

We use a one-dimensional morphodynamic model to analyze the long-term evolution of the lower reaches of the Fly River, Papua New Guinea, from the Everill Junction to the delta mouth. The model shows how the break in the exponential trend of river width triggers deposition, thus producing a tidal region characterized by a higher bed elevation with respect to the river-dominated one. Numerical simulations indicate that the river attains a dynamic equilibrium configuration in which the amount of sediment entering upstream is flushed seaward. A sensitivity analysis is performed, in which the effect of varying solid discharge, tidal harmonics, and initial conditions is discussed. The model shows that an equilibrium configuration results from a delicate balance between the aggrading effect associated with channel divergence (acting mainly during neap tide and at slack water) and the opposite effect of tidal flushing driven by residual water discharge. A physically meaningful morphodynamic equilibrium occurs only for a small range of values of sediment discharge prescribed at the upstream boundary. In particular, an increase in sediment discharge leads to aggradation, while a decrease triggers extensive scour and a deepening of the estuary.

© 2013 Elsevier B.V. All rights reserved.

1. Introduction

Tidal dominated estuaries are key areas that control the fate of sediment from source to sink, starting from the upland areas to the continental shelf (Dalrymple and Choi, 2007; Fagherazzi and Overeem, 2007; Fagherazzi, 2008; Walsh and Nittouer, 2009; Canestrelli et al., 2010a). From a sedimentological perspective, it is important to understand whether a given estuary or delta is infilling, eroding, or tends to attain morphodynamic equilibrium. The latter equilibrium essentially consists of a dynamic state for which all the sediments coming from upstream are flushed out to the ocean without net variations in bed elevation in a neap-spring tidal cycle (Seminara et al., 2010). In these equilibrium conditions, the estuary would then act as a bypassing conduit for dispersal of sediments from the river dominated region to the inner continental shelf (e.g. Wright, 1985).

In the last two decades, several studies have addressed the long-term morphodynamic evolution of dead-end channels, i.e., characterized by a negligible upstream water and sediment discharges. Laboratory experiments (Tambroni et al., 2005), one-dimensional numerical models (Lanzoni and Seminara, 2002; Todeschini et al., 2008; de Swart and Zimmerman, 2009) and analytical methods (Seminara et al., 2010; Toffolon and Lanzoni, 2010) all indicate that these type of channels can reach a long-term equilibrium configuration characterized by a tidally-averaged bed variation that is zero at each location along the channel. In these studies the system was forced by a purely sinusoidal semi-diurnal M2 tide at the channel mouth and the time scales considered were in the order of centuries to thousands of years, during which sea level rise and subsidence effects were neglected. The equilibrium profile was characterized by a shoaling bed with elevations increasing from the inlet to an emerging shore landward. On the contrary, to the authors’ knowledge, no numerical investigations have been carried out on the long-term morphodynamic evolution of tidally-dominated rivers, in which a strong tidal forcing interacts with not negligible riverine discharge and upstream sediment input. To address this research gap, here we provide some results on the hydrodynamic and morphodynamic behavior of the lower reaches of the Fly River, Papua New Guinea, which are affected by both tidal and fluvial processes.

The Fly River originates in the tectonically active highlands of Papua New Guinea (Fig. 1), then crosses an extensive alluvial, low-gradient valley, and finally flows into the Gulf of Papua through funnel-shaped distributaries, forming a typical tidally-dominated delta (Galloway, 1975; Syvitski and Saito, 2007). In the last decades a lot of effort has been devoted to understand the sediment budget...
Fig. 1. Plan view of the Fly River, Papua New Guinea: (a) overall catchment and (b) tidal delta. Courtesy of NASA World Wind. Reprinted from Canestrelli et al. (2010a) with permission of AGU.
and the morphodynamic evolution of the system upstream of the junction with the Strickland, in the distributary delta channels, in the delta front, pro-delta and coastal shelf (see Section 2 for a description of these results). Conversely, little is known about the sediment dynamics in the tidally influenced portion of the Lower Fly, extending for about 400 km from the Everill Junction to the delta apex.

In this work we thus focus on the long-term morphodynamic behavior of the Lower Fly River to understand how the upstream sediment load is conveyed to the river mouth. In particular, we would like to answer the following questions: can a tidally-dominated river reach an equilibrium bed configuration under constant base level (sea level) and constant water discharge? Is the configuration stable or unstable to perturbations in sediment discharge?

To this aim, we use a one-dimensional numerical model solving the di Saint Venant equations coupled with the bed-evolution Exner equation. This approach allows long-term simulations (up to thousands of years) with an acceptable computational effort, filtering out high frequency bed variations due to the presence of estuarine bars, that likely play a minor role in the long-term evolution of the river bed profile (Tambroni et al., 2005; Seminara et al., 2010).

The choice of a cross-sectionally averaged model, notwithstanding its inability to deal with estuarine circulations and stratification dynamics, is justified by field observations (Wolanski et al., 1995a), indicating that the region interested by non-negligible salinity concentrations is a small percentage of the total length of the investigated reach and, hence, likely does not affect the overall morphodynamic behavior of the system.

The rest of the paper is organized as follows. Section 2 provides an overview of the study site. In Section 3, we describe the mathematical model used to investigate the hydrodynamics and morphodynamics of the Lower Fly River, the geometry employed and the boundary conditions used in the simulations. In Section 4 we present the model results, with particular emphasis on the spatial distribution of maximum velocity and sediment discharge. In Section 5 we discuss how the system responds to variations in external forcing, and how the different parameters influence the evolution of the river. Finally, in Section 6 some conclusions are drawn.

2. Study area

The overall river path can be subdivided in three main reaches, the Upper, Middle and Lower Fly, delimited by the confluences with two main tributaries, the Ok Tedi and the Strickland Rivers, joining the Fly at the D’Albertis and Everill Junctions (Fig. 1a).

The Fly River, as well as its tributaries, is a tropical river characterized by remarkably small variations in freshwater discharge, with significant discharge reductions occurring only during infrequent drought periods (Wolanski et al., 1997; Dietrich et al., 1999). The flood period lasts about 40% of the year (Pickup, 1984), the average annual flow being 1900 m$^3$/s below the D’Albertis Junction and around 5500–6000 m$^3$/s at Everill Junction (Wolanski et al., 1997), due to the input of the Strickland river discharge of about 3100 m$^3$/s (Higgins, 1990; Dietrich et al., 1999).

The tidally dominated delta has been the subject of several studies concerning water and sediment dynamics (Wolanski and Eagle, 1991; Wolanski et al., 1995a, 1995b, 1997; Harris et al., 2004), as well as sedimentation patterns, and stratigraphy (Harris et al., 1993; Baker, 1999; Dalrymple et al., 2003).

More recently, the attention has been concentrated upon the lowland reach of the Fly and of the Strickland and the Ok Tedi tributaries, addressing the response of large sandbed rivers and adjacent floodplains to changes in sediment supply, base level, and mean sea level (Dietrich et al., 1999; Aalto et al., 2008; Day et al., 2008; Lauer et al., 2008; Parker et al., 2008; Swanson et al., 2008). Sediment is mainly supplied from the steep headwater regions, characterized by a denudation rate of 3–4 mm/year (Pickup, 1984) favored by the presence of easily weathered volcanic, sedimentary and weakly metamorphosed bedrocks, and the occurrence of high rainfall (>10 m/year, Harris et al., 1993). Before mining activities started in the Ok Tedi and in the Strickland basins (in 1985 and in 1991, respectively), the mean annual sediment discharge at D’Albertis Junction was about 10 Mt/a, while the Strickland carried a mean annual load of roughly 70–80 Mt/a at Everill Junction (Dietrich et al., 1999). Mining is estimated to have caused a 40% increase in the sediment discharge (Eagle and Higgins, 1990; Wolanski et al., 1995a). Despite this, sediment concentration in the Upper and Middle Fly and long-term sedimentation rates over the flanking floodplain are relatively low (Dietrich et al., 1999; Day et al., 2008; Swanson et al., 2008). Most of the suspended load carried by the river (including fine sand) is then conveyed to the sea through three distinct distributary channels departing from the delta apex (Fig. 1b), and delta progradation associated to this suspended sediment load has been estimated to be roughly 6 m/year (Harris et al., 1993).

The tidal regime in the Gulf of Papua is semidiurnal, with a daily tidal inequality and spring-neap variability. Peak to trough fluctuations at the distributary channel mouths are up to 4–5 m during spring tides and about 0.5–1 m during neap tides (Wolanski et al., 1997), implying that the daily tidal flux through the delta is about two orders of magnitude larger than the average freshwater fluvial discharge (Canestrelli et al., 2010a).

Tides propagate inland for about 400 km (i.e., up to Everill Junction) during usual discharge conditions, and can reach Manda, 570 km inland during low-river stages (Dietrich et al., 1999). The Lower Fly can then be considered as an estuary, according to the definition given by Dalrymple and Choi (2007). Hereafter we will use this terminology to indicate the investigated tidal reach of the river.

The transported sediments undergo a significant downstream fining and sediments coarser than fine to very fine sand prevalently deposit in the middle reach, located in the low-gradient part of the foreland basin (Dietrich et al., 1999). The bar deposits at the distributary delta mouths are composed by fine sand, while the distributary channel bed usually consists of fine to very fine sand overlaid by mud deposits, forming after spring tides in the zone of elevated suspended sediment concentrations extending from the distributary-mouth bar area to the delta apex (Wolanski and Gibbs, 1995; Wolanski et al., 1995a; Baker, 1999; Dalrymple et al., 2003).

Sediment transport mainly occurs few days each month, when resuspension processes prevail during spring tides, leading to values of the mean depth-averaged concentration in the range 1–4 g/l (Wolanski et al., 1995a).

During spring tides, the large mixing enhanced by strong tidal fluxes and by lateral shear due to the meandering of the thalweg around the shoals and islands inhibits vertical and horizontal stratification, implying that well mixed conditions are attained throughout the entire delta and tidal river reaches (Wolanski and Eagle, 1991; Wolanski et al., 1995a, 1997). Baroclinic circulation driven by salinity gradients can occur only during neap tides, when suspended sediment tends to settle and sediment fluxes are relatively small. Finally, Coriolis-induced circulations are negligibly small, since the Fly River delta is situated close to the equator.

3. Numerical model

In this section we present the numerical model, the geometry, the initial and boundary conditions used to simulate the morphodynamic evolution of the Lower Fly River. Typical channel depth, h, in the Lower Fly River is of the order of 10 m. The channel width B is about half a kilometer at Everill Junction and increases downstream up to about 40 km at the delta mouth (hydraulically active channel width). The length L of the investigated reach is almost 400 km. As a consequence, the channel can be considered shallow ($h/B \ll 1$) and long compared to its width ($B/L \ll 1$). The long term (i.e., centuries/millennia) behavior of the longitudinal river bed profile can then
be studied by using a one-dimensional approach. Clearly, channel bends occurring in the fluvial reach and the complex cross-sectional geometry characterizing the region between the delta apex and toe produce locally two- and three-dimensional effects which, however, act on much smaller time scales determining the high frequency spatial fluctuations typically exhibited by the cross-sectionally averaged longitudinal bed profile. The specific study of these smaller scale effects would imply the use of two and three-dimensional models, with a consequent huge increase of the computational effort when modeling long-term morphodynamics. Here, we thus focus our attention on the gross morphodynamic behavior of the river, studying the existence of a possible equilibrium configuration of the overall (i.e., cross-sectionally averaged) longitudinal river bed profile.

3.1. Model description

The model here considered consists of the cross-sectionally averaged equations for the conservation of water mass and momentum, and the one-dimensional Exner bed evolution equation. Owing to the large width to depth ratio, the section of the river is schematized as rectangular, with a width that can vary along the downstream x-direction (positive seaward). Moreover, under the hypothesis that the time scale characterizing flow variations (hydrodynamic time scale) is much smaller than the time scale controlling the river bed evolution (morphodynamic time scale), we decouple the two processes and introduce a “morphological factor” \( MF = 200 \) to compute the river bed evolution. We refer the interested reader to Roelvink (2006) for a thorough review of decoupled techniques and for the justification of using the morphological factor approach in tidal environments. Assuming that, as discussed above, baroclinic circulations can be reasonably neglected, the equations governing the flow field read

\[
\frac{\partial B(H-z_{b})}{\partial t} + \frac{\partial Q}{\partial x} = 0
\]  
\[
\frac{\partial Q}{\partial t} + \frac{\partial}{\partial x} \left( \frac{Q^2}{B(H-z_{b})} + \frac{1}{2} \frac{Q^2}{2gB(H-z_{b})} \right) = \frac{g}{2} \left( H-z_{b} \right) \frac{\partial B(H-z_{b})}{\partial x} - gB_{s}L_{f}
\]

in which \( B \) is the channel width, \( H \) is the water surface elevation, \( z_{b} \) is the river bed elevation, \( Q \) is the water discharge, \( B_{s} \) the hydraulic radius and \( g \) is gravity. The friction slope \( S_{f} \) reads:

\[
S_{f} = \frac{Q^2}{B^3H^2C^2R_{h}^{1/2}}
\]

where \( C \) is the Chezy friction coefficient.

The bed evolution Exner equation reads:

\[
\frac{\partial (1-p)Bz_{b}}{\partial t} + \frac{\partial Q_{s}}{\partial x} = 0
\]

with \( p \) the bed porosity and \( Q_{s} \) the total sediment load. The latter, given by the sum of bedload and suspended load, is computed according to Engelund and Hansen (1967) total load relationship:

\[
Q_{s} = 0.05 \frac{u^{5}}{\sqrt{C^{2} \Delta^{2}D_{i}}} \frac{B}{\Delta}
\]

in which \( D_{i} \) is the average sediment diameter and \( \Delta = (\rho_{s} - \rho)/\rho \) the immersed relative density, \( \rho \) and \( \rho_{s} \) being the density of water and sediment, respectively.

Note that in the present formulation of the morphodynamic problem, we consider banks as inerrible, assuming that variations in bed elevation act at a smaller temporal scale than width variations. The planar configuration of the river is then kept fixed when computing the correspondent equilibrium bed configuration. This assumption has been widely used in the study of the evolution of tidal channels (Lanzoni and Seminara, 2002; Todeschini et al., 2008; Seminara et al., 2010; Toffolon and Lanzoni, 2010). Moreover, none or little river shift has been observed in the last 50 years along the lower part of the Middle Fly (Dietrich et al., 1999), and no oxbow lakes are present in the lower two third of the Lower Fly, suggesting a remarkably stable planform pattern. This behavior is in accordance with theoretical findings indicating that tidal meanders are typically more stable than river meanders (Solari et al., 2002).

The model does not consider also the possible flooding of the floodplain adjacent to the river. A discussion on possible effects of the floodplains is presented in Section 3.4.

3.2. Numerical method

A particular attention has been paid on the mathematical form of the system of partial differential Eqs. (1), (2) and (4) to be solved numerically. Often, a non-conservative (primitive) formulation of this system is employed when studying one-dimensional tidal propagation in tidal channels (Friedrichs and Aubrey, 1988, 1994; Lanzoni and Seminara, 2002; Seminara et al., 2010), and the velocity is used as an unknown and conserving variable. This approach, however, gives good results as long as the solution is smooth (i.e. continuous). In the presence of discontinuities (as in the case of a hydrodynamic jump), it has been proven that a non-conservative formulation of the numerical solution leads to an erroneous computation of the height and speed of the discontinuity (Toro, 2001), thus rendering meaningless long-term morphodynamic simulations. In the presence of a strong channel convergence and high amplitude tides, a tidal bore is likely to form, which consists in a discontinuity of the water surface (and for continuity, of the velocity) traveling landward. A tidal bore with height up to 1 m is indeed known to develop during spring tide in the freshwater region of the Fly River upstream of Lewada (Wolanski et al., 1995a). In order to properly address this phenomenon and the related morphodynamics, the PRICE-C scheme of Canestrelli et al. (2009) is used, rewritten in an expanded form as in Canestrelli et al. (2012), together with a modification to ensure mass conservation and well-balancing between fluxes and source terms (Appendix A). This approach has in fact shown to capture the right speed of water and sediment fronts (Canestrelli et al., 2009, 2010b, 2012).

3.3. Model geometry, boundary and initial conditions

In tidally-dominated estuaries, the channel width often follows an exponential law of the type:

\[
B = B_{0} \exp \left( -\frac{x}{L_{b}} \right)
\]

in which \( B_{0} \) is the width at the mouth, \( L \) the estuary length and \( L_{b} \) the convergence length (Savenije, 2005; Seminara et al., 2010; Toffolon and Lanzoni, 2010). Note that the longitudinal coordinate \( x \) starts at the mouth and it is positive landward, contrarily to the coordinate \( x \) appearing in Eqs. (1)–(3), which is here assumed positive seaward and with the origin right downstream Everill Junction. Fig. 2 shows a semi-logarithmic plot of the width distribution on the Lower Fly River. The width has been manually extracted from satellite images, and only the hydraulically active channel width has been considered, i.e. neglecting the presence of islands. This is consistent with the essentially supertidal nature of the larger islands, i.e. the ones located in the delta region (Lawrence, 1995, Fig. 1b).

From Fig. 2 break in the width convergence length is evident: in the first 150 km downstream Everill Junction, i.e. in the river-dominated
of 2650 kg/m³, a total load of 0.1 m³/s is prescribed at the upstream Everill Junction (Fig. 1) are 6 Mt/a and 75 Mt/a, respectively (Markham and Day, 1994). Pickup (1984, see also Higgens et al., 1987, Baker, 1999) estimated that perhaps as little as 10% of this load reaching the Lower Fly River (see Section 5). Pre-mine estimation of sediment discharge of the Middle Fly (see Fig. 1) are 6 Mt/a and 75 Mt/a, respectively (Markham and Day, 1994). Pickup (1984, see also Higgens et al., 1987, Baker, 1999) estimated that perhaps as little as 10% of this load reaching the Lower Fly River is prescribed at the upstream Everill Junction (~2 × 10⁻³, Lauer et al., 2008) has been initially considered in the various numerical simulations, and possible alternative initial conditions are discussed in Section 4.5.

At the upstream boundary, a constant water discharge (= 6000 m³/s) and a constant total sediment load for the non-cohesive fraction have been prescribed. Indeed, the Fly River has a relatively constant freshwater discharge, due to persistent rainfall in the highlands that keeps the river chronically flooded (Pickup, 1984). Moreover, between Kuambit and Obo (see Fig. 1), the presence of broad floodplains and other off-river water bodies connected to the main river through incised channels lead to the dumping of floods (Fig. 5 of Day et al., 2008). This allowed us, as a first approximation, to avoid the use of an intermittency factor. Note that the latter was introduced by Parker et al. (2008) to take into account the fluctuation of water discharge at the upstream boundary of their modeled reach (i.e. the gravel-sand transition, upstream of the D’Albertis Junction), while our upstream boundary is at the Everill Junction. Moreover, note that the concept of intermittency factor implies a steady formative discharge and, hence, the absence of tides. When tidal effects are strong, it is not clear how the intermittency factor should be expressed, since the formative freshwater discharge is likely to control the morphodynamics in the upstream river-dominated reach, while are the flood/ebb discharge peaks that shape the seaward region.

The grain size characteristic of sand in the estuary is about 0.1 mm (Harris et al., 1993) and this value has been chosen for the mean diameter in Eq. (5). Some uncertainty exists on the estimate of sediment load reaching the Lower Fly River (see Section 5). Pre-mine estimates of sediment discharge of the Middle Fly and Strickland Rivers at the Everill Junction (Fig. 1) are 6 Mt/a and 75 Mt/a, respectively (Markham and Day, 1994). Pickup (1984, see also Higgens et al., 1987, Baker, 1999) estimated that perhaps as little as 10% of this sediment is fine to very fine sand. Assuming a typical grain density of 2650 kg/m³, a total load of 0.1 m³/s is prescribed at the upstream boundary. We also assume that the mud can be treated as wash-load (Parker et al., 2008) and it mainly deposits seaward of the delta front, not contributing to the morphodynamics of the tidal river.

At the downstream boundary the tidal signal is assigned using the tidal harmonic constituents provided by the National Tidal Centre, Australia, referring to a gauge located at Umuda Island. Moreover, the total load transport is assumed to be in equilibrium with the local hydrodynamic conditions.

Some indications on the value of the Chezy coefficient to be used in the momentum equation can be obtained from the study of Lauer et al. (2008), according to which C = 66 and 38 m¹/² s⁻¹ in the Strickland and the Middle Fly, respectively. In the Lower Fly, we thus assume that C increases linearly from a value of 38 m¹/² s⁻¹ at Everill junction to a value of 66 m¹/² s⁻¹ at the delta mouth. This choice is justified by two observations: first, the meander wavelength increases downstream, leading to minor dissipations due to secondary currents closer to the delta; second, the mud content increases seaward, therefore bed friction is likely to decrease in the distributary delta channels, as observed by King and Wolanski (1996).

Finally, a sensitivity analysis was carried out to determine the optimal grid size. Starting from a grid made of 100 computational elements, their number was progressively doubled until the morphodynamic results became independent from the grid size, and this occurred for a grid of 400 elements, implying a cell size of about 0.9 km. If fewer elements (i.e. 200) are employed, some hydrodynamic features (e.g., the tidal bore) are not well resolved and the final bed configuration is slightly changed. If more elements (i.e. 800) are used, the tidal bore is sharper but the river topography does not appreciably modify. Therefore, the final choice is a good compromise between computational speed and accuracy.

3.4. Model limitations

Several approximations are embodied in our numerical model. First of all, a simplified and fixed planform geometry is employed, especially in the delta region, to better approximate the real geometry of the river. A two-dimensional model should be employed in the funnel-shaped downstream region to reproduce the diffusion and deposition of sediments toward the banks and therefore a possible redistribution of the hydraulic section. This model should be then coupled with suitable bank erosion formulations, able to deal with hydraulic erosion (Julian and Torres, 2006), as well as with bank collapsing due to excess of pore pressure (Darby et al., 2007). However, an analysis of aerial photographs indicates no detectable changes to the combined surface area of islands and distributary channels in the Fly delta over the past 50 years (Hughes and Baker, 1996). Even though the observation period was relatively short, this would imply that, at least close to the mouth, the hydraulic section is relatively stable.

Another approximation characterizing the model consists in neglecting the presence of intertidal areas (either tidal flats or floodplains) flanking the main channel, which are known to strongly change the hydrodynamics of the system, increasing ebb-dominance (Speer and Aubrey, 1985; Friedrichs and Aubrey, 1988). The reliability of this approximation is supported by the recent observation of Pickup and Marshall (2009), who pointed out that remnants of fluvial terraces in the Lower Fly are higher than the extensive floodplains and backswamps present in the Middle Fly. The absence of significant meanders migration is another clue suggesting the presence of stable banks flanking the tidal dominated reach.

An equilibrium configuration with a main channel flanked by stable tidal flats and contained within supratidal floodplains is also unlikely. A balance needs to establish between tidally driven deposition on the shallow area and wind-wave driven erosion in order for tidal flats to be preserved (Fagherazzi et al., 2006). This is not present because waves are of limited importance in the estuary and the accretional intertidal flats are rare at the margins of distributary channels (Dalrymple et al., 2003). Instead, the estuary is dominated by sandy sediments and sub-tidal sandbanks (Harris et al., 1993). Moreover, in order to favor an ebb-dominated flux able to flush out the sediments accumulated in the estuary flare, tidal flats should occur mainly in the central part of the Lower Fly reach, i.e. where the maximum deposition occurs. There the section is relatively narrow, broad equilibrium tidal flats are not evident and are unlikely to form even with an increase in sediment discharge.

Finally, it has been assumed that only sandy sediments contribute to determine the equilibrium profile while muddy sediments can be...
treated as wash-load and, hence, do not deposit at the river bed (Parker et al., 2008). The mud, indeed, cannot deposit in the upper portion of the tidal river due to tidal currents at spring and the freshwater discharge at neap. Conversely, mud deposition is favored at slack water near to the mouth, where the salinity front enhances flocculation.

4. Results

4.1. River hydrodynamics with a linear bottom profile

As described above, the initial bed configuration consists of a linear profile with slope $2 \times 10^{-3}$ (see Section 3.3). For this morphology, Fig. 3a,b show the computed instantaneous water elevation and velocity at four instants of a spring tidal cycle: when the water surface at the mouth is maximum ($t=0$), when it is equal to mean sea level and the tide is flooding ($t=T/4$), when it is minimum ($t=T/2$) and when it is equal to the mean sea level and the tide is ebbing ($t=3T/4$). Note that $T=12$ h is the characteristic period of the tidal cycle. The velocity at the upstream boundary is about 0.8 m/s and its downstream variations depend on the instant under consideration (Fig. 3b). Due to the significant length of the reach, the numerical results indicate that the flow is never flooding or ebbing throughout the entire system. When the mouth is at high water slack ($t=0$), a drawdown, downstream-accelerating profile establishes in the upstream river-dominated part, ending with a tidal bore (see inset in Fig. 3a) implying a sudden increase of water surface and a sudden decrease of water velocity. Moving further downstream, the flow depth progressively increases while the velocity changes direction and it is finally directed landward in the remaining part of the estuary. At $t = T/4$ the velocity changes direction twice along the river, being the previous flood wave is close to the upstream boundary and the river discharge leads to a large dissipation of the tidal wave (mainly by friction, Horrevoets et al., 2004), while the mouth is at the end of the ebb phase, with a lag with respect to low-water slack due to inertia. Finally, at $t = 3T/4$ the seaward part of the estuary is flooding while the landward part is ebbing.

The numerical results reported in Fig. 3a,b clearly shows that, as the tide propagates along the estuary, it experiences a distortion leading to the non-linear growth of harmonics different from the principal astronomic constituents (e.g. Dronkers, 1964; Pingree and Griffiths, 1978). Distortion is the result of finite amplitude effects and therefore is enhanced for large tidal range to depth ratios. The main effect is that the crest of the tidal wave travels faster than the trough, since the current is enhanced for large tidal range to depth ratios. The main effect is that the crest of the tidal wave travels faster than the trough, since friction is larger for smaller depths. In the case of dead-end tidal channels with negligible freshwater discharge this effect would give rise to a flood duration shorter than the ebb (Friedrichs and Aubrey, 1988). For continuity, in fact, the velocity must be larger during flood, this effect being usually magnified by the increase in tidal range associated with channel funneling. Since the total sediment transport, as a first

![Fig. 3. Water elevation (a) and velocity (b) at four instants during spring tide. Maximum and minimum values along the channel for water surface (c), velocity (d), water discharge (e) and total load (f). In the inset of a the tidal bore is shown. Positive values of the velocity corresponds to ebb conditions, while negative values refer to flood conditions. The bed configuration consists of the linear profile used as initial topography.](https://doi.org/10.1016/j.sedgeo.2013.06.009)
approximation, increases with the fifth power of the velocity (see Eq. (5)), dead-end tidal channels are usually flood dominated and the residual transport is landward. In the case of the modeled reach, although the distortion of the tidal wave is evident, flood dominance cannot establish owing to the presence of a not negligible freshwater discharge. The tidal range increases from the mouth landward (Fig. 3c) and the tidal wave deforms at a point that the wave breaks forming a tidal bore (inset of Fig. 3a). Moving further landward, the tidal range decreases and the tidal bore dissipates in the riverine part of the modeled estuary. The entire estuary is ebb-dominated, with velocity peaks much larger in ebb than in flood in most of the upper 200 km (Fig. 3d). Ebb dominance persists also near the mouth, where the velocities are of the same order, but the ebb velocity is slightly larger (Fig. 3b). The calculated water discharge exhibits an exponential trend similar to the variations in estuary width, with a maximum value at the mouth about two orders of magnitude larger than the average freshwater fluvial discharge (Fig. 3e). Finally, the calculated total-load follows a trend (Fig. 3f) similar to that exhibited by the maximum/minimum velocities, but strongly amplified since sediment transport rate is proportional to the fifth power of the velocity (Eq. (5)).

4.2. Morphodynamics and equilibrium hydrodynamics

In the present series of numerical experiments the model has been run until an equilibrium bed configuration was reached. The tidal reach was considered at equilibrium (and the numerical simulation was stopped) when bed variations in each computational cell did not exceed 1 mm/year.

Our goal was to seek an equilibrium bed profile compatible with the observed width distribution and the prescribed boundary conditions. Even if this solution can be reached only in the presence of constant external forcing (i.e., neglecting variations in mean sea level, sediment discharge, and water discharge), our results shed light on the morphodynamic trajectory of the system and the time scales of the dominant process (Seminara et al., 2010). To this aim we are facilitated by the fact that the Lower Fly is chronically flooded with a relatively constant freshwater discharge.

Fig. 4 shows the time evolution of the computed river bed. The longitudinal profile initially displays a depositional sediment front propagating downstream in the tidal-dominated reach and a scour in the river-dominated reach. After about 273 years, the river-dominated reach starts to slowly infill and equilibrium is reached at about 7000 years. This evolution can be explained by considering that at the beginning of the simulation the constant slope profile is out of equilibrium, since the maximum total load has a large peak at the width break (Fig. 3d). As a result, a positive gradient of sediment discharge establishes upstream, while a negative one forms downstream. The flow responds to these gradients by scouring the upstream part of the reach and depositing the scoured material downstream of the width break, forming a sediment front that propagates toward the river mouth (Fig. 4, t = 100 and 200 years).

As the downstream portion of the river infills with sediment (Fig. 4, t ≥ 5000 years), the water depth decreases and friction, which scales with the inverse of depth (Eq. (4)), increases. Larger dissipation leads to a reduction of the tidal range (with respect to the initial topography) as the tide propagates along the channel. This is evident also in Fig. 5a–b, in which the maximum and minimum water level and velocity are plotted along the channel at the instant at which the bed level in the upstream part of the channel is lowest, i.e. after about 273 years, for spring-tide conditions. The tidal range decreases (Fig. 5a) as the tide dissipates over the shallower downstream region, and the corresponding ebb velocity (i.e., directed seaward) slightly decreases in the central part (i.e., in the interval 140–250 km), while it tends to increase in the upstream portion of the reach (i.e., in the first 140 km). The flood velocity, on the other hand, tends to decrease in the upstream region where bed degradation occurred. These two effects both lead to less tidal flushing, i.e. to larger sediment trapping in the upstream part of the tidal river with respect to the initial conditions. Neap tide conditions, instead, are typically characterized by an overall flattening of the water level along the river for a period of 2–3 days (Snowy Mountains Engineering Corp., 1983). These conditions can then be analyzed, as a first approximation, by plotting steady water and velocity profiles that establish when no tide is prescribed downstream (Fig. 5c,d).

After 273 years, in the absence of tides, in the part of the river with constant width where scour occurs, the depth increases seaward (Fig. 5c), and therefore the velocity decreases in the same direction (Fig. 5d). The opposite is true at the beginning of the simulation. Therefore, both the reduction of tidal flushing during spring tide and the negative gradient of velocity during neap tide in the upper part of the reach lead to a slow aggradation of the estuary (Fig. 4). As a consequence, a decrease of velocity gradient occurs and an equilibrium is reached after about 7000 years. Owing to aggradation within the estuary, the equilibrium configuration is characterized by a further decrease in tidal range, velocity and discharge along the reach (Fig. 6a–e). Note that the simulation has actually been run for a total time of 30,000 years but not noticeable changes occur after the equilibrium condition was attained. At equilibrium, the maximum total load is constant in the river dominated region (Fig. 6f), while it increases significantly in the tidally-dominated. Finally, note that the gradients of maximum total load present at the beginning of the simulation in correspondence of width break (Fig. 3f) are smoothed out (Fig. 6f).

4.3. Comparison with available data

Detailed bathymetric data are available only for the delta area (Daniell, 2008). For the remaining of the tidal river, the only available data consist of water depths drawn on the navigation charts provided by the Snowy Mountains Engineering Corp. (1983). The relevant values have been manually digitalized and referred to a common datum. Note that since the data were collected for navigation purposes, they refer to the deepest part of the river (thalweg). The total length thus obtained is slightly larger than the one resulting from the model, calculated with respect to the river axis. The digitalized data have been linearly compressed in order to make the total distances coincide.

The data are shown in Fig. 7a, together with the equilibrium bed profile computed for an upstream total load of 0.1 m³/s, corresponding to the pre-mining estimate (see Section 5). For a better comparison, in Fig. 7a also a profile in which the higher frequencies of bed elevations are filtered out is shown. Note that these high-frequency oscillations are likely associated with scours next to the sediment bars typically forming along the entire river reach. Despite some differences (more pronounced in the interval 180–280 km), there is a general agreement between the model results and the observed data, both showing an upstream region with lower elevation and sloping seaward and an aggraded downstream region. In particular, the elevations at the mouth and at the upstream end match quite well. The inset in Fig. 7a
Fig. 5. Upper plots: maximum and minimum values of the along channel water elevation (a) and cross-sectionally averaged velocity (b) for spring tide conditions. Lower plots: (c) Backwater profile and bed elevation at the beginning of the simulation and at the instant of maximum scour (no tide). (d) Velocities at the beginning of the simulation and at the instant of maximum scour (no tide).

Fig. 6. Water elevation (a) and velocity (b) at four instants during spring tide. Maximum and minimum values along the channel for water surface (c), velocity (d), discharge (e) and total sediment transport (f). All the quantities are computed when the bed has reached equilibrium.
indicates the point at which a break in bed elevation occurs, corresponding with the beginning of estuary flare (Fig. 3). The model well captures this sudden downstream increase in bed elevation at that location. The differences in the profile are likely related to the simplifications adopted in the model, and in particular to the use of a smoother longitudinal variation of the width, that however largely simplify the analysis of the different physical processes shaping the estuary. Finally, it is also worthwhile to observe that the bed equilibrium profile shown in Fig. 7a is such that the mean water surface elevation at Everill junction (about 5–6 m a.m.s.l., see Fig. 6a) coincides with the local bank elevation (Day et al., 2008).

4.4. Effect of different exponential width profiles and variations in sediment supply

Fig. 7b shows the comparison between the observed bed configuration and the equilibrium longitudinal bed profile obtained by considering various values of the incoming sediment discharge, i.e., varying it by ±10% (0.09 and 0.11 m$^3$/s), +40% (0.14 m$^3$/s) and −30% (0.07 m$^3$/s) with respect to the reference value. 0.1 m$^3$/s. It is evident that an increase in sediment discharge leads to a larger aggradation in the estuary. On the contrary, a decrease of discharge determines a lower aggradation and eventually triggers scour. A ±10% variation of the total load, however, slightly influences the final equilibrium profile. For larger variations (−30% and +40%), the results are drastically different. A large scour (up to 60 m below m.s.l.) occurs for a 30% reduction of the total sediment load. It is clear that in this case the hypothesis of fixed-width sections might lose its validity, since such large depths would likely trigger bank failure and a consequent enlargement of the section. On the contrary, for a sediment discharge of 0.14 m$^3$/s (+40%), bed aggradation in the funneling region leads to elevations that are close to mean sea level. Note that for this large aggrading conditions, floodplain flooding and the related sediment deposition (not accounted for in the model) would progressively become the dominant processes, enhancing the likelihood of river avulsions (Slingerland and Smith, 2004).

A larger sediment-load could also increase deposition close to the river banks, therefore creating a narrower cross section. To assess this possible effect, three additional simulations were carried out with a reduced estuary flare. In these simulations, the point of transition between the upper, nearly constant width reach and the downstream funneling region was kept fixed, while the width at the delta mouth was reduced by a factor 2, 4 and 8, respectively. Note that these configurations imply that also the convergence length was increased in order to keep fixed the transition point. Fig. 8 shows the equilibrium bed profiles computed with a sediment discharge of 0.14 m$^3$/s for both the original and modified geometries. The differences between the computed profiles are small, indicating that a narrower estuary does not necessarily implies a much lower aggradation. In fact, a decrease of the degree of funneling implies a decrease of the area subjected to tidal excursions, and therefore a smaller tidal prism. This, in turn, implies reduced discharges and tidal flushing. However, due to the smaller cross-sectional area, the freshwater discharge provides a larger contribution to the total discharge, therefore increasing the ebb-dominance of the system. In the configurations considered in Fig. 8, these two counteracting effects approximately balance each other, and the final equilibrium profiles are similar.

4.5. Effect of a different initial conditions

Here we assess the effect of different initial bed profiles on the final equilibrium geometry. We computed the equilibrium bed profile starting from 4 different configurations: two with a constant bed elevation of −10 m and −20 m a.m.s.l., and two with a sloping profile similar to that shown in Fig. 4, but translated 5 m up or down. While the evolution is different, the final equilibrium configuration is the same for all simulations (Fig. 9). In general, the farther the initial condition is from the equilibrium profile, the longer is the time needed to reach equilibrium.

Note that this is strictly true when a single grain size is considered as in the present model. When modeling multiple grain sizes with
 formation of different sedimentary layers, the final results could depend on the initial distribution of sediments.

4.6. Role of tidal constituents

Studies dealing with the long-term morphodynamics of dead end tidal channels usually assume negligible upstream inputs of water and sediment and impose a purely sinusoidal semidiurnal M2 tide at the channel mouth (see, among others, Dastgheib et al., 2008; Seminara et al., 2010; Toffolon and Lanzoni, 2010). In this section we will show that in the case of tidal rivers in which upstream water and sediment discharges are present, neglecting some tidal components can drastically change the final equilibrium configuration. Fig. 10 compares the equilibrium bed profile shown in Fig. 4 to the equilibrium profiles computed by prescribing: a single component M2; the sum of M2 and S2; the sum of M2, S2 and N2; a M2 component with amplitude $A = 2.54$ m (i.e., equal to the sum of the amplitudes of the six principal tidal components observed at the mouth of the Fly River, and reported in Table 1). The resulting equilibrium profiles are radically different. In general, increasing the number of harmonic components leads to a deeper equilibrium bed configuration, due to an increase of tidal flushing. On the contrary, smaller tidal ranges lead to a larger aggradation in the entire estuary. We also note that including further (small) components beside the six considered in this work did not appreciably modify the equilibrium profile.

We have also investigated the possibility of a lumped approach, i.e. replacing the sum of the various components with a single M2 constituent having an amplitude $A$ equal to the sum of the amplitudes of each constituent (in this case 2.54 m), an approximation usually employed when studying tidal channel morphodynamics. Fig. 10 clearly shows that with this approach a very large scour forms within the tidal river, due to the lack of neap tide conditions. In fact during neap tides most of the sediment fed into the system deposits in the wider section of the delta region, i.e. where the velocity drops (Fig. 5b), preventing excessively scour at the bed. A lumped approach is appropriate only in the absence of appreciable upstream water and sediment inputs. In this case, under neap-tide conditions a vanishing small sediment transport establishes in the channel due to the low velocities, and only spring tide conditions, adequately described through a single tidal constituent, are responsible for the shaping of the river bed (Toffolon and Lanzoni, 2010). This tide can then be considered a sort of geomorphologically significant tide, occurring with a particular frequency, analogously to the formative discharge usually employed to model river morphodynamics (Wolman and Miller, 1960).

![Equilibrium bed profiles obtained by imposing different tidal conditions at the river mouth. The reference equilibrium profile is the same shown in Fig. 4, computed by prescribing all the six tidal constituents characterizing the tide at the mouth of the Fly River delta. The considered tidal forcings consist of: the sum of M2, S2 and N2; the sum of M2 and S2; the component M2, a single harmonic M2 with amplitude $A = 2.54$ m. The tidal harmonics are reported in Table 1.](image)

**Fig. 10.** Equilibrium bed profiles obtained by imposing different tidal conditions at the river mouth. The reference equilibrium profile is the same shown in Fig. 4, computed by prescribing all the six tidal constituents characterizing the tide at the mouth of the Fly River delta. The considered tidal forcings consist of: the sum of M2, S2 and N2; the sum of M2 and S2; the component M2, a single harmonic M2 with amplitude $A = 2.54$ m. The tidal harmonics are reported in Table 1.

<table>
<thead>
<tr>
<th>$A$</th>
<th>$\theta$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$M_2$</td>
<td>0.921</td>
</tr>
<tr>
<td>$S_2$</td>
<td>0.590</td>
</tr>
<tr>
<td>$N_2$</td>
<td>0.419</td>
</tr>
<tr>
<td>$K_1$</td>
<td>0.305</td>
</tr>
<tr>
<td>$K_0$</td>
<td>0.160</td>
</tr>
<tr>
<td>$O_1$</td>
<td>0.147</td>
</tr>
</tbody>
</table>

Table 1

Amplitude $A$ and phase $\theta$ of the 6 main tidal harmonic constituents at Umuda Island. Courtesy of the National Tidal Centre, Australia.

Please cite this article as: Canestrelli, A., et al., One-dimensional numerical modeling of the long-term morphodynamic evolution of a tidally-dominated estuary: The Lower Fly River (Papua New Guinea), Sedimentary Geology (2013), http://dx.doi.org/10.1016/j.sedgeo.2013.06.009
5. Discussion

5.1. Morphodynamic equilibrium

Recently, Parker et al. (2008) studied the Fly River response to rising sea-level by using a moving-boundary, one-dimensional numerical model, describing the temporal evolution of the delta topset–foreset break and the forest–subaqueous boundary with the basement. The model includes backwater effects, but neglects tides. Since the last glacial maximum, starting about 20 ka before present, sea level has risen approximately 120 m over about 13 ka (transgressive phase). According to the numerical results obtained by Parker et al. (2008), sea-level rise has forced the river mouth to transgress over 700 km. Transgression was characterized by autotreat (in the sense of Muto, 2001) with abandonment of the river delta (i.e., sediment starvation at the topset–foreset break). In the last 7 ka, characterized by a relatively constant mean sea level (highstand phase), a new delta formed and started prograding outward.

The evolution of the Fly River from the last glacial maximum including tidal processes is outside the scope of the present paper, and would probably require a more complex (e.g., a two- or three-dimensional) model. Moreover, the tidal constituents used here would not be applicable to a period in which the volume of water in the oceans and mean sea level were largely different from the present values. However, as a first step towards the understanding of the effect of tides on the long-term evolution of the mean (cross-sectionally averaged) longitudinal profile of the Fly River, we focused our attention only on the relatively constant mean sea level period (high stand) characterizing the last 7 ka, studying the equilibrium of the estuary around the current altimetric and planimetric configuration.

Parker et al. (2008) argue that the Fly River may be presently prograding seaward into the Fly Estuary. In other words, the beginning of the estuary flare, assumed to coincide with the point reached by the fluvial channel, denotes the present-day position of a bay-head delta. The present results, suggest that the end of the fluvial reach, rather than corresponding to the location of a bay-head delta with a seaward decreasing bed elevation, is characterized by a local increase of river bed, connecting a deeper river-dominated part to a shallower tidal-dominated region. These findings are also supported by the available bathymetric data. Moreover, the present model also indicates that the time needed for the Lower Fly River to reach a configuration close to the present one is of the order of the 6000 years (Fig. 9), i.e. of the same order of the highstand phase duration. Therefore, it is likely that the present cross-section-averaged longitudinal profile of the Lower Fly is close to a condition of dynamical equilibrium (on a thousand-year timescale) in the sense of Seminara et al. (2010), and is actively exporting most of the sediment load to the ocean. Note that, based on the presence of beach ridges located half way between the present delta mouth and the delta apex, Dalrymple et al. (2003) inferred that the mouth transgressed about 40–50 km since the last glacial maximum, pointing out how this estimate is compatible with the present rate of delta progradation (6 m/year, Harris et al., 1993). The extent of this transgression is a relatively small fraction of the total length (360 km) of the river reach here considered, and, hence, can be neglected when studying the river equilibrium profile during high stand. For longer time scales, however, under a relatively constant mean sea level, the tide-dominated region would likely translate seaward as a whole, keeping its exponential shape as dictated by the tidal climate. This could be identified as the dynamical equilibrium at a geological scale for a tidally dominated delta prograding under a constant sea level. In this case, the water surface slope in the entire River system (from highland to the sea mouth) would decrease and some accumulation is likely to occur. This aspect, however, deserves further investigation and calls for long term coupled modeling of estuarine and coastal processes.

5.2. Sediment budget scenarios

The present results indicate that, given a particular planform geometry of the estuary, only for a relatively small range of sediment discharge an equilibrium bed configuration is attained with water depths similar to those observed nowadays. If the sediment discharge is increased, deposition is enhanced and the flow depth progressively reduces, possibly favoring avulsion. As shown in Fig. 8, even a decrease in river width (e.g., by bar accretion near to the banks) would not prevent aggradation. On the contrary, a reduction of sediment discharge would lead to a widespread scour (Fig. 7b) which, in turn, could favor bank collapse and, therefore, an enlargement of the river.

The considerations in Section 4.2 rely on the assumption that the mean sediment load reaching the Lower Fly is about 0.1 m³/s. However, this estimate can vary when considering the information provided in the literature. A general agreement exists on the sediment load entering the Middle Fly River at D’Albertis Junction before mining activities (about 10 Mt/a, half of which flowing from the Ok-Tedi River and half from the Upper Fly (Fig. 1)). A sediment load of 75 Mt/a seems a good estimate for the highlands area of the Strickland River (Markham and Day, 1994). How much of this load is lost in his path to the sea? As for the Middle Fly, sediment is dispersed on the floodplains and through tie/tributary channels and goes into a long-term net storage of the order of 40% of the incoming load (Dietrich et al., 1999; Day et al., 2008). On the contrary, on the Strickland River, the relatively high meander migration (5 m/a) indicates that a balance between overbank deposition and bank erosion may occur (Aalto et al., 2008). Therefore, supposing that 40% of the load entering the Middle Fly is lost, and that in the long term all the load flowing through the Strickland River and retained on the floodplains is eventually returned, the estimate of the load entering the Lower Fly is 81 Mt/a. Following Pickup (1984) estimate that 10% of this is sand and supposing a sediment density of 2650 kg/m³, we obtain the estimate of the sand load entering the Lower Fly and Delta region of 0.1 m³/s. As shown in Fig. 7a, this value provides the most realistic equilibrium profile for the Lower Fly.

A slightly different value results by considering the suspended sediment discharge estimated by Salomons and Eagle (1990) on the basis of concentration samples, suggesting transport rates of 8 Mt/a and 66 Mt/a for the Middle Fly and the Strickland River at the Everill Junction. Supposing that bed-load would only slightly modify this estimate, a sediment discharge of about 74 Mt/a is expected at Gwaa, equivalent to 0.09 m³/s. The associated equilibrium profile, shown in Fig. 7b, is very similar to that obtained with the previously considered sediment input (i.e. 0.1 m³/s). The data provided by Dietrich et al. (1999), however, lead to higher estimates. Indeed, they speculate that 30% and 24–28% percent of the total load delivered by the Middle Fly (10 Mt/a) and the Strickland (75 Mt/a) may be sand. This would result in a very high sand load (0.26 m³/s), leading to an unreasonable high aggradation of sediment within the estuary (not shown). Nevertheless, as recently pointed out by Lauer et al. (2008), Dietrich et al.’s (1999) estimates were developed using only few years of data and are probably rather crude. On the contrary, Lauer et al. (2008) estimate the sediment load in the Middle Fly and Strickland through a backward calculation, ensuring that the sediment discharges selected to drive their numerical model are able to reproduce, on the long term, present-day channel width and slope near the upstream end of the investigated reaches, i.e. where the two river systems are likely close to grade. By using Eq. (5) they estimated sand transport rates for the Strickland and Middle Fly respectively of 8.2 Mt/a and 2.2 Mt/a, equal to 10% and 30% of the total load observed in each reach, the rest being mud. According to these estimates, the annual total load entering the Lower Fly would be about 0.12 m³/s: the corresponding equilibrium morphology can be inferred from Fig. 7b.

Concluding, from the plots of Fig. 7, and supposing that under pre-mining conditions (i.e., as of 1985) the Fly River has nearly
attained a morphodynamic equilibrium, the better estimates of the sand load delivered to the Lower Fly is in the range 0.09–0.11 m³/s. Lower sediment rates would lead to a deeper estuary, and presumably to larger cross sections, whereas larger rates would lead to aggradation of the delta region.

The above estimates are all referring to pre-mining conditions. Mining at the headwaters of the Ok Tedi, started in 1985, causing an increase in sediment discharge from 5 Mt/a to 45 Mt/a (Pickup and Marshall, 2009), implying a sediment load entering in the Middle Fly of the order of 50 Mt/a. The sediment load delivered to the Strickland after the beginning of mining (in 1991) increased of about 10 Mt/a, to a total of 85 Mt/a (Markham and Day, 1994). If we assume that the percentage of material depositing in the Middle Fly and the Strickland floodplains has not changed, the total load that currently enters the Lower Fly is about 115 Mt/a, quite close to the value of 119 Mt/a estimated at Ogwa by Markham and Day (1994) in 1992. Even though the exact amount of fine sediments produced by mining activities is unknown (Day et al., 2008; Swanson et al., 2008), we can however suppose that the percentage of sand reaching the estuary has remained essentially the same (about 10% of the total load): it then turns out that approximately 0.14 m³/s of sand is currently reaching the delta region. The present results suggest that with this sand load, the delta would largely tend to aggrade (see Fig. 7).

6. Conclusions

Our model results suggest that, given a particular planform geometry of a tide dominated river region, there is only a relatively small range of sediment discharges that leads to an equilibrium bed configuration compatible with the depth of the upstream fluvial reach. If the sediment discharge delivered to the system is increased, deposition would produce aggradation that reduces the flow depth such that it can no more match the flow depth in the river dominated reach. The river is not able to flush out the sediments even if the width of the delta region is reduced, since a decrease in tidal flushing would be approximately balanced by an increase of river-driven ebb dominance. On the contrary, a reduction of sediment discharge would lead to widespread scour, possibly associated with bank erosion and widening of the estuary.

Acknowledgments

This research was supported by NSF award # OCE-0948213. The second author was partially funded within the project “Morphodynamics of marsh systems subject to natural forcings and climatic changes” (University of Padua, Progetto di Ateneo 2010).

Appendix A

Eq. (1) is solved in fully divergent form in order to ensure mass conservation up to machine precision (Canestrelli and Toro, 2012). While the original Price-C scheme applied to the shallow water equations for a rectangular prismatic section automatically satisfies the C-property (i.e., it is well-balanced in the sense that it exactly solves a quiescent steady state, see Canestrelli et al., 2009), when the section is spatially varying Eq. (2) must be rewritten in a new form in order to satisfy such a property. The expanded form reads:

\[
\frac{\partial Q}{\partial t} + \frac{2Q}{B(H-z_2)} \frac{\partial Q}{\partial x} - \frac{Q^2}{B(H-z_2)^2} \frac{\partial H}{\partial x} + \frac{Q^2}{B(H-z_2)^2} \frac{\partial z_2}{\partial x} - \frac{Q^2}{B^2(H-z_2)} \frac{\partial B}{\partial x} \frac{\partial H}{\partial x} = -gBR \delta_y.
\]

Moreover, a reduced diffusion version of the Price-C scheme is here employed, in order to limit the artificial diffusion in the continuity equation that prevents well-balancing. The first row of the modified identity matrix \(P \) as defined in Canestrelli and Toro (2012) is multiplied by \( \min(Fr_{1/2} + Fr_{1/2}, F_{\text{lim}}) \), where \( Fr_{\text{lim}} = 0.0001 \) and \( Fr_{1/2} = (Fr + Fr_{1/2})/2 \), i.e. being the computation cell under consideration and \( Fr \) being the Froude number. The results obtained by using this modified version of the Price-C scheme were compared with those obtained by applying the fully upwind Roe-type method of Parés and Castro (2004): no substantial differences between the final bed equilibrium profiles were observed. However, we note that, since the present scheme is centered and does not require the explicit knowledge of the eigenstructure of the system, the computational time is significantly reduced (by a factor of five) with respect to a fully upwind Roe-type scheme, thus significantly decreasing the duration of long term morphodynamic simulations.

The scheme is extended to a second order of accuracy through the 1-D-equivalent of the 2-D MUSCL reconstruction procedure described in Canestrelli et al. (2012).

Finally, note that even though a non-divergent form is used for the momentum Eq. (A1), it can be easily proven that the scheme automatically reduces to a modified conservative FORCE scheme if the underlying partial differential system is a conservation law, i.e. for a prismatic rectangular channel with flat bottom (Canestrelli et al., 2009).

References


