Field and Modeling Studies of Fine Sediment Dynamics in the Extremely Turbid Jiaojiang River Estuary, China

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ABSTRACT


The Jiaojiang River estuary, China, is very turbid with depth-averaged suspended sediment concentration commonly exceeding 10 kg m⁻³ at spring tides. An asymmetry of the tidal currents is caused by shallow water effects. This asymmetry generates, at spring tides, a turbidity maximum in the estuary near the salinity intrusion limit. The rates of erosion and deposition are controlled by the tidal currents and the constants in the equations for those rates were inferred from 12 days of data. A transient sediment layer forms at slack tide and is eroded at a threshold velocity of about 0.3–0.4 m s⁻¹ which is about half the value of the erosion threshold velocity of the consolidated sediment underneath. The data were used to calibrate a sediment dynamics model which was successful at reproducing most of the observations at spring tides. The model suggests that the bulk of the sediment in the estuary is imported from coastal waters. This sediment originates from the Changjiang (Yangtze) River which is located 200 km further North. The Manning's bottom friction coefficient is found to be smaller (by 30%) than that in non-turbid estuaries and this may be due to the presence of a fluid-mud layer near the bottom.

ADDITIONAL INDEX WORDS: Estuary, fine sediment, transport, hydrodynamics, modeling, China.

INTRODUCTION

The Jiaojiang River estuary, China (Figure 1), is located 200 km south from the Changjiang river (Yangtze river). The estuary is only about 35 km long. Mean depth is about 1–3 m at low spring tide. The maximum width is 1.8 km but this width increases rapidly in Taizhou Bay in shallow, coastal waters. The freshwater discharge varies seasonally, the mean discharge is 106 m³ s⁻¹ over the period 1951–1984 but 70% of this water was delivered between July and September during the wet summer monsoon (Bi and Sun, 1984). The Jiaojiang estuary is semi-diurnal macro-tidal. At the mouth the mean tidal range is about 4 m. The maximum tidal range is 6.3 m and the vertically-averaged peak tidal current peaks at 2.0 m s⁻¹. The tidal wave is strongly distorted in the estuary over a short distance; at Haimen the flood tide lasts 5.1 hour while the ebb tide lasts 7.3 hours. The tidal currents also exhibit a similar asymmetry, with maximum flood and ebb currents of, respectively, 2.1 and 1.8 m s⁻¹ (Zhou, 1986).

Engineering and scientific research has focused on the tidal hydrodynamics in the Jiaojiang estuary (e.g. Bi and Sun, 1984; Fu and Bi, 1989), but the lack of data have precluded an understanding of fine sediment dynamics and the development of a numerical model for sediment dynamics. The Jiaojiang River estuary is highly turbid, Li et al. (1993) have measured vertically averaged suspended sediment concentration (SSC) exceeding 5 kg m⁻³. The suspended sediment is mostly silt and clay with a mean particle size of 6 to 8 µm. About 60% by volume of the particles are very fine silt (size range between 4 and 16 µm), the clay (particle size less than 4 µm) accounts for 35% (Li et al., 1993). The sediment riverine inflow is estimated to be 1.23 × 10⁶ tons year⁻¹, corresponding to a mean SSC of 0.18 kg m⁻³, i.e. about 4% the SSC in the estuary. In turn the Jiaojiang sediment discharge is only about 1% that of the Changjiang River (1.2 × 10⁶ tons year⁻¹) (Yu et al., 1987). Some of the fine sediment is carried by coastal current to Taizhou Bay in coastal waters off the Jiaojiang River estuary. The Jiaojiang River estuary silts at a rate of about 0.2 m year⁻¹ and to maintain navigation dredging is required every year (Li et al., 1992).

In this study we present field data on the sediment dynamics of the estuary. Tidal currents, and not baroclinic effects, are the dominant transport mechanism at spring tides. We use these data to calibrate a sediment dynamics model for the Jiaojiang River estuary. The model is a advection-diffusion model with source and sink terms to represent erosion and settling. We show that the major source of sediment is the coastal sediment; this sediment is pumped in the Jiaojiang River estuary by the tidal asymmetry at spring tides.

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and accumulates in a turbidity maximum zone. This coastal sediment in turn originates from the Changjiang (Yangtze) river.

**METHODS**

Vertical profiles of temperature, salinity and SSC were measured at a number of stations along and across the estuary using the CTD-cum nephelometer described by WOŁANSKI et al. (1988). In addition at site M1 (Figure 1) a 2.7 m high frame was deployed, and contained an Inter-Ocean model S4 vector-averaging electromagnetic current meter and six self-logging optical fiber Analite nephelometers. Suspended sediment was also collected and used in the settling column of WOŁANSKI et al. (1992) to measure the settling velocity as a function of SSC in waters of different salinity.

**RESULTS**

Field Observations

Vertical stratification in salinity was negligible during the field study, at spring tides (Li et al., 1993).

The field data show some relationship, though very noisy, between depth-averaged suspended sediment concentration (SSC) and velocities (U) at spring tides (Figure 2) but no relationship seems to exist between SSC and U at neap tides (Figure 3). Further, the relationship at spring tide is asymmetric, the increase of SSC with U being more rapid during ebb tide than flood tide. The relationship shows also an hysteresis due to the lags for erosion, vertical mixing, and deposition (DYER and EVANS, 1989; WEST and SANGODOYIN, 1991).

Field data on currents and SSC (Figure 4) show that at ebb tide the peak velocity reaches 1 m s⁻¹ and varies little for 2 hours. By contrast the flood velocity is stronger (peaking at 1.2 m s⁻¹) but is short-lived; thus a strong tidal asymmetry prevails. The onset of erosion is characterized by a rapid, nearly step-like, increase in SSC after slack high tide when ebb tidal currents increase (Figure 4) to about 0.3-0.4 m s⁻¹. The maximum SSC (about 40 kg m⁻³) occurs shortly thereafter, about 2.5 hour after the slack high water. Later, SSC decreases to much lower values (typically 2-3 kg m⁻³) though the ebb tidal currents remain constant.

At spring tides the sediment forms an unconsolidated fluid mud layer on the bottom at slack tide. This slack-water deposit has high water content and low density and is eroded at the subsequent strong tidal currents. Once this layer is eroded, stronger tidal currents are needed to erode the underlying, consolidated sediment.

Near the end of the ebb tide, the depth-averaged SSC increases again; modeling (see later) suggests that this second increase is due to the advection of water from the turbidity maximum zone located upstream from mooring M1.

At flood tide, the tidal currents also erode the slack water deposit layer, but the currents, being stronger, erode more of the underlying consolidated bottom. The highest SSC occurs at flood tide, 3 hours after the slack low water. These findings show that a slack water deposit forms and has a different critical shear stress than the underlying consolidated bed sediment.

**Sediment Erosion Laws**

The field data enable us to derive the constants in the erosion law for cohesive sediment (ODD and OWEN, 1972)
Figure 2. Relationship between depth-averaged suspended sediment concentration and water velocity during spring tide at station M1.

Figure 3. Relationship between depth-averaged suspended sediment concentration and water velocity during neap tide at station M1.
where \( dE/dt \) is the erosion rate, \( M \) is a constant having units of erosion rate, \( \tau \) is the shear stress and \( \tau_c \) is a critical shear stress. The erosion takes place only if the flow-induced shear stress \( \tau \) exerted on the estuarine bed surface exceeds \( \tau_c \), i.e., \( \tau > \tau_c \). The critical shear stress is determined by the nature of sediment, especially determined by the dry bed density \( \rho_b \), the general express is

\[
\tau_c = \alpha \rho_b^\beta
\]

where, the \( \alpha \) and \( \beta \) are the coefficients. THORN and PARSON (1980) give the values of \( \alpha = 8.42 \times 10^{-7} \) \( \beta = 2.28 \) and OWEN (1970) gives \( \alpha = 6.45 \times 10^{-6} \) \( \beta = 2.44 \). The data of Yangtze estuarine sediment show that the best values are \( \alpha = 1.85 \times 10^{-6} \) and \( \beta = 2.25 \) (unpublished data).

We estimated the thickness of the slack-water deposit as that of the fluid mud layer of concentration > 10 kg m\(^{-3}\), at slack tide, during spring tides; this layer disappears during strong tidal currents when the sediment is mixed in the water column. This thickness was typically 0.3–1.0 m. As the sediment flocs settle on the bed surface, compression and de-watering and consolidation may occur. However the consolidation process is slow; initial consolidation may last 10 hours (OWEN, 1970), which is much longer than the duration of slack water (<1.5 hours). There is, thus, not enough time for consolidation of the slack-water sediment deposit, hence the slack-water sediments are easily eroded. The relationship between SSC and velocities (Figure 2) suggest a threshold velocity of the slack water sediment of 0.35 m s\(^{-1}\) for flood currents, and 0.4 m s\(^{-1}\) for ebb tidal current.

After the slack-water sediment layer is eroded away (2.5 hours after slack high water current, and 3 hours after slack low water current respectively), an analysis of our data suggest that its threshold velocity ranges about between 0.67 m s\(^{-1}\) and 0.88 m s\(^{-1}\).

M remains an empirical constant. Values of M are generally in the range of 0.005 to 0.015 (kg m\(^{-2}\) s\(^{-1}\)). In the Jiaozhou River estuary, an examination of the data in Figure 4 for SSC at flood tide, after the slack tide deposit has been eroded, suggests \( M = 0.15 \), about ten times the value in the upper range of the experimental data of ARIATHUARI and ARULANANDAN (1978).

### Settling Laws

#### Settling in Quiescent Waters

Vertical profiles of SSC were measured at various times with a profiling nephelometer in the settling column described by WOLANSKI et al. (1992). The suspended sediment settling velocity \( w_s \) was calculated from the equation for mass conservation of sediment

\[
\frac{dC}{dt} + \frac{\partial}{\partial z} (w_s C) = 0
\]

where \( t \) is time, \( C \) the SSC and \( z \) is the elevation.

The non-linear dependence of \( w_s \) on SSC (Figure 5) is characteristic of cohesive sediment. Indeed there is a zone of flocculation settling, where \( w_s \) increases with increasing SSC, and a zone of hindered settling, where \( w_s \) decreases with in-
increasing SSC. This cohesive-like property was a priori unexpected since the sediment is mostly silt which is only weakly cohesive. The transitional zone between hindered settling and flocculation settling occurs for SSC values between 1.8 and 4.0 kg m\(^{-3}\). In this transitional zone, \(w_f\) is at a maximum of about 0.003 to 0.008 m s\(^{-1}\). The Jiaojiang River and the Fly River in Papua New Guinea (WOLANSKI et al., 1995; WOLANSKI and GIBBS, 1995) are apparently the only two silt-dominant estuaries for which settling velocities have been determined. The settling velocities of both estuaries are qualitatively similar, though in the Jiaojiang estuary the transitional zone of Jiaojiang estuarine sediment is shifted towards higher SSC (by about 30%), and the settling velocity in hindered zone seems to be larger (by about 25%).

\(w_f\) is also affected by the salinity. For example in the zone of flocculation settling for 0.1 < SSC < 1 kg m\(^{-3}\), \(w_f\) varies between 0.0001 to 0.0003 m s\(^{-1}\) in freshwater but has values ten times larger (typically 0.001 to 0.003 m s\(^{-1}\)) for salinities greater than 0.01. At higher salinities, little variation of \(w_f\) with S occurs. This result may be due to the mean floe size being invariant with salinity for brackish water (Li et al., 1993).

This dependence of \(w_f\) on S and SSC for the silt-dominant Jiaojiang River estuary is thus qualitatively similar to that in clay-dominant estuaries (Dyer, 1986). The following model for \(w_f\) was adopted and reproduces well the laboratory findings

\[
W_f = \begin{cases} 
0.107 C^{2.277}, & \text{for } C > 2 \text{ kg m}^{-3} \text{ and } S > 0.01 \\
0.6(0.5 C)^{0.286}, & \text{for } C < 2 \text{ kg m}^{-3} \text{ and } S > 0.01 \\
0.003, & \text{for } S < 0.01 
\end{cases}
\]  

settling in turbulent flows. Other laboratory experiments (not shown) in a settling column show that \(w_f\) (as calculated from Equation 3) decreases for increasing turbulence generated by vertically oscillating grids. However, the interpretation of this finding is ambiguous since turbulence creates an upwards diffusive flux of suspended sediment, not included in Equation 3, resulting in a smaller value of \(w_f\). This finding is also supported by in-situ findings of Li et al. (1993) that the large flocs (with higher \(w_f\)) in the Jiaojiang River estuary form only for tidal currents smaller than 0.5 m s\(^{-1}\) and are destroyed (hence a much smaller \(w_f\) results) for stronger currents. Cohesive sediment flocs settling on the bed can only adhere to the bed only if the shear stress, \(\tau\), which is exerted on the bed surface by the tidal current is lower than the initial bonding strength of the flocs.

To parameterise the effect of turbulence, we adopt the concept of a limiting shear stress of deposition, \(\tau_d\), so that no deposition occurs for \(\tau > \tau_d\) (Einstein and Krone, 1962). Our data (Figure 4) suggest that in Jiaojiang River estuary, the limiting shear stress corresponds to a velocity of 0.5 m s\(^{-1}\).

To calculate the deposition rate \(dD/dt\), we followed ODD and Owen's (1972) expression

\[
dD/dt = \begin{cases} 
Cw_f(1 - \pi/\tau_d), & \text{if } \tau < \tau_d \\
0, & \text{if } \tau \geq \tau_d 
\end{cases}
\]  

SEDIMENT TRANSPORT MODEL

The sediment transport model consist of two parts, a hydrodynamic model and sediment tracer model.
The Hydrodynamic Model

The modeling domain extends from Sanshan to Laoshuhan and covers 11.5 km along the estuary main channel (Figures 1 and 6). The model includes the straight dike located in the middle of estuary. The clockwise angle from north to the downstream direction of the dike is 100 degrees and this is also taken to be the direction of the x-axis (Figure 6). The grid size is 150 m.

The model follows that of Leenderstee (1967); it is two-dimensional (depth-averaged), finite-difference, and it uses the full unsteady, non-linear equations, including Coriolis forces, eddy diffusion and a non-linear friction law using a Manning's roughness coefficient. The solution algorithm uses the ADE numerical scheme of Flather and Heaps (1975). The observed tide is imposed as water surface elevation at the ocean open boundary. At the riverside open boundary, a tide curve is imposed from observations. The tide at this point is distorted by shallow water effects, and, in addition, the mean water surface is taken to be 3 cm higher because of the freshwater discharge. The horizontal eddy diffusion coefficients are taken to be equal to 10 m² s⁻¹.

The model reproduces faithfully the velocity data at the mooring site (Figure 7). To obtain this best fit between observed and predicted currents, we adopted a value of the Manning's friction coefficient of 0.015, a value 30% smaller than the usual value of 0.025 commonly used in less turbid estuaries (Henderson, 1966). This finding suggests that the suspended sediment generates a density-stratified fluid, so that the bulk of the water is able to essentially slip, with little interfacial stress, over the bottom fluid mud layer. A similar small value of Manning's friction coefficient was found in the turbid estuaries of the South Alligator and Fly rivers (King and Wolanski, 1995).

No vertical salinity gradient, only a horizontal one occur at...
spring tides. Salinity is important only for the settling velocity and is well simulated by a simple empirical formula.

**The Sediment Tracer Model**

The sediment transport model is a Lagrangian solution of the advection-diffusion equation for suspended sediment,

\[
\frac{\partial C}{\partial t} + \mathbf{u} \cdot \nabla C = \nabla \cdot (\mathbf{E} \nabla C) + S
\]

where \( \mathbf{u} (u_x, u_y) \) is the velocity field, \( \mathbf{E} \) the diffusion tensor and \( S \) is a sink or source term. The sink and source terms are taken to be the sum of erosion (Equation 1) and settling (Equation 5). To calculate the transport of sediment in and out of the hydrodynamics model domain, the sediment transport model domain extends about 20 km and 12 km respectively further upstream and downstream, with velocities and depth in these sponge layers equal to those at respectively the upstream and downstream open boundaries of the hydrodynamic model. The velocities and depth at each grid point were taken from the results of the hydrodynamics model; these predictions for water surface elevation and velocities were stored in a file with a time interval of 621 seconds. The discretization schemes of the hydrodynamics and sediment transport models are the same.

The sediment dynamics were simulated by tracking a large number of sediment particles (drogues) each weighing 50 tons. Diffusion is modeled as a random walk process. A particle can deposit or be eroded following Equations 1 and 5. A deposited particle remains stationary until it is entrained in suspension. The position \( \mathbf{X}_i (t) \) at time \( t + dt \), where \( dt \) is the time step, is calculated from its position at time \( t \) following Equation 6 written in a Lagrangian representation

\[
\mathbf{X}_i (t + dt) = \mathbf{X}_i (t) + dt \mathbf{u}_i + \Delta (\cos \theta, \sin \theta)
\]

where \( \Delta \) is a random number between 0 and \( 2\pi \) and \( \Delta \) is a random walk length over time \( dt \)

\[\Delta = a\sqrt{2E_x + 2E_y} dt\]

where \( a \) is a random number between 0 and 1. The velocity field \( \mathbf{u} \) is provided by the hydrodynamics model. This Lagrangian technique is a classical one in oil spill modeling (e.g. Al-Rahem et al., 1989; Al-Rahem and Gunay, 1992; Spaulding, 1988) and has also occasionally been used in sediment transport models (Wolanski et al., 1995).

The threshold erosion velocity was determined separately for the consolided sediment and for the slack-water deposit. The consolidated sediment was assumed, from a visual examination of the field data, to have a threshold erosion velocity of about 0.67 m s\(^{-1}\) to 0.88 m s\(^{-1}\), and this value coincidentally was also that measured in the Yangtze River estuary (Dong, unpubl. data). For the slack-water deposit however, our data suggest a much smaller threshold velocity, about 0.35 to 0.4 m s\(^{-1}\). In the model, the slack-water deposit is initially eroded, this typically lasts 2 hours after high water slack and 3 hours after low water slack, respectively. Therefore the consolidated sediment is exposed and eroded, and this is modeled by increasing the threshold erosion velocity to 0.67–0.88 m s\(^{-1}\) for near surface bed sediment. The threshold velocity increases linearly with depth (or time) in the model to a maximum value of 1.2 m s\(^{-1}\) for the deepest or oldest sediment.

The limiting velocity of deposition is taken to be 0.5 m s\(^{-1}\) most of the time. However during erosion of the slack-water layer, the limiting velocity is set equal to the threshold erosion velocity for that layer.

The predicted SSC at the mooring site reproduce faithfully the observations (Figure 4). This suggests that the model has captured the key physical processes operating in the Jiaojiang River estuary. Note that the model is successful in reproducing the strong tidal asymmetry of SSC, with two peaks at ebb tide but only one peak at flood tide.

The model also shows the presence of a turbidity maximum zone located upstream of Haiman (Figure 8) around the dike; there is supporting evidence for the existence of this turbidity maximum zone in the observations of Li et al. (1993). The model was run without the dike, and the results indicated that the turbidity maximum zone still would exist but would be shifted downstream towards Haiman where it would exacerbate siltation and dredging costs for the harbor.

The model is quite sensitive to the assumptions for \( w_f \) (Equation 4). Sensitivity tests were carried out. For instance, assuming that \( w_f \) is the same in fresh water and salt water, or alternatively assuming a constant \( w_f (0.00007 \text{ m s}^{-1}) \), leads to predictions where in one case SSC decreases too rapidly after reaching a maximum value at peak flood tidal currents, or in the other case SSC values being markedly too small at slack water.

**DISCUSSIONS AND CONCLUSION**

Our model suggests that siltation of the Jiaojiang River estuary is due primarily to the import of coastal sediment by the tidal pumping effect due to the asymmetry of the tidal currents (Drönkers and van Leussen, 1988). The asymmetry of tidal currents results in a maximum flood tidal current at the mooring site 20% larger than the maximum ebb tidal current. The coastal sediment in turn originates from the Yangtze River (Yu et al., 1990). The coastal sediment inflow is estimated by our model to be 3–5 times larger than the inflow of riverine sediment (1.23 × 10^6 tons year\(^{-1}\)). Thus our study suggests that the Jiaojiang estuary is silting primarily with Yangtze River sediment.

A suspended sediment particle entering the Jiaojiang River estuary from Taizhou Bay can readily reach it in a few tidal cycles at spring tides. In addition, in the coastal waters of Taizhou Bay, a vertical stratification in salinity has been observed, and the resulting gravitational circulation would also help pump fine suspended sediment towards the estuary.

The tidal dynamics were modeled successfully by a vertically-averaged hydrodynamics model assuming that the Manning’s friction coefficient is 0.015, a value significantly smaller than the usual value (0.025) for less turbid estuaries. This may be due to the suspended sediment stratifying the water column, so that the bulk of the water is able to slip with little friction over a fluid mud layer near the bottom.

The estuary is highly turbidity estuary, with a maximum depth-mean SSC peaking at 10 kg m\(^{-3}\) at the mooring station.
M1. Two peaks in the SSC occur at ebb tide, but only one at flood tide. The sediment is eroded and deposited at tidal frequency, a slack-water sediment deposit forming at slack currents. This layer can be eroded at fairly low velocity (0.35 m s⁻¹) after slack tide. After this slack water deposit is eroded away, the underlying consolidated bed sediment is exposed and can be eroded but its critical erosion velocity is estimated to be much larger (0.67-0.88 m s⁻¹).

The model reproduces well the observations at the mooring site M1 that at ebb tide the depth-averaged SSC peaks at
about 10 kg m$^{-3}$ 2.5 hours after slack high water current; this first peak in SSC is due to the erosion of the slack water sediment deposit. After this first peak, the depth-averaged SSC decreases to about 2-3 kg m$^{-3}$. The model successfully reproduces also the observation of a second peak of SSC, with values reaching 10 kg m$^{-3}$, and the model suggests this second peak at site M1 is due to the advection of water from the turbidity maximum zone. The model also successfully reproduces the finding of only one peak of SSC at flood tide, and this peak occurs at about the same time as the peak flood tidal current. The model shows the presence of a turbidity maximum zone located upstream of Haiman (Figure 8). Numerical experiments suggest that the dike is important in locating, but not in creating, the turbidity maximum.

Wet season data are unavailable but are needed to assess whether occasional floods can help flush sediment offshore which would happen if salt was flushed out of the estuary. In that case the construction of dams on the river may exacerbate siltation of the estuary. Presumably the fluid mud layer will move down-river during occasional river floods (UNCLES and STEPHENS, 1993) but no data are available to assess the importance of this process for the Jiaojiang River estuary.

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LITERATURE CITED


