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# Two-dimensional modeling of fine sediment transport with mixed sediment and consolidation: Application to the Gironde Estuary, France.

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# 1 **1. Introduction**

2 Located in the southwest of France, the Gironde Estuary is considered as the largest estuary in Western  
3 Europe (180 km in length and 20 km wide, Fig. 1). In the lower and central areas, the Gironde fluvial-  
4 estuarine system is characterized by a pronounced turbidity maximum zone (TMZ) that plays an  
5 important role in the sediment transport and morphodynamic of the study area (Jalón-Rojas et al., 2015;  
6 Sottolichio & Castaing, 1999). In such a highly turbid estuary (1-10 g/L near the bed), the maintenance  
7 of the navigation channel is an important issue which arises with the growth of the ship capacity and  
8 the predominance of maritime transport. Under this context, Grand Port Maritime of Bordeaux  
9 (GPMB) has funded the Gironde XL 3D project, with the support of the European Union (EU). This  
10 project aims to develop numerical tools to safely accommodate larger ships by better anticipating  
11 channel maintenance requirements and operations. For both objectives, hydrodynamic and sediment  
12 transport models are needed over different time scales. On the one hand, the model should be able to  
13 predict in real time water levels, flow recirculation and concentrations of suspended particulate matter  
14 (SPM) over short periods of time (36 hours) in order to maintain a safe under keel clearance for the  
15 ship navigation for given hydrological conditions. On the other hand, the annual time-scale must be  
16 considered for channel maintenance operations.

17 Over the years, several numerical models of the Gironde Estuary have been developed. The models can  
18 be grouped according to their usage with the first group mainly dedicated to hydrodynamics (Denot et  
19 al., 2000; Huybrechts et al., 2012; Laborie et al., 2014). The second group analyzes the turbidity  
20 maximum dynamics via three-dimensional models with a fixed bed configuration (Cancino & Neves,  
21 1999; Li et al., 1994; Sottolichio et al., 2001, Diaz et al., 2018, van Maanen & Sottolichio, 2018). The  
22 last group focuses on two-dimensional, depth-averaged models (2DH) of sediment transport and bed  
23 evolution (Chini, 2007; Huybrechts & Villaret, 2013; Van, 2012; Villaret et al., 2011). Huybrechts and

24 Villaret (2013) simulated a 5 year morphodynamic evolution using three sand classes, while Villaret et  
25 al. (2011) and Van (2012) did numerical simulations considering cohesive sediment and bed  
26 consolidation.

27 The mouth of the estuary is mainly composed of sand (> 90%) while a sand-mud mixture is observed  
28 in the intermediate estuary (75 % of mud upstream the P2 station on Fig. 1 according to the GPMB). In  
29 the former area, some migrating sand banks also are observed (Kapsimalis et al., 2004). A 2DH  
30 modeling approach is selected here as a good compromise between computational cost and hydro-  
31 sedimentary processes. Compared to previous works (Huybrechts & Villaret, 2013; Van, 2012), a sand-  
32 mud approach is chosen in order to better reproduce the variability of the bed composition. In  
33 consequence, even if the model will also be used to predict bed evolution, this paper focuses on the  
34 development and the validation of a hydro-sedimentary model. In this way, the model must provide  
35 reliable predictions of water levels, current velocities, and SPM concentrations to further set up a  
36 Computational Fluid Dynamics (CFD) ship squat model (Ali et al., 2018). To the best of the authors  
37 knowledge, few models combining mixed sediment and consolidation processes have been applied for  
38 the Gironde Estuary.

39 This paper is organized as follows. In Section 2 the main features of the Gironde Estuary are briefly  
40 presented. In Section 3, the hydro-sedimentary mathematical model and its numerical solution are  
41 introduced. In Section 4, numerical model calibration and validation are done for different hydrological  
42 conditions using measurements of water level, current velocity, salinity, and suspended-sediment  
43 concentration collected during the project. In Section 5, a sensitivity analysis for sediment parameters  
44 is done to improve the knowledge and the prediction of suspended-sediment concentrations in the  
45 Gironde Estuary. Ongoing studies also are discussed in the latter section.

## 46 **2. Study area**

47 The Gironde Estuary is a macrotidal and convergent estuary with a tidal range varying from 1.5 m  
48 during neap tides to 5.5 m during spring tides at the estuary mouth (Fig. 1). The propagation of the tide  
49 along the estuary induces an amplification of tidal waves and an asymmetry in the rise and fall of the  
50 water level with values of 4 h and 8 h 25 min, respectively. The limit of the tide propagation is located  
51 about 170 km upstream of the estuary mouth. Dordogne River and Garonne River contributions to the  
52 freshwater discharge are estimated to 35% and 65%, respectively (Sottolichio, 1999). For 2018, the  
53 total daily river discharge varied from 133 to 5560m<sup>3</sup>/s-1 during low and high river discharges periods,  
54 respectively. The annual volume of freshwater delivered to the sea is approximately 2.5x10<sup>10</sup>m<sup>3</sup>. The  
55 highest velocities are observed in the navigation channel for the intermediate estuary and on the side  
56 channel for the lower estuary. Near the bottom, average current velocities are approximately equal to  
57 0.75 m/s-1 during spring tides and never exceed 0.5m/s-1 during neap tides (Castaing, 1981).

58 The Gironde Estuary is considered the largest estuary in Western Europe, with the width varying from  
59 20 km at the estuary mouth to 3 km for the narrowest sections and with length of 70 km from the Bay  
60 of Biscay to the confluence of the Dordogne and Garonne rivers. Based on the bed composition, the  
61 estuary can be decomposed into 3 zones comprising (i) a sandy facies in the estuary mouth; (ii) a mixed  
62 facies dominated by mud (primarily composed of clays) along the central part, and (iii) a fluvial  
63 estuary, in the most upstream parts, characterized by the presence of sand, pebbles, and gravel  
64 (Mélières & Martin, 1969; Jalón-Rojas et al., 2015).

65 Fine suspended-sediment observed in the Gironde Estuary are mainly characterized (> 90%) by  
66 sediment diameters smaller than 16 µm (Jouanneau & Latouche, 1981). A large part of these fine  
67 sediment composes a pronounced Turbidity Maximum Zone (TMZ) with concentrations ranging  
68 between 1 and 10 g/LI-1 (Sottolichio & Castaing, 1999). The TMZ formation is mainly due to the tidal

69 asymmetry and partly to the vertical density gradient which maintains suspended sediment in the TMZ  
70 (Allen, 1972; Castaing, 1981; Sottolichio et al., 2001). Its location along the estuary migrates between  
71 Portets and off the estuary mouth depending on hydrological conditions (Castaing, 1981; Jalón-Rojas et  
72 al., 2015).

### 73 **3. Materials and methods**

#### 74 *1.1. Mathematical models*

75 Hydrodynamics processes are simulated using the continuity and momentum equations solve as the  
76 shallow water equations. This model includes parameterizations for the diffusion term and bottom  
77 friction. For further details, see Santoro et al. (2017). Bed composition is considered as a mixture of  
78 two sediment classes (sand and mud), characterized by their fraction and properties, such as the grain  
79 size, the settling velocity, and the concentration. The depth-averaged sediment concentration equation  
80 for suspended sediment transport is computed as follows:

$$81 \frac{\partial C}{\partial t} + U \frac{\partial C}{\partial x} + V \frac{\partial C}{\partial y} = \frac{1}{h} \left[ \frac{\partial}{\partial x} \left( h \varepsilon_x \frac{\partial C}{\partial x} \right) + \frac{\partial}{\partial y} \left( h \varepsilon_y \frac{\partial C}{\partial y} \right) \right] + \frac{E - D}{h} \quad (1)$$

82 where  $C$  is the depth-averaged concentration of the size class of sediment, in  $\text{kg/m}^3$ ,  $U$  and  $V$  are the  
83 depth-averaged velocity component along the  $x$  and  $y$  directions, respectively, in  $\text{m/s}$ ,  $h$  is the water  
84 depth in  $\text{m}$ ,  $\varepsilon_x$  and  $\varepsilon_y$  are the diffusion coefficients along the  $x$  and  $y$  directions, respectively, in  $\text{m}^2/\text{s}$ . In  
85 Eq. 1, the net sediment flux equals the summation of sediment erosion flux  $E$ ,  $\text{kg/s/m}^2$ , and sediment  
86 deposition flux  $D$ ,  $\text{kg/s/m}^2$ , computed according to the Parthenadies (1965) and the Krone (1962)  
87 formulations, respectively.

88 The bed is discretized with a fixed number of layers in which each layer thickness and sand and mud  
89 classes are initialized at time  $t = 0$ . The state of consolidation is reckoned through a linear relationship  
90 between the critical shear stress for erosion of mud ( $0.5 < \tau_{ce} < 1.5 \text{ N/m}^2$ ) and the mud concentration of

91 each layer ( $75 < C_{mud} < 500$  g/L). Erosion constants,  $E_0$ , are set constant for all layers. The critical shear  
 92 stress for erosion ( $\tau_{ce}$ ) and the erosion rate are computed according to the mud fraction ( $f_m$ ) (Waeles et  
 93 al., 2005). To consider the influence of mixed sediments on bed properties, the authors compute erosion  
 94 rates for mud ( $E_m$ ) and sand ( $E_s$ ) separately and depending to the regime type (cohesive, non-cohesive,  
 95 or mixed). For non-cohesive ( $f_m < 30\%$ ) and cohesive ( $f_m > 50\%$ ) regimes, different formulations for  
 96 the critical shear stress for erosion are used accounting for the sediment type (Eq. 2).

$$97 \quad E = f \cdot E_0 \cdot T^a \quad (2)$$

98 Where depending on the sediment type (mud or sand),  $E_s$  and  $E_m$  are the erosion rates of sand and mud,  
 99  $\text{kg/m}^2/\text{s}$ ,  $f$  is the proportion of sand or mud, between 0 and 1,  $E_{0s}$  and  $E_{0m}$  are the erosion constants of  
 100 sand or mud, respectively,  $\text{kg/m}^2/\text{s}$ ,  $T = (\tau_b - \tau_{ce})/\tau_{ce}$  with  $\tau_b$  the bottom shear stress, in  $\text{kg/m/s}^2$ ,  $\tau_{ce}$  the  
 101 critical shear stress for erosion, and  $a$  is a constant equal to 0.5. For the mixed regime ( $30\% < f_m <$   
 102  $50\%$ ), a weighted average is used to compute the erosion rates of sand (Eq. 3) and mud (Eq. 4):

$$103 \quad E_s = (1 - f_m) \left( E_{0s} + \frac{E_{0m} - E_{0s}}{f_{m,crit*} - f_{m,crit}} (f_m - f_{m,crit}) \right) T^{a + \frac{1-a}{f_{m,crit*} - f_{m,crit}} (f_m - f_{m,crit})} \quad (3)$$

$$104 \quad E_m = f_m \left( E_{0s} + \frac{E_{0m} - E_{0s}}{f_{m,crit*} - f_{m,crit}} (f_m - f_{m,crit}) \right) T^{a + \frac{1-a}{f_{m,crit*} - f_{m,crit}} (f_m - f_{m,crit})} \quad (4)$$

105 Deposition fluxes of sand and mud are computed by including the deposition probabilities for each  
 106 fraction in the Krone equation (Eq. 5). For both classes, the settling velocity  $W_s$  is set to a constant  
 107 equal to  $7.5 \cdot 10^{-4}$  m/s. Following Le Hir et al. (2001) and van Maanen and Sottolichio (2018), a high  
 108 critical shear velocity for deposition of mud is applied ( $u_{cd}^* = 10$  m/s). Sand and mud suspensions can,  
 109 thus, occur during the whole tidal cycle.

$$110 \quad D = \begin{cases} W_s C [(1 - \frac{u^*}{u_{cd}^*})^2] & \text{if } u^* < u_{cd}^* \\ D = 0 & \text{if } u^* > u_{cd}^* \end{cases} \quad (5)$$

111 Where  $u^*$  is the shear velocity, m/s. Consolidation processes are accounted for with an iso-  
112 concentration multi-layer consolidation model for mixed sediment (Villaret & Walther, 2008; Villaret et  
113 al., 2011) that assumes that the vertical flux of sediment between layers is proportional to the sediment  
114 mass in each layer, via the use of mass transfer coefficients. This approach was chosen to consider the  
115 sediment as a mixture in bed consolidation and differs from previous studies which include bed  
116 consolidation using the Gibson theory (Santoro *et al.*, 2017; Thiebot *et al.*, 2011; Van, 2012) with mud  
117 only. The bed is discretized with 20 layers characterized by a sediment concentration, a thickness, and a  
118 mass transfer coefficient set empirically to reproduce consolidation. During the simulation, only layer  
119 thicknesses vary and the mass balance computed for each layer is computed using Eq. 6.

$$120 \frac{M_i(t + \Delta t) - M_i(t)}{\Delta t} = F_i(t) - F_{i+1}(t) = -a_i M_i(t) \quad (6)$$

121 where  $i$  is the layer number,  $M_i$  is the mass per unit surface,  $\text{kg/m}^2$ ,  $F_i$  is the sediment flux at the lower  
122 boundary of the  $i$ -th layer,  $\text{kg/s.m}$ , and  $a_i$  is the mass transfer coefficient  $\text{s}^{-1}$ .

## 123 1.2. Model setup

124 Numerical computations are done using the open source Telemac-Mascaret system  
125 ([www.opentelemac.org](http://www.opentelemac.org)). Hydrodynamics are computed using the module TELEMAC-2D which solves  
126 the shallow water equations using the finite element method (Hervouet, 2007). This module is coupled  
127 at each time step with the module SISYPHE which solves the sediment transport and the bed evolution  
128 for both suspended and bedload transport processes. In the current study, only suspended sediment  
129 transport processes are considered. The domain is discretized with an unstructured triangular mesh of  
130 27,000 nodes that spreads over an area of 2,200  $\text{km}^2$ . The numerical domain extends over the entire  
131 estuary from the Bay of Biscay to the limit of the tide influence 170 km upstream from the estuary  
132 mouth. The study focuses on the intermediate estuary at the confluence of the Dordogne and Garonne  
133 rivers (Fig. 1). Inside this study area, cell lengths vary from 180 m and up to 300 m. Bathymetric data



134 from 2009-2016 collected by the national hydrographic service (SHOM) and the GPMB has been  
135 interpolated on the mesh grid. The computational time requires 2 h to simulate 50 d with a time step of  
136 10 s (using a work station with 12 cores of 2.4 GHz and 48 Go Ram).

137 Offshore boundary conditions are imposed from North East Atlantic (NEA) tidal atlases (Pairaud et al.,  
138 2008) including 46 harmonic constituents for better predictions of water levels (Huybrechts et al.,  
139 2012) and upstream boundary conditions are imposed from daily river discharges for both main  
140 tributaries. Mean sea level is obtained from the annual average of water levels collected at the Verdon  
141 station (Fig. 1) and defined at a value of 0.442 m. This latter value is then applied to the maritime  
142 boundary with also a constant salinity of 35 g/L and null values for the SPM. At upstream boundaries,  
143 fresh water is imposed while SPM values are taken from Jalón-Rojas et al. (2015).

144 The bottom friction is parameterized over 10 zones with a Strickler formulation, with roughness  
145 coefficients varying from 40 to 100  $m^{1/3}/s^{-1}$ , obtained from the calibration of water levels and currents  
146 (Huybrechts et al., 2012 and Ross et al., 2017). The lower value corresponds to sandy areas of the  
147 estuary mouth and is relatively close to the value predicted using the van Rijn formulation considering  
148 the bedform influence (van Rijn, 2007; Huybrechts et al., 2012). Higher values of the bed friction  
149 imposed in the intermediate and upper estuaries correspond to flat muddy beds.

150 One of the main issues to deal with for mixed sediment (sand-mud or graded sand) is to impose the  
151 initial distribution of the mixture over the computational domain and for the bed structure. Due to lack  
152 of measurements, the initial distribution is numerically determined assuming a sandy bed (mean  
153 diameter = 0.4 mm) at the estuary mouth as suggested by Van (2012). At the intermediate estuary, the  
154 mud fraction is predominant in the bed composition with the presence of sand in some locations  
155 (Villaret et al. 2010). The bathymetry is defined by the port around  $1200 \text{ kg/m}^3$  (McAnally et al., 2016,  
156 Xu & Yuan, 2007), corresponding to a concentration layer of  $\sim 330 \text{ g/L}$ . The fluid mud is initially

157 provided on the navigation channel as proposed by Sottolichio (1999). It is distributed along the first  
158 five layers (from 75 to 147 g/L) over an area of 40 km<sup>2</sup> and has a thickness of the order of 0.5 m. A 2.5  
159 m mud thickness is also distributed along layers from 237 to 500 g/L which corresponds to a more  
160 consolidated substrate below the bathymetry definition. Simulations are run without considering bed  
161 evolution, but the layers update in order to distribute the fluid mud along the estuary.

162 In order to simulate mud consolidation, a calibration procedure was applied to obtain the most  
163 representative temporal evolution of the sediment height and associated massive sediment  
164 concentrations (Fig. 2). Vertical profiles computed with the consolidation model are adjusted according  
165 to the values obtained from settling column experiments (Van, 2012). The maximum number of bed  
166 layers has been increased up to 20 layers in order to obtain a better discretization of the bed. Fair  
167 agreement between measured and computed values is obtained after the calibration of mass transfer  
168 coefficients for each layer. The values of the mass transfer coefficients and sediment concentrations  
169 vary from  $3 \times 10^{-3}$  to  $1 \times 10^{-6}$  s<sup>-1</sup> and from 75 to 507 g/L, respectively, over layers 1 to 20. These  
170 concentrations have been defined according to the experiments of Van (2012) and are in agreement  
171 with field observations in the Gironde Estuary (Abril et al., 2000) and other estuaries (McAnally et al.,  
172 2007a, b).

## 173 **2. Numerical model calibration and validation**

174 Numerical model calibration and validation are done by comparing measurements acquired during the  
175 project and simulated values for the following parameters : (i) water level ; (ii) current velocity; (iii)  
176 salinity; and (iv) suspended-sediment concentration.

177 *2.1. Water levels*

178 Huybrechts et al. (2012) calibrated the bed friction coefficient of the model in order to make the water  
179 level difference between measured and simulated lower than 10 cm at the mouth and the central part of  
180 the estuary. The calibration and validation have been done for flow currents and water levels obtained  
181 during field investigations in August 2006 and October-November 2009. These two events were  
182 characterized by low river discharges and calm weather conditions. The hydrodynamic model only  
183 included river and tidal forcings. Ten years later, the morphodynamics of the Gironde Estuary may  
184 have changed. More recent bathymetric surveys have been performed and can be now considered in the  
185 model. It is, thus, necessary to check the validity of the calibration, then to analyze how the model  
186 behaves for different hydrological conditions in the upper estuary where the port terminals are located.  
187 The calibration update and validation of the hydrodynamics are realized during low ( $< 200 \text{ m}^3/\text{s}$ ) and  
188 high river discharges (between 1000 and 2000  $\text{m}^3/\text{s}$ ) for neap and spring tide conditions.

189 During 15 days of low flow conditions (between 08/26 and 09/06/2018), the total river discharge varied  
190 from 161 to 235  $\text{m}^3/\text{s}$ . Comparison between measured and computed water levels shows the ability  
191 of the model to predict the tidal amplitude, as well as, the tidal asymmetry with Root Mean Squared  
192 Error (RMSE) values up to 18 cm at the Bordeaux station (Fig. 3). Better results are found at the  
193 Verdon and Pauillac stations, in the lower and intermediate parts of the estuary (Fig. 1), with RMSEs  
194 below 10 cm. Maximum errors are mainly observed during high and low water conditions at the  
195 Bordeaux station in the upper estuary. Regression analysis shows a very good fit for every station with  
196 coefficients of determination,  $r^2$ , of 0.99 confirming the robustness of the model to predict water levels.  
197 However, these accurate predictions of the water level obtained during low river discharges could be  
198 degraded during flood periods. Indeed, seasonal variation of river discharges induces the migration of  
199 the turbidity maximum (Sottolichio, 1999) which can modify the estuarine bed texture and may have a  
200 significant impact on the bed friction (Jalón-Rojas et al., 2018).

201 Differences observed between measured and simulated water levels can come from two sources: a  
202 calibration using constant bed friction coefficients with hydrological conditions and/or variation of  
203 mean sea level due to tide-surge interactions. To distinguish the influence of the friction calibration and  
204 the impact of mean sea level on numerical results, harmonic analysis (Pawlowicz *et al.*, 2002) is also  
205 performed on a longer time series at Verdon, Pauillac, and Bordeaux stations (Fig. 4). The lack of data  
206 for the year 2018 resulted in the harmonic analysis being done with water levels measured during 2015  
207 over a period of 3 months. As expected, the analysis revealed the predominance of the M2 harmonic  
208 constituent (principal lunar semidiurnal) in the composition of the tidal amplitude for the predicted and  
209 measured values with values varying around 1.55 m (Fig. 4a). The N2 (larger lunar elliptic  
210 semidiurnal) and S2 (principal solar semidiurnal) constituent harmonics complete 90% of the signal  
211 with amplitudes around 0.3 m and 0.5 m, respectively. Moreover, the amplification of shallow water  
212 overtides M4 and M6 is noticed at upstream stations (Fig. 4b-c). Differences between measured and  
213 predicted values of amplitude are small and never exceed 7 cm for all harmonic constituents and for all  
214 stations. For semi-diurnal constituents (M2, N2, and S2) at the Verdon station, amplitude errors are  
215 below 5 cm with a maximum value for the M2 harmonic (4.2 cm).

216 Phase lags are also well reproduced with computed delays of 4 min and of less than a minute for M2  
217 and S2 harmonics, respectively (Fig. 4d-e-f). For M4 and M6 harmonics delays increase up to 16 min.  
218 For overall harmonics, phase lags do not show significant variations during the tidal propagation. In  
219 summary, the harmonic analysis points out that differences in amplitude of the main tidal components  
220 are small. For two-dimensional (2D) modeling, this analysis supports the hypothesis of a seasonal  
221 calibration of the model. Trial and error methodology as done by Huybrechts *et al* (2012) is not suitable  
222 to do regular updates of the bed friction calibration. Therefore, the application of automatic calibration  
223 and optimization procedures is more suitable (Smaoui *et al* 2018).

## 224 2.2. *Current velocities*

225 As also observed for water levels, the accuracy of current velocity predictions decreases in the  
226 upstream direction. According to the Fig. 5a-b, results are good enough at Stations P1 and P4 with  $r^2$  of  
227 0.79 and 0.90, respectively. However, the robustness of the model decreases at Station P5 where  
228 simulations overestimate the strength of ebb currents, particularly during spring tides. For the latter  
229 conditions, the measurements describe a well-marked tidal asymmetry characterized by short flood  
230 tides and long ebb tides. However, at Station P5 where the measured asymmetry is the strongest, mean  
231 velocities of flood and ebb currents are equivalent and equal to 1 and 0.98 m/s, respectively (Fig. 5c).  
232 For the same station, the RMSE reaches the maximum value of 0.14 m/s which is not so high compared  
233 to other stations where the minimum value is 0.09 m/s. The best results are obtained at Station P1  
234 where ebb currents are accurately predicted. For other stations, ebb currents are difficult to predict with  
235 an underestimation of the peak ebb velocity and an overall overestimation of velocities during the fall  
236 of the tide (Fig. 5b-c).

## 237 2.3. *Salinity*

238 Validation of salinity predictions used data collected near the bed during mooring surveys and for  
239 different hydrological conditions (see Section 3.3). The simulation started during neap tides  
240 characterized by low tidal amplitudes of measured salinity (below ~10 psu) for all stations (Fig. 6c-d-  
241 e). Conversely, maximum tidal amplitudes are observed during spring tides between days 9 and 13 with  
242 a maximum value of 17 psu. However, for the last half of the simulation, the overall salinity tends to  
243 decrease in response to a steep increase of the total river discharge of 821 m<sup>3</sup>/s in 6 d (Fig. 6b-c-d).  
244 Despite these changing conditions, the model correctly predicts tidal and fortnightly variations, but  
245 tends to underestimate the amplitude of tidal variations, particularly near the estuary mouth at station  
246 P1 (Fig. 6c). These differences ranging between 3.12 and 6.7 psu are mainly due to the fact that

247 measurements are taken near the bed, while numerical values of salinity correspond to depth-averaged  
248 quantities.

249 A few kilometers upstream, at Station P2, the amplitude is better reproduced, particularly between days  
250 8 and 12 (Fig. 6d). However, for some tidal cycles, measured salinity shows rapid changes not captured  
251 by the numerical model. This observation could be explained by the presence of a strong vertical  
252 density gradient in this area which cannot be reproduced by 2DH model. In the intermediate estuary, at  
253 the Station P4, measured salinity decreases rapidly below 5 psu indicating that the salt intrusion limit is  
254 close under this hydrological condition (Fig. 6e). In this area, where salinity values are generally low  
255 but could increase considerably near the bed, the 2DH model is less accurate but still provides reliable  
256 predictions.

257 For the low river discharges period, the highest flowrates are observed between days 29 and 32 with a  
258 maximum value of  $235 \text{ m}^3/\text{s}$  (Fig. 7b). In this hydrological condition, density stratification is weak and  
259 allows accurate predictions, particularly at Station P4 with  $r^2 = 0.84$  (Fig. 7d). At Station P1, the model  
260 correctly predicts tidal variations of salinity but slightly overestimates salinity during slack waters with  
261 a difference up to 2.7 psu for both high and low water slacks, respectively (Fig. 7c). As observed for  
262 the flood period, surrounding areas of the salt intrusion limit, where salinity is low, are difficult to  
263 predict. At Station P5, conversely to the flood period, the model overestimates salinity, but  
264 measurements seem to be wrong and describe a linear signal since day 30 owing to the malfunctioning  
265 of the probe (Fig. 7e).

#### 266 2.4. *Suspended-sediment*

267 In order to compare predicted and measured values, a correction factor estimated using a simplified  
268 Rouse profile has been applied. The Rouse profile was set using a linear increase of the vertical  
269 diffusivity and a reference altitude  $z_a$  of 1 m. During the flood period, only two stations were available

270 due to the loss of a probe and the saturation of a sensor at the other one. Hydrological conditions are  
271 characterized by higher river discharge periods between days 26-28 and 39-40 (Fig. 8b). However, no  
272 clear relation appears with SPM due to the short duration of the simulation.

273 For both Stations P1 and P2, measured and simulated SPM are strongly related to the fortnightly cycle  
274 with maximum values observed during spring tides between days 30 and 34 ( $> 3$  g/L for  
275 measurements) and null values during neap tides at the end of the simulation (Fig. 8c-d). The dynamics  
276 of SPM is reliably predicted by the model excepted at the tidal scale. At this temporal scale, differences  
277 between ebb and flood maximum resuspension are not substantial, particularly at Station P1 (Fig. 8c).  
278 Although the model tends to underestimate SPM values for both stations, it provides reliable  
279 predictions for high river discharges.

280 During the simulation of low river discharges, flowrates are below  $300 \text{ m}^3/\text{s}$  and spring and neap tides  
281 occurred during days 26-29 and 32-35, respectively. For these hydrological conditions, the TMZ  
282 migrates landward near Bordeaux inducing a decrease of SPM in the lower estuary, as observed at  
283 Station P2 (Fig. 9c). At this station, the model also accurately predicts the fortnightly dynamics,  
284 particularly during neap tides. As observed with high river discharges, tidal variations are still difficult  
285 to predict with no clear delineation between flood and ebb resuspensions. For Stations P4 and P5, SPM  
286 levels are clearly underestimated by the model by approximately 29 and 125%, respectively, and these  
287 underestimations could be explained by a combination of several factors (Fig. 9d-e). First, the model  
288 experiences some difficulties to keep suspended particles in the numerical domain over time due to  
289 deposition in shallow areas. Second, the numerical reproduction of the TMZ with a 2DH model does  
290 not consider the influence of vertical density gradients which act to maintain high suspended-sediment  
291 concentrations above the bed.

### 292 3. Sensitivity analysis

293 A sensitivity analysis is performed to study the influence of several physical parameters such as the  
294 critical shear stress for erosion, the settling velocity, the consolidation on depth-averaged SPM (Fig.  
295 10) and fluid mud thickness (Fig. 11). The latter parameter corresponds to the sum of the first five  
296 layers (< 150 g/L). Results are extracted from the intermediate estuary near the Pauillac station (Fig. 1).  
297 For the initial configuration, constant values are chosen for the settling velocity ( $W_s = 0.75$  mm/s), the  
298 critical shear velocity for deposition ( $u_{cd}^* = 10$  m/s) and the Partheniades constant ( $M = 2 \times 10^{-3}$   
299 kg/m<sup>2</sup>/s). For the erosion law, a linear relation between the critical shear stress and the mud layer  
300 concentration is considered (Sect. 3.1). This choice of relation is similar to those selected by van  
301 Maanen and Sottolichio (2018).

302 The first test compares the influence of erosion parameters: the critical shear stress for erosion and the  
303 Parthenadies constant (Fig. 10a). Reducing of 20% the critical shear for erosion yields an increase in  
304 SPM of 73% and amplifies tidal variations. Conversely, halving the Parthenadies constant induces an  
305 overall decrease of SPM of 53% and expands the availability of sediment for consolidation increasing  
306 the thickness of the fluid mud layer (Fig. 11a).

307 Similar trends are observed with bed parameters including one simulation without bed consolidation  
308 and one simulation with the presence of sand in the intermediate estuary (Fig. 10c). As expected,  
309 without bed consolidation SPM concentrations increase and the fluid mud thickness becomes thicker  
310 (Fig. 11d). During the first month of the simulation, the fluid mud layer firstly becomes thinner until  
311 the mud deposit is completely resuspended and reaches an equilibrium on day 30 (Fig. 11d). However,  
312 for simulations covering a period of time of few months, fluid mud deposits in shallow areas are  
313 overestimated and reached unrealistic values (> 5 m). For short-term simulations (up to 36 h),



314 consolidation processes can be neglected. Figure 10c also shows the strong influence of the assumption  
315 of the sand content. The addition of sand (10% per layer) in the intermediate estuary increases the  
316 critical shear stress for erosion, and, therefore, reduces resuspension. Moreover due to consolidation,  
317 the fluid mud is transferred to more consolidated layers which tends to increase the sand influence on  
318 the first layers. However, imposing 10% of sand in the intermediate estuary and in the entire mud  
319 substrate does not seem realistic. Sand is probably located around some banks or between muddy layers  
320 which may be exposed to sediment transport according to the turbidity maximum migration or dredging  
321 activities. This result points out the importance of monitoring spatial and temporal variations of the bed  
322 material composition in order to provide reliable inputs for the model.

323 The last comparisons check the influence of two deposition parameters: the critical shear velocity for  
324 deposition and the settling velocity. Reducing by a factor of two these parameters leads to higher  
325 suspended-sediment concentrations up to 4.8 g/L-1 and higher tidal variations (Fig. 10b). However, for  
326 the settling velocity no significant differences are noticed in fluid mud deposits since the deposition  
327 flux is proportional to the concentration of suspended sediment and the settling velocity (Fig. 11b). For  
328 the critical shear velocity for deposition, the fluid mud thickness reaches values up to 7-8 cm whereas it  
329 reaches 1-2 cm with the baseline simulation (Fig. 11c). A more energetic exchange, thus, is observed  
330 between the mud layer and the water column.

331 This parametric study illustrates the sensitivity of the model to the values of sediment parameters.  
332 These values are mostly imposed as constants over the computational domain. For the settling velocity  
333 it could be interesting to add spatial and temporal variability considering the influence of the  
334 suspended-sediment concentration. Similar improvement can be made for the critical shear stress for  
335 deposition according to the transport capacity, as suggested by Bi and Toorman (2015).

#### 336 **4. Conclusions**

337 The purpose of this paper is to develop a mixed sediment transport model for the prediction of the  
338 under keel clearance in the Gironde Estuary. The current study allows the hydrodynamics and sediment  
339 transport to be set up and validated with good accuracy. The harmonic analysis of the astronomical  
340 tides reveals (i) a strong distortion of the tidal wave inducing the growth of overtides constituents and  
341 (ii) the non-significant effect of tide-surge interactions for annual-scale predictions. Transport for sand-  
342 mud mixtures is considered for erosion, deposition, and consolidation processes with two sediment  
343 classes (sand and mud) and as a function of the mud fraction. Overall simulations were done with a  
344 fixed bed to validate suspended-sediment dynamics. The model was firstly validated for  
345 hydrodynamics by comparison with measured water levels and current velocities with high coefficients  
346 of determination. Predictions of salinity values also are reliable but show some deviations when the  
347 river discharge increases abruptly. Strong vertical density gradients also can explain observed  
348 differences. Suspended-sediment concentrations are correctly predicted during flood conditions, but  
349 with a less-marked tidal dynamic. When river discharges are low, the model underestimates SPM levels  
350 probably due to the difficulty to maintain the TMZ without vertical density gradients. The sensitivity  
351 analysis has revealed a strong influence of (i) the settling velocity (ii) the critical shear stress for  
352 deposition, and (iii) the chosen erosion parameter for the computation of suspended-sediment  
353 concentrations and mud layer thicknesses in the intermediate estuary. This model will be used in the  
354 near future to compute the bed morphological changes and analyze the influence of bed load and  
355 suspended load transport and dredging/dumping operations in the navigation channel.

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**Fig. 1.** Location map of the Gironde Estuary (France) and its main tributaries the Dordogne and the Garonne rivers. Measurement stations and tidal gages are represented along the estuary by black squares and black triangles, respectively. The bathymetric chart is computed with the inverse distance weight method and used data from surveys collected at different times from 2005 to 2018.

**Fig. 2.** Temporal evolution of the relative sediment height  $z/h$  (where  $z$  is the elevation above the channel bed) for vertical profiles of sediment mass concentration obtained from settling column experiments conducted by Van (2012) (markers) and the semi-empirical multi-layer model (line and markers).

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**Fig. 7.** Time series of (a) water levels at the Pauillac station and (b) discharges of the Garonne and Dordogne rivers. Time series of measured (grey points) and simulated (black line) depth-averaged salinity are shown for stations (c) P1, (d) P4, and (e) P5. The simulation is realized during a low river discharge period (from August to September 2018).

**Fig. 8.** Time series of (a) water levels at the Pauillac station and (b) discharges of the Garonne and Dordogne rivers. Time series of measured (grey points) and simulated (black line) depth-averaged SPM are shown for stations (c) P1, and (d) P2. The simulation is done during a high river discharge period (from March to April 2018).

**Fig. 9.** Time series of (a) water levels at the Pauillac station and (b) discharges of the Garonne and Dordogne rivers. Time series of measured (grey points) and simulated (black line) depth-averaged SPM are shown for stations (c) P1, (d) P4, and (e) P5. The simulation is realized during a low river discharge period (from August to September 2018).

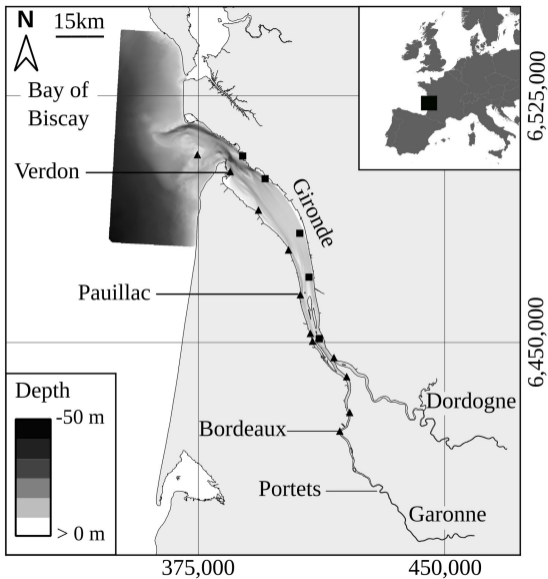
**Fig. 10.** Time series of depth-averaged suspended particulate matter (SPM) computed by the numerical model with different settings for (a) the critical shear stress for erosion ( $\tau_{ce}$ ) and the Parthenadies

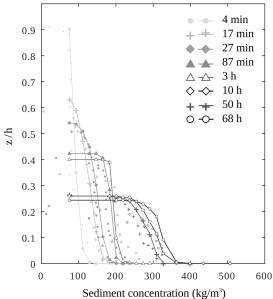
constant ( $M$ ), (b) the critical shear velocity for deposition ( $U_{cd}^*$ ) and the settling velocity ( $W_s$ ) and (c) without consolidation and depending on the sand fraction. Numerical results were extracted near the Pauillac station (Fig. 1) for a period of 60 days.

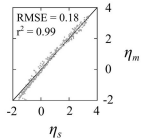
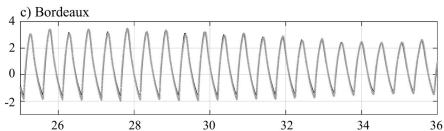
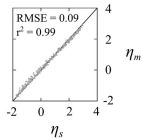
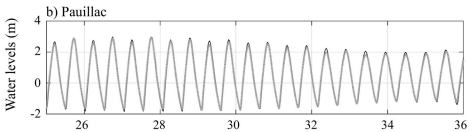
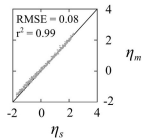
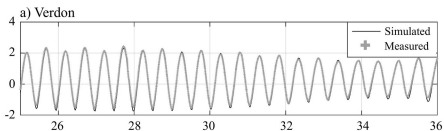
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Elapsed time since 08/01/2018 (days)

