

Development and validation of a three-dimensional morphodynamic modelling system

Accepted in Ocean Modelling, vol. 58-59, 2012

1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
19
20
21
22
23
24
25
26

L. Pinto¹

National Civil Engineering Laboratory, Estuaries and Coastal Zones Division,
Av. do Brasil 101, 1700-066 Lisbon, Portugal
Phone: +351-21-8443613; fax: +351-21-8443016
E-mail: lpinto@lnec.pt

A.B. Fortunato

National Civil Engineering Laboratory, Estuaries and Coastal Zones Division,
Av. do Brasil 101, 1700-066 Lisbon, Portugal
E-mail: afortunato@lnec.pt

Y. Zhang

Center for Coastal Margin Observation and Prediction, Oregon Health and
Science University, 20000 NW Walker Road, Beaverton, OR 97006, USA
E-mail:

A. Oliveira

National Civil Engineering Laboratory, Estuaries and Coastal Zones Division,
Av. do Brasil 101, 1700-066 Lisbon, Portugal
E-mail: aoliveira@lnec.pt

F.E.P. Sancho

National Civil Engineering Laboratory, Estuaries and Coastal Zones Division,
Av. do Brasil 101, 1700-066 Lisbon, Portugal
E-mail: fsancho@lnec.pt

Abstract

The morphology of estuaries is subject to constant changes as a result of the
dynamic imbalance between the forcing actions (tides, surface waves and wind) and
the sedimentary reactions, which are important to understand and predict.
Morphodynamic numerical models are useful tools that can be used to
understanding and predicting morphological developments. The development and

¹ Corresponding author

validation of a three-dimensional morphodynamic model, MORSELF, aiming at simulating rapid morphological evolutions of estuaries (temporal scale of days to months), are presented in this paper. MORSELF results from coupling a three dimensional shallow water model, with an advection-diffusion transport model for the suspended sediments, an empirical formula for the bed load, and a bed updating module. The model only considers the simulation of non-cohesive sediment and do not account with the wave effects. An innovative feature of the model is a new approach to compute the erosive fluxes, which adapts them to the vertical grid resolution. The use of unstructured grids and the implementation in parallel mode make MORSELF particularly adapted to engineering applications. The model was assessed and validated against analytical and experimental test cases, also allowing the inference on the optimum choice of the model parameters, variables and the turbulent closure.

Keywords: Morphodynamic model; Sediment transport; Three dimensional model.

1 Introduction

Morphodynamic models are essential tools to predict the morphological evolution of the coastal, river and estuarine systems. Over the last decades, several morphodynamic process based models have been developed over the world (e.g. de Katopodi e Ribberink, 1992; Nicholson et al., 1997; Cayocca, 2001; Fortunato and Oliveira, 2004; Lesser et al., 2004, Warner et al., 2008; Bertin et al., 2009) with the aim of simulating the processes involve and their interactions, and to predict the morphological evolution of the estuarine and coastal areas. During this period, 2D morphodynamic processed based models have reached a satisfying degree of reliability. Several examples of applications can be found in literature. However, in situations involving complex flows, such as wind and wave induced circulation, density gradients and flows in the adjacent of structures, the deviations of the velocity profile from a logarithmic one disables the use of 2DH models. For t and a 3D model is required.

More recently, 3D morphodynamic process based models were developed (Lesser et al., 2004; Warner et al., 2008) in order to limited ability of the 2D morphodynamic models of simulating the several important 3D flow phenomena

found in nature (). with the objective of improving the representation of depth-varying mechanisms, such as wave induced cross-shore currents or baroclinic circulation..

This paper presents the development and validation of the unstructured grid three dimensional morphodynamic processed based model MORSELF. This modelling system consists in a number of integrated modules which allows the simulation of hydrodynamic flow, non-cohesive sediment transport and morphological changes in estuarine systems, driven by tides, wind and river flows. A detailed description of MORSELF modelling system, with the physical and numerical formulation of the modules, is performed in section 2. Section 3 describes the results of the model validation against analytical and experimental test cases. Section 4 summarizes the most important changes introduced in MORSELF and the main findings of the validation tests.

2 Three Dimensional morphodynamic Model MORSELF

MORSELF is a modeling system consisting in a set of integrated modules that aims to simulate shallow water flows, non-cohesive sediment transport and bottom changes in estuaries (Figure 1). The development of this modeling system was based on existing models, although several changes were introduced to the original models in order to make possible the models connection and integration. The model can be executed in serial mode or on multiple processors using distributed-memory architectures.

For the Hydrodynamics, MORSELF uses the three dimensional (3D) model SELF (Semi-implicit Eulerian Lagrangian Finite Element), an unstructured-grid model designed for the effective simulation of tridimensional baroclinic circulation across river-to-ocean scales (Zhang and Baptista, 2008). The sediment transport equation is treated in two distinct modes: the advection-diffusion component is based on SELF transport formulation, taking advantage of its numerical characteristics such as the option to use different methods to solve the transport equation, while the remain terms are computed based on the ROMS (Regional Oceanographic Modeling System) model sediment transport formulation (Warner *et al.*, 2008). For the bed load transport prediction the model uses empirical formulations (Meyer-Peter and Müller, 1948; van Rijn, 2007). Sediment transport algorithms are implemented for an unlimited number of non-cohesive sediment classes, specified by the users. For the bed elevation changes computation two

different algorithms were combined: the ROMS module was adapted to compute the changes due to suspended sediment transport and the SAND2D (Fortunato and Oliveira, 2004) bottom update model was modified to calculate the changes due to bed load transport.

The MORSELF code is composed by a main code (the Hydrodynamic model code) and several subroutines that allow simulating the sediment transport and bed evolution. The modules coupling through subroutines allow a greater independence of the several modules, making easy the replacement of the codes if a new model version is released. The integration of the several modules was performed via C-preprocessor directives, allowing the user to choose the different components of the model that are included or excluded in the model compilation and execution. This guarantees that only the necessary memory is allocated and relevant algorithms are compiled, generating a more efficient executable file.

2.1 Hydrodynamic module

SELF is a 3D hydrodynamic model developed at the Center for Coastal Margin Observation and Prediction as an open source community model (Zhang and Baptista, 2008, available at <http://www.stccmop.org/CORIE/modeling/selfe/>). SELF uses a semi-implicit finite-element Eulerian-Lagrangian algorithm to solve the Navier-Stokes equations using the hydrostatic and Boussinesq approximations, and transport equations for salt and heat. The primary variables that SELF solves are free-surface elevation, 3D water velocity, 3D temperature and 3D salinity. Forcings include tides, tidal potential, river flow, wind and solar radiation. As turbulence closure model, SELF uses the Generic Length Scale (GLS) turbulence closure of Umlauf and Burchard (2003), which has the advantage of encompassing most of the 2.5-equation closure models.

SELF equations are solved with finite-element and finite-volume techniques. Semi-implicit schemes are applied to all equations to improve stability and maximize the efficiency. The key element in SELF is to decouple continuity and momentum equation via the bottom boundary layer. To treat the advection terms in the momentum equation, SELF uses an Eulerian-Lagrangian method (ELM), allowing the use of high time steps without compromising the precision and the stability.

SELFE treats the advection terms in the transport equations with an ELM, a finite-volume upwind method (FVUM) or a *Total Variation Diminishing* (TVD) scheme. If the ELM is used, the transport equations are solved using a finite-element method. Despite its efficiency, one of the problems of using ELM approach is the potential for mass conservation problems (Oliveira *et al.*, 2000). On the other hand FVUM guarantees mass conservation but are excessively diffuse. To reduce the numerical diffusion a higher-order finite-volume TVD scheme (Sweby, 1984) was implemented in SELFE. The vertical advective fluxes are treated implicitly or explicitly if a FVUM or a TVD is used, respectively.

The horizontal domain is discretized with unstructured triangular grids, while hybrid vertical coordinates (partly terrain-following S coordinates and partly Z coordinates) are used in the vertical direction, allowing higher flexibility in both horizontal and vertical grids. S coordinates are similar to sigma coordinates but with an additional flexibility, the vertical resolution can be adjusted to allow a better resolution near the bottom and surface (Song and Haidvogel, 1994).

The combined horizontal and vertical discretizations result in the whole 3D domain being divided into triangular prisms (Figure 2). A staggering scheme is used to define variables: the surface elevations (η) are defined at element nodes; the horizontal velocities (u, v) are defined at side centers and whole levels; the vertical velocities (w) are defined at the element centers and whole levels as they are solved with a finite-volume method. The locations where the temperatures and salinities are defined depend on the method used to solve the transport equation: they are defined at prism centers if FVUM or TVD ($T_{Up, TVD}$) are used, and at both nodes and side centers if ELM is used (T_{ELM}).

The turbulent closure equations are solved along each vertical column at each node with a finite-element method. The vertical mixing terms and the dissipation term in these equations are treated implicitly, but the production and buoyancy are treated implicitly or explicitly, depending on the sign of their total contribution (Zhang *et al.*, 2004). The advection terms in the turbulence closure equations are neglected in SELFE since they are small when compared to the other terms.

2.2 Sediment transport module

The sediment transport model developed is a three-dimensional model, whose suspension component results from the coupling between the SELFE transport formulation and the ROMS sediment transport formulation (version 2.2). For the bed load prediction two formulations are implemented: Meyer-Peter and Müller (1948) and van Rijn (2007). This model is able to represent an unlimited number of user defined non-cohesive sediment classes. For each sediment class the model solves the horizontal and vertical advection, the vertical diffusion, the bed load transport, the vertical settling velocity and bottom sediment deposition or erosion.

2.2.1 Suspended sediment transport

2.2.1.1 Physical formulation

Sediment suspended in water column is transported by solving the advection-diffusion equation with two additional terms for the vertical settling and for the sediment bed exchange. For a generic tracer (T_r) the transport equation is given by:

$$\frac{\partial T_r}{\partial t} + u \frac{\partial T_r}{\partial x} + v \frac{\partial T_r}{\partial y} + w \frac{\partial T_r}{\partial z} = \frac{\partial}{\partial z} \left(\kappa \frac{\partial T_r}{\partial z} \right) + F_h + \hat{Q}_q \quad (1)$$

where (u, v, w) is the velocity, κ is the eddy diffusivity (m^2s^{-1}), F_h is the horizontal diffusion, $\hat{Q}$ represents point sources or sinks. In the case of sediment, $T_r = C$, where C is the sediment concentration (kgm^{-3}). Eq. (1) is solved for each sediment class q and the term $\hat{Q}_q$ represents the settling velocity and the sediment bed exchange terms given by:

$$\hat{Q}_q = -w_{s,q} \frac{\partial C_q}{\partial z} + E_q \quad (2)$$

where C_q is the sediment concentration of the q class, $w_{s,q}$ is the sediment settling velocity of the q class (ms^{-1}) (positive upwards) and E_q is the sediment erosion flux of q class ($\text{kgm}^{-2}\text{s}^{-1}$).

183 Sediment deposition and erosion

184 For each sediment class q , the sediment settling velocity is determined
185 following Soulsby (1997):

$$w_{s,q} = \frac{\nu_a}{d_{50,q}} \left[\left(10,36^2 + 1,049 D_{*,q}^3 \right)^{1/2} - 10,36 \right] \quad (3)$$

186 where ν_a is the kinematic viscosity of water (m^2s^{-1}) and $d_{50,q}$ is the median grain
187 sediment diameter of class q (m), $D_{*,q}$ is the dimensionless sediment diameter of
188 class q given by:

$$D_{*,q} = \left[\frac{g(s-1)}{\nu_a^2} \right]^{1/3} d_{50,q} \quad (4)$$

189 where g is the acceleration due to gravity, $s = \rho_{s,q} / \rho_0$ is the specific density, $\rho_{s,q}$
190 and ρ_0 are sediment and water reference densities (kg/m^3), respectively.

191 Erosional flux ($\text{kgm}^{-2}\text{s}^{-1}$) is parameterized following Ariathurai and
192 Arulanandan (1978):

$$E_q = E_{0,q} (1-p) f_q \left(\frac{\tau_{sf}}{\tau_{cr,q}} - 1 \right) \quad \text{se} \quad \tau_{sf} > \tau_{cr,q} \quad (5)$$

$$\tau_{cr,q} = \theta_{cr,q} g d_{50,q} (\rho_s - \rho_0) \quad (6)$$

193 where $E_{0,q}$ is a bed erodibility constant ($\text{kgm}^{-2}\text{s}^{-1}$), p is the sediment porosity of the
194 top bed layer (dimensional), f_q is the volumetric fraction of sediment of class q
195 (dimensional), $\tau_{cr,q}$ is the critical shear stress for class q (Nm^{-2}), $\tau_{sf} = \sqrt{\tau_{bx}^2 + \tau_{by}^2}$ is
196 the bed shear stress (Nm^{-2}). $E_{0,q}$ is an empirical quantity, depending on local
197 sediment and bed conditions, with values ranging from 10^{-4} and $10^{-2} \text{ kgm}^{-2}\text{s}^{-1}$ (Xu et
198 al., 2002; Blaas et al., 2007).

199 The dimensional critical shear stress ($\theta_{cr,q}$) is derived from the critical Shields
200 parameter, and its value is determined following Soulsby and Whitehouse (1997):

$$\theta_{cr,q} = \frac{0,3}{1 + 1,2D_{*,q}} + 0,055 \left[1 - e^{(-0,022D_{*,q})} \right] \quad (7)$$

201 Density effects

202 The equation of state used to calculate the density of water as temperature
203 and salinity function is extended to include suspended sediment density effect as
204 follows:

$$\rho = \rho_w + \sum_{q=1}^{Nsed} \frac{C_q}{\rho_{s,q}} (\rho_{s,q} - \rho_w) \quad (8)$$

205 in which ρ is the fluid local density (including the salinity, temperature and sediment
206 effects) (kg/m^3), ρ_w is the water density (including salinity and temperature effects)
207 (kg/m^3), $Nsed$ is the total number of sediment classes.

208 Bottom stress

209 The bottom boundary layer treatment is very important for the sediment
210 transport formulation, since bottom stress determines the bed load transport rate and
211 the sediment resuspension rate to the water column.

212 Bottom friction can be described by linear, quadratic or logarithmic profile
213 formulae. The linear and quadratic drag coefficient methods depend on the
214 horizontal velocity components in the bottom grid cell and the linear (γ_1) and
215 quadratic (γ_2) drag coefficients specified by the user:

$$(\tau_{bx}, \tau_{by}) = \left(\gamma_1 + \gamma_2 \sqrt{u^2 + v^2} \right) (u, v) \quad (9)$$

216 The user can select between linear or quadratic drag by setting γ_1 ou γ_2 to
217 zero, respectively. For the logarithmic formulation, the classic vertical logarithmic
218 profile, defined by a shear velocity u_* and bottom roughness length z_0 , is assumed in
219 the bottom boundary layer:

$$|u| = \frac{u_*}{\kappa_0} \ln \left(\frac{\delta_b}{z_0} \right) \quad (10)$$

where $|u| = \sqrt{u^2 + v^2}$, $u_* = \sqrt{(|\tau_{bx}| + |\tau_{by}|)} / \rho$, $\kappa_0 = 0,4$ is the von Kármán constant, z_0 is the bottom roughness length (m) and δ_b is the bottom computational layer thickness (m). In this formulation, a quadratic drag formulation is used, with $\gamma_2 = (\kappa_0 / \ln(\delta_b / z_0))^2$, and the z_0 value is specified by the user.

2.2.1.2 Numerical formulation

Advection and diffusion

The advection-diffusion component of MORSELF model is based on SELF model transport formulation. The treatment of the advective fluxes depend on the method used to solve the transport equation: for FVUM the vertical advective fluxes are treated implicitly and the horizontal advective fluxes are treated explicitly; for TVD an explicit scheme is used in the advection treatment, due to the non linearity of this method. In both methods, the vertical diffusion term is treated implicitly and the tracer value (called T_{Up} and T_{TVD} in Figure 2) is calculated in the prism center (i,k) . Detailed description of the SELF transport equation numerical formulation is presented in the next lines.

The discretized transport equation for the sediment concentration is given by:

$$C_{i,k}^{n+1} = C_{i,k}^n - \frac{\Delta t'}{V_{i,k}^n} \sum_{j \in S} Q_j C_{j*}^{n+1} + \hat{Q}_{i,k}^{n+1} \Delta t' + \Delta t' \sum_{j=1}^3 (\kappa_h) \frac{C_j^n - C_i^n}{\delta_{ij}} + \frac{A_i \Delta t'}{V_{i,k}^n} \left[\kappa_{i,k} \frac{C_{i,k+1}^{n+1} - C_{i,k}^{n+1}}{\Delta z_{i,k+1/2}} - \kappa_{i,k-1} \frac{C_{i,k}^{n+1} - C_{i,k-1}^{n+1}}{\Delta z_{i,k-1/2}} \right], (k = k_b + 1, \dots, N_z) \quad (11)$$

where j are the five faces of the prism (i,k) , $V_{i,k}$ is the volume of the prism (i,k) , A_i is the area of element i , u_j is the outward normal velocity from prism (i,k) , S_j is the area of face j , $Q_j = u_j S_j$ is the flux through the same face and $\Delta t' \neq \Delta t$ is the time step. The second right hand side term of the Eq. (11) is the advective term, the third term is the sink or source term, the fourth term the horizontal diffusion term and the last terms

241 represents the vertical diffusion. δ_{ij} is the distance between the centers of the prism
242 (i,k) and its adjacent prism (j,k).

243 The only difference between the upwind and the TVD scheme is the definition
244 of the C_{j^*} in the Eq. (11). In the case of a FVUM scheme C_{j^*} is defined as:

$$C_{j^*} = \hat{C}_{j^*} = \begin{cases} C_j, & j \in S^+ \\ C_j, & j \in S^- \end{cases} \quad (12)$$

245 where S^+ are all prism faces where $Q_j > 0$ (outflow faces) and S^- are all prism faces
246 where $Q_j \leq 0$ (inflow faces).

247 Neglecting the non advective terms of Eq. (11) and substituting C_{j^*} given by
248 equation (12), stay:

$$C_{i,k}^{n+1} = C_{i,k}^n \left[1 - \frac{\Delta t'}{V_{i,k}^n} \sum_{j \in S^+} |Q_j| \right] + \frac{\Delta t'}{V_{i,k}^n} \sum_{j \in S^-} |Q_j| C_j^n \quad (13)$$

249 The final equation for the FVUM scheme is obtained substituting the Eq. (13)
250 in the Eq. (11):

$$\begin{aligned} C_{i,k}^{n+1} = & C_{i,k}^n \left[1 - \frac{\Delta t'}{V_{i,k}^n} \sum_{j \in S^+} |Q_j| \right] + \frac{\Delta t'}{V_{i,k}^n} \sum_{j \in S^-} |Q_j| C_j^n + \hat{Q}_{i,k}^{n+1} \Delta t' + \Delta t' \sum_{j=1}^3 (\kappa_h) \frac{C_j^n - C_i^n}{\delta_{i,j}} + \\ & + \frac{A_i \Delta t'}{V_{i,k}^n} \left[\kappa_{i,k} \frac{C_{i,k+1}^{n+1} - C_{i,k}^{n+1}}{\Delta z_{i,k+1/2}} - \kappa_{i,k-1} \frac{C_{i,k}^{n+1} - C_{i,k-1}^{n+1}}{\Delta z_{i,k-1/2}} \right], \quad (k = k_b + 1, \dots, N_z) \end{aligned} \quad (14)$$

251 The FVUM scheme is conditional stable and the Courant number condition is
252 given by:

$$\Delta t' \leq \frac{V_{i,k}}{\sum_{j \in S^+} |Q_j|}, \quad (|Q_j| = S_j |u_j|) \quad (15)$$

253 The right hand side term of Eq. (15) is calculated for all prisms of the domain, $\Delta t'$ is
254 given by the minimum value obtained. Since this is a linear scheme, the vertical
255 advective flux part can be treated implicitly in Eq. (14), and in this case, S^+ does not

256 include the two vertical faces. Since the vertical flux terms imposes most stringent
 257 stability condition in shallow waters, this method results in more efficiency.

258 The second order TVD scheme is obtained by extending the upwind scheme
 259 and adding anti-diffusive fluxes to Eq. (12) as:

$$C_{j^*} = \hat{C}_{j^*} + \frac{\varphi_j}{2} (C_{jD} - \hat{C}_{j^*}) \quad (16)$$

260 where $\hat{C}_{j^*}$ is the upwind value, C_{jD} is the downwind value and φ_j is a limiter function
 261 that depends only of the values at face j ($0 \leq \varphi_j \leq 2$, for $\varphi_j=0$ is a upwind scheme, for
 262 $\varphi_j=2$ is a downwind scheme and for $\varphi_j=1$ is a central scheme).

263 Neglecting the non advective terms of Eq. (11) and substituting C_{j^*} given by
 264 Eq. (16), we have:

$$C_{i,k}^{n+1} = C_{i,k}^n + \frac{\Delta t'}{V_{i,k}^n} \sum_{j \in S^-} |Q_j| \left(1 - \frac{\varphi_j}{2} \right) (C_j - C_i) + \frac{\Delta t'}{V_{i,k}^n} \sum_{j \in S^+} |Q_j| \frac{\varphi_j}{2} (C_i - C_j) \quad (17)$$

265 with

$$\frac{\Delta t'}{V_i^n} \sum_{j \in S^+} |Q_j| \frac{\varphi_j}{2} (C_i - C_j) = \frac{\Delta t'}{V_{i,k}^n} \delta \sum_{j \in S^-} |Q_j| (C_j - C_i) \quad (18)$$

266 where

$$\lambda = \sum_{j \in S^-} |Q_j| (C_j - C_i) \quad (19)$$

$$\delta = \sum_{j \in S^+} \frac{|Q_j| \frac{\varphi_j}{2} (C_i - C_j)}{\lambda} \quad (20)$$

267 Substituting Eq. (18) into Eq. (17), and including the non advective terms, we
 268 obtained the following final equation for TVD scheme:

$$\begin{aligned}
C_{i,k}^{n+1} = & C_{i,k}^n \left[1 - \frac{\Delta t'}{V_i^n} \sum_{j \in S^-} |Q_j| \left(1 - \frac{\varphi_j}{2} + \delta \right) \right] + \frac{\Delta t'}{V_i^n} \sum_{j \in S^-} |Q_j| \left(1 - \frac{\varphi_j}{2} + \delta \right) C_j + \hat{Q}_{i,k}^{n+1} \Delta t' + \\
& + \Delta t' \sum_{j=1}^3 (\kappa_h) \frac{C_j^n - C_i^n}{\delta_{i,j}} + \frac{A_i \Delta t'}{V_i} \left[\kappa_{i,k} \frac{C_{i,k+1}^{n+1} - C_{i,k}^{n+1}}{\Delta z_{i,k+1/2}} - \kappa_{i,k-1} \frac{C_{i,k}^{n+1} - C_{i,k-1}^{n+1}}{\Delta z_{i,k-1/2}} \right], \\
& (k = k_b + 1, \dots, N_z)
\end{aligned} \tag{21}$$

269 The TVD scheme is stable if:

$$\begin{cases} \delta \geq 0 \\ 1 - \frac{\Delta t'}{V_{i,k}} \sum_{j \in S^-} |Q_j| (1 + \delta) \geq 0 \end{cases} \tag{22}$$

270 The δ and r_j values can be obtained using two different formulations, one
271 from the Casulli and Zanolli formulation (Casulli and Zanolli, 2005), and the other
272 given by:

$$\begin{cases} \delta = \sum_{j \in S^+} \frac{\varphi_j}{2r_j} \\ r_j = \frac{\sum_{m \in S^-} |Q_m| (C_m - C_i)}{|Q_j| (C_i - C_j)} \end{cases} \tag{23}$$

273 The first condition of Eq. (22) is easily satisfied. The second condition of Eq. (22)
274 becomes:

$$\Delta t' \leq \frac{V_i}{\sum_{j \in S^-} |Q_j| (1 + \delta)} \tag{24}$$

275 The right hand side term of Eq. (24) is evaluated for all prisms in the domain, and
276 $\Delta t'$ is then given by the minimum value obtain for the entire domain.

277 The second formulation implemented in SELFE for calculate δ is the Casulli
278 and Zanolli (2005) given by:

$$\left\{ \begin{array}{l} \delta = \sum_{j \in S^+} \frac{\varphi_j}{2r_j} \frac{|Q_j|}{\sum_{m \in S^-} |Q_m|} \\ r_j = \frac{\sum_{m \in S^-} |Q_m| (C_m - C_i)}{(T_i - T_j) \sum_{m \in S^-} |Q_m|} \end{array} \right. \quad (25)$$

279 Since $\delta \leq \sum_{j \in S^+} \frac{|Q_j|}{\sum_{m \in S^-} |Q_m|} = \frac{\sum_{j \in S^+} |Q_j|}{\sum_{m \in S^-} |Q_m|} = 1$ then the second condition of Eq. (22), the
 280 courant number conditions, is satisfied if:

$$\Delta t' \leq \frac{V_i}{2 \sum_{j \in S^-} |Q_j|} \quad (26)$$

281 The right hand side term of Eq. (26) is calculated for the entire prisms domain, with
 282 $\Delta t'$ given by the minimum value obtain for the entire domain. Our experience
 283 indicates that the condition in Eq. (25) is usually more efficient.

284 Sediment concentration values are converted to all nodes and levels only for
 285 writing in the output files and visualization. The change of the sediment
 286 concentration value from the prism center to the levels is made thought a weighed
 287 interpolation with the cell thickness. However, two sediment concentration values are
 288 unknown, the concentrations values in the surface (C_n) and in the bottom (C_1) levels.
 289 The surface level is obtained through a null flux condition in this boundary:

$$w_{s,q} C + \kappa \frac{\partial C}{\partial z} = 0 \quad \text{at } z = \eta \quad (27)$$

290 C_n value is obtained by the integration of Eq. (27) between the surface level (n) and
 291 the level below the surface ($n-1$):

$$C_n = \frac{C_{n-1} \kappa_{n-1}}{w_{s,q} \Delta z_n + \kappa_{n-1}} \quad (28)$$

where Δz_n is the surface cell thickness and C_n e C_{n-1} , the tracer values in the surface level and in the level below the surface, respectively.

Bottom tracer value (C_1) is obtained by mass conservation, i.e., assuming that the tracer total mass at the centers (M_c) is equal to the tracer total mass at the levels (M_n), with M_c and M_n given by:

$$M_c = \sum_{i=1}^{nc} C_i \Delta z_i \quad (29)$$

$$M_n = C_1 \frac{\Delta z_1}{2} + \sum_{i=2}^{n-1} C_i \left(\frac{\Delta z_{i-1} + \Delta z_i}{2} \right) + C_n \frac{\Delta z_n}{2} \quad (30)$$

where Δz_1 is the bottom cell thickness, n is the total number of levels, $nc = n - 1$ (total number of centers) and:

$$C_1 = \frac{- \sum_{i=2}^{n-1} C_i \left(\frac{\Delta z_{i-1} + \Delta z_i}{2} \right) - C_n \frac{\Delta z_n}{2} + \sum_{i=1}^{nc} C_i \Delta z_i}{\frac{\Delta z_1}{2} A_e} \quad (31)$$

where A_e is the area of the element e .

Sediment deposition

The vertical advection algorithm for the sediment vertical settling implemented in the MORSELE was adapted from the ROMS model sediment transport module (Warner *et al.*, 2008; <http://www.myroms.org/>). The vertical settling of suspended sediment is computed via a semi-lagrangean advective flux algorithm, that includes a piece-wise parabolic method (PPM; Colella and Woodward, 1984) and a weighted essentially non-oscillatory (WENO) scheme (Liu *et al.*, 1994). PPM use parabolas as interpolation functions and WENO integrates depositional flux over multiple grid cells, so it is not constrained by the Courant-Friedrichs-Lewy (CFL) criterion. Detailed description of PPM and WENO can be found, respectively, in Colella and Woodward (1984) and Liu *et al.* (1994).

311 Sediment erosion

312 Erosional flux obtained through Eq. (5) is computed near the bottom, this is in
 313 a reference height (a) near the bottom, which corresponds to the top of the bed load
 314 transport layer. Several formulations to determine the reference height exist in the
 315 literature (Einstein, 1950; Engelund and Fredsoe, 1976; Smith and Mclean, 1977;
 316 van Rijn, 1984; Garcia and Parker, 1991; Zyserman and Fredsøe, 1994; Cao, 1999).
 317 Zyserman and Fredsøe (1994) state that the reference height should be equivalent
 318 to the diameter of fine sediment grains, in MORSELF is used $a = 3d_{50}$ (Harris and
 319 Wiberg, 2001; Lawson *et al.*, 2007).

320 An innovative formulation to transfer the sediment erosive flux between the
 321 bottom and the center of the bottom computational cell is used in MORSELF. This
 322 transference is made considering that the sediment concentration profile between
 323 the reference height and the cell center follows the Rouse concentration profile given
 324 by:

$$\frac{C}{C_a} = \left[\frac{a(H-z)}{z(H-a)} \right]^b \quad (32)$$

325 where H is the total depth, $b = w_s / \kappa_0 u_*$, u_* is the bed shear velocity, C_a is the
 326 sediment concentration at the reference height a and z is the bottom distance (in this
 327 case is half of the cell bottom computational thickness).

328 For the sediment erosive flux transference to the water column is considered
 329 that the diffusive flux is given by the expression:

$$E_q = \kappa \frac{\partial C}{\partial z} \quad (33)$$

330 with $\partial C / \partial z$ being obtained through derivation of the Eq. (32):

$$\left. \frac{\partial C}{\partial z} \right|_a = b \left(\frac{a(H-z)}{z(H-a)} \right)^{b-1} \left(\frac{-aH}{z^2(H-a)} \right) \quad (34)$$

331 Thus, the erosive flux in the center of the computational cell is obtained
 332 solving the equations system composed by the erosive flux equations for the

333 reference height ($E_{q,a}$) and for the computational cell center located immediately
 334 above the bottom ($E_{q,z}$), being:

$$E_{q,a} = \kappa_a C_a \left. \frac{\partial C}{\partial z} \right|_a \quad (35)$$

$$E_{q,z} = \kappa_z C_a \left. \frac{\partial C}{\partial z} \right|_z \quad (36)$$

335 Replacing the C_a value obtained with the Eq. (35) into the Eq. (36) stay:

$$E_{q,z} = E_{q,a} \frac{\kappa_z}{\kappa_a} \frac{\left. \frac{\partial C}{\partial z} \right|_z}{\left. \frac{\partial C}{\partial z} \right|_a} \quad (37)$$

336 Substituting Eq. (34) into Eq. (37) we get:

$$E_{q,z} = E_{q,a} \frac{\kappa_z}{\kappa_a} \left(\frac{a(H-z)}{z(H-a)} \right)^{b-1} \frac{a^2}{z^2} \quad (38)$$

337 Assuming that the eddy diffusivity profile (κ) near the bottom has a parabolic shape
 338 given by (van Rijn, 1984b):

$$\kappa = \kappa_0 u_* z (1 - z/H) \quad (39)$$

339 and replacing Eq. (39) into Eq. (38), the following equation is obtained for $E_{q,z}$:

$$E_{q,z} = E_{q,a} \frac{(H-z)}{(H-a)} \left(\frac{a(H-z)}{z(H-a)} \right)^{b-1} \frac{a}{z} \quad (40)$$

340 Since a and z values are very small when compared with the H value, we can
 341 consider that $H - z \approx H - a \approx H$ and Eq. (38) is reduced to:

$$E_{q,z} = E_{q,a} \left(\frac{a}{z} \right)^b \quad (41)$$

Thus, the value of the erosive flux in the center of the bottom computational cell is given by Eq. (41) with $E_{q,a}$ given by Eq. (5). This approach, inspired in the approach used to compute the drag coefficient, turn the erosion flux independent of the vertical resolution near the bottom.

2.2.2 Bed load transport

2.2.2.1 Physical formulation

In the literature several formulae exists to calculate the bed load transport ($q_{b,q}$), in MORSELF two formulations are implemented: Meyer-Peter and Müller (1948) and van Rijn (2007) formulae.

For the Meyer-Peter and Müller (1948) formula, the volumetric bed load transport rate is defined as:

$$q_{b,q} = 8(\theta_{sf} - \theta_{cr,q})^{1.5} \sqrt{(s-1)gd_{50,q}^3} \quad (42)$$

where $\theta_{cr,q}$ is the critical Shields parameter given by Eq.(7) and θ_{sf} is the nondimensional Shields parameter given by:

$$\theta_{sf} = \frac{\tau_{sf}}{(\rho_s - \rho_0)gd_{50}} \quad (43)$$

Eq. (42) can be applied to sediment mixtures with $d_{50} > 0.4$ mm.

Van Rijn (2007) defined the bed load transport rate as:

$$q_{b,q} = 0.015 \rho_s \bar{u} H (d_{50,q}/H)^{1.2} M_e^{1.5} \quad (44)$$

where $\bar{u}$ is the depth-averaged velocity and M_e the mobility parameter given by:

$$M_e = (\bar{u} - u_{cr}) / [(s - 1)gd_{50}]^{0.5} \quad (45)$$

358 u_{cr} is the critical depth-averaged velocity for currents based on Shields initiation of
359 motion and is given by:

$$u_{cr} = \begin{cases} 0.19(d_{50})^{0.1} \log(12H/3d_{90}) & \text{for } 0.00005 < d_{50} \leq 0.0005 \text{ m} \\ 8.5(d_{50})^{0.6} \log(12H/3d_{90}) & \text{for } 0.0005 < d_{50} < 0.002 \text{ m} \end{cases} \quad (46)$$

360 Eq. (44) can be applied to $0.05 < d_{50} < 2.0 \text{ mm}$.

361 Assuming that bed load transport occurs in the same direction as the velocity
362 vector in the bottom computational cell, the bed load vector components are given
363 by:

$$q_{bx,q} = q_{b,q} \frac{\tau_{bx}}{|\tau_{sf}|} \quad (47)$$

$$q_{by,q} = q_{b,q} \frac{\tau_{by}}{|\tau_{sf}|} \quad (48)$$

364 where $q_{bx,q}$ e $q_{by,q}$ are the sediment fluxes for a given sediment of class q along the x
365 and y directions respectively.

366 In both formulae presented the magnitude of the bed load transport vectors are
367 calculated assuming a horizontal bottom, thus if a bed slope exist it is necessary to
368 adjust the vectors magnitude.

369 **Bed slope effects**

370 Several approximations exist to estimate the bed load transport on slopes.
371 Three methods are available in MORSELF: Antunes do Carmo (1995), Damgaard
372 *et al.* (1997) and Lesser *et al.* (2004).

373 Antunes do Carmo (1995) proposed the following change to the Meyer-Peter
374 e Müller (1948) formula in order to include bed slope effects:

$$q_{b,q} = 8(\theta_{sf} - \theta_{c,q})^{1.5} \sqrt{(s-1)gd_{50,q}^3} \quad (49)$$

where $\theta_{c,q}$ is the critical bed shear stress for a slope bottom given by :

$$\theta_{c,q} = \theta_{cr,q} \left(\cos\beta_s \cos\beta_t \sqrt{1 - \frac{\tan^2\beta_t}{\tan^2\phi}} + \frac{\sin\beta_s}{\tan\phi} \right) \quad (50)$$

ϕ (radians) is the angle of repose of the sediment, β_s is the bottom slope angle along the flow direction, $\tan\beta_t = \cos\beta_s \tan\beta_n$, β_n is the bottom slope angle perpendicular to the flow direction.

Damgaard *et al.* (1997), based on experimental results, modified Meyer Peter and Müller (1948) formula to consider the gravity effect in the longitudinal slope. The modified formula is given by:

$$q_{b,q} = 8(\theta_{sf} - \theta_{c,q})^{1.5} \sqrt{(s-1)gd_{50,q}^3} fslope \quad (51)$$

where $\theta_{c,q}$ is the threshold Shields parameter on a slopping bed corrected for the slope effect using the Whitehouse (1995) formula:

$$\theta_{c,q} = \theta_{cr,q} \frac{\sin(\phi - \beta_s)}{\sin\phi} \quad (52)$$

and $fslope$ parameter is a correction factor given by:

$$fslope = \begin{cases} 1 + 0.8 \left(\frac{\theta_{cr,q}}{\theta_{sf}} \right)^{0.2} \left(1 - \frac{\theta_{c,q}}{\theta_{cr,q}} \right)^{(1.5 + \theta_{sf}/\theta_{cr,q})} & \text{se } -\phi < \beta_s \leq 0 \\ 1 & \text{se } 0 < \beta_s < \phi \end{cases} \quad (53)$$

Lesser *et al.* (2004) proposed a method based on Bagnold (1966) and Ikeda (1982) works. The influence of bed-slope effects is included in two parts. In the first, a longitudinal slope in the direction of the bed-load transport modifies the magnitude of bed-load vector. This adjustment is calculated following a modified form of the expression suggested by Bagnold (1966):

$$(q_{bx,q}^*, q_{by,q}^*) = \alpha_s (q_{bx,q}, q_{by,q}) \quad (54)$$

390 where

$$\alpha_s = 1 + \alpha_{bs} \left[\frac{\tan \phi}{\cos \beta_s (\tan \phi - \tan \beta_s)} - 1 \right] \quad (55)$$

391 where α_{bs} is a user specified tuning parameter (default = 1) and β_s is the bed slope
392 angle in direction of bed load transport vector.

393 In the second part of the method, the direction of bed load transport vector is
394 adjusted if a bed slope exists in the direction normal to the bed load transport vector.
395 This modification is based on the work of Ikeda (1982) and is computed as follows:

$$q_{bx,q}^{**} = q_{bx,q}^* - \alpha_n q_{by,q}^* \quad (56)$$

$$q_{by,q}^{**} = q_{by,q}^* + \alpha_n q_{bx,q}^* \quad (57)$$

396 where

$$\alpha_n = \alpha_{bn} \left(\frac{\tau_{cr,q}}{\tau_b} \right)^{0,5} \tan \beta_n \quad (58)$$

397 in which α_{bn} is a user-specified coefficient (default = 1,5) and β_n is the bed slope
398 angle normal to the bed load transport vector.

399 **2.2.2.2 Numerical formulation**

400 Bed load fluxes are computed in the cell centers and limited by the availability
401 of each sediment class in the bottom top layer. For all the methods implemented to
402 estimate the bed load transport on slopes [Eqs. (50), (53), (55) and (58)] it is
403 necessary to determine the angles β_s and β_n :

$$\beta_s = \text{atan}\left(\frac{\partial h}{\partial x} \frac{\tau_{bx}}{|\tau_{sf}|} + \frac{\partial h}{\partial y} \frac{\tau_{by}}{|\tau_{sf}|}\right) \quad (59)$$

$$\beta_n = \text{atan}\left(-\frac{\partial h}{\partial x} \frac{\tau_{by}}{|\tau_{sf}|} + \frac{\partial h}{\partial y} \frac{\tau_{bx}}{|\tau_{sf}|}\right) \quad (60)$$

where $\partial h/\partial x$ and $\partial h/\partial y$ are, respectively, the bed slopes in the x and y directions and the angle β_s is limited by 0.9ϕ .

2.2.3 Sediment bed

The sediment bed is modelled as a layer system located above the water column, with a user specified constant number of layers. The system implemented in MORSELF is similar to the one implemented in the ROMS model (see Warner *et al.* (2007) for a detailed description). Each layer is initialized with a given thickness, sediment class distribution, porosity and age. The bed layers are modified at each time step to account for erosion, deposition and stratigraphy.

2.3 Bottom evolution module

In this module the bed changes due to suspended and bed load transports are computed. The total depth variation (Δh) is the sum of the bathymetric variations due to suspended and bed load transports.

The net sediment change due to suspended sediment transport is computed for each sediment class q as follows:

$$\Delta h_{se,q} = \frac{\hat{Q}_q \Delta t}{\rho_s (1 - p)} \quad (61)$$

where $\hat{Q}_q$ represents the settling velocity and the sediment bed exchange terms given by Eq. (2) and Δt is the computational time step.

Eq. (61) is computed in the prisms center and is given by the sum of the depth variations of all sediment classes simulated. The change from the prism centers to the nodes is made as follows:

$$\Delta h_{sn} = \frac{\sum_{e=1}^{nel} A_e \Delta h_{se}}{\sum_{e=1}^{nel} A_e} \quad (62)$$

where Δh_{se} is the element depth variation due to the suspended transport and nel is the number of elements that contain the node n .

To compute the change in the bottom sediment due to bed load transport the bottom evolution module of SAND2D model (Fortunato and Oliveira, 2004, 2007) was adapted. The model computes the resulting bed changes for each sediment class q through the resolution of Exner equation:

$$\Delta h_{b,q} = \frac{1}{1-p} \nabla q_{b,q} \Delta t \quad (63)$$

where $q_{b,q}$ is the bed load sediment flux.

Eq. (61) is solved with a node-centered finite volume technique based on an unstructured triangular grid. Bed load fluxes are computed at the prism center and limited by the availability of each sediment class in the bottom top layer. Flux transference from the prism center to the nodes is made through the computation of the line integral of the flux times the unit normal for each control volume. The control volume is defined through the element center line (Figure 3). Depth variation due to bed load transport is computed at nodes and is given by the sum of depth variations of all the sediment classes simulated.

3 MORSELF model validation

Four tests were performed to validate the MORSELF model. In the first test the formulation used to compute the sediment settling velocity term were validated. Test 2 demonstrates the model ability to simulate velocity and suspended sediment concentration profiles, through the comparison between the model results and the

analytical solution. Model ability to produce a classic suspended sediment concentration profile and the effect of varying turbulence sub models were demonstrated in test 3. In the last test, the two sediment transport components and the morphology component of the model were validated.

3.1 Test 1 – Deposition

Simulations in a square box, with 9 m height and 1 m depth, were performed in order to validate the model formulation used to compute the sediment settling velocity. All the boundaries were closed and inside the domain a null velocity were imposed. This way the sediment inside the box is only subject to the gravity action. Gaussian profile was imposed as sediment initial condition (represented in Figure 4 as initial profile), and simulations with different sediment diameters were performed. We only show the results for the simulation with $d_{50}=0.86$ mm and $w_s=0.1$ m/s (Figure 4) since the results obtained for the others sediment diameters were similar. A time step of 1 s and 47 vertical levels equally spaced were used in the simulation.

Figure 4 shows the evolution of the sediment concentration profile along the simulation in a point located in the box center. The shape of the sediment profile is not modified along the simulation (grey profiles) and the profile settling is equal to the velocity times the time step (0.1 m), showing that sediment settling velocity term is well represented by the model.

3.2 Test 2 – Equilibrium sediment concentration profiles

This test shows the model capacity to correctly simulate velocity and suspended sediment concentration profiles through the comparisons of the model results with the analytical solutions. In this test, the bed load transport is not considered and Eq. (33) is used to compute the erosion flux.

Under equilibrium (stationary) conditions the advection-diffusion equation [Eq. (1)] reduces to

$$Cw_{s,q} + \kappa \frac{\partial C}{\partial z} = 0 \quad (64)$$

as presented by Rouse (1937). This equation can be solved analytically under the assumptions of small concentrations (constant $w_{s,q}$) and a parabolic sediment mixing profile [Eq. (39)], resulting in the expression for the “Rouse” sediment concentration profile [Eq. (32)].

The reference concentration is calculated in accordance with van Rijn (1984b):

$$C_a = 0,015 \frac{d_{50}}{a} \frac{T_a^{1,5}}{D_*^{0,3}} \quad (65)$$

where a is the reference height ($a=0.01H$), D_* is given by Eq. (4) and T_a is the dimensionless bed-shear stress given by:

$$T_a = \frac{\mu_c \tau_{b,c} - \tau_{cri}}{\tau_{cri}} \quad (66)$$

where $\mu_c = f'_c / f_c = (\log_{10}(12H/30z_0))^2 / (\log_{10}(12H/3d_{90}))^2$ is the efficiency factor current, $\tau_{b,c} = \rho_w u_*^2$ is the bed shear stress, $\tau_{cri} = (\rho_s - \rho_w) g d_{50} \theta_{cri}$ is the critical bed shear stress, and θ_{cri} is the threshold parameter, calculated according to the classical Shields curve as modeled by van Rijn (1993).

Analytical velocity profile can be obtained through the vertical integration of the horizontal momentum equation for a uniform and barotropic flow:

$$g \frac{\partial \eta}{\partial x} = \frac{\partial}{\partial z} \left(\kappa(z) \frac{\partial u(z)}{\partial z} \right) \quad (67)$$

Velocity profile expression is achieved through the double integration of Eq. (67), with the first integration between z and H and the second between z_0 and z , resulting:

$$u(z) = -\frac{u_*}{\kappa_0} \left(\ln \left(\frac{z}{z_0} \right) \right) \quad (68)$$

A series of tests were performed to simulate the suspended sediment transport in a long (8000 m) and straight flume (320 m) with a water depth of 5 m

489 (H5) and 10 m (H10), and a constant slope of 3.2×10^{-4} m/m and 1.55×10^{-4} m/m,
 490 respectively. Table 1 shows the model parameters used in the simulations.

491 Simulations with different number of vertical levels (5, 11, 21, 41 and 81
 492 levels evenly spaced and S coordinates) and with two sediment diameters
 493 ($d_{50}=0.2$ mm and $w_s=0.02$ ms⁻¹; $d_{50}=0.8$ mm and $w_s=0.097$ ms⁻¹) were performed. For
 494 the S vertical grids, the values used for the constants that control the vertical
 495 resolution near the bottom and surface were $\theta_b=1$ (both surface and bottom are
 496 solved) and $\theta_f=5$ ($0 < \theta_f \leq 20$ and for $\theta_f \gg 1$ more resolution is skewed toward the
 497 boundaries). The model results correspond to a point located on the longitudinal
 498 centerline of the flume 6160 m from the upstream boundary.

499 **3.2.1 Velocity profiles**

500 The comparison between the analytical and the numerical velocity profiles
 501 were performed through the computation of the root mean squared error (RMSE):

$$RMSE = \sqrt{\frac{\sum_{k=1}^{N_z} (v_{a,k} - v_{n,k})^2}{n}} \quad (69)$$

502 where n is the total number of vertical levels, $v_{a,k}$ and $v_{n,k}$ are, respectively, the
 503 analytical and numerical velocities for a given k level.

504 The analyze of the RMSE values (Figure 5) obtained for the simulations with
 505 different number of vertical levels and the channel with 5 m (H5) and 10 m (H10)
 506 depth shows that velocity errors decrease with the increase of the vertical resolution.
 507 In both channels, the minimum RMSE values were obtained for the simulations with
 508 81 S levels (about 1%). For the H5 channel simulations, similar RMSE values were
 509 obtained for the evenly spaced and S vertical grids, showing that for this depth is
 510 indifferent the type of vertical grid used. For the simulations in the H10 channel,
 511 smaller RMSE values were obtained with the S vertical grids. The difference
 512 between the RMSE values obtained with the two grid types decrease with the
 513 increase of the vertical refinement, with a difference of 0.1 m/s and 0.005 m/s for the
 514 grids with 5 and 81 levels, respectively. Thus, considering only the hydrodynamic, S
 515 grids should be use instead of evenly spaced grids.

3.2.2 Suspended sediment profiles

For the suspended sediment simulations, additional tests were performed with two sediment grain size: $d_{50}=0.2$ mm with $w_s = 0.02$ m/s (fine sand) and $d_{50}=0.8$ mm with $w_s = 0.097$ m/s (medium sand).

To quantify the difference between the analytical and the numerical sediment concentration results relative errors were calculated through the following formula:

$$RE = \sqrt{\frac{\sum_{k=1}^{N_z-1} (C_{a,k} - C_{n,k})^2 \Delta z_k}{H}} \bigg/ \sqrt{\frac{\sum_{k=1}^{N_z-1} C_{a,k}^2 \Delta z_k}{H}} \quad (70)$$

where $C_{a,k}$ and $C_{n,k}$ are, respectively, the analytical and numerical sediment concentration for a given k level, Δz is the k cell thickness.

The error results obtained for the simulation with $d_{50}=0.2$ mm (Figure 6) shows that the sediment concentration errors decrease with the increase of the vertical refinement, with exception for the simulations in the H10 channel and evenly spaced levels where an error of 7% were achieved for all the grids simulated. Generally, acceptable errors were obtained for the sediment concentration, with a maximum error of about 10% for the case H10 and 5 S levels and a minimum error of about 3% for the case H5 and 81 S levels. The test shows that, for this type of sediment, the best results were achieved with S grids, with exception for the simulations in H5 channel and 5 levels and in the H10 channel and less than 21 levels, where the evenly spaced grids presented better results.

For the simulations with $d_{50}=0.8$ mm and less than 21 vertical levels, errors of about 60 % for sediment concentration were obtained (Figure 7). However, with the increase of vertical levels the error is reduced, with a minimum error of about 13% for the case with 81 S levels. This result shows that for the simulations with coarser sediment the use of a higher vertical resolution is necessary to minimize the sediment concentration errors, and in this cases S grids should be use instead of evenly spaced.

To verify if a higher resolution near the boundaries improve the results, simulations for the channel with H5 and S grid with $\theta_f=20$ were performed. The errors obtained with this grid (Figure 7) are similar to the ones achieved with the S grid and

$\theta_f=5$, thus a higher resolution near the surface and bottom do not seem to improve the sediment concentration results.

3.3 Test 3 – Steady uniform channel flow

This test case demonstrates the model ability to simulate vertical profiles of suspended sediment concentration (without bed load) and is a reproduction of the test presented in Warner *et al.* (2005, 2008). The simulation represents the suspended sediment transport for a steady uniform flow in a long (10000 m) and straight (1000 m) channel with a constant slope of 4×10^{-5} m/m. A constant flow rate of $10000 \text{ m}^3\text{s}^{-1}$ is imposed at upstream boundary and a fixed depth of 10 m is maintained at the downstream boundary. The water surface elevation is allowed to vary along the length of the channel. Table 1 shows the model parameters used in the simulations.

Simulation with different turbulence closures were performed using a vertical grid with 21 S levels ($\theta_b=1$ and $\theta_f=3$). Results obtained with the $k\text{-}kl$, $k\text{-}\varepsilon$ ($k\text{-}\varepsilon$), MY (Mellor and Yamada, 1982), UB (Umlauf and Burchard, 2003) turbulence models and an analytical expression are compared. The analytical expression is obtained by stipulating a parabolic shape to the eddy viscosity (ν) profile [Eq.(39)]. In the model the eddy diffusivity is determined from the turbulent Prandtl number and for a stable flow $Pr = 0.39/0.49 = 0.80$ (Kantha and Clayson, 1994). Thus, the algebraic eddy diffusivity is $\kappa = \nu/0.8$.

Model results are compared 1 km from the downstream end. Simulations with different turbulence closures produce different velocity, eddy diffusivity and sediment concentration profiles (Figure 8). Similar velocity profiles were obtained with the different turbulence closures simulated, however when compared with the analytical profile the velocities obtained are slightly higher near the surface and lower near the bottom. Results obtained using the $k\text{-}\varepsilon$ closure match closely the analytical profile for eddy diffusivity and sediment concentration. Similar results were obtained by Warner *et al.* (2005). Suspended sediment profiles are sensitive to vertical mixing and the range of the concentration profiles obtained demonstrates the importance of turbulence parameterization for an accurate simulation of suspended transport.

3.4 Test 4 – Trench migration

This test reproduced the laboratory experiment of van Rijn (1987) and allows validating the two sediment transport components (suspended load and bed load) and the morphologic evolution of MORSELFE model.

3.4.1 Laboratory experiment

The experiment consisted of measuring velocity profiles, sediment concentration profiles and bed level changes in a straight and long channel (length = 30 m, width = 0.5 m) with a vertical depression (trench) in the mobile sand bed. The characteristic diameters of the sediment material used in the experiment were $d_{50}=160\text{ }\mu\text{m}$.

To maintain the equilibrium conditions upstream of the trench (no scour or deposition), sediment of the same size and composition was supplied at a constant rate of 0.04 kg/sm (van Rijn, 1987). The water depth and the mean flow velocity upstream of the trench was held constant ($h_0=0.39\text{ m}$, $u_0=0.51\text{ m/s}$). Figure 9 shows the initial trench dimensions and the monitoring points, denominated E1, E4, E6, E7 and E8.

Flow velocities were measured using a micro-propeller meter and the sediment concentrations were measured by collecting water-sediment samples in the center line of the flume. Simultaneous measurement of the flow velocity and the concentration profiles were performed. Based on the measured velocities and sediment concentrations upstream of the trench, the equilibrium suspended sediment transport rate calculated was $S_s=0.03\text{ kg/sm}$. Based on the analysis of suspended sediment samples, the size of the suspended sediment grain was found to be $160\text{ }\mu\text{m}$ near the bottom and about $120\text{ }\mu\text{m}$ near the surface. The bed roughness was obtained by analysing the measured near-bed velocities upstream of the trench yielding $z_0 = 0.833\text{ mm}$.

3.4.2 Numerical simulations

Numerical simulations were performed in a 30 m long channel with a movable sand bed and a slight bed slope ($4.0\times 10^{-4}\text{ m/m}$) so that the water depth and the depth mean velocity were identical in both channel ends. At the inflow boundary

a constant water level, a “Rouse” sediment concentration profile and a logarithmic velocity profile (with a depth mean velocity of 0.51 m/s) were imposed. Sediment with a characteristic diameter of 140 μm was used in the simulations. Bed load transport was simulated with the van Rijn formula (van Rijn, 2007) with the Lesser *et al.* (2004) method to estimate the bed slope effects. The E_o value was calibrated by comparison with the measured suspended sediment concentration profiles. Table 2 shows the model parameters used in the simulations.

Simulations with vertical grids with 21, 41 and 81 evenly spaced levels were performed. In simulations, flow and suspended sediment are allowed to reach the steady state before morphological evolution is initiated. The total time of simulation is 15 hours and the initial time ($t = 0$) corresponds to 25 minutes of simulation, the time necessary to the model reached the equilibrium. In the simulations, an unlimited supply of sediment is available and the $k-\varepsilon$ turbulence closure model is used. The longitudinal profiles were taken from the channel center.

Figure 10 shows the computed and measured velocities and suspended sediment profiles at the 5 locations at initial time ($t = 0$). The agreement between the measured and computed velocity and suspended sediment profiles is reasonable. The velocity profiles obtained with the different vertical grids simulated are similar except near the bottom, where the grid with high vertical resolution presented high velocity values. Similar behavior is observed in the suspended sediment concentration profiles comparison, showing the importance of the resolution near the bottom for the sediment concentration simulations.

Comparisons between the simulated and measured velocity and suspended sediment concentration profiles were performed through the computation of the RMSE (Eq. (696969)). To dimensionless the errors, for each station the computed RMSE values were divided by the mean measured values.

The velocities errors (Figure 11) analyze show that the errors slightly decrease with the increase of the number of vertical levels, with smaller errors obtained for the stations E1 and E8 (about 10%) and errors of about 20% for the other stations. Similar errors were obtained for stations E4, E6 and E7, but with the major differences between simulated and measured profiles occurring in distinct zones of the water column. At station E4 the major differences occur near the top, at E6 in the middle and near the top of the water column and at E7 between the middle

and the bottom of the water column. The flow deceleration that occurs in the center of the trench (E6) near the bottom is well reproduced by the model.

For the sediment concentration, the errors obtained in some stations do not diminish with the increase of the vertical levels (Figure 11). This behavior can be related with the inexistence of measures near the bottom where the simulated profiles presented the major differences. The comparison of the errors obtained for the different stations shows that the larger errors occur for the stations E7 and E8 and the smaller for station E1. At stations E7 and E8 both suspended sediment concentration and velocity profiles are underestimated near the bottom, thus the sediment concentration errors can be associated with the velocities errors. At station 4 the computed velocity profile shows the correct behavior but the sediment concentration profile is underestimated from mid depth until the bottom. However, near the bottom where the larger difference between the measured and simulated sediment concentration values occurs there are no velocity measurements.

Figure 12 illustrates the channel hourly evolution along the 15 hours of simulation. As the flow progress into the deeper water of the trench, the flow velocity and the bottom stress decrease and the sediment is deposited. The bed load transport converges and the sediment accumulates at the upstream edge of the trench. At the downstream edge of the trench, sediment is picked back up by the accelerating flow and erosion occurs. The trench migrates in the flow direction as the upstream edge is filled and the downstream edge eroded.

Figure 13 shows the measured and computed position of the trench after 15 hours. Inspection of the simulated bottom profiles shows that the trench as been reduced to near half of its initial depth, and has migrated about 2 m downstream. Comparisons between the simulated and measured bottom profile after 15 hours shows that the trench appears to have moved nearly the correct distance, is filled to the correct level and presents an entry bottom slope similar to the measured. Less well simulated is the exit bottom slope, where the model is not able to reproduce the measured values. This discrepancy can be explained by the differences between the measured and simulated velocities at station E7, where the model underestimates the velocity and consequently the suspended sediment transport. The comparison of the bottom profiles obtained for the simulations with only the suspended sediment transport and the total transport (Figure 13) shows, as expected, that the inclusion of

bed load transport significantly increase the quantity of sediment that accumulates in the trench.

This test is frequently used to evaluate the morphodynamic models performance. The results obtained with two other 3D morphodynamic models [DELFT3D (Lesser *et al.*, 2000; Lesser *et al.*, 2004) and ROMS (Warner *et al.*, 2008)] shows that both models have the same difficulty in reproducing the exit bottom slope, and the trench migration seems to be better reproduced by the ROMS model than by the MORSELFE and DELFT3D models.

4 Conclusion

A new 3D morphodynamic model (MORSELFE) has been developed and validated to study the estuaries morphological evolution. MORSELFE consists in a set of integrated modules that aims to simulate shallow water flows, non-cohesive sediment transport and bottom changes in estuaries. The development of this modeling system was based on existing models, although several changes were introduced to the original models not only to make possible the models connection and integration but also to improve the original model formulation. The most important changes introduced in the original models were:

- In the sediment transport model the introduction of a new formulation to transfer the erosive flux from the bottom to the bottom computational cell center; with this approach the erosion flux computation is independent of the bottom layer thickness.
- In the hydrodynamic model the extension of the equation of the state for sea water density to include the density effect of suspended sediment; this inclusion enables the model to simulate processes where sediment density influences hydrodynamics;
- In the sediment transport model the implementation of van Rijn formula (van Rijn, 2007) for computing bed load transport and several formulae to adjust bed load transport for bed slope effects; increasing the users options for the bed load transport computation;
- In the sediment transport model the introduction of a formulation to convert the sediment concentration values from the prism centers to the levels.

701 The model has been validated through the simulation of hypothetical and real
702 sediment transport problems. The validation tests allowed not only to validate the
703 different model components but also to gain some sensibility about some of the
704 model parameters that can influence the simulations and the values that these
705 parameters should take, such as, time step, number of vertical levels and the
706 turbulence closure model to use.

707 The different validation tests performed shows that: (a) the sediment settling
708 velocity term is well represented by the model; (b) the error in the velocity estimative
709 decrease with the increase of the vertical grid refinement and that *S* vertical grids
710 should be used instead of evenly spaced; (c) *S* grids are more appropriated to
711 simulate the sediment concentration, particularly for the coarse sediment; (d)
712 acceptable representation of the sediment concentration can be obtained with 20 *S*
713 vertical levels; (e) for the simulations with coarse sediment is necessary to have a
714 high vertical resolution to minimize the sediment concentration errors; (f) simulations
715 with different turbulence closures produce different velocity, eddy diffusivity and
716 sediment concentration profiles; (g) model is able to reproduce the channel trench
717 migration. Comparisons between the simulated bottom profile and the measured
718 bottom profile after 15 hours shows that the trench appears to have moved nearly
719 the correct distance, is filled to the correct level and presents an entry bottom slope
720 similar to the measured. However, MORSEFE is not able to reproduce the exit
721 bottom slope, similar results were obtained with other models (Lesser *et al.*, 2004;
722 Warner *et al.*, 2008).

723 In MORSEFE, the suspended transport, the bed load transport and the
724 morphological changes are computed using the same time step as the flow model.
725 One factor that contributes for the model efficiency is the use of *S* coordinates in the
726 vertical discretization. With a larger vertical resolution near the bottom is possible to
727 reduce the number of vertical levels needed to obtain a good representation of the
728 sediment concentration. From the testes performed, 20 *S* levels appear to be
729 required to obtain a reasonable representation of the sediment concentration. The
730 combination of the vertical *S* coordinates representation and the possibility of using
731 multiple processors in the simulations allows morphological simulations to be
732 covered within acceptable run times.

733 This model is a useful tool that can be use to improve the knowledge of
734 morphological processes, namely in fast evolution systems and systems subject to

extreme events (floods). The several simulations performed showed that for a three dimensional model MORSELFE is robust and the morphological simulations can be performed with acceptable computational times. Next steps, include the model application to a real system and the coupling to a wave model.

Acknowledgments

This work was partially by the Fundação para a Ciência e a Tecnologia (PhD fellowship SFRH/BD/10404/2002) and by Laboratório Nacional de Engenharia Civil. The authors thank Prof. António Melo Baptista, from CMOP (Center for Coastal Margin Observation and Prediction) for the SELFE model and the ROMS team for the sediment transport module.

References

- Antunes do Carmo, J.S., 1995. Contribuição para o estudo dos processos morfodinâmicos em regiões costeiras e estuarinas. PhD Thesis, University of Coimbra (in portuguese).
- Ariathurai, R., Arulanandan, K., 1978. Erosion rates of cohesive soils. ASCE Journal of Hydraulic Division, 104 (2), 279-283.
- Bagnold, R.A., 1966. An approach to the sediment transport problem from general physics. Geological Survey Professional Paper, vol. 422-I, US Geological Survey, DOI, Washington.
- Blaas, M., Dong, C., Marchesiello, P., McWilliams, J.C., Stolzenbach, K.D., 2007. Sediment-transport modeling on Southern Californian shelves: a ROMS study. Continental Shelf Research, 27, 832-853.
- Canuto, V.M., Howard, A., Cheng, Y., Dubovikov, M.S., 2001. Ocean Turbulence I: one-point closure model. Momentum and heat vertical diffusivities. Journal of Physical Oceanography, 31, 1413-1426.
- Cao, Z., 1999. Equilibrium near-bed concentration of suspended sediment. Journal of Hydraulic Engineering, 125 (12), 1270-1278.

762 Casulli, V., Zanolli, P., 2005. High resolution methods for multidimensional
 763 advection-diffusion problems in free-surface hydrodynamics. *Ocean Modelling*, 10,
 764 137-151.

765 Collella, P., Woodward, P., 1984. The piecewise parabolic method (PPM) for
 766 Gas-Dynamical simulations. *Journal of Computational Physics*, 54, 174-201.

767 Damgaard, J.S., Whitehouse, R.J.S., Soulsby, R.L., 1997. Bed-Load
 768 Sediment Transport on Steep Longitudinal Slopes. *Journal of Hydraulic Engineering*,
 769 vol. 123, no.12, 1130-1138.

770 Einstein, H.A., 1950. The bed-load function for sediment transportation in
 771 open channel flow. Technical Bulletin No. 1026, U.S. Department of Agriculture,
 772 Washington, D.C.

773 Engelund, F., Fredsoe, J., 1976. A sediment transport model for straight
 774 alluvial channels. *Nordic Hydro.*, 7(5), 293-306.

775 Fortunato, A.F., Oliveira, A., 2004. A modeling system for tidally driven long-
 776 term morphodynamics. *Journal of Hydraulic Research*, 42 (4), 426-434.

777 Fortunato A.F., Oliveira, A., 2007. Improving the stability of a morphodynamic
 778 modeling system. *Journal of Coastal Research*, Special Issue 50, 486-490.

779 Galperin, B., Kantha, L.H., Hassid, S., Rosati, A., 1988. A quasi-equilibrium
 780 turbulent energy model for geophysical flow. *Journal of Atmosphere Science*, 45, 55-
 781 62.

782 Garcia, M., Parker, G., 1991. Entrainment of bed sediment into suspension.
 783 *Journal of Hydraulic Engineering*, 117 (4), 414-435

784 Harris, C.K., Wiberg, P.L., 1997. Approaches to quantifying long-term
 785 continental shelf sedimenttransport with an example from the Northern California
 786 STRESS mid-shelf site. *Continental Shelf Research*, 17 (11), 1389-1418.

787 Ikeda, S., 1982. Lateral bed-load transport on side slopes, *ASCE Journal of*
 788 *Hydraulic Division*, 108 (11), 1369-1373.

789 Kanta, L.H., Clayson, C.A., 1994. An improved mixed layer model for
 790 geophysical application. *Journal of Geophysical Research*, 99 (25), 235-266.

791 Lane, E.W., 1955. Design of stable channels. Transactions, ASCE, 120,
792 1234-1260.

793 Lawson, S.E., Wiberg, P.L., McGlathery, K.J., Fugate, D.C., 2007. Wind-
794 driven Sediment Suspension Controls Light Availability in a Shallow Coastal Lagoon.
795 Estuaries and Coasts, 30 (1), 102–112.

796 Lesser, G.R., Roelvink, J.A., van Kester, J.A.T.M., Stelling, G.S., 2004.
797 Development and Validation of a three-dimensional morphological model. Coastal
798 Engineering, 51, 883-915.

799 Lesser, G.R., Roelvink, J.A., van Kester, J., 2000. On-line Sediment transport
800 within DELFT3D-FLOW. Delft Hydraulics Report Z2899,IHE, Delft, The Netherlands.

801 Liu, X.D., Oscher, S., Chan, T., 1994. Weighted essentially non-oscillatory
802 schemes. Journal of Computational Physics, 115, 200-212.

803 Mellor, G.L., Yamada, T., 1982. Development of a turbulence closure model
804 for geophysical fluid problems. Reviews of Geophysics and Space Physics, 20 (4):
805 851-875.

806 Meyer-Peter, E., Muller, R., 1948. Formulas for bed-load transport. Proc.
807 IAHR, Stockholm.

808 Oliveira, A., Fortunato, A.B., Baptista A.M., 2000. Mass conservation in
809 Eulerian-Lagrangian simulations of transport in estuaries. ASCE Journal of Hydraulic
810 Engineering, 126 (8), 605-614.

811 Oliveira, A., Baptista, A.M., 1998. On the role of tracking on Eulerian-
812 Lagrangian solutions of the transport equation. Advance in Water Resource, 21,
813 539-554.

814 Pond, S., Pickard, G.L., 1998. Introductory Dynamic Oceanography.
815 Butterworth-Heinmann.

816 Rodi, W., 1984. Turbulence Models and Their Application in Hydraulics – a
817 State of the Art review. International Association of Hydraulic Research, Delft,
818 Netherlands.

819 Rouse, H., 1937. Modern conceptions of mechanics of turbulence.
820 Transactions of American Society of Civil Engineers, v. 102, p. 436-505.

821 Smith, J.D., Mclean, S.R., 1977. Spatially averaged flow over a wavy surface.
822 Journal of Geophysical Research, 82(12), 1735-1746.

823 Song, Y., Haidvogel, D., 1994. A semi-implicit ocean circulation model using
824 a generalized topography-following coordinate system. Journal of Computational
825 Physics, 115, 228-244.

826 Soulsby, R.L., 1997. Dynamics of marine sands, a manual for practical
827 applications. Thomas Telford, H.R. Wallingford, England.

828 Soulsby, R.L., Whitehouse, R.J.S.W., 1997. Threshold of sediment motion in
829 coastal environments. Stive, M.J.F., et al (Eds.), Proceedings of the Pacific Coasts
830 and Ports'97 conference. University of Canterbury, New Zealand, Christchurch, 149-
831 154.

832 Sweby, P.K., 1984. High resolution schemes using flux limiters for hyperbolic
833 conservation laws. SIAM Journal of Numerical Analysis, 21(5), 995-1011.

834 Umlauf, L., Burchard, H., 2003. A generic length-scale equation for
835 geophysical turbulence models. Journal of Marine Research, 6, 235-265.

836 van Rijn, L.C., 2007. Unified view of sediment transport by currents and
837 waves. I: initiation of motion, bed roughness, and bed-load transport. Journal of
838 Hydraulic Engineering, 133 (6), 649-667.

839 van Rijn, L.C., 1984. Sediment transport, part II: suspended load transport.
840 Journal of Hydraulic Engineering, 110 (11), 1613-1641.

841 van Rijn, L.C., 1993. Principles of sediment transport in rivers, estuaries and
842 coastal seas. Aqua publications, Amsterdam.

843 Warner, J.C., Sherwood, C.R., Arango, H.G. Signell, R.P., 2005.
844 Performance of four turbulence closure models implemented using a generic length
845 scale method. Ocean Modelling, 8, 81-113.

846 Warner, J.C., Sherwood, C.R., Signell, R.P., Harris, C.K., Arango, H.G.,
847 2008. Development of a three-dimensional, regional, coupled wave, current, and
848 sediment-transport model. *Computers & Geosciences*, 34, 1284-1306.

849 Wilcox, D.C., 1998. Reassessment of scale determining equation for advance
850 turbulence models. *AIAA Journal*, 26, 1299-1310.

851 Whitehouse, R.J.S., 1995. Observations of the boundary layer characteristics
852 and the suspension of sand at a tidal site. *Continental Shelf Research*, 15 (13),
853 1549-1567.

854 Xu, J.P., Noble, M., Eittreim, S.L., 2002. Suspended sediment transport on
855 the continental shelf near Davemport, California. *Marine Geology*, 181, 171-193.

856 Zeng, X., Zhao, M., Dickinson, R.E., 1998. Intercomparision of bulk
857 aerodynamic algorithms for the computation of sea surface fluxes using TOGA
858 COARE and TAO data. *Journal of Climate*, 11 (10), 2628-2644.

859 Zhang, Y.-L., Baptista, A.M., Myers, E.P., 2004. A cross scale model for 3D
860 baroclinic circulation in estuary-plume-shelf system: I. Formulation and skill
861 assessment. *Continental Shelf Research*, 24, 2187-2214.

862 Zhang, Y.-L., Baptista, A.M., 2008. SELFIE: A semi-implicit Eulerian-
863 Lagrangian finite-element model for cross-scale ocean circulation. *Ocean Modelling*,
864 21 (3-4), 71-96.

865 Zyserman, J.A., Fredsoe, J., 1994. Data analysis of bed concentration of
866 suspended sediment. *Journal of Hydraulic Engineering*, 120 (9), 1021-1042.

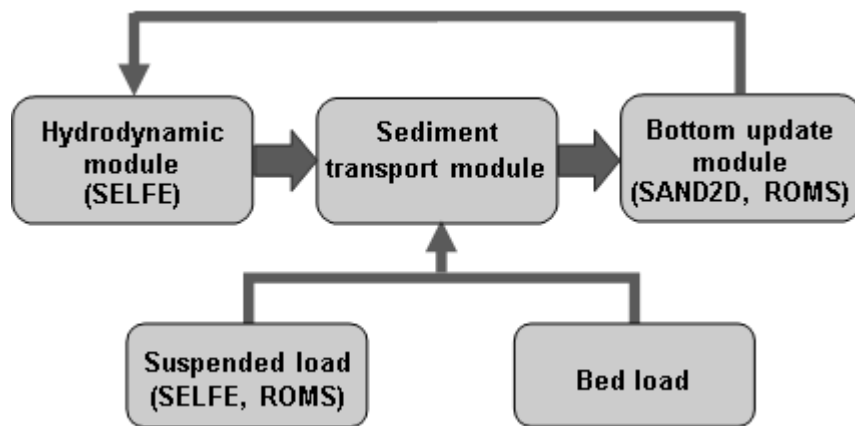


Figure 1 - MORSELFE model scheme.

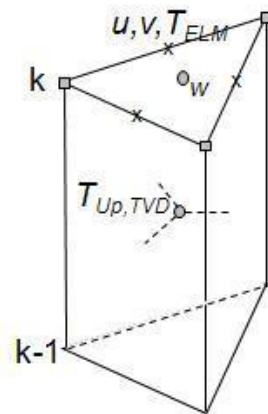


Figure 2 - Triangular prism and variables scheme.

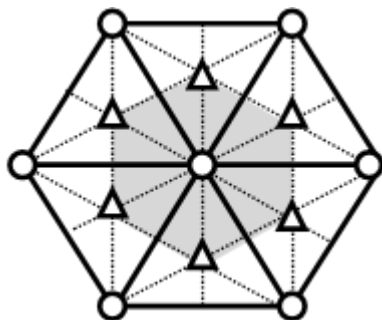


Figure 3 - The control volume (grey area) is defined by center lines (dotted lines) of the elements (lines). Circles and triangles correspond to nodes and element centers, respectively.

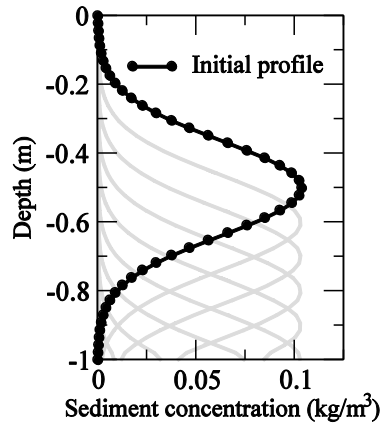


Figure 4 - Sediment concentration profile evolution for the simulation with $d_{50}=0.86$ mm and $w_s=0.1$ m/s.

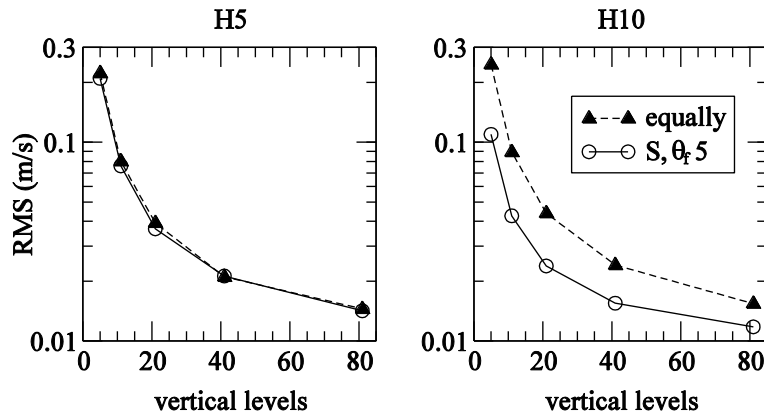


Figure 5 - Velocity RMSE for the different vertical grids.

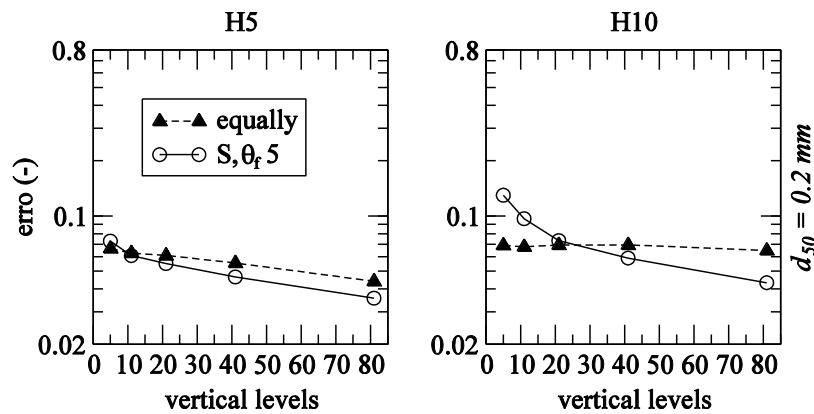


Figure 6 - Sediment concentration relative error for the different vertical grids for the H5 and H10 channel and for the $d_{50} = 0.2$ mm.

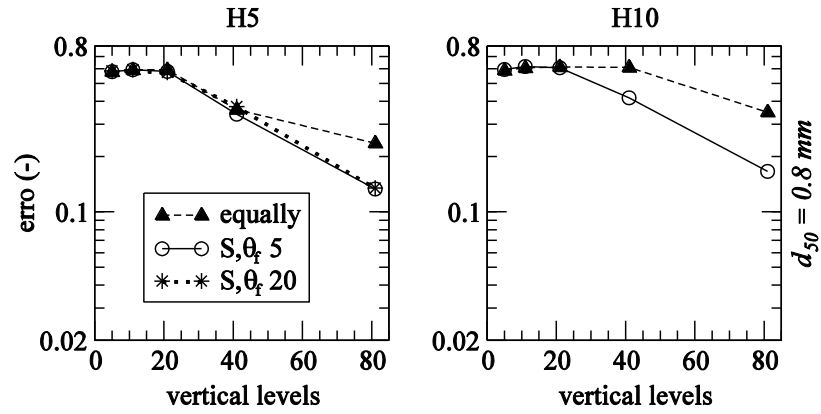


Figure 7 - Sediment concentration relative error for the different vertical grids for the H5 and H10 channel and for the $d_{50} = 0.8$ mm.

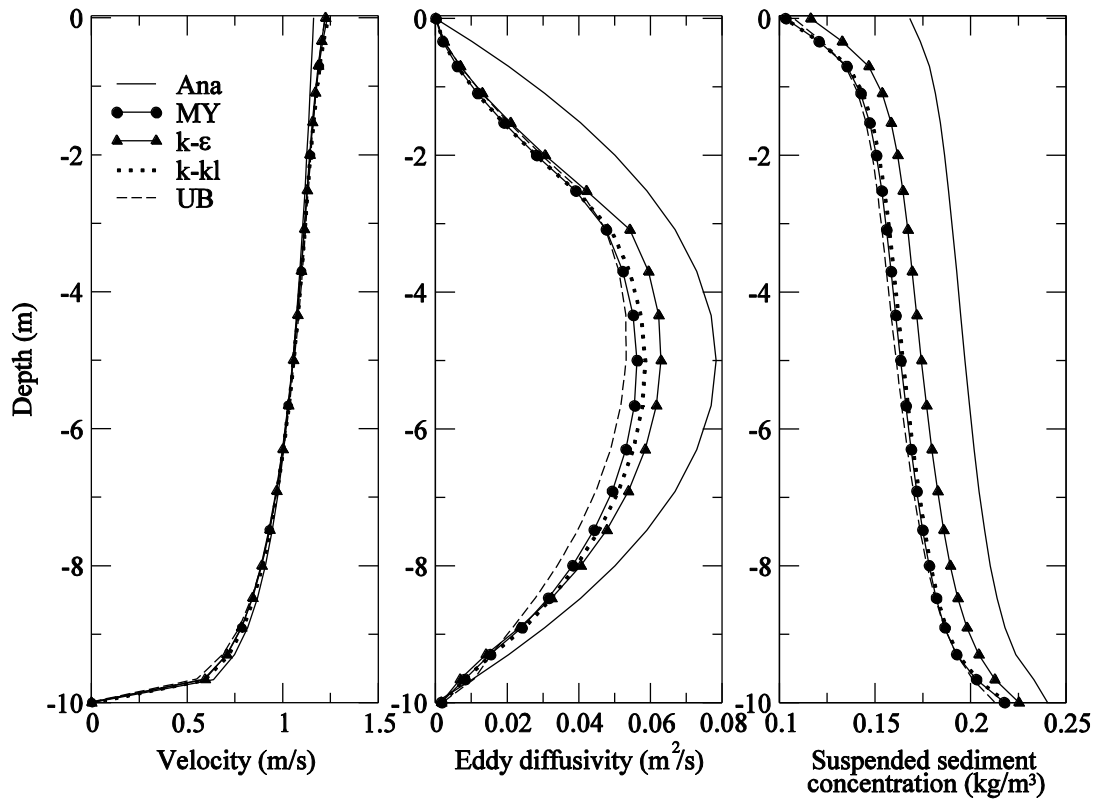


Figure 8 - Vertical profiles of velocity, eddy diffusivity and suspended sediment concentration for the four turbulence closures and for the analytical expression.

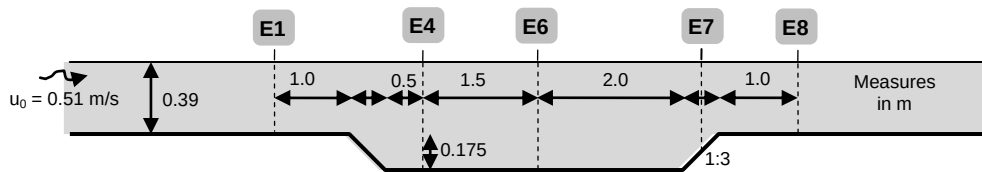


Figure 9 - Initial trench dimensions and location of the monitoring stations.

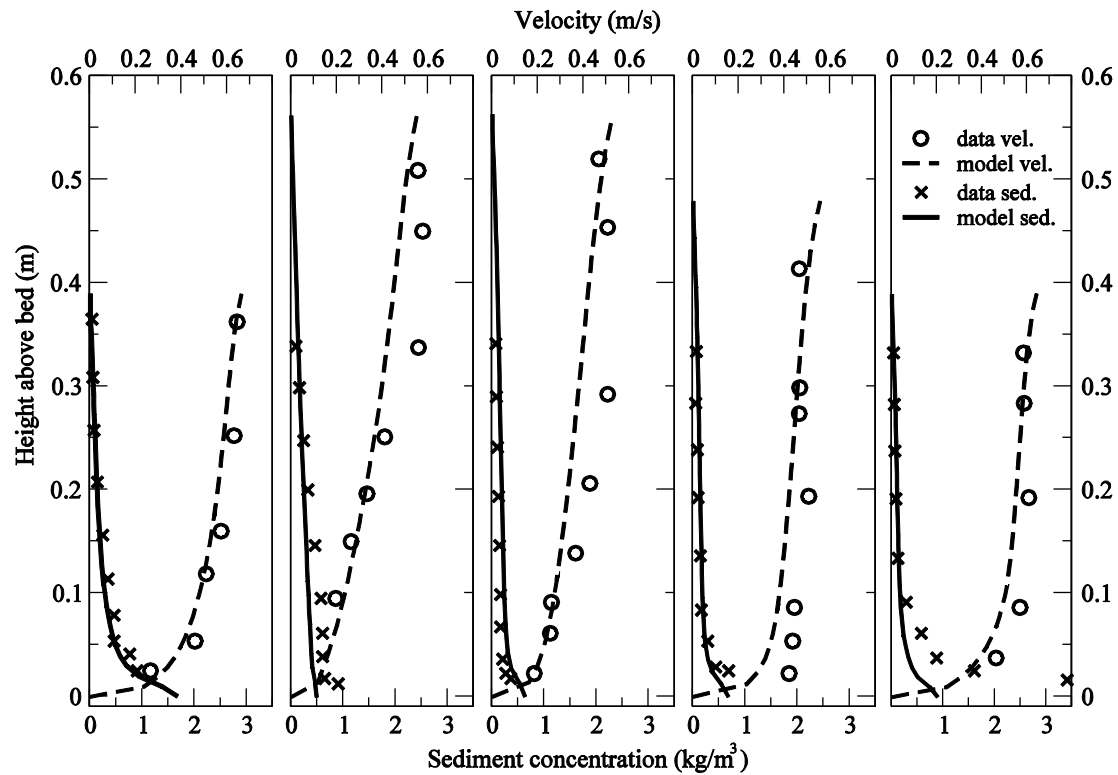


Figure 10 - Measured and simulated velocity and suspended sediment concentration profiles at the 5 monitoring stations at initial time ($t=0$).

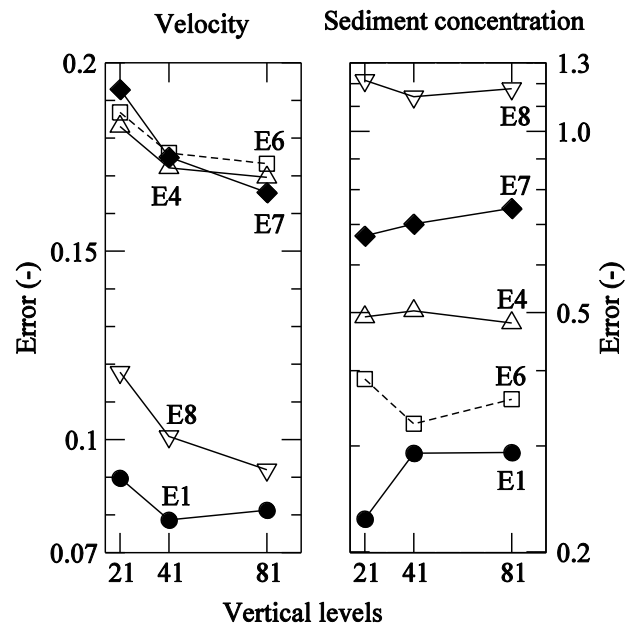


Figure 11 - Velocity and suspended sediment concentration errors for the different vertical grids simulated (21, 41 and 81 levels) and monitoring stations (E1, E4, E6, E7, E8).

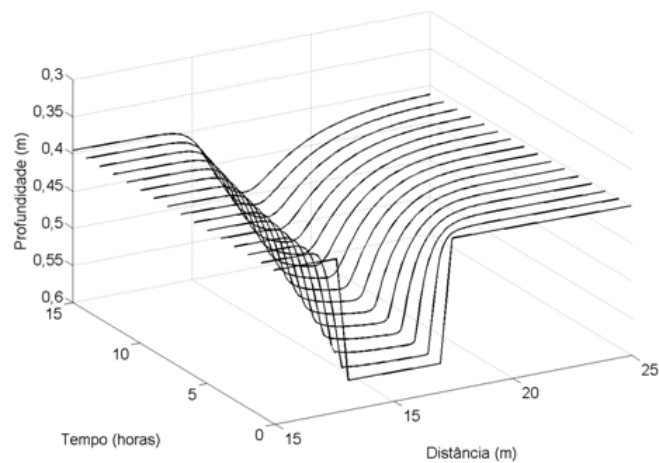


Figure 12 - Trench bottom evolution along the 15 hours of simulation.

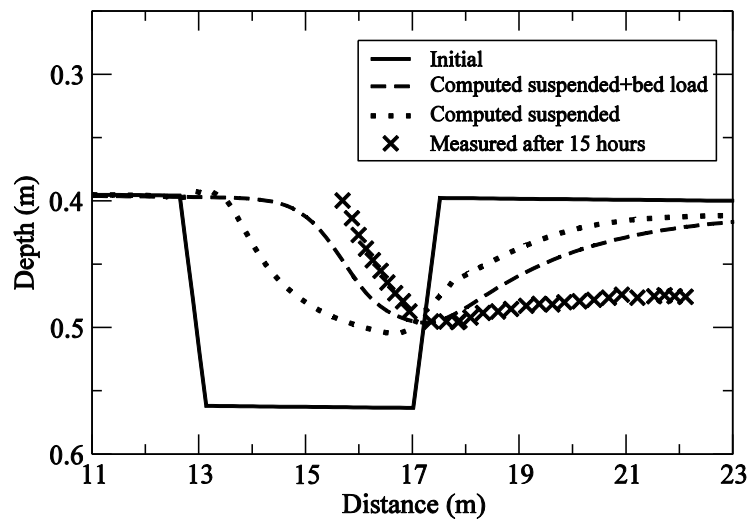


Figure 13 - Influence of the number of layers [where is this in the fig?] on computed bottom evolution.

Table 1 – Model parameters for test 2.

Parameter	Variable	Value
Grid spacing	$\Delta x, \Delta y, \Delta z$	80 m; 80 m; variable
Bottom roughness	z_o	0.0033 m
Settling velocity	w_s	0.02 ms^{-1} ; 0.097 ms^{-1}
Erosion rate	E_o	Variable
Reference height	a	0.01 H
Bed slope	S_o	$3.2 \times 10^{-4} \text{ m/m}$ (H5) $1.55 \times 10^{-4} \text{ m/m}$ (H10)

Table 2 - Model parameters for test case 2.

Parameter	Variable	Value
Grid spacing	$\Delta x, \Delta y, \Delta z$	100 m, 10 m, 21 s levels
Bottom roughness	z_o	0.00053 m
Settling velocity	w_s	0.001 ms^{-1}
Erosion rate	E_o	$5 \times 10^{-5} \text{ kgm}^{-2} \text{ s}^{-1}$
Critical stress	τ_{cr}	0.05 Nm^{-2}
Porosity	p	0.9
Bottom slope	S_o	$4 \times 10^{-5} \text{ m/m}$

Table 3 - Model parameters for test 4.

Parameter	Variable	Value
Grid spacing	$\Delta x, \Delta y, \Delta z$	0.12 m; 0.125 m; variable
Bottom roughness	z_0	0.000833 m
Settling velocity	w_s	0.011 ms^{-1}
Sediment diameter	d_{50}	140 μm
Erosion rate	E_0	$1.6 \times 10^{-2} \text{ kgm}^{-2}\text{s}^{-1}$
Critical stress	τ_{cr}	0.11 Nm^{-2}
Porosity	p	0.4
Bottom slope	S_0	$4 \times 10^{-4} \text{ m/m}$