Implementation and assessment of hydrodynamic, salt and heat transport models: The case of Ria de Aveiro Lagoon (Portugal)

J.M. Dias*, J.F. Lopes

Departamento de Física, Universidade de Aveiro, CESAM, 3810-193 Aveiro, Portugal

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Abstract

Ria de Aveiro is a shallow coastal lagoon located in the Northwest coast of Portugal. The implementation and the assessment (calibration and validation) of the hydrodynamic, salt and heat transport models for Ria de Aveiro lagoon is presented. During the calibration the models parameters were adjusted to give the best fit of the model results to the field data. The hydrodynamic model calibration was carried out by comparing measured and predicted time series of sea surface elevation for 22 stations. The root-mean square of the difference between those values was determined and harmonic analysis was performed in order to evaluate the model ability to reproduce the tide propagation along the main channels of Ria de Aveiro. Validation of the hydrodynamic model was performed comparing measured and predicted sea surface elevation values for 11 stations, as well as velocities in the main flow direction for 10 stations and water flows for 6 stations. The salt and the heat transport models were calibrated comparing measured and predicted time series of salinity and water temperature for 7 stations. These models were validated comparing the model results with an independent field data set. Results indicate that the hydrodynamic, the salt and the heat transport models for Ria de Aveiro were successfully calibrated and validated.

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1. Introduction

The mathematical modeling of physical processes in shallow waters has undergone a large development during the last decades. Due to the complexity of the processes involved, this development will pursue for a long time. However, it can be considered that some models are already in a mature stage, as it is the case of two-dimensional vertically integrated (2DH) models. Therefore, 2DH models can be used as reliable and useful tools to study shallow coastal waters (Neves, 1985; Cheng et al., 1993; Inoue and Wiseman Jr., 2000; Dias et al., 2003; Umgiesser et al., 2003).

A mathematical model is, by definition, an approximate reconstruction of a real phenomenon. The approximations and the parameterizations used for the synthesis of the model lead to discrepancies and deviations of the model results from nature. The optimization of the model operation is a complicated task and before a numerical model can be applied in a reliable study, the model must be implemented and should be verified, calibrated and validated. Only then the model can be used as a research tool for investigations of a given process. However, there is no widely accepted procedure for carrying out these tasks (Cheng et al., 1991). Model calibration and validation appear in various forms, dependent on data availability, characteristics of water body, and most of all, the
perceptions and opinions of the modelers (Hsu et al., 1999). Typically, the calibration is accomplished by qualitative comparison of short time series of predicted values with field data for the same location and period of time (Cheng et al., 1993). Although the root mean squared (rms) errors between the model results and the field data can be computed to measure the model performance, more often the calibration and validation procedures involve some degree of subjectivity. In fact, there is a large disadvantage on the direct comparison of rms errors, since phase errors and amplitude errors are considered together. For example, two dataset with no error in amplitude and a small error in phase can lead to a large r.m.s. With harmonic analysis it doesn’t happen. Additionally, this approach requires that the field data must be concurrent with the time period of model simulation. This requirement imposes severe limitations on using available field data for model calibration. Perhaps this approach is more useful in model validation than calibration.

In the case of modeling tidal hydrodynamic phenomena, another method to calibrate a numerical model is to compare the values of harmonic constants of the tides and tidal currents generated by the model with the respective harmonic constants calculated from field data. Although studies for other areas revealed that the harmonic constants may be not really independent of time (Smith, 1977), several authors in their numerical modeling calibration studies consider the harmonic constants as invariants characteristics of the astronomical tide at a specific site (Cheng et al., 1993; Martins et al., 2002). A long term analysis of the harmonic constants at Ria de Aveiro inlet revealed that meaningful changes are only observed at a scale of years (Tomás and Dias, 2004). Since the harmonic constants may be therefore considered independent of time, if reliable harmonic constants are available, concurrent field data are not required for model calibration and validation. Because in this case the model results and data are compared based on frequency domain properties, there is no time distinction between the model calibration and the model validation (Cheng et al., 1993). Therefore, the distinction must be made based on the time series location. The model parameters are adjusted in order to obtain the best fit between the harmonic constants of the model and that of the data. This method of model calibration is often more effective because concurrent data are usually scarce. A major limitation of frequency analysis is that large amount of data are needed. Therefore, this method requires the use of a robust and efficient numerical model to generate time-series records of at least 29 days (Cheng et al., 1993).

In this paper, two methods are used for the hydrodynamic model calibration. An initial qualitative calibration followed by a quantitative calibration. Direct comparison of scatter plots and root-mean square errors of model results and field data is used in this work to calibrate and validate the hydrodynamic model and the transport models of salt and heat. The second and more accurate method compares harmonic constants of observed and computed sea surface elevation data.

In some model studies, time series data of velocities, or their harmonic constants, are included in the hydrodynamic calibration process. However, in the case of this work there is sufficient sea surface elevation data available (Instituto Hidrográfico, 1991) to discern the spatial variation of the tidal cycle, and therefore the comparison of model calculated velocities with field data is treated as model validation (Hsu et al., 1999).

Although the comparison performed in the frequency domain may be also considered as hydrodynamic model validation, considering that the data used in the calibration are from 1987/88 and that the Ria de Aveiro is permanently subjected to bathymetric changes, all the models are validated through direct comparison of a more recent field data set.

The aim of this paper is to present the implementation of hydrodynamic, salt and heat transport models for Ria de Aveiro lagoon, as well as perform their calibration and validation for the area.

2. Study area

Ria de Aveiro (Fig. 1) is a shallow vertically homogeneous lagoon with a very complex geometry, located on the northwest coast of Portugal (40°38’N, 8°45’W). It is 45 km long and 10 km wide and covers an area of 83 km² at high tide (spring tide), which is reduced to 66 km² at low tide. It is characterized by narrow channels and by large areas of mud flats and salt marshes.

Ria de Aveiro is a mesotidal lagoon (Davies, 1964), and the tides, which are semidiurnal, are the main forcing action. The lagoon receives freshwater from two main rivers, Antuã (5 m³ s⁻¹ average flow) and Vouga (50 m³ s⁻¹) (Moreira et al., 1993; Dias et al., 1999). Boco river, at the southern end of Ilhavo channel, has a negligible flow, as well as the Caster and the Gonde rivers, discharging at the north end of S. Jacinto channel. There is another freshwater source at the southern end of Mira channel that consists in a small system of ponds and rivers, from which flow is not well known. During periods that last from a few hours to a few days, the wind, which is very significant in Aveiro, induces very important effects on the lagoon circulation. Extreme conditions of strong wind may induce particular circulation patterns mainly in shallow areas and wide channels (Dias, 2001). Wind residual circulation is directed along the main channels, S. Jacinto and Espinheiro, and tends to form a closed loop between these two channels (Dias et al., 2003).
The estimated maximum and minimum tidal prism of the lagoon is 136.7\!\times\!10^6 m^3 and 34.9\!\times\!10^6 m^3, respectively, for the maximum spring tide and the minimum neap tide (Dias, 2001). The total estimated freshwater input in a spring tide is very small (about 1.8\!\times\!10^6 m^3 during a tidal cycle) when compared with the mean tidal prism at the mouth (about 70\!\times\!10^6 m^3).

A previous hydrological characterization of Ria de Aveiro (Dias et al., 1999) leads to the conclusion that the lagoon can be considered as vertically homogeneous.

3. The numerical models

One of the objectives of this work is to implement a system of mathematical models able to simulate tidal flows and transport in Ria de Aveiro. Considering the Ria de Aveiro characteristics, a 2DH model was considered as the right choice to simulate the hydrodynamics and the transport of salt and heat (Dias, 2001; Dias et al., 2003).

3.1. Hydrodynamic model

A two-dimensional vertically integrated hydrodynamic model based on finite differences techniques was applied. This model was developed from the SIMSYS2D model (Leendertse and Gritton, 1971; Leendertse, 1987) and solves the second order partial differential equations for the depth-average fluid flow derived from the full three-dimensional Navier-Stokes equations. This results in a system consisting of an equation for the mass continuity and two horizontal momentum equations that are vertically averaged by integrating from the bottom to the surface:

\begin{align}
\frac{\partial \zeta}{\partial t} - \frac{\partial (HU)}{\partial x} + \frac{\partial (HV)}{\partial y} &= 0 \tag{1} \\
\frac{\partial U}{\partial t} + U \frac{\partial U}{\partial x} + V \frac{\partial U}{\partial y} &= -g \frac{\partial \zeta}{\partial x} + \frac{\tau^e_x - \tau^b_y}{H \rho} + A_h \nabla^2 U \tag{2} \\
\frac{\partial V}{\partial t} + U \frac{\partial V}{\partial x} + V \frac{\partial V}{\partial y} &= -g \frac{\partial \zeta}{\partial y} + \frac{\tau^e_y - \tau^b_x}{H \rho} + A_h \nabla^2 V \tag{3}
\end{align}

where $U$ and $V$ are the depth integrated velocity components in the $x$ (eastward) and $y$ (northward) directions, respectively, $\zeta$ is the surface water elevation, $H$ is the water height ($H = h + \zeta$; $h =$ water depth), $t$ is the time, $f$ is the Coriolis parameter, $g$ is the acceleration of gravity, $\rho$ is the water density, $A_h$ is the kinematic constant turbulent horizontal viscosity, $\tau^e_x$ and $\tau^e_y$ are, respectively, the magnitude of the wind stress on the water surface and the magnitude of the shear stress on the bottom.

The relationship between the wind stress and the surface wind may be expressed by the empirical formula first proposed by Ekman (Dronkers, 1964; Leendertse and Gritton, 1971):

\begin{align}
\tau^e_x &= \rho_o C_D W^2 \sin \alpha \quad \text{and} \quad \tau^e_y = \rho_o C_D W^2 \cos \alpha \tag{4}
\end{align}
where $W$ is the wind speed (generally measured 10 m above the sea surface), $\rho_a$ the density of the air, $\alpha$ the wind direction and $C_D$ an appropriate drag coefficient.

The bottom stress is assumed proportional to the square of the horizontal velocity (Dronkers, 1964; Leendertse and Gritton, 1971):

$$
\tau^h_x = \rho g \frac{U(U^2 + V^2)^{1/2}}{C^2} \quad \text{and} \quad \tau^h_y = \rho g \frac{V(U^2 + V^2)^{1/2}}{C^2} \tag{5}
$$

where $C$ is the Chézy coefficient. This coefficient, which is usually experimentally determined, depends on the bottom roughness, its composition and on the height of the water column. In the present work the Chézy coefficient is determined from the Manning roughness coefficient, $n$ (Chow, 1959):

$$
C = \sqrt[6]{H \over n} \tag{6}
$$

The system of Eqs. (1)–(3) was discretized using a finite difference method and the difference equations solved by the ADI method (Alternating Direction Implicit), using a space-staggered grid (Leendertse and Gritton, 1971; Dias, 2001).

With appropriate boundary and initial conditions, this system of equations constitutes a well-posed initial boundary value problem whose solution describes the depth-averaged circulation in a tidal basin. For barotropic models driven by tidal forcing the boundary conditions from experimental data include both the reflected and the incident waves, therefore and by its simplicity are adopted extrapolation formulas at open boundaries in this model. A dynamic water elevation at the ocean open boundary was imposed by using pre-determined tidal harmonics constituents. Constant current velocity was imposed at the rivers boundaries. The initial conditions were horizontal level and null velocity in all the grid points. Along the solid boundaries a null normal velocity was imposed and a free slip condition was assumed.

The model treats also shallow water flats in a mass consisting way. The shallow water flats during a tidal cycle are sometimes dry, and occasionally covered with water. During the dry periods the grid cells representing these areas are taken out of the algebraic system and are later added again, once the adjoining water level is higher than the water inside the dry grid cell. The specific implementation here adopted conserves the mass in each grid cell.

3.2. Salt and heat transport models

3.2.1. General equations

From the knowledge of the computed horizontal velocity field it is possible to determine the salt and the temperature distributions, solving the transport equations of salt and heat (Dias, 2001):

$$
\frac{\partial HS}{\partial t} + \frac{\partial (HUS)}{\partial x} + \frac{\partial (HVS)}{\partial y} = K_S H \nabla^2 S \tag{7}
$$

$$
\frac{\partial HT}{\partial t} + \frac{\partial (HUT)}{\partial x} + \frac{\partial (HVT)}{\partial y} = K_T H \nabla^2 T - \frac{1}{\rho C_p} F \tag{8}
$$

where $S$ and $T$ are the depth integrated salinity and water temperature, $K_S$ and $K_T$ are the horizontal turbulent salt and heat diffusivities, respectively, $F$ is a source term representing the heat flux through a horizontal surface and $C_p$ is the water specific heat at constant pressure.

The conservation of the transported constituent is essential in the advection-diffusion equations, where production or dissipation rates are small. In Eqs. (7) and (8) the advection term is written in the form of the divergence of the advection flow. In this case the advection terms may be approximated by finite-differences in a conservative form (Roache, 1985). The model uses the Flux Corrected Transport algorithm (FCT) in order to discretize the advection term (Leonard, 1979).

Precipitation and evaporation are not considered in the establishment of the salt transport equations. It is also assumed in these models that the mass flows through the water surface and through the bottom are null.

Initial conditions for the transport models are salinity and temperature fields obtained by interpolation of available data. Salinity and temperature condition for the ocean open boundary corresponds to variable ocean salinity and temperature in each time step of the computation. Different boundary river conditions were imposed for the diverse cases studied in this work, according to the environmental conditions. The atmospheric parameters needed to compute the latent and sensible heat fluxes, as well as the radiation fluxes (relative humidity, air temperature, wind speed and cloudiness) were supplied by the meteorological station of the Aveiro University and are specified in each time step of the computation.

3.2.2. Heat and radiative fluxes parameterizations

The parameterization adopted for the heat and radiative fluxes has been developed from the heat budget equation (Hsiung, 1985). The source term in Eq. (8) is obtained from an energetic balance at the water-atmosphere interface:

$$
F = Q_s - (Q_h + Q_w + Q_r) \tag{9}
$$

where $F$ is the heat flux through a horizontal surface, $Q_s$ is the short wave radiation, $Q_h$ is the long wave
planetary radiation, \(Q_h\) is the sensible heat flux and \(Q_e\) is the latent heat flux.

The conductive heat exchange with the underlying sediments is not considered in this work, because it is assumed negligible compared with the other terms. The sensible, \(Q_s\), and latent, \(Q_e\), heat flux are calculated using the Bulk Aerodynamic type formulas, as suggested by Priestley and Taylor (1972):

\[
Q_h = C_p \rho C_w (T_s - T_n)
\]

\[
Q_e = L_v \rho C_w (q_s - q_n)
\]

where \(T_n\) is the air temperature measured at 10 m above the surface, \(T_s\) is the water surface temperature, \(q_s\) is the air specific humidity and \(q_n\) is the saturated specific humidity corresponding to \(T_s\). \(L_v\) is the latent heat of evaporation and \(C_p\) and \(C_w\) are, respectively, the latent and the sensible heat flux coefficients. The sensible heat flux is related to the water-air temperature difference, whereas the latent heat flux is related to the vapor pressure deficit. Air in direct contact with the surface of the lagoon is assumed saturated and at the temperature of the water. In the present work the values \(C_p = C_w = 1.3 \times 10^{-3}\), adopted by Sultan and Ahmad (1994), were used, which are based on monthly averages of water-air temperature differences and wind speeds.

The outgoing net long wave radiation, \(Q_{b0}\), is estimated from the Stefan-Boltzmann equation using a formulation given by Budyko (1974), including corrections for back radiation to the atmosphere (Swinbank, 1963) and for cloud cover (Reed, 1976):

\[
Q_b = \varepsilon \sigma T_s^4 \left( 0.39 - 0.05 \varepsilon^{1/2} \right) (1 - S_a N^2) + 4 \varepsilon \sigma T_s^4 (T_s - T_w)
\]

where \(N\) is the cloud cover expressed in fractions of ten, \(\varepsilon\) is the emissivity of the water surface relative to a black body, \(\sigma\) is the Stefan-Boltzmann constant and \(S_a\) is a function of latitude, varying from 1.0 at the poles to 0.5 at the equator (used \(S_a = 0.8\)).

The incoming solar radiation, \(Q_i\), is calculated using (Reed, 1977):

\[
Q_i = Q_o (1 - 0.6N + 0.0019\alpha)(1 - A)
\]

where, \(A\) is the oceanic albedo, \(Q_o\) is the radiation received on the surface and \(\alpha\) is the solar noon altitude (in degrees).

4. Numerical bathymetry

Bathymetry is probably the most important among many factors that affect the flow properties in shallow systems like Ria de Aveiro. Previous modeling experience indicates that bathymetry controls the spatial variability of current magnitude and direction, constituting a factor that assures the reality of the numerical model. Thus, an accurate bathymetric representation is one of the most important and fundamental requirements in successful modeling (Cheng et al., 1991). This is particularly true for the Ria de Aveiro lagoon, where the bathymetry is very complex. Furthermore, the grid must accurately represent the bathymetric characteristics of the lagoon and guarantee model stability. The task of defining the grid-depth relations for the computational grid points can be extremely time consuming and tedious (Burau and Cheng, 1989).

The flow simulation in this complex domain requires the use of the most refined grid. The model grid must be sufficiently refined to resolve the essential features of the depth and the geometry variations, but as the grid resolution is refined the total number of grid points and the computation time increase geometrically. For this domain the ideal cell dimension would be around 50 m, but the compromise solution was to develop a grid with dimensions \(\Delta x = \Delta y = 100\ m\), resulting in 160 cells in the \(x\)-direction (eastward) and 393 cells in the \(y\)-direction (northward) (Fig. 1). In this case the narrower channels had its width exaggerated, but lowering their depth it was guaranteed that they maintained their water volume.

The numerical bathymetry used in this study was developed from data concerning depth obtained from a general survey carried out in 1987/88 by the Hydrographic Institute of Portuguese Navy (IH). This database comprises a collection of a large number of water-depth data (93269), uniformly distributed in space, including the areas until 2.5 m above the hydrographic null. Changes in the bathymetry may be expected due to sediment dynamics and to the dredging operations performed since 1988.

The water depth at each grid point was determined from the water volume of the cell, calculated using a Monte Carlo cubature method (Dias, 2001).

5. Models calibration

A large number of sensitivity tests were initially carried out in order to analyze the hydrodynamic and transport models sensitivity to the variations in the initialization parameters (time step, initial water level, horizontal viscosity coefficient, and horizontal eddy salt and heat diffusivities). The comparison between model results and analytical and typical solutions were presented in Dias (1993, 2001), where the validity of the numerical solution of these models was proved, resulting in the use of the values of \(\Delta t = 40\ s\), \(A_s = 20\ m^2\ s^{-1}\) and \(K_s = K_A = 5.4\ m^2\ s^{-1}\).

The model calibration is defined as an operation in which specific values, or distributions, or a range of
variation are given to the floating free model parameters, so that the model results fit best to a set of field observations (Koutitas, 1994). During this operation, the model sensitivity to the variation of each free parameter is analyzed by, either formal sensitivity studies, or a series of numerical experiments. The model calibration is subsequently based on the determination and the adjustment of the parameters to which the model is most sensitive.

5.1. Hydrodynamic model

5.1.1. Response to tidal forcing

The hydrodynamic model calibration was carried out with sea surface elevation (SSE) data measured in 1987/88 at 22 stations (Fig. 1) distributed throughout Ria de Aveiro (Instituto Hidrográfico, 1991).

From the observation of the SSE data it is concluded that friction dissipates energy as the tidal wave propagates landward from the lagoon mouth. This behavior is a common feature in estuarine environments (Hsu et al., 1999) and was analyzed and quantified in Ria de Aveiro by Dias and Fernandes (2005). Considering the lagoon geometry and the tidal range at the mouth, the magnitude of the bottom friction coefficient determines the tidal range variation along the lagoon channels. The remaining parameter subjected to adjustment during model calibration is therefore the bottom friction in the Manning-Chézy formulation for the bottom stresses (Eq. (5)). Previous modeling experiments (Burau and Cheng, 1988) as well as results of data analysis (Cheng and Gartner, 1985) suggests a strong influence of the water depth in the bottom stresses than the Manning-Chézy relation does. This effect can be introduced into the computations by allowing Manning’s coefficient $n$ to vary as a function of water depth (Cheng et al., 1993). In this model, as a first approach the Manning’s $n$ values are assigned to a range of water depth rather than assigning them to every point. The relation adopted between $h$ and $n$ was based in the values presented by Cheng et al. (1993) for San Francisco Bay and in the assumption that the flow must be damped in the intertidal areas of Ria de Aveiro (Table 1). The model results are not very sensitive to the absolute values of $n$, but the introduction of depth dependent $n$ is necessary to achieve the best fit to the available field data. Fernandes et al. (2001) observed a similar pattern on their modeling studies of Patos Lagoon.

As the bottom roughness also depends on the type of the bottom sediments, the Manning’s $n$ values were then locally adjusted until the model outputs agreed satisfactorily with the field data. In this procedure it was assumed that an increase in the friction coefficient will lower the tidal ranges both in that zone of the channel and its upstream. Another assumption for the fine tune calibration is that an increase of the friction coefficient will increase the phase lag for high tide and decrease for low tide (Fry and Aubrey, 1990).

Table 1

<table>
<thead>
<tr>
<th>Bottom friction coefficients</th>
<th>Manning’s $n$ value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water depth (m)</td>
<td></td>
</tr>
<tr>
<td>$-2.5 \leq h &lt; -2.0$</td>
<td>0.042</td>
</tr>
<tr>
<td>$-2.0 \leq h &lt; -1.5$</td>
<td>0.038</td>
</tr>
<tr>
<td>$-1.5 \leq h &lt; -1.0$</td>
<td>0.034</td>
</tr>
<tr>
<td>$-1.0 \leq h &lt; -0.5$</td>
<td>0.030</td>
</tr>
<tr>
<td>$-0.5 \leq h &lt; 0.0$</td>
<td>0.027</td>
</tr>
<tr>
<td>$0.0 \leq h &lt; 0.5$</td>
<td>0.024</td>
</tr>
<tr>
<td>$0.5 \leq h &lt; 1.0$</td>
<td>0.022</td>
</tr>
<tr>
<td>$1.0 \leq h &lt; 3.0$</td>
<td>0.020</td>
</tr>
<tr>
<td>$3.0 \leq h &lt; 10.0$</td>
<td>0.018</td>
</tr>
<tr>
<td>$h \geq 10.0$</td>
<td>0.015</td>
</tr>
</tbody>
</table>

Fig. 2 shows the comparison between the simulated and observed SSE time series for 6 stations used in the model calibration (the calibration procedure involves 22 stations). The evaluation of the relation between measurements and predictions was carried out using the root-mean square ($RMS$) of the difference between the observed and predicted surface elevation:

$$RMS = \left\{ \frac{1}{N} \sum_{i=1}^{N} \left[ \zeta_{o}(t) - \zeta_{m}(t) \right]^2 \right\}^{1/2}$$

where $\zeta_{o}(t)$ and $\zeta_{m}(t)$ are the observed and the modeled SSE, respectively, and $N$ is the number of measurements in the time series. The $RMS$ values were computed for each station and are presented in each plot.

In general there is a good agreement between the predicted and the observed SSE for all the stations (Fig. 2), revealing the model’s ability to reproduce nature. However, there are some features that must be addressed. It is necessary to mention that the agreement between the predicted and the observed values at station A (mouth of the lagoon – not shown) is not perfect, as would be expected. In fact, the tide synthesized for station A was specified at the ocean open boundary. However, the open boundary is located 2700 m westward of station A, and a delay of about 3 min is estimated for the tide propagation between the two locations. This delay associated with the natural difference between the observed and synthesized SSE, lead to a root-mean square error of 9.6 cm for station A, which is about 5% of the mean tidal range at the mouth. This error may explain the following errors found for almost all stations. In fact, for stations located at the S. Jacinto and the Mira channels and at the beginning of Ilihavo channel the $RMS$ errors are around 5% of the local tidal ranges.

The highest disagreement between predicted and observed sea surface elevation was found for stations F, G and H at the Ilihavo channel, and for stations U
and V, with the root-mean square errors of about 10% of the local tidal range. The large errors found for stations located at Ilhavo channel may be explained by some inaccurate definition of the bathymetry at specific areas of this channel. In fact, between stations H and I there is a strong constriction in the channel, with a width of about 10 m and where very strong currents occur. This region is very difficult to represent in the numerical bathymetry using a cell 100 m wide, and therefore it was found impossible to improve the agreement between the predicted and the observed surface elevation for the stations located upstream of this constriction. No justification was found for the errors determined for stations U and V, and it was also not possible to improve the agreement between the predicted and the observed data in these stations without deteriorating the results for the other stations. However, the root-mean square errors determined for all the other stations proved that the influence of these results is minimal in the overall lagoon hydrodynamic.

Kuo and Park (1985) demonstrated that proper simulation of the times for high and low tides are important to evaluate the proper modeling of tidal fluxes in domains with large intertidal areas. If the side storage area is not properly accounted for, the model can reproduce the mean tidal range but the phase cannot be properly simulated. The analysis of Fig. 2 reveals that the time lag between model results and observations is low in almost all stations. This time lag will be quantified through harmonic analysis.

Harmonic analysis (Foreman, 1977; Foreman and Henry, 1989) was performed on 29 days length time series of observed and predicted SSE for all the stations. Results for six of the major tidal constituents ($M_2$ - 12.42 h; $S_2$ - 12 h; $N_2$ - 12.9 h; $P_1$ - 24.07 h; $K_1$ - 23.93 h; $O_1$ - 25.82 h) determined for 7 of the stations are presented in Table 2. The difference between the calculated harmonic constants for the observed and the predicted data is also presented in this table. The agreement between the time series is rather good both in amplitude and in phase for the semidiurnal constituents, which are the major tidal constituents in the Ria de Aveiro. For the $M_2$ constituent, which has the highest amplitude, the mean difference of the amplitudes is about 7 cm, with an error of 1.2 cm. The mean phase difference is 4°, with an error of 2°. Results for the other semidiurnal constituents as well as for the diurnal ones, are not so accurate, but still reveal a good agreement between the predicted and observed constants. The comparison between these values reveals that the amplitude of the major constituents may be considered well represented by the numerical model for the entire lagoon, with average differences lower than 4 cm. The predicted phase also represents well the tide propagation for the $S_2$ and $N_2$ constituents for all the stations. For the diurnal constituents the amplitude agreement may be considered good for all the stations and the phase agreement may also be considered good, except for stations F, G, H and J. Upstream of the constriction in Ilhavo channel (Fig. 1) the predicted phases for the diurnal constituents are not reliable, due to the difficulty in correctly represent the constriction in the numerical bathymetry as stated before.

5.1.2. Response to the wind forcing

As previously stated in Section 2, the wind may be very important in the forcing of Ria de Aveiro hydrodynamics during short periods. The calibration of the model response to the wind data against an observed data set is almost impossible, requiring the concurrent measurement of wind and SSE data during episodes of strong wind. This difficulty is increased in
the case of Ria de Aveiro due to the wind characteristics in this region, which direction and intensity are highly variable during short periods. Therefore, the model response to the wind forcing was optimized through an accurate parameterization of the wind drag coefficient through comparison between model results for ideal channels and results from an analytical model.

The tangential wind stress is dependent on the wind intensity and the drag coefficient (Eq. (4)). The value of the drag coefficient, $C_D$, depends on the height at which the wind speed, $W$, is measured, on the stability of the lowest few meters of the atmosphere and on the roughness of the water surface (generated by waves) (Bowden, 1983). As the value of $C_D$ also depends on $W$ itself, the dependence of the tangential wind stress on the wind speed is not strictly quadratic.

For a neutral atmospheric boundary layer, Hicks (1972) found that the water surface dimension was an insignificant factor and that small lakes have drag coefficient similar to those found in the ocean. Smith et al. (1992) obtained several formulations for the drag coefficient as a function of the wind intensity under different conditions (HEXOS program). An average of the coefficients obtained in the different formulations may be used as a first approach to determine the drag coefficient:

$$10^3 C_D = 0.5643 + 0.0873 W$$  (15)

In order to calibrate the response of the hydrodynamic model to the wind forcing, several simulations were performed and the numerical results were compared with results from an analytical model. With this purpose were

<table>
<thead>
<tr>
<th>Station</th>
<th>B</th>
<th>C</th>
<th>G</th>
<th>I</th>
<th>L</th>
<th>O</th>
<th>V</th>
</tr>
</thead>
<tbody>
<tr>
<td>$M_2$ Amplitude (m)</td>
<td>0.890</td>
<td>0.491</td>
<td>0.505</td>
<td>0.840</td>
<td>0.782</td>
<td>0.340</td>
<td>0.768</td>
</tr>
<tr>
<td>Model</td>
<td>0.971</td>
<td>0.586</td>
<td>0.610</td>
<td>0.904</td>
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defined several channels with dimensions similar to that of Ria de Aveiro. This comparison was performed for 14 rectangular flat bottom channels, which were 500 m wide and with depths between 1 and 6 meters (with an interval of 1 meter). Two different cases with 7 experiments each were defined, considering the length of the test channels as 7 and 11 km for each case, respectively. For these simulations the model was only forced with a stationary strong wind (20 m/s of intensity) blowing along the main channel axis (x direction).

The analytical model used in this case is based on the shallow water equations, which are simplified in order to represent the balance between the pressure gradient and the tangential wind stress (Bowden, 1983). The transversal components can be neglected due to the wind direction, as well as the horizontal acceleration and the Coriolis term, which are very small in these circumstances. According to Bowden (1983), in these conditions the bottom stress lies between zero and \( \tau_z/2 \), so that the analytical model equation is:

\[
\frac{\partial \xi}{\partial x} = C_t \frac{\tau_z^*}{\rho g (h + \xi)}
\]  

(16)

where \( h \) is the water depth and \( C_t \) is a constant between 1 and 1.5. Eq. (16) has the following solution:

\[
\Delta \xi = -h \pm \sqrt{h^2 + \frac{2 C_t \tau_z^* L}{\rho g}}
\]  

(17)

where \( \Delta \xi \) is the water elevation difference between the channel extremities and \( L \) is the channel length.

Fig. 3 compares the numerically computed SSE for each test channel and for each one of the 7 depths considered with the correspondent analytical model results determined from Eq. (17), for extreme values of the constant \( C_t \).

According to Merzi and Graf (1985) and Johnson and Vested (1992) the wind drag coefficients at the shallow depths, as the ones found in Ria de Aveiro, are higher than those obtained in deep water for the same wind intensity. From the numerical experiments it was found that a drag coefficient 15% higher than the one determined by Eq. (15) leads to a very good agreement between the numerical and the analytical models (Fig. 3). Therefore, the following equation is used to compute the wind drag coefficient:

\[
10^3 C_D = 0.6489 + 0.1004 W
\]  

(18)

5.2. Salt and heat transport models

In most lagoons and estuaries with a significant freshwater discharge, salinity may serve as an ideal natural tracer for calibration of transport processes. In these environments the tidal current, the freshwater discharge, the density circulation, as well as the turbulent mixing processes affect salinity distribution. Therefore, the salinity distribution reflects the combined results of all these processes, and in turn it controls the density circulation and modifies the transport processes. Assuming that the barotropic flows (tidal and freshwater flows) have been calibrated and validated, the procedure to calibrate the salinity transport model is to match the observed and the computed salinity time series. In this calibration procedure, the analysis of predicted distributions of salinity is used to guide the adjustment of calibration constants through comparison with the typical horizontal salinity distribution observed in Ria de Aveiro.

Once the salinity transport model is considered calibrated, the transport processes may be considered well represented by the model. Therefore, the calibration of the heat transport model is related only to the heat and radiative fluxes parameterization.

A set of salinity and water temperature data measured between 9/7/96 and 28/7/96 is available for comparison with model results. These data include long time series of salinity and water temperature measured each 10 minutes at the mouth of the lagoon (station A). It also includes 25 hours length time series of hourly measured values at stations B, H, I, L, M, O and V. These data were measured after a long period without precipitation, a typical situation of the Summer season, where the freshwater inputs from the rivers are expected to be low.

The long time series of the salinity and the water temperature were imposed at the ocean open boundary. The freshwater inflows through the upstream boundaries are not known, and in this study are used as a...
The imposed flowrate at the rivers boundaries resulting from the calibration procedure were the following: Vouga river – 2 m$^3$ s$^{-1}$; Antuá river – 0.5 m$^3$ s$^{-1}$; Boco river – 0.1 m$^3$ s$^{-1}$; Mira – 0.2 m$^3$ s$^{-1}$; Caster river - 0.08 m$^3$ s$^{-1}$; Gonde river – 0.01 m$^3$ s$^{-1}$.

The freshwater was specified with salinity equal to 0, freshwater temperature for the Vouga and Antuá rivers as 23°C, and the remaining freshwater temperatures as 25°C. In the calibration of the heat transport model the sensible and the latent heat fluxes coefficients are also used as calibration parameters and were considered constant in all the simulations. It was found that the agreement between the predicted and the observed temperature is good when using the values determined by Sultan and Ahmad (1994) ($C_t = C_e = 1.3 \times 10^{-3}$).

Figs. 4 and 5 show the comparison between the predicted and observed salinity and water temperature, respectively, for 6 stations. The patterns observed are essentially dependent on the tidal transport (Dias et al., 1999). The RMS values were computed for each station, and are presented at each plot. The agreement between the predicted and the observed salinity may be considered good for all the stations, the salinity time evolution and amplitude variation being well represented by the model. The maximum absolute RMS value was determined for station O, with a value of 0.64, which represents about 10% of the local salinity amplitude. The RMS values for the other stations are also around 10% of the local salinity amplitude. In general there is a good agreement between the predicted and the observed water temperature values.

Figs. 4 and 5 show the comparison between the predicted and observed salinity and water temperature, respectively, for 6 stations. The patterns observed are essentially dependent on the tidal transport (Dias et al., 1999). The RMS values were computed for each station, and are presented at each plot. The agreement between the predicted and the observed salinity may be considered good for all the stations, the salinity time evolution and amplitude variation being well represented by the model. The maximum absolute RMS value was determined for station O, with a value of 0.64, which represents about 10% of the local salinity amplitude. The RMS values for the other stations are also around 10% of the local salinity amplitude. In general there is a good agreement between the predicted and the observed water temperature values.

<table>
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<td>15 20 1 6 11</td>
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<td>23/7/96</td>
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Fig. 4. Time series of salinity for stations B, H, I, L, O and V, used in the salt transport model calibration (● data; — model).

Fig. 5. Time series of water temperature for stations B, H, I, L, O and V, used in the heat transport model calibration (● data; — model).
(Fig. 5). The RMS values are typically about 5% of the local water temperature amplitude.

According to these results it may be considered that the transport processes in the Ria de Aveiro are well represented by the transport models. The heat transfer between the atmosphere and the water surface may also be considered well represented in the heat transport model.

6. Models validation

The model validation is defined as a procedure consisting in comparing the model output with available field or laboratory data to prove the model efficiency. The measurements used for the validation have to be independent from the set used for calibration.

6.1. Hydrodynamic model

When the harmonic constants are compared in the model calibration, the net result also represents properties of model validation (Cheng et al., 1993). Therefore, it is considered that in these cases it is not necessary to validate the model results by comparing them to an independent field data set. However, considering that the data used in the calibration of this model are from 1987/88 (in a large temporal scale the harmonic constants have slightly changed at the inlet (Tomás and Dias, 2004)) and that the Ria de Aveiro is a system with a highly dynamic morphology, the authors consider essential to validate the model against recent data. Thus, to complement the test of model validation a very wet period in June 1997 was simulated. These conditions were used to investigate the model’s response to the interaction of tidal forcing and varying river discharge.

Under such conditions, the vertically well-mixed assumption may be violated close to the rivers mouth.

The hydrodynamic model validation was carried out with data measured in June 1997. Measurements of current velocity at stations B, C, F, G, H, I, L, M, O and V and of SSE at the same stations plus station D were compared with model predictions. The simulation was performed without changing the values of the friction coefficients adjusted during the calibration. The ocean boundary condition corresponds to the tide synthesized for the considered period of measurements. The freshwater discharges imposed at the rivers boundaries were typical values for this season: Vouga river – 450 m$^3$ s$^{-1}$; Anúu river – 46 m$^3$ s$^{-1}$; Boco river – 2.5 m$^3$ s$^{-1}$; Mira – 42 m$^3$ s$^{-1}$; Caster river – 2.0 m$^3$ s$^{-1}$; Gonde river – 1.2 m$^3$ s$^{-1}$. The remaining input variables were left unchanged.

Unlike surface elevation, current velocity varies rapidly in space both in magnitude and direction, from point to point, reflecting the irregular lagoon geometry, producing intrinsic differences between the model results and the velocity field data. The model results are related to the mean value over the vertical and over a horizontal spatial domain corresponding to the grid size, while the field data is the value at a single point or, at best, an average over several points.

In order to easily compare the velocities, the main flow directions of the observed and predicted current velocities were determined at each station and it was verified that they are almost coincident. The predicted and observed current velocities were projected along these directions.

The comparison between the predicted and the observed SSE and along flow direction velocities for 6 of the stations is plotted in Figs. 6 and 7, respectively. The RMS of the SSE values is around 5% of the local

![Fig. 6. Time series of SSE for stations C, G, I, L, O and V, used in the hydrodynamic model validation (● data; — model).](image-url)
tidal range for all the stations. The accord between the values is rather good, revealing that a general agreement can be observed between model predictions and field measurements.

From the analysis of Fig. 7 it is verified that there are some discrepancies between the model and the field data when assessing the velocities point-by-point. The RMS values are very high, ranging from about 7% of the current amplitude in the case of the better adjustment (Station V) to 35% in the worse case (Station G). However, the model does properly simulate the temporal variation of the velocity in terms of the current phase. This comparison also reveals that the agreement between the predicted and the observed values is rather good for stations L, O and V. These stations are located at the center of wide channel sections, and therefore it can be considered that the observed velocities are at least representative of the current in an area correspondent to the grid cells. Therefore they may be directly compared with the model results obtained for the equivalent grid cells. The other stations are located at narrow sections, and therefore there are intrinsic differences between the predicted and the observed values of the currents. In these narrow sections, from the observed and predicted SSE and current velocity values, the water flows were estimated over the cross-section that includes the station location. Fig. 8 shows the estimated water flows for 6 stations. The predicted and the observed values are in agreement as shown by the low RMS values relative to the flow amplitude, revealing that the

![Fig. 7. Time series of along flow direction velocities SSE for stations C, G, I, L, O and V, used in the hydrodynamic model validation (● data; — model).](image)

![Fig. 8. Time series of water flow SSE for stations C, F, G, H, I and O, used in the hydrodynamic model validation (● data; — model).](image)
model correctly simulates both temporal evolution and magnitude of the water flows in the narrow channels. Based on the results, it may be considered that the hydrodynamic numerical model for the Ria de Aveiro has been successfully validated.

6.2. Salt and heat transport models

The salt and the heat transport models were validated comparing the model results with an independent data set. The salinity and water temperature values were measured in June 1997 at the same 11 stations referred in the hydrodynamic model validation. These data correspond to a very wet period and a situation of high rivers runoff. Therefore, the transport models are validated in very different conditions from that observed during their calibration.

The salinity and water temperature values measured every 10 minutes at the mouth of the lagoon were used as model inputs at the ocean open boundary. At the rivers boundaries were imposed freshwater discharges typical of situations of high rivers runoff (the same specified at Section 6.1) and the water temperature was specified as equal to that used for the calibration. The remaining input variables were left unchanged.

The comparison between the predicted and observed salinity and water temperature for 6 of the stations is plotted in Figs. 9 and 10, respectively, and reveals...
a good agreement between the model results and the observed values. The salinity RMS values range from around 4% (Station C) to 9% (Station B) of the local salinity amplitude. The water temperature RMS values are slightly higher than those found during calibration procedure, ranging from about 7% (Stations B and C) to about 17% (Station I) of the local temperature amplitude.

From these results the salt and the heat transport models may be considered validated, reproducing well the transport and the heat transfer processes in Ria de Aveiro, even in situations of high rivers freshwater input.

7. Conclusions

The hydrodynamic model and the transport models of heat and salt were successfully implemented for the Ria de Aveiro lagoon. Results show that the calibration of the models was successfully carried out, revealing a good agreement between measurements and predictions. The validation tests showed that the models can reproduce an independent observed data set.

The developed and applied models to the Ria de Aveiro seem to make use of an adequate bathymetry. However, differences do exist, and they might be the result of several factors, including: inaccurate definition of the bathymetry in the model for that region; very narrow channels not well resolved by the model horizontal grid; and uncertainties in the field data.

The results show that the models can accurately reproduce the barotropic flows and simulate adequately the salt and heat transport in Ria de Aveiro, even in conditions of high freshwater inputs from the rivers. The models can, therefore, be considered as a new important tool for future studies of the Ria de Aveiro hydrodynamics and water quality.

The procedure presented here for Ria de Aveiro takes modelers through the basic steps involved in the models implementation and evaluation, and can therefore be used as a guideline for similar studies.

Acknowledgements

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References


