Non-Steady Effects in Sand Transport

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ABSTRACT


It has been recognized, through recent laboratory experiments in both rippled and plane bed conditions, that sediment reacts to the flow in a complex manner, involving non-steady processes resulting from memory and settling/entrainment delay effects. The main goal of the present work is to highlight in which field conditions the non-steady effects become important in the quantification of sediment transport in the near shore zone. For that purpose, we have parameterized the non-steady effects due to a settling delay of sediments particles in the form of a function that can be easily computed in terms of known hydrodynamic parameters and sediment characteristics. The evaluation of this function for a range of hydrodynamic conditions representative of the coastal zone revealed that non-steady effects became important for increasing values of the waves orbital velocity and waves asymmetry, for decreasing values of the waves period, in rippled beds and for fine sediment particles. The importance of non-steady effects in the development and evolution of a cross-shore sand bar is also addressed.

ADDITIONAL INDEX WORDS: Sediment transport models, phase lags, cross-shore beach profile.

INTRODUCTION

For the general purposes of morphodynamic computations in the coastal zone, modelers usually employ simple formula-based models to evaluate sediment transport. Sediment transport is computed as a function of the bottom shear stress or the near-bed flow velocity and it is assumed that the sediment particles react immediately to changes in flow conditions (quasi-steady hypothesis). Recently, laboratory experiments performed in oscillating water tunnels (OWT), in plane bed (sheet flow) and rippled bed regimes, have shown that in some conditions the sediment reacts to the flow in a much more complex manner, involving non-steady processes. These processes might be important in the cross-shore direction, where sediment transport is mainly caused by the oscillatory motions induced by surface short gravity waves. The presence of unsteady effects in sediment transport is evident in WATANABE and ISOBE (1990) rippled bed experiments. They show that, in some conditions, oscillatory flow over a rippled bed produces a net sediment transport in the opposite direction of the mean current. These effects are due to the vortices formed over a ripple that retain the sediments entrained from the bed. RIBBERINK and CHEN (1993) and DOHMEN-JANSEN et al. (2002) have shown that even in sheet flow plane bed conditions the unsteady effects in sediment transport are evident for fine sands. The experiments of RIBBERINK and CHEN (1993) considered regular 2nd order Stokes waves with a median grain sediment size, d, equal to 0.013 cm. It was observed that the net sediment transport during the wave cycle was in the opposite direction of the waves for values of the root mean square value of the orbital velocity greater than 0.7 m/s. The experiments of DOHMEN-JANSEN et al. (2002), considered sinus waves and collinear currents with a 0.013 cm median grain diameter, show that the net sediment transport during the wave cycle decreased with a decreasing in the waves period and that the net transport rates obtained with this fine sediment were, in the same hydrodynamic conditions, inferior to the ones obtained with courser sediments (d, = 0.021 and 0.032 cm). Moreover, DIBAJNIA and WATANABE (1992) have examined the validity of the quasi-steady transport model of GRANT and MADSEN (1976) in predicting the net transport rates measured in a series of experiments conduct under asymmetric oscillations with short wave periods and superimposed currents with a median grain size of 0.02 cm. It was found that this method fails to predict the magnitude and direction of the net transport rates measured because the transports at successive half cycles of oscillations are not independent from each other.

The non-steady effects in sediment transport result from memory and settling/entrainment processes and traduce themselves as phase-lags between the sediments concentration and the near bed flow velocities. This mechanism can be explained as follows: if the sediments that are entrained from the bottom at high shear stress remain in suspension at flow reversal then they will be transported by the flow in the succeeding wave half cycle in the opposite direction. On the other hand, the entrainment of sediment particles into the flow is not an instantaneous process in phase with the velocity field. Both these mechanisms cause a reduction of net sediment transport.

The main goal of the present work is to highlight in which field conditions the non-steady effects, as described above, become important in the quantification of sediment transport in the near shore zone. For that purpose, we have parameterized the non-steady effects due to a settling delay of sediments particles in the form of a simple function. The different parameters that intervene in this function are the waves period, T, the median grain size, d, the root mean square value of the near bed orbital velocity, u*, the magnitude of a steady current, U, and the 'horizontal' asymmetry of the waves velocity profile, r. Then, it is possible to establish the conditions upon which the values of these parameters lead to important phase lag effects and quantify the resultant sediment transport. This work has been based in an improved version of DIBAJNIA and WATANABE (1992) sediment transport formula, presented by SILVA et al. (2001). This transport model will be presented in detail in section 3.

SEDIMENT TRANSPORT MODELS

Different model concepts are presently in use for the prediction of sediment transport in the coastal zone, i.e., in wave or combined wavecurrent flow conditions. These range from the bottom boundary layer models to empirical or theoretical transport formulas.

The bottom boundary layer models are complex mathematical models involving high order turbulence closure schemes that solve numerically the momentum equations and the sediments balance equation in the wave or combined wave-
current bottom boundary layer. They describe the intra-wave structure of the flow and sediments distribution over plane (1DV models) and rippled beds (2DV models). The instantaneous transport rate of sediments is computed by integrating the sediment fluxes obtained at each level in the vertical. The main emphasis in developing these models is to describe and understand the physical processes occurring in the bottom boundary layer and serve as reference models to test the more simple parametric formulas. DAVIES et al. (1997) and DAVIES et al. (2002) present a comparison of different 1DV and 2DV models.

In morphodynamics computations more practical sand transport models are commonly use. The sediment transport is computed as a function of bottom shear stress or the near bed velocity and the quasi-steady hypothesis is normally assumed (e.g., BAILARD, 1981; RIBBERINK, 1998). As said before, the processes related to flow unsteadiness are not accounted for within these kinds of models. Therefore, the quasi-steady approach can only be suitable to conditions where the sediments are confined to a thin layer near the bottom such that the settling time of the sediment particles is much smaller than the wave period. In spite of these limitations, the proposed formula of RIBBERINK (1998), based in the well know Meyer-Peter and Muller formula, shows a good fit with data for a wide range of oscillatory and steady flows in flat bed conditions (sheet flow) for medium and course sand (d₃₂>0.02 cm).

To adequately describe the experimental results where non-steady effects are important, DIBAJNIA and WATANABE (1992) have proposed a formula that takes into account the time lag between the suspended sediment particles and the flow. This is done by an exchange of sediment flux between the two half cycles in a wave period due to a delayed settling of sediment particles. SILVA et al. (2001) improved the model of DIBAJNIA and WATANABE (1992) and present a comparison between the numerical results achieved with this formula and an extensive data base. As an example, figure 1 compares the computed net transport rates predicted by BAILARD (1981) formula and the SILVA et al. (2001) formula with the Series H experimental data of DOHMEN-JANSSEN et al. (2002). These experiments were performed in the OWT of WLU/Delft Hydraulics with a fine sand of d₃₂ = 0.013 cm with sinuous oscillatory motion with a superimposed current. The results show that for the conditions where phase-lag effects were observed (uₚ>0.78 m/s) the present model describes correctly the tendency of the experimental results, while the quasi-steady model of Bailard overpredicts.

DOHMEN-JANSSEN (1999) (see also DOHMEN-JANSSEN et al., 2002) has proposed an extension of RIBBERINK (1998) model, where an analytical diffusion model for sediment concentration is used for modeling the phase lag effects. It should be noted that the Dohmen-Janssen model introduces a correction parameter to the solutions of Ribberink model, which modify the magnitude of the net sediment transport rate, but not its direction. Therefore, a priori, this model cannot simulate accurately situations where the unsteady effects

change the direction of the net transport (e.g. rippled beds).

A more concise review of sediment transport models is given in the papers cited above and in SOULSBY (1997).

**PARAMETERIZATION OF THE NON-STEADY EFFECTS**

Consider the flow condition where a regular wave propagates to the shore in the presence of a collinear steady current with mean velocity Uᵢ. The near bed velocity, u(t), has a positive and a negative half cycles with time duration Tᵢ and Tᵢ, respectively. For each one of these half cycles, we can define an equivalent sinusoidal velocity amplitude, uᵢ and uᵢ as:

\[ uᵢ = \frac{2}{Tᵢ} \int_0^{Tᵢ} uᵢ(t) dt, \quad uᵢ = \frac{2}{Tᵢ} \int_0^{Tᵢ} uᵢ(t) dt \]  

(01)

(the indices c stands for the crest and t for the trough). If the waves velocity profile, u(t), is a known function, we can express analytically the quantities Tᵢ and uᵢ (i = c, t) as a function of uₚ, rᵢ = Uᵢ/uₚ, and the wave asymmetry, r, given by:

\[ r = \frac{u_{max} - u_{min}}{u_{max} + u_{min}} \]  

(02)

uₚ, max and uₚ, min represent the maximum (absolute) values of the near bed orbital velocity during the positive and negative half cycles, respectively. For a second order Stokes wave,

\[ uᵢ(t) = uᵢ \cos(\omega t) + uᵢ \cos(2\omega t) \]  

(03)

the following expressions can be derived for uᵢ and uᵢ:

\[ uᵢ = uᵢ + \gammaᵢ \]  

(04)

\[ uᵢ = uᵢ + \gammaᵢ \]  

(05)

With

\[ \gammaᵢ = 2(1 + rᵢ^2 + \gammaᵢ rᵢ) \]  

(06)

\[ \gammaᵢ = 2(1 + rᵢ^2 + \gammaᵢ rᵢ) \]  

(07)

rᵢ represents the dimensionless value of Tᵢ (tᵢ = Tᵢ/T). The factor γᵢ in the equations (6)-(7) is given by:

\[ γᵢ = \frac{1}{6} \sin(\pi Tᵢ) \left[ 3rᵢ + 19rᵢ + X(8rᵢ + 1) \right] \]  

(08)

where,

\[ X = \cos(\pi Tᵢ) \]  

(09)

and

\[ rᵢ = \frac{1 + r²}{2} rᵢ \]  

(10)

for the particular case of a sinus wave (r=0):

\[ X = \cos(\pi Tᵢ) = -rᵢ \]  

(11)

The time duration of each half cycle, Tᵢ and Tᵢ, are found by solving equations (9) or (11).

The net sediment transport rate during the wave period, qₛ, is computed accordingly to:

\[ qₛ = \frac{uₚ}{\sqrt{(s-1)} g d₃₂^3} \]  

(12)
where $\alpha$ and $\beta$ are two empirical constants ($\alpha = 3.2$; $\beta = 0.55$), $s$ the relative density, $g$ the gravity acceleration and $\Gamma$ is given by the following expression:

$$\Gamma = \frac{u_T T_s (\Omega_3^3 + \Omega_9^3) - u_T T_s (\Omega_3^9 + \Omega_9^9)}{u_T T_s + u_T T_s}$$  \hspace{1cm} (13)$$

Accordingly to equations (12) and (13), the net transport rate, $q_N$, is computed in terms of the difference between the sediments transported during the positive half cycle and the negative half cycle. In equation (13) the quantities $\Omega$ and $\Omega^*$ represent, respectively, the amount of sediments which are entrained, transported and settled in the $i$ half cycle, and the amount of sediments still in suspension from the $i$ half cycle, which will be transported in the next half cycle. The non-steady processes are taken into account through the exchanges of sediment fluxes between the two half cycles ($\Omega^*$ quantities). This exchange mechanism is controlled in the model by a parameter $\omega$, defined for each half cycle, which depends on the ratio between the settling time of the sediments particles, $T_{set}$, and the duration of each half cycle, $T$:

$$\omega = \frac{T_{set}}{T} - \frac{1}{2} \frac{u^2}{(s-1)g\alpha_{10}}$$  \hspace{1cm} (14)$$

In the last equation $\omega$ represents the sediments settling velocity. When the value of $\omega$ exceeds a threshold limiting value, $\omega^*$, part of the sediment that is entrained during the $i$ half cycle remains in suspension and is carried into the opposite direction by the velocity of the succeeding cycle. Therefore, this mechanism may enhance or reduce the transport in the wave direction. The values of $\omega^*$ were determined empirically as a function of the skin Shields parameter defined for each half cycle, $\theta_i$, in order to take into account both the flat bed and rippled bed regime:

$$\omega^* = \max \{0.035; 0.8(0.3 + \tanh(0.750 - 0.5))\}$$  \hspace{1cm} (15)$$

The quantities $\Omega$ and $\Omega^*$ in equation (13), are computed as:

$$\Omega = \theta_i \min(1, \frac{\Omega_3^3}{\Omega_9^9}) \hspace{1cm} \Omega^* = \theta_i \max(0, 1 - \frac{\Omega_3^9}{\Omega_9^9})$$  \hspace{1cm} (16)$$

where $\theta_i$ represents an equivalent Shields parameter defined for each half cycle as:

$$\theta_i = \frac{0.5 f c w u^2}{(s-1)gd_{10}}$$  \hspace{1cm} (17)$$

$f c w$ represents the wave-current friction factor for each half cycle.

The function defined in equation (13) can be express by a product of two independent functions:

$$\Gamma = G \Gamma_N$$  \hspace{1cm} (18)$$

The values of $\Gamma_N$ are computed assuming that the phase lag effects are not considered in the model (quasi-steady approach). $G$ is a function that translates the effect of the unsteady processes described above in sediment transport. Assuming that the wave-current friction factor is constant through the wave period we can derive the following expressions for $\Gamma_N$ and $G$:

$$\Gamma_N = \theta_i \frac{1 - \alpha_s \delta_s}{1 + \alpha_s \delta_s}$$  \hspace{1cm} (19)$$

$$\Gamma = Z^3 + \alpha_s \delta_s \frac{(1 - Z^9)}{1 - \alpha_s \delta_s}$$  \hspace{1cm} (20)$$

The quantities $\alpha_s$, $\delta_s$, $Z$, and $Z^9$, in equations (19)-(20) are given by:

$$\alpha_s = \frac{1 - \alpha_s}{\Gamma_c} \hspace{1cm} \delta_s = \frac{1 - \delta_s}{\Gamma_c} \hspace{1cm} Z_c = \frac{\alpha_c}{\delta_c} \hspace{1cm} Z^9_c = \frac{\alpha^9_c}{\delta^9_c}$$  \hspace{1cm} (21)$$

Figure 2. Contour lines of $F$ as a function of $r_i$ and $r$.

When the exchange mechanism is not effective $(\omega^* > \omega_s)$, the values of $Z$ and/or $Z^9$ are equal to one in equation (20).

According to equations (12) and (18), the presence of non-steady effects modifies the quasi-steady values of the net sediment transport through a function $F$:

$$F = G \left| \text{sign}(G) \right|$$  \hspace{1cm} (22)$$

**EVALUATION OF THE NON-STEEADY EFFECTS**

To establish the sensitivity of the non-steady effects, and therefore of the net sediment transport, in the known parameters of the problem ($u_{rms}$, $T$, $d_{rms}$, $r_i$, and $r$), we have defined a standard test case and evaluate the function $F$ for an adequate range of those parameters representative of the near-shore coastal zone.

The standard test case condition considers a 2^nd order Stokes wave with $r = 0.3$, $u_{rms} = 1$ m/s, $T = 7$s and a median grain size $d_{rms} = 0.025$ cm, corresponding to a settling velocity of 3.57 cm/s according to Soulsby (1997). The local depth, $h$, was set constant and equal to 3 m.

Figures 2-5 represent the computed values of $F$ when different values of $r_i$, $r$, $d_{rms}$, $T$ and $u_{rms}$ are considered, while holding the other input values at standard values. In figures 2-4 the value of $u_{rms}$ is constant and equal to 1 m/s. Therefore, $r_i$ represents a variation of $U_i$ between -0.5 and 0.5 m/s. This range of values encompasses the cross-shore mean velocities associated with the undertow and streaming. The values of $F$ are shown considering four different classes: F=0; 0< $F$< 1; $F$ = 1; and $F$ >1. The non-filled white areas correspond to the conditions where the non-steady effects are not effective ($F=1$).

In each figure, the thicker line indicates the conditions for which the quasi-steady net sediment transport, $q_{w}$, computed from

![Figure 3. Contour lines of F as a function of r_i and d_{rms}.](image3.png)
equation (19) is null. To the right of this line the values of \( q_{on} \) are positive (onshore) and to the left the values of \( q_{off} \) are negative (offshore).

Sheet flow conditions were observed for all test conditions considered in figures 2-4: in figure 2 the skin Shields parameter range between 2 and 2.65, in figure 3 between 1.2 and 5 and in figure 4 between 1.9 and 3. In figure 5, for \( u_{rms} < 0.5 \) m/s the values of range between 0.3 and 0.6, indicating the occurrence of rippled bed conditions. For the highest value of \( u_{rms} \) considered the computed value of is equal to 4.2. It should be stressed that the dependence of \( \theta \) as given in equation (15) was established for a dataset for which the \( \theta \) range between 0.1 and 4 and only three experimental conditions were used for \( \theta \geq 2.5 \). Therefore, the results shown for higher skin Shields parameter, in particular for \( d_{50} < 0.0125 \) cm, should be viewed with some restrictions.

Figure 2 illustrates the situation where the \( r \) and \( r_{rms} \) values were changed from the standard values. The quasi-steady net transport rates computed using equation (19) indicate that the values of \( q_{on} \) increase with the asymmetry of the waves velocity profile, \( r \). However, as can be seen in figure 2, the non-steady effects can be expected for values of \( r > 0.2 \) and for the highest positive values of \( r, U \), reducing the onshore quasi-steady net sediment transport. This occurs because the high values of velocity in the positive half cycle entrains a huge amount of sediment, part of which remains in suspension at flow reversal and are transported in the opposite direction by the negative flow velocities. During the negative half cycle, the sediments entrained from the bottom have time to settled during this half cycle.

In figure 3 the values of \( d_{50} \) considered range from very fine sediments (\( d_{50} = 0.01 \) cm) up to course sediments (\( d_{50} = 0.05 \) cm). The computed values of the quasi-steady net sediment transport depend slightly in \( d_{50} \), increasing for decreasing values of \( d_{50} \). The analysis of figure 3 reveals that the non-steady effects became important as \( d_{50} \) decreases, therefore changing this picture. For values of \( r > 0.2 \) and \( d_{50} < 0.03 \) cm there is an exchange of sediments between the positive and the negative half cycle that reduces or even reverse the direction of the onshore quasi-steady transport. For \( r < 0.2 \) and \( d_{50} < 0.02 \) cm the same kind of process occur that enhance the net sediment transport offshore, while for the largest (negative) values of \( r \), the exchange mechanism acts in the opposite sense (from the negative to the positive half cycle) decreasing the offshore sediment transport. These non-steady effects are due to a decrease of the sediments settling velocity but, as pointed out by DOHMEN-JANSSEN et al. (2002), they also result from an increase of the sheet flow layer thickness observed for the very fine sediments. For median and courser sediments the transport generally behaves in a quasi-steady way. However, it should be pointed out that for courser sediments than 0.05 cm ripples may form and non-steady effects became evident.

Figure 4. Contour lines of \( F \) as a function of \( r \) and \( T \).

The patterns of function \( F \) shown in figure 4 do not differ substantially from those in figure 3. In general, we may say that the decrease in the wave period traduces in a higher probability of the occurrence of non-steady effects. These effects are especially important for \( r > 0.2 \) and \( T < 4 \) s where the quasi-steady onshore sediment transport is reversed.
Finally, in figure 5 the values of the orbital wave velocity and the mean flow velocity were derived from the standard values. According to equation (19), the values of $q_{\text{rms}}$ increase with $u_{\text{rms}}$. However, as depicted in figure 5, changes from this behavior are to be expected for the high flow regime ($u_{\text{rms}} > 1 \text{ m/s}$) or when the bed is rippled.

In order to illustrate the importance of the non-steady effects in the development of the cross-shore beach profile, a Boussinesq wave model (Antunes do Carmo and Seabra Santos, 1996) was used to simulate wave propagation in the beach profile illustrated in figure 6a. The bathymetry was represented by means of data from a summer survey of Areão beach that is located south of Aveiro, Portugal, in an extensive sandy shore. The mean water level is at $z = 3.2$ m.

The model computes the water surface elevation and the depth average velocity from the wave conditions specified in the offshore open boundary located at $x = 0$ m. In the simulations, a sinus wave with 1.0 m height and period equal to 4.6 s was considered. At the shore a partial reflection of the waves was considered. The model also describes wave breaking and computes the undertow velocities.

The model results were analyzed to compute the values of $u_{\text{rms}}$, $r$, and $r_{\text{rms}}$. The quasi-steady transport, $q_{\text{rms}}$, and the values of the function $F$ were then evaluated from equations (19), (20), and (22) at different mesh points. Figures 6b represents the computed values of $u_{\text{rms}}$, $r$, and $r_{\text{rms}}$ along the cross-shore direction. For values of $x$ greater than 150 m the values of $u_{\text{rms}}$ decrease due to wave breaking. The waves velocity asymmetry increases with decreasing distance to the shore, namely at the sand bar. In figures 6c-e the solid and dashed lines represent, respectively, the quasi-steady ($q_{\text{rms}}$) and the corrected values of the net transport rates ($q$) for three different median grain sizes: 0.05; 0.025; and 0.013 cm. In these figures the symbols represent the computed values of the Shields parameter. Plane bed conditions ($\theta = 0.6$) are expected all over the beach profile for the finest grain size, while for the more courser sands, ripples are to be expected mainly over the sand bar.

The computed values of the quasi-steady net sediment transport rate indicate that in the region up to $x = 190$ m there is a positive transport to the shore with bed erosion for all $d_{\text{rms}}$ considered. Shoreward, the gradient of $q_{\text{rms}}$ induces accretion at the sand bar. The simulations performed with the mean and courser sediment, respectively, with $d_{\text{rms}} = 0.025$ cm and $d_{\text{rms}} = 0.05$ cm, show that phase-lag effects become important only after the shoaling zone, i.e., for $x = 150$ m. These phase lag effects arise due to the increasing values of $r$ and due to the presence of ripples (note that the $u_{\text{rms}}$ values decrease in this region) and reduce the amount of sediments transported onshore. In these conditions, accretion is to be expected further offshore of the sand bar. For the finest sediment grain size ($d_{\text{rms}} = 0.013$ cm) the computed corrected values of $q$ are all negative (offshore), promoting offshore bar migration and beach erosion.

**CONCLUSIONS**

The non-steady effects due to a delay settling of sediment particles were parameterized in the form of a simple function $F$, which can be easily computed in terms of known hydrodynamic parameters and sediment characteristics. The evaluation of this function for a range of hydrodynamic conditions representative of the coastal zone revealed that non-steady effects became important for increasing values of the waves orbital velocity ($u_{\text{rms}}$) and waves asymmetry ($r$), for decreasing values of the waves period ($T$) and in the case of rippled beds. The median grain size $d_{\text{rms}}$ of the sediment also plays an important role. It was seen that for the finest sediment considered, non-steady effects are important, while for medium and courser sands transport will generally behave in a quasi steady except in the high flow regime and when ripples are present in the bed.

The occurrence of non-steady effects strongly depend on how $d_{\text{rms}}$ is specified in terms of the Shields parameter. The formulation presented in equation (15) was established by optimizing the computed values of net sediment transport in wave and combined wave current flows in sheet flow and rippled beds (see Silva, 2001 for a detailed discussion). It was noticed that this formulation was unable to predict correctly the direction of sediment transport in 20% of the data considered. On the other hand, only a few data were available at high shear stress ($\theta > 2.5$). Therefore, further validation/improvements of the formulation were needed, a task that is being done within the EU program SANDPIT.

The application of the sediment transport practical model described to the Areão beach, while considering simplified wave boundary conditions, has revealed the importance of non-steady effects, especially evident for the finest sediment, in the development and evolution of a bar system.

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**LITERATURE CITED**


