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Review

Cross-Shore Sediment Transport in the Coastal Zone: A Review

François Marin * and Mélanie Vah

LOMC, UMR 6294 CNRS, Université Le Havre Normandie, Normandie Université, 25 rue P. Lebon, 76600 Le Havre, France; melanie.vah@univ-lehavre.fr
* Correspondence: francois.marin@univ-lehavre.fr

Abstract: This paper presents a review of cross-shore sediment transport for non-cohesive sediments in the coastal zone. The principles of sediment incipient motion are introduced. Formulations for the estimation of bedload transport are presented, for currents and combined waves and current flows. A method to consider the effect of sediment heterogeneity on transport, using the hiding–exposure coefficient and hindrance factor, is depicted. Total transport resulting from bedload and transport by suspension is also addressed. New research is encouraged to fill the knowledge gap on this topic.

Keywords: sediment transport; waves; current; coastal zone; bedload; suspension

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1. Introduction

Due to the importance of practical applications, such as coastal erosion, backfilling of dredged channels, and changes in near-shore morphology, numerous studies have been performed on sediment transport in the coastal zone. The vulnerability of coastal areas is particularly high in the context of climate change, with a predicted rise in sea levels, e.g., [1,2]. The coastal zone is generally considered as the transition zone where the land meets water. We focus, in this paper, on the area extending offshore to the continental shelf break and onshore to the most seaward point of the surf zone. The surf zone extends from the water line, the intersection of the land with the still water level, out to the most seaward point of the zone at which waves propagating towards the land begin to break. Waves and currents may occur in the coastal zone. Under the action of hydrodynamic forcing, sediments at the seabed may be transported. This paper presents a review of cross-shore sediment transport for non-cohesive sediments. Bedload transport and suspension transport may occur, depending on the physical properties of the particles and the flow conditions. In bedload transport, particles roll, shift, or make small jumps over the seabed but stay close to the bed. In the case of the suspended load mode, grains are lifted from the seabed and transported in suspension by the flow. The suspended load is primarily supported by fluid turbulence. As sediments generally exhibit size heterogeneity, some authors, e.g., [3,4], have considered the impact of such heterogeneity on transport. The sediment supply may be limited [5]. Despite numerous previous works, our understanding of the physical processes governing sediment transport in the coastal zone is still incomplete. The interaction between hydrodynamic forcing and sediments is highly complex. The modelling of sediment transport is thus largely based on empiricism. For sediments to be transported, conditions such as the motion threshold have to be exceeded. This is considered in Section 2. Bedload transport is presented in Section 3 for currents and combined waves and current flows. Total load transport including bedload transport and suspension transport is addressed in Section 4, while Section 5 is devoted to conclusions and future directions.

2. Sediment Incipient Motion

In view of the emphasis on the sediment movement threshold in coastal engineering, many works have been carried out on this subject. Sediment grains can be set in motion when the destabilising forces acting on the granular medium exceed the stabilising forces. The destabilising forces are linked to the bed shear stress τ_0 induced by the flow when the stabilising forces can be considered as frictional forces related to the submerged weight of the particles. The ratio between disturbing and stabilising forces on sand grains may be considered with the Shields parameter [6] θ, defined as follows:

$$\theta = \frac{\tau_0}{(\rho_s - \rho)gD} = \frac{u_*^2}{(s-1)gD}, \tag{1}$$

where ρ is the fluid density, ρ_s is the sediment density, g is the acceleration due to gravity, D is the median grain size, s is the relative density of sediment, and u_* is the shear velocity at bed ($u_* = \sqrt{\frac{\tau_0}{\rho}}$). The critical Shields parameter θ_c is the value of θ for $\tau_0 = \tau_{0c}$, where τ_{0c} is the critical value of the bed shear stress for the sediment initial movement. If the conditions are such that $\theta < \theta_c$, the sediments remain immobile, whereas they are set in motion when $\theta > \theta_c$. Soulsby and Whitehouse [7] proposed the following formulation for the estimation of θ_c as a function of D_*, the dimensionless grain diameter:

$$\theta_c = \frac{0.30}{1 + 1.2D_*} + 0.055[1 - exp(-0.020D_*)], \tag{2}$$

where $D_* = \left(\frac{(s-1)g}{\nu^2}\right)^{1/3} D$, with ν being the kinematic viscosity. This formulation, which can be applied for waves, currents, and combined waves and currents flows, is still widely used. Equation (2) assumes homogeneous sediment sizes. However, it is clear that, in nature, different sediment sizes are present. The movement of grains depends significantly on the presence of different particle sizes in the area under consideration. The presence of fine particles facilitates the movement of coarser particles, while the presence of coarse particles generally decreases the mobility of finer particles. The presence of fine grains leads to a decrease in the porosity of the medium. In the case of armouring, where the surface layer containing the largest grains prevents the mobilisation of finer grains in the substrate, the critical shear stress corresponds to the threshold for the coarsest particles.

A state of knowledge may be found in [8,9] for steady unidirectional streamflows. It is clear that the estimation of sediment incipient motion can be improved if future research is performed with size-heterogeneous sediments. Grain incipient motion is not easily defined, and Vah et al. [10] proposed three entrainment thresholds based on a new visual method in flume without human interference. The first threshold for very low shear stress conditions was described as the nascent movement threshold. It corresponds to the transition between the static bed state and the initial movement of a few grains. The second threshold was attributed to the bedload threshold, and the third one was defined as the bedform threshold. It is clear that relic bedforms affecting the processes of sediment incipient motion may be present in the field.

3. Bedload Transport

When the condition for sediment incipient motion is satisfied, the sediments can be transported by bedload and/or suspension. Bedload transport is the part of the total transport where intergranular forces are dominant, and the grains remain more or less in contact with the bed during this phase.

3.1. Bedload Transport Induced by Currents

Sediment transport is dominated by bedload transport when $u_*/W_s < 0.4$ for $D_* > 10$, where W_s is the settling velocity of isolated sediment grains, and when $u_*/W_s < 4/D_*$ for $D_* < 10$ [11]. It is clear that low values of shear velocity at the bed and high values of grain settling velocities promote bedload transport, while suspension transport is favoured for high values of u_* and low values of W_s.

Let us consider the dimensionless sediment transport given by

$$\phi = \frac{Q_s}{[g(s-1)]^{1/2} D^{3/2}}, \quad (3)$$

where Q_s is the dimensional transport (m/s^2). Meyer-Peter and Müller [12] proposed the following relationship based on the excess of bed shear stress $\theta' - \theta_c$, where θ' is the effective Shields stress calculated from the grain size for the bed roughness, a relationship which is still widely used for the bedload transport:

$$\phi = 8(\theta\prime - \theta_c)^{3/2}. \quad (4)$$

This relationship is mainly applicable for $D > 0.4$ mm and $D < 29$ mm. Nielsen [13] suggested the following Equation (5):

$$\phi = 12\theta\prime^{1/2}(\theta\prime - \theta_c). \quad (5)$$

Then, Ribberink [14] proposed a modification of the coefficients of the Meyer-Peter and Müller formula, based on extensive experimental data:

$$\phi = 10.4\left(\theta' - \theta_c\right)^{1.67}. \quad (6)$$

For Equation (6), the estimation of the equivalent bottom roughness, k_s, is performed as follows

$$k_s = 3D_{90} \text{ for } \theta' < 1, \quad (7)$$

$$k_s = D_{50}[1 + 6(\theta - 1)] \text{ for } \theta > 1. \quad (8)$$

In Equations (7) and (8), D_{90} is the grain diameter exceeded by 10% of the weight of the sample, and D_{50} is the median grain diameter.

From new data sets, van Rijn [15] proposed the following relationship, expressing the bedload transport in a dimensional form as a function of the excess of the current velocity compared to the critical value of this velocity for the sediment incipient motion, when the flow rate $\overline{U}$ is greater than 0.6 m/s:

$$Q_{sb} = 0.015\overline{U}d(D_{50}/d)^{1.2}M^{1.5}, \quad (9)$$

where d is the water depth, Q_{sb} is the dimensional bedload transport, and $M = \left(\overline{U} - \overline{U_{cr,c}}\right)/[(s-1)gD_{50}]^{0.5}$, with $\overline{U_{cr,c}}$ being the critical flow rate for the sediment incipient motion, which can be estimated [15] with:

$$\overline{U_{cr,c}} = 0.19(D_{50})^{0.1} log\left(\frac{12d}{3D_{90}}\right) \text{ for } 50\ \mu\text{m} < D_{50} < 500\ \mu\text{m} \quad (10)$$

$$\overline{U_{cr,c}} = 8.5(D_{50})^{0.6} log\left(\frac{12d}{3D_{90}}\right) \text{ for } 500\ \mu\text{m} < D_{50} < 2000\ \mu\text{m}. \quad (11)$$

The sediment supply may be limited at the seabed [16,17]. This has received little attention when compared to the literature dedicated to unlimited sediment supply conditions. Tuijnder and Ribberink [18] have shown that bedload transport is significantly dependent on the proportion of the bed covered with sediment in this case. Vah et al. [5] suggested using Equation (12) for the estimation of bedload transport for limited supply conditions ϕ_{lim}:

$$\phi_{lim} = (1 - p)\phi, \quad (12)$$

where $(1 - p)$ is the fraction of the bed covered by sand. Bedforms such as ripples or dunes are common on seabeds [19,20]. From flume data sets, Vah et al. [5] proposed to estimate the dimensionless velocity $c/u\prime_*$ of bedforms under limited sediment supply conditions, when bedload is the dominant mode of sediment transport, in the following way:

$$\frac{c}{u\prime_*} = \frac{2.5}{1-p}\left(\theta' - \theta_c\right)^{3/2}, \tag{13}$$

where c is the bedform migration velocity, and $u\prime_*$ is the effective bed shear stress velocity. Areas where the sediment supply is limited are not uncommon, and new works are needed to investigate sediment transport and bedform migration under limited sediment supply conditions.

The above formulae do not consider graded sediments, even if nonuniform sediments widely exist in the coastal zone. A commonly used method to take into account size heterogeneity for sediments is to introduce a hiding–exposure coefficient. The principle is to consider that it is more difficult to move small grains due to the presence of coarse grains and that coarse grains are more easily entrained if there are no finer grains. When sediments are too large for incipient motion and when they are surrounded by mobile fine grains, it is recommended to use a hindrance factor H_f to take into account the reduction in the bedload transport of small particles, in addition to the hiding–exposure coefficient [21]. Wu et al. [3] developed a semi-empirical formula to estimate dimensionless fractional bedload rates from flume experiments and alluvial river data. This formula, which is still widely used today, is recommended for 0.2 mm $< D_{50} <$ 50 mm and for a standard deviation between 1.28 and 9.91. The fractional bedload sediment flux $Q_{sb,i}$ is expressed in the following way, taking into account a hindrance factor:

$$Q_{sb,i} = 0.0053 H_f p_i \left[\frac{\tau\prime}{\tau_{c,i}} - 1\right]^{2.2} \sqrt{\left(\frac{\rho_s}{\rho} - 1\right) g D_i^3}, \tag{14}$$

where p_i is the proportion of each size class in the composition of the bed surface layer, $\tau\prime$ is the effective bed shear stress, $\tau_{c,i}$ is the critical value of the bed shear stress for size class i incipient motion, and D_i is the median particle diameter of the size class i in the surface layer. The critical bed shear stress is given by

$$\tau_{c,i} = (\rho_s - \rho) g D_i \theta_c \xi_i, \tag{15}$$

where $\theta_c = 0.03$, and ξ_i is the hiding–exposure coefficient. The formulation proposed by Egiazaroff [22] for this coefficient may be used:

$$\xi_i = \left[\frac{log(19)}{log(19\frac{D_i}{D_{50}})}\right]^2. \tag{16}$$

Kleinhans and van Rijn [21] recommend the following expression for the hindrance factor in Equation (14):

$$H_f = 1 - exp\left[-7.2\left(\frac{D_{90,bedload}}{D_{90}}\right)\right]. \tag{17}$$

Sediment transport by bedload for all size classes can be obtained by summing the transport by size class, taking into account their proportions:

$$Q_{sb} = \sum p_i Q_{sb,i}. \tag{18}$$

Van Rijn [23] proposed a fraction-wise approach for bedload transport and suspensions, for currents, waves, and combined waves and currents flows. He recommended such an approach for $D_{90}/D_{10} > 4$ and the use of about six to eight fractions. The range of application of the proposed formulation is the range of sands and cohesive sediments (8 µm $< D_{50} <$ 2000 µm). The sediment transport is obtained for each size class using a classical single-class method, replacing the grain median diameter with the mean diameter of the considered fraction and using the Egiazaroff [22] hiding–exposure coefficient as well as an adaptation for the effective bed shear stress. For $D_{50} > 2000$ µm, it is essential to use a stochastic model approach instead of this deterministic approach.

Durafour et al. [24] have shown that taking into account the sediment shape can improve the prediction of bedload transport for sediments that are heterogeneous in size. An adjustment of the Wu et al. [3] bedload formula, based on the introduction of a length characterizing the particle size and circularity, was found to significantly enhance the estimations of bedload transport rates for graded sediments. However, the literature about the effect of sediment shape on sediment transport is scarce, despite sediment particles (quartz grains, shells, etc.) exhibiting very different shapes.

3.2. Bedload Transport Induced by Waves and Combined Waves and Currents Flows

For waves, bedload transport is dominant when $W_s/u'_* > 1$ [25], where $u'_* = \sqrt{\tau'/\rho}$. It is well known that water waves induce steady streaming [26,27]. Although this streaming is weak compared with the oscillatory component of velocity, it has a significant effect on sediment transport.

Let us consider expressions for the estimation of bedload transport induced by waves plus currents. Soulsby [28] proposed a formulation for bedload transport in the case of current-dominated flows and in the case of wave-dominated flows. This formulation results from an integration over a wave cycle of the bedload transport formula proposed by Nielsen in the case of currents alone (Equation (5)):

$$\phi_{x1} = 12\theta_m^{\frac{1}{2}}(\theta_m - \theta_c), \tag{19}$$

$$\phi_{x2} = 12(0.95 + 0.19cos2\phi_a)\theta_w^{1/2}\theta_m, \tag{20}$$

where ϕ_{x1} is the transport in the direction of the current for current-dominated flows, ϕ_{x2} is this transport for wave-dominated flows, θ_m is the mean value over a wave cycle of the Shields number, θ_w is the amplitude of the Shields number for waves alone, and ϕ_a is the angle between the current direction and the waves' propagation direction. Perpendicular to the current direction, the bedload transport is expressed by

$$\phi_y = \frac{12(0.19\theta_m\theta_w^2 sin2\phi)}{\theta_w^{3/2} + 1.5\theta_m^{3/2}}. \tag{21}$$

Ribberink [14] adapted his formulation for currents (Equation (6)) to the cases of waves and combined waves and current flows, using the results of more than 150 laboratory experiments. For oscillatory flows, the time-dependent bedload transport was treated in a quasi-steady way. The time-averaged dimensionless transport is given by

$$\overline{\phi(t)} = 10.4\overline{(|\theta'(t) - \theta_c|)^{1.67}\frac{\theta'(t)}{|\theta'(t)|}} \text{ for } |\theta'(t)| \geq \theta_c, \tag{22}$$

and $\overline{\phi(t)} = 0$ for $|\theta'(t)| < \theta_c$, where $\theta'(t)$ is the instantaneous Shields parameter:

$$\theta'(t) = \frac{0.5\rho f_w |u_\infty(t)| u_\infty(t)}{(\rho_s - \rho) g D_{50}}, \tag{23}$$

with $u_\infty(t)$ being the horizontal flow velocity just outside the bottom boundary layer, and f_w representing the friction coefficient based on the Swart [29] formula. This coefficient is expressed as follows:

$$f_w = exp\left[-5.98 + 5.2\left(\frac{k_s}{a}\right)^{0.194}\right] \text{ for } \frac{k_s}{a} < 0.63, \tag{24}$$

$$f_w = 0.3 \qquad \text{for } \frac{k_s}{a} \geq 0.63, \tag{25}$$

where a is the orbital amplitude of fluid just outside the bottom boundary layer, and k_s is the bed roughness length:

$$k_s = D_{50} \qquad \text{for } \overline{|\theta|} < 1, \tag{26}$$

$$k_s = D_{50}\left[1 + 6\left(\overline{\theta} - 1\right)\right] \qquad \text{for } \overline{|\theta|} \geq 1, \tag{27}$$

where

$$\overline{\theta} = \frac{\overline{|\tau_0(t)|}}{(\rho_s - \rho)gD_{50}}, \tag{28}$$

and

$$\overline{|\tau_0(t)|} = \frac{1}{2}\rho f_w \overline{u_\infty^2(t)} = \frac{1}{4}\rho f_w U_\infty^2, \tag{29}$$

with U_∞ representing the amplitude of the horizontal component of the fluid particle velocity just outside the bed boundary layer.

For combined waves and current flows, Ribberink [14] adapted his formulation for waves alone (Equation (22)):

$$\overline{\vec{\phi}(t)} = 11\overline{\{|\theta'(t)| - \theta_c\}^{1.65}\frac{\vec{\theta'}(t)}{|\theta'(t)|}}, \tag{30}$$

where

$$\left|\theta'(t)\right| = \sqrt{\theta'^2_x(t) + \theta'^2_y(t)}. \tag{31}$$

The coordinates x and y correspond to coordinates in an horizontal plane $x - y$. The dimensionless bed shear stress is estimated as follows:

$$\vec{\theta'}(t) = \frac{\vec{\tau_0}(t)}{(\rho_s - \rho)gD_{50}} = \frac{0.5\rho f_{cw}|u_\infty(t)|\vec{u_\infty}(t)}{(\rho_s - \rho)gD_{50}}, \tag{32}$$

where

$$|u_\infty(t)| = \sqrt{u_{\infty x}^2(t) + u_{\infty y}^2(t)}, \tag{33}$$

$$\vec{u}_\infty(t) = \vec{U}_{z=\delta} + \vec{u}_{\infty,osc}(t), \tag{34}$$

with $\vec{U}_{z=\delta}$ representing the current velocity at $z = \delta$, which is the wave boundary layer thickness, and $\vec{u}_{\infty,osc}(t)$ representing the oscillating component of the velocity vector for $z = \delta$. The friction coefficient in combined waves and current flows f_{cw} is estimated by

$$f_{cw} = \alpha f_c + (1 - \alpha) f_w, \tag{35}$$

where f_c is the friction coefficient due to the current, corresponding to the quadratic friction coefficient C_D, and $\alpha = U_{z=\delta}/(U_{z=\delta} + U_\infty)$. The roughness length k_s is determined by

$$k_s = max\left\{3D_{90}, D_{50}\left[1 + 6\left(\overline{|\theta|} - 1\right)\right]\right\} \tag{36}$$

for the current flow, and by Equations (26) and (27) for the waves part of the flow. The physical reasons for this difference in roughness length for steady and unsteady flows might be attributed to the different nature of the bedload motion in current flow and unsteady flows. In oscillatory sheet flows, grain–grain flow interactions occur in an important part of the wave boundary layer, while in steady flows, the flow boundary layer generally extends over the full water depth and the sheet flow layer is quite thin.

We have

$$\overline{|\theta\prime|} = \frac{\overline{|\tau_0|}}{(\rho_s - \rho) g D_{50}}, \tag{37}$$

where

$$\overline{|\tau_0|} = 0.5 \rho f_c U_{z=\delta}^2 + 0.25 \rho f_w U_\infty^2. \tag{38}$$

Using an intrawave approach, van Rijn [15] suggested an extension of his formulation developed for current flows to the case of combined waves and current flows; the same equation (Equation (9)) can be used if the following relation is used for the parameter M:

$$M = \left(U_e - \overline{U_{cr}}\right) / [(s-1) g D_{50}]^{0.5}, \tag{39}$$

where the effective velocity $U_e = \overline{U} + \gamma U_{\infty,p}$, with $\gamma = 0.8$ for regular waves and $\gamma = 0.4$ for irregular waves, and $U_{\infty,p} = \pi H_s / \left[T_p sinh(kd)\right]$, the peak orbital velocity based on the significant wave height H_s and the most probable period T_p. The critical flow rate for the sediment incipient motion depends on the critical flow rate for current flows alone, $\overline{U_{cr,c}}$, and on the critical flow rate for waves alone, $\overline{U_{cr,w}}$:

$$\overline{U_{cr}} = \beta \overline{U_{cr,c}} + (1 - \beta) \overline{U_{cr,w}}, \tag{40}$$

where $\beta = \overline{U} / \left(\overline{U} + U_{\infty,p}\right)$. The critical flow rate for current flows alone may be estimated with Equations (10) and (11). The critical flow rate for waves alone can be determined from the following formulations:

$$\overline{U_{cr,h}} = 0.24 [(s-1) g]^{0.66} D_{50}^{0.33} \left(T_p\right)^{0.33} \text{ for } 50\ \mu\text{m} < D_{50} < 500\ \mu\text{m} \tag{41}$$

$$\overline{U_{cr,h}} = 0.95 [(s-1) g]^{0.57} D_{50}^{0.43} \left(T_p\right)^{0.14} \text{ for } 500\ \mu\text{m} < D_{50} < 2000\ \mu\text{m}. \tag{42}$$

Van Rijn [15]'s formulation is recommended for dominant current conditions and values of effective velocities U_e greater than approximately 0.5 m.s^{-1}.

As far as total sediment transport is concerned, some formulae split the transport into transport by bedload, on the one hand, and by suspension, on the other (see Section 4). These formulae can therefore also be used for the estimation of bedload transport. Let us consider the total load transport.

4. Total Load Transport

4.1. Total Load Transport Induced by Currents

The total load transport is the sum of bedload and suspension transport. Engelund and Fredsoe [30] suggested Equation (43) for the estimation of the total load transport when 0.15 mm < D < 4 mm, a formulation which does not involve the threshold value of the Shields parameter for incipient motion:

$$\phi = \frac{0.08 {C_h}^2}{2g} \theta^{5/2}, \tag{43}$$

where C_h is the Chézy number:

$$C_h = \overline{U} \sqrt{(\rho g / \tau_0)}\,. \tag{44}$$

Equation (43) assumes that the grains move at a velocity proportional to the friction velocity $\overline{u_*}$.

Based on a dimensional analysis and a comprehensive analysis of laboratory experiments, Ackers and White [31] proposed the following empirical formulation for the total load transport:

$$Q_s = C_{AW} \overline{U} D \left(\frac{\overline{U}}{\overline{u_*}}\right)^n \left(\frac{F_{AW} - A_{AW}}{A_{AW}}\right)^m, \tag{45}$$

where

$$F_{AW} = \frac{\bar{u}_*^n}{[g(s-1)D]^{1/2}} \left[\frac{\bar{U}}{2.46Ln(10d/D)} \right]^{1-n}. \quad (46)$$

For $1 < D_* \le 60$, we have

$$n = 1 - 0.243LnD_* \quad (47)$$

$$A_{AW} = \frac{0.23}{{D_*}^{1/2}} + 0.14 \quad (48)$$

$$m = \frac{6.83}{D_*} + 1.67 \quad (49)$$

$$C_{AW} = exp\left[2.79LnD_* - 0.426(LnD_*)^2 - 7.97\right]. \quad (50)$$

For $D_* > 60$, these coefficients are expressed as follows:

$$n = 0 \quad (51)$$

$$A_{AW} = 0.17 \quad (52)$$

$$m = 1.78 \quad (53)$$

$$C_{AW} = 0.025. \quad (54)$$

Using a parameterisation of a detailed grain saltation model representing the basic forces acting on a bedload particle for bedload transport and a depth integration of the product of the local concentration and flow velocity for the suspended load, van Rijn [15,32] proposed the following formulation for the estimation of the total load transport for sand under currents alone for a water depth of 1 to 20 m and a flow rate of 0.6 to 5 m·s^{-1}:

$$Q_s = Q_{sb} + Q_{susp} = 0.015\overline{U}d(D_{50}/d)^{1.2}M^{1.5} + 0.012\overline{U}d\left(\frac{\overline{U} - \overline{U_{cr,c}}}{[(s-1)gD_{50}]^{1/2}}\right)^{2.4}\left(\frac{D_{50}}{d}\right)D_*^{-0.6}, \quad (55)$$

where Q_{susp} is the transport by suspension, and $\overline{U_{cr,c}}$ can be obtained from Equations (10) and (11). Measured concentration profiles for calibration have also been used to obtain Equation (55).

For graded sediments, and following Wu et al. [3], the total sediment transport may be estimated by

$$Q_{s,i} = 0.0053H_f p_i\left[\frac{\tau\prime}{\tau_{c,i}} - 1\right]^{2.2}\sqrt{\left(\frac{\rho_s}{\rho} - 1\right)gD_i^3} + 0.0000262p_i\left[\left(\frac{\tau}{\tau_{c,i}} - 1\right)\frac{\overline{U}}{W_S}\right]^{1.74}\sqrt{(\frac{\rho_s}{\rho} - 1)gD_i^3}, \quad (56)$$

where the first term on the right of Equation (56) corresponds to bedload transport, and the second term refers to transport by suspension. Zhang and Xie's [33] formulation is recommended to estimate the settling velocity in Equation (56):

$$W_s = \sqrt{(13.958/D_i)^2 + 1.09(s-1)gD_i} - 13.95\nu/D_i. \quad (57)$$

Similarly to bedload transport, the total sediment transport for all size classes is obtained as follows:

$$Q_s = \sum p_i Q_{s,i}. \quad (58)$$

4.2. Total Load Transport Induced by Waves and Combined Waves and Currents Flows

For combined waves and current flows, waves significantly increase sediment suspension.

From an energetic approach, and taking into account the effect of bottom slope on transport, Bailard [34] proposed an expression for the sediment transport in vector form:

$$\vec{Q}_s = \vec{Q}_{sb0} + \vec{Q}_{sbp} + \vec{Q}_{susp0} + \vec{Q}_{suspp}, \tag{59}$$

where $\vec{Q}_{sb0}$ is the bedload transport over horizontal bed, $\vec{Q}_{sbp}$ is the bedload transport resulting from the bed slope, $\vec{Q}_{susp0}$ is the transport by suspension over the horizontal bed, and $\vec{Q}_{suspp}$ is the transport by suspension due to the bottom slope. These four terms are expressed by Equations (60)–(63):

$$\vec{Q}_{sb0} = \frac{0.1 f_{cw}}{g(s-1)tan\varphi}\overline{|u|^2\vec{u}} \tag{60}$$

$$\vec{Q}_{sbp} = \frac{0.1 f_{cw} tan\beta}{g(s-1)tan^2\varphi}\overline{|u|^3}\vec{i} \tag{61}$$

$$\vec{Q}_{susp0} = \frac{0.02 f_{cw}}{g(s-1)W_s}\overline{|u|^3\vec{u}} \tag{62}$$

$$\vec{Q}_{suspp} = \frac{4.10^{-4} f_{cw} tan\beta}{g(s-1)W_s{}^2}\overline{|u|^5}\vec{i}, \tag{63}$$

where

$$\vec{u}(t) = u_x\vec{x} + u_y\vec{y}, \tag{64}$$

where the particle fluid velocities close to the bed are represented as

$$u_x = \overline{U} + U_\infty cos\phi_a cos\omega t \tag{65}$$

$$u_y = U_\infty sin\phi_a cos\omega t, \tag{66}$$

where $\vec{x}$ is the current direction, β is the bottom slope angle, $\vec{i}$ is the unit vector in the slope direction, ω is the wave pulsation, and φ is the internal friction angle of the sediment.

Assuming linear waves, we have

$$\overline{|u|^2\vec{u}} = \vec{\overline{U}}\left(\overline{U}^2 + \frac{1}{2}U_\infty{}^2\right) + \vec{U}_\infty\left(\vec{\overline{U}}\cdot\vec{U}_\infty\right). \tag{67}$$

Then,

$$\overline{|u|^2\vec{u}}_x = \overline{U}^3 + \overline{U}U_\infty{}^2\left(1 + \frac{1}{2}cos2\phi_a\right) \tag{68}$$

$$\overline{|u|^2\vec{u}}_y = \frac{1}{2}\overline{U}U_\infty{}^2 sin2\phi_a. \tag{69}$$

For waves propagating in the same direction as the current, we have

$$\overline{|u|^3\vec{u}}_x = \frac{1}{8}\left(24U_\infty^2\overline{U}^2 + 8\overline{U}^4 + 3U_\infty^4\right) \tag{70}$$

$$\overline{|u|^3\vec{u}}_y = 0. \tag{71}$$

For a sandy rippled bed, Soulsby and van Rijn (Soulsby [28]) suggested the following formulation for the sediment transport in the current direction:

$$Q_s = A\bar{U}\left[\left(\bar{U}^2 + \frac{0.018}{C_D}U_\infty{}^2\right)^{1/2} - \overline{U_{cr,c}}\right]^{2.4}(1 - 1.6tan\beta), \tag{72}$$

where

$$A = A_{sb} + A_{susp} \quad (73)$$

$$A_{sb} = \frac{0.005d(D_{50}/d)^{1.2}}{[(s-1)gD_{50}]^{1.2}} \quad (74)$$

$$A_{susp} = \frac{0.012D_{50}D_*^{-0.6}}{[(s-1)gD_{50}]^{1.2}}. \quad (75)$$

$$C_D = \left[\frac{0.40}{1 + Ln(z_0/d)}\right]^2 \quad (76)$$

is the drag coefficient due to current alone,

$$z_0 = 6 \quad (77)$$

mm is the bed roughness length, and $\overline{U_{cr,c}}$ is the critical flow rate for the sediment incipient motion, which can be estimated with Equations (10) and (11). The decomposition of the parameter A in two components (Equation (73)), one for bedload (A_{sb}) and one for suspension (A_{susp}), makes it possible to estimate the quantities of transported materials for each of these two types of transport. This formulation includes slope effects and a threshold velocity.

The total sediment transport may also be estimated using van Rijn's [15,35] formula by adding bedload transport (Equation (9) with Equation (39) for the parameter M) and the following expression for the transport by suspension:

$$\vec{Q}_{susp} = 0.012\overline{U}D_{50}M^{2.4}(D_*)^{-0.6}\vec{x} + \frac{0.1V_{asym}}{\rho_s}\int_{z_a}^{\delta} Cdz\vec{x_w}. \quad (78)$$

The first term in the right side of Equation (78) depicts the transport by suspension due to the mean current in the current direction $\vec{x}$, taking into account the waves' effect on sediment agitation. The second term in the right side of Equation (78) represents the wave-induced transport by suspension in the wave propagation direction $\vec{x_w}$ due to the asymmetry of fluid particle motion close to the bed. In Equation (78), C is the sediment volumetric concentration, z is the vertical coordinate, and V_{asym} is the asymmetry factor given by

$$V_{asym} = \left[(U_{on})^4 - \left(U_{off}\right)^4\right] / \left[(U_{on})^3 + \left(U_{off}\right)^3\right], \quad (79)$$

where U_{on} is the maximum orbital velocity directed towards the shore, and U_{off} is the maximum orbital velocity directed towards the open sea. The thickness of the near bed suspension layer is

$$\delta = 6\left[1 + (H_s/d - 0.4)^{0.5}\right]\delta_w, \quad (80)$$

where the boundary layer thickness due to waves is

$$\delta_w = 0.36A_\delta(A_\delta/k_s)^{-0.25}, \quad (81)$$

with A_δ equalling the maximum orbital excursion based on the significant wave height H_s, and k_s representing the equivalent bed roughness of waves. The reference height z_a in Equation (78) is the maximum value between the roughness half lengths induced by the current and by waves, with a minimum threshold value of 1 cm.

For wave-dominated flows, van der A et al. [4] suggested a formulation where sediment transport is depicted as the difference between the sediment quantities transported each half-cycle. The effects of asymmetric acceleration on transport are considered. The total transport is given by

$$\vec{\phi} = \frac{\sqrt{|\theta_1|}T_1\left(\Omega_{11} + \frac{T_1}{2T_{11}}\Omega_{21}\right)\frac{\vec{\theta_1}}{|\theta_1|} + \sqrt{|\theta_2|}T_2\left(\Omega_{22} + \frac{T_2}{2T_{21}}\Omega_{12}\right)\frac{\vec{\theta_2}}{|\theta_2|}}{T}. \quad (82)$$

In this equation, θ_1 and θ_2 are the characteristic Shields numbers for the half0cycles corresponding to the passage of wave crest and trough, respectively, T_1 and T_2 are the duration of the half-cycles for the passage of wave crest and trough, respectively, T_{11} and T_{21} are the flow acceleration durations for the half-cycles T_1 and T_2, respectively, and T is the wave period. The contributions to the transport are Ω_{11} for the sand quantity entrained during the half-cycle of the wave crest and transported during this half-cycle, Ω_{12} for the sand quantity entrained during the half-cycle of the wave crest and transported during the next half-cycle of the wave trough, Ω_{22} for the sand quantity entrained during the half-cycle of the wave trough and transported during this half-cycle, and Ω_{21} for the sand quantity entrained during the half-cycle of the wave trough and transported during the next half-cycle of the wave crest. The way to estimate these terms is detailed in van der A et al. [4].

Van der A et al. [4] developed a multi-class version of their formulation, and the total transport can be obtained using:

$$\vec{\phi} = \sum_{i=1}^{N} p_i \frac{\vec{Q}_{s,i}}{\sqrt{(s-1)gD_i^3}}. \quad (83)$$

Wu and Lin [36] adapted Wu et al.'s formula for graded sediment transport for current alone to the case of combined flow. Taking into account the hindrance factor suggested by Kleinhans and van Rijn [21], we obtain

$$Q_{s,i} = 0.0053H_f p_i \frac{T_c}{T}\left[\frac{\tau'_{on}}{\tau_{c,i}} - 1\right]^{2.2}\sqrt{\left(\frac{\rho_s}{\rho} - 1\right)gD_i^3} + 0.0053H_f p_i \frac{T_t}{T}\left[\frac{\tau'_{off}}{\tau_{c,i}} - 1\right]^{2.2}\sqrt{\left(\frac{\rho_s}{\rho} - 1\right)gD_i^3} + 0.0000262p_i\left[\left(\frac{\tau}{\tau_{c,i}} - 1\right)\frac{\overline{U}}{W_S}\right]^{1.74}\sqrt{\left(\frac{\rho_s}{\rho} - 1\right)gD_i^3}, \quad (84)$$

where T_c is the duration between the waves zero up- and down-crossing, $T_t = T - T_c$, and $\tau\prime_{on}$ and $\tau\prime_{off}$ are the onshore and offshore effective bed shear stress, respectively. The first two terms on the right side of Equation (84) correspond to bedload transport, while the third term corresponds to transport by suspension.

Hu and Chen [37] proposed an analytical model for graded sediment transport in velocity and acceleration skewed oscillatory sheet flow. The composition of transported sediment is shown to be different from the initial bed composition. The fining of sediment above the initial bed and the coarse sediment below the initial bed are reproduced by the model, which reveals vertical sorting processes.

Finally, Aasgaard et al. [38] provided a review of recent experimental advances on suspended sediment transport in the surf zone, and Chen et al. [39] described practical models of sand transport in the swash zone around the water line.

5. Conclusions and Future Directions

Numerous studies have been carried out on cross-shore sediment transport in coastal areas, as shown in Table 1 for bedload transport and total load transport induced by currents, waves, and combined waves and current flows. The different proposed relationships for sediment transport do not always involve the same parameters and can lead to different estimations. This results from an insufficient understanding of the physical processes governing sediment transport. Future research is needed concerning bedform migration and sediment transport in the case of limited sediment supply, a case which is not uncommon. New works are also desired for size-heterogeneous sediment conditions. It is well known that a mixture of grains of different sizes does not behave in the same way as size-homogeneous particles, but the way to properly tackle this problem is still unclear. Particle shape is also recognized to be a significant factor in the hydrodynamic behaviour of grains, but the involved mechanisms are not fully understood. Particles are generally characterised by 2D information, and future research is necessary to study the impact of the third dimension on transport through the analysis of 3D grain shape factors. It is therefore crucial that further work is carried out to improve our knowledge of sediment transport in the coastal zone. This subject has very important practical applications, for example, for coastal erosion, especially in the context of climate change with a predicted rise in sea level.

Table 1. Cited references for sediment transport (bedload/total load transport) induced by currents, waves, and combined waves and current flows.

	References
Bedload transport induced by currents	Wu et al. [3], Vah et al. [5], Meyer-Peter and Müller [12], Nielsen [13], Ribberink [14], van Rijn [15], Tuijnder and Ribberink [18], van Rijn [23], Durafour et al. [24]
Bedload transport induced by waves and combined waves and currents flows	Ribberink [14], van Rijn [15], van Rijn [23], Soulsby [28], Bailard [34], Wu and Lin [36]
Total load transport induced by currents	Wu et al. [3], van Rijn [15], van Rijn [23], Engelund and Fredsoe [30], Ackers and White [31], van Rijn [32]
Total load transport induced by waves and combined waves and currents flows	van der A et al. [4], van Rijn [15], van Rijn [23], Soulsby [28], Bailard [34], van Rijn [35], Wu and Lin [36], Hu and Chen [37]

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