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SEDIMENT TRANSPORT IN THE COASTAL ZONE

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Abstract

The paper provides basic information on the description of sediment transport in the coastal zone of the sea. It explains the mechanisms of sediment movement under the influence of the waves and currents interactions, characteristic for the coastal zone. It presents models describing the movement of sediment in the regime of flat bottom, making their divided in accordance with the method of description of vertical structure of sediments transport. Particular emphasis is placed on modeling of graded sediment transport. It presents basis for the three-layer model of graded sediment transport. This allows, among others, the analysis of the variability of particle size distribution in the whole area of sediment movement. The model assumes that the movement of sediment is divided into three layers: bedload layer, contact load layer and outer flow region, as a result of the shear stress influence on the bottom.

Introduction

The primary factor leading to the reconstruction of the profile of the seabed is a wave caused by the wind and the return current which is caused by this wave. Wind-induced waves are generated in the deep-water off-shore area, where the seabed does not affect the nature of the waves. As a result of decreased depth, the wave undergoes a transformation process leading to the increase of the wave crest height and shortening its length, each wave through becoming shallower and longer. In the area of deepwater it is necessary to use the sinusoidal wave approximation, whereas along with the decrease of the depth, the 2nd Stokes approximation is used and then the cnoidal approximation

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and solitary waves (DRUET, KOWALIK, 1970). In the coastal zone, for the purpose of describing sediment transport, the 2nd Stokes wave approximation is most commonly used (KACZMAREK 1999). It is characterized by shortened and steep crest and elongated and flattened through, in comparison to the sine wave.

In the coastal zone there is an interaction between the hydrodynamic forces and sediment and bathymetric profile of the bottom. The fluid stream due to friction forces, runs the sediment from the bottom and the sediment material is transferred over a certain distance. So, moving water causes the bottom sediment transport, and the spatial variability of sediment transport volume causes changes in morphology of the bottom, which in turn affects the water movement change. The question of what controls the rate and method of sediment transport and the nature of changes in the level of the bottom occupied researchers for over a century. Numerous relations between the level of bottom elevation and the rate of sediment transport and flow parameters, including a number of empirical and far less theoretical relations have been proposed. Empirical relations generally can not be applied beyond the limited conditions for which they were formulated, and the majority of theoretical propositions depends on arbitrary assumptions with only a little or no physical credibility (ALLEN 1977). Comprehensive description of such a complex system is very difficult and therefore only by successive approximations and simplifications it is possible to describe the phenomena occurring in the coastal zone of the sea. Despite that in recent years there has been a significant progress, still many issues remain unresolved.

The objective of this paper is the review of sediments transport models. It has been indicated that the primary advantage of the two- and three-layer models in comparison with classic models, based on empirical and semi-empirical relations, is the fact, that such models allows to evaluate a concentration of sediments at any level, depending on the instantaneous hydrodynamic force. The proposed three-layer theoretical model of graded sediments transport providing a complete theoretical description of the vertical structure of transportation of the sandy sediment of a various grain size. Using this model, it is possible to predict an amount of moving sandy sediments of various grain size, as well as there is a possibility to predict the size of the transport of individual fractions.

Sediment transport in the coastal zone

Hydrodynamic processes (waving and wave driven currents) are the driving force of the sediment transport and the evolution of seabed. Actual parameters of morphodynamic and lithodynamic processes depend on the kind of sediment which residues in the seabed and on the supply of these fractions of sediment that are susceptible to the effects of water flow in the bottom layer (transportation in the form of bedload transport and suspended as a result of shear bottom stresses impact). First of all, the parameters of the litho – and morphodynamic processes depend on the wave climate, the bathymetric bottom arrangement and hydrotechnical facilities in the coastal zone of the sea.

In the traditional division (GRADZIŃSKI et al. 1986) adopted in considerations concerning the transport of sediment, the sediments are transported in three layers, starting from the lowest point: bedload, saltation and in the layer of suspended sediments. The bedload layer covers an area below the bottom of the theoretical level of very high concentration of sediment particles set in motion under the influence of the shear stress impacting the bottom surface. Shearing the bottom layer being the result of shear stress causes only a slight increase in the space between the particles of sediment. Surface friction and intermolecular collisions cause energy transfer between individual molecules of the sediment. The thickness of the bedload layer ranges from one to several tens of sediment grain diameters (NIELSEN 1992, O'DONOGHUE, WRIGHT 2003). In saltation layer with a thickness of about few centimeters, the sediment particles are transported and they are raised from the bottom as a result of turbulent pulsations and punching through the falling sediment particles on the small height above the bottom. Due to the short time of the particles staying in suspension, their transport depends on the oscillation velocity of the wave motion (GRADZIŃSKI et al. 1986). Suspension layer with a thickness of about meters covers an area over the saltation layer up to the free surface of the water column. In this layer, sediment particles reside for a period longer than the period of the wave, and their resultant transport is mainly related to the return current which is characteristic for the coastal zone of the sea. To describe the transport of sediment in suspension it is applied the theory of compensating return current (SVENDSEN 1984).

Sediment transport may take place in two basic regimes. In the regime in which the bottom is covered with bottom forms, e.g. in the form of ripples or in the regime where the bottom is flat with a mobile "carpet" of sediments or flat – stationary. The ability of occurrence of these sediment movement variants depends primarily on the type of sediments building the sea bottom and on the intensity of the waving impact on the bottom. To determine the size of the driving force causing the movement of sediment, most commonly the Shields parameter is used – $\Theta_{2.5}$, and this defines the dimensionless friction. It is assumed, that the beginning of the sediment movement occurs when $\Theta_{2.5} = 0.05$ (NIELSEN 1992).

The resultant sediment transport rate along the transverse profile of the sea coast is the result of the coexistence of wave motion asymmetry and compensating return current. The consequence of the wave motion asymmetry is the asymmetry of orbital velocity of water at the bottom between the wave crest and the through (CHEN et al. 2010, ZHENG 2007, YHENG et al. 2008). There is a far greater impact on the bottom during the wave crest than during the through. As a result, the individual fractions of the bed sediment are more intensively transported in landward than in seaward direction (Fig. 1). Despite the fact that the duration of the wave through is longer than the crest, only small diameter grains participate in the motion. During the wave crest, the wider spectrum of sediment particles is transported.

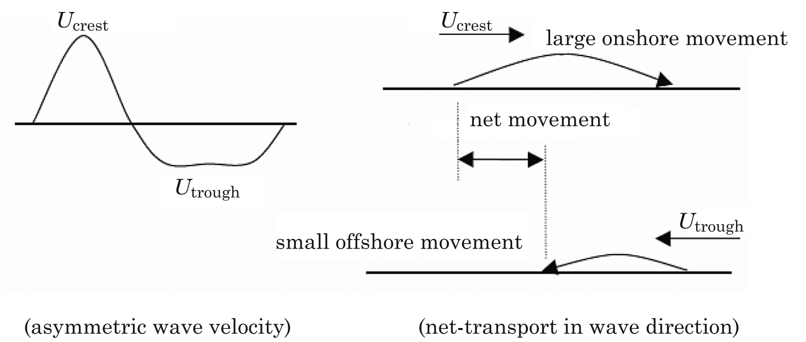

Fig. 1. Resultant sediments transport rate under the influence of the asymmetric waves
Source: HASSAN (2003).

As demonstrated by laboratory tests, carried out in the oscillatory tunnel (RIBBERINK, AL-SALEM 1994), the resultant sediment transport rate with the representative median diameter of d_{50} = 0.21 mm is in landward direction, i.e. compatible with the direction of wave propagation. For fine-grained sediment of d_{50} = 0.13 mm, RIBBERINK and CHEN (1993) obtained the opposite (from the coast) transport direction. It turns out that under conditions of strong wave functions, the resultant sediment transport changes its direction (Fig. 2). The reason for this is the fact that the suspension during the wave through (in seaward direction of the water orbital velocity) of fine grains of sediment, previously raised from the bottom during the wave crest, moves in seaward direction. Sediment particles remain in suspension for the duration of the wave through due to the low speed of their descent. The phase lag (arising as a result of the above situation) between the maximum speed of the water and the maximum concentration of sediment can be particularly well noted for the fine sediment, transported by the strong and short-term waves (DOHMEN-JANSSEN, HANES 2002, JANSSEN 1995, O'DONOGHUE, WRIGHT 2004, VAN DER A et al. 2010). As a result, both the effect of the phase lag and the impact of return current result that the resultant transport of the finest fractions of sediment is directed towards the sea. The magnitude of the volumetric sediment transport intensity depends thus on: the intensity of the impact of wave motion on the seabed, the asymmetry of wave motion, bed sediment grain size composition, as well as on the value of the return current.

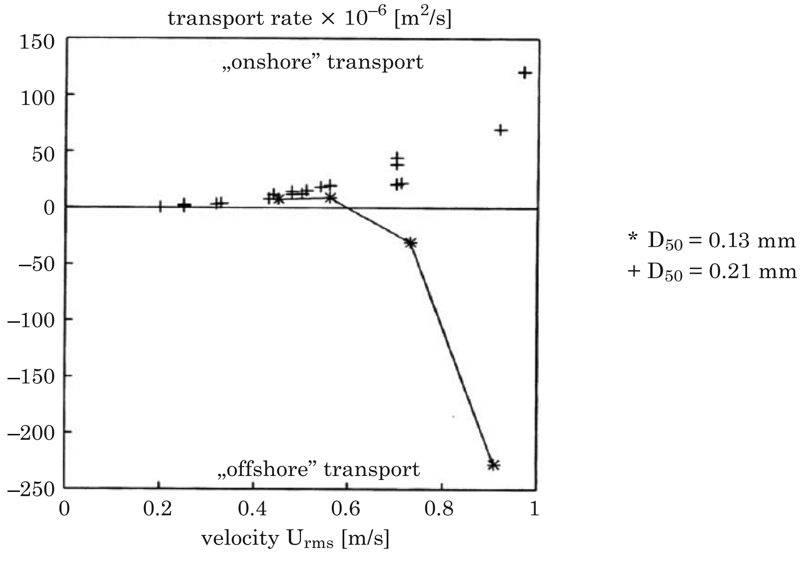

Fig. 2. Resultant sediments transport under the influence of the asymmetric waves
Source: on the basis of: (*) RIBBERINK, CHEN (1993), (+) RIBBERINK, AL-SALEM (1994); following: HASSAN (2003).

The average size of the shallow sea sediment particles decreases with increasing distance from the coast. As a result of observations (PAWLUK 1984, PAWLUK 1990) it was found, that the largest sediment sizes do occur in the area of wave shoaling and it decreases with the increase of water depth. The nature of the sediment which builds the bottom in the profile perpendicular to the shoreline is closely related to the morphology of the coastal zone. Basically, in the submerged bar throughs there are sediments with a lower median size than on the crests of submerged bars (GUILLEN, 1995), on which the sediment is not well sorted out and often characterized by bimodal particle size distributions, without shapes typical for regular bimodal distributions. Bimodal distributions with a clear dominance in the field of fine fractions and mode of coarse fractions as a form of long and flat tail – otherwise known as bimodal distributions negatively sloped – occur on slopes of submerged bars (PRUSZAK 1998). On the seaward slopes of submerged bards, the nature of sediment is usually unimodal and well sorted out. Sometimes in the distribution of sediment grain size it is possible to distinguish such a small mode within the fine fractions and dominant mode of coarse fractions (bimodal distribution positively sloped).

Modeling the sediment transportation in the flat bottom regime

The sediment transport rate, under the action of waves and currents is defined as integrated (summed by depth) product of velocity and sediment concentration. The sediment transport rate is highly variable during the period of the wave. Stationary models of sediments transport models base on the analytical or numerical solution of basic equations of momentum (amount of motion) and the continuity of the fluid and the equation mass conservation – equation of diffusion-advection. Stationary models describe the transport of bottom sediment in the simplified manner. In this case, the empirical or quasi-empirical formulas are used. They are based mainly on the characteristic parameters of the wave motion, such as: the maximum (in wave period) water velocity at the bottom or the maximum bottom friction value. Sediment transport models can be divided into three main groups, according to the criterion of how to describe the vertical structure of the sediment transport:

- models describing suspended sediment transport,
- models describing bedload sediment transport,
- two- and three-layer models, i.e. describing the transport of suspended and bedloaded sediments; possible third layer describes transition region between the layers and corresponds to the saltation layer.

Suspended sediment transport models put the emphasis on the solution of the equations of water momentum and advection – diffusion equation describing the instantaneous concentration of sediment being suspended. In case of flat bed (1DV), i.e. when only the variability of flows in vertical is taken into consideration, it is possible to express as follows:

$$\frac{\partial C}{\partial t} = \frac{\partial}{\partial z}\left(\varepsilon_s \frac{\partial C}{\partial z} + w_s C\right) \tag{1}$$

where:

$C = C(z, t)$ – instantaneous volumetric concentration of the suspended sediment [m^3/m^3],

w_s – sediment grains falling speed [m/s],

$\varepsilon_s = \varepsilon_s(z, t)$ – factor of sediment diffusion turbulence in the vertical profile [m^2/s], whereas the volumetric concentration is possible to be defined as follows:

$$C = \frac{V_s}{V_p + V_s} \tag{2}$$

where:
V_s – volume of the ground control volume,
V_p – pores volume.

To determine the sediment concentration distribution in the vertical profile it is necessary to know the size of the concentration at reference level, i.e. at the lower limit of the modeling. In simplified models, it is generally assumed that the turbulent diffusion factor is constant in time – mostly during the wave period. In addition, it is assumed its constancy throughout the whole depth range. In addition, sometimes, additional simplification may also relate to the concentration at reference level. It can be determined using empirical or semi-empirical functions, making the concentration dependent on the friction along the bottom surface (e.g. ENGELUND, FREDSØE 1976, GLENN, GRANT 1987, GRANT, MADSEN 1986, MADSEN, GRANT 1976, ZYSERMAN, FREDSØE 1994).

A detailed review of models of suspended sediment transport in the regime of flat bottom can be found in the works: AMOUDRY, LIU (2010), DAVIES, VILLARET (2002), DAVIES et al. (1997), DOHMEN-JANSSEN (1999).

In the description of sediment transport, in the bedload layer, strong influence of interactions between sediment particles causes the necessity of a different approach to the description of transport than the concept of diffusion, used in the suspension layer. BAGNOLD (1956) introduced the postulate of "dispersion stress" in the bedload layer, which is the result of sediment grains presence in the soil and water mixture and he showed that as a result of shear of soil and water layer, due to the presence of sediment grains in the mixture, there is an additional normal (causing the relaxation of water and soil mixture) and tangential stress generated. SAYED and SAVAGE (1983), using the results of experimental researches on the mechanics of granular media, proposed the constitutive equations to describe the state of stress and strain, as well as associated relations used to describe normal and tangential stresses resulting from the momentum transfer between particles of sediment on their mutual collisions.

Holistic modeling – describing transport within the whole area of sediment movement, i.e. taking into account the total of bedload and suspended sediment transport in the regime of flat bottom, requires a description based on a two-layer model (e.g. AMOUDRY et al. 2008, ASANO 1990, BAKHTYAR et al. 2010, DONG, ZHANG 2002, LI, SAWAMOTO 1995, LIU, SHEN 2010). The first two-layer models were based on the classical bagnold approach (ASANO 1990, KOBAYASHI, SEO 1985).

KACZMAREK (1991) adapted the stationary description of sediment layer motion, proposed by SAYED and SAVAGE (1983) for wave conditions. In order to determine the instantaneous value of the sediment transport rates in bedload layer, he used the instantaneous stresses at the upper surface of the bedload layer. The resulting solution of the equations concerning movement in bedload layer was in the next step "stitched" with the solution obtained in the suspension layer, and obtained on the basis of diffusion – advective model (1DV). KACZMAREK and OSTROWSKI in 2002 joined the third layer, so-called contact load layer – between a bedload and suspension layers – using for its description the proposal of DEIGAARD (1993). Deigaard's description, adapted by KACZMAREK and OSTROWSKI (2002) enabled the analysis of the momentum exchange between the elements of water and sediment grains and the mutual exchange of momentum (by way of chaotic collisions) between grains of sediment. In turn, a more simplified version of 1DV modelling, rather more similar to the model of KACZMAREK (1991) was presented by MALARKEY et al. (2003).

The primary advantage of the two- and three-layer models in comparison with classic models (formulas), based on empirical and semi-empirical relations, is the fact, that such models allows to evaluate a concentration of sediments at any level, depending on the instantaneous hydrodynamic force. Furthermore, they allow for the description of the vertical profiles of velocity and sediment concentration throughout the entire area of sediment motion.

To describe the transport of sediments in the vertical profile, the various models are used, e.g. UNIBEST-TC (BOSBOOM et al. 1998, RENIERS et al. 1995) and CROSMOR2000 (VAN RIJN 2000, VAN RIJN, WIJNBERG 1996) to describe the bedload transport they use the quasi-stationary model of RIBBERINK (1998), while for the description of the suspended transport – model of van RIJN (1993) providing the size of averaged transport volume within the period of the wave. COSMOS model (NAIRN, SOUTHGATE 1993, SOUTHGATE, NAIRN 1993) is based on the Bailard's model (1981), BEACH models (O'CONNOR, NICHOLSON 1989, O'CONNOR et al. 1998), CIRC (RIVERO, SANCHEZ-ARCILLA 1993, SIERRA, SANCHEZ-ARCILLA 1999) use the model of WATANABE (1980). These models, except the CROSMOR2000 model do not take into account the graded sediment. In CROSMOR2000 model, it is a priori assumed that some diameters of sediment grains move within the bedload layer – to the coast and other within the suspension layer – from the coast. Such an approach does not take into account the fact, that along with the increase of the waving impact on the bottom, the range of sediment diameters, transported from the coast in the suspension is changed. Among others, for this reason, CROSMOR2000 model does not describe the changes in grain size distribution in the coast cross profile which is the result of the storm activity.

Modeling of graded sediment transport

Graded sediment description is performed using the results of the granulometric analysis, while the following relationship applies:

$$\sum_{i=1}^{N} n_i = 1 \tag{3}$$

where:
n_i – value describing the percentage size (fractional) of the i-th fraction in the mixture of sediment which builds the bottom,
N – number of fractions in the mixture.

The biggest simplification used in the modeling of graded sediment transport is the assumption that the intensity of transport (q) of all fractions of the sediment is the sum of independent transport intensities ($n_i q_i$) of individual fractions (BIEGOWSKI 2006):

$$q = \sum_{i=1}^{N} n_i q_i \tag{4}$$

The values of q [m^2/s] and q_i [m^2/s] mean the grain skeleton flow intensity (excluding the porosity) of a given flow volume per unit width and time. Value of qi is the transport rate of homogenous sediment with a diameter of the i-th fraction. It is assumed that this value does not depend on the presence of other fractions in the mixture. Thus, the equation (4) means, that the simplified model of mixture was adopted, assuming no interaction between the different fractions of the sediment. In fact, the sediment fractions interact with each other and the contribution of the various factions in the intensity of sediment transport varies in relation to the situation when there is no interaction between the fractions.

HASSAN (2003) presented the possibility to implement a quasi – stationary model, proposed by BAILARD (1981), to describe the intensity of graded sediment transport. It is assumed the sediment transport intensity in the bedload layer depends on the representative diameter (median d_{50}) of the sediment of the bottom. However, in the suspension layer, because of the variable fall velocity, depending on the particle diameter, the flow of sediment transport depends on the diameter of individual sediments fractions. Thus, with respect to the suspended sediment transport rate, there is a simplification described by the equation of (4).

Within the frames of the simplification, described by the equation of (4), RIBBERINK (1998) made the volume of individual deposit fractions transport dependent on the dimensionless friction, the most frequently described by the Shields parameter, determined for each diameter of the sediment. In turn, DOHMEN-JANSSEN (1999) has adopted the RIBBERINK'S (1998) quasi-stationary model of graded sediments transport and she also introduced a modification, taking into account the effect of the phase shift between suspended sediment concentration and instantaneous velocity of water. In this way, it was possible to achieve the simplified, two-way selective description of sediments transport, wherein the coarser fractions move towards the coast, and finer fraction move away from the coast.

However, the quasi-stationary models of graded sediments transport (BAILARD 1981, RIBBERINK 1998) insufficiently describe the intensity of sediment transport due to hydrodynamic forcing. What's more, they very poorly describe the bidirectional effect of sediments transport within the coastal zone of the sea (BIEGOWSKI 2006). In the last case, in the calculations carried out with model of DOHMEN-JANSSEN (1999), it is possible to obtain better results, although the best results are obtained only when the particle size of the sediment is described with four fractions (HASSAN 2003). This description makes it obviously impossible to characterize a natural sediment, deposited on the bottom of the sea.

The three-layer model of graded sediments transport

Complete description of sediment transport, i.e. the description of grain heterogeneous sediment transport within the entire area of motion may be made on the basis of the three-layer model of graded sediments transport (KACZMAREK et al. 2004). The model assumes that the movement of sediment is carried out in three layers (Fig. 3): bedload layer, contact load layer and outer flow region, as a result of the shear stress influence on the bottom. In the area of each layer there is a different character of the deposits movement and the momentum exchange between the water and sediment particles and therefore, they are described with various equations. At the contact of layers, there is a "stitching" of solutions, so as to ensure the continuity of sediment movement description.

The model assumes that all fractions in the bedload layer move at the same speed in the form of a dense water and soild mixture and sediments sorting is carried out in this layer. It was assumed that the interactions between the sediment fractions are so strong, that finer fractions are slowed down by the coarser ones, and finally all fractions move at the same speed. Thus, this layer does not apply to the simple summation of transport flow for individual fractions, treated as the homogeneous sediment.

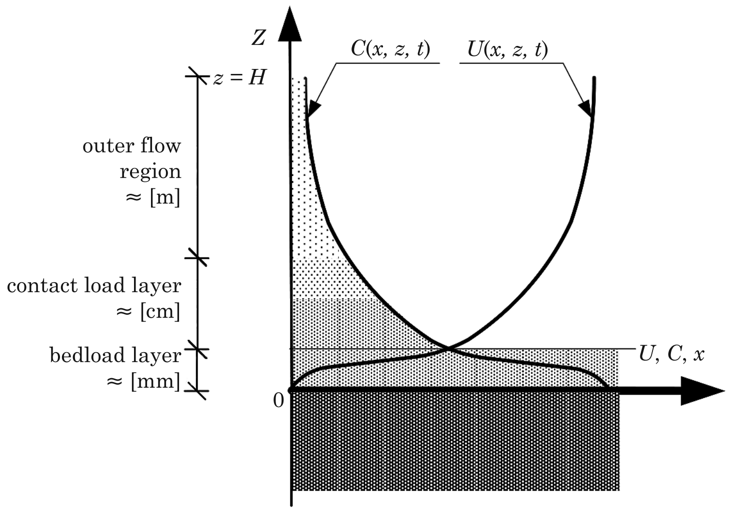

Fig. 3. Scheme of the three-layer sediment transport model

The intensity of sediment transport in the bedload layer (KACZMAREK et al. 2004) towards the coast (for the duration of wave crest) q_{bx}^{+} and away from the coast q_{bx}^{-} respectively equals (see SAWCZYŃSKI 2012, SAWCZYŃSKI et al. 2011):

$$q_{bx}^{+} = \frac{1}{T}\int_{0}^{T_c}\left(\int_{0}^{\delta_{br}^{+}} u_{br}^{+}(z',t)\ c_{br}^{+}(\mathrm{z}',\mathrm{t})\ dz'\right)dt \tag{5}$$

and

$$q_{bx}^{-} = \frac{1}{T}\int_{T_c}^{T}\left(\int_{0}^{\delta_{br}^{-}} u_{br}^{-}(z',t)\ c_{br}^{-}(\mathrm{z}',\mathrm{t})\ dz'\right)dt \tag{6}$$

where:
T_c, T – wave crest duration, wave period,
$\delta_{bx}^{+/-}$ – bedload layer thickness,
$u_{bx}^{+/-}$ – sediment velocity in the bedload layer,
$c_{bx}^{+/-}$ – sediment concentration in the bedload layer,
z' – elevation, while z' axis directed vertically down.

The mathematical modeling takes into account the fact that the most intensive vertical sorting takes place in the process of raising grains in the contact load layer over the bottom. In the contact load layer, turbulent and chaotic pulsations of collision between particles cause very strong differentiation transportation of individual fractions of the sediment. Very close to the bottom – in a sublayer, where in the distribution of the *i*-th fraction of sediments, the slip speed is strongly revealed – there is a very strong interaction between the individual fractions, resulting in mutual chaotic collisions. In a further distance from the bottom, these interactions between fractions are subject to weakening. However, the concentration of the *i*-th fraction is so high as to cause turbulence suppression, while this suppression is dependent on the grain diameter di. It was assumed, that each *i*-th fraction, as a result of interactions, moves with its own speed, and it is characterized by its own concentration. Coarser fractions velocities and concentrations, calculated in the contact load layer are greater than the size of which would have these fractions if the bottom is homogenous and made of only one, corresponding fraction. This increase in speed in the mixture is the result of interactions between fractions where coarser ones are accelerated by the finer ones.

The intensity of sediment transport in the contact load layer (KACZMAREK et al. 2004) towards the coast q_{cx}^{+} and away from the coast q_{cx}^{-} may be respectively described as (see SAWCZYŃSKI 2012, SAWCZYŃSKI et al. 2011):

$$q_{cx}^{+} = \sum_{i=1}^{N} n_i \left(\frac{1}{T} \int_0^{T_c} \left(\int_0^{\delta_{br}^{+}} u_{ci}^{+}(z,t)\ c_{ci}^{+}(\mathrm{z,t})\ dz \right) dt \right) \tag{7}$$

and

$$q_{cx}^{-} = \sum_{i=1}^{N} n_i \left(\frac{1}{T} \int_{T_c}^{T} \left(\int_0^{\delta_{br}} u_{ci}^{-}(z,t)\ c_{ci}^{-}(\mathrm{z,t})\ dz \right) dt \right) \tag{8}$$

where:

δ_{br} – contact load layer thickness,
n_i – percentage content of the *i*-th fraction,
$u_{ci}^{+/-}$ – velocity of the *i*-th fraction in the contact load layer,
$c_{ci}^{+/-}$ – concentration of the *i*-th fraction in the contact load layer.

In the outer layer, over the contact load layer, it is assumed that there is no change in the particle size distribution of the transported sediment. The vertical distribution of concentration in this layer is described by a power function.

The transport flow in the outer layer (KACZMAREK et al. 2004) can be expressed with the following formulas (see SAWCZYŃSKI 2012, SAWCZYŃSKI et al. 2011):

$$q_{ox}^{-} = \int_{\delta_{cr}}^{H} U_{ox}^{-}(z) C_0(z) dz \tag{9}$$

$$q_{oy} = \int_{\delta_{cr}}^{H} U_{oy}(z) C_0(z) dz \tag{10}$$

where, C_0 means the averaged in time value of suspended sediment mixture concentration and U_{ox}^{-} and U_{oy} are current velocities, resulted from waving (in the coastal zone of the sea) of the respectively return current (directed towards the sea) and alongshore current, generated as a result of waves propagation at the angle to the cross profile of the coast.

BIEGOWSKI (2006) in his doctoral dissertation, using a three-layer model of graded sediments transport, conducted, among others, the analysis of grain distribution variability on the profile, exhibiting at the same time a usefulness of the model. He explained the essence of formation of a various grain distributions (unimodal and bimodal) during a short-term storm, depending on the position of the calculation point on the profile (from the sea or from the coast) and the occurring phenomenon of erosion or accumulation as a result of the bi-directional stream of deposits. This bi-directional flow is related to the transport of sediment at the bottom, faced towards the coast as a result of the asymmetry of wave motion and transport of sediments carried away into the sea by the return current.

Summary and conclusions

As a result of the review of sediments transport models, presented in this paper, it has been indicated that the primary advantage of the two- and three-layer models in comparison with classic models (formulas), based on empirical and semi-empirical relations, is the fact, that such models allows to evaluate a concentration of sediments at any level, depending on the instantaneous hydrodynamic force. Furthermore, they allow for the description of the vertical profiles of velocity and sediment concentration throughout the entire area of sediment motion. The three-layer theoretical model of graded sediments transport details the bedload, contact load layer and outer flow region. Character of interactions between water and sediments is different in each of the above layers and that is why they are described in different equations, while at the contact between the layers there is a stitching of solutions, providing a complete theoretical description of the structure of transportation of the sandy sediment of a various grain size. Using this model, it is possible to predict an amount of moving sandy sediments of various grain size, as well as there is a possibility to predict the size of the transport of individual fractions. As a result, the knowledge of grain size may be crucial, e.g. in planning and conducting works associated with artificial sand supply. The silt material extracted from the navigation channel is often used to reinforce the edge in the vicinity of ports and therefore knowledge of the particle size distribution is very important. Based on the three-layer model of graded sediment transport, it is possible to determine the bathymetry changes taking place under the influence of variations in the flow of sediment transport of non-uniform grain size, and to determine the mutual impact of changes in grain size of sediments and evolution of the bottom profile.

References

ALLEN J.R.L. 1977. *Fizyczne procesy sedymentacji.* PWN, Warszawa.

AMOUDRY L., HSU T.J., LIU P.L.-F. 2008. *Two-phase model for sand transport in sheet flow regime.* Journal of Geophysical Research, 113, C03011, doi:10.1029/2007JC004179.

ASANO T. 1990. *Two-plane flow model on oscillatory sheet-flow*. Proc. 22nd Int. Conf. Coast. Eng., ASCE, p. 2372–2384.

BAGNOLD R.A. 1956. *The Flow of Cohesionless Grains in Fluids*. Phil. Trans. Roy. Soc., Ser. A, 249: 235–297.

BAILARD J.A. 1981. *An energetics total load sediment transport model for a plane sloping beach.* J. Geophys. Res., 86(C11): 10938–10954.

BAKHTYAR R., BARRY D.A., YEGANEH-BAKHTIARY A., LI L., PARLANGE J.-Y., SANDER G.C. 2010. *Numerical simulation of two-phase flow for sediment transport in the inner-surf and swash zones*. Advances in Water Resources, 33(3): 277–290.

BIEGOWSKI J. 2006. *Dynamika osadów morskich o niejednorodnym uziarnieniu w świetle teorii i eksperymentu*. PhD thesis, IBW PAN.

BOSBOOM J., KLOPMAN G., RENIERS A, STIVE M.J.F. 1998. *Analytical model for wave-related sediment transport in flat bed sheet flow regime*. Proc. 26th ICCE, ASCE, Reston VA, p. 2573–2586.

CHEN X.W., ZHENG J.H., ZHANG C., YANG X. 2010. *Evaluation of diffraction predictability in two phase averaged wave models*. China Ocean Engineering, 24(2): 235–244.

DAVIES A.G., RIBBERINK J.S., TEMPERVILLE A., ZYSERMAN J.A. 1997. *Comparisons between sediment transport models and observations made in wave and current flows above plane beds.* Coastal Engineering, 31: 163–198.

DAVIES A.G., VILLARET C. 2002. *Prediction of sand transport rates by waves and currents in the coastal zone.* Cont. Shelf Res., 22(18–19): 2725–2737.

DEIGAARD R. 1993. *Modelling of sheet flow: dispersion stresses vs. the diffusion concept*. Prog. Rep., 74, Inst. Hydrodyn. and Hydraulic Eng., Tech. Univ. Denmark, p. 65–81.

DOHMEN-JANSSEN M.C. 1999. *Grain size influence on sediment transport in oscillatory flow.* FEBO-DRUK BV, Enschede, The Netherlands.

DOHMEN-JANSSEN C.M., HANES D.M. 2002. *Sheet flow dynamics under monochromatic nonbreaking waves.* Journal of Geophysical Research, 107(C10), doi:10.1029/2001JC001045.

DONG P., ZHANG K., 2002. *Intense near-bed sediment motions in waves and currents.* Coastal Eng., 45(2): 75–87.

DRUET C., KOWALIK Z. 1970. *Dynamika morza*. Wydawnictwo Morskie, Gdańsk.

ENGELUND F., FREDSØE J. 1976. *A sediment transport model for straight alluvial channel.* Nordic Hydrology, 7.

GLENN S.M., GRANT W.D. 1987. *A suspended sediment stratification correction for combined wave and current flows*. J. Geophys. Res., 92(C8): 8244–8264.

GRADZINSKI R., KOSTECKA A., RADOMSKI A., UNRUG R. 1986. *Zarys sedymentologii*. Wydawnictwo Geologiczne, Warszawa.

GRANT W.D., MADSEN O.S. 1986. *The continental-shelf bottom boundary layer.* Annu. Rev. Fluid Mech., 18: 265–305.

GUILLEN J. 1995. *Sediment response in a littoral system, Island of Terschelling, The Netherlands.* Report 90–20, Dep. Of Phys. Geography, Univ of Utrecht, The Netherlands.

HASSAN W.N. 2003. *Transport of size-graded and uniform sediments under oscillatory sheet-flow conditions*. PhD Thesis, Univ. of Twente, The Netherlands.

JANSSEN C.M. 1995. *Sand Transport in Oscillatory Sheet-flow. A literature review*. Communications on Hydraulic and Geotechnical Engineering, Rep. No. 95–6, Delft Univ. of Technology, Faculty of Civil Engineering.

KACZMAREK L.M. 1991. *Mathematical model for oscillating sheet flow*. Proc. Euromech 262 Colloquium on Sand Transport in Rivers, Estuaries and the Sea, A.A. Balkema/Rotterdam/Brookfield, p. 197–202.

KACZMAREK L.M. 1999. *Moveable Sea Bed Boundary Layer and Mechanics of Sediment Transport.* IBW PAN, Gdańsk.

KACZMAREK L.M., BIEGOWSKI J., OSTROWSKI R. 2004. *Modelling cross-shore intensive sand transport and changes of grain size distribution versus field data.* Coastal Engineering, 51: 501–529.

KACZMAREK L.M., OSTROWSKI R. 2002. *Modelling intensive near-bed sand transport under wave-current flow versus laboratory and field data*. Coastal Engineering, 45(1): 1–18.

KOBAYASHI N., SEO S.N. 1985. *Fluid and sediment interaction over a plane bed*. J. Hydraul. Eng., 111(6): 903–921.

LI L., SAWAMOTO M. 1995. *Multi-phase model on sediment transport in sheet-flow regime under oscillatory flow*. Coastal Eng. Jpn., 38(2): 157–178.

LIU C., SHEN Y.M. 2010. *A three dimensional* k-ε-Ap model for water-sediment movement. International Journal of Sediment Research, 25(1): 17–27.

MADSEN O.S., GRANT W.D. 1976. *Quantitative description of sediment transport by waves*. Proc. 12th Int. Conf. Coast. Eng., ASCE, p. 1093–1112.

MALARKEY J., DAVIES A.G., LI Z. 2003. *A simple model of unsteady sheet flow sediment transport.* Coastal Eng., 48(3): 171–188.

NAIRN R.B., SOUTHGATE H.N. 1993. *Deterministic profile modeling of nearshore processes.* Part 2. *Sediment transport and beach profile development*. Coastal Eng., 19: 57–96.

NIELSEN P. 1992. *Coastal bottom boundary layers and sediment transport.* Advanced Series on Ocean Engineering, 4.

O'CONNOR B.A., NICHOLSON J. 1989. *Modelling changes in coastal morphology*. In: *Sediment Transport Modeling*. Ed. S.S.Y. Wang. ASCE, pp. 160–165.

O'CONNOR B.A, PAN S., NICHOLSON J., MACDONALD N., HUNTLEY D.A. 1998. *A 2D model of waves and undertow in sufr zone*. Proc. Coastal Eng. ASCE, p. 286–296.

O'DONOGHUE T., WRIGHT S. 2004. *Flow tunnel measurements of velocities and sand flux in oscillatory sheet flow for well-sorted and graded sands*. Coastal Engineering, 51(11–12): 1163–1184.

PAWLUK K. 1984. *Segregacja materiału osadowego*. Prace wewnętrzne IBW PAN, Gdańsk.

PAWLUK K. 1990. *Materiał piaszczysty strefy brzegowej*. Studia i Materiały Oceanograficzne, 55.

PRUSZAK Z. 1998. *Dynamika brzegu i dna morskiego*. IBW PAN Gdańsk.

RENIERS A.J.H.M., ROELVINK J.A., WALSTRA D.J.R. 1995. *Validation study of UNIBEST-TC; validation against the LIP 11D experiment*. Report H2130, Delft Hydraulics.

RIBBERINK J.S. 1998. *Bed load transport for steady flows and unsteady oscillatory flows*. Coastal Engineering, 34: 59–82.

RIBBERINK J.S., AL-SALEM A. 1994. *Sediment transport in oscillatory boundary layers in cases of rippled beds and sheet flow*. Journal Geoph. Res., 99(C6): 2707–2727.

RIBBERINK J.S., CHEN Z. 1993. *Sediment transport of fine sand under asymmetric oscillatory flow*. Report H840, part VII, WL-Delft Hydraulics.

RIVERO F.J., SANCHEZ-ARCILLA A. 1993. *Propagation of linear gravity waves over slowly varying depth an currents.* In: *Waves'93* Eds. O.T. Magoon, J.M. Hamsley, Proc. 2nd International Symposium Ocean Wave Measurement and Analysis, New Orleans, ASCE, p. 518–532.

SAWCZYŃSKI S. 2012. *Bathymetry changes and sediment sorting within coastal structures: a case of the silting-up of navigation channels*. PhD thesis, University of Technology in Koszalin (in Polish).

SAWCZYŃSKI S., KACZMAREK L.M., BIEGOWSKI J. 2011. *Bathymetry changes and sand sorting during silting up of the channels*. Part 2. *Modelling versus laboratory data*. Technical Sciences, 14(2): 171–192.

SAYED M., SAVAGE S.B. 1983. *Rapid gravity flow of cohesionless granular materials down inclined chute.* J. Applied Mathematics and Physics (ZAMP), 34: 84–100.

SIERRA J.P., SANCHEZ-ARCILLA A. 1999. *CIIRC-LIM runs for Egmond Pilot Experiment.* Report RR-CIIRC/AHC-99-1, University of Catalunya, Spain.

SOUTHGATE H.N., NAIRN R.B. 1993. *Deterministic profile modelling of nearshore processes.* Part 1. *Waves and currents.* Coastal Engineering, 19(1–2): 27–56.

SVENDSEN I.A. 1984. *Mass flux and undertow in a surf zone*. Coastal Engineering, 8: 347–365.

ZHENG J.H. 2007. *Depth-dependent expression of obliquely incident wave induced radiation stress.* Progress in Natural Science, 17(9): 1067–1073.

ZHENG J.H., MASE H., DEMIRBILEK Z., LIN L. 2008. *Implementation and evaluation of alternative wave breaking formulas in a coastal spectral wave model.* Ocean Engineering, 35(11–12): 1090–1101.

VAN DER A D.A., O'DONOGHUE T., RIBBERINK J.S. 2010. *Measurements of sheet flow transport in acceleration-skewed oscillatory flow and comparison with practical formulations*. Coastal Engineering, 57Ł 331–342.

VAN RIJN L.C. 1993. *Principles of sediment transport in rivers, estuaries and coastal seas*. Aqua Publications, The Netherlands.

VAN RIJN L.C. 2000. *General view on sand transport by currents and waves*. Z2899.30, Delft Hydraulics, Delft, The Netherlands.

VAN RIJN L.C., WIJNBERG K.M. 1996. *One-dimensional modelling of individual waves and wave-induced longshore currents in the surf zone*. Coastal Engineering, 28: 121–145.

WATANABE A., RIHO Y., HORIKAWA K. 1980. *Beach profiles and on-offshore sediment transport.* Proc. 17 Conf. Coastal Eng. ASCE.

ZYSERMAN J.A., FREDSØE J. 1994. *Data analysis of bed concentration of suspended sediment.* J. Hydraul. Res., 120(9): 1021–1042.