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A Stratigraphic Soil Model for Coastal Morphodynamics

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ABSTRACT: Morphodynamic processes in coastal areas are strongly influenced by tidal forcing. The constant change in current velocities leads to a regular alternation of erosion and deposition of sediments. To model the transport of these sediments under tidal conditions, it is important to describe the spatial and temporal variation of the sediment with the appropriate level of detail. The soil model presented here classifes the soil according to differences in sediment distribution. It is grouped in soil textures, which describe the sediment distribution according to the amounts of e.g. sand, silt and clay. In contrast to common morphodynamic models, an exchange layer is not used to describe transport processes, because the information is stored in the layer stratigraphy. First results for a 1D-Z model with a depositional setting are shown.

Keywords: Coastal morphodynamics, Stratigraphy, Sediment transport

1 INTRODUCTION

1.1 *Characteristics of sediment transport in coastal areas*

Morphodynamics in coastal areas is to a large part driven by the energy of the tide. Under the tidal influence current velocities in the estuaries are changing constantly, interrupted by brief periods of very low, near zero velocities. Maximum velocities can go up to 2 *m/s* in the fairway, while being considerably lower in shallow water regions like tidal flats or side arms. Since the hydrodynamic conditions are the main driving force for sediment transport and morphodynamic processes, a periodic alternation of erosion and deposition is common in tidally influenced coastal areas.

Therefore estuaries are a very dynamic and diverse system with regards to sediment distribution and movement. In coastal areas and tidal flats fine sands are quite common and bedload transport plays an important role (Albers, 2012). Inside the estuaries finer sediments transported in suspension are more dominant. A common feature of most estuaries is the existence of a turbidity zone with sediment concentrations ranging from a few hundred *mg/l* up to a few *g/l*. These sediments are to a large part settling during slack water and are then resuspended. In all German estuaries this sequence happens twice a day. Meanwhile tidal flats and side arms tend to be depositional zones due to the lower current velocities. Sediment is transported there in suspension but once it has settled in these areas, there is not sufficient energy available to resuspend the full amount which leads to net deposition.

1.2 *Development of a stratigraphy due to tidal forcing*

This behaviour can lead to characteristic depositional patterns in areas with low currents and a high sediment supply. These patterns are called tidal rhythmites and have been found in many geological studies in areas that still are or were under tidal influence. They are caused by different sediment distributions and amounts being transported during different phases of the tidal cycle. Sediment cores with such a stratigraphy can provide information about the tidal characteristics during the time of deposition (Mazumder 2005).

The diurnal inequality of the tide leads to a difference in the amounts of sediment that is deposited, which is reflected in different layer thicknesses with varying sediment distribution. During the dominant part of the cycle, the share of coarser sediments is higher due to more energy being available for transport. The same can be said regarding the neap-spring cycle and the different amounts of tidal energy leading to different thicknesses of the sediment layers. But such a pattern can only develop in areas that are sheltered from high erosion events that would break up these structures and that are supplied with a large amount of sediment. Daily depositional heights of 1 to 1.5 mm per day can occur under these situations. Fig. 1 shows a schematic example derived during one study (Choi, 2000).

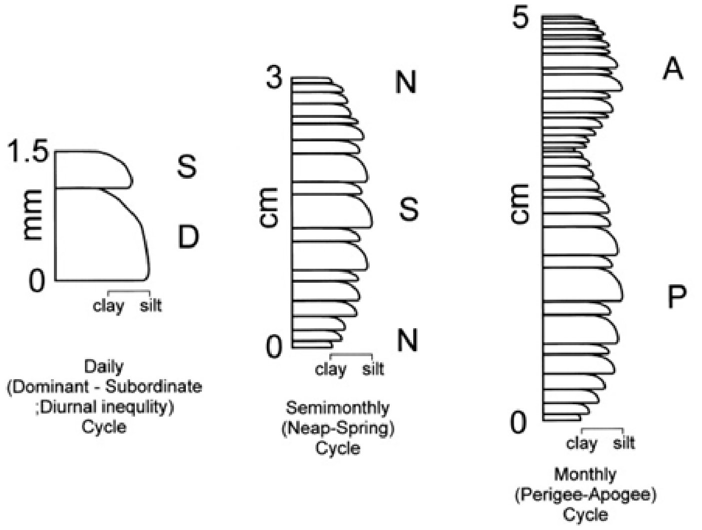

Figure 1. A Schematic principle of the formation of rhythmites and the influence of different periods of the tidal cycle on layer thickness and composition (Choi, 2000).

2 NUMERICAL REPRESENTATION OF TIDAL SOILS

2.1 *Challenges in describing a changing soil stratigraphy*

To model the sediment transport in tidally influenced areas correctly, it is important to store the temporal variation of deposited sediments, so that during erosion phases lastly deposited sediments are being eroded first. Since the magnitude of these transport processes can vary spatially and temporarily, creating a user-defined vertical discretization that displays the stratigraphy sufficiently is very hard to do. In most existing models an unphysical mixing of different sediments in the upper layer occurs, which leads to a loss of information. When two distinct deposition events with different sediment characteristics occur and the information is stored in just one user-defined layer where it is perfectly mixed, the information about the chronological sequence of the deposition is lost.

The soil model presented here is able to create a vertical discretization that is based solely on the sediment inventory and the differences in its vertical distribution. The goal is to represent the vertical variation of sediments over time and space as unaltered as possible. This principle is visualized in fig. 2.

The model is capable of deciding when the changes of the distribution of new sediments are distinctive enough that they should be deposited in a new layer so no unwarranted mixing occurs.

2.2 *Classifying layer structure according to changes in the sediment distribution*

The sediment and its locally varying distribution is described with a set of classifications. First, there are soil classes. These are groups in which the sediment fractions are sorted based on their grain size. They consist of upper and lower boundaries for the grain size, so each sediment fraction can be assigned to a soil class. Typical soil classes are sand, silt und clay. For each soil class, all the individual masses of the associated fractions are summed up. Second, there are soil textures. They describe the mass distribution of the soil classes and are defined by a validity range for each soil class. E.g. the soil texture sandy silt describes a distribution of sediments in the range of 50 – 80 % silt, 12 – 50 % sand and 0 – 8 % clay.

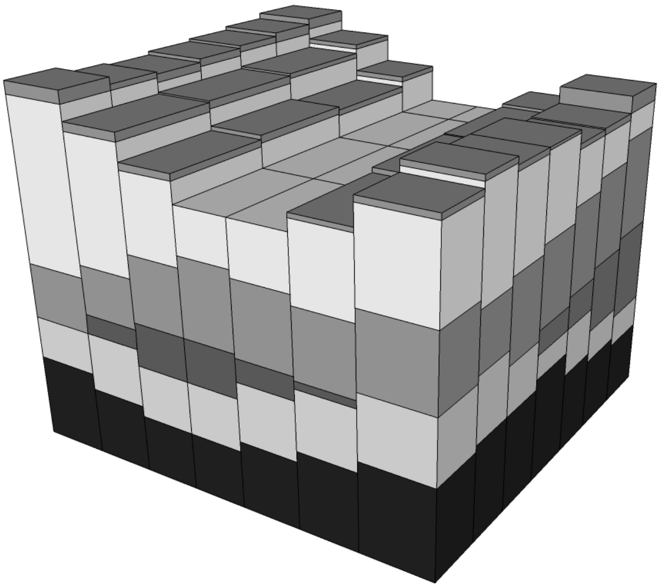

Figure 2. Schematic view of a soil in the model presented here. Brighter colors indicate a higher fraction of sands. The soil can be structured in each grid element with the number of layers that are necessary to describe the vertical differences in sediment distribution.

The sediment distribution of each layer as well as that of the depositional flux can be described with this classification. As long as the composition of the depositional flux has the same soil texture as the one of the top layer, the sediment is deposited in that layer. But if the soil texture of the depositional flux differs from that of the top layer, a new layer, in which the sediment is deposited, is created. By that rule, the chronological sequence of the sediment distribution is stored with very little loss of information.

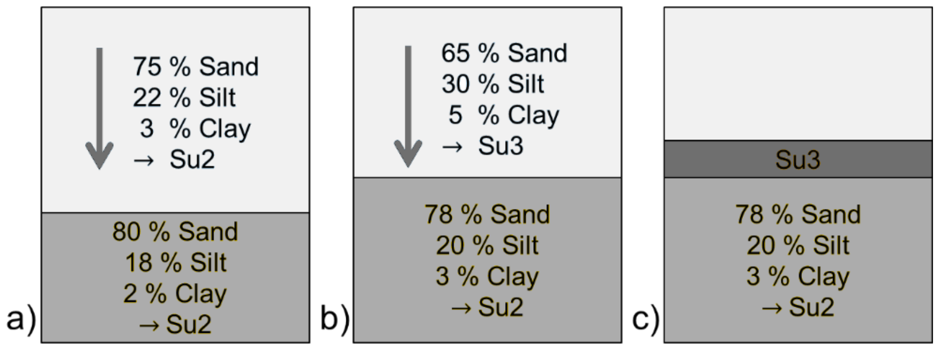

Figure 3. Formation of a new layer: a) Deposition of sediments that differs only slightly from the sediment in the top layer (Su2 = soil texture poor silty sand); b) Deposition occurred in existing layer; Sediment distribution of the depositional flux differs enough from the sediment in the top layer (Su3 = medium silty sand); c) Deposition occurred in new layer.

For data storage purposes a maximum amount of possible layers has to be specified. If this number is reached, two layers have to be merged. This is not done by simply merging the two oldest layers, but by looking for two adjacent layers with a similar sediment distribution, so the stratigraphy remains in its original state as long as possible. This concept is visualized in fig. 4.

A significant difference towards existing morphodynamic models is the absence of a transport or exchange layer. The main function of an exchange layer is to describe the sediment available for transport processes and to provide the basis for the calculation of important variables that describe the interaction between water column and soil, mostly bed roughness and resulting shear stresses. It serves as a thin layer which keeps track of short-period sediment transport processes (Malcherek 2005).

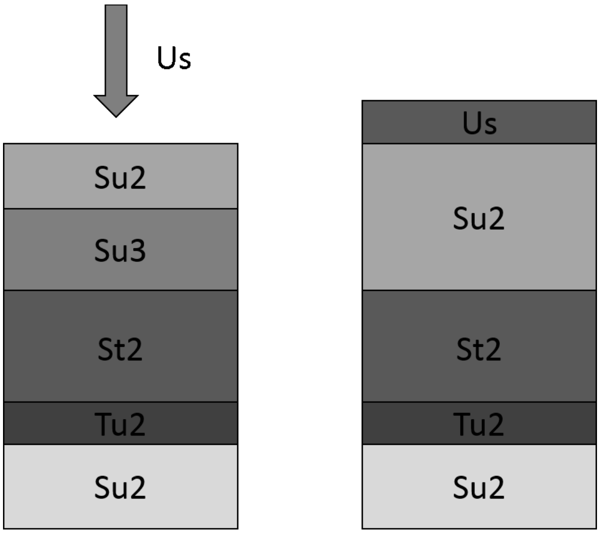

Figure 4. Concept for merging layers when maximum number of layers is reached. The two adjacent layers with the most similarity in sediment distribution in the soil column are merged.

This structure is also limiting the sediment available for erosion per time step, because only the sediment in the exchange layer is available for transport. If the calculated erosion mass is higher, it is reduced to the amount available in the exchange layer. There are approaches to determine the thickness of such layers from sediment parameters like the d50 as well as hydrologic parameters like current velocity or shear stress. The idea behind these approaches is to calculate how deep the energy of the water current reaches into the soil (Warner et al. 2008).

In the model presented here such a thin transport layer is not implemented because the changes in the sediment distribution are accurately described in the layer structure. Therefore the erosion process has to be treated differently from existing models. It is solved as an iterative process in case that there is not enough sediment available in the top layer to match the calculated erosion mass of one of the used sediment fractions for the time step. Instead of reducing the eroded sediment mass, the time is calculated that is necessary to erode all the available mass of the sediment fraction in the top layer with the calculated erosion rate. Now this time span is used to calculate the erosion mass for each fraction according to their respective erosion rates. Then the properties of the top layer are updated and the process is repeated for the remaining duration of the time step.

2.3 *Mass balances and sediment fluxes*

For balancing purposes sediment masses are used. A decrease of sediment mass can only occur in the top layer through erosion. The sediment is either moved to the water column in case of suspended sediments or to an adjacent grid element in case of bedload material. An increase of sediment mass can happen in the top layer through deposition caused either by settling or bedload transport.

If two layers are merged and it has been determined which adjacent layers are going to be merged according to the sequence detailed in fig. 4, the masses of each fraction of the two layers are added to form a new layer.

The thickness of each layer is the result of the sum of the mass of all sediment fractions, their grain densities and how dense they are stored which is described by the porosity. Therefor a change in porosity by consolidation processes leads to a change in layer thickness without affecting sediment mass balance (Sanford 2008).

$$\Delta z_{ilay} = \frac{\sum_{ised=1}^{Nsed} m_{ilay}^{ised}}{A \times rho_s^{ised} \times (1-n)} \tag{1}$$

where Δz = layer thickness, m = sediment mass, A = area of grid element, rho_s = grain density, n = porosity, $Nsed$ = maximum number of sediment fractions and *ised* and *ilay* counters for number of sediment fraction and layers.

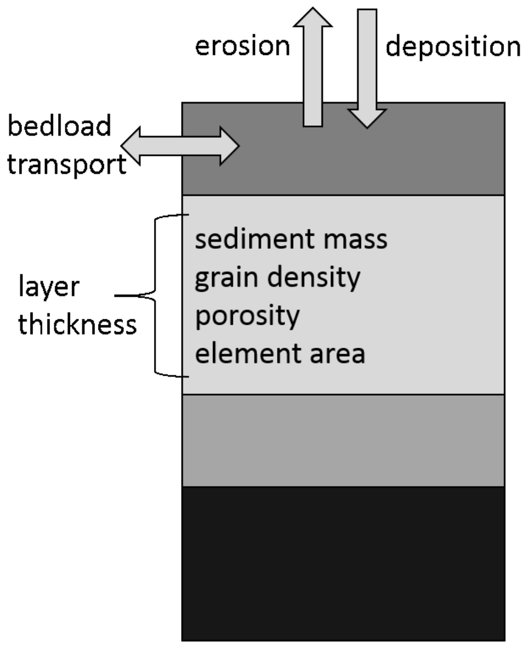

Figure 5. Mass fluxes and stored layer information for calculation of soil evolution.

The total soil elevation at a certain grid element is then the result of the sum of all layer thicknesses starting from a non-erodible starting point.

$$z_{top} = z_0 + \sum_{ilay=1}^{Nlayer} \Delta z_{ilay} \tag{2}$$

where z_{top} = top elevation, z_0 = non-erodible horizon, *Nlayer* = maximum number of layers.

3 APPLICATION OF THE SOIL MODEL

3.1 *The 1D-Z Model*

A 1D-Z model was used to test the discretization algorithms that govern the creation and merging of the layers. The soil is structured according to the outlined principles while the water column consists of one cell. Each sediment fraction has a constant settling velocity assigned to it, which is used to calculate the depositional flux for each fraction.

$$\Phi depo_{ised} = c_{ised} * sv_{ised\ ilay} \tag{3}$$

where Φ_{depo} = depositional flux, c = sediment concentration in water column, sv = settling velocity.

3.2 *Model setup for a depositional setting*

A setup was used that resembles the condition of a sheltered depositional area in which rhythmites can develop. Four sediment fractions are used, that are all transported via suspension. Their characteristics are detailed in table 1. The stated settling velocities are calculated according to Stokes. In the initial state there is one layer, which consists of the same mass of all four sediment fractions.

Table 1. Properties of the used sediment classes

Name	Grain size [*mm*]	Settling velocity [*mm/s*]	Soil class
very fine sand	0,094	7,8	Sand
coarse silt	0,0465	1,9	Silt
medium silt	0,0235	0,518	Silt
coarse clay	0,003	0,0081	Clay

The hydrodynamic forcing is created with a regular sinus function for the bed shear stress with a period of 12 hours. The maximal amplitude varies over a span of two weeks to simulate a neap spring cycle with maximum shear stresses of 0.2 N/m^2 during neap tides and 0.4 N/m^2 during spring tides. In addition to the erosion of the existing sediment in to the water column, additional sediment is added dependent on the shear stress to simulate the transport of sediment from a high erosion region to this depositional area. The sand fraction is only added during shear stresses higher than 0.25 N/m^2 while the other fractions are added while the shear stress is higher than 0.15 N/m^2. Two neap spring cycles are simulated.

3.3 *Results*

Fig. 6 shows the resulting layer structure at the end of the simulation period. The two spring tide periods are visible through a higher amount of sandy sediments that is stored in separate layers for each tidal cycle during that period. These periods can be seen around depths of 0.09 *m* and 0.13 *m*. During the spring tide periods greater thicknesses are visible, resulting directly from a greater amount of sediment in the water column. The result is a similar structure like the one depicted in Fig. 1.

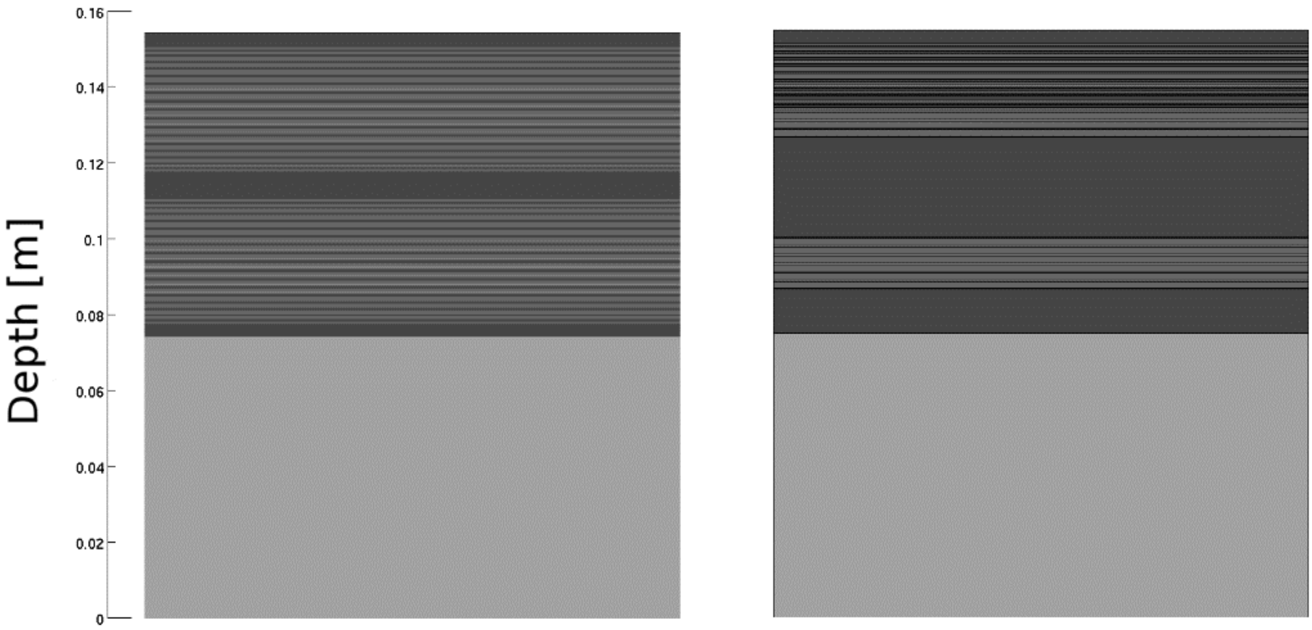

Figure 6. Results from simulation of depositional area after two neap spring cycles. Brighter layers contain a higher fraction of sand. Left picture shows a simulation with a maximum number of layers to store all information. The right picture shows a simulation with a maximum number of layers of 50. Merging of layers has taken place.

In the right picture of fig. 6 the user-defined maximum number of layers is reached during this simulation, which leads to a compression of the first neap-spring cycle. Only the more diverse information containing the sandy depositions during spring tide are preserved, while layers containing mostly silts with different amounts of sand are merged to form larger blocks of silty sediments during the periods of neap tides.

4 CONCLUSIONS

The soil model presented here is capable of forming a soil stratigraphy created by a tidal forcing. The sequence of the temporal variation in the distribution of the deposited sediments is plausible.

Additional work has to been done on layer formation during erosion processes, since armouring of the bed due to the erosion of finer sediments is not accounted for so far.

NOTATION

Φ_{depo}	depositional flux [kg/ms^2]
c	sediment concentration [kg/m^3]
sv	settling velocity [m/s]
Δz	layer thickness [m]
m	mass of sediment fraction [kg]

A	area of grid element [m^2]
rho_s	grain density [kg/m^3]
n	porosity [/]
$Nsed$	maximum number of sediment fractions [/]
$Nlayer$	maximum number of layers [/]
z_{top}	top elevation [m]
z_0	non-erodible horizon [m]

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