Deltares

Morphodynamic modelling over alluvial and non-alluvial layers

Literature review, update to Tuijnder concept

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Executive summary

Rijkswaterstaat is responsible for managing the river system in the Netherlands. As such, it is of crucial importance to predict morphodynamic changes. In the first part of this report, an overview and historical perspective of the possible approaches for predicting morphodynamic changes is given. This aims to help river managers in discerning which method to use depending on their specific needs.

One particular morphodynamic process related to fixed layers is currently of special relevance for *Rijkswaterstaat* . Modelling of this process using DELFT3D is currently done in a simplified manner which is not able to capture some particularly important features. A model that aims at improving the modelling of this process was developed, but it presented several shortcomings. In the second part of this project, this model is carefully analyzed and the shortcomings clearly identified. Based on this, an alternative model is proposed and tested against laboratory data and a field case. The model satisfactorily predicts morphodynamic development in the presence of fixed layers.

Points for improvement of the model are identified and further testing against other laboratory data is recommended. In a subsequent project, the model developed here will be applied to a field case relevant for *Rijkswaterstaat* .

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1 Introduction

Rijkswaterstaat aims at accurately predicting the morphodynamic development in the Rhine branches for the sake of safety against flooding and allowing a navigable river, among other things. There are several different approaches for predicting morphodynamic development. In this report, we first provide an overview of the different approaches (Section 2). The review does not claim to be fully comprehensive, as such a full review falls beyond the scope of this project, but rather provides the reader with the basic understanding of morphodynamic prediction.

A particular physical process that is necessary to consider under some circumstances for accurately predicting morphodynamic development is the formation and break-up of immobile sediment layers (also known as “semi-fixed layers”). Immobile sediment layers develop by vertical sorting processes in the top layer of the bed. Under low-flow conditions, only the finest sediment-size fractions present at the bed surface are mobile. Winnowing and transport of fine sediment causes the formation of a layer of coarse sediment over which fine sediment is transported. During high-flow events, these coarse layers can break-up, suddenly entraining sediment from below. Immobile layers are considered relevant for predicting stability of the river bed in the bifurcation areas of the Rhine, in the Meuse (mostly Common Meuse) and for “smart nourishment” operations.

Tuijnder and Ribberink (2010); Tuijnder *et al.* (2011); Tuijnder and Ribberink (2012) developed a model for predicting the formation and break-up of immobile sediment layers. This concept is an extension of the classical active-layer model Hirano (1971) used for considering mixed-size sediment processes. Implementation of this model into the software package DELFT3D has not given satisfactory results and the reasons for this are unclear.

In the second part of this report (Section 3), the limitations of the existing model for immobile sediment layers are unravelled for subsequently proposing and testing an alternative model. The alternative model is applied to the flume experiments of Struiksma (1999) and the field case at the fixed layer at Nijmegen. The alternative model shows that it has the limit of the Struiksma (1999) concept for partially immobile conditions and Hirano (1971) when all sediment is mobile.

2 Methods to determine morphodynamic development

In this section, different methods to determine morphodynamic development are presented, namely, expert knowledge, scale models, sediment mobility, and subsequently short term and long term responses of the river bed.

2.1 Expert knowledge

The first source of knowledge about the future of a river that became available was the judgement of an expert. River management has strongly depended on the experience of people for deciding on interventions since *Rijkswaterstaat* foundation. About the first engineers in *Rijkswaterstaat* , Bosch (2014) mentions that:

> Another shared characteristic was the impressive empirical knowledge and know-how that these experts had accumulated in practice. This knowledge was not only fundamental for sustaining routine water management, but also served as the foundation of a broad, national knowledge system and was the future basis of the professionalization of centralized water management.

As late as at the beginning of the 20$^{\mathrm{th}}$ century during the planning of the Zuiderzee reclamation, *Rijkswaterstaat* Engineer H. E. de Bruijn (1841–1915) said that the high tide water level would double (De Bruijn, 1911). The opinion was not based on experimental work or on calculations. As De Bruin mentioned, “one has to sense it, as it were, based on experience gained elsewhere and on relevant research” (Disco and Van den Ende, 2003).

A drawback of expert knowledge is that extrapolation of results to different conditions from those under which knowledge is obtained is difficult. Similarly, it is difficult to predict the effect of interventions that have never taken place in the past. The lack of method makes difficult to convince other experts and reach consensus. Despite of the great achievement of the first *Rijkswaterstaat* engineers, they were unable to arrive at a consensus regarding river interventions due to, in part, a lack of understanding of the river dynamics (Bosch, 2014). An essential component for reaching consensus is reproducibility of results, which is inherently impossible if based on knowledge expert, as there are never to cases which are the same.

Within the current framework of the Rivierkundig Beoordelingskader, a river manager has the option to determine to which degree the impact of a measure should be researched, and this could be based on expert knowledge.

2.2 Scale models

[1] A new paradigm in morphodynamic predition arose with the use of scale models to help in the design of interventions. The work by Fargue (1894) was an early example of the use of scale models. Fargue (1894) conducted 21 mobile bed experiments in an approximately 60 m long outdoor curved flume to generalize the observations he had

[1]This section is an excerpt from Chavarrías (2019).

done on the Garonne river as regards to flow in bends (see also Hager (2003)). In the Netherlands the increase in use of laboratory experiments occurred hand in hand with the foundation of the *Waterloopkundig Laboratorium* (WL | Delft Hydraulics) by Dr. Johannes Th. Thijsse (1893–1984) in 1927 (Vreugdenhil *et al.*, 2001). An example is the scale model of the Dutch Rhine-Meuse branches constructed in the centre of Delft in the 1950's (Figure 1). The insight from these scale experiments was crucial in providing understanding of the processes underlying fluvial dynamics as well as engineering solutions to water problems (e.g. Disco and Toussaint, 2014). Drawbacks of scale experiments are the cost in terms of space, time, and labour, and the fact that scale models cannot easily be modified. More importantly, a scale model generally suffers from scale effects, as it is technically difficult to keep all ratios between the relevant forces in the prototype (e.g., inertia, gravity, viscosity, surface tension, pressure, *et cetera*) equal to the equivalent ratios in the scale model. Furthermore, when the same fluid is used in the model and in the prototype, as usually occurs in morphodynamic laboratory experiments, only one ratio between forces can be identical and scale effects are unavoidable (Heller, 2011).

Figure 1 Picture of the scale model of the Dutch Rhine-Meuse branches built by the *Waterloopkundig Laboratorium* in the centre of Delft in 1950. The model was built in the Schuttersveld and the picture was taken from Het Raam. The church on the top right corner is the Lutherse Kerk (also known as Saint George's Chapel) and the windmill on the top left is Molen de Roos. Flow goes from bottom to top. The right-hand branch is the Lek. The second branch starting to count from the right is the Waal. The third one is the Meuse. The upstream boundary is approximately at Wijk bij Duurstede and Tiel. The Biesbosch is visible in the centre of the domain on the left.

2.3 Sediment mobility

2.3.1 Initiation of motion

A crucial step forward in morphodynamic prediction was conducted by Shields (1936), who conducted a broad experimental study to understand the conditions under which sediment starts moving (Figure 2).

While the Shields (1936) criterion provides a clear motion boundary, sediment movement is better understood from an stochastic perspective given the turbulent nature of flow (Paintal, 1971). In fact, sediment becomes mobile with non-dimensional bed shear stress between 0.03 and 0.07 (Breusers and Schukking, 1971, 1976).

While the most famous, the Shields (1936) criterion based on non-dimensional bed shear stress is not the only one. Another widely used criterion is velocity or

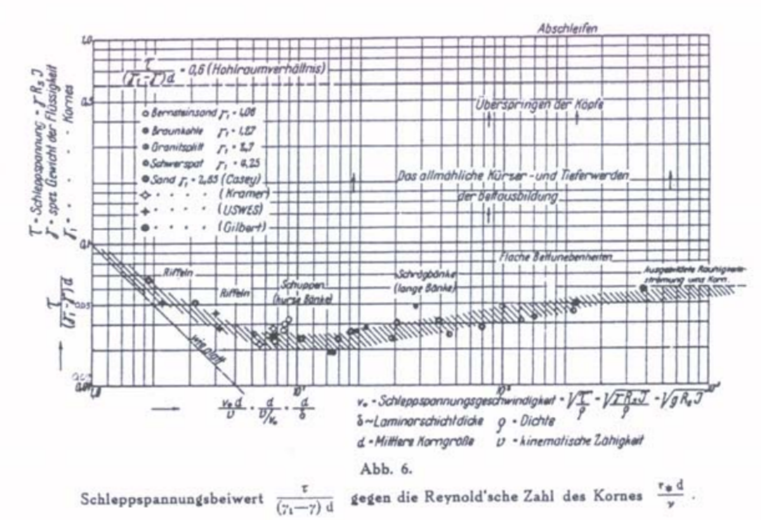

Figure 2 Shields (1936) diagram indicating motion of sediment (above the band) or rest (below the band) as a function of dimensionless bed shear stress (vertical axis) and particle Reynolds number (horizontal axis).

non-dimensional velocity (Izbash and Khaldre (1970), see CIRIA *et al.* (2007)). Other formulations focus on armoured conditions for the design of rip-rap and revetment (e.g., Pilarczyk, 1995; Escarameia and May, 1995).

Shields (1936) considered unisize sediment in his renowned experimental work, which is a limitation for predicting sediment mobility in natural rivers, which can rarely be considered to be formed by sediment of the same size. When the river bed is formed by sediment of different sizes, the sediment particles that are larger than average experience larger flow forces than if the bed would all be formed by sediment of their size, as they protrude from the bed surface and are more exposed to the flow. On the contrary, the sediment particles that are smaller than average experience smaller flow forces than if the bed would all be formed by sediment of their size, as they hide behind larger particles. This effect is known as hiding-exposure (Einstein, 1950).

Egiazaroff (1965) was the first who mathematically formalized the understanding of the hiding-exposure effect. He proposed an equation in which the critical bed shear stress depends on the relative size of a sediment particle with respect to the average (Figure 3). Several other authors propose different equations to account for the same effect and provide experimental evidence (e.g., Ackers and White, 1973; Fenton and Abbott, 1977; Dhamotharan *et al.*, 1980; Parker *et al.*, 1982; Misri *et al.*, 1984; Komar, 1987b,a; Kuhnle, 1993; Buffington and Montgomery, 1997).

Studies on the initiation of motion continue to be of interest and a myriad of them exist which broaden the conditions under which the resulting equations can be used. While crucial for morphodynamic prediction, initiation of motion does not give an indication of the future morphodynamic trends that one can expect. Sediment may be mobile at a particular location and either aggradation or degradation can occur. For relating sediment motion to morphodynamic trends, it is necessary to first understand the amount of sediment that is in motion when the critical threshold has been exceeded.

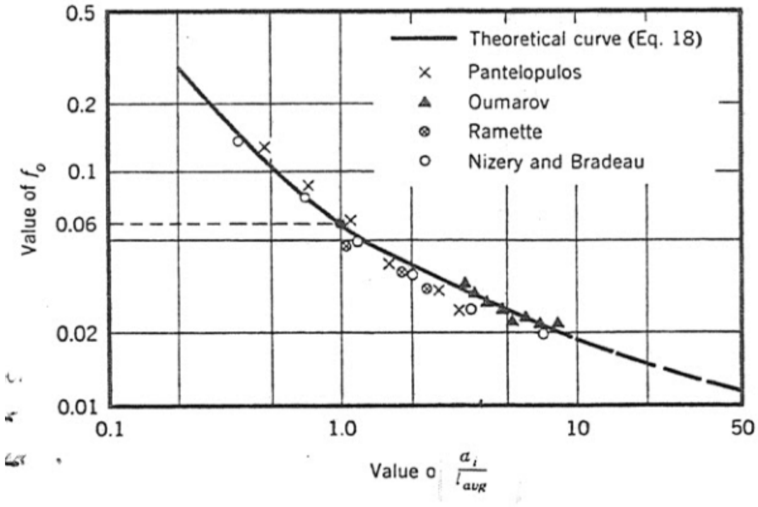

Figure 3 Egiazaroff (1965) theoretically-derived equation and experimental data by other researchers indicating the critical non-dimensional bed shear stress for a particular sediment particle in a mixture (vertical axis) as a function of the grain size relative to the average grain size (horizontal axis).

2.3.2 Sediment transport

A particle of sediment transported at some location by a river may have been entrained from the bed surface somewhere close upstream (tens or hundreds of times the particle size) of the current location. This is classified as bed material transport. This sediment may be transported by rolling or sliding (bed load) or in suspension for some time in the fluid (suspended load). The other possibility is that a particle of sediment in transport has had no contact with the bed surface. The size of this particle of sediment is much finer that the bed surface and its origin is found far upstream of the current location. This sediment is classified as wash load and it is always in suspension (Figure 4). Each of these modes have their own approach to estimate the sediment transport rate. Wash load is relevant for floodplain processes and deposition in the river mouth, but it is not relevant for main-channel morphodynamics. Hence, in the following, we focus on bed material transport. In this case, the sediment transport depends on the capacity of the flow to mobilize and transport the sediment present in the bed.

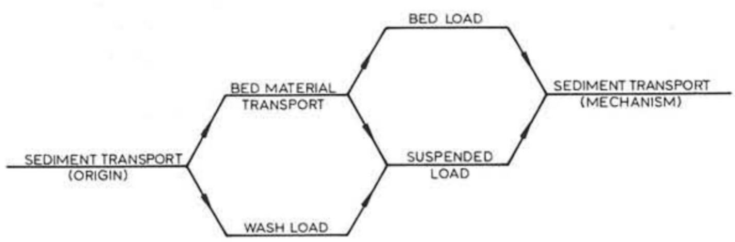

Figure 4 Classification of sediment transport according to its origin and mechanism. Figure from Jansen *et al.* (1979).

A great deal of equations exists that relate flow and the sediment properties to the sediment-transport capacity. One category is based on energy concepts, in which the sediment transport rate is a function of the stream power (e.g. Bagnold, 1973; Einstein, 1950). A second category is based on shear stress, in which the sediment transport rate is a function of the force exerted by the flow on the bed. This category can be further subdivided between the ones including a critical shear stress above which motion (and sediment transport) occurs and the ones in which there is no such a threshold. The paradigm equation of the first type is that by Meyer-Peter and Müller (1948). Other examples are Ashida and Michiue (1971); Fernandez-Luque and Van Beek (1976); Van Rijn (1984a). In the second group, the most widely used relation is probably that by Engelund and Hansen (1967) and other examples include Grass (1970); Wilcock and Crowe (2003); Parker *et al.* (1982).

For computing the sediment transport rate in the presence of a mixture of sediment of different sizes, the concept of sediment transport capacity is frequently used (Deigaard and Fredsøe, 1978; Ribberink, 1987; Armanini, 1995). The sediment transport rate for each size fraction is computed individually (i.e., the capacity) and the total amount is an averaged weighted according to the presence of that particular size fraction in the bed surface. Obviously, the hiding-exposure effect as a modification of the critical bed shear stress is applicable only in those equations containing a critical bed shear stress. Nevertheless, there are other type of functions that account for the hiding-exposure effect which do not include a critical bed shear stress, such as the one used by Wilcock and Crowe (2003). In this case, the hiding modifies a reference, rather than critical, bed shear stress.

Other approaches for computing mixed-size sediment transport rate include those in which the total load is computed independently from the grain size distribution of the load (e.g. Recking *et al.*, 2016).

Rather than relating flow and sediment properties to sediment transport rate, one can relate flow and sediment properties to the entrainment rate, which is the number of particles that are set into motion per unit of bed area and time. In combination with a particle velocity and step length closure relations (e.g. Nakagawa and Tsujimoto, 1980b,a; Nakagawa *et al.*, 1982; Sekine and Parker, 1992; Sekine and Kikkawa, 1992; Niño *et al.*, 1994; Hu and Hui, 1996a,b), one derives the sediment transport rate. This is for instance what is done by Fernandez-Luque and Van Beek (1976); Seminara *et al.* (2002); Parker *et al.* (2003).

It is important to be aware of the fact that all existing sediment transport relations are semi-empirical. Based on some physical principles (e.g., the fact that transport depends on excess bed shear stress) or similarity conditions (as in Engelund and Hansen (1967)) a relation between parameters is derived. However, all relations heavily depend on calibration and the range of applicability is restricted to the conditions in which calibration has been conducted. Moreover, a significant amount of scatter exist even for each calibrated sediment transport relation (Figure 5).

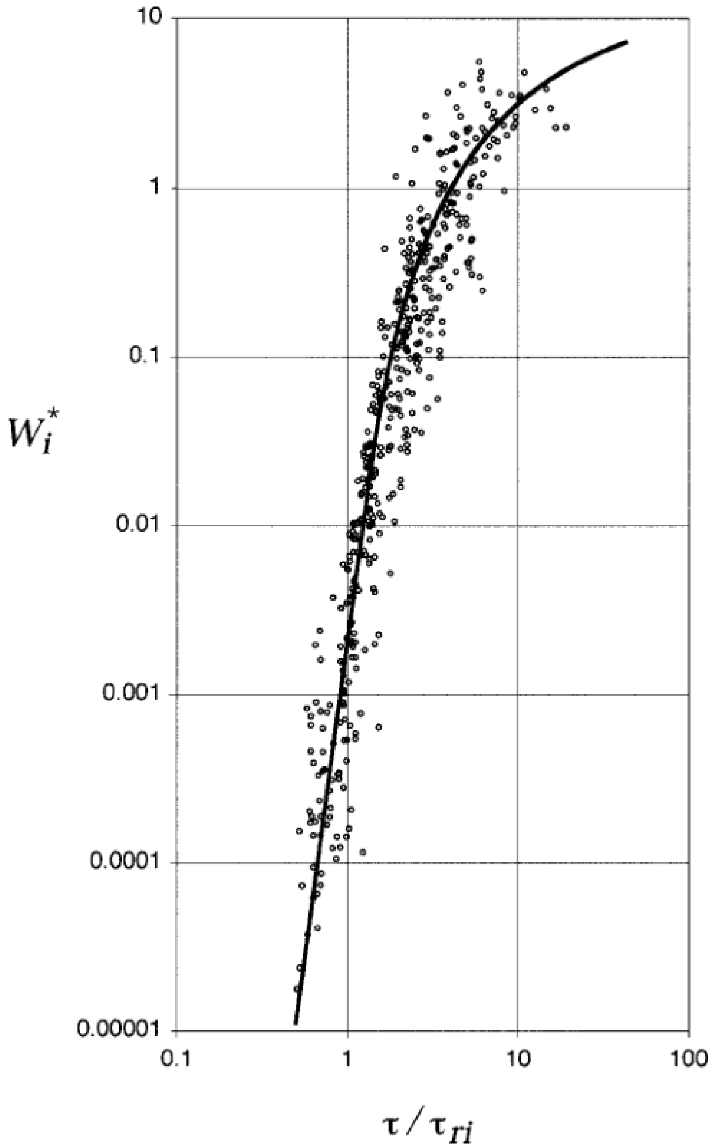

Figure 5 Experimental data and sediment transport relation by Wilcock and Crowe (2003). The vertical axis is the non-dimensional transport capacity and the horizontal the non-dimensional bed shear stress relative to the non-dimensional reference bed shear stress.

2.3.3 Sediment transport direction on sloping beds

As sediment particles are set in motion and transported by the flow, the sediment transport direction mainly follows the direction of the flow. For suspended sediment this is evidently the case, as sediment is part of the flow. The direction of the bedload sediment is not only depending on the flow direction, but it is also affected by the local bed slope. Due to the effect of gravity on sediment particles moving over the bed, sediment is diverted in a down slope direction. Considering this physical process is not only relevant for proper modelling, but it is actually a *sine qua non* condition for obtaining a well-posed two-dimensional or three-dimendional model (Chavarrías *et al.*, 2019).

There is no clear distinction between bed load and suspended load and one needs to consider that most of the suspended load occurs close to the bed, where it may also be affected by the local bed slope. Hence, proper modelling of suspended load would also account for the effect of the bed slope maybe in the entrainment function as well as by modifying its direction.

In river bends, secondary or helical flow transport sediment towards the inner bend, increasing the bed slope. This process is counteracted by gravitational pull, which leads to an equilibrium transverse bed slope (e.g. Engelund, 1974; Struiksma *et al.*, 1985). The model of the bed slope effect is crucial in determining the equilibrium bed slope. Similarly, the bed slope effect plays a major role in river bifurcations (Bolla Pittaluga *et al.*, 2003; Sloff and Mosselman, 2012; Bolla Pittaluga *et al.*, 2015), as well as bar pattern (Crosato and Mosselman, 2009; Schuurman *et al.*, 2013).

Different models have been developed for considering the bed slope effect. Van Bendegom (1947) is the first one who developed such a model based on a force balance on a single particle for wide bends in which secondary flow is small compared to the primary flow (see also Allen (1978)). Several laboratory experiments have been conducted to validate theoretical findings about the transverse bed slope (e.g. Engelund, 1974; Zimmerman and Kennedy, 1978; Ikeda *et al.*, 1981; Koch and Flokstra, 1981; Ikeda, 1984). The main difference between the models is found in the deviation of the bed shear stress direction relative to the depth-average direction, which varies depending on the assumptions used in developing the models.

In general terms, all expressions to model the strength of the gravitational pull are of the form:

$$g_{\mathrm{s}k} = A_{\mathrm{s}} \theta_k^{B_{\mathrm{s}}} \; , \tag{2.1}$$

where $g_{\mathrm{s}k}$ [-] inversely weights the bed slope, θ_k [-] is the Shields (1936) parameters on size fraction k, and A_{s} [-] and B_{s} [-] are parameters. The dependence on the bed shear stress is found by assuming that the bed slope effect depends on the fluid drag force (Koch and Flokstra, 1981).

Talmon *et al.* (1995) conducted a set of experiments on a straight flume in which the bed was initially tilted in transverse direction. Under the conditions of the experiments, an exponential decrease in time of the transverse slope is expected in which the time scale is related to $g_{\mathrm{s}k}$. Measuring the time scale and curve-fitting a function with the form of Equation ((2.1)), they found that the best fit is found for parameters $A_{\mathrm{s}} = 1.7$ and $B_{\mathrm{s}} = 0.5$.

Talmon *et al.* (1995) discussed that the height of bedforms appears to be important in estimating the bed slope effect. As bedform height is related to the grain-size-to-flow-depth ratio and the Shields stress (Van Rijn, 1984a), a third parameter is introduced to such end C_{s} [-]. Finally, the grain size distribution may play a role, which introduces a fourth parameter D_{s} [-] leading to an equation of the form:

$$g_{\mathrm{s}k} = A_{\mathrm{s}} \theta_k^{B_{\mathrm{s}}} \left(\frac{d_{50}}{h} \right)^{C_{\mathrm{s}}} \left(\frac{d_k}{d_{50}} \right)^{D_{\mathrm{s}}} \; , \tag{2.2}$$

where h [m] is the flow depth, d_{50} [m] is the 50% percentile grain size of the sediment mixture, and d_k [m] is the characteristic grain size of fraction k. This equation is implemented in DELFT3D , although the effect of parameters C_{s} and D_{s} has not been intensively studied. Parameters C_{s} and D_{s} were initially implemented in DELFT3D to have a general relation that allows a pragmatic way of considering hiding, dunes, or grain size. However, there is no strong theoretical framework supporting its use.

Based on numerical computations for natural rivers and laboratory flumes, Talmon *et al.* (1995) found the relation A_s = 9, B_s = 0.5, C_s = 0.3 and D_s = 0.0 to reasonably model the transverse slope effect.

Seminara *et al.* (2002) and Parker *et al.* (2003) extend the analysis to conditions in which the bed slope is large and Francalanci and Solari (2007, 2008); Francalanci *et al.* (2009) provide experimental evidence supporting their analysis.

Wiesemann *et al.* (2006) conducts a set of laboratory experiments with different bedform types finding that with dunes, the transverse sediment transport is reduced.

In general, the topic of the effect of the bed slope on the sediment transport is far from being mature. For instance, Ottevanger *et al.* (2013) reanalysed the experimental results of Zimmerman and Kennedy (1978) and found that a B_s of 0.25 also provides reasonable results, and fits within the scatter of the experimental observations. Baar *et al.* (2018) provided an overview of parameters according to different authors showing large variability depending on the condition under which the experiments to derive the parameters were conducted. They also investigated the development of transverse slopes in graded sediment mixtures. Moreover they found that certain sediment transport formulations require a slope treatment to ensure their stability (Baar *et al.*, 2019).

The recent research and continuous development is a symptom that substantial understanding of this physical process is ongoing.

2.4 Changes in the river bed

2.4.1 Exner equation

The next degree of complexity in predicting morphodynamic evolution after estimating the sediment transport rate is to estimate changes in bed level. This is essentially done by analysing the sediment fluxes in a control volume and realizing that the change in bed elevation is equal to the divergence of the sediment transport rate. In other words, if there is more sediment going in than out, the bed level will rise and vice-versa. This is known as the Exner (1920) equation. In fact, bed elevation changes depend also on other processes such as subsidence and uplift and compaction and dilation (Paola and Voller, 2005), although these are generally neglected.

In reality, the original equation by Exner related changes in bed elevation to the flow divergence and not the sediment transport. Here we will use the term Exner equation in a general form as in (Paola and Voller, 2005).

Mass conservation is usually rewritten in terms of conservation of volume assuming that porosity is constant. If porosity is variable in space or time, it needs to be considered (Section 2.4.5).

2.4.2 Active-layer model

The previous principle, encompassed in the Exner (1920) equation, does not take into consideration the type (i.e., size) of sediment that enters or exits the control volume. Hence, there is no information about fining or coarsening of the control volume. Prediction of changes in bed composition was first achieved by Hirano (1971) who considered the fluxes of sediment per size fraction in a control volume, similar to Exner (1920). In plain words, the main concept underlying the model by Hirano (1971) (i.e., the active-layer model) is that the bed surface will become finer if there is more fine sediment entering than exiting the control volume (Figure 6). This idea crucially depends on the thickness of the control volume (i.e., the active layer thickness). For a given bed surface area and sediment fluxes, a thicker active layer will lead to smaller changes in bed surface composition, as the sediment fluxes become relatively smaller.

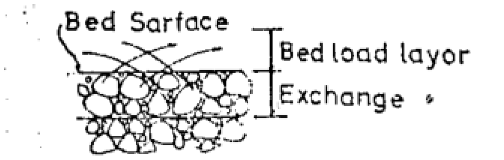

Figure 6 Original sketch by Hirano (1971) showing the top part of the bed (exchange or active layer) interacting with the bed load.

The idea that sediment interacts with the flow means that sediment in the active layer (1) provides bed friction, (2) can at any moment be set into transport if the bed shear stress is large enough, (3) affects the transport of all other sediment particles due to hiding and exposure. The list may not exhaustive, but these are at least some properties of the concept of sediment interacting with the flow. Sediment can be entrained from the active layer only and sediment is deposited in the active layer only. The active layer is assumed to be homogeneous (i.e., sediment in this layer is mixed). Contrary to the active layer, the substrate may be stratified.

One of the critical aspects of the active-layer model is the fact that the vertical extent of the active layer, or active layer thickness, shall be a priori assigned. However, it cannot be physically measured, as it stems from the above schematic representation (Siviglia *et al.*, 2017; Church and Haschenburger, 2017). The active layer thickness is related to the time scale of the process under consideration (Bennett and Nordin, 1977; Rahuel *et al.*, 1989; Sieben, 1997; Wu, 2007). In plane bed conditions and short time scales the active layer thickness is assumed to be proportional to the size of a characteristic coarse fraction in the bed, for instance, D_{84} or D_{90} (e.g., Petts *et al.*, 1989; Rahuel *et al.*, 1989; Parker and Sutherland, 1990). If bed forms are predominant and the time scale under consideration involves the mixing induced by the passage of several bed forms, the active layer thickness is typically related to a characteristic bed form height (e.g., Deigaard and Fredsøe, 1978; Lee and Odgaard, 1986; Armanini and Di Silvio, 1988). The active layer thickness may vary over space and time, although often it is assumed to be a uniform constant.

In the original active-layer model, only gradients in transport cause an increase in bed elevation. Particles that suddenly stop moving do not. This assumption is valid for small concentrations of moving particles and when the adaptation time scale to changing flow conditions is fast with respect to changes in bed elevation (e.g., Armanini and Di Silvio, 1988; Garegnani *et al.*, 2011, 2013).

In a situation in which there are no gradients in the sediment transport rate per size fraction (and as such the mean bed elevation is constant), the only process that can lead to a change in surface grain size distribution is a lowering of the interface between the active layer and the substrate due to, for instance, an increase in the active-layer thickness. This is a limitation of the active-layer model, as several processes are inadequately described in this manner. For instance, dune growth under normal flow conditions (i.e., without change in mean bed elevation) causes the formation of a coarse layer underneath migrating dunes (Blom *et al.*, 2003). Lee-face sorting causes the deposition of coarse sediment at the dune troughs. As it often happens, the coarse sediment is immobile and dunes become composed of the fine fractions only. The coarse layer inhibits the entrainment of fine sediment and limits the sediment transport rate. Although the formation of such a coarse layer is not modelled by the active-layer model, the active-layer model does account for the transport of some of the sediment fractions while some other sediment fractions remain immobile. Whether a particular sediment fraction is mobile or not depends on the closure relation for the sediment transport rate (considering hiding-exposure) and the amount of the particular sediment fraction relative to the total sediment at the bed surface (i.e., in the active layer). The reduction in sediment transport is intrinsic to the fact that there is sediment in the active layer which is not mobile.

2.4.3 Multi-layer and vertically continuous models

Ribberink (1987) included a third layer between the active layer and the substrate to model the effects of dunes exceptionally larger than the average dune height. His model still crucially depends on defining the active layer thickness.

To overcome the problem of setting the active layer thickness, Parker *et al.* (2000) developed a stochastic framework without the need for a distinction between the active and inactive parts of the bed. Blom and Parker (2004), Blom *et al.* (2006), and Blom *et al.* (2008) developed a model that accounts for dune sorting and the variability of bed elevation based on the stochastic framework developed by Parker *et al.* (2000). The model associates a probability of grain size selective entrainment to all elevations within the bed, and hence allows for sediment at any elevation to be entrained and contribute to the bedload discharge. Viparelli *et al.* (2017) developed a simplified vertically continuous model assuming slow changes in bed elevation and a steady probability distribution of entrainment, deposition, and bed elevation, which make their model suitable for large space and time domains.

2.4.4 Fixed-layer modelling

For modelling morphodynamic development in the presence of a non-erodible or fixed layer, Struiksma (1999) developed a model that reduces the sediment transport rate as a function of the thickness of sediment on top of a fixed layer relative to the alluvial thickness. The alluvial thickness is defined as the minimum thickness of sediment for the fixed layer to influence morphodynamic development.

The model by Struiksma (1999) was defined and tested under unisize sediment conditions and presents a conceptual issue when extending it to account for mixed-size sediment conditions. While a fixed layer is built with coarse sediment, it is not modelled as such. The fixed layer is represented by an unerodable surface. Independently of the flow conditions, a fixed layer remains always immobile, i.e., acting as a bed rock surface or concrete surface. While this may seem sensible, two problems arise. First, the properties of the sediment forming the fixed layer have no influence on the sediment transport. Worded differently, hiding and exposure effects are not accounted for. Second, man-made fixed layers may be made with such coarse sediment that indeed it never moves, but fixed layers may also form naturally, and sediment may become mobile only under certain flow conditions. This is known to occur in the Pannerdensch Kanaal and the Grensmaas. Moreover, one would like to predict morphodynamic development in the case that, for instance, a naturally formed coarse layer breaks.

In principle, using the active-layer model one can account for a fixed layer by simply modelling a coarse sediment size fraction. Nevertheless, this leads to physically unrealistic results as immobile sediment does not move in streamwise direction but does move vertically due to mixing (Chavarrías and Ottevanger, 2019). Solving for this problem would require modification of the aggradation flux of sediment from the active layer to the substrate. This flux should first be formed out of coarse immobile sediment to prevent it from moving upwards.

Tuijnder and Ribberink (2010); Tuijnder *et al.* (2011); Tuijnder and Ribberink (2012) developed a model which is an extension of the classical active-layer model approach that intends to allow for a transition from full alluvial (active layer) to a fixed bed approach (Struiksma concept) and back. Implementation of this model into the software package Delft3D has not given satisfactory results and the reasons for this are unclear.

2.4.5 Porosity modelling

All the models presented above assume that porosity remains constant with time. Porosity can change as a result of, for instance, compaction of cohesive sediment or fines infiltrating a matrix of coarse sediment. It may be worthwhile to investigate the role of porosity with regards to the morphodynamic development of the river bed, but for this study, the literature on this topic has not been further evaluated. Some of the most recent advances can be found in Frings *et al.* (2011); Uchida *et al.* (2020).

2.4.6 Armouring

Although a of an outsider in the topic of changes of the river bed, the topic of armouring is of special relevance and complexity. Klaassen (1990) provides an extensive and comprehensive overview of the topic. Although research has advanced the understanding of the topic since his work was completed, his study is still a milestone on armouring.

It is important to differentiate between a static and a mobile armour (Jain, 1990). In the former, coarse particles are never mobile and they form by winnowing transport of fine particles. This is usually what occurs downstream of dams. In the latter, coarse sediment is mobile under high-flow conditions. The bed surface remains coarse nonetheless due to the different mobility of coarse and fine sediment (Parker and Klingeman, 1982).

Break-up of armour layers is an important process to be considered by river managers. When the surface of a river is armoured, the sediment transport of the fine sediment is relatively low due to hiding. If the armour breaks, a sudden increase in sediment transport occurs due to (1) sudden availability of sediment in the substrate and increase sediment transport due to decrease of hiding-exposure effect.

Klaassen (1987) showed in a series of laboratory experiments that break-up of the armour layer in a gravel-sand bed occurs as dunes composed of fine sediment winnowed from below the armour layer form dunes on top of the coarse layer (Figure 7). Turbulence intensity increases downstream of the lee face, where flow reattaches, which causes a larger mobility of coarse sediment. When a flood recedes, the armour layer is found at a lower elevation as lee-face sorting transports the coarse sediment to the dune troughs (Blom *et al.*, 2003).

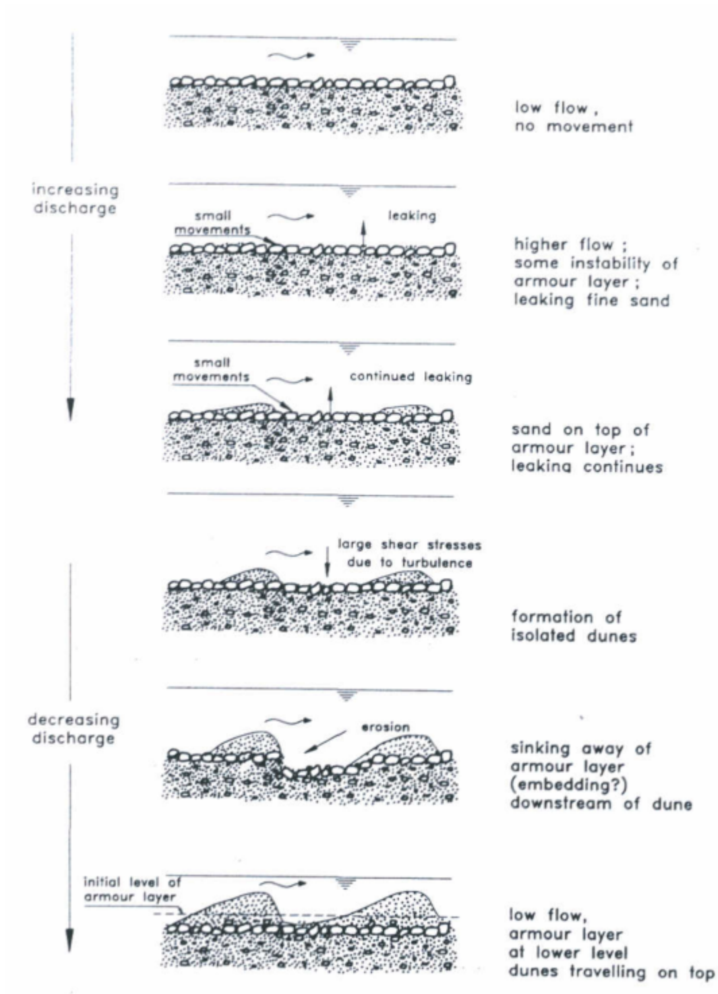

Figure 7 Armour layer break-up and reformation (Klaassen, 1990).

A significant amount of researchers endeavoured to predict the degree of armoring eventually reached given an initial bed and flow conditions (e.g. Gessler, 1965; Little, 1972; Shen and Lu, 1983; Parker and Sutherland, 1990; Chin *et al.*, 1994). It is worth mentioning the research by Marion and Fraccarollo (1997), who conducted laboratory experiments in WL | Delft Hydraulics on the formation of a mobile armour showing the importance of hiding-exposure effect. Duizendstra (2001) shows measurements and estimations of the bedload in the *Grensmaas*, an armoured river.

The work above mentioned focuses on equilibrium conditions and does not intend to predict the changes with time. In order to predict changes with time, Sieben (1999) present an analytical model of armouring in a degradational river obtained by simplifying and linearizing the set of equations until a relaxation equation is found.

Further complexity can only be achieved by numerical solution of the active-layer model. Karim *et al.* (1983); Karim and Holly (1986) present both a computation of the volume of static particles as degradation occurs and a numerical solution of the active-layer model for predicting armouring. At that time, those computations where restrictive in terms of power and memory, and Berezowsky and Jiménez (1994) presents a simplification of the method by Karim *et al.* (1983); Karim and Holly (1986) based on assuming a certain distribution of the sediment.

Bettess and Frangipane (2003) solves the active-layer model for predicting armouring and study different assumptions as regards to the interface between the active layer and the substrate. For instance, one may assume that the active layer thickness remains constant, which implies that sediment from the substrate is transferred to the active layer as the bed degrades, or that the interface between the active layer and the substrate remains constant, which implies that the active layer thickness decreases with time.

In all cases, the active layer is homogeneous and the sediment there present can be entrained according to their volume fraction content. Borah *et al.* (1982) introduces a slightly different concept that takes into account that entrainment of a large particle leaves room for entrainment of small particles hidden by the large one. In computing the composition of the active layer when degradation occurs, the active layer is treated in a similar manner as the substrate, by discretizing it into several layers. However, the sediment is not entrained by layers from top to down, but, for instance, entrainment of a large sediment particle in the topmost layer is followed by entrainment of fine sediment particle in the second layer that was hidden by the coarse particle.

2.5 Short-term response

The first thing that a river manager may be interested in knowing is the short-term response to be expected in a river. This is to know the locations in which aggradation and degradation is expected to happen. Two methods are discussed in the following sections.

2.5.1 Flow pattern analysis

Before carrying out a full-morphodynamic modelling approach, it is possible to make a first estimate of the short-term morphological response by just analysing the flow pattern and its impact on sediment transport. The flow is used to compute local sediment transport rates. The divergence is computed and using the mass balance equation (also known as Exner (1920) equation) one find the initial rate of change in bed elevation (Figure 8). This approach is particularly used and easy to handle for width-average (i.e., one-dimensional) analyses.

The same approach can be followed to estimate expected initial change in bed surface composition. In this case, sediment transport per size fraction must be computed based on the flow pattern for solving the Hirano (1971) equation and obtain the rate of change of volume fraction content of each size fraction at the bed surface.

These analyses (both for bed elevation and bed surface composition) are valid on a short timescale only, as they do not take into account the fact that a change in bed elevation and bed surface composition will change the flow pattern, which will change the sediment transport pattern and eventually the rate of change of bed elevation and bed surface composition. In other words, the short term response does not take into account the interaction between changes in the bed and changes in the flow.

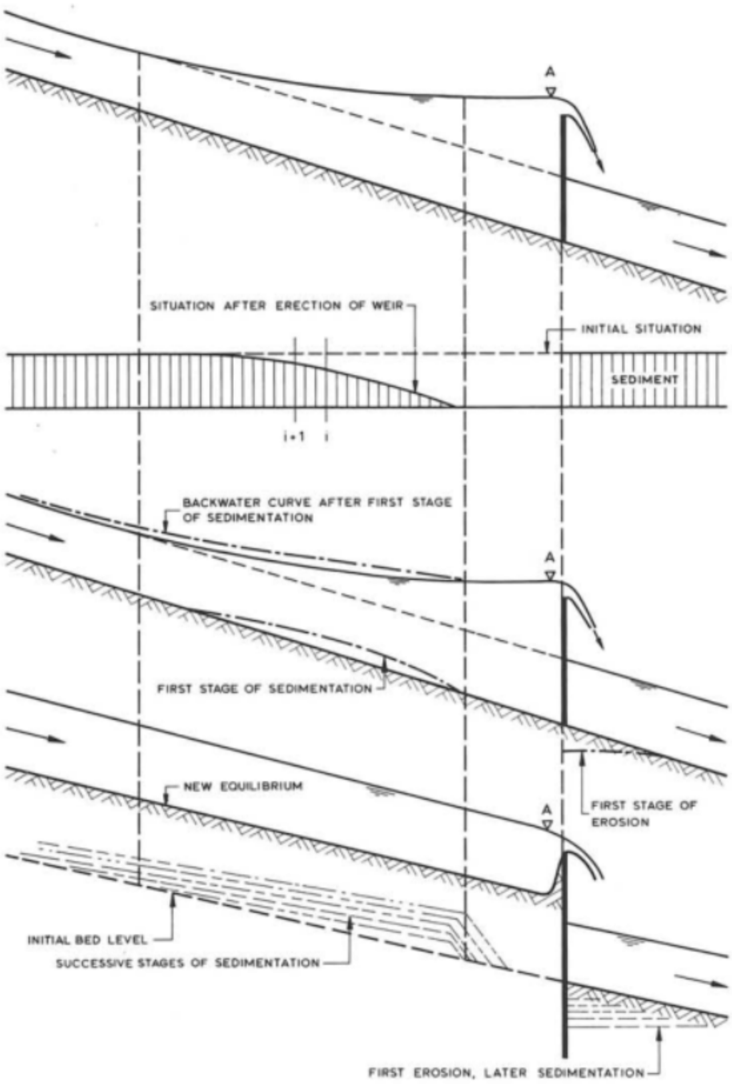

Figure 8 Change of water level and bede level after erection of a fixed weir (from Jansen *et al.* (1979)).

2.5.2 WAQmorf

Another version of flow-pattern analysis is provided by WAQmorf. WAQmorf is a tool for analysing the short-term response to river interventions (Sieben, 2010). Based on the flow velocity pattern of a two-dimensional WAQUA model before and after an intervention and the flow depth before an intervention, WAQmorf computes the mean annual change in bed elevation as well as the change at the end of the low-flow season and the change at the end of the high-flow season. An example application and comparison to Delft3D was conducted by Paarlberg (2009).

Assuming an infinitesimal intervention, such that the equations can be linearized and the water level can be assumed to not be affected by the intervention, the Exner equation (hence only considering unisize conditions) is integrated over a characteristic length scale to reduce it to a relaxation equation. The timescale is a crucial parameter which is specifically derived for the Dutch river system and the tool is built for dealing with three characteristic discharges, also relevant for the Dutch river system. The method implicitly assumes that the bed is initially under equilibrium conditions. Only local changes can be computed and hence the propagation of a morphodynamic feature (such as a nourishment) cannot be assessed.

WAQMorf can serve as a first rapid assessment to see whether morphological interventions have significant effects or not. If they do not present a significant effect, no further morphological investigations are necessary. If they do have significant effects, those effects should be assessed through either expert judgement or DELFT3D computations (Mosselman, 2013; Van der Mheen and Prins, 2015).

2.6 Long-term response

The main caveat of the short-term analysis is that the feedback between the bed and the flow is considered. The next straight forward step in order to be able to predict morphodynamic changes on a long timescale (years to decades) is to consider this mechanism.

2.6.1 Solution of the system of equations

Ideally, one would like to solve the equations of flow and bed analytically for finding the future state. Unfortunately, this is not possible and numerical techniques must be used for finding the solution.

In the short-term response, starting from a flow pattern, the rate of change in bed elevation is computed, by the methods explained in section 2.4. The essence of computing long-term response is to account for this rate of change to compute a new bed level (and surface composition). Using this new bed level, a new flow pattern is computed, which allows for subsequently finding a new bed level. In other words, the flow and bed equations are solved in a time-loop.

It is relevant to mention that the description we have given of the solution of the system of equations (i.e., first solve for the flow, then solve for the bed, and repeat the process) is valid only if the flow and bed equations weakly interact with each other from a mathematical perspective. This is to say that the effect of the bed equation on the flow equations is “small”, which occurs for a small Froude number (De Vries, 1973; Needham, 1990; Zanré and Needham, 1994).

2.6.2 Delft3D

Van Bendegom (1947), working for *Rijkswaterstaat* , developed world’s first two-dimensional morphodynamic model in the 1930’s (Allen, 1978). His work was pioneering in several aspects, including the notion that transverse bed slopes and helical flow affect the direction of sediment transport. Due to World War II, the work was partially destroyed and could only be published in 1947.

As previously mentioned, morphodynamic prediction including the feedback effect of the bed in the flow requires the use of numerical techniques for finding the solution, which is associated to using computers. While this is currently associated to using a machine computer, Van Bendegom (1947) used human computers. He was not the first, as another Dutchman also commissioned by *Rijkswaterstaat* , Nobel laureate Dr. Hendrik A. Lorentz (1853–1928), pioneered when numerically computing the effect of the construction of the *afsluitdijk* (Lorentz, 1926; Hazewinkel, 2004).

Two-dimensional morphodynamic computations using machine computers are relatively recent and a key step was conducted by Struiksma *et al.* (1985). Delft3D 4 (Lesser *et al.*, 2004) stems from their work and allows for computation on structured curvilinear grids. Currently, computation on unstructured grids is possible thanks to development of Delft3D FM De Goede (2020).

3 Improving of Tuijnder and Ribberink (2010) model in DELFT3D

3.1 Introduction

Struiksma (1999) developed a model for predicting morphodynamic development under the presence of a fixed layer (i.e., a sediment layer which is always immobile). The basic concept behind his approach is to reduce the sediment transport rate when the layer of sediment above the fixed layer is below a certain user-specified thickness labelled "alluvial thickness" (Figure 9). In this manner, the bed level cannot degrade below the fixed layer and there is a gradual transition between fully-alluvial conditions and the situation in which the fixed layer significantly affects morphodynamic development.

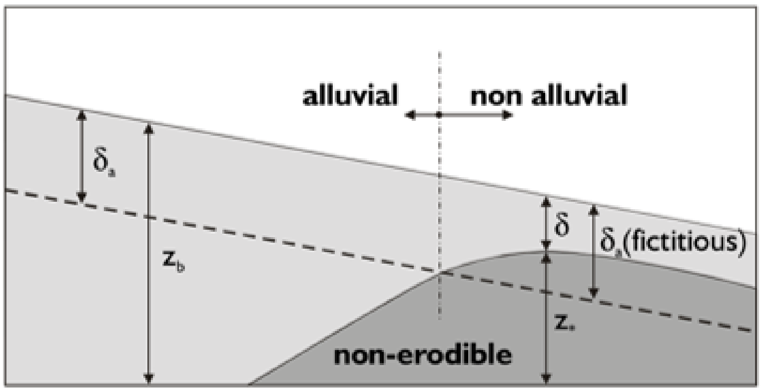

Figure 9 Sketch of the model by Struiksma (1999) (Figure by Struiksma (1999)).

His concept is particularly suitable for modelling the effect of engineered layers made with very coarse sediment (i.e., rip-rap) which is virtually immobile under all foreseeable flow conditions. It is also applicable for real immobile bed-rock or concrete beds. This model was successfully applied in designing the fixed layer constructed at Sint Andries (the Netherlands) in 1998. The original calculations by Struiksma *et al.* (1994) can be compared with the actual development shown, for instance, by Havinga (2020).

As the sediment forming a fixed layer in Struiksma's model is always immobile, it cannot model the formation and break-up of a layer of immobile sediment. This limitation lead to Tuijnder and Ribberink (2010) (also available as Tuijnder and Ribberink (2012)) to develop a model capable of handling such a situation. In this model, coarse sediment is modelled as such. Hence, this model considers several grain-size fractions and is an extension of the active-layer model (Hirano, 1971).

Two main points lead to the need for assessing the model performance and status. The first point is that while the model was aimed at reproducing the fixed-layer behaviour of Struiksma's model, it actually does not capture the experimental results by Struiksma (1999) Tuijnder *et al.* (2011). The second point is that the model was implemented in a research branch of DELFT3D Tuijnder *et al.* (2011), but Ottevanger (2015) was not able to reproduce work on the bifurcation region reported by Tuijnder *et al.* (2012) which may be due to numerous reasons such as code version, unclear model input, FORTRAN compiler, etcetera.

3.2 Review of Tuijnder concept

In this section we discuss caveats of the semi-fixed layer developed by Tuijnder and Ribberink (2010) and of its implementation in Delft3D. The limitations we currently find are:

1 Transfer of immobile sediment out of the active layer
2 Dune height adaptation
3 Dune height under alluvial conditions
4 Mixing of coarse and active layer sediment
5 Sediment mobility

These are explained in the following sections.

3.2.1 Transfer of immobile sediment out of the active layer

In the model by Tuijnder and Ribberink (2010), immobile sediment particles sink into the coarse layer below the active layer. This is not incorrect in principle, but a matter of model development and conceptualization. Whether only mobile sediment must be part of the active layer or not is a modelling choice with profound implications in the interpretation of results. Whatever the choice, the model must not contain contradictions or cause physically unrealistic results.

In the model by Tuijnder and Ribberink (2010), only sediment in the active layer interacts with the flow, which limits the model applicability to cases in which at least one sediment size fraction is mobile. This point is further discussed in Section 3.6.1.

3.2.2 Dune height adaptation

In the model by Tuijnder and Ribberink (2010), the active layer thickness tends to the alluvial active layer thickness, which depends on the alluvial dune height under equilibrium conditions. This causes the dune height to not be affected by the dune migration model.

Consider an alluvial case in which all sediment is mobile. Assume that the case is under equilibrium conditions. A certain dune height is found which can be predicted by a closure relation (e.g., Engelund and Hansen, 1967; Fredsoe, 1982; Van Rijn, 1984b). A sudden change in flow occurs (e.g., a flood wave). The equilibrium dune height under high-flow conditions is larger than initially. Nevertheless, the dunes do not instantaneously adapt to the new equilibrium value, but slowly adapt in a manner best modelled by an advection equation with a source term. (Depending on whether dunes increase or decrease in size the source is negative, i.e., it is a sink term). This is the dune-migration model.

In the model by Tuijnder and Ribberink (2010), the dune height is a surrogate of the active layer thickness. The active layer thickness tends to its alluvial value and is modelled by a relaxation equation. Hence, the advective character of the dune-migration model is neglected. The time-dependent relaxation is approximated, and arguably for most practical applications the advective behaviour has less relevance than the temporal development of the bed forms. However, for nourishment- or dredging-sections, the adjustment of bed forms depends highly on the dunes propagating (advected) from upstream.

3.2.3 Dune height under alluvial conditions

In the model by Tuijnder and Ribberink (2010), while the active layer thickness tends to the alluvial active layer thickness, the dune height does not tend to the alluvial dune height. This a model inconsistency.

Tuijnder and Ribberink (2010) states that the dune height Δ [m] is equal to:

$$\Delta = \Delta_0 \left[1 - \exp \frac{-L_\mathrm{a}}{0.39\Delta_0} \right] \ , \tag{3.1}$$

where Δ_0 [m] is the alluvial dune height and L_a [m] is the active layer thickness. Under alluvial conditions, the active layer thickness is equal to its alluvial value L_{a_0} [m] which modelled as (Tuijnder and Ribberink, 2010):

$$L_{\mathrm{a}_0} = \mu \Delta_0 \ , \tag{3.2}$$

where they propose the value $\mu = 1.5$. Substituting in Equation (3.1) we find:

$$\Delta = \Delta_0 \left[1 - \exp \frac{-1.5}{0.39} \right] \approx 0.9786\Delta_0 \ . \tag{3.3}$$

The error is 2.5% for their proposed value of μ. The error increases to 28% if one uses the more common assumption that the active layer thickness is half the dune height (Ribberink, 1987; Blom, 2008), which then does not match with the concept of Tuijnder and Ribberink (2010), it is however possible to adapt this in the model input, which may lead to unwanted behaviour.

Fortunately, though, this issue does not feedback into the computation as the dune height is only used as output parameter.

3.2.4 Mixing of coarse and active layer sediment

In the model by Tuijnder and Ribberink (2010), sediment in the coarse layer does not mix with sediment in the active layer when all sediment in the active layer is mobile. This can lead to physically unrealistic results.

Tuijnder and Ribberink (2010) describes an incomplete coarse layer (ICL) as:

> a layer containing coarse immobile fractions below the bedforms (dunes) that still allows further entrainment of mobile sediment into the active layers above. The rate of entrainment of mobile grains decreases with increasing concentration of immobile grains in the ICL. The ICL can develop to an immobile layer, that does not allow further entrainment of mobile grain size fractions or it can be buried below a completely developed transport layer with mobile sediment (due to sedimentation or insufficient immobile fractions for CCL [Complete Coarse Layer] formation). During its development stage the ICL remains partly exposed to the flow, whereas the exposure decreases to zero if it is buried. In that case the ICL does not cause any supply limitation any more for bedforms and sediment transport. This is an alluvial situation, although the (former) ICL material remains present just below the alluvial layer.

The corollary of the above definition is that, if all sediment is mobile, there is no mixing between active layer and coarse layer, but if immobile sediment is present in the active layer, both mobile and immobile sediment from the active layer mix with the coarse layer sediment. This leads to the following contradictory results.

Suppose a case in which the sediment mixture is characterized by three sediment size fractions (fine, medium, and coarse). The three size fractions are mobile and the bed surface has 90% of fine sediment and 10% of coarse sediment (i.e., no medium sediment). The case is under equilibrium conditions. In this case, the coarse layer and substrate composition do not affect the results, as there is no mixing between the active layer and the coarse layer because all sediment in the active layer is mobile. Assume the substrate to be composed entirely of the medium fraction for the sake of completeness. The result is that no fining or coarsening is observed.

Consider a second case identical to the first one but for the fact that the size of the coarse fraction is slightly larger such that it is immobile. This triggers the transfer of both mobile and immobile sediment from the active layer to the coarse layer. Mass conservation implies that medium-sized sediment is transferred to the active layer. The final equilibrium situation is that the active layer has no immobile sediment, less than 90% fine sediment (e.g., 82%) and 18% medium sediment. Depending on the grain sizes, the result is that there is actual coarsening of the bed surface, which is opposite to the expected result of applying the semi-fixed layer model. When using the semi-fixed layer model, one would expect that the immobile sediment in the surface layer sinks, causing fining of the bed surface.

The above cases are not only thought experiments but have been corroborated by means of Delft3D simulations. Figures 10 and 11 show the initial and final condition, respectively. Contrary to what one would expect, the bed surface is coarser in the final situation after immobile sediment settles than in the initial state.

These consequences are of course related to pragmatic or simplified concepts that have been applied for the exchange fluxes between the layers and can only be solved with more generic formulations. These specific experiments were not considered during the derivation of the theory. The question remains how important these simplifications are for more practical cases with well mixed beds and fluctuating discharges, and how these will affect the accuracy under these conditions.

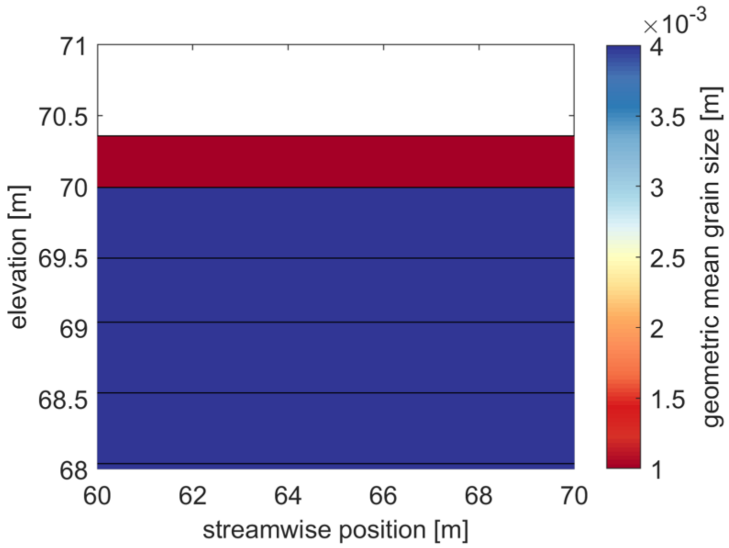

Figure 10 Longitudinal profile showing grain size in the initial state.

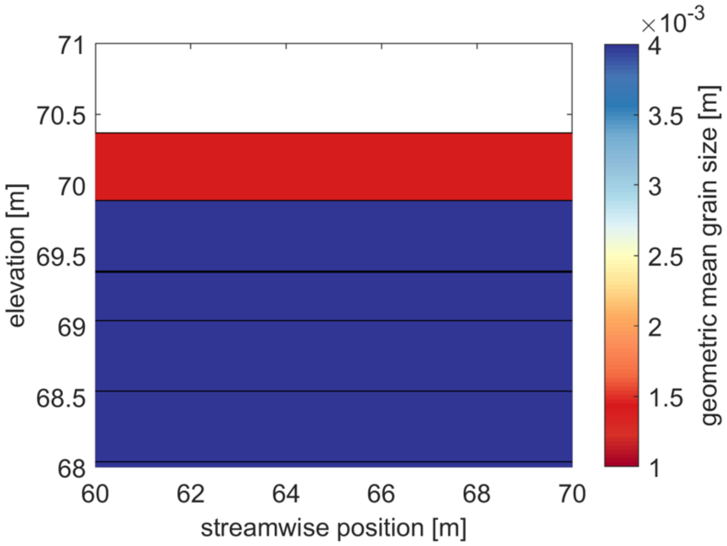

Figure 11 Longitudinal profile showing grain size in the final state.

3.2.5 Sediment mobility

In the model by Tuijnder and Ribberink (2010), the sediment mobility is estimated using the theory by Wilcock and McArdell (1997). This leads to physically unrealistic results.

A crucial point in the model by Tuijnder and Ribberink (2010) is discerning between mobile and immobile sediment particles. Initially, the Shields (1936) criterion was used but it was rejected as it lead to unstable results due to the discrete nature of the sediment mobility (i.e., sediment is either mobile or it is immobile). The theory by Wilcock and McArdell (1997) was eventually used to solve this problem, instead of considering a numerical solution that allows maintaining the original and more simple Shields criterion. Wilcock and McArdell (1997) proposes that sediment of the same characteristic grain size can be partially mobile. For example, one may say that a only 30% of the sediment with characteristic grain size equal to 0.001 m is mobile. The sediment mobility is described by an error function (i.e., a sigmoid curve) centred around the bed shear stress that mobilizes the characteristic grain size, which is found by a power relation of the grain size (Figure 12).

Several problems arise with this interpretation applied to the semi-fixed-layers model. As mobility is described by an error function, technically all sediment is neither fully mobile (i.e., mobility equal to 1) nor completely immobile (i.e., mobility equal to 0). As a corollary, one may say that all sediment is partially immobile. This implies that all sediment is transferred to the coarse layer until the coarse layer is complete. In essence, the mobility criterion is inconsistent with the model concept.

In practice, this does not occur in Delft3D due to floating point arithmetic. A large value of mobility eventually rounds up to 1 and it is considered mobile. Nevertheless, the result of the model is left to the floating point precision and a different result would be obtained if single precision is used rather than double precision, as it is now.

A second limitation is the fact that the theory of Wilcock and McArdell (1997) does not account for hiding, which is specially relevant under the conditions in which the semi-fixed-layers model is to be applied. Related to this problem is the general problem already mentioned that the criterion for mobility does not match the sediment

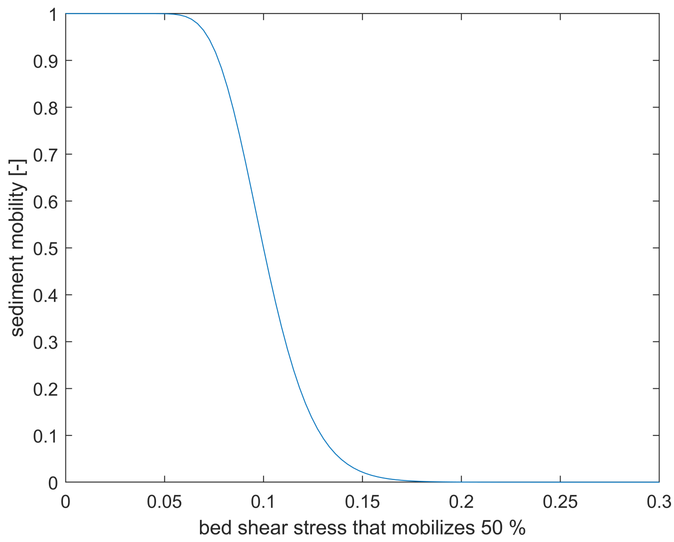

Figure 12 Example of the sediment mobility for a particular grain size according to Wilcock and McArdell (1997).

transport relation. Sediment can be in motion and at the same time be considered immobile and sink. As these shortcomings all arise from the choice of not using a Shields criterion for immobile sediment, this seems to require some repairs that go beyond the scope of this project for now (e.g., starting with a fix of the unstable behavior of the Shields criterion).

Finally, the continuous nature of the sediment mobility causes the following result. Consider a two-sediment-size-fractions case in which the grain sizes are almost the same but one is considered fully mobile (due to round up) while the other sediment is 99.99% mobile. After a long time when equilibrium conditions are reached, all the "immobile sediment" (it is only 0.01% immobile) has sunk and is not transported nor forms part of the active layer.

The above case is not only a thought experiments but has been corroborated by means of Delft3D simulations. Figure 13 shows the mobility of sediment in a two-size fractions case. Figures 14 and 15 show the initial and final state, respectively. While coarse sediment is more than 90% mobile, it disappears from the active layer in the long term.

The conditions in this example are not likely to be found in the more practical simulations for the Rhine branches, but they indicate specific limitations of the concepts that may have an impact on the evolution of the river bed. It is very well possible that these uncertainties affect the long-term solutions, even in a fully dynamic (discharge varying) simulation with a poorly sorted river bed.

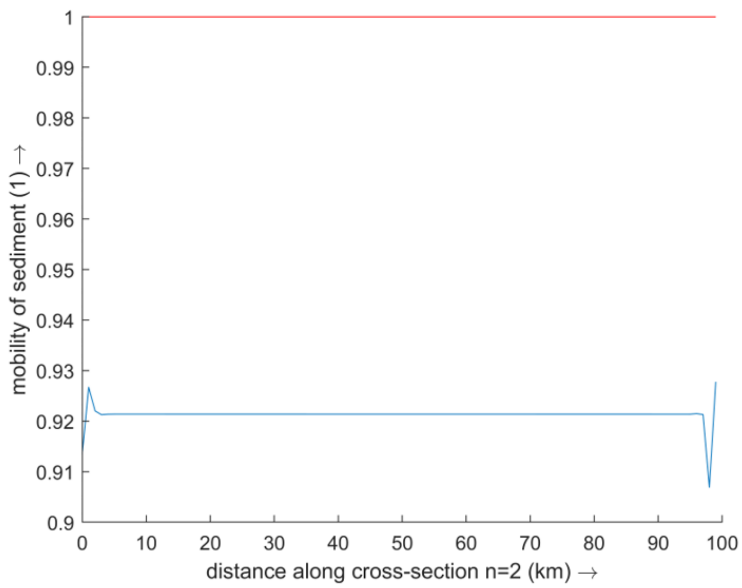

Figure 13 Sediment mobility in a two-size-fractions case.

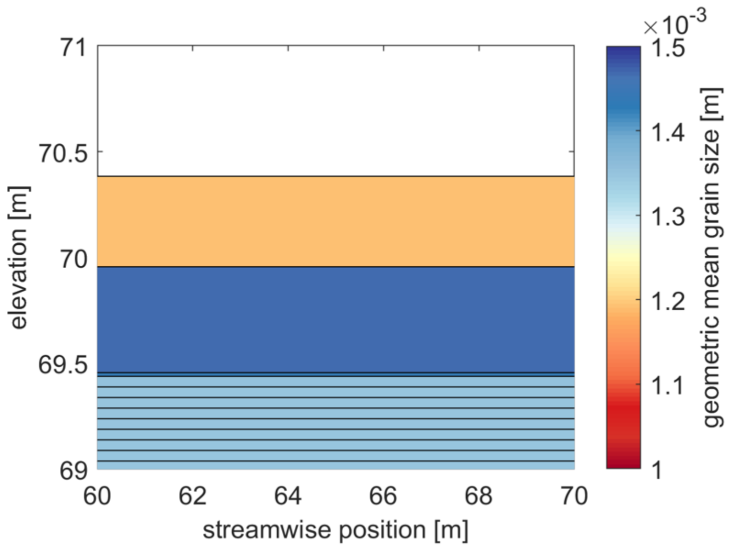

Figure 14 Longitudinal profile showing grain size in the initial state.

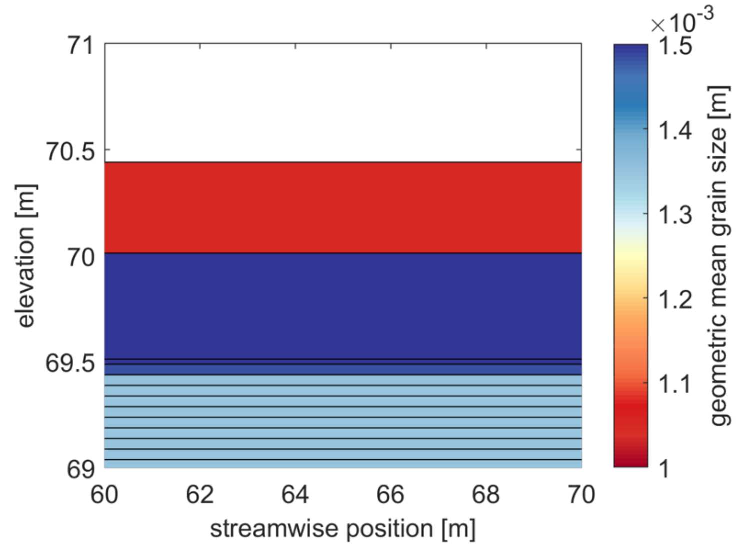

Figure 15 Longitudinal profile showing grain size in the final state.

3.3 Model adaptation

In this section, the model adaptations and desired model behaviour are discussed.

The first and most important point is whether immobile sediment must be transferred out of the active layer or not. Both options are valid if no inconsistencies are introduced in the model. This modelling decision has severe implications on the understanding and judgement of results. This point is later discussed in Section 3.6.1. After discussion with *Rijkswaterstaat* , it was decided that immobile sediment should not form part of the active layer as this is in agreement with the current model of Tuijnder and Ribberink (2010).

The relaxation mechanisms and dune-dependencies in the model developed by Tuijnder and Ribberink (2010) are excessively specific for the conditions in which their model was developed and are removed in the adapted model for the sake of simplicity and generality.

The active layer thickness tends to a value which can be a constant, as well as dune-height dependent, obtained solving the advection-diffusion equation of the dune-evolution model. Adaptation to this value is instantaneous if sediment is available in the coarse layer or via the sediment transport divergence. The relaxation behaviour is intrinsic in the dune-evolution model.

The coarse layer thickness is set to a constant value. This is a simplification which needs to be addressed in the future. It would be realistic to link this length to the variability in deep-through elevation or thickness required to form an armour layer (i.e., related to grain size).

Immobile sediment in the active layer is transferred to the coarse layer. The depositional rate depends on the sediment transport rate, as it is thought to occur due to the passage of dunes. As a consequence, if the sediment transport rate is equal to 0 (as it happens when all the sediment in the active layer is immobile), the flux of immobile sediment is also 0. Hence, immobile sediment present in the active layer remains there. This is a model inconsistency which limits the model applicability to conditions in which at least some sediment is mobile. This limitation implies that in this case immobile sediment may still form part of the active layer. In a dynamic environment this will probably not pose problems.

The immobile sediment that is transferred from the active layer to the coarse layer is replaced by mobile sediment present in the coarse layer. As a consequence, an unrealistic coarsening of the bed surface may still happen in specific cases (see Section 3.2.4). This point is further discussed in Section 3.6.2. If no (or not enough) mobile sediment is found in the coarse layer, the active layer thickness reduces.

When the interface between the active layer and the coarse layer elevates (e.g., aggradation under alluvial conditions), the sediment in the active layer is transferred to the coarse layer. This is the same mechanism present in the active-layer model between active layer and substrate.

When the interface between the active layer and the coarse layer lowers (e.g., degradation under alluvial conditions), mobile sediment from the coarse layer transfer to the active layer. This is the same mechanism present in the active-layer model between active layer and substrate but restricting the flux to mobile sediment only.

The interface between the coarse layer and the substrate is modelled in the same way as the interface between the active layer and the substrate in the active-layer model. That is, both mobile and immobile sediment are transferred without distinction.

3.4 Adapted model equations

In this section, the system of morphodymamic equations of the adapted model are described. The numerical discretization and implementation are not treated.

3.4.1 Bed discretization

We consider a one-dimensional channel with streamwise coordinate x [m] in which variables change with time t [s].

The sediment mixture is discretized into N [-] size fractions with characteristic grain sizes d_k [m] ordered in increasing size, where $k \in [1, N]$ [-] is an index specifying the grain size fraction. Index k_{imm} indicates the smallest immobile size fraction.

Sediment of size fraction k is transported at the bed surface at a rate (including pores) $q_{\mathrm{b}k}$ [m^2/s]. The total sediment transport rate (including pores) q_{b} [m^2/s] is:

$$q_{\mathrm{b}} = \sum_{k=1}^{N} q_{\mathrm{b}k} \ . \tag{3.4}$$

The bed extends from a fix datum z_0 [m] to the bed surface z_{b} [m]. The bed is discretized in 3 layers. From top to bottom these are the (1) active layer, (2) coarse layer, and (3) substrate. The active layer has a thickness L_{a} [m] and extends from the interface between the coarse layer and the active layer z_{cl} [m] to the bed surface. The coarse layer has a thickness L_{cl} [m] and extends from the interface between the substrate and the coarse layer z_{s} [m] to the interface between the coarse layer and the active layer. The substrate has a thickness L_{s} [m] and extends from the fix datum to the interface between the substrate and the coarse layer. The substrate is numerically discretized into N_{s} [-] layers of thickness l_{s} [m] for bookkeeping the stratification. The substrate layers are subject to the constrain:

$$L_{\mathrm{s}} = \sum_{l=1}^{N_{\mathrm{s}}} l_{\mathrm{s}} \ , \tag{3.5}$$

where l is an index indicating the substrate layer from top to down.

The volume fraction content of sediment of size fraction k in the active layer, coarse layer, and substrate layer are $F_{\mathrm{a}k}$ [-], $F_{\mathrm{cl}k}$ [-], and $f_{\mathrm{s}kl}$ [-], respectively. These are constrained by equations:

$$\sum_{k=1}^{N} F_{\mathrm{a}k} = 1 \ , \tag{3.6}$$

$$\sum_{k=1}^{N} F_{\mathrm{cl}k} = 1 \ , \tag{3.7}$$

$$\sum_{k=1}^{N} f_{\mathrm{s}k,l} = 1 \ . \tag{3.8}$$

Given the thicknesses and the volume fraction contents, the volume of sediment of size fraction k in the active layer, coarse layer, and substrate per unit of bed area (which are the conserved variables per size fraction) are:

$$M_{\mathrm{a}k} = F_{\mathrm{a}k} L_{\mathrm{a}} \ , \tag{3.9}$$

$$M_{\mathrm{cl}k} = F_{\mathrm{cl}k} L_{\mathrm{cl}} \ , \tag{3.10}$$

$$m_{\mathrm{s}kl} = f_{\mathrm{s}kl} l_{\mathrm{s}l} \ . \tag{3.11}$$

3.4.2 Model behaviour

Immobile sediment in the active layer is transferred to the coarse layer at a certain rate, which causes a decrease of the active layer thickness. Only mobile sediment is transferred from the coarse layer to the active layer. This process increases the active layer thickness. If the active layer is below a certain threshold L_{a0} [m], which is the active layer thickness under alluvial conditions, there is a flux of mobile sediment (if present) from the coarse layer to the active layer at a certain rate.

If the active layer thickness has reached its alluvial value (hence there is no immobile sediment in it) and aggradational conditions are present, there is a transfer of mobile sediment from the active layer to the coarse layer such that the thickness of the active layer remains equal to its alluvial value.

If the active layer thickness has reached its alluvial value (hence there is no immobile sediment in it) and degradation conditions are present and mobile sediment is available in the coarse layer, there is a flux of mobile sediment from the active layer to the coarse layer such that the thickness of the active layer remains equal to its alluvial value. In the case there is only immobile sediment available in the coarse layer, the active layer thickness reduces.

The coarse layer has constant thickness, which simplifies the system of equations at the expenses of not capturing the complexity of the armouring process in detail. Further extension of the model should consider link the coarse layer thickness to, for instance, the thickness of the active layer or the properties of the sediment within it (cf. Figure 7).

The flux of sediment between the coarse layer and the substrate is such that the thickness of the coarse layer is preserved.

3.4.3 Conservation of mass

In this section the equations dealing with mass conservation of the entire sediment mixture are considered. For simplicity of presentation, we present the equations in a one-dimensional form. The two-dimensional equations are equivalent.

Considering a control volume from the fix datum to the bed surface, mass conservation of the entire sediment mixture is described by equation:

$$\frac{\partial z_{\mathrm{b}}}{\partial t} + \frac{\partial q_{\mathrm{b}}}{\partial x} = 0 \; . \tag{3.12}$$

In which qb is the total bed load transport expressed as sediment volume including pores.

Considering a control volume from the interface between the coarse layer and the active layer to the bed surface, mass conservation of the entire sediment mixture is described by equation:

$$\frac{\partial L_{\mathrm{a}}}{\partial t} + \frac{\partial q_{\mathrm{b}}}{\partial x} + \Phi_{\mathrm{cl}} - \Phi_{\mathrm{imm}} = 0 \; , \tag{3.13}$$

where Φ_{cl} [m/s] is the flux of sediment from the coarse layer to the active layer and Φ_{imm} [m/s] is the flux of immobile sediment from the active layer to the coarse layer (including pores).

Considering a control volume from the interface between the substrate and the coarse layer to the interface between the coarse layer and the active layer, mass conservation of the entire sediment mixture is described by equation:

$$\frac{\partial L_{\mathrm{cl}}}{\partial t} - \Phi_{\mathrm{cl}} + \Phi_{\mathrm{s}} + \Phi_{\mathrm{imm}} = 0 \; , \tag{3.14}$$

where Φ_{s} [m/s] is the flux of sediment from the substrate to the coarse layer (including pores).

Considering a control volume from the fix datum to the interface between the substrate and the coarse layer, mass conservation of the entire sediment mixture is described by equation:

$$\frac{\partial L_{\mathrm{s}}}{\partial t} + \Phi_{\mathrm{s}} = 0 \; . \tag{3.15}$$

3.4.4 Conservation of mass per size fraction

In this section the equations dealing with mass conservation per size fraction are considered. We assume a constant porosity of the bed (independent of the sediment mixture).

Considering a control volume from the interface between the coarse layer and the active layer to the bed surface, mass conservation per size fraction is described by equation:

$$\frac{\partial M_{\mathrm{a}k}}{\partial t} + \frac{\partial q_{\mathrm{b}k}}{\partial x} + \Phi_{\mathrm{cl}k} - \Phi_{\mathrm{imm}k} = 0 \quad k \in [1, N] \; , \tag{3.16}$$

where $\Phi_{\mathrm{cl}k}$ [m/s] is the flux of sediment of size fraction k from the coarse layer to the active layer and $\Phi_{\mathrm{imm}k}$ [m/s] is the flux of immobile sediment of size fraction k from the active layer to the coarse layer. The grain-size-dependent fluxes are constrained by equations:

$$\Phi_{\mathrm{cl}} = \sum_{k=1}^{N} \Phi_{\mathrm{cl}k} \; , \tag{3.17}$$

$$\Phi_{\mathrm{imm}} = \sum_{k=1}^{N} \Phi_{\mathrm{imm}k} \; . \tag{3.18}$$

Considering a control volume from the interface between the substrate and the coarse layer and the interface between the coarse layer and the active layer, mass conservation per size fraction is described by equation:

$$\frac{\partial M_{\mathrm{cl}k}}{\partial t} - \Phi_{\mathrm{cl}k} + \Phi_{\mathrm{imm}k} + \Phi_{\mathrm{s}k} = 0 \quad k \in [1, N] \; , \tag{3.19}$$

where $\Phi_{\mathrm{s}k}$ [m/s] is the flux of sediment of size fraction k from the substrate to the coarse layer which is constrained by equation:

$$\Phi_{\mathrm{s}} = \sum_{k=1}^{N} \Phi_{\mathrm{s}k} \; . \tag{3.20}$$

3.4.5 Sediment flux from the active layer to the coarse layer

The flux of immobile sediment from the active layer to the coarse layer depends on the amount such sediment in the active layer and the time scale at which it is transferred:

$$\Phi_{\mathrm{imm}} = \frac{L_{\mathrm{a}} \sum_{k=k_{\mathrm{imm}}}^{N} F_{\mathrm{a}k}}{T_{\mathrm{imm}}} \; , \tag{3.21}$$

where T_{imm} [s] is the time scale at which immobile sediment is transferred (This is not to be confused with the adaptation time scale of the alluvial dune height). Following Tuijnder and Ribberink (2010), the transferring mechanism is deep dune troughs. Hence the time scale is associated to dune celerity and dune length, which are related to the sediment transport rate. Then:

$$T_{\mathrm{imm}} = \frac{L_{\mathrm{a}} \Lambda_{\mathrm{cl}}}{q_{\mathrm{b}}} \; , \tag{3.22}$$

where Λ_{cl} is average length between the dune troughs which reach the coarse layer which is equal to:

$$\Lambda_{\mathrm{cl}} = \alpha_{\Lambda} \Lambda \tag{3.23}$$

where α_{Λ} [-] considers the effect of the active layer thickness in estimating the dune length:

$$\alpha_{\Lambda} = \min\left(\alpha_{\mathrm{m}}, 1 + (\alpha_{\mathrm{m}} - 1) \frac{L_{\mathrm{a}}}{L_{\mathrm{a0}}}\right) \; , \tag{3.24}$$

where α_{m} [-] is the maximum value, which is a parameter assumed to be equal to 20 (Tuijnder and Ribberink, 2010). Unless otherwise specified, we use a value $\alpha_{\mathrm{m}} = 1$ which decreases the model complexity. The implications of equation for α_{Λ} should be carefully assessed.

The flux per size fraction is:

$$\Phi_{\mathrm{imm}k} = \begin{cases} 0 & \text{for} \quad k \in [1, k_{\mathrm{imm}} - 1] \\ \Phi_{\mathrm{imm}} F_{\mathrm{a}k} & \text{for} \quad k \in [k_{\mathrm{imm}}, N] \end{cases} \; , \tag{3.25}$$

3.4.6 Sediment flux from the coarse layer to the active layer

Under the conditions that the active layer thickness is below the maximum alluvial value, the flux of mobile sediment from the active layer to the coarse layer depends on the amount such sediment in the coarse layer and the time scale at which it is transferred T_{cl} [s]. Under the conditions in which the active layer thickness has reached its alluvial value and aggradation occurs, sediment is transferred from the active layer to the coarse layer such that the thickness of the active layer is constant. This implies that the mass flux is equal to the aggradational rate. Under the conditions in which the active layer thickness has reached its alluvial value and degradation occurs, sediment is transferred from the coarse layer to the active layer such that the thickness of the active layer is constant, if this is present in the coarse layer. If sediment is not present, no flux occurs:

$$\Phi_{\mathrm{cl}} = \begin{cases} \frac{L_{\mathrm{cl}} \sum_{k=1}^{k_{\mathrm{imm}}-1} F_{\mathrm{cl}k}}{T_{\mathrm{cl}}} & \text{for} \quad L_{\mathrm{a}} < L_{\mathrm{a},0} \\ \frac{\partial q_{\mathrm{b}}}{\partial x} & \text{for} \quad L_{\mathrm{a}} = L_{\mathrm{a},0} \wedge \left(\frac{\partial z_{\mathrm{b}}}{\partial t} > 0 \vee \left(\frac{\partial z_{\mathrm{b}}}{\partial t} < 0 \wedge \sum_{k=1}^{k_{\mathrm{imm}}-1} F_{\mathrm{cl}k} > 0 \right) \right) \\ 0 & \text{for} \quad L_{\mathrm{a}} = L_{\mathrm{a},0} \wedge \sum_{k=1}^{k_{\mathrm{imm}}-1} F_{\mathrm{cl}k} = 0 \end{cases} , \tag{3.26}$$

The flux per size fraction depends on the volume fraction content at the interface between the coarse layer and the active layer $f^{\mathrm{I,cl}_k}$ [-]:

$$\Phi_{\mathrm{cl}k} = \Phi_{\mathrm{cl}} f_k^{\mathrm{I,cl}} , \tag{3.27}$$

where:

$$f_k^{\mathrm{I,cl}} = \begin{cases} F_{\mathrm{cl}k} & \text{for} \quad \left(L_{\mathrm{a}} < L_{\mathrm{a},0} \vee \frac{\partial z_{\mathrm{b}}}{\partial t} < 0 \right) \wedge k \in [1, k_{\mathrm{imm}} - 1] \\ F_{\mathrm{a}k} & \text{for} \quad L_{\mathrm{a}} = L_{\mathrm{a},0} \wedge \frac{\partial z_{\mathrm{b}}}{\partial t} > 0 \end{cases} . \tag{3.28}$$

Note that in the cases not specified, the flux is already zero so the volume fraction content is irrelevant.

In the current set-up, the time scale T_{cl} is assumed to be infinitely small, such that if the active layer thickness is below the maximum alluvial value and there is mobile sediment in the coarse layer, this mobile sediment will be instantly mobilized and hence transferred to the active layer. From a numerical point of view, the time scale is equal to the time step and in one time step all mobile sediment is transferred to the active layer.

3.4.7 Sediment flux from the substrate to the coarse layer

The sediment flux from the substrate to the coarse layer guarantees that the thickness of the coarse layer remains constant:

$$\Phi_{\mathrm{s}} = -\Phi_{\mathrm{cl}} . \tag{3.29}$$

The flux per size fraction depends on the sediment at the interface between the substrate and the coarse layer $f^{\mathrm{I,s}_k}$ [-]:

$$\Phi_{\mathrm{s}k} = \Phi_{\mathrm{s}} f_k^{\mathrm{I,s}} , \tag{3.30}$$

where:

$$f_k^{\mathrm{I,s}} = \begin{cases} f_{\mathrm{s}k1} & \text{for} \quad \frac{\partial z_{\mathrm{s}}}{\partial t} < 0 \\ F_{\mathrm{cl}k} & \text{for} \quad \frac{\partial z_{\mathrm{s}}}{\partial t} > 0 \end{cases} . \tag{3.31}$$

3.5 Application

In this section, the adapted model equation are applied to a flume case (Section 3.5.1) and to a field case (Section 3.5.2).

3.5.1 Struiksma flume experiment

Struiksma (1999) introduced the concept of the reduction factor for sediment transport over a fixed layer. Laboratory experiments were done in a flume with a length of 11 m and a width of 0.2 m to test the theory. The flume was filled with a depth roughly 0.15 m of fine sediment with a mean diameter D_{50} of 0.45 mm and a D_{90} of 0.90 mm. A fixed layer with a length of 3 m was constructed halfway the flume using medium gravel 0.08 - 0.16 m. In the upstream portion of the flume a trench was excavated with a depth of 4 cm and a length of 2 m.

An initial bed composition is shown in a transect over the length of the flume in Figure 16. The interface between the active layer and the coarse layer is shown in green and between the coarse layer and the substrate layer is shown in cyan. When the model has no coarse layer, the cyan line shows the thickness of the first substrate layer.

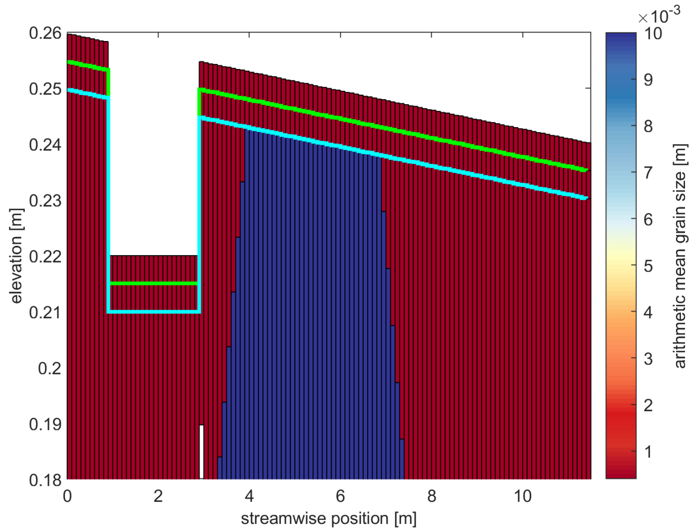

Figure 16 Initial bed level and arithmetic grain size for the T2 experiment of (Struiksma, 1999). In this figure, the green and cyan lines only indicate the thickness of the layers in the initial condition, which are not the active layer and coarse layer thicknesses.

Seven simulation have been run for testing the modified system of equation using different model concepts (Table 1). From this seven simulations, the Simulations S6 and S7 are implementation tests which are discussed in Appendix A.

The first simulation result is shown in Figure 17. This is the case where the immobile sediment is simply removed from the simulation and the concept of Struiksma is applied. This is the behaviour we seek when using the coarse layer implementation. The final equilibrium situation is equal to the initial one, which is crucial as otherwise the results are physically unreallistic (Figure 18).

Figure 19 shows the result of the active-layer model (Hirano, 1971) (i.e., without additional sorting fluxes of immobile sediment). The fixed layer is modelled as coarse immobile sediment, rather than as a lack of sediment. As the trench passes over the immobile sediment layer immobile sediment enters the active layer, as this concept allows for this. After the trench has passed and the sediment transport gradients vanish arriving at equilibrium conditions, the bed level rises above the initial elevation

Case	Model	Sediment fraction(s)	Mobility	α_{m}
S1	Struiksma	fine	-	-
S2	Hirano	fine & coarse	-	-
S3	Adapted coarse layer	fine & coarse	Shields (discrete)	1
S4	Adapted coarse layer	fine & coarse	Shields (continuous)	1
S5	Adapted coarse layer	fine & coarse	Wilcock and McArdell (continous)	1
S6	Adapted coarse layer	fine & coarse	Shields (discrete)	20
S7	Adapted coarse layer	2 x fine & coarse	Shields (discrete)	1

Table 1 Simulations to model the T2 flume experiment of Struiksma (1999).

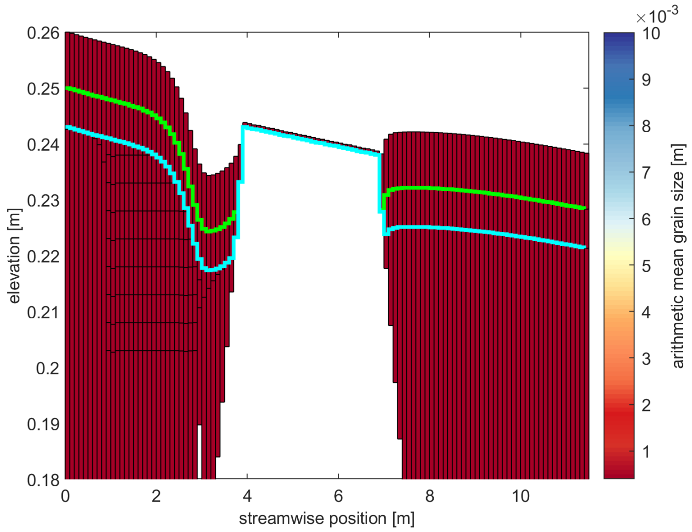

Figure 17 S1 bed level and arithmetic grain size after 1.65 h. The cyan line now does not denote the coarse layer but simply the second substrate layer.

which is unphysical (cf. Figure 20). This is caused by the fact that immobile sediment in the active layer is transported upwards as the bed level increases. Hiding enhances the unphysical aggradation, as it reduces the transport of fine sediment which implies that a larger velocity (i.e., a shallower flow and a larger bed level) is necessary for reaching equilibrium.

Simulation S3 employs the sorting of immobile sediment to the coarse layer. As in simulation S2, the fix layer is modelled using coarse immobile sediment. The trench passes on top of the fix layer without entraining immobile sediment. Worded differently, immobile sediment does not enter the active layer at any time. Using this model we correctly capture the “Struiksma behaviour” (i.e., immobile sediment forming a fix layer is not part of the active layer). As expected, the final equilibrium situation is equal the initial one (Figure 22). It is worth mentioning that there is a slight mixing of sediment in substrate bookkeeping layers at the edges of the trench. This behaviour is expected as immobile sediment mixes with the layers above when the trench passes. This is not a modelling problem but a resolution problem. A larger number of bookeeping layers would reduce this spurious effect. The propagation of the trench in the case of the S3 simulation appears to be slightly slower that the S1 case. We currently have no explanation for this effect.

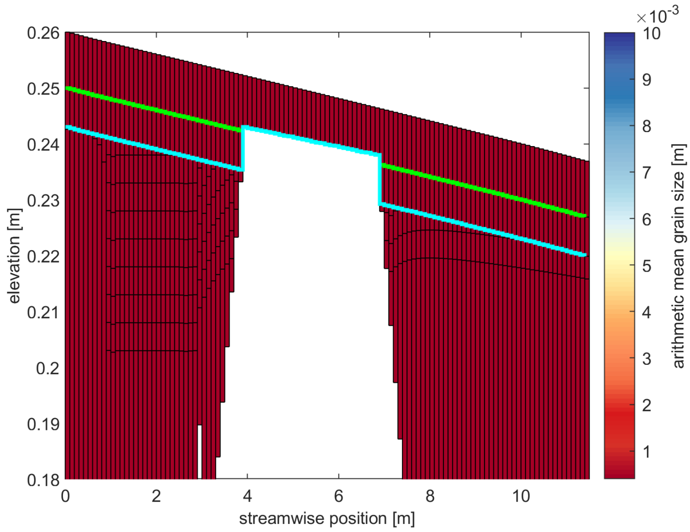

Figure 18 S1 bed level and arithmetic grain size after 7.00 h. The cyan line now does not denote the coarse layer but simply the second substrate layer.

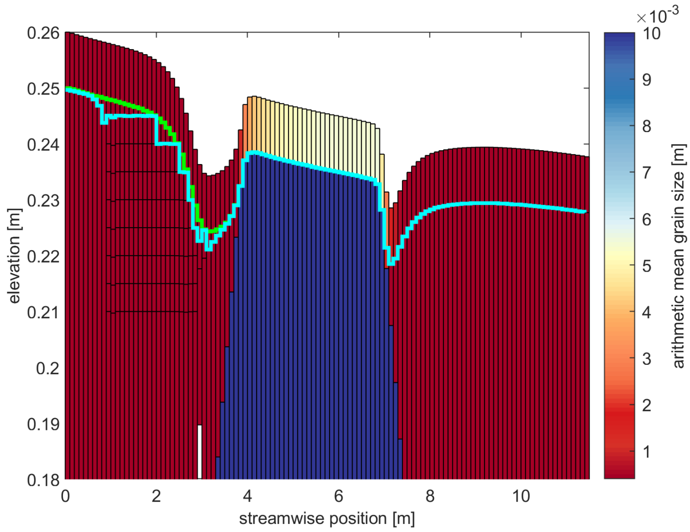

Figure 19 S2 bed level and arithmetic grain size after 1.65 h.

Simulations S4 and S5 show the effect of considering a continuous function for mobility rather than the discrete Shields (1936) formulation (Simulation S3). In the case of a continuous formulation of Shields (1936) (Simulation S4), although continuous, the formulation estimates a mobility of the coarse sediment so low that it is numerically rounded to 0. Similarly, the mobility of the fine sediment is rounded to 1. Hence the results of the continuous formulation provide the same results as the ones of the discrete formulation (Figures 23 and 24).

On the contrary, the formulation of Wilcock and McArdell (1997) provides different results. As the trench passes, “immobile” (it is partially mobile according to Wilcock and McArdell (1997)) sediment initially forming the fix layer is entrained into the active layer (Figure 25). This behaviour is contrary to the “Struiksma effect” we intend to capture. As the trench passes, part of the coarse sediment moves upward with the active layer as in Simulation S2, eventually leading to an unrealistic final state (Figure 26). Running for a long time does not lead a realistic final state. This is because, as coarse sediment sinks into the coarse layer, sediment from the coarse layer is

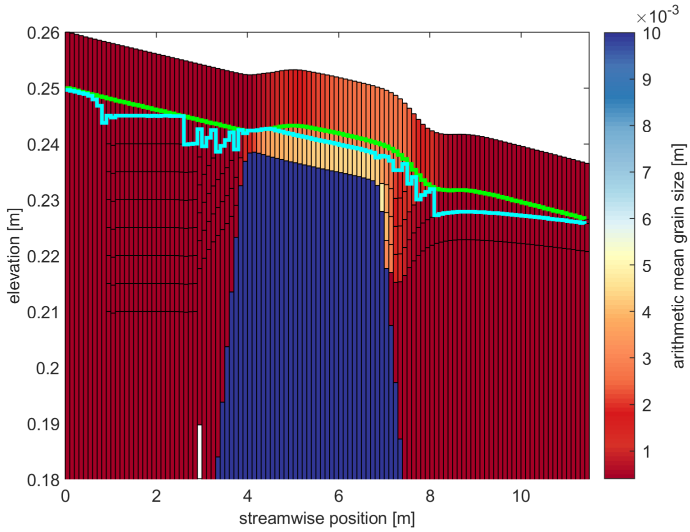

Figure 20 S2 bed level and arithmetic grain size after 7 h

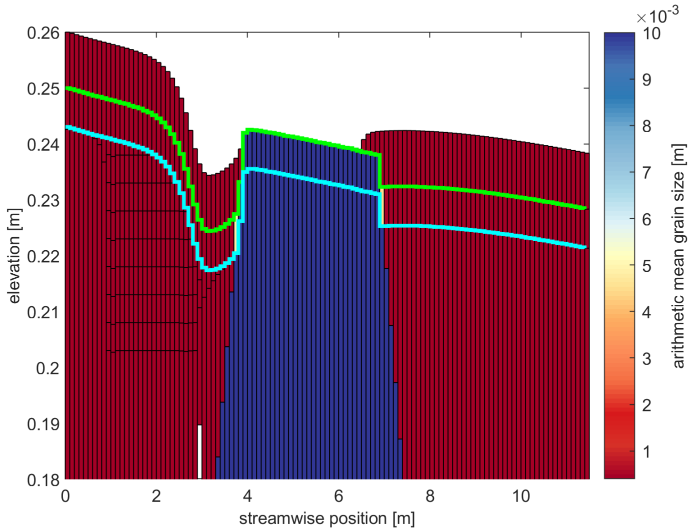

Figure 21 S3 bed level and arithmetic grain size after 1.65 h.

transferred to the active layer (see Sections 3.2.4 and 3.6.2). It is necessary to account for a discrete mobility formulation to properly capture the “Struiksma effect” of a fixed layer.

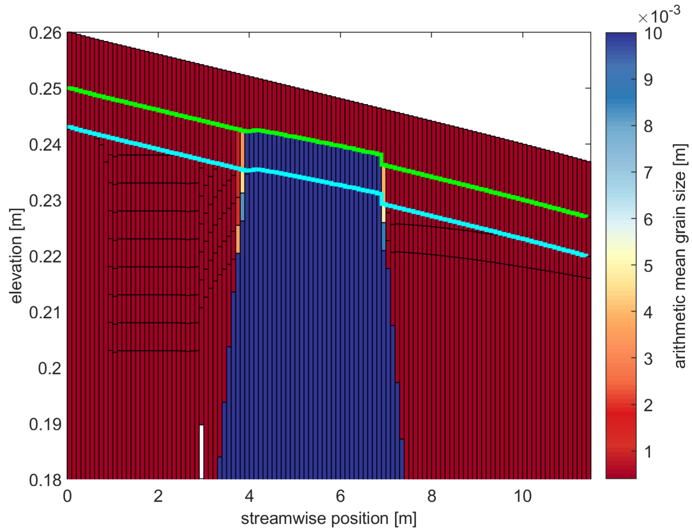

Figure 22 S3 bed level and arithmetic grain size after 7 h

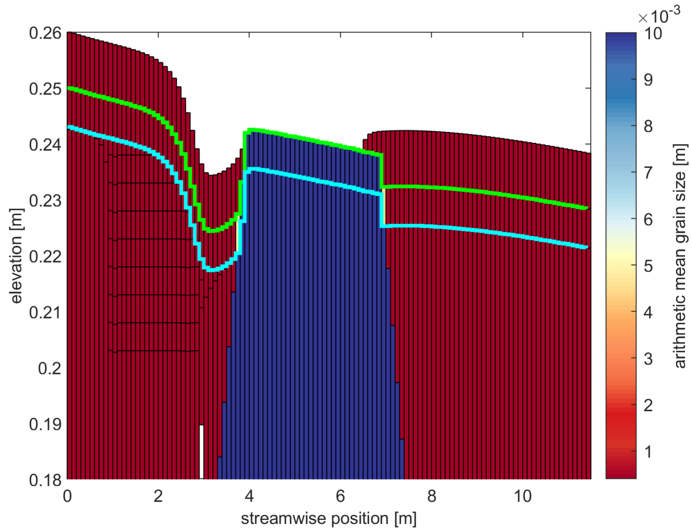

Figure 23 S4 bed level and arithmetic grain size after 1.65 h.

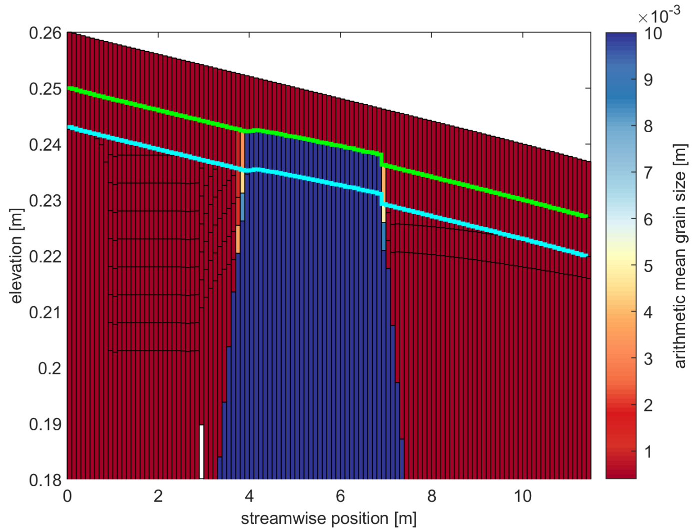

Figure 24 S4 bed level and arithmetic grain size after 7 h

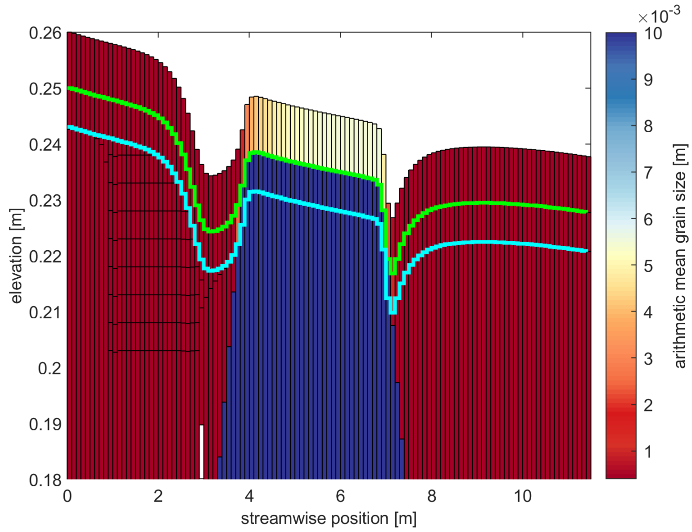

Figure 25 S5 bed level and arithmetic grain size after 1.65 h.

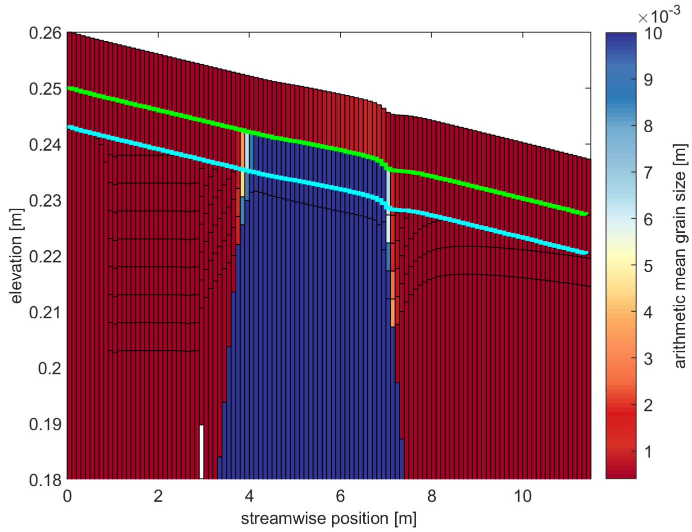

Figure 26 S5 bed level and arithmetic grain size after 7 h

3.5.2 Fixed layer at Nijmegen

In this section, the modelling concepts are applied to a field scenario.

3.5.2.1 Simulation plan

Morphodynamic development along the Upper Waal, which includes the fixed layer at Nijmegen is modelled. Four simulations in which we model the fixed layer as:

- N1: an unerodable layer (i.e., Struiksma),
- N2: coarse sediment using the active-layer model,
- N3: coarse sediment using the active-layer model without considering hiding,
- N4: coarse sediment using the adapted set of equations.

3.5.2.2 Simulation set-up

The model domain “wl2a” of the DVR model schematization (Ottevanger *et al.*, 2015) is selected (Figure 27), and boundary conditions for a constant discharge at Lobith equal to 3053 m^3/s are created based on results from the full DVR model (Figure 28). The upstream morphodynamic boundary condition that we impose is fixed bed level and compositions. The sediment mixture is composed of 10 sediment size fractions with characteristic grain sizes between 63 µm and and 64 mm. In the cases in which the fixed layer is represented by coarse sediment, the coarse sediment forming the fixed layer has a representative diameter equal to 0.35 m (I en W, 2018).

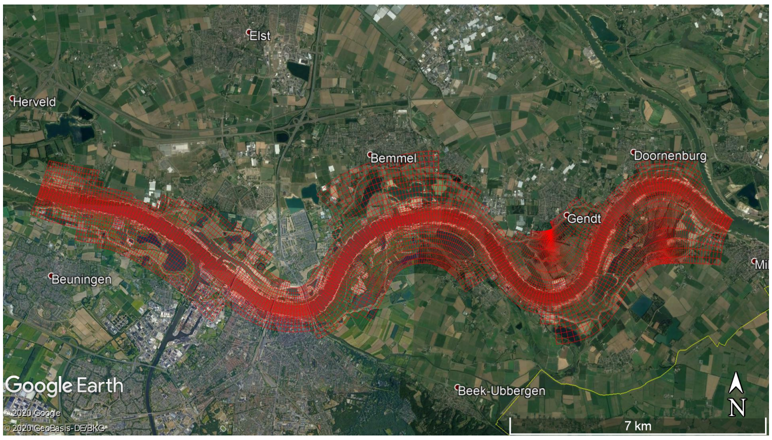

Figure 27 Simulation domain and grid.

The sediment transport rate is modelled using the generalized Meyer-Peter and Müller (1948) sediment transport relation accounting for the hiding effect by means of the closure relation by Egiazaroff (1965) as modified by Ashida and Michiue (1972).

By using a morphodynamic acceleration factor equal to 240, we are able to model 1000 days morphodynamic development. We have tested that the morphodynamic acceleration factor causes errors below 1% in propagation and dampening of infinitesimal perturbations. The initial grain size distribution corresponds to the situation in 2016.

The active layer thickness is set to 0.5 m, which is representative of the mixing processes due to 1 m height dunes. As such, Struiksma’s factor for reducing the sediment transport is made equal to the active layer thickness (Chavarrías and Ottevanger, 2019).

The sediment thickness is everywhere in the main channel sufficiently thick such that there is no lack of sediment, except for the location where there is the fixed layer in this is modelled considering Struiksma’s approach.

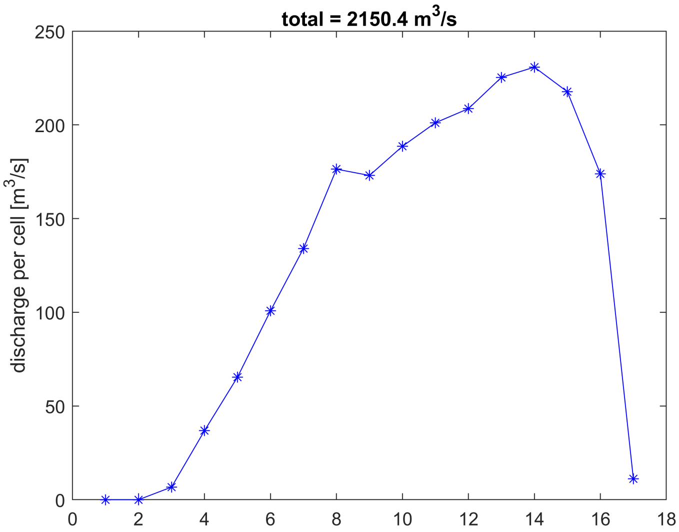

Figure 28 Discharge per cell for a discharge at Lobith equal to 3053 m^3/s.

Based on the latest measurements and data analysis by De Jong and Ottevanger (2020), the elevation of the fixed layer is updated (Figure 29). The points from the data analysis in which the bed elevation is below −1.5 m are considered to be error and are filtered out. The data analysis shows that the elevation of the fixed layer is in general slightly higher than it was in the previous model schematization (Figure 30). We consider that initially there are 5 cm of sediment above the fixed layer. Note that, as the model does not resolve individual bed elevation perturbations such as dunes and ripples, the bed elevation represent the mean value after filtering for those bed elevation perturbations. Hence, the fact that we consider that initially there 5 cm of sediment does not mean that in the field the fixed layer is fully cover but that, in average, there are 5 cm of sediment. It is unknown to us what the real mean sediment thickness is, but such a thickness could possibly also be determined based on the data by De Jong and Ottevanger (2020) (.e.g mean bed level - minimum bed level).

3.5.2.3 Results

Figures 31 and 32 show the initial condition along a longitudinal section for the case in which the fixed layer is modelled using Struiksma's model and coarse sediment, respectively. Note that in the first case there is no sediment below the top 5 cm of sediment. In the second case, below the top 5 cm with the original grain size distribution lays the fixed layer formed by coarse sediment.

As the active layer thickness is set equal to 0.5 m, at the start of the simulation the top 5 cm are mixed with 45 cm of coarse sediment. This is physically realistic, as it represents a situation in which most of the sediment exposed at the bed surface and interacting with the flow is coarse sediment (Chavarrías and Ottevanger, 2019).

Figures 33 and 34 show the final condition for Simulation N1 and N2, respectively. In both cases, an aggradational wave passes over the fixed layer. While in the simulation using Struiksma's concept (N1), the effect of the passing aggradational wave is not visible in the final condition, in the simulation in which the fixed layer is modelled using coarse sediment (N2), the bed has significantly aggraded

This unrealistic effect can be explained from the combination of two effects. The first is that the presence of the coarse sediment in the active layer reduces the mobility of the

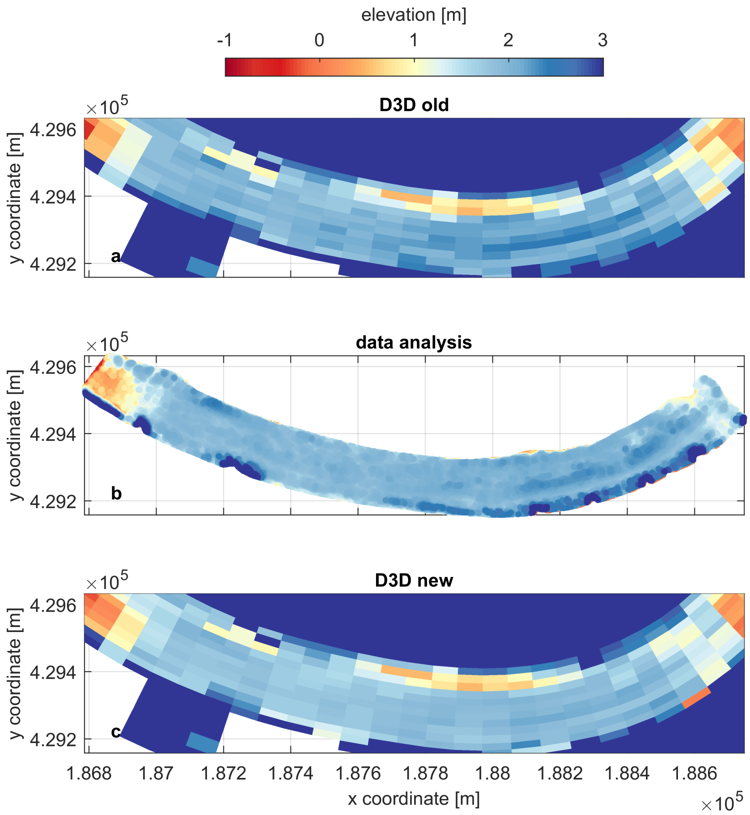

Figure 29 Bed elevation of the fixed layer in the old schematization (top), according the the data analysis (centre), and in the new schematization (bottom).

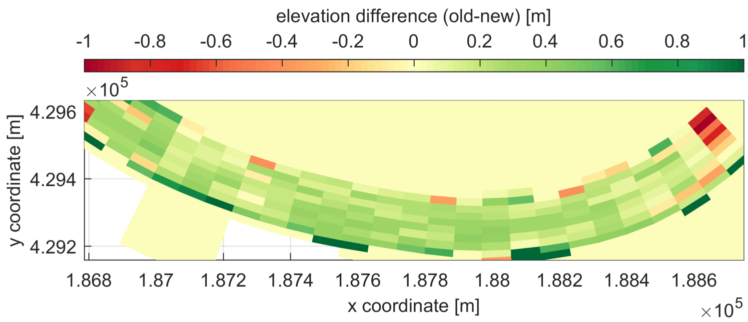

Figure 30 Difference in elevation between the old and new schematization.

fine sediment due to hiding. Thus, further aggradation occurs until the amount of coarse material at the bed surface is relatively small and fine sediment becomes mobile again.

The second factor is the unrealistic transfer of coarse sediment in upward direction under aggradational conditions. As the bed aggrades, sediment in the active layer is transferred to the substrate. The aggradational flux to the substrate does not differentiate between mobile or immobile sediment, as the flux has the same

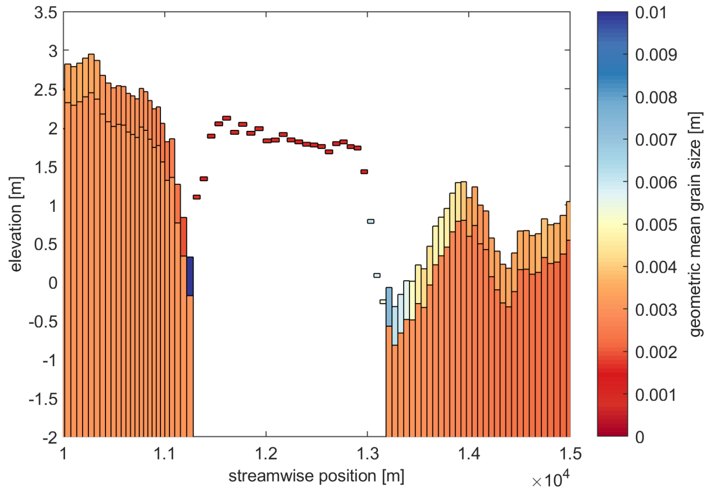

Figure 31 Initial condition in Simulation N1 (Struiksma).

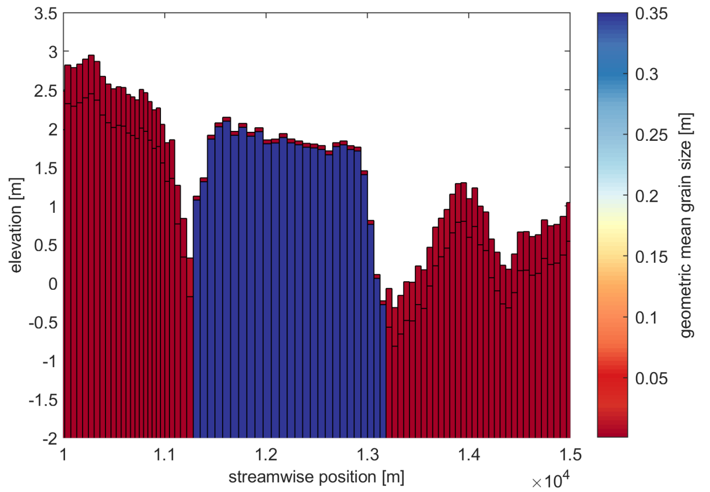

Figure 32 Initial condition in Simulation N2 (Hirano).

composition than the active layer. For this reason, in the aggradational process fine sediment is transferred area that was initially composed of immobile sediment only, and immobile sediment in the active layer moves in the upward direction.

Not accounting for the hiding effect solves part of the problem (Figure 35). In this case, the coarse sediment in the active layer does not reduce the mobility of the fine fraction and the bed aggrades significantly less than in with hiding. Nevertheless, the second problem is not resolved, and coarse sediment moves in the upward direction unrealistically.

Simulation N4 using the same morphophonemic acceleration factor as the other cases results in physically unrealistic bed elevation changes. At this moment this behaviour remains unexplained. From a numerical point of view, bed level changes are computed first and in the same way in all modelling concepts and then sorting of the underlayers is performed. The morphodynamic acceleration factor modifies the bed level changes only. The flux of sediment from the active layer to the coarse layer depends on the sediment transport rate. It may be possible that we are unaware of a

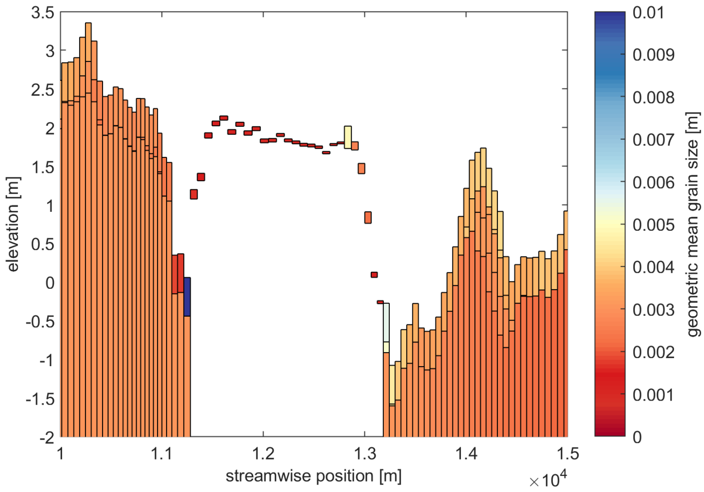

Figure 33 Final condition in Simulation N1 (Hirano).

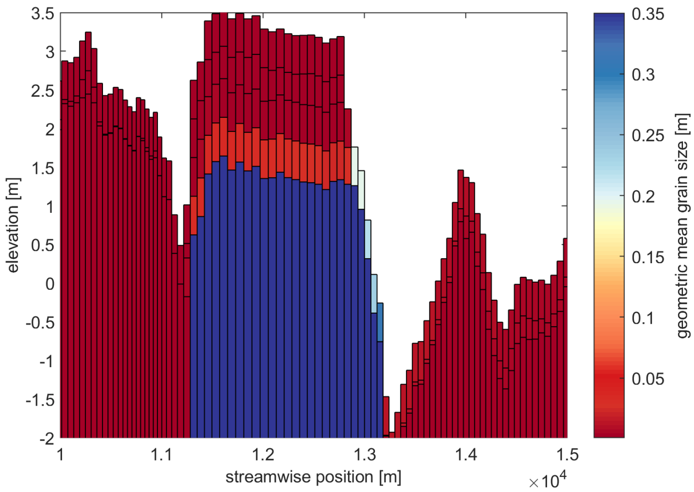

Figure 34 Final condition in Simulation N2 (Hirano).

dependency on the morphodynamic acceleration factor throughout the computation of the underlayer fluxes.

Simulation N4 is run using a morphodynamic acceleration factor equal to 1. This prevents from running the same simulation time as the other cases. Figure 36 shows the results after 1 week. The behaviour is as expected. Immobile sediment does not enter the active layer, which at the locations shown has a thickness equal to zero on top of the fixed layer, indicating that there is no mobile sediment at the bed surface.

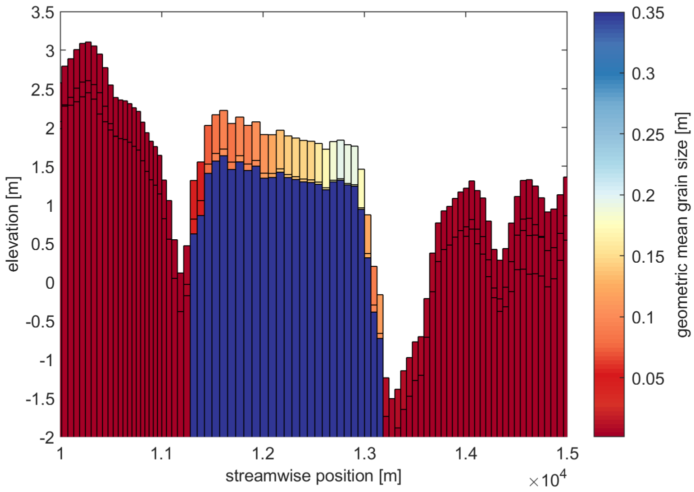

Figure 35 Final condition in Simulation N3 (Hirano no hiding).

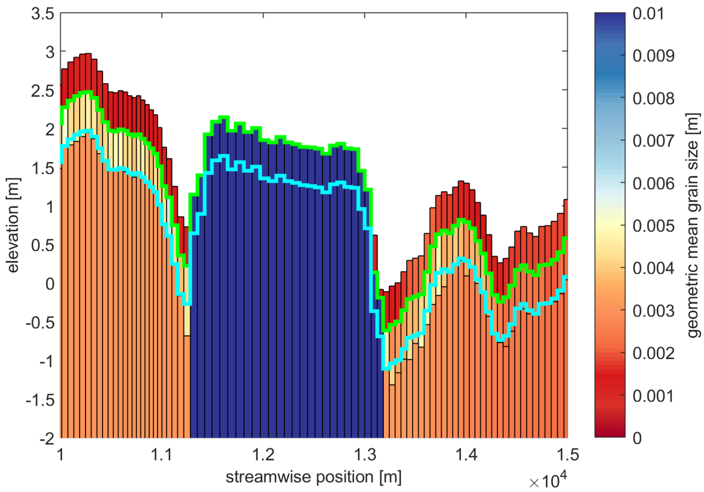

Figure 36 Final condition in Simulation N4 (adapted model).

3.6 Discussion

3.6.1 Presence of immobile material in the active layer

The active layer represents the part of the bed that interacts with the flow. “Interact with the flow” is a vague concept. What it means is that sediment in the active layer:

1 provides friction,
2 can at any moment be set into transport if the bed shear stress is large enough, and
3 it affects the transport of all other sediment particles due to hiding and exposure.

The list is not exhaustive. These are at least some properties of the what “interacting with the flow” means.

Clearly, an immobile sediment particle at the bed surface does all of the above. Hence, it must be part of the active layer. A way of determining if a sediment particle interacts with the flow is to assess if the particle is exposed to the flow. As there exist bedforms which are not resolved in the active-layer model, the concept of being exposed to the flow is better studied in a probabilistic manner and the part of the bed that is exposed to the flow can be said to be the thickness of the bed covering 95% of bed elevation perturbations (Ribberink, 1987).

The choice of separating the mobile and immobile sediment, and leaving in the active layer, is not a straightforward decision. The difficulty comes from the lack of a well defined conceptual frame work. It can be illustrated that dealing with mobile sediment requires usage of elements that have been tuned and developed together, rather than combining separately developed concepts. This is shown in the following examples. If only mobile sediment is part of the active layer, several conceptual issues appear, for instance:

- Suppose a case in which all sediment is immobile. What is the active layer thickness? The active layer thickness is not to be confounded with the bedload layer. Although all sediment is immobile, sediment is available to be transported and must be part of the active layer. The active layer thickness cannot be 0 from a theoretical perspective as this would lead to changes in the bed composition which travel infinitely fast. Moreover, there is always a part of the bed that interacts with the flow. The concept of Struiksma reducing the risk of this occuring by reducing the transport in such a case. The current implementation does something similar, but as there is a flux of immobile sediment to the coarse layer, the risk of this occuring may be larger.
- Suppose a case in which there is large proportion of coarse sediment and some fine sediment, and only the fine is mobile. Do you compute the transport of fine sediment neglecting the effect of the coarse? Do you neglect the coarse sediment in computing friction? If only mobile sediment plays a role, the sediment transport capacity of the fine sediment would be the same regardless of the size of the coarse sediment. In addition, the interaction between the coarse and fine sediment may play a role (cf. Section 3.6.4).
- What happens when using a sediment transport closure relation in which all sediment is always mobile (e.g., Engelund and Hansen, 1967; Wilcock and Crowe, 2003)? A criterion for incipient motion can be used to assess whether sediment is mobile. However, this leads to the contradictory reasoning that sediment is transported and nevertheless it sinks into the coarse layer and stops being transported. A way to overcome this limitation is to restrict the usage of the model to sediment transport relations with a criterion for motion, although this severely limits general applicability.

It is relevant to comment on the behaviour of the model of Struiksma (1999) to deal with fixed layers. His model was derived for unisize conditions, in which there is no active layer as such and the question of which part of the bed interacts with the flow is irrelevant, as all the sediment has the same properties. In the extension to mixed-size sediment conditions of the model of Struiksma (1999) in Delft3D, the active layer is made equal to sediment layer on top of a fixed layer, when this exists. Hence, also in this case, immobile sediment forming the fixed layer does not interact with the flow and does not contribute to the hiding-exposure effect or friction. For practical situations with large rocks, bed rock or even concrete it is still possible to make reasonable simulations as shown in previous sections, especially considering that also several other components (such as sediment-transport relations and roughness predictors) are also not valid for simulating the conditions that occur above such fixed layers. This again also reflects the need to think towards application in an integral sense, rather than developing very advanced models for each of the independent components.

On the other hand, immobile sediment in the active layer is not transferred out of it. The situation in which there is a fixed layer and the active layer above contains immobile sediment is possible. Hence, part of the problems mentioned above are overcome, but others are still open and may require new concepts and further research.

3.6.2 Mobile sediment flux due to immobile sediment deposition

In both the original and the adapted Tuijnder and Ribberink (2010) model, a flux of immobile sediment from the active layer to the coarse layer is replaced by a flux of mobile sediment from the coarse layer to the active layer. This leads to the contradictory results shown in Section 3.2.4. From our perspective, immobile sediment should simply remain in the active layer when using the proposed concepts (Section 3.6.1). Nevertheless, if immobile sediment is transferred out of the active layer, we advocate for not replacing it with mobile sediment in the coarse layer.

Consider a case representing a situation in which immobile sediment in the active layer can be thought to be transferred out of the active layer. For instance, flow decrease after a flood wave. During this stage, dune height decreases and a sediment fraction previously mobile becomes immobile. The deposition of immobile sediment does not “push” sediment from below upwards. Sediment below what was considered the active layer (i.e., below 95% of dune height) during the high-flow is not made available for transport due to deposition of immobile sediment. Under low flow conditions, only sediment that was at the bed surface during high-flow condition is at the bed surface during low flow conditions. Hence, only sediment that was in the active layer must form part of the active layer.

Immobile sediment deposition in this approach must therefore lead to a decrease in active layer thickness without a “return flux”.

3.6.3 Use of the active-layer model under conditions with immobile sediment

The current model available for predicting morphodynamic development under mixed-size sediment conditions is the active-layer model (Hirano, 1971). The implementation of this concept in Delft3D is developed and valid for conditions where all sediment fractions in the active layer contribute to sediment transport, including immobile fractions. However, the implemented concept has not been developed for undersupplied conditions over coarse immobile sediment. Improving this concept for such conditions requires several adjustments to subcomponents as well. The approach as presented in this report above can be considered as an attempt to extend the active layer model, but to such an extent that some of the generic character of the active-layer model has been lost.

In the active-layer model in Delft3D, the predicted bed elevation and grain size distribution represent values averaged over the passage of several bedforms. Worded differently, the effect of individual bedforms is not modelled. Changes in bed elevation are predicted on the basis of gradients in the total sediment transport rate and changes in grain size distribution of the bed surface are predicted on the basis of gradients in the sediment transport rate of each size fraction. The bed is discretized in two parts: the active layer and the substrate. Sediment in the active layer interacts with the flow, while sediment in the substrate is assumed to be too deep to be part of morphodynamic development.

The idea that sediment interacts with the flow means that sediment in the active layer (1) can provide hydraulic friction, (2) can at any moment be set into transport if the bed shear stress is large enough, (3) it affects the transport of all other sediment particles due to hiding and exposure. The list may not exhaustive, but these are at least some properties of the concept of interacting with the flow.

The active-layer model is dependent on a crucial parameter: the active-layer thickness. This parameter sets the part of the bed that interacts with the flow. This is essentially different from the bed load layer thickness (e.g. Van Rijn, 1984a; Luu *et al.*, 2004; Wu and Yang, 2004; Colombini, 2004; Colombini and Stocchino, 2005) which represents the part of the bed that is moving. The sediment in the active layer is available for being transported and sediment in transport deposits in the active layer. For this reason, the thickness of the bed load layer in a fully immobile bed is equal to zero, but not the active-layer thickness. Under plane-bed conditions and on short time scales, the active-layer thickness is usually associated to a characteristic grain size of the bed surface, and under bedform-dominated conditions the part of the bed that interacts with the flow is usually related to a characteristic bedform height. In general, the active-layer thickness covers a significant percentage (e.g., 95%) of the probability distribution of bed elevation fluctuations around the mean value (averaging over the passage of several bedforms) (Ribberink, 1987; Blom *et al.*, 2003). The dependence on the alluvial active-layer thickness and the lack of proper predictors for a thickness is one of the major weak points of the active-layer concept when using it in practice.

The sediment transport rate is modelled by means of a closure relation that predicts the sediment transport rate per size fraction based on the composition of the bed surface. This closure relation needs to account for the fact that fine grains in a mixture of sediment of different sizes hide behind coarse grains and are thus less exposed to the flow than under unisize conditions and viceversa (Einstein, 1950). As the sediment transport rate is modelled by means of a closure relation rather than a conservation equation accounting for the fluxes, the mass of sediment in transport is not conserved. Particles that suddenly stop moving do not cause an increase in bed elevation, only gradients in transport do. For small concentrations of moving particles and when the adaptation time scale to changing flow conditions is fast with respect to changes in bed elevation, this assumption is valid (e.g. Armanini and Di Silvio, 1988; Garegnani *et al.*, 2011, 2013). Many of the predictors are based on highly scattered data from laboratory or field measurements, and do have limited accuracy when used outside the range of conditions for which they were developed. Furthermore, they have not been developed as an integral part of the morphodynamic model and all its other components (such as active-layer model, flow model, roughness, etc.). All these processes interact in a very complex way with various time scales, as can be seen already in this study.

In a situation in which there are no gradients in the sediment transport rate per size fraction (and as such the mean bed elevation is constant), the only process that can lead to a change in surface grain size distribution is a lowering of the interface between the active layer and the substrate due to, for instance, an increase in the active-layer thickness. This is a limitation of the active-layer model, as several processes are inadequately described in this manner. For instance, dune growth under normal flow conditions (i.e., without change in mean bed elevation) causes the formation of a coarse layer underneath migrating dunes Blom *et al.* (2003). Lee-face sorting causes the deposition of coarse sediment at the dune troughs. As it often happens, the coarse sediment is immobile and dunes become composed of the fine fractions only. The coarse layer inhibits the entrainment of fine sediment and limits the sediment transport rate. Although the formation of such a coarse layer is not modelled by the active-layer model, the active-layer model does account for the transport of some of the sediment fractions while some other sediment fractions remain immobile, but without accounting specifically for the undersupplied phenomena (e.g., development of barchan dunes and mixed friction). Whether a particular sediment fraction is mobile or not depends on the closure relation for the sediment transport rate (considering hiding) and the amount of the particular sediment fraction relative to the total sediment at the bed surface (i.e., in the active layer). The reduction in sediment transport is intrinsic to the fact that there is sediment in the active layer which is not mobile

From the above explanation it is clear that when using this active-layer approach the immobile sediment at the bed surface should form part of the active layer. If this would not be the case, several conceptual issues would appear as explained in Section 3.2.1.

The active-layer model can account for situations of partial mobility to a great extent. When using it for modelling an immobile sediment layer the behaviour depends on the thickness of the top layer of coarse sediment. The first option is that, initially, the top layer of coarse sediment is larger than the variability in bed elevation due to bedforms. In this case, the active layer thickness is thinner than the top coarse layer and the fine sediment does not play a role. When a flood occurs, the dunes grow. This is modelled by an increase in active layer thickness, which eventually causes entrainment of fine sediment. The active-layer model can account for this effect. When the flood recedes, the bed surface is composed by both fine and coarse sediment. The coarse sediment in the bed surface affects the transport of fine sediment, which is physically realistic. The situation in which dunes composed of fine sediment travel on top of coarse sediment is implicitly accounted for by the fact that the active layer contains both fine and coarse sediment, but the sediment transport rate is composed by fine sediment only. Obviously in these conditions the sediment-transport rate and hiding and exposure functions cannot be randomly picked from the available functions. It is very relevant to consider functions that have been developed for a range of conditions that apply in these situations. For instance, the adapted Egiazaroff function has been developed for poorly sorted mobile mixtures, and will fail to grasp the correct behaviour of partial mobile undersupplied conditions.

The second option is that, initially, the top layer of coarse sediment is thinner than the the variability of bed elevation. In this case, from the start, the active layer is composed of both fine and coarse sediment and the same processes as above occurs.

Under aggradational conditions when the coarse sediment is immobile, there is an unrealistic result due to the simplifications of the active-layer model Chavarrías and Ottevanger (2019). Immobile sediment moves in the upward direction. Solving for this problem would require modification of the aggradation flux of sediment from the active layer to the substrate. This flux should first be formed out of coarse immobile sediment to prevent it from moving upwards.

3.6.4 Hiding on a fixed layer

While not considering hiding limits excessive aggradation when modelling a fixed layer using coarse sediment and the active-layer model, this is not a viable solution. This is for two reasons. First, hiding plays a role in the sediment transport on top of a fixed layer and it should be considered. It is a fact that the current expressions are not suitable for modelling mixtures with differences in grain size as large as we are considering but hiding needs to be considered. Moreover, currently hiding is a general property of the model and disabling it implies that nowhere in the model hiding is taken under consideration, which is physically unrealistic.

This is not a problem when using the adapted model, as immobile sediment sinks to the coarse layer and does not cause the excessive aggradation. However, the transport of fine sediment is unaffected by the fact that it is actually transported on top of coarse sediment (see Section 3.6.1). A possible solution is to compute hiding based on a mean grain size that considers the sediment at both the active layer and the substrate.

3.6.5 General modelling approach

Various processes are closely intertwined when modelling morphodynamic changes. Model concepts for each process separately cannot be simply be combined to provide the right overall behaviour. Ideally, it is best to blend and adjust all these components together when modelling the specific conditions of, for instance, fixed layers and undersupplied situations. The ideal model is integrating versions of all the submodels that are treated separately in each of the sections explaining the methods for determining morphodynamic development (Section 2). The model-approaches developed by Blom *et al.* (2003) and Tuijnder and Ribberink (2010) are examples of attempts to develop such integrated approaches.

3.7 Conclusions and recommendations regarding model development

Currently, there are three different manners to predict morphodynamic changes in the presence of immobile sediment. The first approach is to assume that immobile sediment represent an unerodible layer (i.e., Struiksma (1999) approach). The main limitation of this approach is that it prevents entrainment of sediment which is immobile only under certain conditions.

The second approach is to model immobile sediment using the standard mixed-size sediment model (i.e., the active-layer model (Hirano, 1971)). The limitation of this approach is that immobile sediment can be transported upwards leading to physically unrealistic results. In addition, specific sub-models in Delft3D (sediment transport, hiding and exposure, friction, etc.) need to be revisited for the undersupplied conditions because the existing ones do not grasp the physics fully (they were derived for fully alluvial conditions).

The third approach is to model immobile sediment using the model developed by Tuijnder and Ribberink (2010). This approach presents several limitations (3.2).

An alternative model formulation that simplifies the model by Tuijnder and Ribberink (2010) has been proposed. Using the alternative model, the laboratory experiment by Struiksma (1999) is successfully modelled for the first time considering the fixed layer as coarse sediment. Application of the model to a field case shows the expected behaviour, although a morphodynamic acceleration factor cannot be used, which strictly limits the model applicability. Moreover, the modelling concept is subject to some of the same limitations of the model by Tuijnder and Ribberink (2010).

In preventing the limitations previously discussed (Section 3.6), our proposed solution is to allow immobile sediment to be part of the active layer and to modify the aggradational flux to the substrate to take into consideration that, if there is immobile sediment in the active layer, this sediment is the first being transferred to the substrate, as it occurs in reality. This would prevent that immobile sediment is unrealistically transported upwards. As immobile sediment is part of the active layer, it interacts with the flow contributing to friction and hiding. Moreover, it can be transported if the flow conditions allow it. All the limitations regarding checking sediment mobility are non-existent. Overall, the model is simple, which provides robustness, but it contains all the essential ingredients. We expect that the approach will provide a first step towards a robust tool for modelling the situations that involve armoured layers.

Under conditions in which a fixed layer is made with very coarse sediment compared to the sediment in transport, using the state-of-the-art hiding relations may yield unrealistic results. Nevertheless, this is not a symptom that one should not consider that immobile sediment forms part of the active layer but a symptom that the hiding relation is being applied under conditions in which it has not been derived for. Thus, adaptation of the hiding function may be necessary. This also hold for several other submodels, e.g. the hydraulic roughness predictor, the sediment-transport predictor and the predictor of dune dimensions.

4 Conclusions and recommendations

The current report provides a literature review of different methods of determining morphodynamic development. These ranged from exper tknowledge, concepts of sediment mobility, and subsequently short and long term response.

Next a review of the Tuijnder and Ribberink (2010) model was done. Different shortcomings were identified and after a discussion with *Rijkswaterstaat* a plan was made to adapt the current implementation. The new model theory was written up, which required some serious rethinking to write down the processes as implemented in the code. It is recommended to rename variables in the code to match the variables as described in Section 3.4.

Finally, the model was applied to the flume experiment of Struiksma (1999), however now the experiment was modelled using a coarse sediment fraction to model the immobile sediment. An application to the fixed layer at Nijmegen showed good results, however it was not possible to run this simulation with a morphological factor larger than 1. This indicates there is probably still something to investigate related to the morphological factor. For the model, two suggestions for improvement are a model for the coarse layer thickness and a keyword to disable the entrainment of mobile sediment from the active layer in the case of sinking immobile sediment.

It is recommended to summarize the model findings in a peer reviewed journal paper, also as the theroretical framework has been expanded on.

It is recommended to implement the code in the main line of Delft3D-FM, which will enable that the less time needs to be spent on recompiling the code and making it available for others. Possibly the River Lab platform is an interesting option in this regard.

For the next year, it is recommended to apply the model to a field case (e.g. the bifurcation of Pannerden). It would even be nicer to have a case in which the formation and break up an armour layer can be modelled and compared to observations.

Furthermore there could be an opportunity to look into previous experiments of Tuijnder (2010) andBlom *et al.* (2003), to see if the current adapted model succeeds in modelling this. Discussion on the role of immobile sediment in the active layer could possibly be finalised. If the current experiments are not enough to finalise the discussion, new experiments could be thought up and perhaps executed in conjunction with the university (e.g. hiding and exposure in bimodal, trimodal mixtures)

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A Implementation tests

Simulation S6 is an implementation test. This simulation is equal to Simulation S3 (cf. Section 3.5.1) except for the fact that α_{m} is larger. This parameter increases the flux of immobile sediment from the active layer to the coarse layer. As no immobile sediment is initially present in the active layer and it cannot enter it because mobility is discerned in a discrete manner, this parameter should have no effect on the simulation compared to Simulation S3. We only find negligible differences which we assume are due to numerical discretization of the fluxes (Figures 37 and 38).

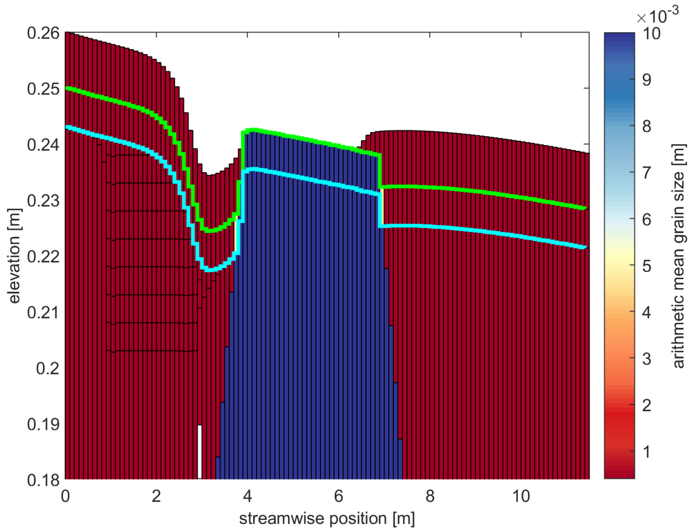

Figure 37 S6 bed level and arithmetic grain size after 1.65 h.

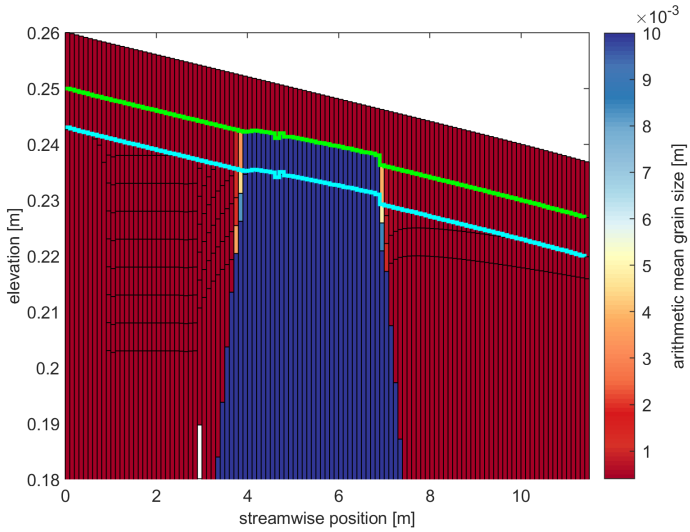

Figure 38 S6 bed level and arithmetic grain size after 7 h

Simulation S7 is an implementation test. This simulation is equal to Simulation S3 except for the fact that two mobile fine fractions with equal frain size are considered. The results should be the same as those of Simulation S3 and, apart from negligible numerical effect, this is the case (Figures 39 and 40).

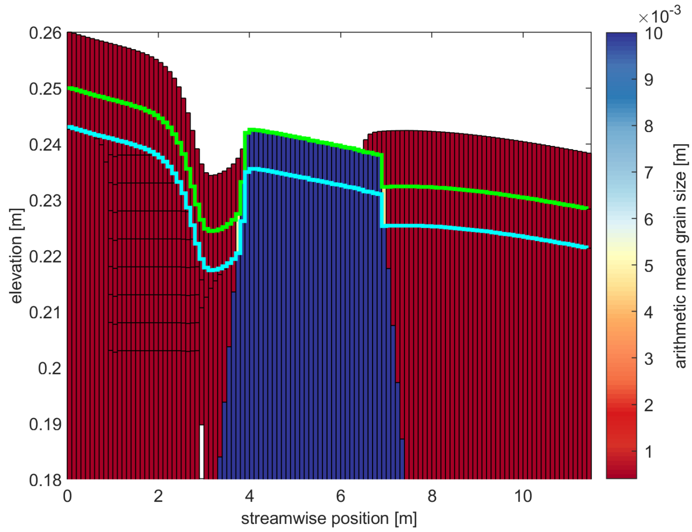

Figure 39 S7 bed level and arithmetic grain size after 1.65 h.

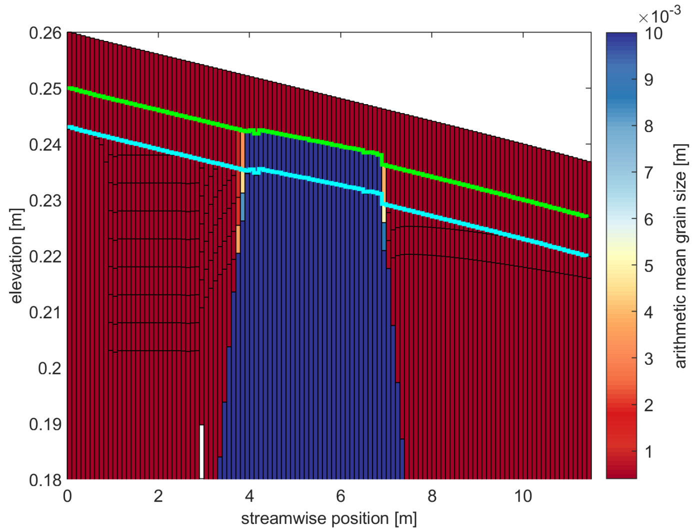

Figure 40 S7 bed level and arithmetic grain size after 7 h