Sediment Exchange

between

Flats and Channels

in tidal Inlets

Msc. thesis B.B. van Marion

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Preface

This thesis for the degree in civil engineering at Delft University of Technology concerns the morphology of tidal inlets. A tidal inlet can be divided into several parts with different properties, including barrier islands, the outer delta, the inlet channel and the flood basin. This thesis focuses on the dynamic interaction of flats and channels in the flood basin.

I would like to thank my parents and all other people who supported me during my study. My acknowledgement concerns also Delft Hydraulics and Prof. Dr. Ir. G.S. Stelling for making use of their software. Last but not least I would like to thank the examination committee consisting of Prof. Dr. Ir. H.J. de Vriend, Prof. Dr. Ir. G.S. Stelling and Dr. Ir. Z.B. Wang for their advice.

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Abstract

The subject of this thesis is the sediment exchange between channels and flats in tidal inlets. This topic is part of a research scheme which must yield more insight into the dynamic equilibrium of tidal inlets. An important aspect of the sediment exchange is the drying and flooding of the flats during the tide. The oscillating water level gives rise to velocity variations which cause stirring and settling of sediment.

In nature, tidal inlets are large and complex systems, the dimensions of which vary continuously due to natural effects (tide, storms, sea level rise). For the purpose of this study a schematised tidal inlet is considered, with dimensions as occurring in the Dutch Wadden Sea. In this schematisation, the bottom of the channel and the tidal flats is fixed. The schematisation is used as a starting point to model the water and sediment motion during a few tidal cycles. Subsequently the output of each model run, i.e. velocities, water depths and concentrations, is used to estimate the sediment transport during ebb and flood, respectively. The transport is calculated in points along a line perpendicular to the channel axis. Model runs are executed in one and two dimensions.

Sediment transport rates according to power law formula (1-D and 2-D cases) and according to a suspended sediment equation (1-D cases only) are compared. The suspended sediment fluxes at the flat agree well with the sediment transport rates determined by the power law formula.

In some of the 1-D runs, it appears that problems occur around the transition from the flat to the channel. Very large values of the transport rate, much larger than those in the adjacent cells, result from a thin water film with relatively high velocities. Therefore, the sediment transport model is provided with a water depth threshold preventing this. The threshold hardly influences the sediment transport rates during flood, either suspended or by power law, because the front of the wave is rather steep. The sediment transport rates during ebb, however, for which the threshold was meant, are reduced, such that the transport rate approaches the values in the surrounding points.

The parameters which have been chosen to vary in the 1-D model runs are:

- the slope of the tidal flat, in order to investigate whether an equilibrium profile of the tidal flat is possible, and to investigate the net transport over the landward edge of the flat if the flat is relatively steep,
- the (a)symmetry of the imposed water level in order to compare the results of a symmetric boundary condition to the results of a flood-dominant condition; the water level in the channel changes with time towards the end of the channel,
- the tidal amplitude and the length of the tidal flat, in order to assess how they are related to the sediment transport.
Abstract

Sediment exchange between flats and channels in tidal inlets

Conclusions from 1-D modelling:

- The flood sediment transport rate per tidal period decreases almost linearly over a flat. It reaches its maximum near the main channel and it goes to zero at the end of the flat. The rate of decrease depends on the tidal range and on the length of the flat. A larger tidal range results in a larger maximum flood velocity and thus in a larger sediment transport rate. A longer flat does not result in a higher maximum flood velocity, but only in a longer duration of high velocities.

- If the flat is horizontal, the sediment transport during flood is distinctly larger than during ebb. If the slope is given a value of $10^{-3}$, the flood-velocities are reduced. The sediment transport rates of flood and ebb are almost equal, but still flood-dominant. If the slope is taken $2.0 \cdot 10^{-3}$ or $3.0 \cdot 10^{-3}$, however, the highest part of the flat erodes, because it becomes ebb-dominant. The residual sediment transport is in the direction of the main channel.

- The sediment balance is used to conclude that the slope of a tidal flat approaches an equilibrium state when there is only tidal activity. The equilibrium slope is between $1.0 \cdot 10^{-3}$ and $2.0 \cdot 10^{-3}$ (see previous conclusion). If the flat has a slope less than $10^{-3}$, the rate of accretion of the higher parts of a flat is larger than that for the lower parts. That is why the slope of the flat becomes steeper.

- The tidal (a)symmetry also influences the equilibrium profile. The previous conclusion applies to a flood-dominant boundary condition (asymmetrical). If the boundary condition is symmetrical, the flat becomes less steep and ebb-dominant as far as the sediment transport is concerned. So if the $M_2$-component does not have the opportunity to distort the water level, milder slopes of the flats are found.

Beside the 1-D also 2-D model runs are executed. It appears that the 1-D model runs are overruled by the 2-D model runs. The sediment transport rate during flood is larger than during ebb. It appears from a contour plot that most area of the schematised tidal inlet accretes.

Conclusions from 2-D modelling:

- Sedimentation of the flat area nearest to the channel is larger than sedimentation of the flat area at the back of the flat. Because the flat will accrete substantially throughout this will lead to milder slope of the flats and a higher level of the flats.

- The equilibrium profile of a flat should be verified with a morphological model, because the shape of the profile does not become clear from the 2-D simulations, executed in this thesis. Suggestions for this profile is a profile with overdepth of the flat above mean sea level.
1 Introduction

1.1 General

Tidal inlets are largely unexplored areas in morphological modelling. There are several empirical formulae available for the morphological equilibrium state, but the theoretical basis of morphodynamic modelling of tidal inlets is still weak. Sediment exchange processes between channels and flats are insufficiently understood. Flats have a big influence on the tidal inlet behaviour because of the large amount of water which is stored on top of them. The amount of water stored in the embayment during high water is several times larger as during low water. As the flats are considered to be an important factor controlling the inlet behaviour, their morphological evolution is worth investigating with a model.

1.2 Description of the problem

Tidal inlets are of paramount importance to the ecosystem, recreation, navigation and industry. The dynamic interaction of channels and flats is a key element in the morphological evolution of inlet systems. The flats form a storage area where a lot of water and sediment can be stored.

Erosion and deposition of sediment are caused by natural effects, viz. wind, waves and tides. The influence of the tide on the sediment transport at the flats is investigated with a 1-D and 2-D schematisation. If the sediment transport is known as a result of the hydraulic circumstances, the initial changes of a mobile bed can be predicted.

1.3 Objectives of the study

The objectives of the study are:

- to study the water movement on the flats with a model. In this model the threshold for drying and flooding of the flat is set to zero, so that the process of drying and flooding is continuous. During ebb only a (very) thin layer of water remains on the dried flat areas.
- to investigate how the sediment transport rates in the direction perpendicular to the main channel behave over a vast flat through the tidal cycle.
- to assess the influence of changing certain parameters on the water movement and, consequently, on the sediment transport rate. These parameters are a symmetrical instead of an asymmetrical water level boundary, the slope and the length of the tidal flat and the imposed amplitude of the M2-component.
- to predict on basis of the sediment balance the erosion or sedimentation tendencies of the bottom.
1.4 Assumptions

The area of interest is the transition main channel/flat and the flat itself. Although nature has three dimensions, 2-D and especially 1-D models are used in order to keep the model simple. Therefore, a highly schematised tidal inlet is considered.

- The tidal inlet is idealised to a rectangular basin with one channel and one flat. Dimensions and bathymetry are described in Sections 3.1 and 3.2.
- The water level varies at one boundary, whereas the other boundaries are closed.
- The tidal amplitude is first chosen to be one metre. This amplitude is representative for a meso-tidal inlet.
- The properties of the water and the sediment are homogeneous throughout the model. Only in the 2-D runs the Chézy coefficients of the channel and the flat are different in order to model the roughness of shallow areas.

During the study, the following additional assumptions were made:

- The tidal asymmetry imposed in the 1-D model runs is flood-dominant, because flood-dominant tidal inlets occur more frequently than ebb-dominant ones.
- In determining the influence of the velocity on the sediment transport rates, the coefficient of proportionality, $m$, in the transport formula $(s = m \cdot v')$ is taken 1.0. In fact, this coefficient depends on local parameters like the median grain diameter, the relative density and the Chézy coefficient. In order to know the real value of the sediment transport rate, one has to determine $m$ and multiply it with $v'$.
- The sediment transport in the 2-D runs (TRISULA) is estimated halfway the main channel ($x = \frac{1}{2}L_{channel}$). In the other program (TIFLAT) it is possible to build in a subroutine, so that the transport is everywhere known.
- The threshold for reducing the ebb sediment transport rates is chosen such, that a smooth curve of sediment transport is obtained.
- The influence of the tidal amplitude is considered for meso-tidal inlets. Model runs are executed with a maximum amplitude of 1.75 m.

1.5 Outline of the report

In this report the following subjects are covered. Chapter 2 contains a general introduction to the theory underlying the studied and other closely related subjects. Chapter 3 describes the software that is used to model the channel and the flat in 1-D. The results are discussed in terms of the sediment transport rate during a tidal cycle, in various cross-sections at the flat. Section 4.1 describes the same runs executed with TRISULA, a software package of Delft Hydraulics for 2-D and 3-D current modelling. Section 4.2 describes the 2-D runs made with TRISULA. Chapter 5 describes how the suspended sediment concentration is modelled in the 1-D runs, in order to investigate the influence of space and time lags in the sediment concentration. This is done to compare the transport by a power law and the suspended sediment transport. Chapter 6 explains why and how the sediment volumes during ebb in certain cross-sections are corrected.
In chapter 7 other parameters are varied, such as tidal (a)symmetry, tidal amplitude and the length of the flat. Chapter 8 completes the two dimensional picture of the sediment transport. In this chapter the sediment transport in the channel, where lateral velocities are very small, is also analysed. Sediment exchange both in x and y-direction is known and offers the opportunity to make a contour plot of areas with erosion or sedimentation. Finally, the conclusions are drawn in the last chapter. The appendices present results, computer codes and vector plots.
2 Related subjects, theories and literature

2.1 General

The movement of sediment and water in a tidal inlet is rather complex, because there is a lot of exchange between the various sections (barrier islands, the outer delta, inlet channels and flood basin) under varying conditions. The conditions (velocity and water level) varying in time and space are caused by the tide, waves (weather conditions) and climatological effects. This thesis, focuses on the influence of the tide. To make a distinction according to the tidal range, tidal inlets are generally split up into the categories micro-, meso- and macro-tidal. One speaks of micro-tidal inlets when the tidal range is less than 2 m. Meso-tidal inlets have a tidal range between 2 and 4 m and macro-tidal inlets have a range larger than 4 m.

The tide, waves and climatological effects have different time scales. Due to these different time scales there is a variety of models like database, empirical, process-based and (semi)-empirical long-term models. There is a possibility to discern short-, medium- and long-term models [de Vriend 1996]. Short-term models describe processes like waves, currents and sediment transport, but only simulate initial morphological changes. Medium-term models have a larger time scale and describe the morphodynamic interaction. Long-term models try to describe slowly changing trends or long-term stability of a tidal inlet. This thesis concerns the first category of models, because the influence of the tide on the sediment transport is studied during one tidal cycle.

Sediment transport can occur in various modes. The extreme modes are purely bed load and purely suspended load. In reality suspended load is always accompanied by a certain amount of bed load. Bed load means that grains are jumping, rolling, etc. on the bottom. Suspended load means that grains are transported in suspension, with the grains distributed over the vertical. The general equation for the suspended sediment concentration is:

$$\frac{\partial c}{\partial t} + u \cdot (\nabla c) - w_s \mathbf{e}_z \cdot (\nabla c) = \nabla \cdot (K \nabla c)$$  \hspace{1cm} (2.1)

in which:
- $c$ : concentration
- $t$ : time
- $u$ : water particle velocity vector
- $w_s$ : particle settling velocity
- $\mathbf{e}_z$ : vector of unity in the vertical direction
- $K$ : diffusion coefficient

If one moves in the direction of the current, the concentration exhibits a retarded adjustment to its equilibrium value. Characteristic length and time scales of the adjustment process are:
\[ L_c = O\left( \frac{u a}{w_s} \right) \quad \text{and} \quad T_c = O\left( \frac{a}{w_s} \right) \]

in which:  \( u \): horizontal water particle velocity
\( a \): water depth

In this paragraph values of these parameters will be derived for an idealised flood-basin (dimensions see figure 3.1). Under the condition that the Stokes regime holds (\( Re_D < 1 \)), one may use for the fall velocity [Battjes, 1990]:

\[ w_s = \frac{1}{18} \frac{\Delta g D^2}{\nu} \quad \text{Re}_D = \frac{w_s D}{\nu} \]

in which:  \( \Delta \): relative density bed material
\( g \): acceleration due to gravity
\( D \): particle diameter
\( \nu \): kinematic viscosity
\( \text{Re}_D \): Reynolds number

For \( D = 200 \mu m \), \( w_s = 0.036 \text{ m/s} \) and \( \text{Re}_D = 7.2 \). This is outside the Stokes regime. Hence the fall velocity reduces to about 0.03 m/s. In the channel \( L_c \approx 1.0/10.0/0.03 \approx 300 \text{ m} \) and \( T_c = 10.0/0.03 = 333 \text{ s} \approx 5 \text{ min} \). At the flats \( L_c \) and \( T_c \) reduce rapidly because the depth decreases.

During flood, water with a certain sediment concentration moves from the channels onto the flats. On the flats the suspended sediment settles due to decreasing velocity. This is enhanced during slack water. During ebb, as the water moves back from the flats into the channels, the sediment concentration takes time to pick up and therefore lies below the concentration of the corresponding flood velocity. Eventually less sediment will be transported back into the channel with the ebb current (unless the ebb current is much stronger than the flood current). This is a mechanism of accretion of the flats by the tide. In situations with high wave activity erosion occurs, because the waves stir up the sediment by exerting a high bottom shear stress. The concentration in the vertical increases and the ebb current will carry the increased quantity of suspended sediment into the channel [de Vriend et al., 1989].

### 2.2 Tidal asymmetry: the hypsometry effect

This section is based on theory of Friedrichs [Friedrichs et al., 1992]. The tidal response of an inlet is asymmetrical due to several interactions. These asymmetries are characterised by unequal duration of the ebb and flood and by their unequal amplitudes. Distortion of the tide can be represented by the non-linear growth of harmonics of the astronomical tidal constituents. The type of distortion is determined by the relative phase difference \( \Phi, \Phi = 2\Theta_f - \Theta_b \), between the semi-diurnal component \( M_2 \) and his first harmonic \( M_4 \). \( \Phi = -90^\circ \pm 30^\circ \) corresponds to flood-dominance and \( \Phi = 90^\circ \pm 30^\circ \) to ebb-dominance, if the tidal components are written as:
\[ \zeta(t)_{M2n} = A_i \cos(\omega_i t + \Theta_i), \quad \text{for} \ i \in \{1,2\} \]

in which:  
- \( \zeta(t) \): water level  
- \( A_i \): amplitude of \( i \)th component  
- \( \omega_i \): tidal frequency of \( i \)th component  
- \( \Theta_i \): phase of \( i \)th component

If a tidal inlet is flood-dominant, this means that the duration of the ebb phase is longer than that of the flood phase. This results in higher velocities during flood. For ebb-dominant systems the opposite is true.

The 1-D momentum equation for the channels reads:

\[ \frac{\partial Q}{\partial t} + \frac{\partial}{\partial x} \left( \frac{Q^2}{A_c} \right) + gA_c \frac{\partial \zeta}{\partial x} + c_d \frac{Q}{A_c} \left| \frac{Q}{P_c} \right| = 0 \]

and the corresponding continuity equation is:

\[ \frac{\partial A}{\partial t} + \frac{\partial Q}{\partial x} = 0 \]

in which:  
- \( Q \): Discharge  
- \( x \): distance along horizontal axis  
- \( A_c \): area of main channel  
- \( \zeta \): water level  
- \( c_d \): drag coefficient  
- \( P_c \): wetted perimeter of main channel  
- \( A \): total area

Non-linear effects are produced by the friction term and the advection term in the momentum equation (2.5) and also by the geometrical profile of the channel in the continuity equation (2.6). In the latter case the tidal flat area plays an important role, because in the tidal system it functions largely as a storage area.

One can distinguish two types of channels. Channels with a relatively small flat area (rectangular channel or a channel with a trapezoidal profile) and channels with a large flat area (a great time variability of the cross-section or large flat area in the back of the basin). The first show flood-dominant behaviour. This means that sediment is imported and that the tidal flat area increases. Field data [Pethick, 1980] and models [van Dongeren and de Vriend, 1993] show that increasing the tidal flat area turns the asymmetry from flood-dominant into ebb-dominant. These experiments and measurements suggests an equilibrium flat area.
Field observations and model results [Friedrichs et al., 1992] show that if the ratio of the tidal amplitude and the water depth $A_y/a > 0.3$ the inlets are characterised by longer falling than rising tides, so higher flood currents, unless vast tidal flat area is absent. Deeper inlets with $A_y/a \sim 0.1 - 0.2$ will show longer rising than falling tides. It is of great importance to know whether a channel is flood- or ebb-dominant, because this determines the direction of the net sediment transport ($s \sim u''$).

There are two ways to consider the non-linear behaviour of tidal inlets. To explain the unequal duration of ebb and flood, Lamb [1932] used Airy's [1845] wave theory to derive an expression for the phase velocity of the tidal wave:

$$c_w = \sqrt{ga} \left( 1 + \frac{3}{2} \frac{\zeta(t)}{a} + O \left( \frac{\zeta_0}{a} \right)^2 \right)$$

in which:
- $c_w$: wave celerity
- $\zeta_0$: water level at the boundary of the inlet

Friedrichs approaches the problem in a different way. According to Friedrichs et al. [1992] the friction term in the momentum equation dominates the acceleration terms. Frictional effects are according to him the primary cause of the non-linear behaviour. Moreover, the acceleration terms often change sign during the tidal cycle, whence their net contribution will be small. Only at slack water this is not true, because then the current velocities are almost zero, and so is the friction term ($O(u^2)$). These assumptions are justified because we consider a tidal inlet with an $A_y/a$-ratio $> 0.1$ and velocities of the order of 0.5 m/s.

Because the friction term in equation (2.5) is larger than the acceleration terms, Friedrichs omits the latter. The momentum equation then reduces to:

$$\frac{\partial \zeta}{\partial t} + n_M^2 R^{\frac{1}{3}} |u| u = 0$$

Equation 2.8

Together with the continuity equation, the model can be reduced to the following partial differential equation:

$$\frac{\partial \zeta}{\partial t} = \kappa \frac{\partial^2 \zeta}{\partial x^2} \quad \text{with} \quad \kappa = \frac{b_c a^{\frac{2}{3}}}{b n_M} \left| \frac{\partial \zeta}{\partial x} \right|^{-\frac{1}{2}}$$

Equation 2.9

in which:
- $b_c$: width of main channel
- $b$: total width of tidal inlet
- $n_M$: friction coefficient of Manning

There cannot exist a solution at slack water for this problem, because then $\kappa$ is not defined. With the boundary conditions $u(L_{ch},t) = 0$ and $\zeta(0,t) = \cos(\omega t)$ a solution with a lot of terms is found, which turns to be a partially progressive and partially standing wave with exponential decay. The decay and also the propagation speed are determined by $L_f$, the friction length, which appears in the arguments of the solution together with $\psi$. 
\[ L_f = \sqrt{\frac{2\kappa}{\omega}} \wedge \psi = \frac{2L_{ch}}{L_f} \]  

in which:  
\( \kappa \) : see above  
\( \psi \) : phase  
\( L_{ch} \) : length of tidal channel

If \( L_{ch}/L_f \ll 1 \) a standing wave occurs. For \( L_{ch}/L_f \gg 1 \) a diffusive propagating wave occurs. For intermediate values of \( \psi \), systems with a length of circa 20 km, Friedrichs finds for \( L_f \):

\[ L_f \approx 2^{1/6} \pi^{1/3} \left( \frac{b}{b_0} \right)^{2/3} \left( \frac{a}{a_0} \right)^{10/9} \left( \frac{\xi}{\zeta_0} \right)^{1/3} \left( \frac{n_m \omega}{\omega_0} \right)^{2/3} \]

The total width, depth and tidal amplitude at the mouth of the channel and the friction coefficient influence the amplitude decay and the wave speed. The values of \( b \) and \( a \) vary constantly during a tidal cycle, so that according (2.11) \( L_f \) also varies and consequently the wave speed does. The values of the variables at high water and low water. A dimensionless parameter, the ratio between \( L_{f\text{ high}} \) and \( L_{f\text{ low}} \), i.e. the friction length for high and low water, respectively, helps concluding whether a basin is flood- or ebb-dominant. It is:

\[ \Lambda = \frac{L_{f\text{ high}}}{L_{f\text{ low}}} = \left( \frac{b_{\text{ high}}}{b_{\text{ low}}} \right)^{2/3} \left( \frac{a_{\text{ high}}}{a_{\text{ low}}} \right)^{10/9} \]

If \( \Lambda < 1 \) the tidal area of concern has a deep main channel and/or vast tidal flat area and it shows ebb-dominant behaviour and for \( \Lambda > 1 \) the channel has little geometry variability during the tidal cycle and it shows flood-dominant behaviour.

If we consider a real-life tidal inlet, it consists of a complicated tree-type network of channels and flats. Small channels dry almost completely during low tide. These channels show flood-dominant behaviour [Lincoln 1988, Speer 1991].

Thus, there are two effects of dominancy which one has to consider and which lead to 2-D modelling of tidal inlets. On the one hand there is the behaviour in the main channel and on the other hand the behaviour in the flat area. The latter is obvious when the forced boundary in the main channel has a shorter flood-duration. The duration of the flood on the flat is also shorter. But when the forced boundary has a shorter ebb-duration, there are points near the boundary of the flat with ebb-dominance and there are points further away from the boundary showing flood-dominancy. So somewhere on the flat there is a transition of dominancy.
2.3 Uniform bottom shear stress and equilibrium hypsometry of a flat

This section is based on the work by Friedrichs et al. 1994. There are some empirical formulae which give a relationship between the flat area and other properties of a tidal inlet. These relationships are all approximative, because the flat area changes all the time due to the wind/wave climate. Thus, these relationships actually refer to the dynamic equilibrium state of the flat area. If the slope of a flat is constant, the bottom shear stress is non-uniform, so morphological changes will occur.

The distribution of horizontal area of a landmass with respect to elevation [Strahler 1952] is a way to illustrate a profile or geometry of a flat. The geometry of a flat with a straight coastline can look like the profiles in figure 2.1. The dashed line imagines a linear profile. The other profiles are convex or concave. From observations along the German Bight it is concluded that the type of hypsometry (convexity or concavity) is determined by tidal range, exposure of the shore to wind waves and patterns of accretion or erosion [Dieckmann et al., 1987]. Macro-tidal tidal inlets have a convex hypsometry and low meso-tidal inlets a concave one. One can also say that the ratio between tidal current activity and wave activity is responsible for the hypsometry. A high ratio implies a convex hypsometry, a low ratio a concave one.

![Figure 2.1: Left figure shows convexity; right figure concavity.](image)

When the weather is calm, the tidal component is dominant and the tidal velocity can be derived from the continuity equation. If we assume the flat to be horizontal, it acts as a storage reservoir. In a slice of constant width, the tidal velocity can be found from the pumping mode approximation:

\[
\int_0^{L_y} \frac{\partial a}{\partial t} \, dy + \mathbf{v} \cdot \mathbf{a} = 0
\]

in which:
- \(L_y\): length of the flat
- \(y\): distance along horizontal axis
- \(v\): velocity at the edge of the flat

\[
da / dt = 2\pi / T = 2\pi / 43200 \approx 1.5 \times 10^{-4} \Rightarrow v = 1.5 \times 10^{-4} \times 106.25 / 1 \approx 0.4 \text{ m/s.}
\]

Values of 0.6 m/s were found from the more extensive models described later in this thesis.
Friedrichs et al. [1994] first considered a straight shoreline, preceded by a flat with a constant slope. Until halfway the flat the maximum flood velocity $V_T$ is found to be uniform and closer to the shore $V_T$ decreases to zero. The outer half of the flat will therefore be morphologically stable and at the inner half deposition occurs. However, during ebb the same velocities occur, but in reversed direction. During slack water the suspended sediment will settle, so that the sediment is transported back into the channel. If there is no erosion, e.g. due to waves, the tidal action will lead to a convex profile of the flat.

Friedrichs et al. attempt to find the equilibrium flat profile for a straight shoreline, such that $V_T$ is the same throughout the flat. To that end, the flat has to have a convex profile (see figure 2.1).

The maximum orbital velocity in shallow water waves is given by:

$$V_w = \frac{H}{2a} \sqrt{ga}$$  \hspace{1cm} 2.14

in which: $H$ : wave height  
$a$ : water depth  
g : acceleration due to gravity

An indication of the magnitude of $V_w$ at the boundary of the flat follows from $H_w = 0.5$ m; $a = 1.0$ m. $V_w \approx 0.8$ m/s. Waves can have a significant effect on the bottom shear stress ($\tau \sim v^3$). This must certainly have its influences on morphology. Due to seasonal weather conditions the tidal flat will therefore never be at rest (static equilibrium).
3 Schematisation of a tidal inlet

3.1 Introduction

A highly schematised tidal inlet, with dimensions which are characteristic of the inlets in the Dutch Wadden Sea, is used in order to get insight into the drying and flooding of flats and to know the velocity in several cross-sections of the schematised inlet, so that the sediment transport can be calculated. The schematisation of the channel and the tidal flat is shown in figure 3.1. Adjacent of the main channel, with a depth of 10 m, is a one kilometre flat area without a slope or with a small slope in y-direction, which dries during low water.

First 1-D runs are executed with the program TIFLAT to get a better understanding of the water movement and how it will behave at the flats. Subsequently the runs in TIFLAT will be compared with the same 1-D runs in TRISULA, a software package of Delft Hydraulics. After that some 2-D modelling in TRISULA is done. In the 1-D and 2-D runs, attention is focused on the y-direction. The variation of the water level in the main channel will cause a wave in y-direction. In order to describe the sediment transport in the cross-sections, the water level and velocity distributions over the flat have to be known.
3.2 1-D modelling with TIFLAT

3.2.1 Introduction

TIFLAT, a program written by Prof. Stelling [1990] is able to describe bores and other long hydraulic waves phenomena. This program was used to model the tidal motion in y-direction over the flat, assuming uniformity in x-direction. In the sections below the constituting equations and the input of the program are described.

3.2.2 Constituting equations

The continuity and momentum equations solved in TIFLAT read:

\[
\frac{\partial \mathbf{v}}{\partial t} + \mathbf{v} \cdot \nabla \mathbf{v} + g \frac{\partial \zeta}{\partial y} + g \frac{\mathbf{v} \cdot \mathbf{v}}{C^2 R} = 0
\]

\[3.1\]

\[
\frac{\partial \zeta}{\partial t} + \frac{\partial va}{\partial y} = 0
\]

\[3.2\]

\[a = h + \zeta\]

\[3.3\]

This set of equations is discretised on a staggered grid, to yield:

\[
\frac{v_{i+\frac{1}{2}, j+\frac{1}{2}}^{t+1} - v_{i-\frac{1}{2}, j+\frac{1}{2}}^{t}}{\Delta t} + \frac{v_{i+\frac{1}{2}, j-\frac{1}{2}}^{t+1} - v_{i-\frac{1}{2}, j-\frac{1}{2}}^{t}}{\Delta y} + g \frac{\zeta_{i+1}^{t} - \zeta_{i}^{t}}{\Delta y} + g \frac{v_{i+\frac{1}{2}, j+\frac{1}{2}}^{t+1}}{C^2 R} = 0
\]

\[3.4\]

\[
\frac{\zeta_{i}^{t+1} - \zeta_{i}^{t}}{\Delta t} + \frac{a_{i+\frac{1}{2}, j+\frac{1}{2}}^{t+1} - a_{i-\frac{1}{2}, j+\frac{1}{2}}^{t+1}}{\Delta y} = 0
\]

\[3.5\]

in which:

- \(v\) : horizontal water particle velocity
- \(t\) : time
- \(y\) : distance along horizontal axis
- \(g\) : acceleration due to gravity
- \(\zeta\) : water level
- \(C\) : Chézy coefficient
- \(R\) : hydraulic radius
- \(i\) : grid index
- \(l\) : time index
- \(a\) : water depth
- \(h\) : mean water depth
3.2.3 The input file

The input file is the tool for the user to define his model. An example of an input file is printed in Appendix I and will be explained in the next paragraph. The input file consists of a number of rows with the following elements:

i) The number of nodal points in x- and y-direction.

ii) The step size in x- and y-direction in metres.

iii) A matrix with the bathymetry (in metres) in the nodal points of the computational grid. A small program written in FORTRAN and inserted in Appendix II describes how the desired bathymetry is achieved. Figure 3.2 shows the bathymetry of a channel and a flat, partly (from $y_{26}$ to $y_{67}$) without a slope.

![Figure 3.2: The bathymetry of the bottom versus the nodal points.](image)

iv) The initial water level in metres. This value should preferably be chosen at high water, because then it makes the initial water level correspond with the boundary condition, so that the water level does not need much adaptation time in the beginning of the computation.

v) A matrix (2 rows) of which each row describes the kind of boundary to model the wave. The last eight columns correspond respectively to a reference level and an amplitude in metres, a period in seconds and a phase difference in degrees of the $M_2$-component, a damping coefficient and three parameters of the $M_4$-component close the row.

vi) Three values which are: Time step in seconds, from which time step calculated information will be recorded and how many time steps the program will execute.

vii) The value for the Chézy coefficient in m$^{1/2}$/s.

viii) The points in which water levels and velocities will be recorded.

The values of some parameters are explained in the next paragraphs:

ad i) and vii) $\Delta y$ and $\Delta t$ are chosen such that the model is stable. First the space step $\Delta y$ is chosen and after that, the time step $\Delta t$ is determined from the stability condition. The rather steep slope of the channel makes a small time step necessary. A less steep slope is constructed at the transition channel/flat, $y_0$ till $y_{16}$, so that the time step can be enlarged. Finally $\Delta y=25$ m and $\Delta t=25$ s. The geometry and the space step force to use 66 nodal points in which water level and velocity will be calculated. Eventually the amount of nodal points is adjusted to 106 because of resonance.
ad vii): There is one boundary condition necessary at each side of the model. The water level is imposed at the channel side and the velocity at the landward edge of the flat. The velocity there is set equal to zero. There is a possibility to model also the $M_2$-component of the tide.

ad viii): For simplicity, and because of vertical mixing the density of the water is everywhere equal to 1000 kg/m$^3$. The Chézy factor is chosen to be constant along the $y$-axis. To give an approximation of the Chézy factor the formula of White-Colebrook is used. The mean water depth in the model is chosen to be one metre and a bed roughness $k$ of 0.05.

\[
C = 18 \log \left( \frac{12 \alpha}{k} \right) = 18 \log \left( \frac{12 \cdot 1}{0.05} \right) \approx 45 \text{ m}^2/\text{s}
\]

in which: $\alpha$ mean water depth $k$ bed roughness

3.3 Description of the runs

For each run with TIFLAT, an extension to the name of the input file has to be given. This extension consists of three characters, $xyz$. In the administration system used in this thesis, the first character, $x$, describes the slope of the flat. The second, $y$, describes the imposed (a)symmetry. The last one, $z$, describes the tidal amplitude.

1. Run 0yz corresponds to a slope of the flat 0.0. So the tide will behave like a bore. Run 1yz is a model with a slope of 0.5 $\times$ 10$^{-3}$. Run 2-, 3- and 4yz have a slope of respectively 1.0 $\times$ 10$^{-3}$, 2.0 $\times$ 10$^{-3}$ and 3.0 $\times$ 10$^{-3}$.

2. The water level boundary can be written as:

\[
\zeta(t) = A_{M2} \cos(\omega_{M2} \cdot t + \Theta_{M2}) + A_{M4} \cos(\omega_{M4} \cdot t + \Theta_{M4})
\]

in which: $A_i$ amplitude of $i^{th}$ component $\omega_i$ tidal frequency of $i^{th}$ component $\Theta_i$ phase of $i^{th}$ component

Asymmetry can be realised by attributing values to the $M_2$-components of the water level. Normally the amplitude of the $M_2$-component is much smaller than the $M_2$-amplitude.

Run $x0z$ has no imposed asymmetry. Only the $M_2$-component is imposed. $A_{M2} = 1.0$ m. Run $x1z$ has a shorter ebb duration, so $\Theta_{M4} = -90^\circ$. Run $x2z$ on the contrary has a shorter flood duration. The amplitude of the $M_2$-component has a value of 0.1 m. Books for coastal engineering [e.g. Pugh 1987] report of $M_2$-components with an amplitude of 0.05-0.15 m. In order to create a shorter flood, the value for the phase difference is chosen:

\[
\Phi = 2 \Theta_{M2} - \Theta_{M4} = -90^\circ; \quad \Theta_{M2} = 0^\circ \quad \Rightarrow \quad \Theta_{M4} = 90^\circ.
\]

Substituting all the magnitudes in the arguments of the cosine term specifies $\omega_i$: 
\( \omega_{M_2} = 2\pi/T_{M_2} \). In which \( T_{M_2} = 43200 \text{ s} \) is the semi-diurnal tidal period.

\[ T_{M_2} = \frac{\sqrt{2}}{2} T_{M_2}. \]

3. Run \( xyI \) has a \( A = 1.0 \text{ m} \). This value is attributed of course to the \( M_2 \)-component. Run \( xy2, 3 \) and \( 4 \) have respectively a tidal amplitude of \( 1.25, 1.5 \) and \( 1.75 \text{ m} \).

### 3.4 The results

In this section the water levels and velocities in some cross-sections are plotted in diagrams for several runs. These runs have been adjusted, because earlier ones showed resonance. This phenomenon is most obvious for a non-sloping flat (run \( 0y2 \)). The resonance becomes weaker, if the slope of the flat increases. This seems logical, because the water is slowed down by the slope and the changed bathymetry of the flat reduces the propagation speed of the tidal wave in y-direction. Yet, resonance occurred for runs with a slope unequal to zero. The following computation shows that resonance can occur in this situation. It concerns run \( I02 \), a run with \( A = 1.5 \text{ m} \) and a plot of the velocity during the tidal period is in Appendix IV.

From the figure it is observed that the tidal period and the resonance period correspond with:

\[ T \approx 13.4 \text{ cm}; \quad 3 T_{res} \approx 1.4 \text{ cm} \quad \Rightarrow \quad T_{res} = 43200 \cdot 1.4/(3 \cdot 13.4) = 1504 \text{ s}. \]

\[ c_{mean} = (gh_{mean})^{\frac{1}{2}} = (9.81 \cdot 1.2)^{\frac{1}{2}} \approx 3.43 \text{ m/s} \quad \Rightarrow \quad L_{res} = c \cdot T = 5160 \text{ m}. \]

Resonance occurs in a basin with a closed boundary (\( u(L_I, I) = 0 \)) if the length of the basin is an uneven multiple of a quarter of the wave length. Although the above method to show resonance is rather crude, the ratio between \( L_{fl} \) and \( L_{res} \), (0.29) suggests that a near-resonant condition occurs.

In order to reduce resonance, another set of runs has been made with a different geometry. Some adjustments have been made to the bathymetry of the input, so that an extension between \( y_{07} \) and \( y_{106} \) (see figure 3.2), is added to move the reflections farther away from the area of interest and to reduce resonance. From the point of view of continuity, one can say that the duration of the highest velocities will become longer, since more water must be transported to fill the increased storage capacity of the flat. The amplitude of the flow velocity will remain the same, because the wave propagation speed in the parts close to the channel will not alter. Hence, this also has consequences for the sediment transport.

The adjustments are effective. The plots of the velocity depicted in the figures at the next pages are smoother, but there are still some very small fluctuations in the \( 02I \) run. In the other runs the resonance has disappeared. In the next paragraph the results of the flood-dominant runs are discussed.
Run 021: Flat with a horizontal slope; flood-dominancy; $A = 1.0$ m (figure 3.3). There is no slope present. Therefore the tide behaves like a bore. The tidal front rushes over the dry flat. Consequently, the maximum flood-velocity, $V_{T,f}$ in cross-sections 2-5 is larger than the maximum ebb-velocity, $V_{T, ebb}$. This is why the duration of the flood is shorter than that of the ebb. This phenomenon is enhanced because the forcing water elevation boundary is fastly rising and slowly falling.

\begin{figure}[h]
\centering
\includegraphics[width=0.8\textwidth]{water_level.png}
\caption{Water level and velocity in five cross-sections of Run 021: a horizontal flat; flood-dominancy; $A = 1.0$ m; water level and velocity with TIFLAT.}
\end{figure}
Run 121: Flat with a slope of $0.5 \cdot 10^{-3}$, flood-dominancy, $A = 1.0$ m (see figure 3.4). The slope does not influence the asymmetry, imposed at the boundary, because $V_{T_{fl}}$ in cross-sections 2-5 is larger than $V_{T_{eb}}$.

**figure 3.4:** Velocity in five cross-sections of run 121; flat with a slope of $0.5 \cdot 10^{-3}$; flood-dominancy; $A = 1.0$m; water level shows similarities with the water level of run 021.
Run 221: Flat with a slope of $1.0 \times 10^{-3}$; flood-dominancy; $A = 1.0$ m (see figure 3.5). The slope of the flat is larger in comparison with run 121. Despite the imposed asymmetry at the boundary $V_{T,fb}$ is smaller than $V_{T,ebh}$ in the cross-sections. The slope of the flat does influence $V_{T,fb}$.

![Image](image_url)  
*Figure 3.5: Velocity in five cross-sections of run 221; flat with a slope of $1.0 \times 10^{-3}$; flood-dominancy; $A = 1.0$ m; water level shows similarities with the water level at the previous page.*

### 3.5 Sediment Transport

The results of the previous chapter will be used to determine the sediment transport rates during ebb and flood. The sediment transport can be written as:

$$s = f(v(t)) = m \cdot v(t)^n$$  \hspace{1cm} 3.8

in which:  
- $s$ : sediment transport rate per unit width  
- $m$ : proportion coefficient  
- $n$ : power coefficient

Velocities and water depths are gathered in several output files and during run time in several arrays. The velocity is calculated in time steps of 25 seconds. The flood transport rate can be calculated by integrating (3.8) over the flood part of a tidal period.

$$V_{fb} = \int_{t_{f0}}^{t_{f\text{ end}}} (m \cdot v(t)^n) \, dt$$  \hspace{1cm} 3.9

in which:  
- $V_{fb}$ : sediment transport rate per unit width per flood part of the tide  
- $t_{f0}$ : time the flood starts  
- $t_{f\text{ end}}$ : time the flood ends
The same goes for the sediment transport rate during ebb. The coefficient $m$ is hard to predict, because it depends on several local quantities, such as the grain diameter, the relative density and the Chezy coefficient. For Engelund & Hansen $n$ equals 5 and $m$ can be determined as is done below. Transport formulae are often expressed in terms of the dimensionless parameters $X$ and $Y$:

$$X = f(Y) \quad \text{with} \quad X = \frac{s}{D_{50}^{3/2} \sqrt{g \Delta}} \quad \text{and} \quad Y = \frac{\Delta D_{50} C^2}{\mu v^2}$$ \quad (3.10)

in which: $D_{50}$: the grain diameter that is exceeded by 50% of the grains
$\Delta$: relative density bed material
$\mu$: the ripple factor: $\mu = \left( \frac{C^2}{g} \right)^{3/5}$

E&H: $X = 0.084 \cdot Y^{\frac{3}{2}}$ \quad $\Leftrightarrow$ \quad $s = 0.084 \frac{v^5}{g^{\frac{3}{2}} \Delta^2 D_{50} C^2}$ \quad (3.11)

$s = mv^n = 5.4 \cdot 10^{-4} \cdot v^5$ \quad (3.12)

in which: $m \sim g^{-n} \Delta^{-2} D^{-1} C^{-3}$

$g = 9.81 \text{ m/s}^2$
$\Delta = 1.65$
$D_{50} = 200 \mu \text{m}$
$C = 45 \text{ m/s}$

Because $m$ is a coefficient of proportionality which depends on the local circumstances, its value in this analysis is for simplicity reasons 1.0. The value of coefficient $n$ is: $3 \leq n \leq 5$. In steps of 0.5 there has been tried to find the influence of the power $n$ in (3.13) on the sediment transport. The above considerations are translated into FORTRAN code in the subroutine in Appendix III. The sediment transport rate is determined in every nodal point. The sediment transport rates (output) for the various runs are depicted in the next three figures. The diagram plots the nodal points (distance in $y$-direction) versus the sediment transport rate of ebb and flood for a constant value 5 of the exponent $n$ (E&H). Smaller values of $n$ result in similar plots, but will give larger sediment transport rates, because $|V_x| < 1.0 \text{ m/s}$. Some remarks follow below.

First the legend of the diagram is explained. The black line is the sediment transport rate determined according the power five in TIFLAT. The red line is of the suspended sediment transport rate in TIFLAT (see chapter 5). Dots represent the sediment transport rates which result from post-processing the velocity in some nodal points from TRISULA 1-D and 2-D runs (see chapter 4).
Figure 3.6: Sediment transport rates of Run 021; a horizontal flat; flood-dominancy; $A = 1.0m$.

Figure 3.7: Sediment transport rates of run 121; flat with a slope of $0.5 \cdot 10^3$; flood-dominancy; $A = 1.0m$. 
The transport rate is modelled by a power law. Between the 23rd and 27th cell the sediment transport rates for the ebb part of the tide are very large, because during a relatively long time a water film with a certain velocity (see figure 3.9 - results from run 021 at 23rd till 27th cross-section) occurs. It is the location where flat just has turned into channel. Perhaps the flow is critical? Therefore the Froude number is determined. Where \( v \) is largest, \( Fr = \frac{v}{\sqrt{g \cdot d}} \approx 0.74/(9.81 \cdot 0.05)^{1/2} = 1.06 \) and where \( d \) becomes very small \( Fr \approx 0.35/(9.81 \cdot 0.011)^{1/2} = 1.06 \).

The flow is critical during a large part of the ebb period. Does it make sense to use the power law during the whole tidal period? In case it does not, a threshold could prevent transport in very thin layers.
figure 3.9: Water depth and velocity of run 021 to show the water film; cr. 1 to 5 correspond with $y_{23}$ to $y_{27}$.

Via the sediment balance for one tidal period:

$$\frac{\partial z}{\partial \tau} + \frac{\partial S}{\partial y} = 0$$  \hspace{1cm} (3.14)

in which:

- $z$: bottom level
- $\tau$: time co-ordinate for variations at time scales much larger than the tidal period
- $S$: tidally averaged sediment transport rate per unit width
One can tell whether a profile is in equilibrium or not. If the partial derivative of the sediment transport \( \frac{\partial S}{\partial y} \) is positive/negative, the bottom will tend to erode/accrete. A positive derivative corresponds to erosion of a cell and negative to sedimentation. The derivative of the sediment transport in a cell is determined by discretising (3.14):

\[
\frac{\partial z}{\partial \tau} + \frac{S_{j+1} - S_{j-1}}{2\Delta y} = 0
\]

in which: \( j \) : grid index

From the figures at pages 22 and 22 one can presume that for every exponent \( n \) the curve of the derivatives over the flat is the same. The partial derivatives \( \frac{\partial S}{\partial y} \) of the sediment transport rates for flood and ebb, respectively, are similar for the runs 021, 121 and 221. Thus, the slope does not influence the shape of the derivative curve, but only its magnitude. Suppose that the proportion coefficient \( m \) is unknown, then only the sign of the derivative matters.

In figure 3.10 \( \frac{\partial z}{\partial \tau} \) between the cells 21 and 27 is negative during flood and at the flat \( \frac{\partial z}{\partial \tau} \) is just larger than zero. Sediment is deposited at the flats. During ebb the opposite happens at the flat. The flat seems to erode a little bit. The slope of the channel seems too steep, because sediment is transported from the transition flat/channel (\( \frac{\partial z}{\partial \tau} \) is very large ⇒ erosion) further into the channel, where the velocity decreases because of larger depth and the sediment finally settles.

Adding the derivatives results in the net sediment balance for the whole tidal period (see figure 3.11). The flats accrete and the slope of the main channel becomes less steep. The rate of accretion of the higher parts of the flat is larger than for the parts near the main channel, with the consequence that the flat tends to steepen. Clearly this cannot go on forever, but is the tide without the help of the waves capable of stopping this trend? To answer this question, some extra model runs are discussed in the next section.

![gradient of flood volume](image1)

![gradient of ebb volume](image2)

*figure 3.10: (l) the derivative during flood; (r) ebb; a negative gradient means erosion; positive corresponds to sedimentation.*

Adding these pictures results in the sediment balance for a tidal period.
3.6 Runs with steep slopes of the flat

As mentioned in the previous section model, runs with a steeper slope are executed. At the two next pages, the sediment transport rate and its partial derivative are depicted with slopes of $2.0\times10^3$ and $3.0\times10^3$, respectively. The sediment transport rates become very small, because the velocities are very small. Of course less nodal points are involved in the calculation because of the steeper slopes. For a slope of $3.0\times10^3$ is decided to calculate even with a tidal amplitude of 1.5 m.

From the figures at the next pages it appears that the higher cross-sections of the flat become ebb-dominant from a sediment transport point of view. The sediment balance (figure 3.13 and figure 3.15) for the net sediment transport shows that the higher parts of the flat erode, whereas the lower parts accrete. Thus the relatively steep slope of the flat decreases.

Together with the mechanism of the previous section, this yields a net rise of the tidal flats under the influence of the tide only. First the slope of the flats becomes steeper, because the rate of accretion of sediment out of the main channel for the higher part of the flats is larger than for the lower parts. Subsequently the slope reaches an equilibrium condition: the profile cannot become steeper than a slope between $1.0\times10^3$ and $2.0\times10^3$, because then erosion occurs.
Flood and ebb sediment transport rate

Figure 3.12: Sediment transport rate for a flat with a slope of $2.0 \times 10^{-3}$ and $A = 1.0 \text{ m}$.

Transport gradient

Figure 3.13: Net gradient of the above sediment transport rate.
Figure 3.14: Sediment transport rate for a flat with a slope of $3.0 \cdot 10^{-3}$ and $A = 1.5$ m.

Figure 3.15: Net gradient of the above sediment transport rate.
4 Modelling of a tidal inlet with Trisula

4.1 Trisula 1-D

The same model runs of chapter 3 have been executed with the software package TRISULA, developed by Delft Hydraulics. This package offers the user various menus to choose from. The various menus are ordered in a tree (see figure 4.4 at page 32), which one can use to define his problem and to attribute values to the model parameters. The values at the end of each branch between brackets are input and correspond to run 001, a run in which the Mn-component is omitted.

As mentioned before, comparable runs have been made with TIFLAT and TRISULA. The TRISULA-output in four cross-sections was post-processed and edited into one history file, so that water level, velocity and sediment transport plots of the cross-sections could be made. Water level and velocity plots are shown in the figures at the next pages. The velocity results are not quite as smooth as in TIFLAT. All the time when the flat is dry there are small irregularities. This is caused by the flooding/drying procedure. TRISULA sets the flat dry by building low dams, which prevent the remaining water in the cells of the flat from returning to the channel. When the flood returns these dams are lowered and the water that has stayed backward behind the dam flows in the direction of the main channel. This is not natural and results in negative velocities. A consequence is that \( V_T p \) from both models differ. The returning water from behind the dams might also explain this phenomenon. The drying of the flat looks quite the same in both models. The ebb-amplitudes have approximately the same value, because the water can undisturbedly flow off the flat. But when the slope of the flat increases \( V_T p \) seem to be more equal with \( V_T p \) in TIFLAT.

These differences must have their consequences on the sediment transport. The calculated transport rates, based on data from 1-D TRISULA runs and calculated in the same way as in the subroutine of TIFLAT (see Appendix III), are indicated in the figures at pages 22 and 22 as single points, not connected by a curve. A comparison between the sediment transport rates according to the two models is made below.

Run 021: Due to the strong initial flood-disturbance the flood transport rate from TRISULA is clearly less than that from TIFLAT. The points describing the ebb transport rates lie close to the curve from TIFLAT.

Run 121: The flood- as well as the ebb transport rates from the two models have approximately the same value, because in both models the amplitude of the velocity during flood and ebb are equal. Only in cross-section 2 this does not hold, because of an initial flood-disturbance.

Run 221: For this run goes the same as for the previous, except that the amplitudes are smaller because the slope of the flat is increased. Thus the velocity is decreased.
Figure 4.1: Water level and velocity in four cross-sections of run 021; flat with a horizontal slope; flood dominancy; $A = 1.0\text{m}$. Water level and velocity with TRISULA.
Figure 4.2: Velocity in four cross-sections of run 121; flat with a slope of $0.5 \times 10^3$; flood-dominance; $A = 1.0m$; water level shows similarities with the water level at the previous page.

Figure 4.3: Velocity in four cross-sections of run 221; flat with a slope of $1.0 \times 10^3$; flood-dominance; $A = 1.0m$; water level shows similarities with the water level at the previous page.
Process definition
Main processes
Dimension (106; 3; 1)
Distances Rectilinear (25.0; 25.0)
\((0^\circ \Rightarrow F_{cor}=0)\)
Grid enclosure Interactive (1,1; 106,1; 106,3; 1,3; 1,1)
Depth data Space varying (from file)
Boundary position Interactive (h-rand; H; Z; 1,2; 1,2)
\((u\text{-}rand; H; Z; 106,2; 106,2)\)

Input parameters
Numerical options numerical iterations (2), marginal depth (0.005)
drying procedure (max)
Simulation period \((0-1860; \Delta t = 1.0 \text{ min.})\)
Initial condition Uniform (1.3 m)
Open boundary condition
General parameter Duration of smoothing (0.0)
Hydrodynamics Harmonics \((h_M = 1.0 \cdot \cos(30[^\circ/h] - 0[^\circ]))\)
Physical coeff. Flow General \((g = 9.813 \text{ m/s}^2)\)
\((\rho = 1000 \text{ kg/m}^3)\)
Bedstress \((\text{Chézy})\)
Uniform \((C = 45 \text{ m}^{1/3}/\text{s})\)
Eddy visc. Uniform \((v = 1.0 \text{ m}^2/\text{s})\)

Output Selection
Site definition Stations \(((20,2),(30,2),(40,2),(50,2),60,2))\)
File types Map file \((\Delta t_{map} = 60.0 \text{ min.}, t_{0_{map}} = 1020 \text{ min.})\)
History file \((\Delta t_{his} = 1.0 \text{ min.})\)

*figure 4.4: Tree structure for input in TRISULA with applied parameters between brackets.*
4.2 Trisula 2-D

4.2.1 Model

In this section, the differences from the 1-D runs are described, following the items in the tree structure (figure 4.4) and in figure 4.5.

- The grid is: \( \Delta x = 400.0 \) m and \( \Delta y \) is still the same 25.0 m.
- The depth data change into a 106 x 25 matrix.
- The h-boundary condition (2-D) at the left boundary is imposed in the channel mouth and not at the flat area. The h-boundary condition is not asymmetric anymore.
- The Chézy coefficient in \( x \)- and \( y \)-direction in the channel differs from the Chézy coefficient at the flat. At the flat the Chézy coefficient, \( C_x = C_y = 45.0 \) m\(^{1/2}\)/s, remains the same. For the channel with a depth of 10.0 m according equation (5.1) \( C_x = C_y = 65.0 \) m\(^{1/2}\)/s.
- The cross-sections are situated halfway the total length of the basin. Also some vector plots are made to get an idea of the streamlines.

4.2.2 Results

Again, transport rates in the cross-sections 2, 3, 4 and 5 are calculated from the velocities. These sections are situated at the flat halfway the tidal basin (\( x = 10 \) km). Special attention should be paid that the flow pattern is 2-D (see the velocity maps in Appendix V, realised with 2-D results from TIFLAT). The chronologically placed maps show that the arrows in the same
nodes point in different directions, because the velocity direction alters during the tide. This
has an influence on the transport in y-direction, which is of concern. This time the flood-
sediment transport rate in y-direction is realised in the following way. According to equation
(3.8), the transport along a streamline becomes:

\[ s_{tot} = f(u_{tot}) = m \cdot u_{tot}^n \quad \text{with} \quad u_{tot} = \sqrt{u^2 + v^2} \quad \text{(4.1)} \]

in which:
- \( s_{tot} \): sediment transport rate along a streamline per unit width
- \( u_{tot} \): horizontal water particle velocity along a streamline
- \( u \): horizontal water particle velocity in x-direction
- \( v \): horizontal water particle velocity in y-direction
- \( m \): coefficient of proportionality
- \( n \): exponent

Equation 3.9 becomes:

\[ V_{f, y} = \int_{t_{f, 0}}^{t_{f, \text{end}}} (m \cdot \cos \alpha(t) \cdot u_{tot}(t)^n) \, dt \quad \text{(4.2)} \]

in which:
- \( V_{f, y} \): sediment transport rate per unit width per flood part of the tide in y-
direction
- \( t \): time
- \( t_{f, 0} \): time the flood starts
- \( t_{f, \text{end}} \): time the flood ends
- \( \alpha \): angle between the velocity direction and the y-axis

The same goes for the sediment transport rate during ebb. Again the required sediment
transport rates are solved numerically. The sediment transport rates from the 2-D runs are
shown in the same diagrams (see figures 3.6 through 3.8) as the rates from the 1-D runs, in
order to compare them with each other. The shape of all the curves is the same. Thus only the
values can make a difference. So there are no consequences for the sign of the derivative. The
value of the derivative determines only the rate of sedimentation or erosion.

 Runs without a slope of the flat:
The values of the sediment transport rate according to the 1-D and 2-D models agree with
each other. It implies that only the flood transport rate from TIFLAT is larger than the flood
transport rate from TRISULA. This is caused by the initial flood disturbance for a model with
a horizontal flat.

 Runs with a flat slope of 0.5 \times 10^{-3}:
The sediment transport rates of both TRISULA-runs seem to correspond better with the
results from TIFLAT. The 2-D results show a tendency to increase with respect to the 1-D
results as the slope increases.

 Runs with a flat slope of 10^{-3}:
The slope is larger with respect to the previous run. The values from the two 1-D runs agree
with each other, but the value from the 2-D run for \( y = 30 \cdot \Delta y \) goes astray. The amplitude of
the ebb sediment transport rate in the 2-D model is larger than the flood-amplitude. Thus the
2-D model shows an ebb-dominant behaviour, in contrast to the 1-D model.
5 Calculation of the suspended sediment transport

The material that is transported has a very small grain diameter, so that it can be brought into suspension by high velocities and that it does not settle too easily when the horizontal velocity decreases. The sediment concentration will be distributed over the vertical, but as the concentration is very small, density effects do not play a role.

TIFLAT includes a concentration solver which computes the suspended sediment concentration. The partial differential equation which describes the depth averaged concentration reads:

\[
\frac{\partial \bar{c}}{\partial t} + v \cdot \frac{\partial \bar{c}}{\partial y} = -\frac{w_s}{a} (\bar{c} - \bar{c}_e) \tag{5.1}
\]

The concentration can be solved by discretising (5.1). This results in the following equation:

\[
\frac{c_{i+1}^{l+1} - c_i^l}{\Delta t} + v_i^{l+1} \frac{c_i^l - c_{i-1}^l}{\Delta y} = -\frac{w_s}{a_i^{l+1}} (c_{i+1}^{l+1} - c_e^{l+1}) \tag{5.2}
\]

in which:
- \(c\) : depth average concentration
- \(i\) : grid index
- \(l\) : time index
- \(v\) : horizontal water particle velocity
- \(y\) : distance along horizontal axis
- \(w_s\) : particle settling velocity
- \(a\) : water depth
- \(c_e\) : depth average equilibrium concentration

Because the suspended sediment transport can be written by definition as:

\[
s = v a \bar{c} \tag{5.3}
\]

In the perturbation term, the right hand side of equation (5.1) \(c_e\) is assessed by:

\[
\bar{c}_e = \frac{s}{va} = \frac{m|v_i^{l+1}|^{n-1}}{a_i^{l+1}} \tag{5.4}
\]

in which:
- \(s\) : sediment transport rate per unit width
- \(m\) : coefficient of proportionality
- \(n\) : exponent
The differential equation for the suspended sediment transport requires, just like the set of equations for the water movement (3.1) and (3.2), a boundary condition. For the boundary condition is found, using equation (5.4) with a coefficient $m = 1.0$ and $n = 5$ according Engelund & Hansen, the average concentration during the tide, is 0.02. In fact the velocities in the direction of the main channel, i.e. perpendicular to the plane of figure 5.1 between $y_1$ and $y_{16}$ determine the concentration during the flood phase. A plot of the concentration in five cross-sections through the tidal period in run 021 is shown in figure 5.2.

![Figure 5.1: Cross-section of tidal inlet.](image)

![Figure 5.2: Concentration in five cross-sections for run 121; flat with a slope of 0.5 $10^{-3}$; flood-dominancy; $A = 1.0m$.](image)

Finally, the instantaneous suspended sediment transport in the nodal points on the flat is determined via equation (5.3). The sum of all these instantaneous transports determines the sediment transport rates, both for flood and for ebb. These transports are calculated the same way as the transports described in section 3.5. This way the suspended transport rate is known in every nodal point. It is shown in the same diagrams (figures at pages 22 and 23) as the results, from TIFLAT and TRISULA with the power law formula.
The results from the suspended sediment transport model, in which the equilibrium concentration is estimated according the power law, are at the flat consistent with the transport rates which follow directly from the power law. They both have an ebb-peak and show much similarity, except for a kind of phase lag near the transition channel/flat.

The phase lag is caused by the decay length, \( L_c = \nu \alpha / \omega_a \). Every time step there is a small decay length. The contribution to the phase lag is small, because \( \nu < 1 \) and \( \alpha < 1 \) throughout the tidal period. The instantaneous transport contributions are largest when the velocity is large. But during a tidal period \( L_c \) reaches a maximum in \( y_{26} \) when the velocity has just reached its maximum. \( L_c = 0.54 \cdot 0.45 / 0.03 = 8.1 \) m.

The transport in the main channel is not predicted very well so far. The velocity in the main channel (1-D) is very small. The transport rate according both models is therefore close to zero. Also the calculations with the concentration are not satisfactory, because the small velocities make the concentration decrease rapidly. Runs with a smaller particle settling velocity \( w_s \) do not result in sediment transport rates in the main channel, which are not almost zero. 2-D runs in TIFLAT must yield a better insight into the transport from channel onto the flat, using the transport formula of E&H again.
6 Correction of the ebb sediment transport rate

In section 3.5 it was already stated that it can make sense to introduce a threshold in the subroutine that adds the instantaneous sediment transport rate in a time step $\Delta t$ to the total sediment transport rate. The threshold is an imaginary and artificial value of water depth below which no sediment is transported. The threshold is read from the input file and appears in an IF-THEN-construction (see Appendix III). If the water depth is less than the threshold, the value zero is added to the cumulative sediment transport rate (ebb or flood).

In figure 3.9 water depth and velocity for the nodal points $y_{23}$ till $y_{27}$ were plotted versus the time for a flood-dominant run in which the flat had no slope. This figure confirms the presumption of a very thin water film. It makes clear that calculating the sediment transport rate from the velocity (power law) does not make much sense during approximately one third of the tidal period, because the water depth becomes too small.

What should be the value of the threshold? Again, several runs have been made to figure out what this value should be. For run 021 the threshold becomes 0.12 m, for the other ones 0.05 m is taken. The sediment transport rates with and without threshold, calculated in TIFLAT, are plotted in the figures at the next pages. The black curves correspond to the sediment transport rates determined from the power law formulas, while the coloured ones refer to the suspended sediment model.

![Flood and ebb sediment transport rate](image)

*Figure 6.1: Sediment transport rate for a flat with a horizontal slope and with a threshold of 0.12 m.*
Note that the sediment transport rates during flood are almost the same, irrespective of whether they are computed from the power law formula or from the concentration model. This is explained by the fact that the tidal front is steep and thus the water depth changes rapidly. The sediment transport rates during ebb are reduced so that a more smooth curve is obtained.

Figure 6.2: Sediment transport rate for a flat with a slope of $0.5 \cdot 10^{-3}$ and with a threshold of 0.05 m.

Figure 6.3: Sediment transport rate for a flat with a slope of $1.0 \cdot 10^{-3}$ and with a threshold of 0.05 m.
7 Parameter variation

7.1 Symmetrical water level boundary

As mentioned in Chapter 3 the water level boundary condition is chosen asymmetric. The further from the mouth of the inlet the more the tide is distorted, because of the increasing contribution of the $M_2$-component. At the mouth of the inlet frictional effects have not had much influence on the water movement and therefore the tide at the mouth is more symmetrical. In this section the influence of a symmetrical imposed water level at the boundary on the water and sediment motion is analysed.

The figures below are the velocities in five cross-sections during a tidal cycle. Each figure corresponds to a different geometry of the flat. A comparison between the figures in Section 3.4 and this section leads to the conclusion that the maximum flood velocity decreases and the maximum ebb velocity increases with respect to the maxima in the flood-dominant case. The flood maximum in every cross-section on a horizontal flat is still larger than the ebb maximum. However, some cross-sections are ebb-dominant if the flat has a slope of $0.5 \cdot 10^{-3}$. If the slope is $10^3$ five cross-sections are ebb-dominant. This must have its consequences for the sediment transport and the morphology of the flat.

![Velocity graph]

*Figure 7.1: Velocity in five cross-sections of run 001; flat with a horizontal slope; not an imposed dominancy; $A = 1.0m$; water level is almost a sine between -1 and +1.*
Figure 7.2: Velocity in five cross-sections of run 101; flat with a slope of $0.5 \cdot 10^{-3}$; not an imposed dominancy; $A = 1.0m$; water level is almost a sine between $-1$ and $+1$.

Figure 7.3: Velocity in five cross-sections of run 201; flat with a slope of $1.0 \cdot 10^{-3}$; not an imposed dominancy; $A = 1.0m$; water level is almost a sine between $-1$ and $+1$. 
The sediment transport resulting from one of the three runs is shown in figure 7.4 below. It is the sediment transport for a flat with a slope of $0.5 \cdot 10^{-3}$. The dashed line corresponds to a symmetrical water level boundary. The changed velocities have their consequences for the sediment transport. The transport during flood decreases and the transport during ebb increases. The morphological consequences depend on the dominancy of the sediment transport. A horizontal flat still accretes because the shape of the derivative ($\partial z/\partial t$) does not alter.

The equilibrium profile of a flat which is exposed to a symmetrical tide, is less steep than the profile in the flood-dominant case. This can be concluded from the sediment balance in the figure at the bottom of this page, for a flat with a slope of $0.5 \cdot 10^{-3}$. The lower parts of the flat are eroding.

**Figure 7.4:** Sediment transport rates of the runs 101 and 121; flat with a slope of $0.5 \cdot 10^{-3}$, $A = 1.0m$; the drawn line is a flood-dominant boundary and the dashed line a symmetrical boundary.

**Figure 7.5:** Net gradient of the sediment transport of the runs 101 and 121; flat with a slope of $0.5 \cdot 10^{-3}$, $A = 1.0m$; the black line is a flood-dominant boundary and the red line a symmetrical boundary.
7.2 Tidal range

As mentioned in chapter 2, the tidal range of a meso-tidal inlet is between 2 and 4 m. In this chapter it is attempted to find out whether there is a relationship between the sediment transport according E&H \( n = 5 \), and the tidal range. The transport determined in the \( x2I \)-runs is used as a reference. Of course the transport rates become larger, if the tidal range increases, because the derivative \( da/dt \) is larger. This implies that the velocity becomes larger.

Runs \( x2I, x22, x23 \) and \( x24 \) correspond respectively to \( H_{TR} = 2.0 \) m, 2.5 m, 3.0 m and 3.5 m. The sediment transport rates in the nodal points for the four runs for the same bathymetry of the flat are put into the same diagram (see the following pages).

A flat without a slope (\( x = 0 \)):

The maximum sediment transport rates for the flood occur at the beginning of the flat. The maxima seem to lie at equal distances from each other. If the tidal range increases with steps of 0.5 m, the sediment transport rate during flood increases by 0.8 times the value for \( H_{TR} = 2.0 \) m. This holds for the whole flat, because the sediment transport rates decrease almost linearly from the beginning till the end of the flat.

The ebb sediment transport rates are harder to predict, because the maximum is at the transition flat/channel. However they do show similarity with the sediment transport rates \( H_{TR} = 2.0 \) m.

\[ \text{Flood and ebb sediment transport rate} \]

\[ \text{figure 7.6: Sediment transport rates for several tidal ranges; flat with a horizontal slope.} \]

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A flat with a slope of e.g. $0.5 \cdot 10^{-3}$ ($\chi = 1$):

It is hard to find any relationship between the tidal range and the sediment transport if the flat has a slope. For the two largest tidal ranges it becomes clear that the flood transport is not linear anymore over the flat. This is caused by the bend in the profile, that was made to reduce resonance (see section 3.4). After the bend the sediment transport seems to become linear, again.

![Flood and ebb sediment rate](image)

*Figure 7.7: Sediment transport rates for several tidal ranges; flat with a slope of $0.5 \cdot 10^{-3}$.*
7.3 Length of the flat

Figure 7.8 shows the sediment transport rates for various flat lengths. If the flat length is large, the storage capacity becomes also large. From the pumping mode, equation (2.13), follows that if the length of the flat increases, then the velocities increase. That is why the maxima increase in the plot.

Again, the sediment transport rate during flood is almost linear over the flat. Conspicuous is the fact that the maximum sediment transport rate is achieved for a flat with a length of 4000 m. It looks as if the longer flat (5000 m) is already influenced by the slack water.

Figure 7.8: Sediment transport rates for a horizontal flat and for various lengths of the flat.
8 2-D sediment transport computation with TIFLAT

This chapter describes the 2-D computation of the sediment transport with the program TIFLAT. This is done to yield insight into the sediment transport in the channel and at the transition flat/channel. The results of the 1-D simulation show that sediment transport along the y-axis in the channel is very small, because the velocity in y-direction is small due to the large depth. Actually the velocity along a streamline determines the sediment transport. This requires a contribution of the x-component of the velocity and therefore 2-D modelling is necessary.

The original profile from the schematisation in chapter 3 is used, so that in the 2-D model the geometry extension, that was constructed to reduce resonance in the 1-D model, is omitted. Figure 8.1 shows the water level at five nodal points. The points are situated halfway the tidal inlet ($x = \frac{1}{2}L_{ch}$). Points 1, 2 and 3 are in the channel, point 4 is at the transition flat/channel and point 5 is on the flat.

![Water level](image)

*Figure 8.1: Water level for five points halfway a tidal inlet with a flat with a horizontal slope.*

The flats do not dry, because the water level is above zero. The tidal flat area and the length of the channel are too large to dry the flats, or, the gorge of the channel is too small to drain away the required amount of water to dry the flats. Two adjustments of the geometry are possible in order to achieve the drying of the flats. The first is to increase the area of the gorge, so that the discharge in the mouth increases. The second is to reduce the area of the basin. This option has been executed by reducing the length of the basin from 20 to 15 km and by linearly increasing the length of the flat from zero at the mouth to 1 km at the end of the basin (schematisation [van Dongeren and de Vriend, 1993]).

Results for the water level and the velocities halfway the channel in the five cross-sections are shown in figure 8.2. This time the water level sinks below bankful (cross-sections 1-4) so that the flat almost dries. The water level in cross-section 5 is approximately the level of the flat. It will flood again as the water level rises. Appendix V contains also some vector plots of the velocity, calculated with TIFLAT. These plots show the direction and the magnitude of the velocity in all the calculation points of the tidal inlet every time a tenth of the tidal period has passed. The tidal front during the flood phase is observable from the transition of large to small arrows.

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Figure 8.2: Water level, velocity in x-direction and y-direction for five points halfway a tidal inlet with a flat without a slope.
The net sediment transport, calculated the same way as described in Section 4.2.2, is also depicted in a vector plot, after the velocity plots in Appendix V. Because of the large differences between the magnitudes of the sediment transport in the channel and at the flat, there are two plots. One map is the area of the channel from \( y = 0 \) to \( y = 625 \) m and the other map is the flat area from \( y = 625 \) m to \( y = 1625 \) m. Mind that the vector scale differs in both plots. In the channel is the net sediment transport almost parallel to the channel axis and flood dominant because the \( y \)-velocity is very small and the \( x \)-velocity during flood is larger than during ebb. There must be sediment deposited in the channel. At the flat the direction of the net sediment transport varies, but is directed onto the flat, except at the abrupt bend, where channel turns into flat. Here, the net sediment transport is negative.

In order to verify whether what morphological patterns occur in the tidal basin the 2-D sediment balance for one tidal period is:

\[
\frac{\partial z}{\partial \tau} + \frac{\partial S(x)}{\partial x} + \frac{\partial S(y)}{\partial y} = 0
\]  
8.1

in which:  
\( z \) : bottom level  
\( \tau \) : time co-ordinate for variations at time scales much larger than the tidal period  
\( S(x) \) : tidally averaged sediment transport rate per unit width in \( x \)-direction  
\( S(y) \) : tidally averaged sediment transport rate per unit width in \( y \)-direction

The partial differential equation, discretised in space, becomes:

\[
\frac{\partial z}{\partial \tau} + \frac{S(x)_{j,1} - S(x)_{j,1-1}}{2\Delta x} + \frac{S(y)_{j,1} - S(y)_{j,1-1}}{2\Delta y} = 0
\]  
8.2

in which:  
i : grid index  
j : grid index

A positive derivative \( \partial z/\partial \tau \) corresponds to sedimentation of a cell and a negative one to erosion. Figure 8.3 shows a contour plot at which places sedimentation or erosion occurs. The areas that are coloured yellow and brown accrete. The rate of accretion in the brown area is stronger than in the yellow parts. The white areas do not participate in the flow conditions or will slightly erode. The red areas show more severe erosion.

The highly schematised tidal inlet leads to a strong accreting character of the channel and the flat. Only a part near the gorge, the transition zone channel/flat over the total length of the inlet and a part of the flat area halfway the inlet length are eroding. The last part is very remarkable, because the surrounding flat area is accreting and it looks if this part tries to create a small channel system.

If the resolution of the contour plots is finer - not printed here because of the large amount of colours - it appears that the rate of accretion of the flat area near the channel is higher than at locations near the end of the flat with the consequence that the slope of the flat tends to decrease and that the flat will accrete throughout to rise to above mean sea level. The tide is creating an overdepth for the flats [see van Dongeren and de Vriend, 1993].

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Figure 8.3: Erosion/sedimentation patterns of a tidal basin with a slope of $1.0 \times 10^{-3}$. 
An extra model run with a different geometry has been made to approach reality in a better way. The cross-section of a tidal basin is normally not prismatic with the length. Therefore the gorge of the basin has been widened with 500 m, so that more water can be drained away. Towards the end of the basin the channel gets linearly narrower with the length till the original dimensions of the cross-section.

Figure 8.4 shows the sediment balance for this case. Again, the flats are flood-dominant, except at the back of the channel. The area of the channel that erodes has become larger in respect to the previous model run. The flat area widens in both y-directions (negative and positive), but near the gorge the flat area is almost zero. In order to find out where the rate of accretion of the flats is largest along the channel the flat area at the gorge stretches in the next model run out till \( y = 1625 \) m, again. A finer resolution of the contour plots is shown in figure 8.5.

The finer resolution results in the fact that the sedimentation process at the flats near the gorge is stronger than the process at the flats in the back of the inlet.
Figure 8.4: Erosion/sedimentation patterns of a tidal basin with a funnel shaped channel and a flat slope of $1.0 \times 10^{-2}$. 
Figure 8.5: Erosion/sedimentation patterns of a tidal basin with a funnel shaped channel and a flat slope of $1.0 \times 10^{-3}$. Where is the strongest sedimentation at the flat? Near the tidal gorge.
9 Conclusions and recommendations

9.1 Conclusions

Although the problem of sediment exchange between flats and channels in tidal inlets has a 2-D character, this thesis focuses on a 1-D approximation. The sediment transport rate has mainly been determined from results of 1-D models. Considering the objectives formulated in the introduction, the following conclusions can be drawn:

Conclusions 1-D modelling:

- The 1-D TRISULA-runs for sloping flats yield similar results as the corresponding TIFLAT-runs (equal amplitudes of the velocity during flood and ebb). Consequently, the sediment transport rates have approximately the same value. In the 2-D runs, the transport rates become larger than in the 1-D runs, because of the influence of the water motion in the main channel.

- During flood, the sediment transport rate over a flat with a constant slope decreases almost linearly from a maximum at the beginning of the flat to zero at the end. The maximum depends on the local value of the coefficient \( m \) in the transport formula, the slope of the flat, the tidal range and the length of the flat. If the reservoir capacity of the flat is increased, the sediment transport rate increases due to the prolonged occurrence of relatively high flood-velocities.

- The sediment transport rate during ebb reaches very high values around the transition channel/flat. This is due to a very small water depth, combined with a relatively high velocity, occurring at the edge of the flat during approximately one third of the tidal period. This phenomenon is kept from dominating the ebb transport rate by introducing a threshold in TIFLAT. If the water depth is below this threshold the transport is zero. This results in a more realistic curve of the ebb sediment transport rate around the transition flat/channel. The introduction of this threshold hardly influences the flood sediment transport rate, because of the steep front of the tidal wave during flood.

- All cross-sections for the three flat areas with slopes of 0, 0.5\( \cdot 10^{-3} \) and 1.0\( \cdot 10^{-3} \) are flood-dominant in terms of the sediment transport, if the calculation is performed with a threshold. It means that the imported quantity of sediment is larger than the exported quantity. The dominance becomes less clear as the slope increases.

- The sediment balance for the net transport rate over a tidal period indicates whether initially erosion or deposition occurs. The flats with slopes of 0, 0.5\( \cdot 10^{-3} \) and 1.0\( \cdot 10^{-3} \) accrete and the side slope of the main channel becomes less steep. The rate of accretion of the higher parts of the flat is larger than the rate for the parts near the main channel. As a consequence, sediment from the main channel slope is deposited at the higher parts of the flat and the flat becomes steeper.
Conclusions and recommendations

- It appears that for larger slopes \(2.0 \cdot 10^{-3}\) or \(3.0 \cdot 10^{-3}\) the higher cross-sections of the flat become ebb-dominant in terms of the sediment transport. Thus, the aforementioned increase in the flat slope will not continue forever. Apparently, the equilibrium flat slope lies somewhere between \(1.0 \cdot 10^{-3}\) and \(2.0 \cdot 10^{-3}\).

- This equilibrium slope holds for a flat which is exposed to a flood-dominant tide in the channel. Closer to the mouth of the inlet, the tidal asymmetry is smaller. This results in a less steep equilibrium slope, because the flat is already ebb-dominant for the sediment transport if the slope is \(1.0 \cdot 10^{-3}\). Thus, the steepness of the equilibrium slope decreases as the tidal asymmetry decreases.

Conclusions 2-D modelling:

- 2-D modelling is a necessity to approximate reality, since the inundation of the flats has a distinctly 2-D character. The magnitudes of the sediment transport in the channel - the exchange between channel and flat - become clear, because of the (large) velocity vector, the velocity along a streamline.

- The level of the flats at mean sea level results in transport onto the flat. If the flat level is chosen higher, a stable situation of the flats seems to establish itself, because sedimentation can not go on forever. The flat can be in equilibrium, but the shape of the profile is unknown. Morphological computations can verify whether the theory of convexity/concavity [Friedrichs et al., 1994] in chapter 0 holds or whether the profile will be totally flat, because the rate of accretion of the flat area is largest nearest to the channel.

- The slope of the flats near to the gorge of the basin or the level of those particular flat areas will respectively be larger or higher than the slope or the level of the flats at the back, because the rate of accretion is larger. In the 2-D model runs only the M2-component was modelled.

Conclusions 1-D and 2-D modelling:

- The drying/flooding procedure is critical in the case of a horizontal flat. In the TRISULA model a threshold is used for the water motion. If the water depth is below this threshold, then the point is set dry. For large flat areas problems occur, when the water level rises again. The stored amount of water behind the threshold produces a current in a direction opposite to the ones to be expected for a rising tide.

- Closer to the mouth of the inlet, the tidal asymmetry is smaller. This results in a less steep equilibrium slope (see conclusions 1-D). Thus, the steepness of the equilibrium slope decreases as the influence of the tidal asymmetry decreases. This is opposing the last 2-D modelling conclusion, that says that steeper/higher slopes of the flats occur near the gorge.
9.2 Recommendations for further research

The work presented in this thesis has led to the following observations and recommendations for further research:

- Large flats with a horizontal slope, which dry completely, can be easier modelled in TIFLAT than in TRISULA. If the grid size in both models is equal and not exceptionally small the TRISULA run will cause an initial ebb disturbance of the water movement as the flooding of the flat starts.

- 1-D and 2-D morphological models, with the geometry as described in chapter 3, should be utilised to verify whether the level of the flats will rise, under the influence of the tide only.

- In the case of 2-D modelling, suspended sediment transport should be considered, with special attention for the sediment that is stirred up and brought into suspension by high velocities in the main channel. Further, some amount of sediment is brought onto the flat. In the 1-D model runs the stirring factor of the high velocity in the main channel was absent.

- More realistic geometry of the tidal inlet should be used e.g. a funnel shaped inlet, avoid sharp bends (e.g. the transition flat/channel), an inlet that is more shallow at the back, flat area that borders the end of the inlet, so that no reflection of the incoming tidal wave occurs, etc.
## List of Symbols

<table>
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<tr>
<th>Symbol</th>
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<td>L^2/T</td>
</tr>
<tr>
<td>t</td>
<td>time</td>
<td>T</td>
</tr>
<tr>
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</tr>
<tr>
<td>T_c</td>
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<td>T</td>
</tr>
<tr>
<td>u</td>
<td>horizontal water particle velocity</td>
<td>L/T</td>
</tr>
<tr>
<td>u_{out}</td>
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<td>L/T</td>
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<tr>
<td>v</td>
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<td>L/T</td>
</tr>
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<td>Definition</td>
<td>Dimension</td>
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<td>$V$</td>
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<tr>
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<td>$L/T$</td>
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<tr>
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<td>particle settling velocity</td>
<td>$L/T$</td>
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<tr>
<td>$x$</td>
<td>distance along horizontal axis</td>
<td>$L$</td>
</tr>
<tr>
<td>$y$</td>
<td>distance along horizontal axis</td>
<td>$L$</td>
</tr>
<tr>
<td>$z$</td>
<td>bottom level</td>
<td>$L$</td>
</tr>
<tr>
<td>Greek letters</td>
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</tr>
<tr>
<td>$\alpha$</td>
<td>angle between $U + V$</td>
<td>-</td>
</tr>
<tr>
<td>$\Delta$</td>
<td>relative density bed material</td>
<td>-</td>
</tr>
<tr>
<td>$\zeta$</td>
<td>instantaneous water elevation</td>
<td>$L$</td>
</tr>
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<td>$\zeta_0$</td>
<td>amplitude at the boundary of the inlet</td>
<td>$L$</td>
</tr>
<tr>
<td>$\Theta$</td>
<td>phase</td>
<td>-</td>
</tr>
<tr>
<td>$\Lambda$</td>
<td>dominancy parameter</td>
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</tr>
<tr>
<td>$\mu$</td>
<td>ribble factor</td>
<td>-</td>
</tr>
<tr>
<td>$\nu$</td>
<td>kinematic viscosity</td>
<td>$L^2/T$</td>
</tr>
<tr>
<td>$\rho$</td>
<td>mass density of water</td>
<td>$M/L^3$</td>
</tr>
<tr>
<td>$\sigma$</td>
<td>Courant number</td>
<td>-</td>
</tr>
<tr>
<td>$\Phi$</td>
<td>relative phase difference</td>
<td>-</td>
</tr>
<tr>
<td>$\Psi$</td>
<td>phase</td>
<td>-</td>
</tr>
<tr>
<td>$\omega$</td>
<td>tidal wave frequency</td>
<td>$T^{-1}$</td>
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</table>
References


Appendices

Appendix I

Before the adjustments due to resonance

Input file run 001

```
66  3 A: NMAX,NMAX
25.00 25.00 B: DX,DY
10000 10000 10000 10000 10000 10000 10000 9000 8000 7000
6000 5000 4000 3000 2000 1800 1600 1400 1200 1000 800 600
400 200 0 0 0 0 0 0 0 0 0 0 0
0 0 0 0 0 0 0 0 0 0 0 0 0
10000 10000 10000 10000 10000 10000 10000 9000 8000 7000
6000 5000 4000 3000 2000 1800 1600 1400 1200 1000 800 600
400 200 0 0 0 0 0 0 0 0 0 0 0
0 0 0 0 0 0 0 0 0 0 0 0 0
10000 10000 10000 10000 10000 10000 10000 9000 8000 7000
6000 5000 4000 3000 2000 1800 1600 1400 1200 1000 800 600
400 200 0 0 0 0 0 0 0 0 0 0 0
0 0 0 0 0 0 0 0 0 0 0 0 0

1 D: NOROW
2 2 65 2 3 D
64 D: NOCOL=
2 2 2 1 1 D
3 2 2 1 1 D
4 2 2 1 1 D

↓

63 2 2 1 1 D
64 2 2 1 1 D
65 2 2 1 1 D
0 0 0 0 E: IDX1,IDX1,IDX2,IDX2
1.05 F: ZETA0, INITIAL WATER LEVEL
2 G: NROB(NUMBER OF OPEN BOUNDARIES)
1 2 2 1 1 0.00 1.00 43200 .0 .0 .0
21600 -90.0 G
66 2 1 2 1 0.00 0.00 43200 .0 .0 .0
21600 .0 G
25 1 3100 H: DT,NT0,NT1
1 2 .10E+06 I: ITER1,ITER2,EPS
.00 .00 1000.00 .00 .00 J: VI,FF,RHOM,WINDU,WINDV
.00 .50 2 K: TRSH,DP0,NFLYP
45.00 L: CZDEF
50 20 2 30 2 40 2 50 2 60 2 M: MAP INTERVAL,HISTORY POINTS
```
After the adjustments due to resonance

Input file run 111

106  3  A: MMAX, NMAX
   25.00  25.00  B: DX, DY
10000 10000 10000 10000 10000 10000 10000 8000 7000 6000
5000  4000  3000  2000  1799  1599  1399  1199  999  799  599  399
199  0  0  0  0  0  0  0  0  0  0  0  0
0  0  0  0  0  0  0  0  0  0  0  0  0
0  0  0  0  0  0  0  0  0  0  0  0  0
-687  -718  -749  -781  -812  -843  -874  -906  -937  -968  -999
-1031
-1062 -1093 -1124 -1156 -1187 -1218 -1249
10000 10000 10000 10000 10000 10000 10000 8000 7000 6000
5000  4000  3000  2000  1799  1599  1399  1199  999  799  599  399
199  0  0  0  0  0  0  0  0  0  0  0  0
0  0  0  0  0  0  0  0  0  0  0  0  0
0  0  0  0  0  0  0  0  0  0  0  0  0
-687  -718  -749  -781  -812  -843  -874  -906  -937  -968  -999
-1031
-1062 -1093 -1124 -1156 -1187 -1218 -1249
10000 10000 10000 10000 10000 10000 10000 8000 7000 6000
5000  4000  3000  2000  1799  1599  1399  1199  999  799  599  399
199  0  0  0  0  0  0  0  0  0  0  0  0
0  0  0  0  0  0  0  0  0  0  0  0  0
0  0  0  0  0  0  0  0  0  0  0  0  0
-687  -718  -749  -781  -812  -843  -874  -906  -937  -968  -999
-1031
-1062 -1093 -1124 -1156 -1187 -1218 -1249
10000 10000 10000 10000 10000 10000 10000 8000 7000 6000
5000  4000  3000  2000  1799  1599  1399  1199  999  799  599  399
199  0  0  0  0  0  0  0  0  0  0  0  0
0  0  0  0  0  0  0  0  0  0  0  0  0
0  0  0  0  0  0  0  0  0  0  0  0  0
-687  -718  -749  -781  -812  -843  -874  -906  -937  -968  -999
-1031
-1062 -1093 -1124 -1156 -1187 -1218 -1249
1 D: NOROW
2 2 105  2 3 3 D
0 0 0 0 0 0 0 0 0 0 0 0 0 0 0 0 0 0 0
1.35  F: ZETA0, INITIAL WATER LEVEL
2  G: NROB(NUMBER OF OPEN BOUNDARIES)
0 0 0 0 0 0 0 0 0 0 0 0 0 0 0 0 0 0 0
21600 -90.0 G
0.0  2 1 2 1 0.00  0.00  43200  0.0  0.0
21600 0.0 G
0.0  25 1296 3100  H: DT,NTG,NT1
1 2 .10E+06 1: ITER1, ITER2, EPS
0.0  0.0 1000.00 .00 .00 J: VI, FF, RHOM, WINDU, WINDV
0.0  0.0 2 K: TRSH, DFO, NFLYP
45.00 L: CZDEF
50 20 2 30 2 40 2 50 2 60 2 M: MAP INTERVAL, HISTORY POINTS

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Appendix II

Program, that calculates the bathymetry for the channel and the flat, appropriate for the input file that TIFLAT reads.

```fortran
        dimension idp (3,201)
        character*3  ext
        character*10 output
        character*20 form
        write (*,'(" Geef extentie aan! Naar welke idpres.ext file moet
        *profiel worden weggeschreven: '',$")')
        read (*,'(a3)') ext
        output = 'idpres.'//ext
        open (6, file = output,
        *           form = 'formatted', recl = 3622)
        write (*, '(" verhang geul = 0.04")')
        verh1 = .04
        verh2 = .008
        idp0 = 10000
        write (*, '(" verhang plaat = ")')
        read (5, '(f5.3)') verh3
        write (*, '(" De celafl meting id = ")')
        read (5, '(i2)') idx
        write (*, '(" De diepte in de geul idp0 = ")')
        c       write (5, '(i5)') idp0
        c
        read (5, '(i5)') idx
        ixg = 1500/idx + 6
        write (form, '(()('' 'i2'''i6)'')) ixg
        do 100 n = 1,3
          do 200 k = 1,ixg
            somdx = (k-1)*idx
            if (sodmx .le. 175.) then
              idp (n,k) = idp0
            elseif (sodmx .le. 375.) then
              idp (n,k) = idp0 - (k-175/idx-1)*verh1*idx*1000
            elseif (sodmx .le. 625.) then
              idp (n,k) = 2000 - (k-375/idx-1)*verh2*idx*1000
            else
              idp (n,k) = idp0 - 10000 - (k-625/idx-1)*idx*verh3*1000
            endif
            200          continue
        100          continue
        do 300 n = 1,3
          write (6, form) (idp (n,k),k=1, ixg)
        300          continue
        end
```
Appendix III

Subroutine that calculates sediment transport rates (ebb and flood) according a power law. This is a sum of the instantaneous transports.

\[ V_\beta = \sum_{\beta \leq b} u_n \]

with \( b \in \{3, 3.5, 4, 4.5, 5\} \) and \( n \in \{0, T/\Delta t\} \)

Later formulae are added to calculate the sediment transport rates on basis of the depth averaged suspension transport equation. After that also a threshold has been added.

c subroutine transp(ul ,nmax ,mmax ,h ,c1 ,
  * veb0 ,veb1 ,vvl0 ,vvl1 ,nt
  ,nt0 ,
  * vec0 ,vec1 ,vvc0 ,vvc1 )
c Date : April 16, 1997
c Author : B.B. van Marion
c Agency : DELFT HYDRAULICS, DELFT THE NETHERLANDS
c Copyright : 1997 DELFT van Marion ltd.
c C Function : Calculates transports
C COMMON /NUMECO/ DT ,HDT ,TRSH ,vtresh,WINDU ,WINDV
c
dimension ul (nmax,-1:mmax+1),
  * veb0(5,nmax,-1:mmax+1),vvl0(5,nmax,-1:mmax+1),
  * veb1(5,nmax,-1:mmax+1),vvl1(5,nmax,-1:mmax+1),
  * vec0(nmax,-1:mmax+1) ,vvc0(nmax,-1:mmax+1) ,
  * vec1(nmax,-1:mmax+1) ,vvc1(nmax,-1:mmax+1) ,
  * h (nmax,-1:mmax+1) ,c1 (nmax,-1:mmax+1)
if (nt.ge.nt0.and.nt.lt.nt0 + 43200/dt) then
do 300 k = 6,10
  b = .5*k
  do 300 n = 1,nmax
    do 300 m = 1,mmax
      if (ul(n,m).le..0.and.h(n,m).gt.vtresh) then
        veb1(k,n,m) = veb0(k,n,m) + ul(n,m)*
        (abs(ul(n,m))**(b-1))*dt
        vvl1(k,n,m) = vvl0(k,n,m)
      elseif (ul(n,m).gt..0.and.h(n,m).gt.vtresh) then
        vvl1(k,n,m) = vvl0(k,n,m) + (ul(n,m)**b)*dt
        veb1(k,n,m) = veb0(k,n,m)
      else
        veb1(k,n,m) = veb0(k,n,m)
        vvl1(k,n,m) = vvl0(k,n,m)
      endif
  enddo
  enddo
enddo

300     continue
310     do 400 n = 1, nmax
320       do 400 m = 1, mmax
330         if (ul(n,m) .le. 0. and. h(n,m) .gt. vthresh) then
340           vec1(n,m) = vec0(n,m) + ul(n,m) * h(n,m) * cl(n,m) * dt
350           vvc1(n,m) = vvc0(n,m)
360         elseif (ul(n,m) .gt. 0. and. h(n,m) .gt. vthresh) then
370           vvc1(n,m) = vvc0(n,m) + ul(n,m) * h(n,m) * cl(n,m) * dt
380           vec1(n,m) = vec0(n,m)
390         else
400           vec1(n,m) = vec0(n,m)
410           vvc1(n,m) = vvc0(n,m)
420       endif
430     endif
440     continue
450     else
460       goto 500
470     endif
480 500     return
490 end
Appendix IV

Figure shows resonance.
Appendix V

Contents

Figures 1-11: Velocity vector plots during a tidal period with an interval of a tenth of the period, post-processed from results from a run in the program TIFLAT.

Figure 12: Vector plot of the net sediment transport in the channel. The net sediment transport is calculated in a subroutine in the program TIFLAT.

Figure 13: Vector plot of the net sediment transport at the flat.
t = 0.0 T after H.W. at the gorge at (0, 0)

DELFt HYDRAULICS

fig1
t = 0.1 \text{ T after H.W. at the gorge at (0, 0)}

VECTOR SCALE:
2.0 m/s

Velocity vectors
t = 0.3 T after H.W. at the gorge at (0, 0)

VECTOR SCALE: 2.0 m/s

Velocity vectors
t = 0.4 T after H.W. at the gorge at (0, 0)

VECTOR SCALE:
2.0 m/s → Velocity vectors
$t = 0.5 \, T$ after H.W. at the gorge at $(0, \, 0)$
$t = T$ after H.W. at the gorge at (0, 0)