Turbulent exchange of fine sediments in tidal flow
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PROEFSCHRIFT

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Abstract

The morphodynamic and biodynamic behaviour of estuaries and coastal regions is of great economic and environmental interest. Accumulations of fine cohesive sediments in harbours and navigation channels may hinder navigation and form a source of potential pollution as heavy metals and pesticides are adsorbed by cohesive sediments. Deficiencies in predictive capability of cohesive sediment transport models, which are used as tools in estuarine and coastal management, arise from, among other things, the lack of knowledge of cohesive sediment properties and sediment-induced buoyancy effects on turbulence, inadequate and too short records of hydrodynamic and sedimentary parameters at too few sites.

This study addresses the role turbulence plays in erosion, deposition, and small-scale mixing of cohesive sediment in tidal flows, with emphasis on sediment-induced density stratification resulting in dampening of turbulence. The approach is to quantify the turbulence properties of water and sediment motions in relation to vertical distributions of mean velocities and sediment concentrations. To that end, in-situ high-frequency measurements were made of velocities and suspended sediment concentrations at three levels close to the bed at one location in a tidal channel during a few tidal cycles in June and August 1996.

Flux Richardson and gradient Richardson numbers were calculated for selected parts of the measuring periods. The results showed that effects of sediment-induced density stratification on the turbulence properties occurred at higher levels in the water column and during those parts of the tidal cycle when flow velocities were relatively small or concentrations were relatively high. Effects of density gradients also showed in a decrease in the streamwise integral length scale and a decrease in correlation between streamwise and vertical velocity fluctuations.

Computations with a 1DV mathematical model indicated that rapid settling of suspended sediment towards slack water can be explained from buoyancy effects. The computations also indicated a limited availability of sediment from the bed during high flow velocities. Laboratory experiments carried out in a rotating annular flume with bed material from the research area showed that limited availability was a result of high bed strengths.
Samenvatting

Het morfodynamisch en biodynamisch gedrag van estuaria en kusten is van groot belang voor economie en milieu. Accumulaties van fijne cohesieve sedimenten in havens en vaarwegen kunnen hinderlijk zijn voor scheepvaart en vormen een bron van potentiële vervuiling. Pesticiden en zware metalen hechten zich namelijk bij voorkeur aan cohesieve sedimenten. Tekortkomingen in het voorspellend vermogen van sediment-transportmodellen, die worden gebruikt bij het ontwikkelen van beleid voor estuaria en kusten, komen voort uit onder andere de gebrekkige kennis van de eigenschappen van cohesieve sedimenten en van de door het sediment geïnduceerde gelaagdheidseffecten op de turbulentie, en uit inadequate en te korte meetreeksen van hydrodynamische en sediment-parameters op een te klein aantal locaties.

Deze studie richt zich op de rol die turbulentie speelt bij erosie, depositie en verticaal turbulent transport van cohesieve sedimenten in een getijstroming. De nadruk ligt daarbij op door sediment geïnduceerde stratificatie die resulteert in een demping van de turbulentie. De aanpak is turbulentieparameters te quantificeren in relatie tot verticale profielen van gemiddelde snelheden en sedimentconcentraties. Daartoe zijn hoogfrequente metingen uitgevoerd van stroomsnelheden en sedimentconcentraties dicht bij het bed van een getijgeul gedurende een aantal getijcycelen in juni en augustus 1996.

Flux Richardson en gradient Richardson getallen zijn bepaald voor geselecteerde gedeelten van de bovengenoemde meetperioden. Uit deze resultaten kwam naar voren dat gelaagdheidseffecten van de door sediment geïnduceerde dichtheidstratificatie op de turbulentieparameters het duidelijkst aanwezig waren op hogere locaties in de waterkolom en gedurende de delen van de getijcylus waarin stroomsnelheden relatief gering waren en sedimentconcentraties relatief hoog waren. Gelaagdheidseffecten kwamen ook tot uiting in een reductie van de integrale lengteschaal in de stromingsrichting, en in een reductie van de correlatie tussen de horizontale en verticale snelheidsfluctuaties.

Berekeningen met een ééndimensionaal wiskundig model lieten zien dat een snelle afname van de sedimentconcentratie in de perioden vlak voor kentering verklaard kan worden door gelaagdheidseffecten. Deze berekeningen gaven ook aan dat tijdens maximale stroomsnelheden de hoeveelheid sediment beschikbaar voor erosie, beperkt was. Laboratorium experimenten, met sediment uit het onderzoeksgebied, gaven aan dat een grote bedsterkte de beperkte beschikbaarheid kan verklaren.
The investigations were supported by the Netherlands Geosciences Foundation (GOA) with financial aid from the Netherlands Organisation for Scientific Research (NWO).
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Chapter 1 Introduction

1.1 General
The morphodynamic and biodynamic behaviour of estuaries and coastal regions is of
great economic and environmental interest. Accumulations of fine cohesive sediments
in harbours and navigation channels may hinder navigation and form a source of
potential pollution as heavy metals and pesticides are adsorbed by cohesive sediments.
Large investments are needed for measures to dispose dredging waste. Deposits of
cohesive sediment along shores and on intertidal flats form important breeding grounds
for marine life and feeding grounds for many species of birds. These reaches are to be
preserved and protected against, for example, disposed polluted harbour sludge. Such
demands confront harbour and waterway authorities with many design, maintenance
and management problems.

A better understanding of the processes which control the behaviour of cohesive
sediment is required to improve mathematical models that are used by harbour and
waterway authorities for predicting effects of engineering works on the water quality,
accretion rates of intertidal flats and siltation rates of channels and harbours. Deficiencies in predictive capability of cohesive sediment transport models, arise from,
among other things, the lack of knowledge of cohesive sediment properties, sediment-
induced buoyancy effects on turbulence, inadequate and too short records of
hydrodynamic and sedimentary parameters at too few sites. Short term records do not
usually account for episodic influences like high river run-off, storms, or algae blooms
(Mehta, 1989; Kornman and De Deckere, 1998).

A Dutch interdisciplinary research programme in the Ems/Dollard estuary was
initiated in the early nineties which addressed some of the above mentioned
shortcomings in cohesive sediment transport modelling (BOA theme-project). Field
measurements of factors controlling the behaviour of cohesive sediment were to be
performed at regular intervals on a seasonal time scale. Biological and physical
processes were also to be studied on short time scales and in isolated situations, such
as surface erosion in field flumes on an intertidal flat, flocculation processes and
vertical mixing in a tidal channel, mucus production by algae in small scale laboratory
systems, entrainment of fluid mud in laboratory flumes, etc. All processes were to be
parameterised and integrated in a numerical model which were to be validated against the results of field measurements.

![Diagram of sediment processes](image)

Figure 1.1. Conceptual model of controlling factors and processes that characterise the cohesive sediment behaviour under typical episodic conditions.

Quantifying vertical exchange processes is desirable in order to obtain more insight into the physical and biodynamical behaviour of flat/channel systems, since the horizontal exchange of suspended fine sediment in such systems takes mainly place in the upper part of the water column (see Figure 1.1). Turbulence strongly influences the erosion, flocculation and deposition of cohesive sediments. The present study is part of the BOA theme-project. It addresses the role turbulence plays in erosion, deposition, and vertical mixing of cohesive sediment.

1.2 Literature review
Various aspects of turbulence in stratified shear flows have been studied in laboratory set-ups (e.g. Webster, 1964; Komori et al., 1983) and in atmospheric boundary layers (e.g. Kaimal, 1973), but comparatively few studies were conducted in estuarine flows. A selection of literature on methodology and results of in-situ measurements in estuarine flows is summarised in this section (see also Table 1.1). Emphasis is placed on effects of salinity and sediment-induced density gradients on the turbulence properties. Effects of salinity gradients are included because they may be similar to those of sediment concentration gradients.
The notation used herein is as follows: the instantaneous streamwise velocity component \( U \) is written as the sum of a fluctuating part, \( u' \), and an ensemble mean, \( \bar{U} \). Similar notations are used for the vertical velocity component, \( W \), salinity, \( S \), and the sediment concentration, \( C \). Root mean square (RMS) values of the fluctuating parts are denoted by a prime (e.g. \( u' \)), covariances representing the vertical transports of momentum and sediment, for example, are written as \( \bar{uw} \) and \( \bar{cw} \), respectively.

Bowden and Howe (1963) found from measurements in the Mersey estuary that relative turbulence intensities \( u'/\bar{U} \) decreased with an increase in relative height above the sediment bed. The ratio \( w'/u' \) had a mean value of 0.5 ± 0.1, and the correlation coefficient \( R_{uw} = \frac{\bar{uw}}{u'w'} \) was about 0.2. The turbulence intensity \( u'/\bar{U} \) and \( R_{uw} \) were about half of those measured in the Red Wharf Bay near a bed which consisted of fine sand. The relatively small values of \( u'/\bar{U} \) and \( R_{uw} \) in the Mersey estuary were attributed to the presumed presence of high silt concentrations in the first two metres above the bed.

Heathershaw and Thorne (1985) examined the incipient motion of seabed gravels (diameter 4 - 32 mm) with a hydrophone and a TV-camera. It was found that gravel movements were mainly controlled by peaks in streamwise velocity. It was concluded that coarse sediment movement is better correlated with form induced drag than with the instantaneous shear stress. Effects of this gravel on the turbulence were not mentioned.

West et al. (1986) measured turbulent perturbations of velocities by means of a 55-mm diameter electromagnetic flow meter (EMF) in a partially mixed reach of the Great Ouse Estuary (UK). The objective of this study was, among other things, to determine the usefulness of an EMF for measuring turbulence in estuarine flows. About 100 turbulence time series of about 5 or 10 minutes were recorded, of which 75 were rejected because of obvious spurious noise caused by power cables, measuring frame vibration, earth looping, wind and ship-induced waves and unknown factors. The large number of rejected records shows the problems in making turbulence measurements in the field. It was concluded from the 25 remaining records that the EMF was a suitable instrument for detecting the turbulence properties in the stratified flows of the Great Ouse estuary. For gradient Richardson numbers, \( R_l = -g \left( \frac{\partial \rho}{\partial z} \right) \rho (\partial \bar{U} / \partial z)^2 \) (where \( \rho \) is the density of the water, \( g \) the acceleration of gravity, and \( z \) the vertical coordinate positive upward) smaller than 0.15, the turbulence characteristics of the estuarine flow were, generally speaking, similar to those of homogenous laboratory
Table 1.1. Turbulence measurements of solutes and particles in tidal flow.

<table>
<thead>
<tr>
<th>Source</th>
<th>Instrumentation</th>
<th>Location</th>
<th>Measured ( u ), ( v ), ( w ), ( s ), ( c )</th>
<th>Re</th>
<th>( z/h )</th>
<th>( Ri_z )</th>
<th>Results</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bowden and Howe (1963)</td>
<td>EMF</td>
<td>Mersey estuary, Red Wharf Bay U.K.</td>
<td>( \times ) ( \times ) ( \times ) ( \times )</td>
<td>1.4×10^7</td>
<td>0.1 - 0.7</td>
<td>?</td>
<td>Intensities, correlations, spectra</td>
</tr>
<tr>
<td>Heathershaw and Thorne (1985)</td>
<td>EMF, Underw. TV, Hydrophone</td>
<td>West Solent, South Coast U.K.</td>
<td>( \times ) ( \times ) ( \times )</td>
<td>1.8×10^7</td>
<td>0.01 - 0.02</td>
<td>?</td>
<td>Quadrant analysis</td>
</tr>
<tr>
<td>West et al. (1986)</td>
<td>EMF</td>
<td>Great Ouse estuary U.K.</td>
<td>( \times ) ( \times ) ( \times )</td>
<td>4.5×10^6</td>
<td>0.2 - 0.7</td>
<td>0 - 1</td>
<td>Intensities, correlations, quadrant analysis</td>
</tr>
<tr>
<td>Shiono and West (1987)</td>
<td>EMF</td>
<td>Conway estuary U.K.</td>
<td>( \times ) ( \times ) ( \times )</td>
<td>1.1×10^6</td>
<td>0.1 - 0.2</td>
<td>0 - 0.3</td>
<td>Intensities, correlations, quadrant analysis</td>
</tr>
<tr>
<td>West and Shiono (1988)</td>
<td>EMF, Induction salinometer</td>
<td>Teign estuary U.K.</td>
<td>( \times ) ( \times ) ( \times )</td>
<td>1.0×10^6</td>
<td>0.3 - 0.6</td>
<td>0 - 1</td>
<td>Intensities, correlations, quadrant analysis, mixing lengths</td>
</tr>
<tr>
<td>West and Oduyemi (1989)</td>
<td>EMF, silt meter</td>
<td>Conway estuary U.K.</td>
<td>( \times ) ( \times ) ( \times )</td>
<td>1.1×10^6</td>
<td>0.1 - 0.7</td>
<td>0 - 1</td>
<td>Intensities, correlations, spectra, quadrant analysis</td>
</tr>
</tbody>
</table>
flow (Komori et al., 1983) the relative intensities \( u' / U \) and \( w' / U \) decreased with increasing relative height above the sediment bed, the ratio \( w' / u' \) had a mean value of 0.6 ± 0.15, and the correlation coefficient \( R_{uw} \) had a mean value of 0.4 ± 0.1. From a few records for which \( Ri_g \) was larger than 0.3 it was concluded that ejection events were more pronounced than sweeps, and that \( R_{uw} \) and the streamwise integral length scale \( L_x \) reduced with \( Ri_g \) (at \( z/h \approx 0.6 \)). The energy spectra for these periods were found to behave more or less in a similar manner as those of atmospheric data in the presence of stable stratification (Kaimal, 1973).

Shiono and West (1987) performed turbulence measurements over a flat section of a channel bed approximately 9 km from the mouth of the Conway estuary for approximately 1.5 hours during both flood and ebb tides. The purpose of these measurements was to investigate the turbulence conditions closer to the bed than during the Great Ouse field measurements (West et al., 1986). At some distance from the bed the velocity profiles were far from logarithmic, which was thought to be caused by a combination of secondary flow and longitudinal density-gradient effects. The salinity data showed fairly well-mixed conditions near the bed for the flood tide, but relatively large vertical density gradients over nearly the whole water column during ebb tide.

A very prominent feature of the flow were longer-period perturbations, which covered lower frequencies for flood tides than for ebb tides. A number of possible causes were mentioned, e.g. the generally lower vertical density gradients during flood tide. The longer-period perturbations accounted for about 20% of \( u' \) and \( uw \).

Although an appreciable density gradient was present at the measurement point during ebb tides with \( Ri_g \) values up to about 0.3, short period (< 68 s) perturbations near the bed behaved in a manner which was very similar to the bed-generated turbulence found in unstratified flow in laboratory flumes; (1) \( u' / u_* \) was similar for flood and ebb and was approximately 1.8 for \( z/h \) between 0.15 and 0.2, (2) a logarithmic velocity distribution was present close to the bed (\( z < 0.45 \) m), (3) the probability density distributions showed a good agreement with laboratory data, (4) a quadrant analysis gave results similar to laboratory flow data (Nakagawa and Nezu, 1977). The values of \( R_{uw} \) were 0.33 on average and showed little variation with \( Ri_g \). The distinct differences between flood and ebb profiles led to friction factors, \( C_d = \tau_b / (\rho_d U_f^2) \) where \( \tau_b \) is the bed shear stress and the subscripts \( d \) denote depth averaged values, of \( 3.8 \times 10^{-3} \) and \( 1.8 \times 10^{-3} \) for flood and ebb, respectively. The streamwise integral length scale \( L_x \) at ebb tide was about half of that at flood tide.
West and Shiono (1988) measured turbulence fluctuations of salinity and horizontal and vertical velocity components in the centre of an ebb dominated channel in the Teign estuary. The objectives of this study were to determine the vertical transports of solutes and momentum in stably stratified flows. The relative intensities \( u' \bar{U} \) and \( w' \bar{U} \) and hence \( w'/u' \) were not significantly affected by stability effects, but depended mainly on the relative depth. The values of the relative intensities \( u' \bar{U} \) and \( w' \bar{U} \) decreased with increase in relative height above the sediment bed; the ratio \( w'/u' \) had a mean value of 0.6 ± 0.15 (similar results as for the Ouse data, see West et al. 1986). The correlation coefficient \( R_{uw} \) was dependent on both \( z/h \) and \( R_{l_g} \). \( R_{uw} \) showed a trend from 0.35 at \( z/h = 0.2 \) to 0.2 at \( z/h = 0.8 \) for relatively low values of \( R_{l_g} \) (< 0.3). The correlation coefficient \( R_{uw} \) was generally smaller than 0.2 for \( R_{l_g} > 0.3 \). The correlation coefficients \( R_{uw} = |\bar{sw}|/|sw'| \) and \( R_{su} = |\bar{s}u|/|su'| \) were found to decrease and increase with \( R_{l_g} \), respectively (in agreement with Komori et al., 1983). The ratio \(-\bar{sw}/|su|\) was approximately 0.5 for neutral conditions and decreases to 0.1 for \( R_{l_g} > 1 \).

Conditional sampling analysis (Lu & Willmarth, 1973) led to the conclusion that wave-like motions with \( w \) and \( s \) being \( \pi/2 \) out of phase, and \( u \) and \( s \) being \( \pi \) out of phase, may explain the simultaneous decrease in \( R_{uw} \) and increase in \( R_{su} \).

Vertical mixing-lengths decreased with increasing \( R_{l_g} \) and the solute mixing-length decreased more rapidly than the momentum mixing-length. This behaviour is in qualitative agreement with the expressions of Munk & Anderson (1948). Despite the decrease in solute mixing-length and subsequently in the eddy diffusivity, the flux \( \bar{sw} \) varied only slowly with increasing \( R_{l_g} \), for the observed range in \( R_{l_g} \), because \(-\partial \bar{s}/\partial z\) increased.

West and Oduyemi (1989) conducted turbulence measurements of sediment concentrations and velocities in the Cowney and Tamar estuaries. The main objectives of those measurements were to extend the previous work to include the turbulent fluctuations of suspended solids concentration and to elucidate the vertical turbulent flux of suspended solids.

The relative intensities \( u' \bar{U} \) and \( w' \bar{U} \) were reduced by buoyancy effects. The concentration fluctuation \( c \) was found to have more or less similar spectral characteristics as those of \( u, w, \) and \( s \) (West and Shiono, 1988). An approximate -5/3 power law was found to exist in the scalar spectra for \( c \) and -7/3 and -5/2 power laws were found in the co-spectra of \( cw \) and \( cu \), respectively (in agreement with Kaimal, 1973).
The variations in correlation coefficients $R_{uw}$, $R_{uw}$ and $R_{uw}$ with relative depth $z/h$ and $R_{ig}$, respectively, are summarised in Table 1.2. Table 1.2 shows that the correlation coefficient $R_{uw}$ is approximately constant for $R_{ig}$ up to 0.4 and then decreased with $R_{ig}$ for $R_{ig} < 0.4$. West and Shiono (1988) found that the correlation coefficient for streamwise turbulent transport of salt, $R_{uw}$ increased with $R_{ig}$. From this difference in behaviour between $R_{uw}$ and $R_{uw}$, it was concluded that contributions of wave-like motions to $\bar{uw}$ were less than those to $\bar{su}$.

Table 1.2. Summary of correlation coefficients (West and Oduyemi, 1989).

<table>
<thead>
<tr>
<th>$z/h$</th>
<th>$R_{uw}$</th>
<th>$R_{uw}$</th>
<th>$R_{uw}$</th>
<th>$R_{uw}$</th>
<th>$R_{uw}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.1 - 0.4</td>
<td>$R_{ig} &lt; 0.2$</td>
<td>0.5</td>
<td>0.1</td>
<td>0.25</td>
<td>0.0</td>
</tr>
<tr>
<td>0.4 - 0.7</td>
<td>$R_{ig} &lt; 0.2$</td>
<td>0.4</td>
<td>0.1</td>
<td>0.2</td>
<td>0.1</td>
</tr>
</tbody>
</table>

The analysis of effects of stratification presented by West and Oduyemi (1989) is as follows. It is suggested that stresses are unaffected by stratification during periods of higher flow velocities in a tidal cycle, because the bed shear stress effectively balances the longitudinal surface slope term in the one-dimensional depth averaged momentum equation. As the vertical density gradient increases, mixing-lengths decrease and the velocity gradients increase, and consequently, at some point in the water column, $\bar{U}$ increases for a given value of the bed shear stress. As a result the ratios $u/\bar{U}$, $w/\bar{U}$ may reduce while bottom shear stress and transport of momentum are virtually unaffected.

Except for the difference in behaviour between $R_{uw}$ and $R_{uw}$ with $R_{ig}$, effects of sediment-induced gradients on the turbulence properties were similar to those of salinity induced gradients. Correlation coefficients, $R_{uw}$, $R_{uw}$ and $R_{uw}$, and vertical mixing-lengths decreased with increasing $R_{ig}$. The auto- and co-spectra were found to behave more or less in a similar manner to those of atmospheric data in the presence of stable stratification.

1.3 Objectives, approach and scope

The objective of the present study is to increase the understanding of the role of turbulence in the transport of suspended sediment under tidal conditions with emphasis on the effects of suspended sediment-induced density gradients on the turbulence properties. Sediment-induced sediment gradients may reduce turbulent kinetic energy by buoyancy destruction. In turn a reduction in turbulent kinetic energy may reduce the
turbulent exchange of suspended sediments and thereby cause even larger sediment-induced density gradients and an increase in $Ri_s$.

The approach is to quantify the turbulence properties of the velocity and concentration fields in a tidal channel in relation to mean velocity gradients, suspended sediment-induced mean density gradients, settling velocities, erosion and deposition of sediment and sediment properties. To that end, in-situ high-frequency measurements were made of suspended sediment concentrations (SSCs) and velocities at three levels close to the sediment bed at one location in a tidal channel in the Ems/Dollard estuary during a few tidal cycles. Several quantities were determined from the measurements, such as the turbulent transports of sediment and momentum.

Simulations of SSCs and vertical transports of sediment and momentum were made with a 1DV mathematical model made available by WL Delft Hydraulics which accounts for sediment settling, deposition, erosion and turbulence damping by suspended sediment-induced density gradients. The simulation results were used for analysis and interpretation of the measurements. Various formulations for sediment settling and erosion were examined.

Laboratory experiments were carried out in order to assess the erosion characteristics of sediments from the research area. The sediment beds tested were either directly placed, or formed from deposition.

The scope of this research was partly determined by the circumstances encountered in the field during the relatively short periods of measurements. These circumstances concerned, for example, variations in water depth, wave activity, maximum flow velocities, strength of sediment bed, sediment type, etc. It is inherent to local measurements that the influence of upstream processes on the measurements is not always clear. The consequences of this difficulty for the interpretation of the measurements and the validity of the conclusions are discussed.

The outline of this thesis is as follows: Chapter 2 describes the geographical area of the field research; Chapter 3 describes the methodology of the field measurements used and Chapter 4 the results; in Chapter 5 the measurements are analysed through comparison with the 1DV mathematical model mentioned; Chapter 6 presents the results of the laboratory experiments and discusses their relevance for the interpretation of the field measurements, and in Chapter 7 the results of the previous chapters are integrated, final conclusions are drawn and recommendations are made for future research.
Chapter 2  The Ems/Dollard estuary: a physical system description

2.1 Introduction
The meso-tidal estuary of the River Ems is situated in the north-eastern part of The Netherlands (Figure 2.1) between the tidal limit at the weir, about 50 km upstream of Emden, and the seaward boundary, which is formed by the islands Rottumeroog and Borkum, and the stagnant waters behind these islands. It has been chosen as the area for investigation because it is characteristic of tidal inlets in the Netherlands and because extensive field surveys already have been carried out in the area. The transport of mud in this estuary has been investigated by Eisma (1986), De Jonge (1992) and Van Leusen (1994), so that much is known of the physical behaviour of this estuarine system.

The field work of the BOA theme-project was carried out at two locations: in the middle of the Dollard on the Heringsplaat and in the adjacent channel Groote Gat. This means that the BOA theme-project focused mainly on the temporal variability and less on spatial variability. Nevertheless a physical description of the estuary is required to be able to compare it with other estuaries and to get a thorough understanding of the scale and the impact of mud transport. Therefore in this chapter the estuary is characterised by describing its topography, the tidal behaviour, and the distribution of typical estuarine properties comprising salinity, the composition of the sediment and the tidal averaged suspended sediment concentration. Special attention is paid to the transport of fine sediments in the estuary. Appendix A presents a summary of the physical properties of the Ems/Dollard estuary.

2.2 Topography
In this section the topography and morphology of the Ems/Dollard estuary is described and a short historical overview is presented because much of the present day topography is the result of natural processes like flooding and siltation, and of human activities comprising land reclamation, dredging activities, diking and regulation works like the training wall “Geise dam” along the Emden Fahrwasser. The human
involvement is most dominant in the Dollard reach and the Emder Fahrwasser, the area of interest for our understanding of accumulation and transport of fine sediments.

**Present day morphology**
The Ems/Dollard shows typical properties of both an estuary and an intertidal basin. It has a more or less funnel shaped river mouth, but unlike most estuaries it has a relatively large brackish tidal basin called the Dollard reach. The latter is more typical for intertidal reaches, nevertheless the Ems/Dollard will be referred to as an estuary.

![Diagram of Ems/Dollard estuary](image)

Figure 2.1. The Ems/Dollard estuary and the BOA field site where two measuring platforms are located: (1) a measuring platform on the intertidal flat *Heringsslaat* and (2) the RWS pole 208 in the adjacent tidal channel *Groote Gat*. Sand flats in the inner part of the estuary can be covered by a layer of soft mud which can reach a thickness of few decimetres along the borders of the flats.
The estuary is about 110 km long when measured along the main tidal channel, it covers 470 km² and comprises roughly 50% intertidal flats. It can be divided into three reaches. The most seaward reach extends from the islands Rottumeroog and Borkum to Eemshaven. It has a total area of 215 km². In this reach the most notable aspect of the complex topography is the two-channel system divided by a series of shoals. The middle and more narrow reach extends from Eemshaven to Mond van de Dollard. The total area is 155 km² and only 35% consists of tidal flats. The Hondpaap is the most important tidal flat; it divides this reach into two channels. The uppermost reach extends from Mond van de Dollard to the mouth of the Eems, near Pogum. The total area is about 100 km² and 80% consists of intertidal flats. The morphology of the Dollard reach is very characteristic for tidal reaches but relatively simple. It consists of one main tidal channel, the Groote Gat, surrounded by tidal flats and creeks.

The composition of the intertidal flats is highly dynamic and complex. It can change locally on a typical time scale of one week from muddy to sandy. This process is affected by creeks and by bioturbation, through erosion and burrowing of the upper layer of sediment, respectively; both give the tidal flats a battered appearance.

A short overview of recent history
The Ems/Dollard estuary was formed during the Holocene with the river Ems more or less at the present position (Roeleveld, 1974). Due to a number of floods in the 14th and 15th centuries the Dollard was formed, which obtained its maximum size of 350 km² in about 1500. Land reclamation by diking began pretty much at the same time and new polders succeeded each other over a relatively short period. The accretion at the time was probably large because it was a custom to wait until the salt pasture was overgrown before dikes were built.

The Ems flowed initially as a meandering stream directly past the old town of Emden. However, its course was broken in 1509 by a storm surge and despite great efforts in 1585 to restore its old route the Ems kept its new position: Emden became difficult to reach for larger vessels. In order to prevent further siltation of the Ems, and to fix its position, the training wall “Geise Leitwerk” was constructed between 1872 and 1900: the Ems, or “Emder Fahrwasser”, became separated form the Dollard.

In the fifties plans were made for opening Emden harbour for large ore-tankers. A 12 kilometre long Geisedam was constructed 400 m north of the old “Leitwerk” in 1962, a 2.3 km long “Leitwerkes Seedeich” was constructed on the north side of the Emden Fahrwasser in the same period.
Accumulation of sediment in the Dollard reach

Recent research by means of different measuring techniques reveals various accumulation rates ascribed to either low accuracy of the measurements, or incorrect interpretation of the results.

Smit et al. (1960; in Wiggers, 1974) reconstructed the accumulation from about 1500 when the Dollard reached its largest expansion. They found an annual accretion rate of 1.7 cm for the polders reclaimed in the 17th century and 1.3 cm for polders reclaimed in the 19th century. The decrease in accretion rate was explained from the composition of the sediment bed, which showed that in the beginning of the accretion old clay layers contributed to the availability of mud besides the fine sediment from the adjacent North Sea.

An accretion of approximately 1 - 3 mm/year over the last century was found by means of a 210Pb-method and a pollen method (Heijns, 1987). For the 210Pb-method only six samples were used and for the pollen method only one sample. The results of these measurements should therefore be interpreted with care. On the other hand it strengthens the idea supported by some researchers, that the development of the morphology of the Dollard reach came to an end. They refer to the fact that nautical charts from the last 50 years show hardly any changes in the morphology of channels and intertidal flats (Abrahamse, 1976; Wiggers, 1974).

![Figure 2.2](image-url)

Figure 2.2. Development of the Dollard reach over the period 1952 - 1990: (a) cumulative contributions of areas at different levels; (b) The mean water depth, that is the “wet volume” (I) divided by the “wet surface” (O) at various water levels (0 = N.A.P. Dutch ordnance datum; Source: Lowijs, 1995).

Sounding measurements carried out over the last 40 years show a different insight into the development of the Dollard. The accretion rate of the intertidal flats were estimated at 1 cm/year during the period 1952 - 1985. Figure 2.2a shows that the area
at levels between 0.5 and 1.35 m N.A.P. increased over the past decades at the expense of the area between -1.0 m and 0.5 m N.A.P. Figure 2.2b shows the development of the mean water depth for various water levels: since 1952 the water depth increased for water levels between -2 m and 0.5 m N.A.P. because the tidal flow is confined to the deeper parts of the Dollard for water levels below 0.5 m N.A.P. Minimum water depth approximately coincides with the flooding of the tidal flats, the water level at which this happens increased from about N.A.P. +0.0 m in 1952 to N.A.P. +0.5 m in 1990. The area-elevation volume provided by the Dollard reach showed a decrease of approximately 10 % over the period 1960-1990 (GRAN, 1990).

2.3 Meteorology
The annual mean air temperature is 9 °C. Temperature fluctuations are regulated by the seasonally varying insolation and the moderating influence of the North Sea especially along the coastline: the air temperature along the coast is always a few degrees higher in winter and a few degrees lower in summer compared to the more inland situated reaches. The water temperature in the upper reach differs even more from the sea water temperature since the water depth is much smaller.

The annual precipitation is 675 mm near the barrier islands and 700 mm near Emden with a standard deviation of approximately 40 mm. Its excess over the evaporation is approximately 150 mm.

The annual mean wind direction is from the south westerly directions. The annual mean wind speed is 6.5 m/s near the barrier islands and 4.5 m/s in the Dollard reach. The strongest winds are mostly from south westerly direction in winter, and the weakest are north-east to south-east mostly during summer (Climate atlas of the Netherlands, 1972).

2.4 Hydrodynamics and fresh water input
As for most Dutch estuaries, the tidal range and the tidal asymmetry in the Ems/Dollard increase in the landward direction. The latter phenomenon is an important property of an estuary for the direction of the sediment transport (Dronkers, 1986). In this chapter the tidal behaviour, the fresh water input at various locations and the salinity distribution are presented, and the tidal mixing is discussed.

Tide
The mean tidal range increases from 2.2 m near Borkum via 3.0 m at Delfzijl to 3.3 m in the Dollard reach. Very prominent for this estuary is the diurnal inequality which can
reach 0.3 m at Delfzijl. The distortion of the tide as it propagates from the North Sea into the Ems/Dollard can be modelled by a super-position of $M_2$ (semi-diurnal lunar) and $M_4$, the first harmonic (Friedrichs and Aubrey, 1988):

$$h(t) = h_{M2} \cos(\omega t - \theta_{M2}) + h_{M4} \cos(2\omega t - \theta_{M4})$$  \hspace{1cm} (2.1)

where $h$ is the water level, $t$ is the time, $\omega$ is the tidal frequency, $h_{M2}$ is the amplitude and $\theta_{M2}$ is the phase. The phase of $M_4$ relative to $M_2$ is defined as:

$$2\theta_{M2} - \theta_{M4}$$  \hspace{1cm} (2.2)

A direct measure of the non-linear distortion is the ratio of amplitudes:

$$h_{M4}/h_{M2}$$  \hspace{1cm} (2.3)

Non-linear friction results in greater frictional dampening in shallow water, slowing the water levels changes around low tide relative to high tide. The result inside the estuary is a longer ebb and a shorter flood with the highest velocities occurring during flood. (Dronkers, 1986; Friedrichs and Aubrey, 1988).

In the Ems estuary $h_{M4}/h_{M2}$ is about 0.05 in the outer reach and 0.1 near Emden, thus showing the increasing importance of non-linear distortion in the shallower inner part of the estuary. $2\theta_{M2} - \theta_{M4}$ is about 160° in the outer reach, 180° in the Emden Fahrwasser, and decreases upstream of Pogum (Van Leussen, 1994). These values for $2\theta_{M2} - \theta_{M4}$ imply that the duration of the flood is shorter than the duration of ebb in the outer reach and upstream of Pogum, and that in the Emden Fahrwasser the duration of ebb and flood are approximately similar, as shown in Table 2.1.

<table>
<thead>
<tr>
<th>Location</th>
<th>Mean tide (m)</th>
<th>Spring tide (m)</th>
<th>Duration (h:min)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>HW</td>
<td>LW</td>
<td>Tidal range</td>
</tr>
<tr>
<td>Delfzijl</td>
<td>1.3</td>
<td>-1.7</td>
<td>3.0</td>
</tr>
<tr>
<td>Knock</td>
<td>1.3</td>
<td>-1.7</td>
<td>3.0</td>
</tr>
<tr>
<td>Emden</td>
<td>1.4</td>
<td>-1.8</td>
<td>3.2</td>
</tr>
<tr>
<td>Pogum</td>
<td>1.5</td>
<td>-1.8</td>
<td>3.3</td>
</tr>
<tr>
<td>Nw.St.zijl</td>
<td>1.4</td>
<td>-1.9</td>
<td>3.3</td>
</tr>
</tbody>
</table>

Van Leussen (1994) and Van de Kreeke et al. (1997) found that peak velocities during flood tide are larger than those during ebb tide, and therefore, generally
speaking, the Ems/Dollard estuary can be addressed as flood dominant. However, the Dollard reach forms an exception. It is shown in Chapter 4 that peak velocities during ebb were larger than those during flood during June and August 1996. The ebb dominance in the Dollard reach might be explained from the phase lag between the incoming tidal wave and the reflected tidal wave. This may lead to lower water levels on average and subsequently higher current velocities during ebb (Ridderinkhof, 1998, personal communication).

**Annual mean water levels and tidal ranges**

The annual mean water level and the annual mean tidal range were analysed by Bossinade et al. (1993). The annual mean water level and the annual mean tidal range on the seaward boundary increased 0.2 cm/year, averaged over the period 1931-1990. This trend is in agreement with the increase in the annual mean high water level of 0.25 cm/year for the Wadden-sea over the same period. Differences up to about 5 cm between succeeding years are largely explained by differences in wind strength and wind direction. In the inner part of the estuary the tidal range increased according to the increase at the seaward boundary, although since about 1960 the tidal range in the estuary increased even more (GRAN, 1990). Additional to this observation, the flood reaches Delfzijl 8 minutes earlier, and Nieuwe Statenzijl 15 minutes earlier, when compared to the sixties (GRAN, 1990). These findings point to a reduced drag on the tidal wave caused by either natural or artificial changes in the morphology of the estuary. They are in agreement with the increase in mean water depth over the last decades mentioned in Section 2.2 (see Figure 2.2).

**Wave climate**

The wave climate is dominated by locally generated wind waves. Only in the outer reaches wave penetration from the adjacent North Sea is present. A fetch of a few kilometres in combination with relatively small mean water depths results in small wave heights in the range of 0.1 to 0.5 m. The impact of waves on the sediment beds of intertidal flats, on the other hand, is substantial when compared to that of the tidal currents (Dronkers, 1986).

The wave growth is mainly limited through white capping, which is often observed for wind speeds exceeding 4 Beaufort. Surf zones on the borders of tidal flats are absent in the inner part of the estuary. This can be explained by the considerable reduction of the fetch when tidal flats become exposed.
Fresh water discharge
The main fresh water input is coming from the Ems river. The Ems is a rain fed river, its discharge shows a significant variation over the seasons: a high peak in winter, a relatively low discharge in the remaining seasons (see Figure 2.3). The discharge in the winter of 1996 was exceptionally low.

![Discharge Graph](image)

Figure 2.3. Discharge at the weir 50 km upstream of Emden (Source: Wasser- und Schiffahrtsamt Emden).

More rivers and channels are entering the Ems/Dollard estuary which are altogether an important contribution to the total fresh water input. The distribution of the total mean annual fresh water discharge, about 150 m³/s, can be estimated as follows: 80% enters the estuary via Emdr Fahrwasser, 10% via Groote Gat, and 10% via Delfzijl harbour and other locations (GRAN, 1990).

Longitudinal salinity distribution
The longitudinal salinity distribution shows a gradual decrease from about 30% near the barrier island Borkum to the point of maximum salt intrusion near the town of Leer, about 20 km upstream of Emden. The salinity is a good indicator for typical estuarine processes such as flocculation of fine sediment particles. Van Leussen (1994) measured the salinity distribution along the axis of the estuary during June 1991. He found that the salinity distribution shifted seaward over a length of approximately 15 km as a result of an increase of the river discharge from about 25 m³/s to 60 m³/s. The maximum longitudinal salinity gradient is approximately 1×10⁻³ ppt/m (≈ 0.8×10⁻³ kg/m³/m). The entire salinity distribution shifted under the influence of the tide. In the greater part of the Dollard the surface salinity gradient is small (of the order of 0.1
ppt/m, Helder and Ruardij, 1982), and gravitation circulation is considered of minor
importance.

**Tidal mixing**

An estuary can be classified as stratified, partially mixed or well mixed depending on
the vertical differences in salinity. A measure for the mixing in an estuary is provided
by the Estuary Richardson number \( Ri_E \) (Fisher, 1979):

\[
Ri_E = \frac{\Delta \rho gh Q_f}{\rho A u_t^3}
\]  

(2.4)

where \( \Delta \rho \) is the density difference between the fresh water and the sea water, \( g \) is the
acceleration of gravity number, \( h \) is a typical water depth, \( Q_f \) is the fresh water
discharge, \( \rho \) is the density of the (sea) water, \( A \) is the cross sectional area at the mouth
and \( u_t \) is the root mean square velocity in the mouth. The numerator is a measure for
the power that is needed to mix the fresh discharge over the whole water column and
the denominator is a measure for the available power. For partially mixed estuaries \( Ri_E \)
is between 0.08 and 0.8.

The values for \( Ri_E \) for parts of the estuary, presented in Table 2.2, show that even
at high river discharge only the Ems-Dollard estuary is partly stratified. These rough estimations are in agreement
with various measurements of vertical salinity distributions showing only small variations in salinity except for the Ems-Dollard estuary at the turbidity maximum and in the Southern part of the Dollard (Dorrestein, 1960; Helder and Ruardij, 1982; Van
Leussen, 1994).

<table>
<thead>
<tr>
<th>Location</th>
<th>( h ) (m)</th>
<th>( A ) (m²)</th>
<th>( u_t ) (m/s)</th>
<th>( Q_{f,low} ) (m³/s)</th>
<th>( Q_{f,high} ) (m³/s)</th>
<th>( \Delta \rho / \rho )</th>
<th>( Ri_{E,low} )</th>
<th>( Ri_{E,high} )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mond v.d. Dol.</td>
<td>10</td>
<td>11000</td>
<td>1</td>
<td>30</td>
<td>300</td>
<td>0.02</td>
<td>0.005</td>
<td>0.05</td>
</tr>
<tr>
<td>Emden Fahrw.</td>
<td>10</td>
<td>5100</td>
<td>0.8</td>
<td>30</td>
<td>300</td>
<td>0.02</td>
<td>0.023</td>
<td>0.23</td>
</tr>
<tr>
<td>Turbidity max.</td>
<td>5</td>
<td>2900</td>
<td>0.8</td>
<td>30</td>
<td>300</td>
<td>0.02</td>
<td>0.010</td>
<td>0.10</td>
</tr>
<tr>
<td>Groote Gat</td>
<td>10</td>
<td>2700</td>
<td>0.6</td>
<td>5</td>
<td>40</td>
<td>0.01</td>
<td>0.008</td>
<td>0.07</td>
</tr>
</tbody>
</table>

**2.5 Fine sediment characteristics**

Sediment is characterised in the first place by its grain size: sand (> 63 \( \mu \)m), silt (3 \( \mu \)m
- 63 \( \mu \)m), and clay (< 3 \( \mu \)m). The sand and silt consist mainly of quartz, feldspar and
mica, the clay minerals are mainly illite, kaolinite and montmorillonite. The cohesive properties, such as “true cohesion” and flocculation ability, are basically determined by the mineralogical composition and the organic coatings of the particles.

The distribution of sand and mud is shown in Figure 2.1. In the upper reaches most flats contain large amounts of mud, except for the northern part of the Heringsplaat and the Maanplaat. In the middle reaches the flats are partly covered with a mixture of sand and mud, and in the seaward reaches only small amounts of mud are found. Tidal channels with a muddy sediment bed are the Ender Fahrwasser, Groote Gat and the Bocht van Watum (Van Heuvel, 1992). On flats along the German shore and in the tidal channel Bocht van Watum the percentage of mud is probably increased by the dumping of dredging material from the Ender Fahrwasser.

The sand/mud distribution is continuously changing during the year as a result of amongst others storm events, variable organic activity and dredging activities. The overall transport of the mud and thereby its distribution in the Ems/Dollard over the different seasons is not completely understood. Generally speaking, more mud is found on the tidal flats in summer than in winter. It is not known if the mud is actually leaving the estuary in winter.

**Physicochemical properties**

A summary of the most important physicochemical properties related to the transport of cohesive sediment was determined in the framework of the EC Mast-I Research Programme (see Table 2.3; EC MAST-I, 1993). It will be used here as a guideline for presenting the characteristics of the sediments in the Ems/Dollard estuary.

**Table 2.3 Physicochemical properties as discussed in the EC MAST-I report (1993)**

<table>
<thead>
<tr>
<th>1. Physicochemical properties of the over flowing fluid</th>
<th>3. Water bed exchange processes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Chlorinity, Oxygen content, pH, Sodium Adsorption Ratio (SAR), Temperature, Redox potential, ion-composition</td>
<td>Critical shear stresses for erosion and deposition, erosion rates, settling velocity, rheological properties of the mud</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>2. Physicochemical properties of the mud</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>See 1, plus: density, grain size distr., Specific surface area (SSA), Cation Exchange Capacity (CEC), Organic content</td>
<td></td>
</tr>
</tbody>
</table>

Physicochemical properties of suspended sediment and the water have been measured at various locations along the estuarine axis of the Ems/Dollard estuary (Eisma, 1986; Van Leussen, 1994). Van Leussen (1994) showed that some of these,
such as the organic content, pH, Redox Potential and the Oxygen content, are related to biological processes, which are typically season dependent. For a comprehensive description of the physicochemical and biological processes related to these properties the reader is referred to Van Leusen (1994) or Van der Lee (in prep.). Some results obtained from an extensive field survey in June 1991 (Van Leusen, 1994) are summarised herein.

The Sodium Adsorption Ratio (SAR) is proportional to the ratio of the exchangeable cations of low valency and high valancy, when equilibrium is assumed. A high content of high valancy cations, or a low SAR, reduces the thickness of the diffusive double layer surrounding the clay particles and thereby reduces the repulsive forces between them. The SAR distribution found in June 1991 confirms the general findings of high SAR values (about 160) in the salinity region and low SAR values (about 5) in the fresh water reach of the estuary.

The cation exchange capacity (CEC) is a measure of the maximum number of cations that can be attracted. Its distribution showed an increasing trend in the landward direction: from 20 meq/100 g in the seaward reach to 50 meq/100 g in the river Ems. The specific surface area (SSA) is closely related to the CEC value and a clear indication for the particle surface available for adsorptive processes. Values for the SSA for the suspended fine sediment ranged from 100 m²/g in the seaward reach to about 200 m²/g in the fresh water reach of the estuary. The increase in the SSA was in agreement with the increase in the CEC, and with the increasing amount of clay incorporated in the suspended sediment.

The repulsive force between small particles/microflocs is determined by the surface charge which is reflected by the Electrophoretic Mobility (EM). The EM seems to be a good indicator of the flocculation ability. It is determined by the organic coatings of the particles and the availability of cations (high CEC, low SAR). The longitudinal distribution of EM shows an increase in magnitude in landward direction which corresponds well the decrease in settling velocity and floc size, and with results from other estuaries such as the Gironde and the Loire estuaries (Van Leusen, 1994; see also Hunter and Liss, 1982).

The mineralogical composition of sediment samples taken at various positions in the Ems/Dollard is presented in Table 2.4. The compositions of the samples from Delfzijl harbour and from Eems harbour hold for the complete sediment sample, whereas the remaining ones are composition analyses of the clay fraction. The samples show that illite and smectite form the largest fraction of the clay minerals. In the fluvialite suspended sediment from the River Ems kaolinite is absent, which has
implications for the origin of the Ems/Dollard mud (see section 2.6). The organic content of the mud from Delfzijl is small compared to the organic content of the mud from Eems harbour, but probably shows a strong seasonal variation.

Table 2.4 Composition of clay minerals and quartz (percentages by weight)

<table>
<thead>
<tr>
<th>Location</th>
<th>Illite + smectite</th>
<th>kaolinite</th>
<th>montmor.</th>
<th>Quartz</th>
<th>organic</th>
</tr>
</thead>
<tbody>
<tr>
<td>Favejee, 1960 (&lt; 0.5 μm):</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>West Dollard</td>
<td>80</td>
<td>5 - 10</td>
<td>5 - 10</td>
<td>ca. 5</td>
<td></td>
</tr>
<tr>
<td>Ems mouth</td>
<td>80</td>
<td>5 - 10</td>
<td>5 - 10</td>
<td>ca. 5</td>
<td></td>
</tr>
<tr>
<td>Ems low tide banks</td>
<td>80 - 90</td>
<td>ca. 5</td>
<td>ca. 5</td>
<td>ca. 5</td>
<td></td>
</tr>
<tr>
<td>Fluvial</td>
<td>90 - 100</td>
<td>-</td>
<td>&lt; 3</td>
<td>&lt; 1</td>
<td></td>
</tr>
<tr>
<td>Marine</td>
<td>80</td>
<td>5 - 10</td>
<td>5 - 10</td>
<td>ca. 5</td>
<td></td>
</tr>
<tr>
<td>Winterwerp et al., 1992:</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Delfzijl</td>
<td>10</td>
<td>&lt; 2</td>
<td>-</td>
<td>55</td>
<td>2</td>
</tr>
<tr>
<td>Eems Harbour</td>
<td>20</td>
<td>&lt; 10</td>
<td>-</td>
<td>35</td>
<td>9</td>
</tr>
<tr>
<td>Van Leussen, 1994:</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Seaward reach (Borkum)</td>
<td>80</td>
<td>20</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>River Ems</td>
<td>85</td>
<td>15</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Water-bed exchange processes**

The distribution of the settling velocities is related to the sediment concentration, the turbulence structure of the flow, and the physicochemical surface properties of the sediment, which are in turn influenced by salinity, dissolved organic substances and the origin of the water/sediment sample. For a comprehensive description of settling velocities along the axis of the Ems/Dollard estuary the reader is referred to Van Leussen (1994), for settling velocities in the Dollard the reader is referred to Van der Lee (in prep.). A short summary will be presented herein.

From laboratory experiments it was found that the flocculation ability of suspended sediment shows an increase in the seaward direction, a minimum value at the point of maximum salt intrusion and a small increase in the fresh water reach of the estuary (Van Leussen, 1994). This trend in flocculation ability corresponds well with the trend in EM. It is suggested that in the high salinity reach of the estuary the sediment sub-fraction smaller than 25 μm is flocculated, forming microflocs. These microflocs are linked together by polysaccharides, for example, and form macroflocs with high settling velocities (Van Leussen, 1994).
The general increase in flocculation ability with increasing salinity is in agreement with the increase in the macrofloc sizes measured with an underwater video system “VIS” (Van Leussen, 1994) and with results reported by Eisma et al. (1991). The drop in floc size in the low saline reach of the estuary can be explained from the dissolving of muco-polysaccharides (Eisma, 1986). VIS measurements in the Dollard showed a strong variation in floc size and settling velocities (Van der Lee, in prep.). In the smallest measured size classes, around 100 μm, settling velocities varied between 0 and 5 mm/s. In the larger size classes the settling velocities converge to a smaller range approximately between 2 and 4 mm/s. Floc densities were calculated from the measured floc sizes and settling velocities by means of Stokes’s settling formula (Van Rijn, 1993). They show a strong variation in the smaller size classes, from highly organic to almost sand grain densities, whereas the densities of the larger flocs were all very low. Maximum settling velocities were found during periods of maximum sediment concentration just after maximum flow velocity (see also Van Leussen, 1994). It is mentioned herein that these results apply to the upper regions of the water column. Other studies have shown that in the region close to the sediment bed floc break-up by turbulence might result in a reduction in settling velocity at high flow velocities (Van der Lee, 1998).

Critical shear stresses for erosion were measured by means of a field flume on the intertidal flat Heringsplaat in the Ems/Dollard estuary, in 1996 and in 1997 (Kornman and De Deckere, 1998). The results showed that in 1996 the critical shear stress for erosion changed from approximately 0.2 N/m² in March via 0.5 N/m² in April and May to finally 0.1 N/m² in June 1996. These variations could be explained from biological activity on the intertidal flats: sediment stability increased with the increase in diatoms and carbonhydrates in April 1996, whereas the stability diminished as a result of bioturbation and grazing by Corophium volutator at the beginning of June 1996. In 1997 little variation in biological parameters was determined, and consequently, little variations in critical shear stress were observed (Termaat, in prep.).

2.6 Sediment transport
The tide-induced sediment transport in the Ems/Dollard is determined by the hydrodynamic conditions, comprising tidal asymmetry, current strength and wave climate, the settling behaviour of the sediment and the amount of sediment available for transport (Ridderinkhof, 1997). Availability of fine sediment for transport is, amongst other things, determined by biological activity and consolidation. The impact of dredging activities on the total sediment transport is substantial, as will be shown. In
order to obtain more insight into the net transport direction of the coarse sediment fraction in the Ems/Dollard estuary, some results of the McLaren method (see below) and the source(s) of the Ems/Dollard sediments are discussed.

**Siltation rates, dredging and dumping**

The turbidity maximum is situated in the Emder Fahrwasser and is the area of largest siltation rates. According to Rokosch (1995) the major part of the sediment that is entering the Emder Fahrwasser, about 7000 ton dry material per tide, is deposited and has to be removed by dredging. From 1950 until 1965 the amount of dredged material was approximately $1.6 \times 10^6$ m$^3$. From 1965, when the Emder Fahrwasser was deepened by 1 m on average, the amount of dredged material became approximately $5 \times 10^6$ m$^3$ per year (GRAN, 1990).

From 1954 until 1990 about $4 \times 10^6$ m$^3$ material with unknown density was yearly dredged from Emden harbour and pumped into upland disposal areas. These large quantities were unacceptable for both economical as well as environmental reasons. The solution for this problem was to allow for a thick layer of fluid mud, with a density of about 1150 kg/m$^3$, in the outer harbour, thereby creating a natural barrier against new sedimentation. The annual mean quantities pumped ashore reduced to about $1.4 \times 10^6$ m$^3$ (Rokosch, 1995).

Other locations of dredging activities are the region between the Oude Wester Eems and Doeke Gat, Delfzijl harbour and Eems harbour. The material is dumped in Mond van de Dollard, the tidal channel Bocht van Wattum, and other locations. The amount of material that is dredged from Delfzijl harbour and dumped in the Mond van the Dollard ranges from $0.3 \times 10^6$ m$^3$ to $1.5 \times 10^6$ m$^3$ per year.

If the annual mean dredged volumes were spread out over the intertidal flats in Dollard reach they would make a layer of mud of approximately 0.1 m thick. Dredging and dumping activities should therefore be considered when modelling the morphological development of the estuary.

**Suspended sediment concentration**

The suspended sediment concentration (SSC) increases towards the turbidity maximum near Pogum and then decreases further upstream. The sharp density interface often observed in the turbidity maximum points to the presence of a layer of fluid mud. The sediment fluxes are therefore relatively high in this region: tens of tons of dry material per meter channel width in both directions (Van Leussen, 1994).
Figure 2.4. Time series of 10 minute averaged values of water levels, wind velocities, and depth averaged SSCs above the intertidal flat Heringsplaat and in the channel Groote Gat during two spring-neap tidal cycles from September 8 until October 6, 1996.

Figure 2.4 shows some results of field measurements of water levels, wind velocities and depth averaged SSC above the intertidal flat Heringsplaat and in the adjacent tidal channel Groote Gat (See also Ridderinkhof et al., in prep.). Suspended sediment concentrations are generally in the range of 0.1 - 1 g/l during periods of maximum flow velocity. The highest SSCs are found during spring tide. Effects of high wind velocities (> 10 m/s) on the SSCs are most pronounced above the tidal flat. Not shown is the influence of biology on the sediment bed stability. It was found that during May 1996 the SSCs reduced to values well below 0.1 g/l, which could only be explained from biological activity (Kornman and De Deckere, 1998).

The net flux of sediment is strongly influenced by the tidal asymmetry. It can readily be derived that the tide-induced bed load transport is in the direction of the strongest bottom currents, which are possibly enhanced by gravitational circulation in
the Emder Fahrwasser. The strongest bottom currents and thereby the bed load transport are directed upstream for most parts of the flood dominant Ems/Dollard estuary. The net flux of suspended sediment is influenced by the hysteresis in the (Lagrangian) SSC - velocity curve (Groen, 1967). Ridderinkhof (1998) showed, by means of a simplified numerical model of the Dollard, that during calm weather conditions a net flux of suspended sediment in the landward direction may exist of up to about 50% of the total flux. A strong reduction in the net flux can be anticipated when wave activity on tidal flats inhibits sediment settling. Gravitational circulation was not considered in Ridderinkhof’s study since it was assumed to play a minor role in the Dollard reach (see also Sections 2.4 and 5.3).

McLaren Method
The McLaren method is based on trend analyses of sediment particle sizes. It is assumed that the strongest correlation between the sediment size histograms coincides with a transport path. The sediment transport directions are found from the skewness of the histograms (Van Heuvel, 1992). The transport paths for sand all show net accretion which means that the estuary is filling up with sand (Van Heuvel, 1992).

The usefulness of the McLaren method for determining transport paths and directions of the fine sediment fraction is disputable. Field measurements in the seaward and middle reach of the Ems/Dollard have shown that little variations in size histograms of the fine sediment fraction are present (e.g. Favejee, 1960; Van Leussen, 1994). This is thought to be caused by flocculation. Fine sediments stick together to form microflocs which remain intact during transport (see also Section 2.4).

Fine sediment sources
Favejee (1960) showed that the size histograms of the fine sediments from the Ems/Dollard are comparable with that of marine sediments. These fine sediments contain 5 to 10% kaolinite, which mineral is fully absent in the riverborne material (see Table 2.4). These findings point to a marine origin of the major part of the fine sediments. The large amount of marine sediments (> 70%) on the low tide banks of the River Ems could originate from fossil marine sediment layers eroded by the Ems. The conclusion is therefore that the fine sediments in the Ems/Dollard estuary originate either from recent marine deposits or from eroded fossil marine sediment layers.
Chapter 3 Description of the field measurements

3.1 Introduction
Concentrations of suspended sediment and water velocities were measured for periods of several tides in 1995 and in 1996 in the tidal channel “Groote Gat” in order to study their interaction (Figure 3.1). The tides were selected from spring and neap tides of different seasons to investigate variations in flow velocities and SSCs (see also Section 2.6). An experimental set-up was developed for this purpose in collaboration with Rijkswaterstaat Meetdienst Noord and Utrecht University. It consists of a pole RWS208 with sets of high-frequency flow meters and high-frequency fibre optical turbidity sensors attached at three levels above the sediment bed.

Figure 3.1. The Ems/Dollard Estuary and the measuring pole RWS208 equipped with a rigid frame (to the right) for turbulence measurements.
The fibre optical turbidity sensors were newly developed by W.L. Delft Hydraulics, and they were tested and adjusted in collaboration with this institute. The main advantage of this experimental set-up is that turbulence properties, the local vertical transport of sediment and momentum for instance, can be measured directly instead of deducing them from the averaged velocity and concentration profiles.

Section 3.2 describes the measuring facilities and the instruments used in the field and Section 3.3 explains the methodology of the field measurements, which includes the testing of the fibre optical turbidity sensors and the measuring technique in 1995. On the basis of the experience gained from these tests, a field program for 1996 was developed, which is presented in Section 3.4. Section 3.5 presents the quality assessment of the time series of velocities and SSCs obtained in 1996.

3.2 Measuring facilities and instrumentation
Two measuring frames were attached to the measuring pole RWS208: a Rijkswaterstaat frame for long term turbidity and velocity measurements, and a rigid frame for turbulence measurements. These frames are shown in Figure 3.1 to the left and to the right of the pole, respectively. Pole RWS208 was located in a straight reach of the channel approximately 30 m to the east of the mean low water level (about -1.7 m N.A.P.) at 7°09'43''E 53°17'14''N in 1995 and 7°09'43''E 53°17'15''N in 1996. The average bottom elevation was 3.3 m below N.A.P. and the channel width 600 m. Visual observations of the borders of the Heringsplaat during low low water spring tide showed that the bed surface was very smooth. The bed level showed only small variations of typically 0.05 m over a 10 m distance. The slope of the bank perpendicular to the channel axis was approximately 1:30. Sediment samples taken from positions directly adjacent to the measuring pole showed that the channel bed was composed of silt and clay.

Electromagnetic flow meters (EMFs)
The EMFs used in this study are discoid twin-axis type electromagnetic flow meters manufactured by W.L. Delft Hydraulics. They are coded as D232, D233 and D334, have 5 cm diameter sensing heads and operate with a cut-off frequency of approximately 7 Hz.

This type of EMF has been found to be suitable for measuring two normal components of the velocity fluctuations under field conditions (Soulsby, 1980; West et al., 1986; French and Clifford, 1992). One of the limitations of the instruments is their spatial resolution. Soulsby (1980) found severe attenuation of measurements of vertical
velocity intensities due to sensor averaging under typical field conditions, whereas measurements of horizontal intensities and of the Reynolds stress in the vertical plane in flow direction were relatively unaffected. This resulted in too low values of the ratio \( w/u_* \), where \( u_* \) is the friction velocity.

The EMFs D232-D234 were calibrated in April 1994 by \( \text{WL} \) | Delft Hydraulics and in February 1996 by the Laboratory of Hydromechanics of Delft University of Technology. The linear responses (\( \approx 1.0 \text{V/(m/s)} \)) measured in 1996 did not differ significantly from the original responses measured in 1994 except for the D333 X-channel, the response of which was found to be 1.5% higher. A possible explanation for this inconsistency could be the relatively large scatter in the original towing tank data of the D333 X-channel. The linear responses of the 1996 calibration were used for the data-processing. The maximum "root mean square values" or noise levels measured in still water were approximately 0.004 m/s.

The offsets of the EMFs were measured in the laboratory in a large tank, and in-situ in advance of each measuring period. An appreciable difference of 0.05±0.02 m/s was found between the offsets measured in February 1996 in the Laboratory and in April 1996 in-situ. This difference was most likely due to bias errors in the laboratory estimates resulting from electrical or magnetic fields which presumably were not present in the field. From April to August 1996, long term changes in measured offsets in-situ were found to be 0.02±0.01 m/s at most. It was therefore decided to use the in-situ measured offsets, instead of the manufacturer's offsets, for data-processing.

**Fibre optical turbidity meters (FOSLIMs)**

Reports on high-frequency recordings of SSC are scarce and are mostly based on acoustic back scatter techniques (Thorne et al., 1996), or on back scatter of infra-red light (Kawanisi and Yokosi, 1993). West and Oduyemi (1989) and Darbyshire and West (1993) do not mention the principle of their measuring technique, nor the limitations or advantages.

FOSLIMs use the principle of light attenuation for measuring SSCs. They were applied in this research because sediment flocs composed of clay, silt and biological components absorb light instead of reflecting it, contrary to sand particles, for example, and therefore back scatter techniques are less useful.

The FOSLIMs have been manufactured by \( \text{WL} \) | Delft Hydraulics. A sketch of the original sensor head of the FOSLIM is shown in Figure 3.2a. The sensor consists of two glass fibres mounted on a rigid rod in such a way that both alignment of the fibres and the distance between the fibres can be altered. One of the fibres is connected to a
light emitting diode and sends an infra-red light beam through the measuring volume (see Figure 3.2). The fibre opposite to the transmitted light emitting fibre collects the light and passes it on to a photo diode where the light intensity received is measured. The difference between the emitted and received infra-red light intensities is a measure for the water turbidity, which is related to the SSC through in-situ calibration (see Section 3.3.2). A daylight filter prevents influences of daylight on the turbidity measurement. The FOSLIM output voltage is approximately linearly related to the concentration of fine sediments.

![Figure 3.2. The original sensor head (a) used until April 1996, and the adapted sensor head (b) used after April 1996.](image)

A prototype FOSLIM was tested in the early nineties by M. Christie and K. Dyer of Plymouth University. It was found not suitable for field measurements (personal communication). Long wiggling fibres presumably caused losses of light in the fibres which were incorrectly interpreted as turbidity fluctuations.

Different prototypes were purchased in July 1995. They had short fibres which were tightened thoroughly when attached to the measuring frame. However, this design also proved to be not suitable for field measurements: during the April 1996 measurements the fibre heads metal coating corroded, which allowed the fibres to twist. Slight changes in the positions of the fibre heads alter the characteristics of the instrument completely. From then on we deft hydraulics fixed the fibre heads with a synthetic resin at a separating distance of approximately 6 mm (see Figure 3.2b).

The adapted FOSLIMs were tested in the laboratory in June 1996. The measuring range was limited, as a result of fixing the fibre heads, to approximately 0-30 g/l for mud from the Heringsplaat. The upper range, 20-30 g/l, showed increased noise levels and was therefore removed. This was done by increasing the amplification in such a way that the maximum output signal of 10 Volt was obtained at 20 g/l. The noise level
measured in clear water did not exceed $0.8 \times 10^{-4}$ V. The offset drift measured over a period of 90 hours in still tap water was approximately 0.03 g/l per hour. The accuracy of the SSC measurements is mainly determined by the calibration procedure.

Under field conditions it is quite possible that the fibre heads remain clear from fouling in periods of relatively high water velocities, but the heads may foul during low and high water slack. It was found that seaweed can easily cover the small fibre heads (see also Section 3.5). Therefore, FOSLIMs are calibrated in-situ since fouling and suspended sediment characteristics depend on the in-situ conditions. The in-situ calibration procedure is described in Section 3.3.2.

Additional instrumentation
The Rijkswaterstaat Pole 208 is equipped with three turbidity meters of the MEX-type (w | delft hydraulics, MEX-3 RD-10/5 sensor) and three spherical EMFs fixed at 0.3 m, 0.7 m and 1.4 m (1.0 m in 1995) above the bed. The frame is shown in Figure 3.1 to the left of the pole.

The MEXs are used in this study for long-term turbidity measurements. They provide fouling correction through a two-way measuring system. The light attenuation is measured over two different path ways and the values obtained are subtracted, which compensates for contributions of the fouling (Van Rijn, 1993). Furthermore the relatively large measuring volume and the sensor shape make the MEX less sensitive to fouling, when compared to the FOSLIM. Therefore the MEX is considered as a robust measuring device for low-frequency turbidity measurements. A data-logger calculates and stores the 10 minute averages and standard deviations of the MEX and EMF signals.

The water level was recorded at the Rijkswaterstaat stations “Dollard Noord” and “Skansker Diep”, 2 km north and 1 km south of Pole 208, respectively. The wind velocity was obtained from a measuring platform (BOA Measuring Bridge) located on the Heringsplaat (+0.2 m N.A.P.), 400 m south west of Pole 208. Turbidity, water velocity, salinity, water level and the temperature were also recorded.

3.3 Methodology
3.3.1 Data collection
Three combinations of EMFs and FOSLIMs were fixed at approximately 0.1 m, 0.4 m and 1.0 m above the channel bed, the distance between a FOSLIM and EMF fixed at the same level being approximately 0.05 m (see Figure 3.3). The small separation distance between a FOSLIM and an EMF allowed both velocity and turbidity
fluctuations to be measured in approximately the same “measuring volume”. A consequence of this set-up was that the velocity measurements were hindered by the turbidity sensor heads during flood and vice versa during ebb for the smaller scales of turbulence (< ~0.1 m). If this mutual hindrance had a significant effect on the measurements, it would have resulted in differences in the turbulence parameters during ebb and flood. The largest differences would occur at the lowest sensor position where small scales become important. In Section 4.4 and Appendix B it is shown that noise contributions to the velocity signals were independent of flow velocities and flow direction and appeared to be most pronounced at the highest level of 1.0 m. These noise contributions could therefore not be linked to the presence of the FOSLIM sensor head. The relatively large Reynolds stresses measured during flood in June and August 1996 at the 0.1 m level is explained from the presence of a scour hole which is further discussed in Section 4.4.

Figure 3.3. Sketch of the rigid frame for turbulence measurements with the combinations of EMFs and FOSLIMs attached at three heights above the bottom.

The measuring frame was lowered along the measuring pole during low water (LW). The sensors were aligned visually with the direction of the flow. The signals were digitized, sampled and stored in files on a PC by the data-log system DASY-Lab (DASYTEC GmbH, Mönchengladbach, Germany). DASY-Lab supports a “sample and hold” system which avoids time lags between the signals. The sampling frequency was set at 20 Hz, which was about twice the cut-off frequency of the instruments. After each measurement the files were saved on computer-tapes which were transported to Delft University of Technology.
The offsets of the EMFs were measured in-situ during LW in advance of each measurement (see also Section 3.2). The measuring frame was lifted and a bucket was placed over an EMF transducer head and fastened to the rigid frame so as to minimize water flow along the transducer. Then the frame was lowered into the channel and rotated in such a way that the flow ran into the bottom of the bucket. The signals of the X and Y channels of the EMF were sampled for a period of approximately 5 minutes. This procedure was repeated until all offsets had been determined.

Suspended sediment samples for calibration of the FOSLIMs and MEXs were taken on a regular basis during the measurements. The samples were pumped from the sensor positions perpendicular to the direction of the flow by means of small tubes (4 mm diameter) and a peristaltic pump. The intake velocity was approximately 0.1 m/s. Sampling errors due to the inertia of large suspended particles were negligible because the suspended sediments consisted mainly of silt and clay. Following Crickmore and Aked (1975, in Van Rijn, 1993) sampling errors were estimated at 10%. Sampling times were registered in advance of the suspended sediment sampling; the sampling took about 1-2 minutes. The samples were stored in the dark in order to avoid primary production. The SSCs were determined in the Laboratory of Hydromechanics through filtering using mixed celluloid ester filters with 0.15 μm pore size. The errors made in determining the SSCs were found to be much smaller than the sampling errors. For a description of the filtering procedure the reader is referred to De Wit (1992).

For a comprehensive description of all available information about the estuary during the measuring periods in 1995 and 1996, such as water levels, wind velocities, discharges at Pogum and Nieuwe Statenzijl etc., the reader is referred to Ridderinkhof et al. (in prep.).

3.3.2 Data processing
The time records of the EMFs and FOSLIMs were processed with the software packages MATLAB 5.0 and EXCEL 7.0. Three processing stages are explained in this section: spike identification and removal, calibration, and determination and correction of possible sensor tilt.

Spikes are typically identified as a small number of outliers in a signal which result from other processes than the process under investigation, such as instrument vibration, interference etc. These processes probably influence the signal to a certain extent in the range of the signal itself, but this is neither easily noticed nor remedied afterwards. Spikes have a relatively large impact on some of the turbulence properties. The procedure followed in this study is therefore to identify the spikes and to check
their influence on the parameters under investigation. If removal of the spikes alters the
conclusions drawn, it will be brought up for discussion.

The identification of spikes is based on estimates of the probability density
distributions of representative parts of the signals. The probability density distributions
showed that temporal turbulence fluctuations seldom exceeded a threshold of about
five times the standard deviation. This finding is in agreement with probability density
distributions presented by Kwanisi and Yokosi (1993), French and Clifford (1992) and
West and Shiono (1985). It was therefore decided to use the threshold of five times the
standard deviation away from the mean as a criterion for spike identification. Each
signal was divided into 10 minute records of which standard deviations were
determined. Peaks were removed from the records if they exceeded the threshold. A
complete 10 minute record was rejected if spike removal lowered the standard
devation by more than 10%.

The EMF responses were obtained from laboratory calibrations (Section 3.2) and
the offsets were determined in-situ (Section 3.3.1). Using the calibrations, a computer
program calculated the horizontal and vertical velocity components \( U \) and \( W \)
respectively) and the SSC \( C \).

The FOSLIMs were calibrated in-situ by taking samples of suspended sediment at
the FOSLIM positions at time intervals of approximately 15 minutes (see also Section
3.2). The one minute averaged FOSLIM signals were compared directly to the SSCs
obtained from sampling. If necessary, small corrections were made for the sampling
times: always in the same direction and over the same time interval for all turbidity
meters. Offsets and responses were obtained from least squares fits through the plots
of SSC against FOSLIM output voltages. For the calibration of the MEXs a similar
procedure was followed.

In order to correct for possible tilt of the EMF sensor heads the EMF axes were
rotated over an angle \( \varphi \) such that there was no longer a correlation between \( U \) and \( W \)
when evaluated over the measuring period (see also Darbyshire, 1993). The velocities
were corrected according to:

\[
U = U_m \cos \varphi + W_m \sin \varphi
\]  \hspace{1cm} (3.1)

\[
W = -U_m \sin \varphi + W_m \cos \varphi
\]  \hspace{1cm} (3.2)

where the subscript \( m \) denotes the measured values.
3.3.3 Feasibility test in July 1995

After some small scale laboratory tests an in-situ measurement was conducted in the tidal channel "Groote Gat" in July 1995 in order to test the feasibility of the instruments for measuring the turbulence parameters of interest: turbulence intensities, Reynolds stress, turbulent transport of sediment, and related parameters. Delft hydraulics took part in this test because of their expertise in the matter of turbidity measurements with the FOSLIM and field measurements in general.

Data collection and processing

The test was carried out at pole RWS208 for a two hour period during flood tide at 7 July 1995 starting from 14:54h local time. The water depth changed from 2.5 m to 4 m during the measuring period. The weather conditions were very moderate and during the measuring period the wind from the south decreased from 5 to 3 m/s.

A single prototype FOSLIM and a single EMF of Delft hydraulics were mounted on the measuring frame and were lowered to a level of approximately 1 m above the sediment bed. The time series of the SSCs were obtained through calibration of the one minute averaged FOSLIM signal with 7 suspended sediment samples. All sampling times were shifted backward for 5 minutes in order to get better agreement between the FOSLIM signal and the samples. The calibrated signal and the sediment samples are shown in Figure 3.4.

The manufacturer's calibration data of the EMF were used: 10 V/(m/s) for the responses and zero offsets. The EMF axes were rotated such that there was no correlation between $U$ and $W$ evaluated over the measuring period, see Section 3.3.2. In this way it was found that the velocity data had been collected at a tilt of -2.9°.

A visual inspection of the signals revealed that some spikes were present in the X-channel of the EMF. The next section addresses, amongst other things, the influence of their removal on some of the turbulence properties.

Data analysis and discussion

The time series of one minute averaged longitudinal velocity component and SSC, $\bar{U}$ and $\bar{C}$, respectively, are presented in Figure 3.4. The agreement between the samples and the calibrated signal is fair, except for a sample taken at 16:15 hours which had an extremely large standard deviation. This points to errors made during filtration and this sample was therefore omitted. The measured time series of $U$ and $C$ seemed realistic, which enhanced confidence in the measuring technique and in the performance of the FOSLIM.
Turbulence properties and spectral analysis

After calibration the data were divided into 10 data-records of 10 minutes. Trends were computed and subsequently removed from the records. The turbulence intensities were computed from the variances, and the Reynolds stress and the vertical turbulent transport of suspended sediment were computed from the covariances, of $u$, $w$, and $c$. The stationarity of each 10 minute record was examined by dividing it into 1 minute segments and then applying a run test thereby following a standard procedure (see Section 4.5 or Bendat and Piersol, 1971).

![Graph showing SSC and longitudinal velocity over time]

Figure 3.4. Test measurement during flood, July 7 1995. The vertical bars denote the estimated sampling errors in the samples.

Table 3.1. Outcome of the applied run test.

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*S = stationary, **N = non-stationary.

Table 3.1 shows, for example, that the first three records of $c'$, and the sixth record are non-stationary. This is in agreement with the trends in the averaged SSC during
these periods shown in Figure 3.4. Records 4 and 7 can be considered stationary for "all" turbulence properties and are selected for spectral analysis.

Auto-spectral density functions (auto-spectra) and the cumulative auto-spectra and co-spectra are calculated for records 4 and 7 (see also Section 4.2). The auto-spectra for $u$ and $c$, normalized with their variances, plotted against the wave number, $k$, are shown in Figure 3.5, and the cumulative spectra are shown in Figure 3.6. The wave number $k = 2\pi f / \bar{U}$ where $f$ is the frequency. For computational aspects of the spectra the reader is referred to Section 4.5.

![Figure 3.5. Normalised auto-spectra of records 4 and 7: (a) normalised auto-spectrum $S_{uu}$ of $u$; (b) normalised auto-spectrum $S_{cc}$ of $c$; ---, record 4; ---, record 7; the wave number is defined as $k = 2\pi f / \bar{U}$, $f$ is the frequency.](image)

The sharp decrease of $S_{cc}$ at $k = 50$ m$^{-1}$ is the result of analogue filtering beyond the cut-off frequency of $\sim 3$ Hz of this FOSLIM prototype. The FOSLIMs used in the measurements of August 1995 and of 1996 had cut-off frequencies of 10 Hz.

The slopes of $S_{uu}$ over $1$ m$^{-1} < k < 100$ m$^{-1}$ for both records 4 and 7 are in agreement with the expected -$5/3$ behaviour in the inertial subrange (Nieuwstadt, 1992; Hinze, 1975). The slope of $S_{cc}$ is about -$5/4$ whereas the -$5/3$ behaviour is expected for spectra of passive scalars (Hinze, 1975). The slope of -$5/4$ is more or less in agreement with auto-spectra of $c$ presented by West (1989). The difference between the "scalar" and "suspended sediment" slopes is addressed in Section 4.5.

The cumulative spectra shown in Figure 3.6 are plotted against $kz$, in which $z$ is elevation above the bottom. According to Soulsby (1980) it can be assumed that through this scaling the normalised spectra measured at different heights in the water
column collapse into a single curve. Nezu and Nagagawa (1993) show that better results are obtained if \( k \) is multiplied by the integral scale \( L_x \). This subject is further discussed in Section 4.5. In this Chapter we follow the work of Soulsby (1980) which is common practice in field research.

![Graphs](image_url)

Figure 3.6. Cumulative spectra for record 4: (a) cumulative normalised auto-spectra; (b) cumulative normalised co-spectra.

High-frequency losses are corrected for and are computed by extrapolating the tails of the spectra, according to the \( k^{-5/3} \) and the \( k^{-7/3} \) dependence for the auto-spectra and co-spectra respectively, down to the Kolmogorov wave number \( k_d \), defined as \( 2\pi/\eta \), where \( \eta = (v^3/\varepsilon)^{1/4}, \) \( v \) is the kinematic viscosity, and \( \varepsilon \) is the turbulence dissipation rate. The turbulence dissipation rate \( \varepsilon \) is determined assuming local equilibrium between turbulence production and dissipation so that it may be calculated from the multiplication of the measured Reynolds stresses and the mean velocity gradient, which is here calculated from a log-velocity distribution.

The highest losses are 15% for \( c^2 \), 8% for \( \bar{cw} \) and 6% for \( w^2 \). The loss of \( c^2 \) is mainly due to the relatively low cut-off frequency of about 3 Hz of the prototype FOSLIM. The spectral losses for these measurements are therefore considered small.

If measurements are made closer to bed, the losses increase due to cut-off losses and sensor size. If, for example, the measuring height \( z \) is reduced from 1 m to 0.2 m it can be derived (from Figure 3.6) that cut-off losses are approximately 20% for both \( c^2 \) and \( \bar{cw} \), for infinitely small sensor size and 30% if the “sensor size” of the combination of EMF and FOSLIM is estimated at 0.1 m (\( kz \approx 13 \)).
Figure 3.7 shows the corrected Reynolds stress and turbulent transport of sediment. The trend in Reynolds stress shows on average an increase over the measuring period which is in agreement with the increase in the velocity $\bar{U}$ shown Figure 3.4.

![Graph showing Reynolds stress and turbulent transport](image)

Figure 3.7. Time series of Reynolds stress and the turbulent transport of sediment (flood tide July 7); $z/h$ varies from 0.4 to 0.2, where $h$ is the water depth.

At the start of the measuring period, at 15:05h, a decrease in Reynolds stress is observed. This decrease cannot be explained from changes in the velocity $\bar{U}$ since $\bar{U}$ is almost constant during this period. The trend in $\bar{cw}$ shows an increase until 15:30h, while Figure 3.4 shows that $\bar{C}$ already starts to decrease at 15:10h. This can be explained from a limited amount of sediment available for transport. The sediment is resuspended in the first part of the flood but remains close to the bed thereby creating a density gradient. This gradient could affect the turbulence structure. When the flood velocities become larger the available amount of sediment becomes homogeneously distributed over the entire water column. The SSC near the bed then decreases (see also Chapter 5).

The correlation coefficients for the Reynolds stress and the turbulent transport of sediment are defined as:

$$ R_{uw} = \frac{\overline{u'w'}}{u'w'} $$

$$ R_{cw} = \frac{\overline{c'w'}}{c'w'} $$

(3.3)  
(3.4)
These coefficients were also corrected for high-frequency losses as explained above. The noise levels of the EMF and the FOSLIM (see Section 3.2) do not change the turbulence intensities significantly (increase less than 5%) and are neglected.

Spike removal was found to have no significant influence on the turbulence parameters except for \( \bar{u} \) and \( R_{sw} \). \( \bar{u} \) decreases by 40% at 15:00h and by approximately 10% at 15:30h and 15:40h, whereas \( R_{sw} \) increases from 0.32 to 0.38 at 15:00h. The first 10 minute data-record of \( \bar{u} \) is therefore rejected.

Figure 3.8. The correlation coefficients for the Reynolds stress and the turbulent transport of sediment.

Figure 3.8 shows the time histories of \( R_{sw} \) and \( R_{cw} \) for this test. The trends of \( R_{sw} \) and \( R_{cw} \) are very similar. The small values of \( R_{sw} \) and \( R_{cw} \) directly after the start of the measuring period can tentatively be explained from stratification effects. This would be in agreement with the decrease in Reynolds stress during this period whilst \( \bar{U} \) remains constant (see Figures 3.4 and 3.7). Stratification effects are considered in greater detail in Section 4.7. The maximum values of approximately 0.40 for \( R_{sw} \) and 0.35 for \( R_{cw} \) would then represent correlation coefficients for unstratified flow. The magnitudes as well as \( R_{sw} \) being larger than \( R_{cw} \) are in agreement with measurements of Komori et al. (1983) and West and Oduyemi (1989). Komori et al. (1983) found from laboratory experiments values for \( R_{sw} \) of about 0.4 and for the correlation coefficient for the heat flux, \( R_{uth} \), about 0.25. West and Oduyemi (1989) found from in-situ measurements in the Cowory and Tamar estuaries values for \( R_{sw} \) of about 0.5 and for \( R_{cw} \) of about 0.25 (see Section 1.3).

This test demonstrated the feasibility of combining a FOSLIM and an EMF for measuring turbulence parameters in-situ, so that insight into the turbulence structure in
a tidal channel can be obtained. The results of this test were promising, though tentative, since the measuring period covered only two hours and measurements were made at only one level in the water column.

3.3.4 Test measurement of August 1995

This measurement was made to test the complete measuring system as shown in Figure 3.3. In addition to this set-up, two MEX turbidity sensors were mounted at $z = 0.7$ m and $z = 1.3$ m. The installation of this heavy equipment was carried out using the vessel Regulus of Rijkswaterstaat Meedienst Noord.

The test was carried out at pole RWS208 during a three and a half hour period during flood on Thursday August 31, 1995 starting from 11:31h local time, and during a nine hour period on Friday September 1, 1995 starting from 11:51h local time. The tidal range was 3.3 m. The weather conditions were moderate: cloudy but no rain, relatively large wind speeds from the north ranged from 8 m/s to 12 m/s (Beaufort 5 - 6).

These measurements showed that the complex measuring system, consisting of equipment and instrumentation from the BOA Measuring Bridge, from the Laboratory of Hydromechanics of the Delft University of Technology, and from Utrecht University needed further development before it could be used successfully. The FOSLIMs and MEXs worked well but the EMFs did not function properly, probably because of a connector failure between the EMFs and the datalog-computer. In the winter of 1996 a new, smaller, datalog-system was purchased replacing some of the heavy BOA equipment. This made the installation of the measuring system much simpler. The MEX sensors were omitted for the sake of convenience. Their task, performing long term turbidity measurements, was taken over by the three MEX sensors of the Rijkswaterstaat measuring frame.

Figure 3.9 shows the calibrated signals of two FOSLIMs and two MEXs located at approximately the same level during flood at August 31, 1995. Even the short term variations in the FOSLIM signals are in agreement with those of the MEXs, especially at $z = 0.7$ m. The data do not show significant offset drift.

Twelve suspended sediment samples were taken in total, four at each level. The errors of the calibrations are relatively small for all instruments despite the small number of sample points. The errors for the FOSLIM calibrations were even smaller compared to those of the MEX calibrations (Figure 3.9). This can be explained from the fact that the suspended sediment samples were taken exactly at the FOSLIM positions. These errors being small also indicates that the sampling errors were
probably somewhat smaller than the estimated 10% (see Section 3.3.1). These results show that FOSLIMs are suitable instruments for measuring suspended sediment concentrations up to a few grams per litre in the field.

![SSC graph with error markers](image-url)

**Figure 3.9.** Comparison of one minute averaged data of FOSLIM and MEX turbidity sensors during the test measurement August 31, 1995: ———, MEX at \( z = 0.7 \) m; ——–, MEX \( z = 1.3 \) m; ◦, FOSLIM at \( z = 0.7 \) m; △, FOSLIM at \( z = 1.1 \) m.

### 3.4 Measuring periods in 1996

Measuring periods during neap and spring tides of different seasons were selected in order to study variations in maximum flow velocities and SSCs. The precise dates were determined in consultation with the colleague researchers of the BOA-theme project for reasons of logistics and synchronism of the measurements. Some mutual interests were to be harmonised. The EMFs, for example, were not only needed for turbulence measurements in the channel “Groote Gat” but also for flow velocity measurements on the adjacent tidal flat “Heringsplaat” during joint field measurements of physicists and biologists. Because simultaneous channel and flat measurements would greatly enhance the value of the total BOA data-set, a compromise was made such that the turbulence measurements were carried out simultaneously with in-situ settling velocity measurements in the tidal channel (Van der Lee, in prep.) and directly after the joint measurements on the tidal flat.

**Actual measuring periods**

Dates of the field surveys carried out in the Groote Gat in 1996 together with the prevailing conditions are listed in Table 3.2.
### Table 3.2. The measurements made in 1996.

<table>
<thead>
<tr>
<th>Season</th>
<th>Period</th>
<th>Tide, max.(ebb) velocity</th>
<th>Concentration range</th>
<th>Wind conditions</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>Spring</td>
<td>16-20 April (5 tides)</td>
<td>spring, 0.8 m/s</td>
<td>0.1 - 0.5 g/l</td>
<td>4 - 5 m/s, S</td>
<td>- VIS* measurement</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>- malfunction of FOSLIM</td>
</tr>
<tr>
<td>Summer</td>
<td>25-28 June (4 tides)</td>
<td>neap, 0.7 m/s</td>
<td>0.2 - 1.2 g/l</td>
<td>0 - 8 m/s, N-W</td>
<td>- adapted FOSLIMs</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>- small amounts of seaweed</td>
</tr>
<tr>
<td></td>
<td>4-6 July (3 tides)</td>
<td>spring</td>
<td>0.3 - 3.0 g/l</td>
<td>5 -12 m/s, SW</td>
<td>- large amounts of seaweed</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>7-9 August (3 tides)</td>
<td>neap, 0.7 m/s</td>
<td>0.2 -0.8 g/l</td>
<td>3 - 8 m/s, S-SW</td>
<td>- VIS measurement</td>
</tr>
</tbody>
</table>

* Video In-situ, system for floc size and settling velocity measurements.

Table 3.2 shows that a considerable number of tides could be measured in 1996. This offered the possibility to select parts of the data-set with relatively high quality (see Section 3.5) and to verify certain findings by applying similar analyses to comparable parts of the data-set. The differences in maximum velocities during spring and neap tides were not particularly large. The SSCs were not very high, especially not for the spring tide of April.

### 3.5 Quality of 1996 data

High-quality data are required for the analysis of effects of stratification on turbulence properties. High-quality data in this study stands for data which are free of unwanted influences, such as imperfections of the measuring system and unsuitable measuring conditions for this type of research. Criteria for the assessment of the quality of the data concern, for example, the number of spikes, high-frequency losses, sensor tilt, wave activity, SSC etc. In this section parts of the data-set are selected for further analysis. Both the performance of the measuring system and the measuring conditions are evaluated. Not all considerations are presented herein: only some discussions about the quality of the data of June 1996 are presented. This discussion is representative of the assessment of the quality of the data of other measuring periods. A summary of the data qualities of measurements made in 1996 is presented at the end of this section.
Discussion of the quality of the data of June 1996

Figure 3.10 shows the time series of SSCs and velocities close to the channel bed during June 27. The FOSLIMs were calibrated in-situ during ebb and the MEXs were calibrated during flood on June 26 (coefficients of determination \( R^2 > 0.95 \)).

Small differences between the MEX results and FOSLIM results shown in Figure 3.10 can be attributed to the different measuring sites (separating distance is 5 m) and to the errors of approximately 0.05 g/l which result from the calibration procedure. Large differences (> 0.1 g/l) can only be explained from large gradients in the SSC or from sensor malfunctioning. The peak at 19:00h in the FOSLIM signal is in agreement with the peak at 12:00h and might be explained from the build-up of high SSCs close to the bottom as the result of the reduction in the tidal flow velocity. However, the MEX-signal does not show such a peak at 19:00h. Moreover, the peak in the FOSLIM signal that occurs at 21:00h is unrealistically high. These large increases in turbidity are most likely caused by fouling owing to seaweed.

![Graph](image)

**Figure 3.10.** Some results of the field measurements during neap tide on June 27, 1996: (a) SCC near the bed; \(\Delta\), FOSLIM at \(z = 0.1 \text{ m}\); \(---\), MEX at \(z = 0.3 \text{ m}\); (b) Velocity at 1.0 m; \(\square\), EMF at \(z = 1.0 \text{ m}\).

The wind from the north reaches speeds of 8 m/s (4-5 Beaufort) and is directed along the channel. The LW period is marked in Figure 3.10 by a low velocity, and a short increase in the suspended sediment concentration. Auto-spectra of \(u\) during LW show distinct peaks at about 0.5 Hz, the peak at 1.0 m being larger than the peak at 0.1 m above the bottom (see Figure 3.11). These peaks are attributed to wave activity.
The wave activity disappears from the auto-spectra, when the flow velocity increases from approximately zero to 0.5 m/s and the water depth increases from 1.8 m to 2.5 m. However, it is mentioned herein that the spikes in the tails of the spectra remain present. This subject is further discussed in Appendix B.

Figure 3.11. Auto-spectra of 10 minute records at June 27, 1996: ---, $S_{uu}$ at 14:45u, $\bar{U} < 0.1$ m/s, $h = 1.8$ m; ---, $S_{uu}$ at 15:45u, $\bar{U} = 0.5$ m/s, $h = 2.5$ m; (a) $S_{uu}$ at $z = 0.1$ m; (b) $S_{uu}$ at $z = 1.0$ m.

To be on the safe side, only those parts of the June data are included in the data-set for further research which display only small differences between the recordings of MEX and FOSLIM, and which show no evidence of wave activity.

**Summary of data quality of measurements made in 1996**

The data of the April measurements showed large differences in quality. The velocity time series were of relatively good quality: only a limited number of spikes and minor wave activity were present. The SSCs were extremely low so that suspended sediment-induced stratification effects were fully absent. Time series of SSC of the FOSLIMs were unreliable due to malfunctioning of these instruments (see also Section 3.2 and Section 3.4). The data of the April measurements were therefore not included in the data-set for further research.

The data of the June measurements are of considerably better quality, when compared to the April data. Good in-situ calibrations were available for MEX and FOSLIM sensors and the velocity time series were of good quality. Nevertheless, not all data are suitable because of the influence of seaweed and wave activity already mentioned.
The data of the July measurements are of low quality, since large amounts of seaweed were found on the instruments when the measuring frame was lifted. The FOSLIM at $z = 1.0$ m was damaged in such a way that it could not be fixed.

![Graphs](image)

Figure 3.12. Some results of the field measurements during neap tide on August 7-8, 1996: (a) SCC near the bed; $-\Delta-$, FOSLIM at $z = 0.1$ m; $-\ldots-\ldots$, MEX at $z = 0.3$ m; (b) Velocity at 1.0 m; $-\square-$, EMF at $z = 1.0$ m.

The data of the August measurements, on the contrary, are of good quality as seaweed was no longer present and most instrumentation worked properly. Some parts of the FOSLIM signals, which were recorded at $z = 0.1$ m and $z = 0.4$ m, were rejected because spikes removal resulted in relatively large changes in the turbulence intensities (> 10%, see Section 3.2.2). Some influence of wave activity was present during LW. However, the major part of the August data set was used for further analysis (see Chapter 4). Figure 3.12 shows an example of the time histories of SSCs and velocities.

3.6 Conclusions
The test measurements in June and August 1995 showed that the FOSLIM is a useful high-frequency device for measuring SSC. Direct comparison with another robust turbidity sensor of the MEX type showed that drifts in offset and response were almost absent during the test period. It was found that the accuracy of the FOSLIM depends to a large extent on the in-situ calibration procedure.

It has been shown that by combining a FOSLIM and an EMF it is possible to measure vertical turbulent fluxes of fine sediments. The value of the correlation coefficient $R_{ew}$ for neutral flow conditions was about 0.35 which is in agreement with

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values presented in literature. High-frequency losses, estimated from the co-spectra, are relatively small, about 15% for $c^2$ and about 8% for $cw$. According to Soulsby (1980), higher losses should be anticipated if measurements are made near the bed.

The data sets obtained from the field measurements in 1996 showed large quality differences. In April 1996 SSCs were extremely low so that suspended sediment-induced stratification effects were fully absent. These low SSCs were ascribed to biogenic stabilisation of the sediment beds (Kornman and De Deckere, 1998). Part of the measurements made in June and July 1996 were hindered by the presence of seaweed. The data obtained in August 1996 showed a good quality except for some parts of the FOSLIM signals, which contained a large number of spikes; these parts were excluded. Part of the June data and the major part of the August data were used for further analysis, the results of which are presented in Chapters 4 and 5.
Chapter 4 Results of the field measurements

4.1 Introduction
This chapter presents and discusses some typical aspects of estuarine flows based on results of the field measurements conducted in the tidal channel Groote Gat during June and August 1996. The focus is on stratification effects caused by vertical gradients in sediment concentration. First a general picture is drawn of the prevailing hydrodynamic conditions at the measuring location (see also Section 3.1). Some considerations about the influence of, among other things, flow acceleration and transverse shear, are presented in Sections 4.3. The principal results are presented in Sections 4.4 to 4.7. Section 4.4 shows results of the turbulence measurements such as: almost complete records of turbulent transports of sediment in horizontal and vertical directions over a period of two tidal cycles. This type of measuring results contributes significantly to the insight into the local transport processes of suspended sediment and were rarely published. In Section 4.5 spectral analysis is used in order to obtain insight into the spatial scales of velocities and sediment concentrations. Results of the computation of stratification levels and effects of the stratification on a number of turbulence parameters are presented in Sections 4.6 and 4.7, respectively. Finally, the various results of the turbulence measurements are integrated and discussed in Section 4.8.

4.2 Water levels, mean velocities and mean concentrations
Some typical results of the measurements conducted in June and August 1996 are presented in Figures 4.1 and 4.2. The figures show the water levels and their time derivatives, the flow velocities in longitudinal direction, and the SSCs.

The prevailing conditions were moderate during both measuring periods, that is, moderate wind speeds and small river discharge. Under these conditions little effects of wave activity are expected, except for periods of low water, and of salinity induced density stratification in the Dollard reach.

The water levels in June were monitored at the Rijkswaterstaat measuring pole Dollard Noord, 2 km north, and in August at the Rijkswaterstaat pole Schanskerdiep, 1 km south of the measuring location. The flow velocities and 10 minute averaged
suspended sediment concentrations were registered at three different heights above the channel bed using the turbulence measuring frame and the Rijkswaterstaat frame, respectively (see also the legends of Figures 4.1 and 4.2). Of the suspended sediment concentrations only the values obtained at the highest and lowest positions are shown. For more details about the prevailing experimental conditions and the experimental setup the reader is referred to Chapter 3.

Figure 4.1. Results of the field measurements during June 26 - 28, 1996 (10 minute averaged values): (a) water level, $h$, and time derivative of the water level, $\partial h/\partial t$; (b) longitudinal velocities, $\bar{U}$ ($>0 = \text{flood}$); (c) suspended sediment concentrations, $\bar{C}$.

**Water levels and mean velocities**

Figures 4.1a,b and 4.2a,b show that at maximum and minimum water levels the flow velocities are approximately zero. The phase difference of about 90° between the water level and the flow velocity is typical of the semi-enclosed tidal basin behaviour of the Dollard (Dorrestein, 1960).
A comparison between the measured water levels in August and the tide-table for the Dollard shows that the mean water level was approximately 0.2 m lower than expected. This set-down was probably caused by the easterly winds at sea in a period preceding the measurements.

![Graphs showing water level and velocity variations.](image)

Figure 4.2. Results of the field measurements in August 7 - 9, 1996 (10 minute averaged values): (a) water level, $h$, and time derivative of the water level, $\partial h/\partial t$; (b) longitudinal velocities, $\bar{U}$; (c) suspended sediment concentrations, $\bar{C}$.

The velocities in Figure 4.1b and 4.2b show a significant ebb dominance: about -0.6 m/s on average for the maximum ebb velocity compared to about 0.5 m/s on average for the maximum flood velocity, the absolute velocities in June being slightly larger during ebb compared to August. Ebb dominance is also present in the velocity measurements made on the adjacent tidal flat Heringsplaat, 400 m south west of Pole 208 (Ridderinkhof et al., in prep.).

The flow velocities in the Dollard basin are affected by the presence of the tidal flats, as the free surface area increases drastically when the tidal flats become

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submerged. This is clearly shown in Figure 4.1: when the time derivative of the water level starts to decrease, the flood velocities continue to increase for about three to four hours. The tidal flats get submerged when the water level exceeds about 0.2 m.

**Suspended sediment concentrations**

The time series of 10 minute averaged suspended sediment concentrations of June and August 1996 are shown in Figure 4.1c and 4.2c, respectively. The maximum measured SSCs in June were approximately 1.2 g/l and in August 0.8 g/l. The amount of suspended sediment and distribution of the suspended sediment in the water column is the result of a number of sediment transport processes such as erosion, settling, turbulent mixing, advection, and the availability of sediment for erosion. Interpretation of the suspended sediment concentration measurements is hampered by the lack of knowledge of upstream processes. Therefore, the explanation of the time series of suspended sediment concentrations can best be done through analysis with a numerical model in which the dominant processes are represented in a well-balanced way. In Chapter 5 the concentration time series are analysed by means of a 1DV mathematical model (Uittenbogaard et al., 1996). Some comments are made herein about typical features of the time series of suspended sediment concentration, such as the relatively large peaks observed and the time lags between the velocity and concentration curves.

The general relation between velocities and sediment concentrations is obvious: larger velocities tend to imply larger concentrations. If the time series of velocities and suspended sediment concentrations are inspected in more detail, the relation between them is less clear. The SSCs during flood are similar to the SSCs during ebb, whereas the maximum flow velocities during flood are smaller than those during ebb (ebb dominance). Numerical studies of the Dollard show that the small differences between ebb and flood SSC may be explained from a limited amount of sediment available for resuspension (see also Section 5.4; Ridderinkhof, 1998). The scour lag is the time lag between the increase in velocity and in SSC. It is about 1 hour on average. The settling lag is the time lag between the decrease in velocity and in SSC. Figures 4.1c and 4.2c show that there is hardly any settling lag: the sediment already settled when velocities are still relatively high. In most cases the concentration peaks during flood in Figure 4.1 and to a lesser extent in Figure 4.2, precede the maximum flood velocities, but those during ebb lag behind the maximum ebb velocities.

A typical feature of the time series of concentration in June is that during flood the onset of resuspension takes place at the moment that the velocity drops for a short time (June: 6/26 14:30; 6/27 3:30; 6/27 17:00; 6/28 5:00), that is at approximately at h
-0.7 m N.A.P. (see also the dot-dash lines in Figures 4.1 and 4.2). An explanation for this could be that the sediment becomes available for resuspension when the water level reaches the borders of the tidal channel. At these borders large amounts of soft deposits of cohesive sediment can be found. Activity of waves could also contribute to the availability of sediment for resuspension in these periods of the tide. The measuring location is relatively close to the border of the Heringsplaat.

On a number of occasions this material is not mixed over the entire depth. The buoyancy effect resulting from the relatively large gradients of suspended sediment concentration may have affected the turbulence structure (see Section 4.6). The simultaneous increase and decrease of concentration at the 0.3 m and 1.4 m level, respectively, towards low water slack, (June: 6/27 13:00h; 6/28 1:00h; August: 8/7 22:00h; 8/8 10:00h) show that the distribution of sediment over the depth changes from homogeneous to high near-bed concentrations and low concentrations in the higher parts of the water column.

**Determination of vertical velocity and concentration gradients**

In order to estimate the suppression of turbulence owing to density stratification, resulting from vertical concentration gradients, gradient Richardson numbers have been calculated. The time series of the Richardson number are presented in Sections 4.6 and 4.7, together with turbulence parameters also showing the influence of stratification.

Mean vertical gradients of velocity and concentration are needed to calculate the Richardson numbers. Mean velocity and concentration profiles were determined through curve fitting using the least-squares method. The empirical Van Veen profile proved suitable for fitting to the mean velocities. This profile is given by:

\[
U(z) = (n+1)U_d \left( \frac{z-z_b}{h} \right)^n
\]  

(4.1)

where

\[
\begin{align*}
 n &= \frac{1}{4.7} \left( \frac{d_r}{h} \right)^{0.06} \\
 U_d &= \text{the depth averaged velocity, } n \text{ is a parameter which depends on the effective roughness } d_r. \text{ The height } z_b \text{ is a correction height which represents the difference between the bed level measured directly under the pole and the average bed level upstream. This difference was caused by scour near the bed pole. A sketch of the situation is presented in Figure 4.3. The correction height } z_b \text{ was adapted in such a manner that Eq. 4.2 yielded more or less realistic values for } d_r \text{ on average (about 0.1}}
\end{align*}
\]
m). The “new” sensor heights with respect to the undisturbed bed level are summarised in Table 4.1. The three levels of measurement above the channel bed are referred to as the 0.1 m level, the 0.4 m level and the 1.0 m level throughout this thesis.

![Figure 4.3. Sketch of the flow situation near the experimental set-up.](image)

Table 4.1. Overview of the sensor heights measured in-situ and after correction for the scour hole.

<table>
<thead>
<tr>
<th>Aimed heights</th>
<th>Measured in-situ</th>
<th>After correction for scour</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>June</td>
<td>August</td>
</tr>
<tr>
<td>0.1 m</td>
<td>0.1 m</td>
<td>0.15 m</td>
</tr>
<tr>
<td>0.4 m</td>
<td>0.4 m</td>
<td>0.5 m</td>
</tr>
<tr>
<td>1.0 m</td>
<td>1.0 m</td>
<td>1.0 m</td>
</tr>
</tbody>
</table>

The results for the velocity profiles of the June period are shown in Figure 4.4. During flood a value of 0.18 is found for the parameter $n$ and during ebb it varies between 0.2 and 0.25: the velocity profiles during flood are steeper than during ebb. During the August period the situation is reversed: $n = 0.19$ for the flood period and $n = 0.13$ for the ebb period. The differences found between the ebb and flood profiles can be explained from uncertainties in the offsets of the EMF's of 0.01 m/s (see also Section 3.2) and from differences between the bed level in the seaward and landward directions. For the 0.4 m levels and 1.0 m levels the uncertainties in $n$ have only small effects on the velocity gradient.

Figure 4.4 also shows that during slack water periods the time series of parameter $n$ shows large fluctuations. The velocity gradients obtained from the Van Veen profile during these periods are not valid. However, the absolute differences between the “fitted” Van Veen profiles and the actual measurements are often small ($< 0.02$ m/s). Outside these periods the average error in velocities is approximately 5%.
Figure 4.4. Depth averaged velocities ($U_d$), and the parameter $n$ of the Van Veen profile during the June measuring period (Eq. 4.1): $--$, $U_d$; $----$, $n$.

Figure 4.5. Fitted velocity profiles from maximum flood to maximum ebb. The times shown in the figure denote the time from high water slack (19:30h August 7, 1996): $+$, $U$ measured (10 minute average); $----$, "Van Veen fit".

Figure 4.5 presents some velocity profiles during a period of flow deceleration from maximum flood to HW, and during flow acceleration from HW towards maximum ebb velocity. The flow acceleration during ebb is much larger than the deceleration during flood (compare velocity profiles at HW -1:20h and HW +1:10h). The Van Veen fit is not appropriate for the velocity profile at HW +0:30h, the measurements indicate that the velocity profile should bend backwards in the upper part of the water column. This effect may be due to acceleration effects and is in qualitative agreement with the models of Soulsby and Dyer (1981) and of Kuo et al. (1996) and is discussed in more detail in Sections 4.3 and 5.3.

The Rouse profile was used for fitting to the mean concentrations. The Rouse profile is given by:

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\[ C(z) = C_0 \left( \frac{z-z_b}{h-(2-z_b)} \right)^{-\alpha} \]  \hspace{1cm} (4.3)

where

\[ \alpha = \frac{w_s}{\kappa |u_*|} \]  \hspace{1cm} (4.4)

$C_0$ is a coefficient equal to the concentration at mid-depth, $w_s$ is the settling velocity in the case of a single grain size diameter, $\kappa$ is the Von Kármán constant and $u_*$ is the friction velocity. The Rouse profile is a solution of the balance equation for the suspended sediment provided a number of conditions are met. These are: steady-state conditions, zero horizontal concentration gradients, parabolic distribution of the eddy diffusivity, and, as already mentioned above, a single grain size and unhindered settling. These conditions are too stringent in this case. Nevertheless, in view of the poor spatial resolution (only three points) and the relatively large measuring errors, the assumption of the Rouse distribution for the concentration seems acceptable. However, the values found for the coefficients $C_0$ and $\alpha$ from fitting to the data have limited physical meaning unless the conditions mentioned above are approximately met during certain phases of the tide.

![Graph showing concentration at mid-depth ($C_0$), and the parameter $\alpha$ of the Rouse profile (Eq. 4.3) during the June measuring period: $\ldots$, $C_0$; $\ldots$, $\alpha$.]

The time history of $C_0$ shows strong similarities with the time histories of concentration presented in Figure 4.1. The time histories of $\alpha$ show large variation and sometimes even negative values. Large values of $\alpha$ coincide with periods of low suspended sediment concentrations. From Eq. 4.3 and 4.4 the settling velocity $w_s$ can be calculated. The settling velocity $w_s$ varies between 0.5 and 1.5 mm/s for periods with appreciable suspended sediment concentrations ($C_0 > 0.1$ g/l). Except for the periods of slack water, the average errors of the concentration “fit” are about 10\%.

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These errors are comparable to the errors which resulted from the calibration of the MEX’s (see Section 3.1.1).

4.3 Analysis of equation of motion

The equation of motion for the longitudinal direction in well mixed estuaries may be presented as (McDowell & O’Connor, 1977, Shiono and West, 1987):

\[
\frac{\partial \bar{U}}{\partial t} + \bar{U} \frac{\partial \bar{U}}{\partial x} + \bar{V} \frac{\partial \bar{U}}{\partial y} + \bar{W} \frac{\partial \bar{U}}{\partial z} = -g \frac{\partial \eta}{\partial x} - \frac{g}{\rho} (h - z) \frac{\partial \rho}{\partial x} - \frac{\partial \bar{u}^2}{\partial x} - \frac{\partial \bar{w}^2}{\partial z} + f \bar{V} \tag{4.5}
\]

Some of the variables were already introduced in the previous sections but for the sake of completeness all variables are defined: \(x\) (positive in the landward direction), \(y, z\) (positive upward) are the longitudinal, transverse and vertical co-ordinates. \(\bar{U}, \bar{V}\) and \(\bar{W}\) are the ensemble averaged velocity components, \(g\) is the acceleration due to gravitation, \(\eta\) is the surface elevation, \(\bar{\rho}\) is the ensemble averaged density, the terms \(\bar{u}^2, \bar{w}^2\) and \(\bar{uw}\) represent Reynolds stresses and \(f_c\) is the Coriolis-parameter of approximately 1.17\(\times10^{-4}\) s\(^{-1}\). The terms on the left hand side of Eq. 4.5 are the contributions of advection of momentum and local acceleration. The first two terms on the right hand side are the expressions for the longitudinal pressure gradient; the first term represents the water level gradient in longitudinal direction and the second term the longitudinal density gradient. The last terms are the Reynolds stress gradients in longitudinal, transverse and vertical directions and the Coriolis force. The turbulent pressure term \(\frac{\partial \bar{u}^2}{\partial x}\) can be neglected because it is considered small compared to the other pressure terms. Similar equations can be derived for the other co-ordinate directions.

In this study only the vertical and longitudinal velocity components were measured. Therefore only two normal stresses (\(\bar{u}^2\) and \(\bar{w}^2\)) and one shear stress (\(\bar{uw}\)) can be calculated. An order of magnitude estimation is used to compare \(\bar{u}v\) to \(\bar{uw}\). This is done based on the simplifying assumptions:

\[
\frac{\partial \bar{uw}}{\partial z} \approx \frac{U^2}{h} \tag{4.6}
\]

\[
\frac{\partial \bar{uv}}{\partial y} \approx \frac{U^2}{B} \tag{4.7}
\]
\frac{\partial u^2}{\partial x} = \frac{U^2}{L}, \text{ and } L \gg h, B \gg h \quad (4.8)

\text{U is a characteristic turbulence velocity scale, } B \text{ and } L \text{ are characteristic length scales for the vertical, transverse and longitudinal directions, respectively. Typical of the hydrodynamic conditions in the Dollard is that } h \text{ is much smaller than the other two length scales: } h \text{ is of the order of a few meter while } B \text{ and } L \text{ are several hundreds metres.}

\text{Keeping only the largest turbulent stresses, and neglecting advective acceleration because of low Froude numbers justifies the depth-integrated form of Eq. 4.5 (McDowell & O' Connor, 1977):

\begin{equation}
\frac{h}{\partial x} \frac{\partial U_d}{\partial x} + gh \frac{\partial n}{\partial x} + \frac{gh^2}{2 \rho_d} \frac{\partial \rho_d}{\partial x} + \frac{\tau_b}{\rho_d} = 0
\end{equation} \quad (4.9)

\text{The subscript } d \text{ denotes the depth averaged values. For the calculations of the bed shear stress (} \tau_b \text{) the following formula was applied:

\begin{equation}
\tau_b/\rho_d = C_d U_{10} |U_{10}|
\end{equation} \quad (4.10)

\text{where } C_d \text{ is a friction factor and } U_{10} \text{ is the longitudinal velocity at the level of } 1.0 \text{ m above the sediment bed.}

\textbf{Longitudinal density gradient}

\text{The longitudinal density gradient is often referred to as a driving force for the landward transport of suspended sediment: the so-called estuarine circulation (see Section 1.3). For a salinity gradient with decreasing salinities towards the head of the estuary, water with a relatively high salinity is forced in the landward direction in the near-bed region (see also Eq. 4.5). For reasons of conservation of mass less saline water is moving seaward. For a longitudinal sediment concentration gradient with higher levels in the landward direction the flow situation is reversed.

\text{Salinities were recorded in 1996 during high water periods at the BOA measuring platform on the Heringsplaat at approximately +0.2 m N.A.P.. Between April and September 1996 the salinity varied (gradually) between 20 \%o and 25 \%o. The lower values were found in April and September when the discharge of the River Ems was relatively large. These small variations at least indicate that the salinity distribution during this part of the year is not very sensitive to variations in tidal range, short term variations in river discharge, wind direction, lateral mixing etc. Figure 4.7 shows some}
results of these measurements for four tidal cycles. As expected, only small variations in salinity are present during the high water periods on the tidal flat.

Dorrestein (1960) reports that a substantial part of the Dollard water is well mixed vertically and horizontally. In August 1954 a homogeneous water mass of ca. 13.5 % salinity was present and in August 1956 the salinity was approximately 7.8 %. The differences between these salinities and the salinities shown in Figure 4.7 can be explained from the differences in river discharge at Versen: about 130 m$^3$/s in August 1954, 300 m$^3$/s in August 1956, and only 25 m$^3$/s on average in the summer of 1996. Thus even at relatively high river discharges a considerable amount of Dollard water remains well mixed. At low river discharges this amount is most likely to increase.

![Figure 4.7. Salinities, S, and water levels, h, recorded during high water periods in June at the BOA measuring platform on the Heringsplaat 400 m South West of measuring location in the channel: -- , S; --- , h.](image)

An upper limit of the longitudinal density gradient calculated from the salinity gradient measured in the Emden Fahrwasser can be estimated at approximately $10^3$ kg/m$^3$/m (Van Leussen, 1994). Numerical simulations of Robaczewska (1992) show that the longitudinal density gradient is probably less. Even if this upper limit is applied, approximations of the different terms of Eq. 4.9 show that density driven currents are only expected to occur during periods of high water (see also Figure 4.8). The water depths in the Dollard are too small on average for the longitudinal density gradients to have a significant effect on the water motion. However, during some parts of the tide, especially during periods of low current velocities, some stratification can be present in the upper part of the water column resulting from fresh water input from the sluices of Nieuwe Statenzijl during LW or from the Ems via the Geisedam or Mond van de Dollard during HW.
Acceleration effects

Figure 4.8 shows the different terms in Eq. 4.9 for a tidal period during the measuring period in June 1996. The water level gradient was calculated from the total sum of the other terms of Eq. 4.9. Figure 4.8 shows that, generally speaking, the bed friction balances the pressure caused by the water level gradient. However, acceleration is not negligible, especially not during HW at the beginning of the ebb.

Measurements have shown that accelerating tidal flows have lower turbulence intensities and lower Reynolds stresses than decelerating tidal flows, resulting in a loop of hysteresis curves with the stress lagging behind the velocity as much as 22.5 to 27 degrees. (Anwar and Atkins, 1980). In Anwar’s experiment the acceleration ranged from $1.4 \times 10^3$ to $3.8 \times 10^3$ m/s², and the channel depth was 0.4 m.

![Figure 4.8. The terms of Eq. 4.9 during a representative tidal period in June 1996 using only the tidal components $M_2$, $M_4$, $M_6$, $S_2$ and $K_1 + O_t$ of the flow velocities and the water levels obtained at the Rijkswaterstaat pole Dordrecht. For the longitudinal density gradient an upper limit was estimated (see previous section).](image)

Kuo et al. (1996) showed seemingly contradicting results. Kuo et al. solved Eq. 4.5 numerically for an hypothetical estuary of 500 m width and 10 m depth with a second order model for the turbulence closure. The length of the estuary was 160 km for a $M_2$ constituent forcing and 40 km for a $M_4$ constituent forcing. The maximal acceleration was $1.4 \times 10^4$ m/s². Kuo et al. (1996) found that the bed shear stress is stronger during the acceleration phase of the tidal cycle than during the deceleration phase, and that the stress leads the velocity. These findings are in qualitative agreement with models described by Soulsby and Dyer (1981) and Lavelle and Mofjeld (1983). Kuo et al. explain the differences between his simulations and the data of Anwar and Atkins from the large spatial gradient of the unsteady boundary turbulence structure: during acceleration the Reynolds stresses increase monotonically from a certain (small) value at the surface to the maximum value near the bed, during deceleration the maximum Reynolds stress is away from the bed (See also Anwar and Atkins, 1980). Therefore
the phase shift between surface velocity and the shear stress depends to a large extent on the height where the value of the shear stress is measured. The shear stress measured near the bed leads the surface velocity and the shear stress measured at some distance from the bed lags behind the surface velocity.

Figure 4.9. The shear stresses measured at the elevations 0.1 m and 0.4 m ($z/h < 0.2$) versus the measured velocities at 1.0 m during the first complete tidal cycles of the August period: -o-, acceleration; -x-, deceleration.

In this research the stresses were measured in the near-bed region ($z/h < 0.2$). Figure 4.9 shows the measured shear stresses versus the velocity measured at 1.0 m. Lag effects are hardly visible. One could conclude that the shear stresses are somewhat smaller during acceleration than during deceleration at the 0.4 m elevation, but the differences between the stresses are within the error band. This finding would not contradict Kuo's (1996) and Soulsby and Dyer's (1981) models.

4.4 Reynolds stresses, turbulence intensities and turbulent sediment transport
Turbulence intensities and Reynolds stresses have been measured in previous research work on open channel flow. The results of this work were summarised by Nezu and Nakagawa (1993). Most of the research was carried out in well controlled laboratory flumes. It is expected that laboratory research can be used for estimations of turbulence properties in full scale tidal flows on basis of the universality of turbulence, provided that the Reynolds number in the laboratory is large enough. However, tidal flows always show unsteadiness, large ratios of horizontal to vertical scales, and carry various kinds of sediment. Salinity or temperature induced density stratification make the turbulence structure even more complex. Various inertial sub-ranges may exist as
the horizontal shear flow is characterized by the channel width and the vertical shear flow by the water depth (Yokoshi, 1967, in Nezu and Nakagawa, 1993).

**Shear stresses**

Figures 4.10a and 4.10b show the Reynolds stresses (10 minutes averages) obtained from the measurements made in June and August 1996, respectively. Positive values coincide with the flood parts of the tidal cycles. High-frequency losses were found to be negligible (< 1%). Low-frequency cut-off losses were examined by lowering the record length from 10 minutes to 1 minute. It was found that for the 0.1 m and 0.4 m level, respectively, approximately 94% and 87% of the shear stresses was recovered (averaged over June and August). For the 1.0 m level this value was 92% (June period). The random errors, shown in Figure 4.10 as vertical bars, are the standard deviations of the mean (SDOM; Bendat and Piersol, 1971) evaluated over 10 minute periods. One minute averaged Reynolds stresses were used as independent estimates. The x-channel of the highest positioned EMF sensor showed significant interference in August and therefore results for $z = 1$ m are not shown in Figure 4.10b.

![Figure 4.10](image)

Figure 4.10. Time series of 10 minute averaged Reynolds stresses during neap tides: (a) $-\overline{uw}$ at 0.1 m, 0.4 m and 1.0 m, June 1996; (b) $-\overline{uw}$ at 0.1 m and 0.4 m, August 1996.
On average the shear stresses measured in June are somewhat larger than the shear stresses in August. This is in qualitative agreement with the slightly larger velocities in June (see Figures 4.1 and 4.2). In Fig. 4.10a the maximum shear stresses of the first and third flood are somewhat larger than the maximum stresses of the second and fourth flood. These variations in shear stress are in agreement with the variations of the flow velocity (see Figure 4.1). Moreover, in Figure 4.10b, the maximum shear stresses of the first and the third ebb are somewhat larger than that of the second ebb which is in agreement with the variations in the ebb velocities (see Figure 4.2).

Figure 4.10 shows that the absolute values of the Reynolds stresses at the lowest sensor positions \(|\bar{uw}|_{0.1}\) are the largest for both measuring periods. The small difference in \(|\bar{uw}|_{0.1}\) between ebb and flood is not in agreement with the ebb domination in the tidal velocities (see Section 4.2). Since the ebb velocities are about 20% larger than the flood velocities the stresses during ebb are expected to be about 40% to be larger than the stresses during flood at the same level. The shear stresses measured at the elevations of 0.4 m and 1.0 m show the expected ebb dominance. Consequently, the shear stresses at the higher elevations are not in agreement with \(\bar{uw}_{0.1}\). For quasistationary conditions a linear distribution of the shear stress is expected:

\[
\bar{uw}(z) = -\bar{u}_e |\bar{u}_e| (1 - z/h) \tag{4.11}
\]

If the water depths as determined from Figures 4.1 and 4.2 are substituted in Eq. 4.11, the stresses at the elevations 0.4 m and 1.0 m are approximately 85% and 70% of \(\bar{uw}_{0.1}\), respectively. During flood in June and August these percentages are of the order of about 50%. The large vertical gradient of the shear stress in the near-bed region during these periods can only be explained from a local scour hole enhancing the shear stresses at 0.1 m during flood, or from erroneous measurements. Both possibilities are discussed below. For now, it is assumed that the bed shear stress evaluated from the shear stress measured at 0.4 m and Eq. 4.11 is more representative of the bed shear in the tidal channel than \(\bar{uw}_{0.1}\).

Figure 4.11 shows that the bed shear stress \(\bar{u}_e |\bar{u}_e|\) calculated from the measured shear stress in the 0.4 m level and Eq. 4.11 coincides with the squared velocity \(U_{1.0}\) if multiplied by a factor \(C_d = 1.8 \times 10^{-3}\). This value for \(C_d\) corresponds with a roughness height of approximately 2 mm which is in agreement with roughness heights for flow over a cohesive sediment bed (Van Rijn, 1993). Bassoulet et al. (1992) found values for \(C_d\) of 1\times10^{-3} to 2\times10^{-3} for a tidal flat in the Penzé estuary (Brittany, France), Shiono and West (1987) found values for \(C_d\) of 2\times10^{-3} to 4\times10^{-3} for a flat bed composed of fine sediments (with \(U_d\) instead of \(U_{1.0}\)). Visual observations of the
borders of the tidal flat *Heringsplaat* during LW spring showed that bed forms were almost absent at levels below N.A.P. (See also Section 3.2).

![Graph showing bed shear stress](image)

Figure 4.11. The bed shear stress $u|\nu|_y$ calculated from $\bar{uw}_{0.1}$, and the bed shear stress calculated from Eq. 4.10 ($C_d = 1.8 \times 10^3$), for two representative tides of the June and August measuring periods.

As stated before, the deviating values of $\bar{uw}_{0.1}$ can be explained from at least two different effects, i.e. a scour hole, or from erroneous measurements. In Section 4.2 it was concluded from the velocity profiles that the lowest EMF was not completely outside the scour hole surrounding the pole. This may have enhanced the turbulence intensities and shear stresses. The scour hole is confined to a small zone surrounding the pole, with typical dimensions of a few times the diameter of the pole, say 2 m. If the boundary layer growth $\partial \delta / \partial x$, where $\delta$ is the boundary layer thickness, is estimated at 1/20 to 1/50 (Schlichting, 1968), it is clear that the sensors at elevations of 0.4 m and higher were outside the influence of such a scour hole. Not explained is the fact that the deviating values of $\bar{uw}_{0.1}$ occur only during flood. A conclusive answer cannot be given because of the lack of visual observations of the bed area directly underneath the measuring frame.

Deviating values of $\bar{uw}_{0.1}$ could also result from bias errors. Bias errors originate, amongst other things, from interference between $u$ and $w$ signals and sensor rotation. Some effects, which might be due to interference, are shown in Figure 4.10a: the shear stress at the 1.0 m level shows a constant offset of approximately $-7 \times 10^{-5}$ m$^2$/s$^2$ during periods of slack water. This effect is negligible for the other sensors at 0.1 m and 0.4 m. In August this offset was $-2 \times 10^{-4}$ m$^2$/s$^2$ (not shown). The origin of this effect is discussed in Appendix B.
Sensor rotation can have a significant effect on the values of the Reynolds stresses (Haines and Gelfenbaum, 1996). The shear stress \( \overline{uw} \) in the horizontal level can be calculated from the following expression (analogue with Mohr’s circle, see also Hinze 1975):

\[
\overline{uw} = \frac{\overline{u^2} m + \overline{w^2} m}{2} \sin 2 \varphi + \overline{uw}_m \cos 2 \varphi
\]  
(4.12)

The subscripts \( m \) denote the measured turbulence and \( \varphi \) is angle of rotation. Semi-theoretical distributions for the turbulence intensities are (Nezu and Nakagawa, 1993)

\[
u'/u_* = 2.30 \exp (-z/h)
\]
(4.13)

\[
w'/u_* = 1.27 \exp (-z/h)
\]
(4.14)

The constants 2.30 and 1.27 were obtained from laboratory experiments over smooth beds and proved to be independent of the Reynolds number (in the range \( 3 \times 10^4 \) to \( 2.5 \times 10^5 \)) and the Froude number (in the range 0.5-3.1). Using Eq. 4.12, 4.13 and 4.14, an expression for the bias error caused by rotation \( (\varepsilon_{br}) \) can be derived for small angles \( \varphi \):

\[
\varepsilon_{br} \approx \frac{\overline{uw}_r - \overline{uw}_m}{u_*^2} \approx -3.7 \varphi \exp (-2 z / h)
\]
(4.15)

Anisotropy increases towards the bed and thereby the relative importance of an accurate orientation of the sensor. The uncertainty in sensor angle is estimated at approximately \( \pm 0.1 \) rad. According to Eq. 4.15 this would lead to bias errors of the order of 40% in the near-bed region. An important aspect of sensor rotation is that its effect on the shear stresses is reversed when the flow direction reverses. Application of Eq. 4.12 directly to the second and third tidal cycles of the August measuring period shows that \( |\overline{uw}_{r1}| \) increases by 18% during ebb and reduces by 35% during flood for \( \varphi = -0.1 \) rad. The results are shown in Figure 4.12. The effects of the relatively small value of rotation angle on \( \overline{uw}_{r1} \) are not as large as expected on the basis of Eq. 4.15 but are still appreciable.

Despite the good match between the “rotated shear stress” and the bed shear stress \( \tau_b/\rho_b \), sensor rotation provides no satisfying explanation for the deviating behaviour of the measured values of \( \overline{uw}_{r1} \). Rotation also affects the values for the correlation coefficients \( R_{uw} \). They were approximately 0.4 without rotational correction, but change into 0.3 for flood and 0.5 for ebb, for \( \varphi = -0.1 \) rad. These values for \( R_{uw} \) seem unrealistic. It can be concluded that the Reynolds stresses and the correlation
coefficients $R_{uv}$ can be easily manipulated. It appears likely that the deviating behaviour of $uw_{01}$ is real and is probably due to locally generated turbulence.

Figure 4.12. The effect of sensor rotation on $uw_{01}$ for two representative tidal cycles during August 8, 1996; $\times$, measured values of $uw_{01}$; $\Delta$, $uw_{01}$ for $\varphi = -0.1$ rad.; $---$ $\tau_b/\rho_d = 0.0018U_{10}^2$.

Turbulence intensities

Figure 4.13 shows times series of the turbulence intensities for two tidal cycles during August 1996. The horizontal lines in Figure 4.13 represent the semi-theoretical relations Eq. 4.13 and Eq. 4.14. The turbulence fluctuations were corrected for high-frequency losses which are due to sensor size and limitations of sampling rate. The losses were determined by extending the tails of the spectra of two representative flood, and two representative ebb records from the June measuring period (Soulsby, 1980; see also Section 3.3.3). They were found to be 13%, 8% and 5%, respectively, for $\bar{w}^2$ at the 0.1 m, 0.4 m and 1.0 m levels, and 5%, 3% and 3% for $\bar{u}^2$ at the same levels. High-frequency losses at the 0.1 m level are not as dramatic as expected on the basis of Figure 3.6 (see Section 3.3.3). The reason is that the spectral density functions scale with the integral scale $L_x$ which varies with $\sqrt{x}$ (Nezu and Nakagawa, 1993) instead of $x$. The shift of the spectral density functions towards the higher wave numbers for measurements close to the bed is therefore less than expected. In Section 4.5 this subject is discussed in more detail. The random errors were determined as described in the previous section: the SDOM for $\bar{u}^2$ and $\bar{w}^2$ varied between 5% and 10%, but increased up to 20% for slack water periods.

Figure 4.13a shows the time histories of the longitudinal turbulence intensities $u'/u_*$ and vertical turbulence intensities $w'/u_*$ at the 0.4 m level for two tidal cycles. The
values for $u/u_*$ and $w/u_*$ are comparable with the laboratory values except for periods of slack water. During such periods the turbulence intensities $u/u_*$ and $w/u_*$ increase drastically. Similar effects were reported by French and Clifford (1992). They showed that these effects coincide with a reduction of the correlation coefficient for the Reynolds stress.

![Graphs showing turbulence intensities](image)

**Figure 4.13.** Turbulence intensities of the vertical and longitudinal velocities normalised by the friction velocity $u_*$ of two tides during August 8, 1996: (a) 0.4 m level, $z/h$ varies between 0.1 and 0.2; (b) 1.0 m level, $z/h$ varies between 0.2 and 0.6; $\times$, $u'/u_*$; $\Delta$, $w'/u_*$; ---, Eq. 4.13; -- - - - -, Eq. 4.14.

Figure 4.13b shows the time history of $w'/u_*$ at the 1.0 m level. The relative height of this level varies throughout the tidal cycle: $z/h = 0.2$ for high water and $z/h = 0.6$ for low water. The “theoretical” variation of $w'/u_*$ according to Eq. 4.14 is also plotted in Figure 4.13b. The variation of the measured $w'/u_*$ is very similar to the theoretical variation and the intensities are also in agreement.

Table 4.2 summarises the averaged values of the turbulence intensities for the complete June and August measurements. The large uncertainties in $w'/u_*$ and $u'/u_*$ for the August period at the 0.1 m level result from differences between ebb and flood as discussed in the previous section. The uncertainty of 0.4 for $u'/u_*$ at the 1.0 m elevation results not only from changes in $z/h$ but largely from interference. The values for $u'/u_* = 2.3$ and $w'/u_* = 1.5$ at the 0.1 m level are probably too large because of the extra turbulence generated by the scour hole. Nevertheless, the turbulence intensities
measured in June and August are, generally speaking, in agreement with the intensities obtained from laboratory data.

Table 4.2. Averaged turbulence intensities for the June and August measuring periods ($\bar{U} > 0.1$ m/s), and some laboratory results (Nezu and Nakagawa, 1993).

<table>
<thead>
<tr>
<th></th>
<th>0.1 m level</th>
<th>0.4 m level</th>
<th>1.0 m level</th>
</tr>
</thead>
<tbody>
<tr>
<td>$z/h = 0.01 - 0.02$</td>
<td>$u'/u^*$ = 2.3 ± 0.1</td>
<td>$w'/u^*$ = 1.5 ± 0.1</td>
<td>$w'/u^*$ = 2.1 ± 0.1</td>
</tr>
<tr>
<td>$z/h = 0.08 - 0.13$</td>
<td>$u'/u^*$ = 2.2 ± 0.1</td>
<td>$w'/u^*$ = 1.2 ± 0.1</td>
<td>$w'/u^*$ = 1.1 ± 0.1</td>
</tr>
<tr>
<td>$z/h = 0.22 - 0.34$</td>
<td>$u'/u^*$ = 1.9 ± 0.4</td>
<td>$w'/u^*$ = 1.2 ± 0.1</td>
<td>$w'/u^*$ = 1.1 ± 0.1</td>
</tr>
</tbody>
</table>

Field measurements:

<table>
<thead>
<tr>
<th></th>
<th>0.1 m level</th>
<th>0.4 m level</th>
<th>1.0 m level</th>
</tr>
</thead>
<tbody>
<tr>
<td>June 1996</td>
<td>$u'/u^*$ = 2.3 ± 0.1</td>
<td>$w'/u^*$ = 1.3 ± 0.1</td>
<td>$w'/u^*$ = 1.1 ± 0.1</td>
</tr>
<tr>
<td>August 1996</td>
<td>$u'/u^*$ = 2.4 ± 0.3</td>
<td>$w'/u^*$ = 1.5 ± 0.2</td>
<td>$w'/u^*$ = 1.1 ± 0.1</td>
</tr>
<tr>
<td>Laboratory data1:</td>
<td>$u'/u^*$ = 2.3 ± 0.1</td>
<td>$w'/u^*$ = 1.3 ± 0.1</td>
<td>$w'/u^*$ = 1.1 ± 0.1</td>
</tr>
</tbody>
</table>

1 smooth conditions

Shiono and West (1987) carried out field experiments in the Conwy estuary in Wales. They found, at $z/h = 0.2$, values of 1.8 for $u'/u^*$ and values of approximately 1.0 for $w'/u^*$. Kawanisi and Yokosi (1993) report values 2.5-3.0 for $u'/u^*$ and 1.0-1.1 for $w'/u^*$ at $z/h = 0.03$ - 0.1. French and Clifford (1992) show values of approximately 2.5 ± 0.4 for $u'/u^*$ and 1.1 ± 0.3 for $w'/u^*$ at $z/h = 0.1$ for steady flow conditions. These data in combination with the data presented in Table 4.2 seem to confirm the universality of the expressions Eq. 4.13 and Eq. 4.14.

Some remarks must be made herein. The laboratory data were obtained from experiments with hydraulically smooth beds. Nezu and Nakagawa (1993) report that for “hydraulic rough conditions” the turbulence becomes less anisotropic. The proportionally factor in Eq. 4.13 reduces from 2.3 to 2.0. The values shown in Table 4.2 for the 0.1 m level may therefore be somewhat too large. For separating flows and reattaching flows the ratio $u'/w'$ will reduce to unity. The ratio of $u'/w'$ of 1.5 at the 0.1 m level shows that the turbulence at this level is still representative of boundary layer flow conditions.

**Turbulence characteristics of sediment transport**

Figures 4.14 and 4.15 show some turbulence characteristics of the SSC during two tidal cycles of August 8th 1996. High-frequency losses were corrected for. The standard deviations of the means of $\overline{c^2}$, $\overline{cw}$; and $\overline{cu}$, were approximately 20%, 20% and 10%, respectively. Low-frequency cut-off errors were examined and found to be small for $c^2$ and $cw$ (< 10%). On the other hand, low-frequency cut-off errors for $cu$
were significant. The losses for $\overline{cu}$ varied between 20% and 50% when the data analysis was based on 1 minute data recordings instead of 10 minute recordings. Although contributions to $\overline{cu}$ from flow variations associated with time scales larger than 10 minutes were expected, larger record lengths were not examined since pulsations of the mean flow itself would then almost certainly be incorporated in the ensemble means. Therefore the choice for a record length of 10 minutes for the determination of $\overline{cu}$ is rather arbitrary, in contrast to the other turbulence parameters for which a distinction could be made between turbulence contributions and variations of the tidal current.

Figure 4.14. Quantities related to the suspended sediment concentration for two tidal cycles during August 8th 1996: (a) Mean suspended sediment concentrations measured by the MEX at 0.3 m, 0.7 m and 1.4 m; (b) suspended sediment gradient $\partial \overline{C} / \partial z$ determined from Eq. 4.3; (c) mean squared turbulence fluctuations $\overline{e^2}$ at 0.1 m and at 0.4 m.
Figure 4.14 shows that variations of the mean concentration, the concentration gradient and the mean squared fluctuations of the concentration are closely related. There is clearly a correlation between the peaks in the SSC, gradients and fluctuations. However, it also shows that, for example, at 9:00h August 8th peaks in concentration gradients follow from relatively small variations in the MEX recordings which are well within the error band (plotted in Figure 4.14a). These particular peaks are not reflected in the variation of turbulence fluctuations, which could point to the unrealistic nature of the gradients, or to large dampening of turbulence. Either way, the calculated gradients at 9:00h August 8th have no significance and are therefore not used in the analysis of density stratification effects (Section 4.7). The relative turbulence intensities of the concentrations are between 0.1 and 0.2, and are comparable to those of the longitudinal velocity.

![Graph](image)

Figure 4.15. Turbulent transports of suspended sediment for two tidal cycles during August 8th 1996: (a) vertical suspended sediment transports $\overline{cw}$ at 0.1 m and at 0.4 m; (b) horizontal suspended sediment transports $\overline{cw}$ at 0.1 m and 0.4 m.

The sign of the turbulent fluxes suggested that upward moving fluid parcels, which moved slowly in the longitudinal direction, contained relatively high concentrations, and downward moving fluid parcels which moved fast in the longitudinal direction, contained relatively low concentrations. This finding is in agreement with the concept
of gradient transport type models: "random" motions on average transport the quantity of interest in direction of the lower concentrations. The ratio of the fluxes $\overline{cu}/\overline{cw}$ is between 2 and 4 which clearly demonstrates the anisotropy caused by the influence of the bed. For flows remote from beds or walls this ratio is of the order 1 (Lauder, 1975). In Section 4.8 the nature of the turbulent motions is discussed in more detail.

For some parts of the tidal cycle it can be assumed that the upward turbulent sediment transport ($\overline{cw}$) balances the downward transport of suspended sediment from settling ($w_s C$). In these parts of the tidal cycle the temporal variations in SSC are small. Periods around 11:00h and 17:00h August 8th shown in Figure 4.14 and Figure 4.15 could be representative of such parts. The upward turbulent sediment transport $\overline{cw}$ varies between $0.5 \times 10^{-4}$ kg/m$^2$/s and $1.5 \times 10^{-4}$ kg/m$^2$/s, and SSC varies between 0.2 g/l and 0.5 g/l. From these values it follows that $w_s$ varies between 0.1 and 0.8 mm/s during the periods mentioned. In Chapter 5 a more sophisticated model will be used for estimating important key-parameters of the suspended sediment transport, such as the settling velocity.

4.5 Spectral analysis

Spectral analysis can give some insight into the spatial structure of the turbulence of sediment laden tidal flow. First some definitions are presented, following closely the work of Nezu and Nakagawa (1993) and Nieuwstadt (1992). Second, some remarks are made about the computational method for the determination of the spectral distributions and correlation functions. Third, estimates of spectral distributions for some representative parts of the tidal cycle are presented and discussed.

The spectral distribution $S_q(k)$ is defined as the Fourier transform of the correlation function $R_q(r)$:

$$S_q(k) = \frac{1}{2\pi} \int_{-\infty}^{\infty} R_q(r) e^{-ikr} dr$$

(4.16)

$$R_q(r) = \frac{\xi_q(x)\xi_q(x+r)}{\xi_q^2} = \int_{-\infty}^{\infty} S_q(k)e^{ikr} dk$$

(4.17)

where $r$ is the streamwise lag distance, $k$ is the wave number, $\xi_q(x)$ can be any record of $u(x)$, $w(x)$, or $c(x)$. If $r$ is set to zero, Eq. 4.17 reduces to:

$$R_q(0) = \int_{-\infty}^{\infty} S_q(k)dk = 2\int_{0}^{\infty} \text{Re}(S_q(k)) dk$$

(4.18)
\( R_u(0) \) is equal to 1 for \( j = i \), and between -1 and 1 for \( j \neq i \). If \( k \) is set to zero, and \( \xi(x) = \xi(0) = u(x) \), \Eq{4.16} reduces to:

\[
\int_{0}^{\infty} R_u(r)dr = \pi S_{uu}(0) = I_x
\]

(4.19)

where \( I_x \) is the integral length scale which characterises the dimension of the large eddies. It has been found that \( I_x \) scales with \( \sqrt{z} \) (Nezu, 1977):

\[
I_x = B_1 \sqrt{2h} \text{ for } z/h < 0.6
\]

(4.20)

\( B_1 \) is approximately 1.0 for a Reynolds number \( R_* (= u_h/\nu) \) of 1600, and 1.1 for \( R_* = 600 \).

For isotropic turbulence the dissipation rate \( \varepsilon \) is equal to (Tennekes and Lumley, 1972):

\[
\varepsilon = 15\nu\left( \frac{\partial u}{\partial x} \right)^2
\]

(4.21)

By differentiating \Eq{4.17} twice with respect to \( r \) for \( \xi(x) = \xi(0) = u(x) \), and then setting \( r = 0 \) gives (Nezu and Nakagawa, 1993):

\[
-\frac{d^2 R_u(r)}{dr^2} \Bigg|_{r=0} = \frac{1}{u'^2} \left( \frac{\partial u}{\partial x} \right)^2 = \int_{-\infty}^{\infty} k^2 S_{uu}(k) dk
\]

(4.22)

and from \Eq{4.21} and \Eq{4.22} one obtains:

\[
\int_{-\infty}^{\infty} k^2 S_{uu}(k) dk = \frac{1}{\lambda^2} = \frac{\varepsilon}{15\nu u'^2}
\]

(4.23)

where \( \lambda \) is defined as the Taylor microscale, \( k^2 S_{uu}(k) \) is defined as the dissipation spectrum.

A spectral distribution for the productive and inertial subranges (\( k \ll \eta^{-1}, \eta \) is the Kolmogorov length scale, defined in Section 3.3.3) can be approximated by (Von Kármán’s spectrum):

\[
S_{uu}(k) = \frac{1}{\pi} I_x \left( 1 + \left( \frac{k}{k_0} \right)^2 \right)^{-5/6}
\]

(4.24)

where \( k_0 \) is a relatively small wave number. The estimated spectra obtained from the field measurements are compared to this spectral distribution. If \( k \) becomes
significantly smaller than \( k_0 \), \( S_\omega \) approaches \( L_\omega / \pi \) (see also Eq. 4.19). For large \( k \), \( L_\omega^{-1} \ll k \ll \eta^{-1} \), \( S_\omega(k) \) approaches the \("-5/3 power law":\)

\[
S_\omega(k) = \frac{1}{\pi} \frac{L_\omega}{k_0} \left( \frac{k}{k_0} \right)^{-5/3} = \frac{\alpha_1}{u^2} e^{2/3} k^{-5/3}
\]

(4.25)

\( \alpha_1 e^{2/3} k^{-5/3} \) is the Kolmogorov power spectrum and \( \alpha_1 \) is the universal Kolmogorov constant. From experiments \( \alpha_1 \) was found to be 0.26 (Nieuwstadt, 1992). From Eq. 4.25 it follows that:

\[
s = K \frac{u^3}{L_\omega}
\]

(4.26)

with \( K \) given by Nezu and Nakagawa (1993):

\[
K = \left( L_\omega k_0 \right)^{\frac{1}{2}} \left( \alpha_1 \pi \right)^{-\frac{3}{2}} \approx 0.7 \text{ for } R_L (= L_\omega u'/v) > 5 \times 10^4
\]

(4.27)

\( L_\omega k_0 \) is then approximately 0.77 (see also Hinze, 1975). Von Kármán's spectrum (Eq. 4.24) only depends on the constant \( B_1 \) if this value for \( L_\omega k_0 \) is used together with Eq. 4.20. \( B_1 \) can be found from fitting estimated spectral distributions \( S_\omega \), obtained from the measurements at different heights above the channel bed, to Von Kármán's spectrum. Eq. 4.20, Eq. 4.23, Eq. 4.26, \( u' \) and \( B_1 \) can be used to estimate the dissipation rate \( \varepsilon \) and Taylor microscale \( \lambda \) from the spectra.

**Computation of the spectral distributions**

In practice we have a sequence of \( N \) samples of \( \xi_t \) at discrete values of \( t = t_0 + n \Delta t \) denoted as \( X_{t_n} \). Estimates for the (continuous) power spectra can be obtained from the modified periodogram \( I_{N,q}(\omega_p) \). The latter can be calculated from the Fourier transforms of \( X_{t_n} \) (Bendat and Piersol, 1971):

\[
I_{N,q}(\omega_p) = \frac{\Delta t}{2 \pi N} \zeta_{X_t}(\omega_p) \zeta_{X_t}^*(\omega_p) \quad p = 1, \ldots, N
\]

(4.28)

with

\[
\zeta_{X_t}(\omega_p) = \sum_{n=1}^{N} X_{t_n} e^{-i \omega_p n \Delta t} \quad n = 1, \ldots, N
\]

(4.29)

where \( \omega_p = 2 \pi p / T \) is the angular frequency, where \( T = N \Delta t \) is the length of time of the record, \( \zeta_{X_t}(\omega_p) \) is the finite Fourier transform of \( X_{t_n} \). \( \zeta_{X_t}^*(\omega_p) \) is the complex conjugate of \( \zeta_{X_t}(\omega_p) \). \( \zeta_{X_t}(\omega_p) \) was computed by using a “Fast Fourier Transform” algorithm. Estimates of the spectral distributions for records of \( M \) samples, were
determined from an ensemble average of modified periodograms $I_{N_{i},y}(\omega_x)$ which were computed for segments of length $N$, with $N < M$. A “Hanning window” was applied to the sections prior to computing the periodogram in order to minimise “spectral-leakage” and segment overlap was used in order to reduce the variance of the estimates. (Welch’s method, in: Kay, 1988). If the “frozen-turbulence” approximation is applied (Taylor Hypothesis):

$$\frac{\partial}{\partial t} = -\overline{U} \frac{\partial}{\partial x} \quad \text{(4.30)}$$

the relation between the spectral estimate $\hat{S}_y(k_p)$ and $I_{N_{i},y}(\omega_x)$ becomes:

$$\hat{S}_y(k_p) = N \overline{U} \sum_{i=1}^{M/N} I_{N_{i},y}(\omega_x) \frac{X_{i,\ell,x}}{X_{i,t,x}} \quad \text{(4.31)}$$

where

$$k_p = \omega_x \sqrt{\overline{U}}, \quad p = 1, \ldots, N \quad \text{(4.32)}$$

and $X_{i,t,x}$ is the “rms”-value of $X_{i,t,x}$.

In order to minimise the bias of the estimates $S_y(k)$ for small values of $k_p$, the segments should be statistically independent. (The circumflex is dropped for the sake of convenience.) In other words, $R_y(r)$ should go to zero for large values of $r$. This condition does not yield the same results for all components of $R_y(r)$ as, for example, contributions of large scale variations to $R_{uw}(r)$ are expected, but not to $R_{ww}(r)$ (see also Section 4.4). An indicative measure for $T$ can be derived from the integral scale $L_x$. According to Nezu and Nakagawa (1993) $L_x$ can be approximated by (Eq. 4.20):

$$L_x \approx \sqrt{zh} \quad \text{for } z/h < 0.6$$

$L_x$ is about 2 m for $h = 4$ m and $z = 1$ m. Applying Eq. 4.30 yields:

$$T \gg \frac{L_x}{\overline{U}} \quad \text{(4.33)}$$

The flow velocity $\overline{U}$ was larger than 0.2 m/s for the records processed, which means that $T$ should larger than 10 seconds. A value of $T = 51.2$ s was selected.

The number of segments determines to a large extent the reliability of the estimate $S_y(k)$ provided that $\xi(k)$ is stationary. The stationarity of the records of $u$, $w$ and $c$ of length $M$ was checked by dividing them into 0.1$M$ segments and then applying a stationarity test (Bendat and Piersol, 1971; West et al., 1986). The hypothesis of stationarity is accepted at the 95% confidence level if the number of runs (zero
crossings) observed in sequence of the turbulence parameter of interest relative to the median is at least 3 but no more than 8. Figure 4.16 shows an example of the test for \( u' \) over a 600 second period. 7 runs are observed in sequence. Hence, this data record can be considered stationary. The stationary test is not absolute, but is merely a helpful tool for selecting appropriate record lengths for the estimation of spectral distributions and correlation functions.

![Graph](image_url)

Figure 4.16. Stationarity test of a record of \( u' \) at the 0.1 m level (16:11h-16:21h, 6/26/1996).

**Estimates of Spectra**

Examples for the power spectra of \( u \) and \( w \) are shown in Figure 4.17. The spectra are obtained from velocity measurements at three heights above the channel bed during maximum ebb velocity in June 1996. The spectra are normalised by the integral scale \( L_c \) (Eq. 4.20). The Von Kármán’s spectrum and the transverse spectrum for isotropic turbulence are also included in Figure 4.17.

The \( u' \)-power spectra shown in Figure 4.17a collapse to a single line for \( kL_c > 0.5 \) and are, to a certain extent, in agreement with Von Kármán’s spectrum if \( B_1 \) is set to 0.8 \( (R_e = 8.8 \times 10^4) \). This value for \( B_1 \) is representative of other spectral estimates during periods of maximum ebb and flood currents if the conditions are approximately stationary. It is less than the values of about 1 recommended by Nezu (1977) for low \( R_e \) (Eq. 4.20). However, Nezu pointed out that \( B_1 \) decreases slightly as the Reynolds number increases.

Some deviations from Von Kármán’s spectrum can be observed in the productive range \( (kL_c < 1) \) and in the inertial range \( (1 < kL_c < 0.1 L_c/\eta) \). Both the \( u' \)- and \( w' \)-power spectra for \( z/h = 0.01 \) show increases in slopes for \( kL_c \) of about 10 and larger. The ratio of \( L_c \) and \( \eta \) can be estimated as (Tennekes and Lumley, 1972):

\[
\frac{L_c}{\eta} = R_e^{3/4}
\]  

(4.34)
$R_e = L_u u/\nu$ is approximately $3 \times 10^4$ for maximum ebb at the 0.1 m level. The upper limit of the inertial range $kL_x = 0.1 L_u/\eta$ is then approximately $2.3 \times 10^3$ which is appreciably larger than the values for $kL_x$ where the spectra start to bend downward. The increased slopes are not caused by viscous effects but result probably from high-frequency losses due to sensor size.

![Graph](image)

Figure 4.17. Estimated power spectra, for a one hour period during ebb (10:30-11:30h, 6/27/1996): (a) $u$-power spectra ($S_{uu}$), normalised with $L_u$ and Von Kármán’s spectrum; (b) $w$-power spectra ($S_{ww}$), normalised with $L_u$ multiplied by $w^2/u^2$, and the transverse spectrum for isotropic turbulence.

The measured $u$-spectrum for $z/h = 0.3$ shown in Figure 4.17a deflects towards a lower value of $S_{uu}(0)$ when extrapolated than the Von Kármán’s spectrum. This effect is entirely due to the limited record length of 51.2 s. If the record length is increased the deflection can be reduced, although at the cost of reduced accuracy of the spectrum for large $k$. This procedure does not make sense because the spectra lose their validity for $kL_x$ smaller than 1 since the frozen turbulence hypothesis becomes inappropriate (Nezu and Nakagawa, 1993). Apart from the explainable deviations described above, it can be concluded that agreement between the estimated $u$-power spectra and the Von Kármán spectrum is satisfactory.

On the basis of Von Kármán’s spectrum a transverse power spectrum can be calculated for isotropic turbulence (Nieuwstadt, 1992). This spectrum is shown in Figure 4.17b as the solid line. Figure 4.17b shows that the estimated $w$-power spectra
are well described by the transverse spectrum for large values of $kL_x$, indicating that the turbulence becomes isotropic. In the productive range large deviations are expected because the turbulence is strongly non-isotropic in this range. The ratio $w^2/u^2$, which is between 0.3 and 0.4, only partly explains the difference of approximately one decade between the $w$-power spectra and the calculated transverse spectrum. A possible explanation is that because of anisotropy of the turbulence the vertical integral length scale $L_z$ is less than the value of $\frac{1}{2} L_x$ for isotropic turbulence. The value of $L_z$ in the present case is about $\frac{1}{4} L_x$.

Some remarks should be made herein about high-frequency losses of turbulent fluctuations due to sensor size in the near-bed region. In Section 3.3.3 high-frequency losses are estimated using Soulsby's method (Soulsby, 1980). According to this method, spectra are valid over the entire boundary layer when normalised by $z$ with $z$ being elevation above the bed. Figure 4.17 shows that normalisation by $L_x$ yields good agreement between the spectra measured at various heights above the sediment bed. Consequently, normalisation by $z$ is less appropriate. Since the eddies containing a large part of the turbulent energy are of the order of $L_x(= \sqrt{zh})$, high-frequency losses due to sensor size are probably not as dramatic for small $z$ as suggested by Figure 3.6. These remarks are supported by the relatively small errors which result from high-frequency losses presented in Section 4.4.

Examples for the power spectra for $c$ and for the absolute values of the co-spectra for $c$ and $w$ are shown in Figure 4.18. The spectra are obtained from velocity and turbidity measurements at three heights above the channel bed during a period of high suspended sediment concentrations during flood in June 1996. In Section 4.6 it is shown that during these periods density stratification effects on the turbulence structure were present.

The $c$- and $u$-power spectra, shown in Figure 4.18a, are normalised with the integral scale $L_x$. However, in Eq. 4.20 $B_1$ was set to 0.4 in order to obtain agreement of the $u$-power spectra with Von Kármán's spectrum. A similar reduction of the turbulence length scale is found during the period 15:00h-16:45h 26/6 when relatively large vertical gradients of suspended sediment occur. This reduction in the integral scale $L_x$ of about 50% could tentatively be explained by stratification effects (see also West et al., 1986).

The slopes of the $u$-power spectra are well described by Von Kármán's spectrum. In other words: there is no indication of significant dampening of turbulence, despite the assumed stratification. In Section 4.3 it was shown that in the tidal channel the pressure caused by the water level gradient is balanced by bed friction except for
periods of low current velocity. Dampening of turbulence and a subsequent reduction of the eddy viscosity would directly lead to flow acceleration and subsequent increase in the velocity gradients and turbulence production. West and Oduyemi (1989) argue that if stratification effects lead to increased flow velocities, given the water level gradient, the conversion of potential energy into kinetic energy is enhanced. A reduction of the turbulence level, not to mention a complete collapse of the turbulence, is therefore not expected in the near-bed region (see Section 1.3). In the next sections effects of stratification on the turbulence structure are discussed in more detail.

Figure 4.18. Estimated spectra for a one hour period during flood (3:30-4:30h, 6/27/1996): (a) c-power spectra, u-power spectra and Von Kármán’s spectrum; (b) co-spectra (Re(S_{uw}))

Comparison of the c- and u-power spectra shows that the “energy” of the fluctuations of concentration and velocity are quite similarly distributed for kl, smaller than approximately 10. The -5/3 behaviour is also recovered in the c-power spectra except for high wave numbers. The similarity between the c- and u-power spectra is supported by the relatively high correlation between the u and c fluctuations (see also Section 4.7). The deflections of the c-power spectra for high wave numbers are attributed to the measuring technique. The suspension of flocs should be considered as a two phase system at small length scales. The size of large individual sediment flocs can be of the order of the diameter of the measuring light beam, in which case “spikes” in the turbidity signal occur. These spikes appear in the spectra as upward deflections of the tails.
Figure 4.18b shows the co-spectra, with inertial subrange slopes of approximately -3/7. These are similar to the $u\nu$-spectra (not shown), and to co-spectra of sediment laden flow presented by West and Oduyemi (1989) and of atmospheric data (Kaimal, 1973). The spectra presented in Figure 4.18 seem to confirm the conclusions of West and Oduyemi (1989) that the suspended sediment concentration fluctuations respond to the velocity fluctuations. The suspended sediment in the Dollard reach consists mainly of fragile flocs with settling velocities between 0.1 and 5 mm/s (Van der Lee, in prep.). These settling velocities are relatively small compared to the velocities of the energy containing eddies which are typically of the order of 0.02 - 0.05 m/s, and therefore it is very plausible that the motions of the flocs are closely related to the water motions.

The turbulence dissipation rates $\varepsilon$ are estimated according to Eq. 4.20 and Eq. 4.26. $B_1$ is set to 0.8, which might be too high a value for stratified flow conditions, as discussed above. The parameter $u'^3$ follows directly from the measurements of $u$ (Section 4.5). The errors in $\varepsilon$ were estimated at approximately 20% based on the SDOM’s of $u'$ presented in Section 4.4. Figure 4.19 shows some results for a number of flow conditions.

![Graph showing dissipation and production rates](image)

**Figure 4.19.** Some results for the dissipation rate $\varepsilon$ and turbulence production rate $P_q$ normalised by the friction velocity $u_*$ and the water depth $h$. Also plotted is the semi-theoretical relation according to Eq. 4.35.

The semi-theoretical relation according to Nezu and Nakagawa (1993), shown in Figure 4.19, can be obtained from Eq. 4.13, Eq. 4.20 and Eq. 4.26:
\[
\frac{\varepsilon h}{u^3} = \frac{12.2K}{B_1} (z/h)^{-0.5} \exp(-3z/h)
\] (4.35)

with \(K = 0.7\) and \(B_1 = 0.8\). In the present work the production rates were calculated by multiplying the measured shear stresses by the vertical velocity gradients (see Section 4.2). The shear stresses at the 0.1m level \((z/h = 0.01 - 0.02)\) were not used for the calculation of the turbulence production, as discussed in Section 4.5.

Except for two data points, the estimated dissipation rates for unstratified flow conditions seem to confirm Eq. 4.35, and are close to the turbulence production rates. These two data points represent too high estimates of the dissipation rates for stratified flow conditions for \(z/h = 0.3\), at 15:00h June 26 and 4:00h June 27. The validity of these data points is disputable because for stratified flow conditions Eq. 4.26 might not be appropriate. Eq. 4.26 is based on the assumption that \(u'^2\) represents the energy of turbulent fluctuations whereas, for stratified flow conditions a significant part of \(u'^2\) may be attributed to wave-like motions. These motions are not necessarily dissipative and therefore Eq. 4.26 may overestimate \(\varepsilon\).

4.6 Density stratification: gradient Richardson number and flux Richardson number

Density stratification can have significant effects on the structure of turbulence. Parameters which characterise the importance of density effects in shear flow are among others the flux Richardson number \(Ri_f\) and the gradient Richardson number \(Ri_g\). They are explained from the balance equation for turbulent kinetic energy \(Q\) which can be written as:

\[
\frac{DQ}{Dt} = T_q + P_q - B_q - \varepsilon
\] (4.36)

where \(T_q\), \(P_q\) and \(B_q\) represent the diffusive transport of \(Q\), shear production and buoyancy destruction of \(Q\), respectively. \(T_q\) is often neglected for boundary layer flow conditions. This implies that local equilibrium exists between \(P\) and \(\varepsilon\) which is approximately true for open channel flow (Nezu and Nakagawa, 1993). If uniform flow in the x-direction is assumed, Eq. 4.36 reduces to:

\[
\frac{\partial Q}{\partial t} = -u\frac{\partial \bar{Q}}{\partial z} - \gamma \frac{\partial \bar{Q}}{\partial \rho} - \varepsilon
\] (4.37)

where \(\gamma = ((\rho_s - \rho_w)/\rho_s)\), \(\rho_s\) the density of the sediment and \(\rho_w\) density of the water. The first term represents the shear production and the second term the buoyancy
destruction. The ratio of these terms is defined as the flux Richardson number \( R_{if} \) for turbulence sediment transport,

\[
R_{if} = \frac{B_q}{P_q} = \frac{\gamma g \bar{cw}}{- \rho \bar{uw} \frac{\partial U}{\partial z}}
\]  

(4.38)

Eq. 4.37 can now be written as follows:

\[
\frac{\partial q}{\partial t} = -\bar{uw} \frac{\partial U}{\partial z} \left( 1 - R_{if} \right) - \epsilon
\]  

(4.39)

Since both the shear production and \( \epsilon \) are positive, \( q \) certainly decays if \( R_{if} \geq 1 \). Measurements of \( R_{if} \) in stable boundary layer flows showed that for \( R_{if} \sim 0.2 \) turbulence is already completely suppressed (Nieuwstadt, 1992). The gradient Richardson number \( R_{ig} \) is obtained by substituting gradient expressions for the fluxes in Eq. 4.38:

\[
R_{ig} = \frac{-\gamma g \frac{\partial C}{\partial z}}{\rho \left( \frac{\partial U}{\partial z} \right)^2}
\]  

(4.40)

Bottom turbulence is dampened for \( R_{ig} \sim 0.2 \), and free turbulence is dampened for \( R_{ig} \sim 0.5 \) to 1. A relation between \( R_{if} \) and \( R_{ig} \) is found if \( \bar{cw} \) and \( \bar{uw} \) in Eq. 4.38 are approximated using the eddy viscosity/diffusivity concept:

\[
R_{if} = \frac{K_s}{K_m} \frac{-\gamma g \frac{\partial C}{\partial z}}{\rho \left( \frac{\partial U}{\partial z} \right)^2} = \frac{1}{\sigma_T^2} R_{ig}
\]  

(4.41)

\( K_s \) is the eddy diffusivity for fine sediment particles, \( K_m \) is the eddy viscosity and \( \sigma_T^2 = K_m / K_s \) is the turbulent Prandtl number for suspended sediment.

Section 4.5 showed that the motions of the flocs are closely related to the water motions. It is therefore assumed herein that Reynold’s analogy also applies to sediment transport. Reynold’s analogy assumes that there is complete analogy between the transport of momentum and heat. From experiments \( \sigma_T \) was found to be between 0.7 and 0.9, it slightly increases in the wall region \((z/h < 0.2; \text{ Hinze, 1975}) \). From measurements on the distribution of suspended sediments in open channel flow \( \sigma_T^2 \) is found to be approximately 1.0 (Nezu and Nakagawa, 1993). In this study the value for
\( \sigma_f \) is set to 1.0. In Sections 4.2 and 4.4 gradients and transports are derived which are used to calculate \( Ri_f \) and \( Ri_g \). Some results are presented in Figure 4.20 and Figure 4.21 for the June and August measuring periods, respectively. \( Ri_f \) and \( Ri_g \) are deliberately not shown during periods of slack water where they become indeterminable.

![Graphs of Ri_f and Ri_g](image)

Figure 4.20. Time series of the measured gradient Richardson number \((Ri_g)\), the flux Richardson number \((Ri_f)\) and the longitudinal velocity \(\bar{U}\) at 1.0 m, 22:00h June 26 - 8:00h June 27.

Figure 4.20 shows \( Ri_g \) and \( Ri_f \) during a period where they could be determined at all three levels above the channel bed. The vertical bars denote the SDOM for \( Ri_f \). \( Ri_g \) behaves similar to \( Ri_f \), at least for \( Ri_g < 0.1 \), which is to be expected in boundary layer flow. Close to the sediment bed, stratification is relatively unimportant during the entire flood period whereas higher in the water column \((z = 1.0 \text{ m})\) stratification is more pronounced, especially during periods of relatively small flow velocities (see also
Darbyshire and West, 1993). The increase in $Ri_f$ and $Ri_g$, at about 4:00h, coincides with a drop in the flow velocity and a peak in the sediment concentration (see Figure 4.1).

The trend of $Ri_g$ is in agreement with the trend in $Ri_f$. An approximate match between the values of $Ri_g$ and $Ri_f$ at $z = 0.4$ m and $z = 1.0$ m can be obtained, as shown in Figure 4.20, by dividing $Ri_g$ by a factor three (thus $\sigma_r^2 \approx 3$). The values of $Ri_g$ should be interpreted with care because of the limited vertical resolution of the concentration profiles. Furthermore, small variations in offsets of the turbidity sensors during the measuring period can have significant effects on the concentration gradients.

![Graph showing time series of gradient Richardson number ($Ri_g$), flux Richardson number ($Ri_f$), and longitudinal velocity ($U$) at 0.1 m and 0.4 m.](image)

Figure 4.21. Time series of the gradient Richardson number ($Ri_g$), the flux Richardson number ($Ri_f$) and the longitudinal velocity $U$ at 1.0 m, 0:00h August 8 - 1:00h August 9.

Figure 4.21 shows the calculated Richardson numbers at 0.1 m and 0.4 m during two tidal cycles in August. At 0.4 m the stratification is more pronounced than at 0.1 m, which is in agreement with the findings from the June measuring period. The absolute values of both $Ri_g$ and $Ri_f$ are significantly smaller than those of the June measuring period. This effect can be explained from the higher concentrations in June.

4.7 Effects of density stratification on turbulence parameters

In the previous section it was shown that during certain parts of the tidal cycle, effects of density stratification on the turbulence characteristics can be expected. Conditions are preferred which comprise: large variation of $Ri_g$ numbers, nearly stationary flow,
relatively large density gradients, and well defined velocity profiles. Relatively large density gradients are important to reduce the influence of inaccuracy of the MEX calibrations and of offset variations. These conditions are approximately met in periods of temporal decrease of the flow velocity and increase in concentration during flood during the June measuring period, of which an example is shown in Figure 4.20 at about 4.00h (see also Figure 4.1). These periods are denoted as period 1 to 4.

![Diagram](image)

Figure 4.22. The variation of $R_{cw}$ with $Ri_g$ for periods of relatively large density gradients ($> 0.1 \text{ g/l/m}$) and significant horizontal velocity ($U_{10} > 0.2 \text{ m/s}$; $z/h = 0.1 - 0.5$).

The variation of the correlation coefficient $R_{cw}$ with $Ri_g$ is shown in Figure 4.22 for periods 1 to 4 (see legend for the exact times and dates). A similar behaviour of $R_{cw}$ was presented by West and Oduyemi (1989). The scatter of the data displayed by Figure 4.22 is somewhat larger than that of laboratory data (e.g. Webster, 1964), and comparable to that of West and Oduyemi (1989). $R_{cw}$ decreases from an average value of $0.33 \pm 0.05$ for $Ri_g < 0.1$ to approximately $0.26 \pm 0.10$ for $Ri_g$ between 0.2 and 0.8. The vertical bar denotes the SDOM for $R_{cw}$ and the horizontal bar denotes an crude estimation of the error in $Ri_g$. The shifts in horizontal direction between the variations of $R_{cw}$ with $Ri_g$ for each separate period might have resulted from a bias error in $Ri_g$. Of all the data shown in Figure 4.22, the data of the first period (15:00h - 16:45h) are assumed the most reliable because the MEX sensors were calibrated against sediment samples over this particular period. If this assumption is correct, $Ri_g$ is probably estimated too high for the period 2 to 4. Results from the literature show that for extremely stable flow, i.e. $Ri_g > 0.5 - 0.7$, $R_{cw}$ approaches zero or even changes sign (Komori et al., 1983; Ueda et al., 1981).
The estimated values of $Ri_k$ for periods 2 to 4 can be adjusted in the following way. For homogeneous shear flows that are in equilibrium it can be assumed that $\sigma_f^z$ remains finite at all Richardson numbers, and varies as (Schumann and Gerz, 1995):

$$\sigma_f^z \approx \sigma_{f,0}^z \exp \left( - \frac{Ri_f}{\sigma_{f,0}^z Ri_{f,\infty}} \right) + \frac{Ri_f}{Ri_{f,\infty}}$$  \hspace{1cm} (4.42)$$

where $\sigma_{f,0}^z$ is the turbulent Prandtl number for neutral shear flows (set to 1.0, see above), and $Ri_{f,\infty}$ is the flux Richardson number for strong stratification. If Eq. 4.42 is combined with Eq. 4.41 and a value of $Ri_{f,\infty}$ is assumed (here 0.2), a relation between $Ri_f$ and $Ri_k$ is found which is plotted in Figure 4.23 (dashed line), together with $Ri_f$ for periods 2 to 4.

![Figure 4.23](image_url)

Figure 4.23. The variation of $Ri_f$ with $Ri_k$ during periods 2 to 4 and $Ri_f$ calculated from Eq. 4.41 and Eq. 4.42 (dashed line).

The slope of the trend of $Ri_f$ vs. $Ri_k$ shown Figure 4.23 as the solid line, is equal to the measured inverse turbulent Prandtl number $\sigma_{f,0}^z$, which number would then be approximately equal to 3 (see also Figure 4.20). This value seems much too high. A more realistic value for $\sigma_{f,0}^z$ is for small values of $Ri_k$ of the order of one, as was already mentioned in the previous paragraph. The overestimation of $\sigma_{f,0}^z$ and $Ri_k$ for periods 2 to 4 are attributed to errors in concentration gradients caused by fouling of the MEX sensors and to a less extent to errors in velocity gradients. The concentration gradients and the subsequent $Ri_k$ can be adjusted in such a way that the slope of the trend of $Ri_f$ (Figure 4.23) reduces to approximately 1.

Figure 4.24 shows the variation of $R_{uw}$ with the adjusted $Ri_k$, $Ri_k^z$, for periods 2 to 4 together with that for period 1 with $Ri_k$. The variations of $R_{uw}$ for periods 2 to 4 are
now in agreement with the variation of $R_{uw}$ for period 1 and with the trend line of West and Oduyemi's data (1989), also plotted in Figure 4.24.

![Graph showing the variation of $R_{uw}$ with $Ri_k$ for period 1 and with $Ri_{f}$ for periods 2 - 4.]

Figure 4.24. The variation of $R_{uw}$ with $Ri_k$ for period 1 and with $Ri_{f}$ for periods 2 - 4.

![Graph showing the variation of $R_{uw}$ with $Ri_{f}$ for periods 2 - 4.]

Figure 4.25. The variation of $R_{uw}$ with $Ri_{f}$ for periods 2 - 4.

Figure 4.25 shows the variation of $R_{uw}$ with $Ri_{f}$ during periods 2 to 4. No data on $Ri_{f}$ are available for period 1. The horizontal bar denotes the SDOM for $Ri_{f}$. The correlation coefficient $R_{uw}$ reduces from an average value of $0.36 \pm 0.04$ for $Ri_{f} < 0.05$ to approximately $0.25 \pm 0.05$ for $Ri_{f} > 0.15$. The reduction of $R_{uw}$ for increasing stratification seems to confirm the previous results.

Figure 4.26 shows the variation of the correlation coefficient $\frac{\overline{uw}}{w'^{2}}$ normalised with its mean value for $Ri_{g} < 0.1$, with $Ri_{gs}$ obtained from the data during period 1 and from large eddy simulation (LES) results for uniform shear flow with homogeneous turbulence. It can be shown that the assumption of nearly homogeneous turbulence is often satisfied even in inhomogeneous boundary layers (Mellor and Yamada, 1974, in
Schumann and Gerz, 1995). The results again show a decreasing trend of $-\overline{uw}/w^2$ for $0 < Ri_g < 0.3$, which is similar to that in Figure 4.24.

![Graph showing $-\overline{uw}/w^2$ vs. $Ri_g$]

Figure 4.26. The variation of $-\overline{uw}/w^2$, normalised by $-\overline{uw}/w^2$ for $Ri_g < 0.1$, with $Ri_g$ (adjusted) during period 1 and the results of LES of Schumann and Gerz (1995).

![Graph showing $R_{cw}$ vs. $Ri_g$]

Figure 4.27. The variation of $R_{cw}$ with $Ri_g$ during periods 2 to 4 ($z/h = 0.1 - 0.5$), and the trend of West and Oduyemi's data for $z/h = 0.1 - 0.4$ (1989).

Figure 4.27 shows the variation of $R_{cw}$ with $Ri_g$. The scatter of the data is large: $R_{cw}$ decreases from an average value of $0.23 \pm 0.1$ for $Ri_g < 0.1$ to approximately $0.20 \pm 0.05$ for $Ri_g > 0.2$. The variation of $R_{cw}$ with $Ri_g$ shows a similar scatter, but the value of $R_{cw}$ for $Ri_g < 0.1$ is larger: $0.31 \pm 0.1$ and it decreases to $0.20 \pm 0.15$. Measurements of the turbulent sediment fluxes in the August measuring period were of relatively good quality compared to those of the June measuring period (see also Section 3.5). For neutral flow conditions the correlation coefficient $R_{cw}$ is $0.19 \pm 0.1$ and $R_{cw}$ is $0.31 \pm 0.1$ for $z/h < 0.2$. These values of $R_{cw}$ and $R_{cw}$ are in agreement with June
measurements and with data of West and Oduyemi for suspended sediment (1989), and
of Komori et al. (1983) for heat transport in air.

Figure 4.28. The variation of $R_i_f$ with $R_i_g$ during parts of the August measuring period and
$R_i_f$ calculated from Eq.4.41 and 4.42 (dashed line) for $R_{i,f,\infty} = 0.2$ and $\sigma_{i,g} = 1.0$.

Figure 4.28 shows the variation of $R_i_f$ with $R_i_g$ during periods of significant density
gradients ($ > 0.1$ g/l/m) during the August 1996 measuring period (see also the
legend). Some bias errors are present in $R_i_g$, probably due to small offset changes in
the turbidity sensors. Nevertheless, some evidence is found for a linear relation
between $R_i_f$ and $R_i_g$ for small values of $R_i_g$ ($< 0.2$) and a more or less constant value
for $R_i_f$ for large values of $R_i_g$ ($> 0.5$).

4.8 Turbulence structure
The turbulence structure in tidal flows can be described on the basis of the findings
from the turbulence research in laboratory flumes because of the universality of
turbulence. However, the description might require some modification because tidal
flows have significant higher Reynolds numbers compared to laboratory flows, they are
unsteady and are often influenced by stratification. Nezu and Nakagawa (1993)
hypothesise that kolk boil vortices are important for vertical transport of sediment and
momentum in geophysical flows. The fact that these structures are given little attention
in laboratory studies is attributed to difficulties with the analysis of the low-frequency
range of turbulence records and the weakness of kolk-boil vortices over flat beds in
laboratory flumes. This section examines if the findings of the Dollard field
measurements can contribute to a better understanding of the turbulence structure in
large scale tidal flows.
As shown in Section 4.2, the tidal velocities depend to a large extent on the bathymetry of the Dollard and the tidal variations in water level. Tidal variations of the sediment concentration showed that they only roughly correlate with the variations in tidal velocities. It was argued that interpretation of the measurements and the analysis of the role of sediment induced stratification in the horizontal transport of fine sediments, can best be done through analysis with a numerical model in which the dominant processes are represented in a well-balanced way (see Chapter 5).

In Section 4.3 it is shown, by examining the equation of motion for tidal flow, that acceleration and deceleration effects on the turbulence structure are of little importance for the shallow water flows studied (West et al., 1986). The terms representing inertia and the longitudinal density gradient are small compared to the tidal forcing for the Dollard reach. In addition, it is shown that the ratio of the contributions of transverse and longitudinal shear to the equation of motion is of the order of $h/B$ where $h$ is the water depth and $B$ is the channel width. Generally the ratio $h/B$ is much smaller than one. These findings seem inconsistent with the findings of French and Clifford (1992) who stress the importance of flow unsteadiness and the need for a complete description of the Reynolds stress tensor. An explanation can be found from the differences between the measuring locations. French and Clifford conducted their measurements in a relatively small but strong meandering tidal marsh channel of one of the western marshes of Scolt Head Island, Norfolk (UK) during high water, spring tide. This type of channels has strong curvatures and exhibit accelerations an order of magnitude larger than commonly found in estuarine boundary layers. In conclusion, the findings of French and Clifford apply to another class of estuarine flows of which the smaller channels in the outer reaches of the Dollard might be representative but not the main channel Groote Gat where our measurements were conducted.

In Section 4.4 the shear stresses and turbulence intensities are presented. An equivalent roughness height of 2 mm is found from shear stress measurements which is in agreement with commonly proposed roughness heights for flow over a cohesive sediment bed (Van Rijn, 1993). Visual observations of the edge of the Heringsplaat in 1996 showed that bed forms were almost absent at levels below N.A.P. The results presented on turbulence intensities suggest that the turbulence characteristics in the tidal channel are comparable to the turbulence characteristics obtained in laboratory flumes at much smaller Reynolds numbers. Both the variation of turbulence intensities with relative height as well as the ratio of longitudinal to vertical velocity intensities
were in agreement with laboratory data (Nezu and Nakagawa, 1993). Similar conclusions were presented by West and Oduyemi (1989), and Kawanisi and Yokosi (1993).

Concentration fluctuation intensities were coupled to the average suspended sediment concentrations and the vertical concentration gradients. Except during slack water periods when turbulence vanished, the vertical turbulent flux of suspended sediment was upward and the horizontal turbulent flux was against the direction of the main flow. The upward and horizontal turbulent fluxes had maximum absolute values for maximum turbulent concentration fluctuation intensities. The signs of the turbulent fluxes suggest that upward moving fluid parcels, which move slowly in the streamwise direction, contain relatively high concentrations, and downward moving fluid parcels which move fast in the streamwise direction, contain relatively low concentrations. Photographs of upward moving flow structures, taken on April 20, 1996, are presented in Figure 4.29.

Figure 4.29. Boil structure during a re-suspension event (April 20, 1996). The photographs were taken at an interval of a few seconds.

The conditions during April 20, 1996, were quite unique: wind was fully absent, and water was very clear during slack water, which was probably due to biological activity (Kornman and De Deckere, 1998). The "Secchi depth", a measure for the level of turbidity, varied between 0.5 m and 1.0 m. Sediment clouds appeared randomly distributed at the free surface during the start of the flood at 11:00h. The clouds had typical diameters of 0.3 to 1.0 times the water depth (h was between 2 and 3 m). The separating distance between the clouds was estimated at 10 to 30 m. The turbulence structures that caused these re-suspension events could be representative of kolk-boil vortices which are associated with bursting motions, because of their random appearance and the absence of bed forms as discussed above (Nezu and Nakagawa,
1993). The disappearance of the sediment from the free surface region seemed rather quick compared to the settling velocity of the individual particles of approximately 0.1 - 5 mm/s (Van der Lee, in prep.). This might be explained from the unstable stratification which results from the re-suspension event. Fluid that reached the surface had a relatively high concentration compared to the surrounding fluid. It may have moved downward quickly as the result of its larger density, a process known as settling convection.

In Section 4.5 turbulence spectra were presented for some representative parts of the tidal cycle. An important result of the spectral analysis is that the integral length scale $L_x$ is approximately $0.8 \sqrt{zh}$ which is in agreement with laboratory data of Nezu and Nakagawa (1993). This means that even close to the bed the sizes of the turbulence structures remain relatively large in the streamwise direction. The vertical integral length scale ($L_z$) is estimated at $\frac{1}{4}$ of the value for $L_x$ which is significantly smaller than the $\frac{1}{2}L_x$ for isotropic turbulence. These macroscales are probably closely related to the large coherent structures described above (Kawanisi and Yokosi, 1993). Since the coherent structures can be scaled with the outer flow variables, it is no surprise that the water depth $h$ appears in the formulation for $L_x$. The similarity between the spectra of velocity fluctuations and concentration fluctuations suggested that the concentration fluctuations correspond to the turbulent water motions. This aspect supports the visual observations of the kolok-boil vortices as described in the previous paragraph.

In Section 4.6 the Richardson flux and gradient Richardson numbers are calculated for some parts of the measuring periods in June and August 1996. The results show that effects of density stratification on the turbulence structure can be expected for those parts of the tidal cycle where velocities are relatively small and/or where concentrations are relatively high. For the June measuring period such periods occur during flood when the flow velocity decreases for approximately one hour and the sediment concentrations peak (see Figure 4.1 and Figure 4.21). The "high estimated dissipation rates" and the reduced length scale $L_x$ for these periods, as derived in Section 4.5, can therefore be attributed to density stratification. In Section 4.7 the correlation coefficients are presented for the transport of momentum and sediment. The results show decreasing correlations between $u$ and $w$ fluctuations for increasing stratification. This suggests that motions were generated which contributed fully to the velocity fluctuations but to a smaller extent to the transport of momentum. These might be internal waves. How these wave-like motions can be discerned in the velocity
spectra is difficult to predict (Uittenbogaard, 1995). Because of the strong stratification there is no theoretical basis for the observed -5/3 behaviour of the power spectra in the inertial sub-range (see Figure 4.18). Some evidence is found for an increase of the turbulent Prandtl number with increasing stratification. Similar results have been found by other researchers (Ueda et al., 1981; Komori et al., 1983; Schumann and Gerz, 1995).
Chapter 5 Analysis of field measurements through simulations with the 1DV POINT MODEL

5.1 Introduction

Time series of mean velocities, concentrations and turbulence properties, obtained during the June and August 1996 measuring periods, were analysed by means of a one-dimensional numerical model. The 1DV POINT MODEL of wt. | delft hydraulics (Uittenbogaard et al., 1996) was selected for this purpose, because some related experience has already been obtained: Winterwerp et al. (1998) used the model to analyse results of field measurements of fine-sediment transport conducted in the Maasmond area. It was found that, in line with the findings of Chapter 4, buoyancy effects played an important part.

The 1DV POINT MODEL, or 1DV-model for short, computes profiles of mean velocities and concentrations, and turbulence properties in the water column. Water levels, depth-averaged velocities, and sediment parameters form the input for the 1DV-model. All horizontal gradients are assumed zero except the stream-wise pressure gradient. The 1DV-model comprises a $q$-$c$ turbulence closure model (better known as the $k$-$c$ turbulence model) and the Partheniades-Krone formulations for describing erosion and sedimentation. The $q$-$c$ turbulence model allows for the investigation of possible effects of buoyancy on the vertical exchange processes (Violet, 1988; Sheng and Villaret, 1989). The choice of the commonly used Partheniades-Krone formulations follows related research on sediment transport in the Ems/Dollard estuary (Ridderinkhof, 1998). For a comprehensive description of the governing equations and some numerical aspects of the 1DV-model, the reader is referred to Appendix C.

Advection is not incorporated in the 1DV-model, which can have consequences for the analysis of the measurements depending on the property under consideration (see also Section 5.2). Other uncertainties are related to, among other things, the variability of the sediment properties and the crude approximations for the water-bed exchange processes. Therefore, "simulations" is too strong an expression for describing the computations with the 1DV-model. The main purpose of the computations is not to match the numerical results to the measurements, but to learn if some typical aspects of
the fine-sediment behaviour, as presented in Chapter 4, can be explained from
interactions between the vertical transport processes.

Section 5.3 presents results of computations of erosion - sedimentation cycles in the
tidal channel Groote Gat during the June measuring period and compares them to the
field measurements. In Sections 5.4 and 5.5 some results of in-situ studies on the
erodibility of the sediment bed (Kornman, 1998) and on settling velocities (Van der
Lee, 1998) are schematically modelled and added to the 1DV-model. The
consequences for the analysis of the June measurements are discussed. In Section 5.6
the August measurements are analysed and discussed on the basis of the experiences
gained for the June measurements. In Section 5.7 the results of the analyses of the June
and August field measurements are integrated and discussed.

5.2 Parameter selection and computations with the 1DV POINT MODEL
The variations in water level \( h \) and depth-averaged velocities \( U_d \) from June 26 12:15h
until June 28 10:00h are used as input for the computations with the 1DV-model.
Figure 5.1a shows these input variables over the second and third tidal cycles of the
June measuring period. The depth-averaged velocity \( U_d \) is calculated from
extrapolation of the Van Veen velocity profiles up to the water surface followed by
integration over the water depth.

The roughness length \( z_0 \) is prescribed and it was found from the bed friction
parameter \( C_d \) of 1.8\times10^{-3} \) by assuming a logarithmic velocity distribution near the
sediment bed (see Appendix C). This gives \( z_0 = 6.7 \times 10^{-5} \) m. The critical shear stress
for erosion, \( \tau_{cr} \), is set to 0.1 Pa which is an average value of the critical shear stresses
found from in-situ erosion studies on the adjacent tidal flat (Kornman,1998). The
critical shear stress for deposition \( \tau_{cd} \) was arbitrarily set at 0.1 Pa. A fair agreement
between the variations in measured and the computed SSC can be obtained by
adjusting the parameters \( M \) and \( w_s \). For this purpose they were set to 4\times10^{-5} \ \text{kg/m}^2/\text{s}
and 0.5 mm/s, respectively (see Run J1, Table 5.1; Figure 5.1b). The value for \( M \) is in
the same range as those obtained from in-situ field experiments on the tidal flat
Heringsplaat (Kornman, 1998). The value for the mean settling velocity \( w_s \) is in the
same range as those found by Van Leusen (1994) from settling tube measurements in
the Ems/Dollard. It should be mentioned that, according to Van Leusen (1994), the
settling tube measurements underestimate the settling velocities. Small settling
velocities of about 0.1 mm/s led to too small computed differences in SSC between
maximal flow velocities and slack water. Large settling velocities of about 2 mm/s led
to too large computed concentration gradients compared to June and August measurements.

Effects of buoyancy are investigated by incorporating the buoyancy terms in the transport equations for turbulent kinetic energy, $q$, and dissipation, $e$ (Run J2, Table 5.1; see also Appendix C). The critical shear stress for erosion, $\tau_{c,s}$, can be treated as a constant if the top layer of sediment has uniform properties (Kuiper et al., 1989). But if the top layer is eroded in a relatively short period of time and deeper lying layers with higher resistance against erosion become exposed, $\tau_{c,s}$ should be made dependent on depth (Kornman, 1998). This is schematically represented in computation J3 by limiting the SSC to $C_{max}$ and by increasing the erosion rate $M$. Effects of the settling velocity's dependence on SSC and turbulence level are investigated in the computations J4 and J5 (Table 5.1). For a comprehensive description of all parameters the reader is referred to Appendix C.

<table>
<thead>
<tr>
<th>Run:</th>
<th>$w_s$ (m/s)</th>
<th>$M$ (kg/m$^2$/s)</th>
<th>$C_{max}$ (kg/m$^3$)</th>
<th>Buoyancy</th>
</tr>
</thead>
<tbody>
<tr>
<td>J1. Basis</td>
<td>5$\times$10$^{-4}$</td>
<td>4$\times$10$^{-5}$</td>
<td>-</td>
<td>Not incl.</td>
</tr>
<tr>
<td>J2. Plus buoyancy</td>
<td>5$\times$10$^{-4}$</td>
<td>4$\times$10$^{-5}$</td>
<td>-</td>
<td>Included</td>
</tr>
<tr>
<td>J3. Plus limited SSC</td>
<td>5$\times$10$^{-4}$</td>
<td>4$\times$10$^{-4}$</td>
<td>0.5</td>
<td>Included</td>
</tr>
<tr>
<td>J4. $w_s = f(C)$</td>
<td>Eq. C.21/ C.22</td>
<td>4$\times$10$^{-4}$</td>
<td>0.5</td>
<td>Included</td>
</tr>
<tr>
<td>J5. $w_s = f(C,G)$</td>
<td>Eq. C.21/ C.22/ C.23</td>
<td>4$\times$10$^{-4}$</td>
<td>0.5</td>
<td>Included</td>
</tr>
</tbody>
</table>

The computations were made on a non-equidistant computational grid with 21 grid points and with a computational time step of 6 seconds. Convergence was checked with an equidistant grid with 101 grid points and with a computational time step of 0.6 seconds. Convergence was not always found for the gradient Richardson number $Ri_g$ for periods of low velocity (slack water) when it became indeterminable: both numerator and denominator tended to zero. The gradient Richardson number $Ri_g$ is therefore not shown for these periods.

The influence of the initial conditions on the computational results was checked by comparing the model results for the June period for two different SSCs at the start of the measuring period, to wit: 1.0 g/l and 0.0 g/l. The comparison showed that the starting conditions have a limited influence, that is, only on the first simulated tidal cycle. The computational results of the first tidal cycles of the June and August periods.
are therefore not presented herein. The influence of the initial conditions extended to several tides when the settling velocities were set to a small value (< 0.2 mm/s).

Some model results for the SSCs for Runs J1 and J2 are shown in Figure 5.1b for the second and third tidal cycles of the June measuring period. The measured variations of SSC are also plotted in Figure 5.1b.

The variations in the measured and computed SSC at 0.7 m show, generally speaking, the same behaviour: the sediment SSC starts to increase some time after the start of each ebb or flood, and the major part of the sediment settles during slack water. There are important differences, though; during the start of the flood the measured SSC increases rather abruptly, whereas a slow increase is predicted by the model. The peaks in measured SSC during floods are attributed to a local increase in the availability of sediment at the edges of the Heringsplaat during rising tide, as already mentioned in Section 4.2. After the sudden increase, the measured SSC decreases while the flow velocities are still increasing.

![Graph showing model input and SSC during June 1996 measuring period.](image)

**Figure 5.1.** The model input and the SSC during the June 1996 measuring period: (a) The water level, \( h \), and the depth-averaged velocity, \( U_d \) (the bottom elevation was -3.3 m N.A.P.); (b) Computational results for the SSC at 0.7 m: without buoyancy (Run J1) and with the buoyancy taken into account (Run J2), and the measured SSC at the same level.
5.3 Effects of buoyancy on fine sediment behaviour and hydrodynamics

Figure 5.1b shows distinct differences between the computed variations in SSC for the first two parameter sets shown in Table 5.1: the average level of SSC is significantly lower if buoyancy effects are taken into account, especially during slack water. According to the 1DV-model this is caused by the suppression of turbulence by buoyancy effects. The suppression of turbulence cause the rapid decrease in SSC, and subsequently the reduction in settling lag, shown in Figure 5.1b at 13:00h June 27 and 00:00h June 28. The increased importance of buoyancy is reflected in the increase in the gradient Richardson numbers as is shown later on in this section.

Effect of buoyancy on the turbulence properties

Figure 5.2 shows the measured and computed shear stresses at 0.4 m above the sediment bed. Except for the stresses during the first parts of the flood, at 3:00h and 15:00h, and of the ebb at 10:00h, the computed shear stresses are in good agreement with the measured stresses. That is partly due to the fact that the roughness height $z_0$ used in the computations, was derived from the measured Reynolds stresses (see also Appendix C). The overestimation of the bed shear stresses by the model at 10:00h is ascribed to acceleration, as discussed later on.

![Graph showing measured and computed shear stresses at 0.4 m above the sediment bed.]

Figure 5.2. Measured and computed shear stresses at 0.4 m above the sediment bed during the second tidal cycle of the June measuring period (2:00h - 16:00h, June 27).

The shear stresses calculated by the model (at 0.1 m, 0.4 m and 1.0 m) are hardly influenced by buoyancy for absolute flow velocities larger than approximately 0.1 m/s (see Figure 5.2). This finding is in agreement with model results for a placed bed in a
laboratory flume (Sheng and Villaret, 1989) and field measurements of shear stresses close to the sediment bed in a stratified tidal flow (Shiono and West, 1987).

Figure 5.3 shows the turbulent vertical transports of sediment at 0.4 m during the second tidal cycle of the June measuring period. The measured vertical transports show only small differences between ebb and flood while the computed transports show large differences. The peaks in the measured vertical transports of sediment during acceleration are not reproduced by the model. It can be shown, by changing the settling velocity in the 1DV-model, that the major part of turbulent transport is used for balancing the settling flux \( w_s \bar{C} \). Therefore, differences between the computed and measured variations of \( \bar{C} \) (see also Figure 5.1b) are reflected in the turbulent transports (Figure 5.3). Despite the discrepancies the computational results are of importance for the interpretation of the measurements, as shown below.

![Figure 5.3](image)

Figure 5.3. Measured and computed turbulent vertical transports of sediment at 0.4 m above the sediment bed during the second tidal cycle of the June measuring period (2:00h - 16:00h, June 27).

The computed vertical transports are in the same range as the measured values of \( \bar{w} \) for \( w_s \) about 0.5 mm/s. As mentioned before, direct measurements of settling velocities in the upper regions of the water column showed values up to several millimetres per second (Van der Lee, 1998; Van Leusden, 1994). However, if it is assumed that \( w_s \) is this large in the near-bed region, the 1DV-model predicts too large vertical concentration gradients and too large vertical transports compared to the measurements. Mean settling velocities in the near-bed region of 0.5 mm/s could be explained from floc break-up, a large net accretion during the entire measuring period, or from upwelling of water at the measuring location near the border of the tidal flat.
Buoyancy has no large effects on the vertical transports in these computations, except for the periods around slack water. The 1DV-model with the buoyancy effects included (Run J2, Table 5.1) shows strongly reduced vertical transports in the periods before slack water during deceleration. It is this reduction that causes the “rapid settling” shown by the model results in Figure 5.1b at 13:00h June 27 and 00:00h June 28. During acceleration the differences are less pronounced because the sediment concentrations, and subsequently the buoyancy effects, are small (see also Figure 5.6.). Van der Lee (1998) observed a relatively rapid decrease in sediment concentrations in the tidal channel Groote Gat approximately one hour after maximum current velocities during measurements from a research vessel drifting with the tidal current during October 11 1995 and April 11 1996. Van Leussen (1994) observed similar processes in other parts of the Ems/Dollard and referred to them as “rapid settling events”. Van der Lee and Van Leussen explain the sudden decrease in SSC from large sediment flocs which develop during periods of low turbulence intensities. These flocs are thought to have settling speeds of the order of a few mm/s (underwater camera observations). However, the computational results obtained herein show that effects resulting from buoyancy could also provide an effective mechanism for the “rapid settling event”.

![Graph](image)

Figure 5.4. Measured and computed gradient Richardson numbers at 0.4 m above the sediment bed during the second tidal cycle of the June measuring period (2:00h - 16:00h, June 27).

The gradient Richardson number and estimations of the velocity and concentration gradients from the measurements

Figure 5.4 shows computed gradient Richardson numbers and those derived from the measurements (see Section 4.6) for a full tidal cycle. Around slack water periods, around 8:45h and 13:00h June 27, both the 1DV-model and the measurements show increased values of $Ri_g$. During periods of maximum current velocities, around 5:00h
and 11:00h June 27, the values of $Ri_g$ are small ($< 0.05$) which indicates that buoyancy effects are negligible. Differences between the measured and the computed $Ri_g$ numbers can be explained from differences in the measured and computed vertical transports of sediment. For example: at 4:00h peaks are present in the measured values of $cw$ and $Ri_g$, but not in the computed values. The computed rapid decrease in SSC at 0.4 m at 13:00h June 27, as shown in Figure 5.1b, coincides with the computed reductions in $\bar{uw}$ and $cw$ (Figures 5.2 and 5.3), and the increase in the computed $Ri_g$ in Figure 5.4. The latter is in agreement with the increase in the measured $Ri_g$. In Section 4.2 the velocity and concentration gradients were obtained from fitting the Van Veen velocity profile and the Rouse concentration profile to the measured time series of velocities and concentrations, respectively.

![Image](image_url)

Figure 5.5. Velocity profiles during the transition from maximum flood velocity, via high water to maximum ebb velocity (June 27, HW: 8:45h).

Figure 5.5 shows some of the Van Veen velocity profiles, fitted to the measuring points, and the velocity profiles obtained from the 1DV-model with and without buoyancy. The profiles calculated by the 1DV-model show that a “Van Veen profile” can be used for estimating the velocity gradient in the bottom region ($< 0.2h$) except during slack water. The depth-averaged velocity $U_d$ obtained from the Van Veen profiles is not representative of $U_d$ according to the 1DV-model. The computed velocities show clearly a phase lag at higher positions in the water column which is due to acceleration. This computed effect is much more pronounced if buoyancy effects are taken into account. Unfortunately, no velocity measurements were made at higher positions in the water column. The author recommends that in future research velocity measurements are made over the entire water column.

At HW+0:30h the computed velocities near the bottom exceed the measured velocities, which results in too large computed shear stresses during the acceleration.
phase of the ebb current (see Figure 5.2). The measured velocity gradients during ebb are relatively large compared to the computed velocity gradients (HW +1:00h and +2:10h). In August the situation is reversed (see also Section 4.2).

Direct comparisons of the measured and computed gradients in SSC was of little use because of the large differences in SSC between measurements and computations (see also Figure 5.1b). In order to get a general idea of validity of the Rouse profile for estimating the gradients from the measurements, the Rouse profile is fitted to the SSC computed by the 1DV-model at 0.3 m, 0.7 m and 1.4 m. The gradients predicted by the Rouse-fit can be compared to the gradients computed by the 1DV-model. The outcome of the comparison is that Rouse-fits yield smaller gradients than the computed gradients in the near-bottom region, especially during slack water periods (not shown).

Another result is that the coefficients $C_0$ and $\alpha$ of the Rouse profile (Eq. 4.3, Section 4.2) are largely affected by the unsteady-state conditions and by buoyancy effects. The settling velocity estimated from $\alpha$ can be up to 2 mm/s during periods of significant stratification, whereas the prescribed settling velocity was 0.5 mm/s. This finding might explain the discrepancy between the settling velocities of 0.5 - 1.5 mm/s obtained from the time series of $\alpha$ and $u_*$ in Section 4.2, and the much smaller settling velocities of 0.1 - 0.8 mm/s estimated from the time series of $\overline{\overline{\nu}}$ in Section 4.4.

**Effects of buoyancy on the net horizontal transport rates**

Horizontal transport rates are computed by multiplying the SSC and streamwise velocity component, followed by integration over the water depth. Figure 5.6 shows the horizontal transport rates estimated from the measurements and those computed by the 1DV-model. The magnitudes of the computed horizontal transport rates are in agreement with the magnitude of “measured” transports rates since $U_d$ is prescribed, and $M$ and $w_*$ are used as fitting parameters. The total amount of sediment that is transported back and forth through the cross section at the measuring location can be estimated from Figure 5.6 at $5\times10^6$ kg for a channel width of 600 m. Ridderinkhof (1998) showed that the tidal induced net transport in landward direction can be up to 20 to 50% of this amount. If the “drainage area” of the channel Groote Gat upstream of the measuring location is estimated at approximately 15 km², and the bulk density is approximately 1300 kg/m³, the net transport would be equivalent to a net accretion of 0.1 - 0.2 mm/tidal cycle during calm weather conditions. These figures do not contradict the net accretion of approximately 1 cm/year according to GRAN, 1990 (see Section 2.2).
Figure 5.6. Sediment transport during the June measuring period in 1996: 0, "measured"; ---, no buoyancy (Run J1); ---, with buoyancy (Run J2); 2:00h June 27 - 3:00h June 28.

Ridderinkhof (1998) showed with an idealised numerical model of the Ems-Dollard that the parameter settings in the Partheniades-Krone formulations for erosion and deposition (see Appendix C, Eq. C.17 and C.18) do not alter the direction of the net horizontal flux of fine sediments. Net fluxes can be explained from the interaction between tidal asymmetries and the hysteresis in the Lagrangian sediment load-velocity curve. The degree of hysteresis determines the magnitude of the net flux (see also Groen, 1967).

Hysteresis between the tidal velocities and SSC is also present in the Eulerian measurements under consideration (see also Section 4.2, Figures 4.1 and 4.2). Figure 5.7 shows the sediment load derived from the measurements and those predicted by the 1DV-model plotted against the depth-averaged velocities $U_d$. Sediment loads were derived from the concentration measurements by extrapolation of the Rouse concentration profile (Eq. 4.3, Section 4.2) to the water surface, followed by integration over the water depth. The hysteresis between sediment loads and velocities is clearly present in both the measurements and the model results; the sediment load is appreciably lower during acceleration than during deceleration. Differences between the computed hystereses with and without buoyancy are most pronounced for periods of relatively small velocities (< 0.1 m/s) and seem less influenced by buoyancy than expected from the variations in SSC shown in Figure 5.1b. The main reason for this difference is that, although the SSC at higher levels in the water column decreases over a short period of time, the sediment is not deposited according to the 1DV-model as long as the bottom shear stresses exceed $\tau_{c,d}$. The present results were obtained from
Eulerian measurements, but it is noted once more that the net flux depends on the hysteresis of the Lagrangian sediment load-velocity curve.

Figure 5.7. Hysteresis of the Eulerian sediment load-velocity curve during the June measuring period in 1996: (a) 2:00h - 14:30h June 27; (b) 14:30h June 27 - 3:00h June 28; ◊, measurements; ---, no buoyancy (Run J1); ---, with buoyancy (Run J2).

Sediment-induced buoyancy effects in principle have two opposite effects on the degree of hysteresis of the Lagrangian sediment load-velocity curve. The first effect concerns the relatively low concentrations during slack water. Ridderinkhof (1998) shows with a schematic model of the Ems/Dolland that, in case of a limited amount of sediment available for transport, a low concentration during slack water causes an increase in the variations in the sediment load and in an increase in the Lagrangian hysteresis. The second effect concerns the dampening of the turbulence resulting from buoyancy effects, which causes lower concentrations during decelerating flow. This effect reduces the settling lag and thereby the hysteresis.

5.4 Limitation of sediment availability
Parallel to the turbulence measurements in the Groote Gat in June 1996, erosion experiments were carried out on the adjacent tidal flat Heringsplaat by means of an in-situ erosion flume (Kornman, 1998). During the experiments the bed was subjected to a stepwise increasing bed shear stress ranging from 0.06 - 0.7 Pa. Two types of surface erosion occurred: (1) a constant erosion rate at a given bed shear stress, and (2) an initially high erosion rate followed by a decrease to an approximately zero erosion rate after each shear stress increment. The first type is in agreement with the erosion of a fluffy top layer with constant properties (Kuijper et al., 1989). This type of erosion formed the starting point in the previous sections. The second type suggests that the
sediment bed strength increased during the test, i.e. the sediment available for erosion was limited for given bed shear stress. The second erosion type is schematically represented, in the 1DV-model, by limiting the depth-averaged sediment concentration to a maximum value ($C_{max}$) (see Table 5.1 Run J3). This type of erosion forms the starting point in this section. $C_{max}$ depends, among other things, on the spring-neap tide cycle, conditions during exposure, wave activity, and biological activity. Kornman and De Deckere (1998) found that a relationship existed between the sediment availability and biological activity in the Dollard during 1996.

![Diagram](image)

Figure 5.8. Some results from the 1DV-model for the June 1996 measuring period with a limited sediment concentration: (a) Model input: the water level, $h$, and the depth-averaged velocity, $U_d$, (b) Computational results of the SSC at 0.7 m for “limited” and “unlimited” sediment availability.

Figure 5.8 shows the computational results for the SSC at 0.7 m for $C_{max}$ set at 0.5 g/l. The erosion rate coefficient parameter $M$ is set at $4 \times 10^{-4}$ kg/m$^2$/s, ten times larger than in the computational results shown in Section 5.2. Three phases can be distinguished in the time series of the SSC in Figure 5.8, for example: a short erosion phase during 8:45h - 10:00h; a phase of maximum SSC during 10:00h -12:00h; a deposition phase during 12:00h -15:30h. The most important differences from the variations in SSC for an unlimited sediment availability are that the increase in SSC during flood is steeper, and that the SSC during maximum ebb and flood are similar,
despite the ebb dominance. Both aspects improve the agreement with the measurements of SSC which is enabled by the additional “fitting parameter” added to the 1DV-model.

Figure 5.9. Variations of SSC during the second tidal cycle of the June measuring period: (a) Measured SSC at 0.3 m and 1.4 m, (b) Computational results (Run J3): SSC at 0.3 m and 1.4 m; 2:00h - 16:00h June 27, 1996.

Figure 5.9 compares the measured and computed variations in SSC at 0.3 m and 1.4 m above the sediment bed. There is some resemblance between the vertical distributions of SSC: relatively large gradients during acceleration and deceleration, and small gradients during maximum ebb and flood currents. The computations indicate that the measured peaks in SSC at 0.3 m can partly be explained from accumulation of sediment close to the bed during acceleration before entrainment into the water column, and during deceleration before/during deposition. At 13:00h the simultaneous increase in measured SSC at 0.3 m and decrease at 1.4 m suggests a collapse of the concentration profile. A similar effect can be found in the computational results, although at a later moment. Without buoyancy such effects are hardly present in the computational results (not shown).
Figure 5.10. Variations of vertical turbulent vertical transports of sediment during the second tidal cycle of the June measuring period: (a) Measured values of $\bar{c}w$ at 0.1 m, 0.4 m, and 1.4 m; (b) Computational results (Run J3): $\bar{c}w$ at 0.1 m, 0.4 m, and 1.4 m; 2:00h - 16:00h June 27, 1996.

Figure 5.10 shows the vertical turbulent transports of sediment at three levels above the sediment bed. The trends in the computed transports are to a certain extent comparable to the measured transports: large vertical transports in the beginning of the ebb and flood periods, then a period of approximately constant transports during maximal current velocities, and finally a reduction in vertical transports towards slack water. Differences between ebb and flood in $\bar{c}w$ are small compared to those presented in Figure 5.3 for the case of unlimited sediment availability.

Another aspect of the computational results, which is of importance for the interpretation of the measurements, is that $\bar{c}w$ does not vary much over the lowest meter of the water column which is in agreement with the measurements.

Figure 5.11 shows $Ri_g$ at three levels above the bed. The agreement between the measured and computed turbulent vertical transports of sediment is reflected in that between the $Ri_g$ values. Both the measured and computed values of $Ri_g$ show that at higher positions in the water column the stratification is most pronounced.
Figure 5.11. Gradient Richardson numbers during the second tidal cycle of the June measuring period: (a) $R_i$ derived from the measurements at 0.1 m, 0.4 m, and 1.0 m; (b) Computational results (Run J3): $R_i$ at 0.1 m, 0.4 m, and 1.0 m; 2:00h - 16:00h June 27, 1996.

5.5 Settling velocity depending on concentration and turbulence

Settling velocity measurements were carried out in the upper part of the water column of the tidal channel Groote Gat on June 26, 1996, by means of an underwater video camera (Van der Lee, in prep). Floc sizes and settling velocities varied during the tidal cycle: the largest flocs occurred during periods of maximum SSC with settling velocities up to several mm/s. Small flocs predominated during periods of low velocities and low SSC. Van Leusen (1994) suggested that the ultimate settling velocity is the result of a combination of processes where the residence times of flocs in regions of high shear, cohesion, and SSC play an important role. Winterwerp (1998) presented a flocculation model in which these processes are incorporated. The description of such a flocculation model is beyond the scope of this study. Herein some empirical relations are used which describe the settling velocity's dependence on SSC and turbulence (Mehta, 1986; Van Leusen, 1994; see also Appendix C):

$$w_{s,0} = k_1 \overline{C}^m \quad \overline{C} < 1 \text{ g/l}$$  \hfill (C.21)
\[ w_{s,0} = w_{s,\max} (1 - k_2 \bar{C})^\beta \quad \bar{C} \geq 1 \text{ g/l} \]  
\[ w_s = \frac{w_{s,0} (1 + aG)}{1 + bG^2} \]  

where \( k_1, k_2, m, w_{s,\max}, \beta, a \) and \( b \) are coefficients (see Table 5.2). The shear parameter \( G = \sqrt{\nu/s} \) is a measure for the velocity gradients at the dissipation scale. Eq. C.21 describes the dependence of the settling velocity on the SSC, Eq. C.22 describes the hindered settling behaviour, and Eq. C.23 describes the dependence of \( w_s \) on the shear stresses at small length scales (represented by \( G \)). All effects are accounted for by combining the settling velocity found from Eq. C.21 or C.22 and Eq. C.23 (Van Leussen, 1994).

| Table 5.2. Coefficients used in Eq. C.21 through C.23. |
|---|---|---|---|
| Eq. C.21 | \( k_i \) m/s (g/l)^m | \( m \) (-) | Source: Van Leussen, 1994: Obtained from in-situ settling tube measurements in the Ems/Dolland. |
| | 1.5 \times 10^{-3} | 1.2 | |
| Eq. C.22 | \( w_{s,\max} \) (m/s) | \( k_3 \) (g/l)^1 | \( \beta \) (-) | Source: Thorn, 1981, in Mehta, 1986: Lab. experiments with mud from the Severn (U.K.) |
| | 2.6 \times 10^{-3} | 0.008 | 4.65 |
| Eq. C.23 | \( a \) (s^-1) | \( b \) (s^2) | Source: Van Leussen, 1994: Lab. experiments with mud from the Ems/Dolland |
| | 0.3 | 0.09 | |

Accounting for the dependence of the settling velocity on SSC by applying Eq. C.21 and C.22, and ignoring the effects of turbulence, results in relatively strong differences in settling velocity between periods of low and high sediment concentration and between the 0.3 m and 1.4 m level. The latter effect has consequences for the vertical structure of the flow, close to the sediment bed the computed SSC and \( w_s \) are relatively large. If no deposition occurs, relatively large turbulent sediment transports are computed close to the sediment bed to counterbalance the high settling rates. This effect results in large values for \( \bar{c}w \) and large vertical gradients in \( \bar{c}w \) as shown in Figure 5.12c. These large gradients in the vertical turbulent transports and the large values of these transports are not in agreement with the measurements of \( \bar{c}w \) as shown in Figure 5.10 and again in Figure 5.12a.

It can be inferred that no large variations in settling velocity were present in the bottom boundary layer at the measuring site (\( z < 1.0 \text{ m} \)). Eq. C.21 and C.22, with the parameters shown in Table 5.2, are at least incomplete for describing the settling velocity. A reduction of \( k_i \) in Eq. C.21 reduces the heights of the peaks in \( \bar{c}w \) but on
the other hand causes seemingly unrealistic high SSCs during slack water. Van Leussen (1994) found from laboratory experiments a reduction in the parameter \( m \) for increasing levels of \( G \). The latter could partly explain the discrepancy between the computational results and the measurements.

Figure 5.12. Vertical turbulent transports of sediment during the second tidal cycle of the June measuring period: (a) Measured values of \( \bar{c}w \) at 0.1 m, 0.4 m, and 1.0 m; (b) to (d) Computational results (Runs J3, J4 and J5): \( \bar{c}w \) at 0.1 m 0.4 m, and 1.0 m; legends see Figure 5.10; 2.00h-16.00h June 27, 1996.

When only hindered settling is taken into account (Eq.C.22), the computational results do not differ much from those with a constant settling velocity. Near-bed
concentrations do not become high enough for the hindered settling effects to have a significant effect on the deposition rates. However, including hindered settling is essential for modelling the formation of fluid mud layers in deeper parts of estuaries and coastal seas (Winterwerp et al., 1998).

Eq. C.23 was implemented in order to describe the settling velocity's dependence on the turbulence level. At higher levels in the water column, where the values of $G$ are relatively small, the computed behaviour of the settling velocity remains in accord with the observations of Van der Lee (in prep.). At lower levels in the water column, where $G$ can reach large values, the computed settling velocities are partly controlled by Eq. C.23.

The result of implementing Eq. C.23 is that smaller gradients in SSC are found in the bottom region during large flow velocities, which can be explained from a reduction in the settling velocities. The results for $\overline{\omega w}$ are shown in Figure 5.12d. During periods of stratification and low flow velocities the influence of Eq. C.23 is largely reduced which results in overestimations of the vertical transports, similar to those in Figure 5.12c.

It can be concluded that the settling velocity's behaviour described by Eq. C.21 through C.23 is not confirmed by the measurements. The assumption of a constant settling velocity proves to be the best option for the description of the small gradients in $\overline{\omega w}$. A flocculation model which describes the growth and decline of sediment flocs on the appropriate time scales could possibly connect the observations made by Van der Lee (1998) and Van Leussen (1994) in the upper part of the water column and the measurements in the bottom region presented herein.

5.6 Analysis of the August measurements
The analysis of the August measurements was carried out with the 1DV-model that comprises a limited sediment availability and a constant settling velocity. The assumption of a limited sediment availability combined with an increased erosion rate made it possible to simulate important features of the June measurements such as the steep increase in SSC during floods and the approximately similar maximum SSC during ebbs and floods, despite the ebb dominance. The assumption of a constant settling velocity proved to be the best option for describing the small gradients in vertical turbulent transports in the bottom region. Table 5.3 shows some of the fitting parameters used for the computations with the 1DV-model.
Table 5.3. The fitting parameters used in the 1DV POINT MODEL for the August simulation.

<table>
<thead>
<tr>
<th>Run:</th>
<th>( w_s ) (m/s)</th>
<th>( M ) (kg/m²/s)</th>
<th>( C_{max} ) (kg/m³)</th>
<th>Buoyancy</th>
</tr>
</thead>
<tbody>
<tr>
<td>A1.</td>
<td>( 5 \times 10^{-4} )</td>
<td>( 2 \times 10^{-4} )</td>
<td>0.25</td>
<td>Included</td>
</tr>
</tbody>
</table>

\( C_{max} \) and \( M \) were set at 50\% of those used for the June simulations since the SSC during the August period were significantly lower compared to the June period. The reasons for this difference must be sought in changes in processes that have not been modelled, such as biological processes or erosion by surface waves.

![Graph showing water levels and velocities](image)

Figure 5.13. The model input and the SSC during the August 1996 measuring period: (a) The water level, \( h \), and the depth-averaged velocity, \( U_d \) (the bottom elevation was -3.3 m N.A.P.); (b) Computational results for the SSC at 0.7 m (A1), and the measured SSC at the same level.

Figure 5.13 shows the prescribed \( U_d \) and \( h \), and the computed and measured SSC. A remarkable feature of the tidal cycles in the August period are the relatively long flood periods, which is also reflected by the computed variations in SSC.

The 1DV-model cannot reproduce the decrease in the measured SSC towards high water. This decrease is probably caused by horizontal advection effects (the decrease in SSC in the seaward direction). It is noted that if no buoyancy effects are taken into
account, the low SSC during slack water cannot be reproduced unless a large settling velocity is assumed (> 1 mm/s).

Figure 5.14. Measured and computed shear stresses at 0.4 m above the sediment bed; August 8, 1996.

Figure 5.14 shows the measured and computed shear stresses at 0.4 m above the sediment bed. The computed shear stresses are in good agreement with the measured stresses. That is because the roughness height $z_0$ used in the computations, was derived from the stress and velocity measurements (see also Appendix C).

The measured and computed vertical transports are shown in Figure 5.15a and 5.15b, respectively. Variations in the measured $\overline{cw}$ and variations in measured SSC are very similar (compare Figures 5.15a and 5.13; see also Section 4.4 Figures 4.14 and 4.15). This indicates that $\overline{cw}$ represents the turbulent flux that counterbalances the settling of sediment. The peaks in $\overline{cw}$ at the start of the acceleration of the tidal flow are low (see arrows). The computational results show a different picture: over a short period of time the material is eroded and brought in suspension (see the high peaks in Figure 5.15b). After the erosion the computed vertical transports are more or less constant and represent the flux that counterbalances the settling of sediment. The order of magnitude of these fluxes are larger than measured fluxes. The relatively small gradient in vertical fluxes is in agreement with the measured gradients. The latter result confirms the findings obtained from the analysis of the June measurements: no large variations are present in the settling velocities in the bottom region.
Figure 5.15. Variations of turbulent vertical transports of sediment during the August measuring period: (a) Measured values of $\bar{c}_w$ at 0.1 m and 0.4 m; (b) Computational results (Run A1): $-K_m/\sigma_z^2 \partial \bar{C}/\partial z$ at 0.1 m, 0.4 m, and 1.0 m; August 8, 1996.

5.7 Discussion
Additional insight has been obtained into some typical aspects of the fine sediment behaviour by comparing the results of the June and August measurements to a relatively simple one-dimensional vertical model. The model results showed that during periods around slack water the relatively rapid decrease in SSC and high deposition rates can be explained from buoyancy effects: a positive feed back mechanism developed in which decreasing turbulence production caused increasing concentration gradients and vice versa. This process is reflected in the increase in the measured and computed gradient Richardson numbers. In Sections 4.5 and 4.7 it is clearly shown from the reduced turbulence length scales and correlation coefficients for the Reynolds stresses, that during the June measuring period the locally measured density gradients affect the local turbulence structure. During periods of maximum flow velocity, especially close to the sediment bed, buoyancy plays a minor role.
Rapid reductions in SSC and increased settling rates caused by sediment-induced buoyancy have opposite effects on the degree of hysteresis of the SSC-velocity curves. The degree of the Lagrangian hysteresis determines the net transport of sediment. In order to investigate the actual effect of buoyancy on the net transport it should be incorporated in a 3D numerical model for the entire Ems/Dollard estuary. However, it is unlikely that the overall direction of sediment transport is changed by buoyancy effects.

It is mentioned herein that the $q$-$e$ turbulence model possibly overestimates the reductions in shear stress during periods of significant stratification (Uittenbogaard, 1995). The turbulent Prandtl number $\sigma^*_t$ is set to 1.0 in the 1DV-model whereas it may be in fact a function of $Ri_e$ (see also Section 4.7). The large reductions in shear stress in combination with acceleration effects can cause an overestimation of the distortions of the velocity profiles during the turning of the tide (see Figure 5.5). A Munk-Anderson type relation can be applied in order to model $\sigma^*_t$ (Galland et al., 1997; Winterwerp et al., 1998), or Eq. 4.41 and 4.42 (Schuman and Gerz, 1995).

In-situ flume experiments on the Heringsplaat showed basically two types of erosion behaviour: (1) a constant erosion rate for a given bed shear stress, (2) a decreasing erosion rate for a given bed shear stress. Both types are schematically modelled by assuming an unlimited and a limited sediment supply, respectively. The computational results show that only the second erosion type can reproduce the steep increases in SSC, and similar maximum SSCs during flood and ebbs, of the June measuring period. Comparison of the model results to the August measurements shows that advection ten played a major role in the variations in SSC: the reductions in SSC during increasing flow velocities cannot be reproduced by the 1DV-model. No decisive conclusions can be drawn about the type of erosion that dominates in the Dollard during the June and August measuring periods on the basis of the comparison of the computed variations in SSC, since advection is not accounted for.

Settling velocities were estimated from the experiments at approximately 0.5 mm/s. Larger settling velocities led to too large computed vertical turbulent transports of fine sediment and too large concentration gradients compared to the June and August measurements. Smaller settling velocities led to too small computed differences in SSC between maximal flow velocities and slack water.

The mean settling velocities of a few mm/s found by Van der Lee (1998) in the upper part of the water column during maximum SSC do not necessarily contradict the
much smaller settling velocities presented in this study for the near-bottom region. The comparison of the computational results to the measurements showed the limited applicability of Eq.C.21 through C.23 in combination with the $q-c$ turbulence model. A flocculation model such as the one developed by Winterwerp (1998), in which the basic process of floc growth and break-up are described on the appropriate time scales, could possibly aid in connecting observations made by Van der Lee (1998) and the findings presented herein.
Chapter 6 Laboratory experiments

6.1 Introduction
The erosion and strength properties of bed material from the Dollard were studied in the laboratory. A number of experiments were carried out in a rotating annular flume with different flow velocities and consolidation times (see Figure 6.1). The main purpose of the laboratory experiments was to study bed strength and aspects of the water-bed exchange processes under controlled circumstances. These exchange processes received little attention in the previous chapters, mainly because of the lack of proper field equipment to make measurements in the narrow transition zone between sediment bed and water column. The laboratory experiments should therefore be considered as supplementary to the field measurements rather than as replacements. The results of the laboratory experiments are compared to data from similar laboratory experiments on other types of fine sediment (Winterwerp and Kranenburg, 1997) and to results of field measurements (Kornman, 1998; Mitchener et al., 1998).

Figure 6.1. Rotating annular flume of Laboratory of Hydromechanics, Delft University of Technology.

The cohesive sediment used in the annular flume originates from underwater deposits in the Buiten Aa, about 5 km upstream of the measuring location. As the mineral composition of the fine sediment fraction of the deposits shows little variation
within the Ems/Dollard (Section 2.5) it may be assumed that the mineral composition of the fine sediment fraction of these deposits was similar to that of the measuring location. The sediment consolidated and froze during storage after which it was mixed and homogenised again. As a consequence in-situ properties such as stress history and surface layer properties were lost. Biological constituents may have altered during collection, transportation and storage of the sediment and during the preparation of the sediment bed. Nevertheless, it may be assumed that the sediment beds prepared shared important physicochemical characteristics with underwater deposits and fluffy layers as they appear in the Dollard. (e.g. Verbeek et al., 1994). These characteristics comprise, among other things, the ability of the sediment to: flocculate and settle, compact (consolidate), liquefy or swell, and to quickly change its internal structure upon loading resulting in visco-plastic or visco-elastic behaviour (De Wit, 1995; Van Kessel, 1997).

Two types of beds were tested in the annular flume: (I) beds formed from deposition with relatively short consolidation times (3 hours to 4.5 days), and (II) placed sediment beds. Type (I) beds represent fresh underwater deposits which can be formed during slack water periods in tidal flows. They are easily liquefied and transported or re-suspended into the water column (De Wit, 1995; Winterwerp and Kranenburg, 1997; Van Kessel, 1997). Type (II) beds are representative of semi-permanent sediment layers with higher strengths. These layers are formed by compaction where the expulsion of pore water is caused by self-weight consolidation or by evaporation. Such semi-permanent layers were present along borders of the Heringsplaat during the measuring periods in 1995 and 1996. It is not clear whether type (II) beds can be liquefied, or if erosion only occurs as surface erosion (Cornelisse et al., 1992).

Section 6.2 briefly describes the annular flume and the most important properties of the flow in it. An operating method is described, experimental procedures and programme are presented. Section 6.3 presents results of physicochemical and mechanical tests on the sediments collected. The results of the annular flume experiments on re-entrainment and erosion of layers of Dollard mud are presented in Sections 6.4 and 6.5, respectively. In Section 6.6 these results are summarised and implications for the interpretation of the field measurements are discussed.

6.2 Measuring facilities, procedures and program
The experiments were carried out in the annular flume of the Laboratory of Hydromechanics of the Delft University of Technology. A sketch of the annular flume
is shown in Figure 6.2. The radius $R$ of the annular flume is 1.85 m, the width $W$ is 0.3 m and the height $d$ is 0.4 m.

The flow within the flume is driven by a ring-shaped rotating top lid. The flume is rotating in the opposite direction in order to minimise undesired secondary flow cells. The optimal condition for erosion studies is defined as the condition of minimal variation of the bed shear stress. For that condition Booij (1994) found from clear water experiments:

$$\frac{\omega_t}{\omega_f} + 1 = -1.17 \frac{H}{W}$$

(6.1)

where $\omega_t$ and $\omega_f$ are the angular velocities of the top lid and the flume, $H$ is the water depth and $W$ the width of the flume. It is assumed herein that if a layer of mud is present at the bottom of the flume, Eq. 6.1 still approximates the optimal conditions.

The development of the boundary layers at the top lid, side walls and mud layer is suppressed by the rotation of the annular flume (Booij, 1994). This results in plug flow conditions in the core of the water layer. The flow velocity in the core with respect to the flume is referred to as $U_c$ (Figure 6.2, Section A-A).

![Diagram of the annular flume](image)

Figure 6.2. Sketch of the annular flume of Laboratory of Hydromechanics.

The operating method of the annular flume is as follows: both top-lid and flume are slowly accelerated in the same direction in such a way that the velocities of both the water layer and the fluid mud can adapt to the prescribed speed of the flume; the entrainment is initiated by rotating the top-lid in the opposite direction (Winterwerp and Kranenburg, 1997). The shear thus generated produces turbulence in the water layer and, as a consequence, resuspension or erosion of bed material.
Deposited beds
The procedures for the entrainment experiments with deposited beds were similar to those adopted by Winterwerp and Kranenburg (1997), see also Van der Ham (1996). A layer of fluid mud was formed from deposition of a dense suspension, and was re-entrained by the turbulent water layer. This was done for various consolidation times ranging from 3 hours to 4.5 days and mean water layer velocities, $U_c$, ranging from 0.2 m/s to 0.7 m/s. The entrainment programme is summarised in Appendix D.

Flow velocities and concentrations were measured at 0.1 m and 0.2 m above the bottom of the flume by means of electromagnetic flow meters (EMF, 20 mm diameter sensor heads, wt | delft hydraulics) and optical turbidity meters, respectively (OSLIMs and FOSLIMs, wt | delft hydraulics, see also Chapter 3). Parallel to the entrainment experiments, bed density profiles were measured in measuring cylinders by means of an electrical conductivity probe (ECP, wt | delft hydraulics). Strength profiles were determined with a sounding apparatus (Van Kessel, 1997). The results are described in Section 6.3.

Placed beds
Prior to the placed bed erosion experiment, a pilot test was carried out. The sediment bed had a yield strength between 100 and 200 Pa. Except for a small "runnel" at the outer glass wall the bed remained intact, even during the maximal attainable speeds of lid and flume (2 m/s). This runnel probably resulted from scour by small pellets. Based on these experiences, the preparation of the placed bed was adapted. The sediment was homogenised by intensive mixing until all pellets were broken up. Water was added until a slurry was obtained with a yield strength between 20 and 40 Pa (Section 6.3). The slurry was pumped into the annular flume layer by layer. The surface of the slurry was levelled manually. Top-lid and flume speeds were increased stepwise, with $U_c$ ranging from 0.15 to 1.05 m/s. The programme of erosion tests is presented in Section 6.5 and Appendix D.

Six pressure sensors were installed in order to monitor the pressures inside the mud layer: three at 0.03 m, and three at 0.01 m below the water-mud interface. Any flow velocities inside the mud layer were recorded by means of an EMF which was placed at 0.03 m below the water-mud interface. Small size test beds were created in measuring cylinders, following a similar procedure as in the annular flume, in which yield strength profiles and density profiles were measured.
6.3 Properties of fine sediments from the Dollard

The mud was collected by means of a Van Veen grab from the bottom of the tidal channel Buiten Aa 7°13' E, 53°15' N, on December 4 1996. The water temperature was 4° C and the salinity was 12 %. The mud had a grey colour and it had a jelly-like appearance. The firmness quickly reduced when the mud was remoulded.

In previous annular flume studies it was found that the accuracy of annular flume results is too low to correlate the erosion results with any physicochemical bulk parameters other than the density and the sand content (Winterwerp et al., 1992). However, for the sake of completeness, some physicochemical properties are listed in Table 6.1. The CEC values are comparable to those found in other regions of the Ems/Dollard estuary (Van Leussen, 1994). The low sand content of 2.8% is typical of cohesive sediment in the inner reaches of estuaries and tidal basins.

Table 6.1. Physicochemical properties of the mud/pore water from the Dollard.

<table>
<thead>
<tr>
<th>Property</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density of the solids (kg/m³)</td>
<td>2514</td>
</tr>
<tr>
<td>Water content</td>
<td>180 %</td>
</tr>
<tr>
<td>Salinity of pore water</td>
<td>8 %</td>
</tr>
<tr>
<td>Loss on ignition</td>
<td>15 %</td>
</tr>
<tr>
<td>Cation exchange capacity (CEC, meq per 100 g)</td>
<td>27.7</td>
</tr>
<tr>
<td>Percentages: Sand (&gt; 63 μm) / Silt / Clay (&lt;2 μm)</td>
<td>2.8% / 47.9% / 49.3%</td>
</tr>
</tbody>
</table>

1Percentage of the dry mass.

Settling velocities

Settling velocities were determined from the initial settling rates of homogeneous suspensions. It was found that the settling velocity is approximately $5 \times 10^{-4}$ m/s for sediment concentrations of 20 - 40 g/l. It was significantly lower for concentrations of about 60 g/l namely $4.4 \times 10^{-5}$ m/s ± 10%. This reduction in settling velocity is attributed to hindered settling effects (see also Section 5.5). Homogeneous suspensions of about 100 g/l sometimes initially showed no signs of settling at all, but after a certain undefined time, the sediment did start to settle. Surprisingly, each time the settling velocity then was approximately $2.8 \times 10^{-5}$ m/s ± 10%. The settling velocity did not change during the entrainment experiments.

Rheological properties

Some rheological properties of the sediment were determined by means of a rheometer (Physica, VDS200) and a five blade torque vane (height 20 mm, diameter 10 mm). The mud was remoulded before each vane test. The vane-tests started approximately after
one minute rest time ($\Delta T$). The angular velocity of the vane was 0.03 rad./s. Figure 6.3 shows some results for the peak strength and remoulded strength of the Dollard mud for different sediment concentrations.

![Figure 6.3. Peak strength and remoulded strength of remoulded samples of Dollard mud as functions of concentration ($\Delta T = 1$ minute, angular velocity is 0.03 rad./s).](image)

Peak strengths were reached at a rotation angle, $\gamma_p$, of about 0.5 rad. for all concentrations examined. Both peak strengths and remoulded strengths increased by more than one order of magnitude if the dry density increased from 300 g/l to 500 g/l, and also the ratio of peak strength to remoulded strength increased. The latter result suggests that the ability of the mud to quickly regain strength increases with concentration. The peak strengths presented in Figure 6.3 are not necessarily
representative of those of other sediment beds with similar density. They depend to a large extent on the stress history, structure and composition of the bed.

Figure 6.4 shows the strengths of the Dollard mud as a function of time. No large increase in strength at the sediment surface was found, as expected, since no overlying sediment layers causing self-weight consolidation were present. The top layer of the sediment turned brown during the test which is an indication of changes in the chemical structure (oxidation of the iron present in the mud). The angle $\gamma_p$ was approximately 0.5 rad. after 120 minutes rest time, but reduced to about 0.2 rad. after 4000 minutes rest time. The increase in peak strength and rigidity, and decrease in remoulded strength is attributed to thixotropy.

Properties of the beds formed from deposition

![Graph](image)

Figure 6.5. Sediment bed characteristics for increasing consolidation times; (a) dry density profiles, $C_{bed}$ ($z$); (b) strength profiles resulting from sounding tests, $\tau_r$ ($z$), and a theoretical yield strength profile calculated from $C_{bed}(z)$ for 4.5 days consolidation time.

Density profiles and yield strength profiles were determined in measuring cylinders columns in order not to disturb the deposited beds in the flume prior to entrainment experiments. Figure 6.5a shows that the interface height decreased from 0.13 m to 0.06 m over a period of 4.5 days, and the concentration of the bed $C_{bed}$ increased accordingly. A similar settling/consolidation behaviour was found in the annular flume. Figure 6.5b shows that $\tau_r$ is relatively small in the upper part of the sediment bed ($\tau_r < 1$ Pa) for short consolidation times (3 and 6 hours). In the lower part significant values for $\tau_r$ are found ($> 5$ Pa), which result from the weight of the overlying sediment layers. Assuming a zero horizontal effective stress, the maximum theoretical yield strength, apart from cohesive effects, equals one half of the vertical effective stress (Van Kessel, 1997). This maximum value is shown in Figure 6.6 as a dot-dash line.
The results for 4.5 days consolidation time indicates that this theoretical strength was approached gradually.

**Properties of placed beds**

Figure 6.6 shows concentration and yield strength profiles in a small size test bed which is assumed to be representative of the placed bed in the annular flume. The large densities and yield strengths near the surface may have been caused by the levelling procedure. Some yield strength profiles showed evidence of a layered structure (not shown). The yield strength of the bulk is between 30 and 40 Pa, which is significantly higher than the yield strength of beds formed from deposition (see Figure 6.5b), and much higher than the average bed shear stresses in the field and in the annular flume. Generally, these mean bed shear stresses did not exceed 2 Pa.

![Figure 6.6. Sediment bed characteristics of placed beds of Dollard mud: (a) dry density profiles, $C_{sed}(z)$; (b) measured strength profiles, $\tau_s(z)$, at two different positions.](image)

6.4 Entrainment of fluid mud layers

The entrainment of a fluid mud layer was monitored visually at the outer glass sidewall, and by measuring the increase in concentrations in the turbulent water layer. The latter can be translated in a decrease in the thickness of the mud layer $\delta$, as the dry density profile of the fluid mud layer is known (see Figure 6.5a). As an example, the results for entrainment experiment E12 are plotted in Figure 6.7.

The entrainment velocity $w_e$ is defined as $-d\delta/dt$, was constant in an initial stage of the entrainment process, then it decreased after about 15 minutes and became zero after 45 minutes. The remaining mud layer turned out to be too strong to be entrained into the turbulent water layer. The differences between the observed variations in $\delta$ and those calculated from the concentrations during the initial entrainment phase might
have resulted from the centrifugal force, which causes relatively high values of $\delta$ at the outer wall (Winterwerp and Kranenburg, 1997).

Figure 6.7. Results of entrainment experiment E12 (one minute averaged): (a) thickness of the fluid mud layer, $\delta$, and concentrations; (b) tangential velocities $\overline{U}$; (c) smoothed relative turbulence intensities $\sqrt{\overline{u'^2}/U_c}$ at 0.2 m, 0.1 m and 0.03 m.

It appears from Figure 6.7b that a significant velocity difference was present between the turbulent water layer ($\overline{U}$ at 0.2 m) and the upper part of the fluid mud ($\overline{U}$ at 0.1 m) during the initial stage of the entrainment process. No movements of the fluid mud were detected at 0.03 m during the first 20 minutes of the experiment. It can be concluded that during the entrainment process the main part of the fluid mud remained
stationary with respect to the flume. Similar conclusions were drawn from the other entrainment experiments.

Figure 6.7c shows the (smoothed) relative turbulence intensities in the turbulent water layer and in the fluid mud. It is estimated that the high-frequency losses for $u^2$ due to sensor size and the cut-off frequency of 10 Hz, is smaller than 20% at 0.1 m and 0.2 m above the bottom of the flume (see also Booij, 1992, Figure 3.6). Clear water experiments with conditions similar to experiment E12 showed that near the bottom of the flume the relative turbulence intensity is about 0.1 (dotted line in Figure 6.7c). It can be concluded from Figure 6.7c that the turbulence was suppressed at the 0.1 m level. The turbulence suppression is attributed to the presence of the mud. The fact that velocity fluctuations were present at 0.03 m does not necessarily indicate that the flow was turbulent, viscous effects may still have dominated the behaviour of the flow.

Table 6.2 shows some results of the initial stage of the entrainment experiments E11 to E14. The dimensionless entrainment rate $E_*$ is defined as $w_\tau / u_\tau$, the overall Richardson number $Ri_\tau$ is defined as $\Delta \rho g H / \rho_m u_\tau^2$, where $\Delta \rho$ is the density difference between the turbulent water layer and the sediment bed, $H$ is the thickness of the turbulent water layer, $\rho_m$ is density of the turbulent water layer, and $u_\tau$ is the friction velocity at the top-lid where the turbulence is mainly produced. The friction velocity $u_\tau$ and mean water layer velocity $U_c$ were obtained from a streamwise momentum balance applied to the water layer (Booij, 1994; Kranenburg and Winterwerp, 1997).

<table>
<thead>
<tr>
<th>ID</th>
<th>Calculated</th>
<th>Measured</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$U_c$</td>
<td>$u_\tau$</td>
</tr>
<tr>
<td></td>
<td>(m/s)</td>
<td>(m/s)</td>
</tr>
<tr>
<td>E11</td>
<td>0.20</td>
<td>0.012</td>
</tr>
<tr>
<td>E12</td>
<td>0.35</td>
<td>0.022</td>
</tr>
<tr>
<td>E13</td>
<td>0.72</td>
<td>0.046</td>
</tr>
<tr>
<td>E14</td>
<td>0.72</td>
<td>0.046</td>
</tr>
</tbody>
</table>

Entrainment rates in the initial stage

The values of the initial entrainment rates $E_*$ shown in Table 6.2 are plotted in Figure 6.8. Also shown are entrainment rates for Kaolinite and mud from the Caland channel and those calculated with a simplified entrainment model (Winterwerp and
Kranenburg, 1997). The error bars plotted in Figure 6.8 are based on the differences in visually observed $w_e$ and those calculated from the concentrations.

The simplified entrainment formula for the initial stage as presented by Winterwerp and Kranenburg (1997), reads:

$$E_\ast = \frac{C_w}{(2C_d H/W)^{0.5}} \frac{1}{(Ri_\ast + C_q)}$$  \hspace{1cm} (6.2)

where $C_w = 0.07$ and $C_q = 5.6$ are model coefficients which were derived from entrainment experiments with salt and fresh water, and $H/W = 0.55$. It was derived for the following conditions: (1) the upper part of the fluid mud layer is dragged along with the turbulent water layer by viscous effects; (2) settling effects in the turbulent water layer are not important yet; (3) entrainment rates are not (yet) reduced by the yield strength of the mud; (4) turbulence production is only due to shear production at the sidewalls and top lid. Condition (1) cannot be verified by means of the present velocity measurements, because the spatial resolution of the EMF sensor is not sufficient (about 20 mm). Large $Ri_\ast$ are often associated with small $u_\ast$ which makes assumption (3) critical (e.g. experiment E11, Table 6.2). Condition 4 is supported by the turbulence measurements presented in Figure 6.7c.

\[ \text{Figure 6.8. Initial entrainment rates } E_\ast \text{ versus the Richardson number } Ri_\ast. \]

It appears from Figure 6.8 that the trend in $E_\ast$ is in agreement with the data of Winterwerp and Kranenburg (1997) and that the trend seems to be well described by Eq. 6.2, if the error bands shown in Figure 6.8 are taken into account. It should be
mentioned that the data from experiments E1-E8 should be interpreted with care because the acceleration phase of the starting procedure was carried out too fast.

**Influence of yield strength on the entrainment rates**

The yield strength (up to 1 Pa) at the surface of the fluid mud in experiment E11 proved to be large enough to resist the turbulent shear stress of the order of 0.1 Pa. The upper part of the fluid mud showed some wave-like motions and 30 minutes after the start of the experiment cracks appeared at the mud interface. However, no floc erosion occurred and the fluid mud layer consolidated during the experimental period: \( \delta \) decreased from 0.163 m to 0.155 m over a 2.5 hour period. No effective stresses can have developed in the top layer of the mud during the three hours consolidation period and therefore this top layer may have had some similarity with the fresh deposits in tidal areas of several millimetres thickness (fluffy layers). The resistance against erosion of such layers results from the so-called true cohesion.

In experiments E9 and E10 beds with relatively large yield strengths were eroded. The consolidation times were 19 hours and 4.5 days, respectively (see Figure 6.5 for density and strength profiles). The flow velocity was increased stepwise. It was observed that at a certain instant during the erosion experiment, the erosion rate decreased markedly. The bed shear stress at this instant was equated to the critical shear stress for erosion, \( \tau_{ce} \). These critical shear stresses and the corresponding yield stresses are listed in Table 6.3. The bed shear stresses \( \tau_b \) were calculated from \( C_d U_t^2 \), where \( C_d = 1.8 \times 10^{-3} \). Table 6.3 shows that, as expected, some correlation exists between \( \tau_b \) and \( \tau_{ce} \). It can be concluded that \( \tau_{ce} \) is one order of magnitude lower than \( \tau_b \) which is in agreement with the findings of Zreik et al. (1998).

<table>
<thead>
<tr>
<th>ID</th>
<th>( \tau_{ce} ) (Pa)</th>
<th>( \tau_b ) (Pa):</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td>E9</td>
<td>0.2</td>
<td>1.7</td>
<td></td>
</tr>
<tr>
<td>E9</td>
<td>0.5</td>
<td>2.7</td>
<td></td>
</tr>
<tr>
<td>E9</td>
<td>&lt; 0.9</td>
<td>( \sim 15 )</td>
<td>all mud eroded</td>
</tr>
<tr>
<td>E10</td>
<td>1.0</td>
<td>7.8</td>
<td>not completely stable</td>
</tr>
<tr>
<td>E10</td>
<td>&lt; 1.8</td>
<td>( \sim 30 )</td>
<td>all mud eroded</td>
</tr>
</tbody>
</table>

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6.5 Some results on erosion of a placed bed

Placed bed erosion experiments were carried out in order to obtain a better insight into the relation between the yield strength and the erosion characteristics. Turbulent flow induces peak stresses at the bed surface and inside the sediment bed. The latter were calculated from the measured pressure fluctuations at 0.03 m and 0.01 m below the water-mud interface as explained below. Table 6.4 shows some results of the velocities, estimated mean bed shear stresses, and measured pressure fluctuation intensities inside the placed bed. The values of $\tau_b$ are significantly smaller than the values of $\tau_\gamma$ presented in Figure 6.6.

Table 6.4. Velocities, bed shear stresses and rms-values of pressure fluctuations, placed bed experiment.

<table>
<thead>
<tr>
<th>Step</th>
<th>$U_r$ (m/s)</th>
<th>$\tau_b$ (Pa)</th>
<th>$p'$ (Pa)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.14</td>
<td>0.05</td>
<td>&lt; 3</td>
</tr>
<tr>
<td>2</td>
<td>0.34</td>
<td>0.26</td>
<td>&lt; 3</td>
</tr>
<tr>
<td>3</td>
<td>0.58</td>
<td>0.65</td>
<td>3</td>
</tr>
<tr>
<td>4</td>
<td>0.80</td>
<td>1.21</td>
<td>7</td>
</tr>
<tr>
<td>5</td>
<td>0.96</td>
<td>1.95</td>
<td>-</td>
</tr>
</tbody>
</table>

Figure 6.9. Results of placed bed experiments: (a) one minute averaged velocities; (b) suspended sediment concentrations.
Figure 6.9 shows that during the first three velocity steps no erosion occurred. Then the erosion started approximately 5 minutes after the initiation of step four, at 13:07h ($U_c$ then was 0.8 m/s). An entire section of the sediment bed was eroded as a whole over the width of the flume and over a length of some tenths of a metre. Inside the scour hole vigorous circulation patterns developed and part of the eroded material was deposited downstream. At 13:40h a much larger part of the bed broke apart which showed, among other things, in strong reductions in flow velocities in the turbulent water layer. The EMF at 0.03 m became uncovered and velocities increased at this level.

![Graphs](Figure 6.10. Velocity fluctuations $u$ in the middle of the turbulent water layer and pressure fluctuations $p$ inside the sediment bed approximately 0.03 cm below the water-mud interface.)

Figures 6.10a and 6.10c show examples of the velocity fluctuations $u$ in the middle of the turbulent water layer for velocity steps 1 and 3, respectively. Figures 6.10b and 6.10d show examples of the pressure fluctuations $p$ at approximately 0.03 m below the water-mud interface for velocity steps 1 and 3, respectively. The long term variations in $p$ in Figure 6.10 b and d are due to a slight tilt of the flume which resulted in water level variations of $\pm 3$ mm. The short term fluctuations shown in Figure 6.10b cannot be distinguished from the instrumental noise of $\pm 3$ Pa. The short term fluctuations
shown in Figure 6.10d are above the noise level and their rms-value \( p' \) is estimated at 3 Pa. Hinze (1975) suggests that \( p' \approx 3 \tau_b \). This relation is more or less confirmed by the values for \( \tau_b \) shown in Table 6.3 for velocity steps 3 and 4, and shows that the pressure fluctuations at the interface penetrate to a level of 0.03 m below the water-mud interface.

In order to estimate the order of magnitude of pressure induced peak stresses inside the mud layer the following assumptions are made: the bed is at rest, transverse stress gradients are zero, \( \delta \) is much smaller than the length scale of the pressure variations, and the mud is considered impermeable so that a single phase approach suffices. The force balance for the mud layer then reads:

\[
\frac{\partial p}{\partial x} + \frac{\partial \tau}{\partial z} = 0
\]

(6.3)

where \( \tau \) is the shear stress in \( z \)-planes in the tangential direction. Integrating Eq. 6.3 over the mud layer gives:

\[
\tau(z) = (z - \delta) \frac{\partial p}{\partial x} \quad 0 \leq z \leq \delta
\]

(6.4)

If furthermore the "frozen-turbulence" approximation is applied (Eq. 4.30), the maximum value for \( \tau \) at the bottom of the flume, \( \tau_{\text{max}} \) can be approximated by:

\[
\tau_{\text{max}} = \frac{\delta}{U_c} \left. \frac{\partial p}{\partial t} \right|_{\text{max}}
\]

(6.5)

The maximum time derivative of the pressure \( \partial p/\partial t \) at \( \text{max} \) is approximately 50 Pa/s and 100 Pa/s for velocity steps 3 and 4, respectively. This gives values of \( \tau_{\text{max}} \) of about 5 and 8 Pa. These stresses are about 5 to 10 times larger than the average bed shear stress \( \tau_b \) (see Table 6.4). On the other hand they are smaller than the yield strength of the placed bed which was between 20 and 40 Pa (see Figure 6.6). Other effects might have contributed to the breakdown of the sediment bed, such as the intermittency of the shear stress at the mud-water interface, centrifugal forces, weakening of the bed by crack forming or swell. It is probably a combination of these effects that caused the break-up of the placed bed.

6.6 Discussion

The results of the entrainment experiments are in agreement with the findings of Winterwerp and Kranenburg (1997), in that the initial entrainment process is identical to the entrainment process in a two-fluid system. As such, the findings do not lead to
new insights into the entrainment process of layers of fluid mud by a turbulent upper layer. Certain aspects were examined in greater detail: (1) it was shown that at least the major part of the fluid mud remains stagnant with respect to the flume, velocities only increase in the transition zone between mud layer and turbulent water layer; (2) turbulence intensities are reduced in this transition zone compared to those in clear water experiments, which justifies the omission of turbulence production in the transition zone in a simplified entrainment model (Eq. 6.1); (3) very weak beds with very small yield strengths (up to 1 Pa) can resist mean shear stresses of about 0.1 Pa. This result may explain the scour lag shown in Figure 5.3.

The placed bed experiments showed that short term turbulent pressure fluctuations are transferred to deeper levels inside the sediment bed. The bulk erosion of the bed did not occur immediately after the increase in velocity of the lid, which suggests that other mechanisms, such as swell or crack forming, can of importance for the erosion process. The destruction of the stress history of the mud by mixing and pumping caused that the yield strength of the placed bed was relatively low compared to that of layers present in the field. In the field the yield stresses of cohesive sediment beds with similar densities are one to two orders of magnitudes higher than that of the placed bed tested in the flume (Mitchener et al., 1998).

Figure 6.11 show a footprint in a layer of mud in the field. The following relation can be derived from the force balance applied to a triangular section of the sidewall of the footprint (see sketch in Figure 6.11):

$$\frac{\delta}{\sin \psi} \tau_\psi \geq F_0 \sin \psi \approx \frac{1}{2} \frac{\delta}{\tan \psi} \rho_g g \sin \psi$$

(6.6)

which gives,

$$\tau_\psi \geq \frac{1}{2} \rho_g g \delta \sin 2\psi$$

(6.7)

where \(\tau_\psi\) is an average yield stress of the mud layer. The right hand side of Eq. 6.7 is maximal for \(\psi = \pi/4\). For this value of \(\psi\) and, in addition, a bulk density of 1250 kg/m³ and a thickness of the mud layer (as estimated from Figure 6.11) of 0.05 m, the averaged yield stress becomes at least 160 Pa. This value is significantly larger than the yield stresses presented in Figure 6.6b. It is therefore unlikely that such layers collapse directly as a result of turbulent pressure fluctuations. Collapse of these layers owing to wind waves is a more likely mechanism for erosion (Van Kessel, 1997). Bioturbation and grazing by Corophium Volutator, for example, can also contribute to the erosion of semi-permanent layers during periods preceding an algae-bloom (Kornman and De
Deckere, 1998). The assumption made in Section 5.4 of a limited availability of sediment for erosion for the June and August periods therefore seems justified, since storms did not occur and biological activity diminished.

Figure 6.11. Footprint in a layer of mud at the edge of the tidal flat Heringsplaat and diagram of forces.
Chapter 7 Conclusions and recommendations

7.1 Conclusions
The objective of the present study as defined in Section 1.4, was to increase the understanding of the role of turbulence in the transport of suspended sediment in tidal flow. In particular, the dampening of turbulence caused by sediment-induced density gradients was to be examined. For this purpose detailed measurements were made of flow velocities and SSC in the lower part of the water column in a well-mixed tidal channel in the Dollard tidal basin in 1996. The measurement technique used consisted of combinations of electromagnetic flow meters and turbidity meters, and proved successful. The water depth at the measuring location varied between 1.5 m and 4.5 m. Water level variations were dominated by the semidiurnal tide. Maximum absolute flow velocities of about 0.6 m/s were found during ebb tide. Sediment samples taken from positions directly adjacent to the measuring pole showed that the channel bed was composed of silt and clay. Variations in SSC and in vertical transports of sediment and momentum were analysed by means of the 1DV POINT MODEL. The erosive properties of mud collected from the tidal channel were studied in a laboratory flume. In this chapter the major findings of these studies and analyses are summarised.

Turbulence properties under unstratified flow conditions
The turbulence measurements under unstratified flow conditions showed that the turbulence characteristics in the tidal channel are comparable to the turbulence characteristics obtained in laboratory flumes at much smaller Reynolds numbers. The variation of turbulence intensities with relative depth and the ratio of longitudinal to vertical velocity intensities were in agreement with laboratory data (Nezu and Nakagawa, 1993). Similar conclusions were presented by West and Oduyemi (1989), and Kawanisi and Yokosi (1993). A friction factor $C_d$ of $1.8 \times 10^{-3}$ was found from velocity measurements. The equivalent bed roughness height derived from this factor is 2 mm, which is in agreement with commonly proposed roughness heights for flow over a cohesive sediment bed (Van Rijn, 1993). This is supported by visual observations during low water at the edge of the adjacent tidal flat which showed that bed forms were almost absent.
Concentration fluctuation intensities were coupled to the average suspended sediment concentrations and the vertical concentration gradients. Except during slack water periods when turbulence vanished, the vertical turbulent flux of suspended sediment was upward and the horizontal turbulent flux was against the direction of the main flow. The upward and horizontal turbulent fluxes had maximum absolute values for maximum turbulent concentration fluctuation intensities. The signs of the turbulent fluxes suggest that upward moving fluid parcels, which move slowly in the streamwise direction, contain relatively high concentrations, and downward moving fluid parcels which move fast in the streamwise direction, contain relatively low concentrations. Photographs of large upward moving flow structures, as marked by the sediment from the bed, were presented.

An important result of the spectral analysis is that the integral length scale $L_x$ is approximately $0.8 \sqrt{zh}$ which is in agreement with laboratory data of Nezu and Nakagawa (1993). This means that close to the bed the sizes of the turbulence structures remain relatively large and high frequency losses caused by sensor size were not as large as suggested by the work of Soulsby (1980). The vertical length scale ($L_z$) is estimated at $\frac{L}{4}$ of the value for $L_x$ which is significantly smaller than the $\frac{4}{3}L_x$ for isotropic turbulence. This is in agreement with the phenomenon of stretching of the large eddies near walls in the streamwise direction (Komori et al., 1987; Nezu and Nakagawa, 1993). The similarity between the spectra of instantaneous velocity fluctuations and concentration fluctuations suggests that the concentration fluctuations result mainly from the turbulent water motions.

**Effects of stratification on turbulence properties**

Flux Richardson and gradient Richardson numbers, $R_{if}$ and $R_{ig}$, were calculated for some parts of the measuring periods in June and August 1996. The results showed that effects of sediment-induced density stratification on the turbulence structure occurred at higher levels in the water column and for those parts of the tidal cycle where flow velocities were relatively small or where concentrations were relatively high. For the June measuring period such periods occurred during flood when the flow velocity decreased for approximately one hour and the sediment concentrations peaked. The reduction in integral length scale $L_x$ for these periods can be attributed to density stratification. The correlation coefficient $|R_{wu}|$ reduced from an average value of $0.36 \pm 0.04$ for $R_{if} < 0.05$ to approximately $0.25 \pm 0.05$ for $R_{if} > 0.15$. The scatter of the data was somewhat larger than that of laboratory data (e.g. Webster, 1964), and comparable to that of the field data of West and Oduyemi (1989). The scatter of the
data of $R_{cv}$ and $R_{cw}$ was too large to determine a trend in these correlation coefficients for increasing stratification.

The decreasing correlations between $u$ and $w$ fluctuations for increasing stratification suggest that for increased stratification levels, motions are generated which contribute fully to the velocity fluctuations but to a lesser extent to the transport of momentum. These motions might be internal waves, although these motions are difficult to discern in the velocity spectra (Uittenbogaard, 1995). Because of the strong stratification there is no theoretical basis for the observed $-5/3$ behaviour of the power spectra in the inertial sub-range (see Figure 4.18). Some evidence was found for an increase of the turbulent Prandtl number with increasing stratification. Similar results have been found by other researchers (Ueda et al., 1981; Komori and Ueda, 1983; Schumann and Gerz, 1995).

**Effects of stratification on sediment transport**

Computations with the 1DV POINT MODEL indicated that during periods around slack water the relatively rapid decrease in SSC and high deposition rates can be explained from buoyancy effects: a positive feedback mechanism developed in which decreasing turbulence production caused increasing concentration gradients and vice versa. This process is reflected in the increase in the measured and computed gradient Richardson numbers, and reductions in measured correlation coefficients and integral length scale $L_k$.

Sediment-induced stratification is directly related to the settling of particles or flocs. A settling velocity of about 0.5 mm/s was found from fitting the computed vertical sediment flux to the measured one. A settling velocity of 0.5 - 1.5 mm/s was obtained from fitting the Rouse concentration profile to the measured SSC. However, computations with the 1DV POINT MODEL indicated that the SSC gradients are largely affected by the unsteady flow conditions and by buoyancy effects. As a result the Rouse-profile method overestimates the settling velocities by a factor up to four. Comparisons of computational results to the measurements showed the limited applicability of commonly used expressions for the dependence of the settling velocity on SSC and turbulence shear parameter $G$.

Mean settling velocities of a few mm/s were observed in the upper part of the water column during maximum SSC in the same tidal channel (Van der Lee, 1998). This result does not necessarily contradict the smaller settling velocities in the near-bed region found in the present study. The formation of macro-flocs in the upper part of the water column and break-up of these macro-flocs in the lower part of the water
column by turbulence may explain the decrease in the mean settling velocity towards the bed.

Rapid reductions in SSC and increased settling rates caused by buoyancy affect the degree of hysteresis in the SSC-velocity curves. The degree of the Lagrangian hysteresis determines the net transport rate of sediment (see also Groen, 1967; Ridderinkhof, 1998). However, it is unlikely that the overall direction of sediment transport is changed by buoyancy effects.

**Additional results**

Erosive properties of mud collected in-situ were investigated in annular flume experiments in the laboratory. Mud layers were either formed from deposition or directly placed into the flume. The results of the experiments on beds formed from deposition were in agreement with the findings of Winterwerp and Kranenbrug (1997). It was found, among other things, that very weak beds with small yield strengths (up to 1 Pa) can resist mean shear stresses of about 0.1 Pa. This may explain the scour lag between the increase in flow velocity and increase in SSC observed in the field.

From the placed bed experiments it was inferred that it is unlikely that semi-permanent mud layers collapse directly as a result of turbulent pressure fluctuations. Collapse of these layers owing to wind waves is a more likely mechanism for erosion (Van Kessel, 1997). Bioturbation and grazing by Corophium Volutator, for example, can also contribute to the erosion of semi-permanent layers during periods preceding an algae-bloom (Kornman and De Deckere, 1998).

**7.2 Recommendations**

It is recommended that in future field research sediment-induced buoyancy effects on the turbulence are also investigate at higher levels in the water column. Assessment of vertical and horizontal mean velocity and mean SSC profiles is recommended if the detailed turbulence measurements are to be related to the overall properties of the tidal flow. Measurements of SSCs upstream of the measuring location, for example, may aid in discriminating between temporal variations in SSC caused by local sediment-bed exchange processes and those caused by advective transports.

From the field measurements some evidence was found for an increase of the turbulent Prandtl number $\sigma_z$ with increasing stratification, whereas it was set to 1.0 in the 1DV POINT MODEL. A Munk-Anderson (1948) type of relation, or a relation proposed by Schuman and Gerz (1995; Eq. 4.42), can be applied in order to model the turbulent Prandtl number (Galland et al., 1997; Winterwerp et al., 1998).
A flocculation model such as the one developed by Winterwerp (1998), in which the basic process of floc growth and break-up are described on the appropriate time scales, could possibly aid in connecting the observations made by Van der Lee (1998) and the findings presented herein.

In order to investigate the overall effect of buoyancy on the net sediment flux into the estuary, it should be incorporated in a 3D numerical model for the entire Ems/Dollard estuary.
References


Haines, J.W., Gelfenbaum, G., *Turbulent stresses in the surf-zone: which way is up?* Coastal Engrg. 1996, 3453-3466.


Appendix A  Summary of the physical properties of the Ems/Dollard estuary

Climate
- The water temperature is between 10 and 15 degrees Celsius (in summer the temperature gradient is positive in landward direction and in winter it is negative)
- The annular precipitation is 75 cm
- Winds from the west dominate, and are about 7 m/s near the barrier islands and 4 m/s in the Dollard reach on annual basis

<table>
<thead>
<tr>
<th>Topography</th>
<th>upper part</th>
<th>middle part</th>
<th>lower part</th>
</tr>
</thead>
<tbody>
<tr>
<td>(Dollard reach)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>intertidal areas</td>
<td>85%</td>
<td>35%</td>
<td>45%</td>
</tr>
<tr>
<td>total area (km²)</td>
<td>100</td>
<td>155</td>
<td>215</td>
</tr>
<tr>
<td>mean water volume (m³)</td>
<td>120 10⁶</td>
<td>550 10⁶</td>
<td>770 10⁶</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Tide</th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>prism (m³)</td>
<td>115 10⁶ (Dollard) / 75 10⁶ (Emd. F.W)</td>
<td>1000 10⁶</td>
<td></td>
</tr>
<tr>
<td>excursion (km)</td>
<td>12</td>
<td>-</td>
<td>17</td>
</tr>
<tr>
<td>range (m)</td>
<td>3.3</td>
<td>2.8</td>
<td>2.3</td>
</tr>
<tr>
<td>max.velocities (m/s)</td>
<td>1</td>
<td>1.3</td>
<td>1.5</td>
</tr>
<tr>
<td>diurnal inequality</td>
<td></td>
<td>can reach about 0.3 m</td>
<td></td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Sediment size distribution</th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>sand (d &gt; 55 μm)</td>
<td>50% (kg/kg)</td>
<td>67%</td>
<td>87%</td>
</tr>
<tr>
<td>silt (3 &lt; d &lt; 55 μm)</td>
<td>30%</td>
<td>28%</td>
<td>12%</td>
</tr>
<tr>
<td>clay² (d&lt; 3 μm)</td>
<td>20%¹</td>
<td>5%</td>
<td>1%</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Fresh water discharge</th>
<th>Summer</th>
<th>Winter</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ems</td>
<td>50 m³/s</td>
<td>200 m³/s</td>
</tr>
<tr>
<td>Westerwoldsche Aa</td>
<td>5 m³/s</td>
<td>15 m³/s</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Mixing characteristics for dissolved matter</th>
<th>High Discharge</th>
<th>Low discharge</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sea water mean age</td>
<td>14 days</td>
<td>36 days</td>
</tr>
<tr>
<td>Flushing time Ems</td>
<td>12 days</td>
<td>72 days</td>
</tr>
<tr>
<td>water</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dilution dissolved matter c¹ times</td>
<td>18 days</td>
<td>36 days</td>
</tr>
</tbody>
</table>

¹ Clay content in the Dollard reach differs from 5% in the centre to 50% and higher near shore
² The organic content is about 7% of the clay content
Source: De Jonge, 1992
Appendix B  Noise levels in velocity recordings

The noise levels in the velocity recordings of June and August 1996 were higher than expected. Laboratory tests showed that noise levels of $2 \times 10^{-5}$ m²/s² were to be reckoned with (see also Section 3.2), whereas the noise levels found for the June and August measuring period were up to $1 \times 10^{-3}$ and $1 \times 10^{-3}$ m²/s², respectively. Some characteristics of the noise contributing to the velocity signals, are shown in the power spectra Figure B1.

![Figure B.1](Image)

Figure B.1. Power spectra during three different stages of an ebb tide (21:10h - 23:30h June 27th 1996): (a) at 1.0 m, EMF D234; (b) at 0.4m, EMF D233; (c) at 0.1m, EMF D232; ---, $u$-power spectra; ------, $w$-power spectra.
Figure B.1 shows $u$ and $w$-power spectra obtained from EMFs D232-D234 during increasing ebb flow. The plots on the left show power spectra for relatively small velocities, the plots in the middle for intermediate velocities, and the plots on the right show power spectra for large velocities (maximum ebb). Noise appears in the power spectra as one, or some times two peaks followed by a "bump". This spectral shape is more or less similar at all three levels, and it is conserved for increasing velocities. In case of increasing velocities, the locations of the peaks shift towards smaller wave numbers; however, the associated frequencies remain constant and are approximately 4 Hz.

Table B.1. Absolute and relative contributions of the noise ($u^2_p$) to the variance of $u$ ($u^2$).

<table>
<thead>
<tr>
<th>Position/EMF</th>
<th>$u^2_p$ at 21:40h $\times 10^{-4}$, ( %)</th>
<th>$u^2_p$ at 22:05h $\times 10^{-4}$, ( %)</th>
<th>$u^2_p$ at 23:15h $\times 10^{-4}$, ( %)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.1 m /D232</td>
<td>0.01, (0%)</td>
<td>-0.36, (3%)</td>
<td>-1.62, (-5%)</td>
</tr>
<tr>
<td>0.4 m /D233</td>
<td>0.92, (45%)</td>
<td>1.02, (15%)</td>
<td>0.80, (3%)</td>
</tr>
<tr>
<td>1.0 m /D234</td>
<td>1.60, (73%)</td>
<td>1.98, (36%)</td>
<td>2.08, (12%)</td>
</tr>
</tbody>
</table>

The relative contributions of noise to the variances of $u$- and $w$-records show large variations and they depend, among other things, on the velocity of the flow and the EMF. For example: noise contributions are clearly present in the $u$-power spectrum for a relatively low flow velocity at 0.4 m, but are negligible for relatively high flow velocities at the same level (compare the left and the right plots in Figure B.1.b, see also Table B.1.).

Figure B.2. Co-spectra for maximum ebb and maximum flood, at 0.1 and 1.0 m.: ---, $uw$-spectra for ebb (23:15h June 27th); ---, $uw$-spectra for flood (16:15h June 26th).

150
The contributions of the noise to the total variance of the signals can be estimated from Figure B.1 in the following manner. The integrated spectral density is subtracted from a hypothetical integrated density which was found from extrapolating the spectra from 2 Hz, according to the $k^{-5/3}$ dependence. The results for the $u$-power spectra are presented in Table B.1.

In Table B.1 the positive numbers are contributions to the variance, the negative numbers represent reductions of the variance resulting from high-frequency losses being larger than the noise contributions. Table B.1 shows that the absolute noise contributions do not change much for increasing flow velocities.

The co-spectra for $u$ and $w$ for maximum ebb and for maximum flood, are shown in Figure B.2. A remarkable feature of the noise present in $u$ and $w$ is that it is correlated, and that it always gives a positive contribution to the Reynolds stress $\overline{uw}$ (denoted by the arrows). The absolute contributions are $6 \times 10^{-5}$ m$^2$/s$^2$ and $3 \times 10^{-5}$ m$^2$/s$^2$ for ebb and flood, respectively, and are of the same order of magnitude as the seemingly constant offset for $\overline{uw_{10}}$ of approximately $7 \times 10^{-5}$ m$^2$/s$^2$ in June as shown in Figure 4.10. For the August period the offset of $\overline{uw_{10}}$ is approximately $2 \times 10^{-4}$ m$^2$/s$^2$.

The correlation coefficients of the noise contributions can be determined from coherency spectra which are defined as $|S_y(k)/S_u(k)S_w(k)|$. The peaks and bumps of the $u$- and $w$-spectra show correlations between 0.4 and 0.8 for periods of slack water, the highest values are found for EMF D234 at 1.0 m. No coherence is found for the noise between the different levels.

A number of sources for the noise were examined. Sources which arise from wind, waves or tidal flow velocities, which showed large variations over the measuring periods, were excluded because the noise contributions were approximately constant over the measuring periods in June and August. The most likely source appeared to be increased instrumental noise due to wear and fouling of the EMF sensors heads. When the noise contributions could be considered small, they were assumed constant over the measuring period and were deducted from the turbulence intensities and Reynolds stresses. The velocity records of EMF 234 (at 1.0 m elevation) during the August periods were omitted herein, because the noise contributions were considered too high.
Appendix C  Description of the 1DV POINT MODEL

This appendix describes equations and boundary conditions that govern the (vertical) transport of sediment. These equations and boundary conditions together form the basis for the basis for a one-dimensional model (1DV POINT MODEL) for the transport of suspended sediment, in which homogeneous flow in horizontal directions are assumed. Some remarks are made herein about the numerical solution method being used; for a detailed description the reader is referred to Uittenbogaard et al. (1996). The 1DV POINT MODEL is still under development at w.l. | delft hydraulics.

Mean flow equations
The time dependent behaviour of the water/sediment column can be described by the conservation equations for mass and momentum. The Reynolds equations for momentum and sediment are presented herein, and the Boussinesq turbulence closure together with the $q$-$\varepsilon$ turbulence model (better known as the $k$-$\varepsilon$ model). The momentum equation and the $q$-equation are discussed in Sections 4.3 and 4.6, respectively, and they are therefore only repeated in this appendix.

If the transverse velocity $\vec{V}$ and the vertical velocity $\vec{W}$ are set to zero, and transverse gradients and advective transports are omitted, the balance equation for momentum reads:

$$\frac{\partial \bar{U}}{\partial t} = -g \frac{\partial \eta}{\partial x} - \frac{g}{\rho} (\bar{h} - z) \frac{\partial \bar{p}}{\partial x} - \frac{\partial \bar{u} \bar{w}}{\partial z}$$  \hspace{1cm} (C.1)

The balance equation for sediment reads:

$$\frac{\partial \bar{C}}{\partial t} = \frac{\partial \bar{C} \bar{w}}{\partial z} + \frac{\partial \bar{w}}{\partial z} \frac{\partial \bar{C}}{\partial z}$$  \hspace{1cm} (C.2)

where the first term on the right hand side represents the vertical derivative of the turbulent transport of sediment and the second term represents the derivative of the settling flux of the sediment ($w$, is taken positive in the downward direction). The turbulence transports in Eq. C.1 and Eq. C.2 are called the "Reynolds terms" for which additional expressions must be supplied (closure problem).

The Boussinesq turbulence closure is based on the assumption that the deviatoric turbulent stress can be treated in the same way as the deviatoric molecular stress for a Newtonian fluid, which is proportional to the velocity deformation tensor.
\[(\overline{u_i u_j} + \frac{1}{3} \overline{u_i^2} \delta_{ij}) = K_m (\frac{\partial \overline{U_j}}{\partial x_i} + \frac{\partial \overline{U_i}}{\partial x_j})\]  \quad i,j,k = 1,\ldots,3 \quad (C.3)

where $\delta_{ij}$ is the Kronecker-delta. The turbulent pressure $\frac{1}{3} \overline{u_i^2} \delta_{ij}$ can be neglected in most flows. In the same manner:

\[\overline{c u_i} = K_s \frac{\partial \overline{C}}{\partial x_i} = K_m \frac{\partial \overline{C}}{\sigma_i^T \partial x_i}\]  \quad (C.4)

where $K_s$ is the eddy diffusivity and $\sigma_i^T$ is the turbulent Prandtl number for sediment which is discussed in Sections 4.6 and 4.7. The eddy viscosity $K_m$ can be expressed as proportional to the product of a characteristic length scale and a characteristic velocity scale of turbulence, analogous to the molecular viscosity:

\[K_m \sim L \times U\]  \quad (C.5)

This assumption does not imply that the local stress depends on local variables, because the characteristic length scale $L$ is strongly non-local, it is determined by the order of the geometry of the flow. Eq. C.5 is appropriate if only one characteristic length scale exists (Tennekes and Lumley, 1972).

The $q-\varepsilon$ turbulence model

If a time scale $T$ is defined as $L/U$, and $q$ is scaled with $U^2$ and $\varepsilon$ with $U^2/T$, $K_m$ can be defined as:

\[K_m = c_\mu \frac{q^2}{\varepsilon}\]  \quad (C.6)

where $c_\mu$ is a multiplicative constant. Eq. C.6 is the basis of the $q-\varepsilon$ turbulence model. For the derivation of the balance equations for the turbulent kinetic energy $q$ and the dissipation rate $\varepsilon$ the reader is referred to e.g. Nieuwstadt (1992). The balance equation of $q$ reads:

\[\frac{\partial q}{\partial t} + \frac{\partial}{\partial x} \left( \frac{K_m \partial q}{\sigma_i^T \partial x} \right) - \frac{\partial \overline{U_j}}{\partial x_i} - \frac{q}{\rho} \frac{\partial \overline{U_j}}{\partial x_i} = \sigma_i^T \frac{\partial \overline{U_j}}{\partial x_i} - \gamma \frac{q}{\rho} \frac{\partial \overline{U_j}}{\partial x_i} - \varepsilon\]  \quad (C.7)

where $\gamma = ((\rho_s - \rho)/\rho_s)$, the first term on the right hand side of Eq. C.7 is a diffusion term of $q$ ($T_q$ in Eq. 4.36). The remaining terms represent shear production, buoyancy destruction and viscous dissipation. $\sigma_i^T$ is the turbulent Prandtl number for kinetic energy. The balance equation of $\varepsilon$ reads:
\[
\frac{\partial \varepsilon}{\partial t} = \frac{\partial}{\partial z} \left( \frac{K_m}{\sigma_T} \varepsilon \frac{\partial \varepsilon}{\partial z} \right) - c_{1e} \frac{\varepsilon}{\rho} \frac{\partial u}{\partial z} - c_{1e} (1 - c_{3e}) \frac{\varepsilon}{q} \frac{g}{\gamma} \frac{c_w - c_{2e}}{q} \frac{\varepsilon^2}{q} 
\] (C.8)

The production plus dissipation of \( \varepsilon \) in Eq. C.8 is assumed to be related to the production plus dissipation of \( q \) (see Eq. C.7) multiplied by \( \varepsilon / q \). This is a heuristic way of modelling and Eq. C.8 is therefore considered as the less rigorous part of the \( q-\varepsilon \) model (Booij, 1998, personal comm.). More sophisticated turbulence models are not considered herein, since uncertainties concerning modelling of, among other things, settling, erosion etc. and from the omission of advection, exceed those of turbulence modelling with the \( q-\varepsilon \) model. The constants in Eq. C.6 - C.8 have been found from calibration against laboratory data of grid generated turbulence and boundary layer flow (see Rodi, 1980; Uittenbogaard et al., 1996). Eq. C.1 - C.8 form a closed set of equations if the boundary conditions are given.

**Boundary conditions and modelling of erosion, settling and deposition**

The starting point for the boundary conditions are: (a) zero transports through the water surface of momentum, sediment, turbulent kinetic energy and dissipation (wind stress is not considered herein); (b) a constant stress layer with a logarithmic velocity distribution near the bed where source and sink terms are defined for the sediment concentration. Enhancement of bottom turbulence by waves, or suppression of turbulence by flow laminarisation by polymers, flow acceleration or density stratification, for example, could be expressed by adaptations of roughness length \( z_0 \). In this study \( z_0 \) is chosen constant. It can be calculated from the friction coefficient \( C_d \) (see Section 4.4) and Eq. 4.10 when a logarithmic velocity profile is assumed:

\[
z_0 / z_1 = \frac{k}{\sqrt{C_d}}
\] (C.9)

where \( z_1 \) is 1 m. The boundary conditions for Eq. C.1 are:

\[
\left. \frac{\partial \mathbf{u}}{\partial z} \right|_{z=z_0} = u_*^2
\] (C.10)

\[
\left. \frac{\partial \mathbf{u}}{\partial z} \right|_{z=h} = 0
\] (C.11)

with \( u_* \) calculated from the following equation:

\[
u_* = \frac{\kappa \mu(z_0)}{\ln(z_0 / z_0)}
\] (C.12)
where $z_a$ is an arbitrarily chosen height close to the sediment bed ($z_0 < z_a << h$). In the 1DV POINT MODEL $z_a$ is chosen in the centre of the lowest computational layer.

If in the logarithmic layer $\partial q/\partial t$, the buoyancy destruction, and transport of $q$ are assumed negligible, Eq. C.7 yields:

$$\frac{\partial q}{\partial t} = -u'w' \frac{\partial \tilde{u}}{\partial z} \bigg|_{z=z_a} = \frac{u_o^3}{\kappa z_a}$$  \hspace{1cm} (C.13)

Eq. C.13 together with Eq. C.3, C.6 and C.10 gives:

$$q|_{z=z_a} = \left( \frac{K_m(z_a)}{c_\mu} \tilde{c}(z_a) \right)^{0.5} = \left( \frac{\kappa u_o^3}{c_\mu} \frac{u_o^3}{\kappa z_a} \right)^{0.5} = \frac{u_o^2}{\sqrt{c_\mu}}$$  \hspace{1cm} (C.14)

Eq. C.13 - C.16 form the boundary conditions at $z = z_a$ for the balance equations for dissipation and turbulent kinetic energy, Eq. C.8 and C.7.

It is found from measurements that $q$ decreases near the surface because the turbulence production is much smaller than the turbulence dissipation, but $q$ does not tend to zero ($u/u_o$ even increases slightly for $y/h > 0.9$, Komori et al., 1987). It is the correlation coefficient for the Reynolds stress, $R_{oo}$, that decreases strongly near the water surface which explains the small stresses. In order to model the small stresses near the water surface with the $q-\varepsilon$ model, which is an isotropic model, the following boundary conditions are applied:

$$\frac{\partial \tilde{c}}{\partial z} \bigg|_{z=h} = 0$$  \hspace{1cm} (C.15)

$$q \bigg|_{z=h} = 0$$  \hspace{1cm} (C.16)

Eq. C.13 - C.16 form the boundary conditions at the free surface for the balance equations for dissipation and turbulent kinetic energy, Eq. C.8 and C.7.

The boundary conditions for the balance equation for sediment require a pragmatic approach: empirical expressions enter the sediment balance in order to describe the exchange of sediment between the bed and the water column. The sediment enters and leaves the water column according to Eq. C.17 and Eq. C.18, respectively. The combination of Eq. C.17 and C.18 is known as the Partheniades-Krone formulation,

$$E = M \left( \frac{\tau_b}{\tau_{c,e}} - 1 \right) \hspace{1cm} |\tau_b| > \tau_{c,e}$$  \hspace{1cm} (C.17)

$$D = w_s C_b \left( 1 - \frac{\tau_b}{\tau_{c,d}} \right) \hspace{1cm} |\tau_b| < \tau_{c,d} \leq \tau_{c,e}$$  \hspace{1cm} (C.18)
where $E$ is the bed erosion rate, $M$ is an erosion constant, $\tau_{c,e}$ is the critical shear stress for erosion, $D$ is the deposition rate, $C_b$ is the near-bottom concentration, and $\tau_{c,d}$ is the critical shear stress for deposition. The critical shear stress $\tau_{c,d}$ has been determined at several locations on the Heringsplaat by means of an in-situ erosion flume (Houwing and Van Rijn, 1995) in 1996: during the measurement periods in June and August the values for $\tau_{c,d}$ were between 0.05 and 0.15 Pa (Kornman, 1998). Two extreme conditions are considered in this study: the first condition uses the unrestricted form of Eq. C.17 which represents an unlimited availability of sediment. The second condition also uses Eq. C.17 but limits the total amount of sediment available for resuspension. This represents a situation where only a top sediment layer can be eroded. The parameter $\tau_{c,d}$ in Eq. C.18 has not been determined in-situ. It is assumed equal to $\tau_{c,e}$.

The definition of $C_b$ seems important for the total settling flux.

At the water surface the settling and diffusive fluxes are set to zero:

$$\left. w_s C \right|_{z=h} = 0 \quad \text{(C.19)}$$

$$\left. c w \right|_{z=h} = 0 \quad \text{(C.20)}$$

Eq. C.20 is automatically satisfied because of Eq. C.4, Eq. C.6 and Eq. C.16. Eq. C.19 implies that $\overline{C}$ at the surface will go to zero if $w_s \neq 0$.

During the field experiments in 1996 detailed information has been obtained about the settling velocities of sediment flocs in the upper part of water column of the tidal channel Groote Gat. Measurements from a floating vessel were conducted with an underwater video camera (Van der Lee, in prep.). The results showed a large scatter in floc sizes and settling velocities, though it can be concluded that during maximum SSC large flocs are present in the upper part of the water column with settling velocities up to a few millimetre per second, which are absent during periods of small SSC.

The following relation for the mean settling velocity seems to hold in the upper part of the water column ($z/h > 0.5$) (Van der Lee, in prep):

$$w_{s,0} = k_1 \overline{C}^m \quad \overline{C} < 1 \text{ g/l} \quad \text{(C.21)}$$

where $k_1$ and $m$ are empirical constants. Eq. C.21 shows that the settling velocity increases for increasing concentrations due to enhanced flocculation, or to particles with relatively large settling velocity which are brought in suspension during large flow velocities.

Close to the sediment bed effects of hindered settling may affect the sediment flocs and thereby the settling velocity. Winterwerp et al. (1998) showed that for simulating
sediment concentrations in the North Sea near the Rhine estuary, it is crucial to take into account hindered settling, only then the relatively high near bed concentrations could be reproduced. To account for effects of hindered settling Eq. C.21 can be replaced by the following relation (Richardson and Zaki (1954) in Mehta, 1986):

$$w_{s,0} = w_{s,\text{max}} (1 - k_2 \overline{C})^\beta \quad \overline{C} \geq 1 \text{ g/l} \quad (C.22)$$

where $\beta$ is a constant of approximately 5, $w_{s,\text{max}}$ is a reference settling velocity, and $k_2$ is a constant of approximately 0.0125 l/g for marine mud near the Rhine estuary according to Winterwerp et al. (1998) and approximately 0.008 l/g for mud from the Severn Estuary according to Thorn (1981, in Mehta, 1986). Laboratory settling experiments with Dollard mud showed settling velocities of $4.4 \times 10^{-5} \text{ m/s} \pm 10\%$. for $C = 60 \text{ g/l}$ (See Section 6.3). Combining the laboratory values and the in-situ measured velocities yields values of $k_2$ between 0.002 l/g and 0.009 l/g for Dollard mud.

Above the Heringsplaat floc size measurements were made close to the sediment bed at 0.3 m and at 0.6 m elevation by means of a laser diffraction technique (Van der Lee, 1998). Although the technique was not yet calibrated against other size measuring techniques, the qualitative influence of wave activity and flow velocity was clearly demonstrated: large flow velocities and large wave activity correlated with small flocs. Small floc sizes relate to small settling velocities and vice versa (Van Leussen, 1994; Van der Lee, in prep.).

To account for effects of turbulence on floc break-up and associated decrease in the settling velocities, Eq. C.21 and Eq. C.22 are adapted in the following manner (Van Leussen, 1994):

$$w_s = w_{s,0} \frac{1 + aG}{1 + bG^2} \quad (C.23)$$

where $G = \sqrt{\varepsilon / \nu}$ is a measure for the velocity gradients at the dissipation scale. Van Leussen (1994) found from settling experiments with mud from the Dollard: $a = 0.3 \text{ s}$ and $b = 0.09 \text{ s}^2$.

**Some remarks about the computational method**

The water level gradient in Eq. C.1 is adjusted in such a way that the depth integrated horizontal velocity matches a prescribed velocity. The prescribed velocities, which are time dependent, are derived from the results of field measurements (see also Figures 4.1 and 4.2).
An incomplete $\sigma$–coordinate transformation is applied. The vertical geometrical transformation reads:

$$\frac{\partial}{\partial z} = \frac{1}{h} \frac{\partial}{\partial \sigma}$$  \hspace{1cm} (C.24)

where $\sigma$ is the dimensionless vertical co-ordinate. To account for water level changes, a similar transformation should be applied to the term $\bar{W} \partial / \partial z$ of the material derivative of momentum, concentration and turbulence parameters. However, this term is omitted (see Eq. C.1, C.2, C.7 and C.8). If the term $\bar{W} \partial / \partial z$ is incorporated in Eq. C.2, for instance, a reduction in water level would lead to an increase in sediment concentration for reasons of conservation of mass. This is not in agreement with the actual situation in tidal flows where water level changes alone do not affect the sediment concentration because sediment is advected with the tidal flow. By putting $\bar{W}$ to zero, artificially, zero horizontal gradients are accounted for.

A first-order upwind scheme is used for the settling term in Eq. C.2. This scheme introduces a numerical diffusion of the order of $\nu_w \Delta z$, where $\Delta z$ is the computational layer thickness. The numerical diffusion should be small compared to the eddy diffusion (the ratio of the numerical diffusivity to the eddy diffusivity is often referred to as the Peclet number). This condition is not always met close to the sediment bed and near the water surface, and in regions of strong stratification or during slack water. The additional diffusion might reduce the vertical concentration gradients and thereby the effects of stratification on the turbulence. Dependence of the settling flux on the computational grid must be checked by varying the grid size.

**Parameters used in the computations**

<table>
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<tr>
<th>Run</th>
<th>$\Delta t$ (min)</th>
<th>$\Delta t_{rel}$ (min)</th>
<th>No. grid points</th>
<th>$w_r$ (m/s)</th>
<th>$M$ (kg/m²/s)</th>
<th>$C_{max}$ (kg/m³)</th>
<th>Buoyancy</th>
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<td>A1</td>
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<td>$\rho_w$ (kg/m³)</td>
<td>$\tau_{ae}$ (kg/m³)</td>
<td>$\tau_{ad}$ (kg/m³)</td>
<td>$z_0$ (m)</td>
<td>Depth (m - N.A.P.)</td>
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<td>-------------------</td>
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<tr>
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<tr>
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<tr>
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<td>0.1</td>
<td>$6.7 \times 10^{-5}$</td>
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1 See Table 5.2 for coefficients.
## Appendix D  Summary of the laboratory experiments

### Table D1  Erosion experiments, short consolidation times (< 10 h)

<table>
<thead>
<tr>
<th>ID</th>
<th>Settling/cons. period (min)</th>
<th>Mean conc. top of bed (g/l)</th>
<th>Initial δ (m)</th>
<th>Speed of lid (m/s)</th>
<th>Speed of flume (m/s)</th>
<th>Friction vel. lid u* (m/s)</th>
<th>Initial Ri* (-)</th>
<th>Initial w* (m/s)</th>
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</thead>
<tbody>
<tr>
<td>E1</td>
<td>235</td>
<td>142</td>
<td>0.13</td>
<td>-0.88</td>
<td>0.55</td>
<td>0.034</td>
<td>99</td>
<td>4.87E-04</td>
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<td>161</td>
<td>0.11</td>
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<td>0.038</td>
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<td>5.66E-04</td>
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<td>E3</td>
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<td>0.034</td>
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<td>7.62E-04</td>
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<td>0.046</td>
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<td>0.013</td>
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<tr>
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<sup>(1)</sup>Beds formed from a homogeneous suspension of about 100 g/l.

### Table D2  Erosion experiments, long consolidation times (> 10 h)

<table>
<thead>
<tr>
<th>ID</th>
<th>Setting/cons. period (hours)</th>
<th>Mean conc.bed (g/l)</th>
<th>Thick. Mud δ (m)</th>
<th>Speed of lid (m/s)</th>
<th>Speed of flume (m/s)</th>
<th>Bed shear stress τ* (Pa)</th>
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</thead>
<tbody>
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<td>215</td>
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<td>0.04</td>
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<td>0.00</td>
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### Table D3  Placed bed erosion experiments

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<th>Thick. Mud δ (m)</th>
<th>Speed of lid (m/s)</th>
<th>Speed of flume (m/s)</th>
<th>Bed shear stress τ* (Pa)</th>
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</thead>
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<td>-0.70</td>
<td>0.38</td>
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<td>-1.10</td>
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List of symbols

$A$  
cross sectional area at the mouth of the estuary

$a$  
parameter in Eq. C.23

$B$  
characteristic transverse length scale

$B_{\alpha}$  
constant in Eq. 4.20

$B_q$  
buoyancy destruction of turbulent kinetic energy $q$

$b$  
parameter in Eq. C.23

$C$  
sediment concentration

$C_h$  
parameter in Eq. 4.3 (= SSC at $z/h = 0.5$)

$C_{\alpha}, C_q$  
model coefficients in Eq. 6.2

$C_{max}$  
limit for the SSC in the 1DV POINT MODEL

$c$  
fluctuating part of $C$

$c_{\mu}$  
parameter in Eq. C.6

$c_{1e}, c_{2e}, c_{3e}$  
constants in Eq. C.8

$D$  
deposition rate of suspended sediment

$d$  
height of the annular flume

$d_{e}$  
effective roughness height

$E$  
erosion rate

$E_{\ast}$  
dimensionless entrainment velocity of a deposited bed (= $w_{\ast}/u_{\ast}$)

$f_G$  
gravity force acting on a segment of mud

$f$  
frequency

$f_c$  
Coriolis-parameter

$G$  
shear parameter ($= \sqrt{g/\nu}$)

$g$  
acceleration due to gravity

$H$  
water depth in the flume

$h$  
water depth in the estuary

$I$  
wet volume below a given water level

$K$  
parameter in Eq. 4.26

$K_e$  
eddy diffusivity for fine sediment particles

$K_m$  
eddy viscosity

$k$  
wave number

$k_1, k_2$  
parameters in Eq. C.21-C.22

$k_4$  
Kolmogorov wave number ($= 2\pi/\eta$)

$k_0$  
constant in Eq. 4.24

$L$  
characteristic longitudinal length scale

$L_A$  
integral longitudinal length scale ($= \pi S_{\alpha}(0)$)

$L_z$  
integral vertical length scale ($= \pi S_{\omega}(0)$)

$M$  
total no. of samples in a record, erosion constant in Eq. C.18

$m$  
parameter in Eq. C.21

$n$  
parameter in Eq. 4.1

$O$  
wet surface at a given water level

$P_q$  
production of turbulent kinetic energy $q$

$p$  
fluctuating part of the pore water pressure

$R$  
radius of the annular flume

$R_{\alpha}$  
Reynolds number ($= L_{\alpha}u_{\alpha}/v$)
$R_e$ Reynolds number ($= u_e h / v$)
$R_d$ correlation coefficient
$Ri_g$ gradient Richardson number
$Ri_f$ flux Richardson number
$Ri_{f,w}$ flux Richardson number for strong stratification
$Ri_s$ estuary Richardson number
$Ri_r$ overall Richardson number
$r$ streamwise lag distance
$S$ salinity
$S_{uf}$ spectral distribution
$T$ length of time of a record $u,w$ or $c$
$T$ time scale appendix $C$
$T_q$ transport of turbulent kinetic energy $q$
$t$ time
$Q_f$ fresh water discharge
$q$ turbulent kinetic energy
$U$ longitudinal velocity component
$U$ characteristic velocity turbulence scale
$U_c$ velocity in the core of the annular flume
$u$ fluctuating part of $U$
$u^*$ friction velocity
$u_r$ root mean square velocity in the mouth of the estuary
$V$ transverse velocity component
$W$ vertical velocity component, width of the flume
$w$ fluctuating part of $W$
$w_{x,max}$ parameter in Eq. C.22
$w_s$ particle settling velocity
$w_e$ entrainment velocity of a deposited bed
$X_{st}$ discrete values of $x_t$ at time $t$
$x$ longitudinal co-ordinate
$y$ transverse co-ordinate
$z$ elevation, vertical co-ordinate positive upward
$z_h$ correction, vertical co-ordinate positive upward
$z_0$ roughness height in Eq. 4.1
$z_a$ roughness height
centre of the lowest comp. layer in the 1DV POINT MODEL

Greek symbols
$\alpha$ Rouse-parameter in Eq. 4.3
$\alpha_1$ universal Kolmogorov constant (Eq. 4.24)
$\beta$ parameter in Eq. C.22
$\Delta t$ rest time of mud sample, computational time step
$\Delta t_{rel}$ relaxation time
$\Delta z$ distance between two grid points of the comp. grid
$\delta$ thickness of the mud layer
$\varepsilon$ dissipation rate
$\varepsilon_{b,r}$ bias error caused by rotation
$\gamma$ constant in Eq. 1.1 ($= (\rho_c - \rho_m) / \rho_c$)
\( \eta \)  
Kolmogorov length scale, surface level  

\( \varphi \)  
angle of the EMF sensor rotation  

\( \kappa \)  
Von Kármán constant  

\( \lambda \)  
Taylor microscale  

\( \nu \)  
kinematic viscosity  

\( \rho \)  
density  

\( \rho_m \)  
density of the turbulent water layer in the flume  

\( \sigma_T^\prime \)  
turbulent Prandtl number for fine sediments  

\( \sigma_{T,0}^\prime \)  
turbulent Prandtl number for fine sediments for neutral shear  

\( \sigma_T^\varepsilon \)  
turbulent Prandtl number for turbulent kinetic energy \( q \)  

\( \sigma_T^\varepsilon \)  
turbulent Prandtl number for dissipation \( \varepsilon \)  

\( \sigma^\prime \)  
effective stress  

\( \sigma \)  
dimensionless vertical co-ordinate  

\( \tau \)  
shear stress  

\( \tau_p \)  
peak strength of cohesive sediment  

\( \tau_y \)  
yield strength of cohesive sediment  

\( \omega_T \)  
angular velocity of top lid of the annular flume  

\( \omega \)  
angular velocity of the flume  

\( \xi \)  
any record of \( u, w \) or \( c \).  

\( \psi \)  
angle in Eq. 6.6  

\( \zeta_{p,1} \)  
finite Fourier transform of \( X_{ul} \)  

Subscripts  
\( m \)  
measured  

\( c \)  
critical  

\( d \)  
depth averaged, deposition  

\( e \)  
erosion  

\( b \)  
bed  

\( 0.1/0.4/1.0 \)  
aimed sensor heights (Table 4.1)  

\( s \)  
sediment  

\( w \)  
water  

\( ' \)  
prime, denotes a root mean square value

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Acknowledgements

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