Selective transport phenomena in coastal sands
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Chapter 1

Introduction and Basics

1.1 Preface

Understanding sediment transport processes is important for coastal defence, protection of coastal infrastructure like harbours and the maintenance of waterways. Morphological changes of sandy coastlines at a macroscopic scale are caused by the motion of grains at a microscopic scale. Conversely, the motion of grains at a microscopic scale is caused by parameters that are directly related to macroscopic morphological features, such as water depth. This feedback system makes understanding sediment transport a complicated matter.

Measurements in the field, necessary for study of sediment transport, are difficult to perform. Processes take place at spatial scales varying from microns to kilometers where the associated timescales increase accordingly [Oos95]. For the study of morphodynamics one has to monitor large areas for longer periods of time.

Radiometric properties of sand make radiometry a useful measuring tool for the study of sediment transport. In contrast with the time consuming process of sampling followed by tedious analysis in the laboratory, radiometric techniques are a tool for obtaining real time information on sediment characteristics. It is possible to monitor sediment transport without the necessity of using artificial tracers.

For the present work radiometric techniques have been applied to study selective transport phenomena that occur due to differences in size and density of minerals. The special radiometric features of so called heavy minerals (see below) makes radiometry especially suited for such a study because it allows for a (real time) monitoring of the movement of these minerals. The study of
Selective transport leads to a better understanding of transport processes in general.

The sandy sediment of the Dutch coasts consists mainly of quartz (SiO$_2$) and feldspar (e.g., (Ca,Na)AlSi$_3$O$_8$ plagioclase feldspar or AlSi$_3$O$_8$ orthoclase feldspar) and usually a small amount ($\leq 1\%$) of heavy minerals like zircon (ZrSiO$_4$), ilmenite (FeTiO$_3$) and rutile (TiO$_2$). The name of the latter class of minerals reflects the fact that their specific density is higher than the density of bromoform, a heavy liquid ($\rho = 2.8 - 2.9$ kg L$^{-1}$, where L = liter) used in geoscience for the separation of heavy and light minerals, and consequently also higher than that of quartz ($\rho = 2.6$ kg L$^{-1}$), the most abundant detrital mineral in sand. As will be explained later on in this chapter, heavy mineral grains will primarily be found in the smaller size fractions of sediment.

Due to chemical properties heavy minerals are more likely than light ones to incorporate traces of the radioactive elements $^{232}$Th and $^{238}$U within their lattice. The resulting relatively high specific radioactivity (partly emitted as $\gamma$ rays), observed by Bonka in 1980 on the Frisian Island of Norderney [Bon80], makes it possible to detect even very small quantities of heavy minerals by radiometric methods. The actual concentrations of radionuclides are affected by the circumstances at the time of crystallization and hence contain information about the provenance of the sediment.

Radioactivity is a process in which an unstable nucleus decays under emission of radiation. The product nucleus is either stable or radioactive itself; $^{40}$K decays in one step to a stable nucleus but the decay of $^{232}$Th and $^{238}$U proceeds through several radioactive daughter nuclei before stable lead isotopes are reached. This is illustrated in figure 1.1 where the decay schemes of $^{232}$Th and $^{238}$U are shown.

Based on their characteristics $\alpha$, $\beta$ and $\gamma$ radiation are distinguished. Both $\alpha$ and $\gamma$ radiation are emitted with clear distinctive energies typical for a nucleus. The electrons and positrons that are formed with $\beta$ decay have a continuous energy spectrum with only a well defined maximum which makes $\beta$ decay unsuited for the identification of a nucleus. As $\alpha$ particles have a limited range (in the order of micrometers in solids and centimeters in air) they are not suited for field measurements. Gamma rays, however, with an average energy of around 1 MeV are penetrating enough to be able to reach a detector above the ground from depths up till a few decimeters.

In table 1.1 $\gamma$-ray activity concentrations of three natural radionuclides, potassium ($^{40}$K), bismuth ($^{214}$Bi) (used as an indicator of the activity of the parent $^{238}$U) and thorium ($^{232}$Th) in various heavy mineral fractions are shown of a sample collected at ‘De Hors’ at the Dutch barrier island of Texel
Figure 1.1: Decay schemes of $^{232}$Th and $^{238}$U; horizontal and diagonal arrows denote α and β decay, respectively. The most important γ-ray emitters are marked. The half life of the nuclei is indicated in the boxes.

[Mei90]. The average activity concentration of the heavy fraction is more than two orders of magnitude larger than that of the light fraction. The light minerals, such as quartz and feldspar, contain only relatively low concentrations of thorium and bismuth but feldspar is relatively high in potassium. There is much variation in activity concentrations among different heavy mineral fractions; zircon, for example is high in bismuth and ilmenite is relatively high in bismuth as well as thorium. Because also the bismuth to thorium ratio varies, it is possible to identify minerals by measuring their γ-ray activity.

Heavy minerals are commonly found in trace amounts (1 - 2% by weight) in most beach and nearshore sediments [Don95]. At some locations, however, like Ameland and Texel concentrations up to 70% have been observed [Gre89, Mei92]. Such concentrations develop due to selection or sorting processes and have been studied worldwide. Heavy mineral placers found at the beach are generally thought to be the result of reworking processes at the beach where light minerals are winnowed and the heavy fraction stays behind [Eit95, Woo75]. This implies that high concentrations only occur due to erosive processes. However, 25,000 m$^3$ of heavy minerals were found at a location on the beach of Ameland in the winter of 1992, an amount that cannot possibly have originated from the eroded beach volume alone [Bos92, Mei92]. This would imply that heavy minerals must have been concentrated in the coastal
Table 1.1: Relative mass and activity concentrations of various heavy mineral fractions in a sample sand collected at De Hors, Texel. The fractions are: a) mainly magnetite, b) Almost pure ilmenite, c) Mainly ilmenite, some garnets, d) Mainly garnet, still some ilmenite, few epidote, e) Mixture of garnet, epidote, ilmenite and possible rutile, f) Mixture of epidote, staurolite, tourmaline and others, g) Rutile, some epidote, tourmaline and various others, h) Predominantly zircon, i) Average heavy minerals. Fraction j) consists of the light mineral quartz and some feldspar [Mei90].

<table>
<thead>
<tr>
<th>Fraction</th>
<th>Activity concentration (Bq kg⁻¹)</th>
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<td></td>
<td>(^{214}\text{Bi})</td>
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<td>i</td>
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<td>j</td>
<td>2.87 ± 0.10</td>
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</tbody>
</table>

zone and subsequently deposited on the beach or that heavy minerals are continuously deposited on the beach whereas light minerals are transported in the opposite direction.

A logical question is: why study heavy minerals when they normally occur only in trace amounts in coastal sands? Besides the fact that, as will be shown later, accumulations of heavy minerals are not as uncommon as one generally assumes, the study of these minerals is interesting for yet another reason: many transport models are based on experiments with light minerals. Heavy minerals should be considered as a test case for these models. Because models reflect the present state of knowledge, applicability to heavy minerals may show how accurate this knowledge is.

In this thesis the focus will be on three main questions:

1. How do heavy mineral accumulations develop,

2. What can be learned about sediment transport in general from the study
of heavy minerals and

3. What role may radiometric methods play in such studies?

To study the formation of heavy mineral placers at the beach and nearshore of the Netherlands, radiometric methods were utilized in two laboratory and one field experiment. The focus will lie on transport in a direction perpendicular to the coast (cross-shore) under wave-driven conditions. In the small-scale-wave-flume of the Laboratory of Fluid Mechanics of Delft University of Technology selection processes were studied under mild wave conditions within the ripple flow regime. A semi-quantitative model was subsequently developed to calculate grain trajectories (Chapter 2). Experiments with higher fluid velocities in the sheet-flow regime were performed in a wave tunnel at Delft Hydraulics. Here the heavy mineral distribution within the bed was monitored by in situ radioactivity measurements. An estimation of the transport rates for heavy and light minerals was the result (Chapter 3). Radiometric surveys during almost three and a half years at two test sites at the Frisian Island of Ameland gave insight into morphodynamics at a larger scale. An Empirical Eigenfunction Analysis indicated that deposition of heavy minerals occurs regularly in small but detectable quantities (Chapter 4).

Chapter 5 is a synthesis of the three experiments and their implications. It is shown that the study of heavy minerals and selective transport increases the understanding of sediment transport processes in general. Accumulations of these radiogenic minerals are not as exceptional as is commonly believed and their presence tells something about the history of the beach and the local wave energy conditions. Moreover, the possibility is considered of using heavy minerals for coastal protection purposes.

In this chapter the basic principles of sediment transport, natural radioactivity and its detection is discussed. It is meant as an introduction of frequently used parameters and terms relevant to this study. More comprehensive discussions on sediment transport and fluid mechanics can (for example) be found in van Rijn (1990) and (1993): ‘Principles of fluid flow and surface waves in rivers, estuaries, seas and oceans’ and ‘Principles of sediment transport in rivers, estuaries and coastal seas’, respectively [Rij90, Rij93], Fredsøe & Deigaard (1992): ‘Mechanics of coastal sediment transport’ [Fre92] and Batchelor (1967): ‘An introduction to fluid dynamics’ [Bat67].
1.2 Sediment characteristics

Bottom sediment consists of mineral grains that have settled from the transport medium, in this case water. Sedimentary particles are characterised by their specific density, $\rho_s$, the grain size and the settling or fall velocity. Density is a feature of the mineral; grain size and settling velocity reflect the transport and settling conditions where the latter should only be used for sediment deposited from suspension.

**Grain size** Sediment names like clay, silt, sand and gravel refer to the size of mineral grains. Sand is defined as sediment with a grain size between 64 and 2000 $\mu$m. Sand in natural deposits contains grains of a range of sizes. Commonly used values to characterise a sediment are the median particle size, $d_{50}$, and the ‘d-ninety’, $d_{90}$, of the sediment. They correspond to sizes which are exceeded by 50% and 10% (by weight) of the particles, respectively. Particle size of natural sediments normally is described with a log-normal distribution.

There are several methods to determine the grain size distribution of a sediment sample and reported values depend in part on the method chosen. In this study dried sediment samples were put through a stack of sieves with decreasing mesh size. The weight fractions collected on each sieve present a grain size distribution.

In this perspective one should realise that the shape of a distribution is influenced by the density of the grains. For samples containing grains of several minerals the non-uniform density of the minerals should be taken into account. As will be explained later, mineral grains with a high density are found primarily in the smaller fractions. Their relatively large contribution to the weight of these fractions decreases the median grain size of the sample. To correct for this effect one could transform the weight per fraction, $M^O_f$, to a corrected weight, $M^C_f$, according to

$$M^C_f = M^O_f \times \frac{\rho_q}{\rho_f},$$

where $\rho_f$ is the density of the sieve fractions and $\rho_q = 2.65$ kg $L^{-1}$ is the density of normal quartz sand. A corrected grain size distribution will in general result in a somewhat larger $d_{50}$ than the uncorrected distributions. For typical sand containing approximately 30% percent heavy minerals by weight the $d_{50}$ might increase from approximately 150 to 160 $\mu$m.

Another aspect that should be taken into account is that grains are not perfect spheres and the size obtained by sieving is not necessarily the same as
its equivalent spherical diameter. Wang and Komar (1985) suggest that it is better to use the intermediate diameter $D_b$, defined by a tangent rectangle, for evaluations of for example transport rates and settling velocities [Wan85]. They found that $D_b$ is related to the sieve diameter $D_{sv}$ by $D_b = 1.32D_{sv}$.

**Fall velocity** Another method to obtain the grain size distribution of a sample is by measuring the fall or settling velocity $w_s$. The *standard fall diameter* is defined as the diameter of a sphere with a specific density of $\rho_s = 2.65$ kg L$^{-1}$ and the same fall velocity as the particle in still, distilled water at 24° C. Because the fluid drag force is in equilibrium with the gravity force the fall velocity is related to the diameter of a sphere $d$ by:

$$ w_s = \left[ \frac{4(s-1)gd}{3C_D} \right]^{0.5}, $$

(1.2)

where the drag coefficient $C_D$ is a function of the particle *Reynolds number* $\text{Re}_d = w_s d / \nu$ (see below), $s = \rho_s / \rho$ and $\nu$ (m$^2$ s$^{-1}$) is the kinematic viscosity coefficient.

When a sample contains a considerable amount of grains with a higher density than 2.65 kg L$^{-1}$ this method results in a larger median grain size than would be obtained by sieving. For typical sand containing approximately 30% heavy minerals with a median grain size of approximately 150 $\mu$m according to the sieving method (without a correction for density) a value for the $d_{50}$ of 180 $\mu$m is obtained when calculated from the fall velocity.

The fall or settling velocity is an important parameter in sediment transport models as will be discussed later in this section. For small grains (with a diameter $< 100$ $\mu$m) in the Stokes region ($\text{Re}_d < 1$) the drag coefficient is $C_D = 24 / \text{Re}_d$. When the Reynolds number gets larger the drag coefficient increases rapidly and the simple expression disappears. For larger grains the relation between grain size and settling velocity is somewhat more complicated because shape effects play a role. For non-spherical grains the following empirical relations between grain size and fall velocity can be used [Rij93]:

$$ w_s = \begin{cases} \frac{(s-1)gd^2}{18\nu} & \text{for } 1 < d < 100\mu m \\ \frac{10w}{d} \left(1 + \frac{0.01(s-1)gd^4}{\nu^2} \right)^{0.5} - 1 & \text{for } 100 < d < 1000\mu m \end{cases} $$

(1.3)

1This relation is valid for quartz and most heavy mineral grains but not for mica due to its platy shape

2Vertical velocities are denoted by $w$ and horizontal by $u$
Hydraulic equivalence When grains have the same settling velocity they are called hydraulically equivalent in this work. If settling is the sole selection mechanism, like in a settling tube, equation 1.3 implies that minerals with a higher density will be found together with larger-sized fractions of the lighter minerals. The relation between grain size and settling velocity of particles as a function of their respective densities can be obtained from relation (1.3). Although it is true that grains with a higher density are found in the smaller size fractions there are, however, other mechanisms besides settling that are important for the sediment composition. Depositions occur that are not hydraulically equivalent [Sli84]. Selection processes that may cause this will be discussed later in this section.

1.3 Fluid mechanics

Because it forms the bases of sediment transport in this section, some basic fluid mechanics will be discussed.

1.3.1 Basic equations

Bernoulli’s equation Consider a fluid element in a stream with velocity $\vec{v}$, density $\rho$ and a viscosity $\eta$. The equation of motion for such an element is

$$\rho \frac{d\vec{v}}{dt} = -\nabla p + \eta \nabla^2 \vec{v} + \rho (\vec{g} \cdot \vec{e}),$$  \hspace{1cm} (1.4)

where $p$ is pressure, $g$ the acceleration of gravity and $\vec{e}$ the unit vector. Equation (1.4) is known as the Navier-Stokes equation and is the most complete equation for incompressible ($\rho = constant$) viscous flow.

In the case where viscous forces are negligible, Bernoulli’s equation applies. It represents the law of conservation of energy for an ideal fluid:

$$U_{\text{pot}} + U_{\text{kin}} = \text{constant.}$$  \hspace{1cm} (1.5)

Because $U_{\text{kin}} = \frac{1}{2} \|\vec{v}\|^2$ and $U_{\text{pot}} = p + \rho g z$ this gives:

$$\left( \frac{\|\vec{v}\|^2}{2} + \frac{p}{\rho} + g z \right) = \text{constant,}$$  \hspace{1cm} (1.6)

where $\vec{v}$ it the fluid velocity and $\rho$ the density of the fluid. Equation 1.6 also follows from the Navier-Stokes equation under the conditions of steady ($\frac{\partial \vec{v}}{\partial t} = 0$), irrotational ($\nabla \times \vec{v} = 0$) and inviscid ($\eta = 0$) flow.
Turbulence  The flow in coastal waters is usually turbulent. Reynolds found, by applying dimensional analysis, that occurrence of turbulence depends on a dimensionless quantity, now known as the Reynolds number (Re).

The Reynolds number is a measure for the ratio between forces of inertia and viscous forces; a flow in a tube with diameter $D$, for example, with a velocity $V$ has a Reynolds number: $Re = \frac{DV}{\nu}$ where $\nu$ is the viscosity of the fluid. If this ratio exceeds a certain value, depending on the flow conditions, the flow becomes turbulent.

Turbulent flow problems can in principle be described with known equations in which the residual momentum fluxes, caused by turbulence, represented by so-called Reynolds-stress terms, have to be taken into account. This means that the normal and shear stresses consist of a viscous and a turbulent part. The problem of turbulent flow modelling is to relate the Reynolds stresses to the mean velocity components. A variety of approaches to solve this closure problem is described in the literature e.g. Launder & Spalding (1972) [Lau72].

Boundary layer  The Navier-Stokes equations can only be solved for certain simple cases. In general, the equations are hard to solve and one has to use special methods to obtain solutions. For two-dimensional irrotational flows of ideal fluids ($\rho$ is constant and $\eta=0$) the differential equations may be somewhat simplified by introducing a velocity potential $\Phi$:

$$u = \frac{\partial \Phi}{\partial x}, \quad w = \frac{\partial \Phi}{\partial z}. \quad (1.8)$$

Substitution into the equation of continuity leads to a Laplace equation.

However, the velocity potential can not be applied in the neighbourhood of solid surfaces because there the flow can not be regarded as irrotational any longer due to viscosity effects. To overcome this problem the boundary layer concept was introduced, in which it was stated that a turbulent flow with a very high Reynolds number can be divided in two parts with different equations of motion to be solved:

I The layer close to the surface where viscosity has to be taken into account, called boundary layer.

#Footnotes#

3If one defines a typical length $L$ one can write this ratio as follows:

$$\frac{F_{\text{inertia}}}{F_{\text{viscous}}} = \frac{2}{\eta A \frac{\partial V}{\partial z}} = \frac{V^2 L^2}{\nu V L} = \frac{LV}{\nu}, \quad (1.7)$$

with $V = L/T$, $A = L^2$ being the surface and $\eta = \rho \nu$. The exact value of $L$ depends on the problem.
II The part outside the boundary layer where the flow can be regarded in first approximation as inviscid flow.  

With this approach the flow problem can be solved over the majority of the flow domain with the relatively easy potential flow approximation. Inside the boundary layer, which is a very important part for sediment transport due to its large velocity gradients, other solution methods have to be used. For certain situations solutions can be found in the literature. However, in most cases there are no clear-cut solutions available.

1.3.2 Waves

Surface gravity-driven waves may occur at a large variety of time scales; they vary from tidal waves with a period of half a day or day to wind waves with periods in the order of seconds. The wave height, that is the distance between the crest and the trough, ranges from centimeters up to several meters. Short surface waves, in which the fluid pressure is not hydrostatic anymore, (contrary to long surface waves), cause an oscillatory fluid motion. In shallow water, this motion extends down to the bottom and is important for selective transport of sediment.

**Small-amplitude linear-wave theory** Under the conditions of irrotational flow, the basic equations for unsteady flow in terms of the potential \( \Phi \) are the Laplace equation:

\[
\frac{\partial^2 \Phi}{\partial x^2} + \frac{\partial^2 \Phi}{\partial z^2} = 0
\]  

(1.9)

and the time-dependent Bernoulli equation

\[
\frac{\partial \Phi}{\partial t} + \frac{1}{2} \left( \frac{\partial \Phi}{\partial x} \right)^2 + \frac{1}{2} \left( \frac{\partial \Phi}{\partial z} \right)^2 + \frac{p}{\rho} + gz = 0.
\]  

(1.10)

Under the assumption that the amplitude of the surface elevation is small compared to the water depth (small-amplitude waves), and compared to the wave length (low-steepness waves), the non-linear terms may be neglected and equation (1.6) transforms into the linear Bernoulli equation.

The boundary conditions necessary to solve these equations are obtained from the situation at \( z = -h \), where the vertical velocity component \( w = \frac{\partial \Phi}{\partial z} = 0 \), and at \( z = 0 \) where \( w = \frac{\partial \Phi}{\partial z} = \frac{a}{\partial t} \) and the pressure \( p = 0 \). However, because \( a \) is one of the unknown variables a Taylor-series expansions at \( z = 0 \) is needed. Solving the equations of Laplace and Bernoulli with the boundary

\(^4\)In free surface flow, the boundary 'layer' often extends over the entire water column.

\(^5\)For a more elaborate discussion see [Rij90] and references therein.
1.3 Fluid mechanics

conditions, assuming a horizontal bottom and periodicity in space and time, the following expression for $\Phi$ is obtained:

$$
\Phi = -\frac{\omega}{k} \frac{\cosh[k(h+z)]}{\sinh[kh]} \sin(\omega t - kx),
$$

(1.11)

with:

- $z$ = vertical coordinate with $z=0$ at the mean of the water surface
- $\hat{a}$ = amplitude of the (water)surface elevation (m)
- $\omega$ = angular frequency ($\frac{2\pi}{T}$) (s$^{-1}$)
- $k$ = wave number ($\frac{2\pi}{L}$) (m$^{-1}$)
- $L$ = wave length (m)
- $T$ = wave period (s)
- $h$ = water depth for $a = 0$
- $a$ = surface elevation: $a = \hat{a} \cos(\omega t - kx)$
- $\frac{\omega}{k}$ = wave propagation velocity (m s$^{-1}$)

The fluid velocities in the $x$ and $z$ direction are the derivatives of $\Phi$:

$$
u = -\frac{\omega H}{2} \frac{\cosh[k(z+h)]}{\sinh[kh]} \cos(\omega t - kx),
$$

(1.12)

$$
w = -\frac{\omega H}{2} \frac{\sinh[k(z+h)]}{\sinh[kh]} \sin(\omega t - kx),
$$

(1.13)

with $H$ being the wave height $= 2\hat{a}$.

Near the bottom $z \approx -h$ the fluid motion will be mainly horizontal because the vertical component has to go to zero at the bed ($w \approx 0$), in phase with the water elevation. As been explained in section 1.3.1 the potential flow solution is only valid outside the boundary layer. In the case of waves without a significant net current this layer will remain very thin because the flow keeps on changing direction and the layer will not have time to develop. Therefore the velocity at the edge of this boundary layer $u_b$ is approximated by the velocity at $z \approx -h$ with a peak value of:

$$
\hat{u}_b = \frac{\omega H}{2\sinh(kh)}.
$$

(1.14)

This equation should, in principle, not be valid in shallow water, because the wave amplitude is not small compared to the water depth anymore. Measured velocities in shallow water, however, show reasonable agreement with computed values.

Inside the wave-boundary layer the velocity distribution can be solved analytically in the case of laminar flow. For turbulent flow several methods may be used to approximate a solution. The most practical way is to assume a distribution that meets the boundary conditions.
Non-linear small-amplitude wave theory  Taking higher-order terms of the non-steady Bernoulli equation into account, the potential function $\Phi$ can be represented by a power series. The water elevation can be approximated by a Fourier series, which leads to an asymmetric wave profile as shown in figure 1.2. The crests become somewhat narrower and higher and the troughs longer and shallower.

![Figure 1.2](image)

**Figure 1.2:** An example of a higher-order wave profile with $A_1$ being the amplitude of the first-order wave and $A_2$ of the second.

1.4 Sediment transport

1.4.1 Bottom shear stress

Important for sediment transport is the friction exerted by the fluid on the bottom, the bed or bottom shear stress $\tau_b$ (N m$^{-2}$). In case of steady ($\frac{\partial \tau}{\partial t} \approx 0$) uniform ($\frac{\partial \tau}{\partial x} \approx 0$) flow the following relation between bottom friction and flow velocity an empirical relation is used:

$$\tau_b = \rho f_c \bar{u}^2,$$

(1.15)

where $\bar{u}$ is the depth-averaged velocity and $f_c$ the empirical Darcy-Weisbach coefficient based on data for turbulent flow. This coefficient is related to the bottom roughness expressed in terms of the roughness length $k_n$ (the Nikuradse roughness length). There are various expressions in the literature to calculate $k_n$ depending on the nature of the bed morphology. When bed forms, like ripples, are present, $k_n$ is directly related to the height and length of such structures. However, in a flat-bed situation the roughness is fully determined by the sediment type. Used values for $k_n$ in such cases vary from the $d_{50}$ to $3d_{50}$ in case of a moving bed.
For waves an equation for the bed shear stress and its peak value $\tau_b$ is used that is similar to the one for uniform flow:

$$\tau_b = \frac{1}{2} \rho_f w_b |u_b| \text{ hence } \tilde{\tau}_b = \frac{1}{2} \rho f_w u_b^2,$$

where the ‘wave-friction factor’, $f_w$, is introduced [Jon66]. This factor depends on the relative bottom roughness, expressed as the ratio between the Nikuradse roughness length $k_n$ and the semi-excursion length of the water particles $A_b = \hat{u}_b T/\pi$, where $T$ is the wave period. Based on data of Jonsson (1966) the following expression was found:

$$f_w = \exp \left[ -6 + 5.2 (A_b/k_n)^{-0.19} \right],$$

with $f_w \leq 0.3$. To give an idea of the numerical values: a wave with a period $T$ of 6 seconds and a peak velocity $u_b$ in the order of 1 m s$^{-1}$ gives a semi-excursion length of approximately 1 meter. For sheet flow over a sediment bed containing sand with a $d_{50} \approx 200 \mu$m the wave friction factor is in the order of 0.01; when ripples are present with a length of 10 cm and a height of 2 cm, the friction factor $f_w \approx 0.05$.

**Shields parameter** A minimum shear stress $\tau_{b,\text{crit}}$ and therefore a threshold fluid velocity is needed to set a grain into motion. The magnitude of this velocity can be calculated using the *Shields parameter* $\Theta$

$$\Theta = \frac{\tau_b}{(\rho_s - \rho)g d_{50}} = \frac{0.5 f_w u_b(t) |u_b(t)|}{(\rho_s - \rho)g d_{50}}.$$

This parameter expresses the ratio of the bed shear stress $\tau_b$ and the (stabilizing) force of gravity. When the bed shear stress is high enough, $\Theta$ becomes larger than the *critical Shields parameter* $\Theta_c$ and grains will start to move. The magnitude of this critical Shields parameter was determined empirically by Shields as a function of the Reynolds number. This function can be represented as a function of a dimensionless particle diameter that incorporates the density and diameter of the grain.

**Hiding and sheltering** An alternative way to using the critical Shields parameter for calculating the critical value of the shear stress is given by Komar, Li and Wang described in several articles [Kom84, Kom86, Li86, Kom88]. By considering the horizontal directed forces working on a spherical grain resting on a bed of grains with radius $R$ one may write:

$$\tau_{\text{crit}} = (\rho_s - \rho_m) g r \tan(\alpha),$$

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where $\alpha$ is the pivoting angle, $r$ the radius of the grain and $g$ the gravitational constant. A larger angle leads to a larger critical shear stress; consequently a larger fluid stress is needed to initiate motion. From measurements of Miller and Byrne (1966) the following empirical relationship was found:

$$\alpha = c(r/R)^f,$$

(1.20)

in which $c$ and $f$ are empirical coefficients and $R$ the average sediment diameter [Mil66]. Komar and Wang (1984) used the following expressions, valid for grains with a diameter smaller than 1 mm (as in the present study SI-units are used, the numerical values of $c$ and $f$ will differ from those in the references)[Kom84]:

$$\tau_{\text{crit}} = 0.00071(\rho - \rho_m)gr^{0.57}\tan(\alpha)$$

(1.21)

$$\alpha = 61.5(r/R)^{-0.3}.$$  

(1.22)

Equation (1.21) provides a critical shear stress that is both a function of size and density.

More general one can say that in case of non-uniform sediment the (critical) shear stress for individual grains may depend on the surrounding sediment. This is called ‘sheltering’ or ‘hiding’ because grains are ‘sheltered’ by each other from the fluid. Several other authors give correction factors, $\xi_i = \Theta_{c,\text{corrected}}$, to account for ‘hiding’ for either the critical Shields parameter (shear stress) or the effective Shields parameter. An overview is given by Ribberink 1987 and for a more elaborate discussion see references therein [Rib87]. Equations (1.23), (1.24) and (1.25) give expressions developed by several authors.

Based on a theoretical approach Eziagaroff (1957) suggested a correction for the critical bed shear stress:

$$\xi_i = \left(\frac{10\log 19}{10\log 19d_i/d_m}\right)^2$$

(1.23)

where $d_i$ is the grain size of a fraction and $d_m$ is the average grain size of the total sediment $d_m = \sum_i p_id_i$. For $d_i = d_m$ the correction factor is equal to unity. Misri et al. (1984) gives an exposure correction based on theoretical grounds:

$$\xi_i = (d_i/d_m)^{-1}$$

for $\frac{d_i}{d_m} < 1.0$,  

$$\xi_i = \left(\frac{10\log 1.5}{10\log(0.5d_i/d_m) - 0.57}\right)^2$$

(1.24)

for $\frac{d_i}{d_m} > 1.0$. 

1.4 Sediment transport

Day (1980) proposes an exposure correction based on a large number of experimental data

\[ \xi_i = \left( \frac{0.4}{(d_i/d_a)^{0.5}} + 0.6 \right)^2, \]  

(1.25)

where \( d_a \) is representative for the gradation of the mixture:

\[ \frac{d_a}{d_{50}} = 1.6 \left( \frac{d_{84}}{d_{16}} \right)^{-0.38} \]  

(1.26)

All three expressions give a correction smaller than unity for the coarser grains implicating that these grains experience a larger force than in case of uniform sediment. However, some authors suggest a hiding factor that is equal to unity for the coarser fractions [Rib87]. Hiding plays a role in selective transport because small grains require a higher shear stress to be set in motion when they are hided between larger grains.

**Intergranular forces** A grain at the bottom also experiences ‘impact forces’ from saltating grains hitting the bed. In wind this effect is responsible for bringing grains continuously in saltation. In water, however, according to Bagnold (1941) the descending grains are not capable of ejecting other grains by their impact and initiation of motion is caused by the fluid alone [Bag41].

1.4.2 Sediment transport models

In the near coastal region longshore (shore-parallel) and cross-shore (shore-normal) sediment transport may be distinguished; both complicated processes that are not easy to describe. Longshore transport is governed by tidal and/or wave induced currents that transport sediment, stirred up by waves and currents. Cross-shore transport is complex, due to the continuous variation of the flow velocity, as well in magnitude as in direction. Several mechanism play a role: e.g. orbital wave motion, wave asymmetry and generation of long waves. In shallow coastal waters also the spatial variation of the water depth is important (shoaling, refraction). The present study is mainly focussed on sediment transport mechanisms due to waves and is therefore mostly affiliated with cross-shore sediment transport.

We may divide sediment transport into two modes: (I) bed-load and (II) suspended-load transport. In the first mode the grains remain more or less in contact with the bottom or stay in the near vicinity of it. They move by sliding, rolling and ‘jumping’ (saltation). In the suspension transport mode grains have no contact with the bottom for a longer period of time (several
wave periods). However, in practice it is not always possible to distinguish between the two transport modes.

In the low (orbital) velocity regime, with vortex ripples, relatively a lot of sediment is brought into suspension. In the high velocity regime (\( \Theta \gtrsim 0.8 - 1.0 \)) a large amount of sediment is moving in a region very close to the bottom (a few centimeters) and relatively little, higher up in the flow. Under these sheet flow conditions, sediment moves like a sheet over the bottom, a division between bed-load and suspension mode becomes a question of definition.

There are a wide variety of sediment transport models available, based on theoretical as well as empirical considerations and applicable to different conditions. Comprehensive overviews are given by Horikawa (1988) [Hor88] and van Rijn (1993) [Rij93]. As examples three models valid under wave conditions in the sheet flow regime will be discussed.

- **Bailard’s model (1981)** [Bai81]. This model relates transport rates to the work done by the bottom shear stress, \( \tau_0 u \), in which \( u \) is the horizontal fluid velocity. A distinction is made between bed load and suspended load transport. Bailard uses two efficiency factors \( e_b \) and \( e_s \), for bed load transport \( q_b \) and suspended load transport \( q_s \), respectively. These factors account for the fraction of the energy spent to the transport process. Both are obtained by calibration. The intra-wave transport rate in grain volume per unit length and time (m\(^3\) s\(^{-1}\)) at time \( t \) is given by:

\[
q_b(t) = \frac{\tau_0 u}{(\rho_c - \rho)g \tan \beta} = 0.5 f_w e_b \frac{u^3(t)}{\Delta g \beta} \tag{1.27}
\]

\[
q_s(t) = \frac{0.5 f_w e_s}{\Delta g w_u} |u^3(t)|u(t), \tag{1.28}
\]

where \( \Delta = (\rho_c - \rho)/\rho \), \( \beta \) is the internal angle of repose (for sandy sediments \( \beta = 0.6 \)) and \( f_w \) is the wave-friction coefficient (see eq.(1.16)).

- **Ribberink’s model (1993)** [Rib93, Rib96]. In this (bed-load) model the Shields parameter \( \Theta \) (see page 21) is the parameter determining sediment motion. One of the goals of this model was to create a formula that was valid under waves as well as under steady flow conditions. Therefore it uses a type of formula that was empirically obtained by Meyer-Peter and Müller (1948) [Mey48] for steady flows but also showed good agreement with observed transport rates under wave conditions. The basic
1.4 Sediment transport

formula looks like:

\[ \Phi_b(t) = m \left\{ \left| \Theta(t) \right| - \Theta_c \right\}^n \frac{\Theta(t)}{|\Theta(t)|}, \]  

(1.29)

where \( \Phi_b \) is a dimensionless transport parameter \( \Phi_b = \frac{b}{(\Delta g d_0)^{0.3}} \) and \( \Theta_c \) the critical Shields parameter. In case of waves where the Shields parameter is time dependent, \( \Theta = \Theta(t) \), a representative Shields parameter \( \Theta_{repr} \) is defined to be able to make a comparison with steady current.

\[ \left\{ \left| \Theta_{repr} \right| - \Theta_c \right\}^n \frac{\Theta_{repr}}{|\Theta_{repr}|} = \langle \left\{ \left| \Theta(t) \right| - \Theta_c \right\}^n \frac{\Theta(t)}{|\Theta(t)|} \rangle, \]  

(1.30)

where \( \langle . \rangle \) is the half or full wave-cycle time averaging. The coefficients \( m \) and \( n \) were determined by fitting field as well as lab data and are taken to be 9.1 and 1.78, respectively.

- Dibajnia and Watanabe’s model (1992) [Dib92]. This model takes into account the delayed behaviour of sediment in suspension; sediment stirred up under the crest of the wave and which has not settled yet will be transported in the opposite direction under the trough of the wave and vice versa. By considering the fall velocity of the sand and thickness of the sheetflow layer \( \delta_n \) under crest and trough of the wave this ‘time lag’ is incorporated in the model.

The thickness of the sheetflow layer \( \delta_n \) is calculated by an energy concept:

\[ \delta_n = \frac{E_k}{(\rho_n - \rho)Vg} = \left( \frac{0.5 \rho V u_i^2}{(\rho_n - \rho) Vg} \right) \frac{u_i^2 \rho}{2(\rho_n - \rho)g}, \]  

(1.31)

where \( E_k \) is the kinetic energy of a sand particle with volume \( V \) and density \( \rho_n \) moving in its ambient fluid with density \( \rho \). The parameter \( u_i \) can either be \( u_c \) or \( u_t \), and is defined as the equivalent sinusoidal velocity amplitude under crest or trough, respectively

\[ u_i^2 = \frac{2}{T_i} \int_0^{T_i} u^2 dt, \]  

(1.32)

where \( T_i = T_c \) or \( T_t \) is the duration of the positive or negative part of the velocity profile, respectively. The time which a particle needs to settle \( T_{f,i} \) is then equal to the thickness of the layer divided by the fall velocity \( w_n \):

\[ T_{f,i} = \frac{u_i^2}{2 \Delta g w_n}, \]  

(1.33)
wherein $$\Delta = \frac{(\rho_s - \rho)}{\rho}$$. If the ratio $$\omega_i$$ between settling time $$T_{f,i}$$ and the period of the positive (negative) part of the wave velocity profile is larger than unity, sand is still in suspension when the flow changes direction and will subsequently be transported in the opposite direction under the other part of the wave cycle.

$$\omega_i = \frac{T_{f,i}}{T_i} = \frac{u_i^2}{2\Delta g w_n T_i}$$ \hspace{1cm} (1.34)

To estimate the amount of sediment in suspension, the dimensionless parameters $$\Omega_i$$ and $$\Omega'_i$$ are introduced:

If $$\omega_i \leq 1$$ then
$$\Omega_i = \frac{2w_i}{d_{50}}$$ and $$\Omega'_i = 0$$.
If $$\omega_i > 1$$ then
$$\Omega_i = \frac{2w_i}{d_{50}}$$ and $$\Omega'_i = 2(\omega_i - 1)\frac{u_i T_i}{d_{50}}$$ \hspace{1cm} (1.35)

The net non-dimensional transport rate represented by $$\Gamma$$ is given by:

$$\Gamma = \frac{u_c T_c (\Omega_c^3 + \Omega_c'^3) - u_t T_t (\Omega_t^3 + \Omega_t'^3)}{u_c + u_t}$$ \hspace{1cm} (1.36)

where the subscripts $$c$$ and $$t$$ refer to crest and trough, respectively, and $$T = T_c + T_t$$. The parameter $$\Gamma$$ is empirically related to measured non-dimensional transport rates $$\Phi_{bw} = \frac{(\rho)}{u_n d_{50}}$$ by:

$$\Phi_{bw} = 0.001|\Gamma|^{0.55} \frac{\Gamma}{|\Gamma|}$$ \hspace{1cm} (1.37)

The model is capable of predicting a net transport opposite to the direction of the peak velocity in case of sheet flow. According to the authors, it gives a reasonable estimate of the transport rate under a wide range of sheet flow and ripple conditions.

Both Bailard's and Ribberink's model are quasi-steady: they directly connect the sediment motion to the instantaneous fluid velocity. The model of Dibajnia & Watanabe is quasi-unsteady: it incorporates non-linear effects in the coupling between the motion of the fluid and the sediment.

### 1.4.3 Selective transport

Selective transport of sediment occurs when grains react differently to the driving forces wind and water. Hiding, for example, results in a selection based on grain size (see above). Singerland (1984) [Sli84] distinguished four different types of sorting processes:
1.4 Sediment transport

1. entrainment sorting: sorting by differences in the threshold velocity to initiate motion
2. shear sorting: vertical sorting resulting in a layered structure by interactions between grains
3. suspension sorting: separation of grains in suspension by different settling velocities
4. transport sorting: separation of grains due to differences in transport velocity, entrainment and suspension sorting.

In the models presented in section 1.4.2 selection is incorporated since they comprise the sediment characteristics density and the median grainsize. For an inhomogeneous sediment consisting of several sediment types, characterised by \( \rho_{s,i} \) and \( d_{50,i} \), one can calculate the transport rate \( q_i \) for each sediment type separately. Interaction between grains of different sediments can be incorporated by applying hiding factors as discussed in section 1.4.1. Taking into account the fraction \( p_i \) of each sediment type in the total sediment, the fraction within the transported sand \( p_{i,T} \) according to Bailard’s and Ribberink’s model is than given by (without hiding effects)

\[
p_{i,T} = \frac{p_i q_i}{\sum p_i q_i} \tag{1.38}
\]

For the respective models this means

- Bailard’s model (eq. 1.27): In case of bed-load transport

\[
p_{i,b} = \left(1 + \sum_{j=2}^{n} \frac{\Delta i p_j}{\Delta j p_i}\right)^{-1} \tag{1.39}
\]

and for suspended load transport

\[
p_{i,s} = \left(1 + \sum_{j=2}^{n} \frac{\Delta i p_j w_{n,j}}{\Delta j p_i w_{n,i}}\right)^{-1} \tag{1.40}
\]

where \( w_n \) is the fall velocity.

- Ribberink’s model (eq. 1.29): If it is assumed that velocities are high enough such that \( \Theta \ll \Theta_c \) and \( n \approx 1.5 \) then \( p_{i,T} \) is the same as for Bailard’s model in case of bed load transport and given by equation 1.39.
This means that for these two models ‘selectivity’ is independent of the fluid velocity. In the model of D & W the heavy fraction within the sand in motion, does depend on the velocity because it is one of the factors that determines whether or not sand is transported opposite to the direction of the peak velocity.

1.4.4 Heavy mineral placers

Accumulations of heavy minerals (placer) at beaches are mostly found in areas suffering from erosion and have generally been thought to develop due to the winnowing of light material from beach sediments. Extensive studies about the subject have been published by several authors. Frihy et al. (1995) [Fri95] studied spatial variations of the heavy minerals contents along the Nile Delta in Egypt. Areas of erosion were characterised by higher concentrations of heavy minerals like garnet and zircon. The less dense and coarser grains were found in zones of beach accretion where they were deposited after being selectively removed from areas of beach erosion. Woolsey et al. (1975) [Woo75] concluded that backshore heavy mineral concentrations on Sapelo Island (Georgia) were fully the result of dune erosion. The concentration process was most effective during moderate surf energy levels attending storm wane.

Eitner (1995) [Eit95] found entrainment sorting to be very important for development of heavy mineral concentrations at the beach. Heavy minerals in beach samples were hydraulically equivalent with the light minerals but the critical bed shear stress varied. He concluded that because the heavy and light fraction had the same settling velocity, concentration can only be the result of selective removal of the light fraction from the beach. Consequently accumulations of heavy minerals are indicative for beach erosion.

Cordes (1996) [Cor66] studied the origin of heavy mineral concentrations at a beach near Skagen in Denmark. He concluded the main source of heavy minerals to be dune sand, enriched by selective aeolian transport. Accumulations developed by reworking of the beach sediments.

However, Stapor (1973) [Sta73], who studied heavy mineral placers at the coast of Florida, found strong indications for offshore concentration of heavy minerals. Wave tank experiments performed by May (1973) [May73] support this. Slingerland (1984) [Sli84] described the formation of deposition areas at bar crests containing a relatively large heavy mineral fraction with a settling velocity that is greater than that of the light fraction; large light grains are more likely to be entrained than small heavy ones, because they protrude further into the flow. Moreover, hiding effects may play a role. This points to the possibility that heavy minerals are concentrated offshore.
to be subsequently transported to the beach. Also Bonka (1980) [Bon80], who observed the high \( \gamma \)-dose rate coming from heavy minerals at the beach of Norderney, suggests a combination of concentration in water and on land.

The experiments performed by May (1973) in a wave tank were performed to study the formation of accumulations of heavy minerals in the form of discrete lamina. Accumulations did occur with several combinations of water depth and wave height where the best results, with respect to selectivity, were obtained under moderate conditions (peak velocities at the wave boundary layer in the order of 0.2-0.3 m s\(^{-1}\)). It was observed that under these conditions heavy minerals were transported onshore whereas light grains were moved in an offshore direction. The author explained this by defining three classes of sediment, based upon their response to the wave condition:

1. grains that are not moved,
2. grains that are only moved under the crest of the wave, i.e. unidirectional motion,
3. grains that move under the crest as well as the trough of the wave, i.e. bi-directional motion.

If the heavy grains fall into class two and the light ones in class three, the selection is most effective [May73]. This indicates that the sediment is segregated offshore and that the heavy minerals may, under favourable conditions, be transported shorewards to be deposited there. This shows that heavy and light mineral grains are transported in different modes: the small heavy grains move close to the bottom as bedload whereas the larger lighter grains move in suspension.

About the origin of heavy mineral placers the following can be concluded:

- Heavy mineral placers are generally found at beaches that suffer from erosion.
- Entrainment sorting is an important factor for the development of heavy mineral concentrations in water as well as at the beach.
- Heavy minerals may be transported shorewards while light minerals may move in the opposite direction.
1.5 Gamma radiation and its detection

1.5.1 Natural radioactivity

The crust of our earth contains several naturally radioactive elements. Some of them, like $^{40}$K, $^{232}$Th, $^{235}$U and $^{238}$U, are remnants of long lived isotopes and have been there since the origin of this planet. They have a half life that is of the same order of magnitude as the age of the earth.

The differences with which light and heavy minerals incorporate natural radionuclides makes it possible to use radiometric techniques for the study of transport processes. The practical applications are not restricted to detection of accumulations of heavy minerals; the sensitivity of such methods allows for discrimination between sediments based upon their radiometric characteristics.

In figure 1.1 one can see that the half life of the nuclei in the chains are considerably shorter than that of the parent nuclei $^{238}$U and $^{232}$Th. If the system is isolated, meaning that no nuclei can disappear otherwise than through the process of nuclear decay, the state of so-called secular equilibrium is reached. In secular equilibrium the activity concentrations of all the nuclei in the chain becomes equal. Thus, provided the system is closed, measuring the activity concentration of one member of the chain provides information about the presence of all members.

---

**Activities are expressed in becquerel (Bq), where one becquerel corresponds to one decaying atom per second. The activity $A$ of a radioactive nucleus, given in Becquerel (Bq) is given by**

$$A = \frac{N \ln 2}{T_1},$$

(1.41)

where $N$ is the number of atoms and $T_1$ is the half-life time of the radionuclide. In this thesis activity concentrations are determined, which is the activity per unit weight of material (sand), expressed in Bequerel per kilo (Bq kg$^{-1}$). One can calculate, using Avogadro’s number, that 1 ppm = 1 ppb of 1 ppm is equal to an activity concentration of 12.3 and 4.0 Bq kg$^{-1}$, respectively. One percent $K_2O$ contains 257 Bq kg$^{-1}$ $^{40}$K.

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*This should not be confused with the specific activity.*
1.5 Gamma radiation and its detection

The decay series of uranium includes a radium isotope with a long half-life time that is soluble in water. The next product in this series is radon, which is a gas and has the possibility to escape. The principal $\gamma$-ray emitters in the uranium decay series are progeny of these nuclides. Because it is not sure that the system is isolated, the measured activities might not be representative for the actual uranium concentration. To check to what extent mineral grains can be considered as an isolated system, uranium concentrations determined with X-ray fluorescence were compared with measured bismuth concentrations. Within the uncertainties of the measurements there were no indications that the systems were not closed [Mei94]. For quartz, however, it is known that up to 20-30\% of radon formed in the grains escapes [Reg91, Spo96]. Therefore, the activity concentration in the decay series of $^{238}\text{U}$ will be reported as the first nuclide in the chain that emits $\gamma$ rays, being bismuth-214.

From the two main ingredients of the sand found at the Dutch beaches, the light minerals quartz and feldspar, feldspar is a mineral that may contain components that are rich in potassium ($^{40}\text{K}$). This nucleus and several nuclei present in the decay chain of $^{238}\text{U}$ and $^{232}\text{Th}$ which can be incorporated in the crystal lattice of certain heavy minerals, are $\gamma$-ray emitters; their activity concentration can be determined by $\gamma$ ray detection.

1.5.2 Detection of gamma rays

For the detection of $\gamma$ radiation we use the interaction of photons with matter. Three processes play a role: (1) the photo-electric effect, (2) Compton scattering and (3) pair production. For the energy range of interest mainly the first two processes are of importance. In the photo-electric effect the total energy of the photon is transferred to the detector material whereas in compton scattering only part of its energy is transferred. The third process, pair production, is less important for the detection of gamma radiation of natural origin because the average energy of the emitted $\gamma$ rays is approximately 1.0 MeV, which is just over the threshold energy for pair production (1.0 MeV). The three $\gamma$-ray detectors used in this study will be discussed:

- a semi-conducting hyper-pure germanium (HPGe) detector for high accuracy measurements in the laboratory,

- a sodium iodide (NaI(Tl))- and

- a BGO (beryllium germanium oxide) scintillation detector.
Germanium detector  In the laboratory samples were measured on a semi-conducting HPGe detector. This detector is basically a diode, consisting of a hyper-pure germanium crystal between a p type and a n type contact formed on opposite surfaces of the crystal. Incoming photons create electron hole pairs within the depletion layer. By applying a reverse voltage over the diode, these charge carriers cause a change in voltage. Because the number of electron-hole pairs depends linearly on the energy of the incident photon, the magnitude of the voltage change is a measure of the absorbed energy. Semi-conductor detectors have a very high energy resolution because the average energy needed to create an electron hole pair is small (3.0 eV). They are therefore well suited for γ-ray spectroscopy. The semi-conducting detectors have to be cooled to very low temperatures before they can be operated and are not very practical for field measurements; their use is mainly restricted to the laboratory.

![Graph](image)

**Figure 1.3:** γ spectrum of sand from Ameland measured with a HPGe detector during 674 seconds. Note the narrow high photo peaks and a broad continuum of Compton energies.

In the setup at the KVI laboratory the crystal (180 cm$^3$) is placed inside a container with walls of 10 cm thick, made of ‘old lead’ to reduce the background radiation. Samples are dried and sieved to remove coarse parts (> 2 mm), and are placed on the crystal in a sample holder matching the
1.5 Gamma radiation and its detection

sample volume. Signals from the detector are pre-amplified (ORTEC GEM-45200S) and then processed by a Personal Computer Analyzing (PCA) card (IBM PCA from ‘the Nucleus’). This card in combination with a personal computer acts like a multi-channel analyzer in which the pulse height of a signal is converted in a channel number by a analogue to digital converter (ADC) and the contents of the channel is incremented by one. In this way a pulse-height spectrum is generated that corresponds to the energy spectrum of the photons. In figure 1.3 a γ spectrum is shown of a sand sample from Ameland (containing a relatively large concentration of heavy minerals) measured with a HPGe detector during 674 seconds. Note the narrow high photo peaks and the broad continuum of compton energies.

**Portable NaI detector** For onshore field measurements (Chapter 4) a portable γ-ray detector (SCINTREX GIS-5) was used. This scintillation detector contains a $3.7 \times 3.7 \times 5.0 \text{ cm}^3$ thallium activated sodium iodide (NaI(Tl)) crystal, specially ‘ruggedised’ for field measurements.

A scintillation detector is based on the principal that some materials convert an absorbed gamma ray into a light pulse with an intensity proportional to the energy of the incident gamma ray. This light pulse is transformed into an electrical pulse by a photo-multiplier tube (PMT). These pulses are subsequently processed electronically. The scintillation detector is the combination of scintillation crystal and PMT. The energy resolution of scintillation detectors depends on the scintillation material but is in general considerably worse than of semi-conductor detectors. Because they do not need to be cooled and they are very suitable for field measurements.

Pulses from the SCINTREX detector system may sorted according to amplitude by using one of the four windows:

1. Total Counts (TC), all γ energies above 0.05 MeV
2. Bi + Th + K, all γ energies above 1.38 MeV
3. Bi + Th, all γ energies above 1.66 MeV
4. Th, all γ energies above 2.38 MeV

The possible measuring periods are 1, 3, 10, 30 and 100 s; all recordings on the island were ‘total-counts’ measurements with an integration time of 10 seconds. It is assumed that the square root of the measured number is a good estimate of the statistical uncertainty.

The ‘action radius’ of the detector is in the order of meters if held at hip height. Because the penetration length of γ rays in air is large, it is the
solid angle, penetration length in sand plus the activity concentrations of the sediment that are the factors that determine how large the ‘action radius’ is.

**BGO detector** For the measurements within the Large Scale Wave Tunnel (see Chapter 3) the ERG-detector system MEDUSA (Multi-element detector system for underwater sediment activity) was used, which contains a Bismuth Germanium Oxide (BGO) scintillation crystal (5 cm Ø, 15 cm long). This crystal has a high efficiency for $\gamma$ radiation due to the high specific density of the material and the high Z value of bismuth. It is not hygroscopic and is resistive to shock, properties that make it very suited for a detector that has to be towed over the bottom of the sea. Due its high efficiency it allows for accurate measurements of the activity concentrations in a short time.

A disadvantage of using a BGO crystal is that the light output is relatively low resulting in an energy resolution which is worse than a NaI crystal of similar size. Photo peaks in the spectrum are broader, but the peak to Compton ratio is considerably better than for HPGe or NaI. To make optimal use of the high efficiency of this detector the special method of *spectrum deconvolution* is used for analysis as will be described in the next section.

In figure 1.4 a schematic view of the detector system, as used in the tunnel, is given. On the cylindrical crystal a photo multiplier tube (PMT) is mounted together with a voltage divider (for the PMT). They are placed inside a watertight sealed aluminium cylinder. A second cylinder connected to the first by a coaxial cable, contains the high voltage power supply (HVS), an amplifier, an analogue to digital converter (ADC) and additional electronics to transmit the signals through a coaxial cable to a PC for storage and processing.

To shield the crystal from radiation coming from its surroundings (cosmic rays and radiation coming from building materials) the tube with the crystal is placed inside a rectangular piece of lead ($25 \times 17 \times 11$ cm$^3$) with a slit in the bottom side such that the detector faces the sediment bed. The two tubes together with the lead shielding were placed in a specially carriage designed to fit into the tunnel. The carriage could be placed on a rail present in the tunnel and was positioned manually at the desired locations in the tunnel at a chosen distance from the bottom of the tunnel.

### 1.5.3 $\gamma$ ray analysis

Two types of analysis may be distinguished: (1) spectroscopy based on the content of photo peaks in case of $\gamma$ detection with a high energy resolution as with the HPGe detector and (2) spectrum deconvolution in case of detection with a low energy resolution as with the BGO detector.
Spectroscopy The pulse height spectrum from the HPGe detector is analysed using the commercial analysis programme GammaTrac [Gam93]. The contents of each photo peak is determined after subtraction of the background. From this the activity concentration of the corresponding radionuclide is derived, taking into account the mass, the *branching ratio*, which gives the statistical chance that this transition takes place, and the photo-peak efficiency of the detector (the probability of a full energy response of a photon to be detected). Assuming secular equilibrium, the activity concentration of the parent nucleus $^{232}\text{Th}$ (or $^{214}\text{Bi}$) is considered to be the weighted mean of the concentrations of several isotopes from the respective decay series.

In a few cases two $\gamma$-ray energies are not separated within the resolution of the detector. Usually this is not a problem because in the decay series of uranium and thorium several $\gamma$ rays are emitted which may be used to determine the activity concentration. For $^{40}\text{K}$ a problem may occur because its only $\gamma$ ray at 1460.8 keV practically coincides with the 1459.2 keV $\gamma$ ray emitted by one of the nuclei in the decay series of thorium. For samples where K and Th have the same activity concentration the K line is about ten times stronger than the Th line and corrections based on the strength of other Th lines is readily made. In certain cases, however, with a thorium activity concentration that is several orders of magnitude larger, it becomes virtually impossible to determine the $^{40}\text{K}$ content accurately.

To determine concentrations a good calibration of the system is essential. The efficiency of the detector is defined as the number of photons in the photo peak divided by the total number that is emitted by the source. Besides the volume of the crystal and the photon energy it depends on the size and density of the sample and the sample-detector geometry. The last aspect depends on...
the type of sample holder that is used.

The system is calibrated for different types of sample holders with an aqueous solution of man-made radionuclides emitting $\gamma$ rays with energies in the interval from 59 - 1836 keV. The efficiencies for the $\gamma$ energies radiated by the calibration solution have been determined for several sample volumes.

Based on the efficiency $e_w$ of water a density dependent correction can be made for the self absorption in the sample; the efficiency $e_s(E_i)$ for an energy $E_i$ of a sample of sand with a density $\rho_s$ is given by

$$e_s(E_i) = e_w(E_i) \times \frac{S_s}{S_w},$$

where $S$ is called the self-absorption factor. Because this factor is a measure for the fraction of the emitted photons that is not absorbed the name is somewhat misleading. It may be calculated by

$$S_j = \frac{1 - e^{-\mu_j \rho_j d}}{\mu_j \rho_j d},$$

where $d$ is the height to which the sample holder is filled. The attenuation or absorption coefficient $\mu_j$ (cm$^2$ g$^{-1}$) is a function of the Z number of the absorber $j$, i.e. the type of material, and the photon energy. Multiplication with the specific density of the absorber gives the linear attenuation coefficient $\mu$ (cm$^{-1}$), which is a quantity that will be used in the rest of this thesis.

The uncertainties in the measured radionuclide concentrations originate from different sources:

- counting statistics,
- uncertainties in the efficiencies and
- uncertainties arising from inhomogeneities in the samples.

An extra systematic error is caused by the summing effect, when two (or more) photons are absorbed by the detector within the dead time of the detector they may be detected as one instead of two photons. In case of $^{214}$Bi and $^{232}$Th the total uncertainty is estimated to be in the order of 5% - 10%. This is confirmed by comparisons with chemical measurements of the heavy mineral contents of sand samples [Mei90].

The level of accuracy for $^{40}$K depends strongly on the heavy minerals concentration in the sample; high concentrations lead to a decrease in accuracy. This may partly be avoided by improving the statistical accuracy by increasing the measuring time.
**Spectrum deconvolution** In figure 1.5 a typical $\gamma$ spectrum as recorded with the BGO detector is shown. On basis of the contents of the photo peaks alone no accurate activity concentrations can be obtained. It is also evident that the integrated contents of the full spectrum is large compared to that of the photo peaks.

![Gamma spectrum with peaks labeled K, Bi, and Th](image)

**Figure 1.5:** A typical $\gamma$ spectrum (measured of the coast of Terschelling in November 1994) as recorded with the BGO detector with a measuring time of 5 minutes. The location of the prominent photo peaks of $^{40}$K, $^{232}$Th and $^{238}$U are indicated.

The method of spectrum deconvolution assumes that the number of possible sources of natural $\gamma$ radiation from minerals are limited to three: $^{40}$K, $^{232}$Th and $^{238}$U. This means that every natural $\gamma$ spectrum can considered to be a superposition of standard spectra of each of these three sources separately plus a constant contribution from the background. Hence, the count rate $R(i)$ in counts per second as recorded in channel $i$ of a spectrum from an arbitrary sample containing only natural radionuclides can be written as

$$R(i) = C_K * K(i) + C_{Bi} * Bi(i) + C_{Th} * Th(i) + B(i),$$  \hspace{1cm} (1.44)
with $K(i)$, $Bi(i)$ or $Th(i)$ are the count rates in channel $i$ from the standard spectra and $B(i)$ the contribution of the background; the unknown concentrations of the radionuclides are denoted with $C_K$, $C_Bi$ and $C_Th$ and $i = 1, \ldots, N$ is the channel number where $N$ is the total number of channels in the spectrum. The values of $C_K$, $C_Bi$ and $C_Th$ are then determined by a least squares method:

\[
\sum_i \frac{[C_K \cdot K(i) + C_Bi \cdot Bi(i) + C_Th \cdot Th(i) + B(i) - R(i)]^2}{\sigma^2_{S(i)}} = \text{minimal},
\]

where $\sigma^2_{R(i)} = \sqrt{R(i)}$ is the uncertainty in the count rate $R(i)$ [Pat95].

To obtain standard spectra, calibration pads from the British Geological Survey (BGS) were used [Sta93]. These are concrete blocks with a volume of $1 \times 1 \times 0.3$ m$^3$ with a well known concentration of the three radionuclides $^{40}K$, $^{232}Th$ and $^{238}U$ ($^{214}Bi$). They are an approximation of the infinite-plane geometry which occurs at the sea bottom. The spectra were measured with the detector underwater and a geometric correction factor was applied for the limited size of the pads.

With this method activity concentrations can be determined with an accuracy of approximately 10% with a measuring time of 10 seconds. This short measuring time makes it possible to tow the detector over the sea bottom while measuring the activity concentration$^6$. In the experimental setup of the large scale wave tunnel, where the BGO detector was used, measuring periods of 30 sec were taken to increase the accuracy.

$^6$If a towing speed of $\approx 8$ km h$^{-1}$ is assumed this means that the spatial resolution is $\approx 25$ m.