

Sediment Dynamics

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Chapter 4 A User's Guide to Hydraulic Modelling and Experimentation,
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4.1. Introduction

This chapter provides descriptions of essential features of modelling of a broad range of problems involving many classes of sediment, flow, applications and research topics. The experimental reproduction of hydraulic processes involving sediment dynamics is focussed on the sediments and not the hydrodynamic processes, which are covered in other chapters. Since a great deal of experimental research on sediment dynamics remains to be done, laboratory experiments to elucidate process knowledge (research) as well as physical modelling of site-specific applications (consultancy) will be covered.

The following items are not discussed within this chapter:

- Soil mechanics and geotechnics, which constitute a major subject area in their own right
- Fluidisation/liquefaction of the bed by upward flows or wave action
- Permeable flows.
- Neither numerical models nor field measurements are discussed.

A thorough and readable account of physical modelling in coastal engineering, including a substantial chapter on sediment transport models, is given in the book by Hughes (1993). He offered the opinion: "Understanding sediment transport in coastal regions is a perplexing challenge that in all likelihood will continue to frustrate coastal researchers and engineers for generations to come". This opinion could be applied to non-coastal situations such as rivers with equal force, and is as true today as it was in 1993. The present guideline is not intended to duplicate the detailed (and still generally applicable) accounts given in this and other books. Instead the aim is to summarise key results, and add knowledge derived in recent years, particularly that based on the experience of HYDRALAB partners.

Most established laboratories that perform physical modelling have their own particular methods for conducting model studies. Some of the techniques have evolved over time as a result of experience, while other techniques have been acquired from other laboratories doing similar types of modelling. In practice physical modelling procedures in various laboratories vary widely in their methods, such as water and sediment circulation, waves (regular/irregular, long/short-crested), natural or low density sediments, scaling methods, methods of measuring sediment concentrations, transport, bottom topography, analysis techniques, etc.

Considering first **research** into sediment processes, a great deal of understanding has been gained in the past 50 years by means of both laboratory and field experiments. However, there are still important gaps in knowledge and the pace of experimentation carries on undiminished into the future. On the other hand, the consultancy use of physical models for the prediction of site-specific sediment response has decreased somewhat over the years as the capabilities of numerical models have increased. Yet there are still many

classes of application for which a physical model is either the cheapest or the most reliable approach, or is an alternative or complement to a numerical model (composite modelling, the subject of the CoMIBBS JRA), and this is likely to remain true well into the foreseeable future.

The following books, which are widely accepted in the field of physical modelling of sediment dynamics, are used as an underlying basis for this document:

- Yalin, M.S., 1971
- Dalrymple, R.A., 1985
- Hughes, S.A., 1993
- Van Rijn, L.C., 2007.

4.2. Objectives and approach

The following general approach applies to consultancy applications involving physical modelling of sediment dynamics, and to some extent to research experiments as well. Steps 1 to 10 will normally be considered during the preparation of the proposal, and then refined once the work commences. The layout of this chapter broadly follows the steps of this approach.

1. The objectives of the study must be clearly identified, and a written statement agreed with the client. Misunderstandings at this stage are very difficult to correct later. For example, in a beach study is the plan shape, the cross-shore profile, or the longshore transport rate of primary concern? Establish whether the client wants only the model data, or an interpretation of what the data means. Establish what level of Quality Assurance the client requires.
2. The relevant physical processes must be identified and their approximate magnitudes estimated. The dominant processes must be reproduced in the model, and omission (or non-scaled reproduction) of lesser processes must be justified by consideration of the ratios of omitted to include processes (e.g. ratio of terms in the momentum equation).
3. Decide if the problem can justifiably be treated as having one horizontal dimension (1DH), or whether both horizontal dimensions (2DH) must be modelled. In the former case, a (narrow) current or wave flume will suffice. In the latter case, for rivers either a broad current flume or a full 2DH physical flow model is required, and for coastal and offshore problems a wave basin is required. Using a flume will reduce costs compared with 2DH facilities, and/or a larger number of tests could be performed, but at the expense of omitting cross-flume processes.
4. Consider the scaling issues and the scaling approach to be adopted. Decide whether natural density or low-density sediment will be used and what the minimum scale is that will ensure that non-scaled phenomena have a negligible effect. Determine whether a vertically distorted model should/can be used. Choose the geometric scale (and vertical exaggeration, for distorted models) and calculate the scale of other variables (time, velocity, etc.).
5. Decide what the requirements are for flow generation (where appropriate). This includes considering how the flows will be circulated and the necessary measures for straightening the flows and ensuring that entrance conditions are gradual and turbulence levels are natural.
6. Decide what the requirements are for wave generation (where appropriate):
 - a. Regular vs irregular waves
 - b. Long-crested vs short-crested waves

- c. Consideration of wave reflections (use of active or passive absorption)
 - d. Low frequency waves
 - e. Wave velocity- or acceleration skewness
 - f. Re-circulation of water (e.g. from longshore currents)? See Chapter 2.
7. Decide how (and if) sediment will be re-circulated.
 8. Choose the most appropriate model facility, bearing in mind all the above considerations. Decide whether the scaling benefits of using a large facility (e.g. near-full-scale wave flume, oscillating water tunnel) outweigh the considerable added costs, time and staff resource needed.
 9. Decide what needs to be measured and which instruments to use. Consider whether their calibration requirements, accuracy, data logging requirements and data storage.
 10. Plan a test series. Leave adequate time for calibrations and preliminary tests and for turn-around between tests.
 11. Estimate the costs in conjunction with planning the test series. This is best done as a fixed sum for commissioning (and de-commissioning) the facility including preliminary tests and calibrations, plus a unit cost per test. Estimate the time necessary to complete the test series, allowing 10–20% contingency time for breakdowns etc. Consult with the client to ensure that cost and time are in line with his expectations.
 12. Design the model, including moulding of the bed, construction of structures, placement of rock armour or scour protection.
 13. Perform the calibrations of the instruments and the current and/or wave generation facilities. Perform preliminary tests to establish the best routine for the main test series.
 14. Perform the test series. Examine the first few tests particularly carefully, analysing the data as far as possible, and noting the time taken per test. Adapt the procedures and test series if necessary. Log the data in an organised manner.
 15. Keep detailed notes in a dedicated log-book (one book for the whole study, not separate ones kept by different individuals). Ensure that everything is recorded, including sketches where useful. Remember to record water temperature, especially if suspended sediments are involved. Don't trust to memory, or assume that something is too obvious to require a note.
 16. Analyse the data, preferably as the test series proceeds.
 17. Interpret the data, including conversion of model results to prototype scale.
 18. Establish the sources and magnitudes of errors, and quote these together with the interpreted results.
 19. Write a report on the study.
 20. Archive the data and paperwork in a way that will be retrievable over the number of years required by the client or by QA requirements.

4.3. General principles

A large number of interacting physical processes can be operating in the prototype situation. These must be considered, and where possible reproduced, in the physical model. Some or all of the following should be considered, depending on the type of problem being tackled.

- Type of forcing: current, tidal, waves, wind, buoyancy, Coriolis;
- Type of sediment: cohesive, noncohesive, mud, sand, gravel, mixtures;

- Mode of transport: threshold of motion, bedload transport, suspended transport;
- Bedforms: ripples, dunes, megaripples, sand waves;
- Waves;
 - Long-crested, short-crested;
 - Asymmetry, skewness, low-frequency waves;
 - Longshore currents;
 - Undertow, rip currents, rip cells;
 - Boundary-layer streaming, mid-water return flows;
- Longshore drift of sediment;
- Breaker bar formation;
- Permeability;
 - Beaches (sand, shingle, both);
- Soil mechanics/geotechnics;
- Fluidisation/liquefaction of the bed;
- Biological activity.

Some of these are discussed in further detail below. A fuller understanding of the principles and processes involved, especially in the prototype situation, can be gained from standard texts on sediment transport such as Fredsøe & Deigaard (1992), Nielsen (1992), Van Rijn (1993, 1998) and Soulsby (1997) (see Bibliography).

4.3.1. Multi-directional versus uni-directional waves

In deep water the sea-states generated by storms are multi-directional. As these waves approach the coast they shoal and refract and the directional spreading of the sea-states reduces. By the time waves break near the shoreline they are commonly quite long crested and it is common practice to represent the sea-state in a physical model as a uni-directional irregular series of waves. This is a better approximation inshore than offshore.

In recent years there has been an increase in the availability of multi-directional wave paddles and so there is less need to make such an approximation in cases where the incident seastate can reliably be shown to be multi-directional (See Chapter 2).

4.3.2. Wave skewness and asymmetry

In deep water, the time series of surface elevations at a point is commonly assumed to be Gaussian, with a Rayleigh distribution of wave heights (Chapter 2). As waves enter shallow water, their shapes evolve from sinusoidal to peaky (like a Stokes wave) with sharp wave crests separated by broad, flat wave troughs. These waves have positive skewness, but near zero asymmetry (of their surface elevation). It has been hypothesized that the larger onshore velocities under the peaked wave crests transport more sediment than the offshore velocities under the troughs.

As waves continue to shoal and break, they evolve from profiles with sharp peaks to asymmetrical, pitched-forward shapes with steep front faces. Broken, sawtooth waves have positive asymmetry, but near zero skewness (of their surface elevation). Water rapidly accelerates under the steep wave front, producing high

onshore velocities, followed by smaller decelerations under the gently sloping rear of the wave. Large accelerations generate strong horizontal pressure gradients that act on the sediment. Although the precise mechanisms are not fully understood, it has been hypothesized that if accelerations increase the amount of sediment in motion, there will be more shoreward than seaward transport under pitched-forward, asymmetric waves.

It is therefore important that the correct skewness and asymmetry be generated in the laboratory, in order that the correct form of sediment transport may be generated. This can best be done by generating waves in deeper water and giving them sufficient time and space to shoal naturally before the test section with sediment is reached.

4.3.3. Critical bed-shear stress

Particle movement will occur when the instantaneous fluid force on a particle is just larger than the instantaneous resisting force related to the submerged particle weight and the friction coefficient. The degree of exposure of a grain with respect to the surrounding grains (hiding of smaller particles resting or moving between the larger particles) can affect the forces at initiation of motion. In-bed pressure gradients can lift small embedded particles out of the bed. Cohesive forces are important when the bed consists of appreciable amounts of clay and silt particles.

Initiation of motion in steady flow occurs when the dimensionless bed-shear stress (Shields parameter) $\tau = \tau / [(\rho_s - \rho) g D_{50}]$ is larger than a threshold value $\tau_{cr} = \tau_{cr} / [(\rho_s - \rho) g D_{50}]$. Thus: $\tau > \tau_{cr}$, with τ_{cr} = critical bed-shear stress for movement of noncohesive sediment, ρ_s = sediment density, ρ = fluid density, D_{50} = median sediment diameter and g = gravitational acceleration.

The τ_{cr} -factor depends on the hydraulic conditions near the bed, the particle

τ_{cr}

shape and the particle position relative to the other particles. The hydraulic conditions near the bed can be expressed by either the grain Reynolds number $Re^* = u^* D_g / \nu_k$ or by a dimensionless particle size $D^* = D_{50} [(s - 1) g / \nu_k^2]^{1/3}$, $s = \rho_s / \rho$ = relative density, u^* = friction velocity and ν_k = kinematic viscosity. The Reynolds number and the Shields parameter should ideally be the same in model and prototype, but this can usually only be accomplished in facilities which represent the processes at full (or nearly-full) scale, otherwise a compromise must be made.

Many experiments have been performed to determine the τ_{cr} -values as a function of Re^* or D^* . The experimental results of Shields (1936) relating to a flat bed surface are most widely used to represent the critical conditions for initiation of motion (Figure 4.1). The Shields-curve represents a critical stage at which only a minor part (say 1% to 10%) of the bed surface is moving (sliding, rolling and colliding) along the bed. The scaling requirements to meet these conditions are discussed in section 4.4.

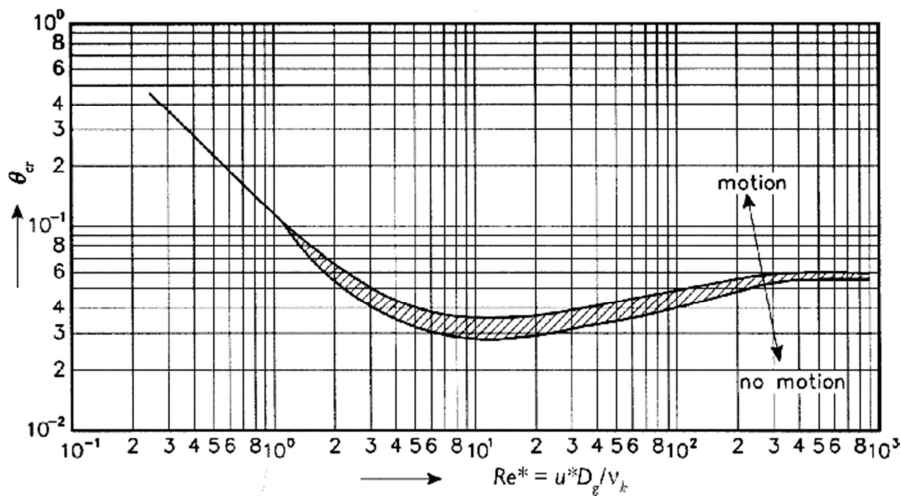


Figure 4.1 Initiation of motion according to Shields (1936), as presented by Van Rijn (1993).

4.3.4. Bed forms in steady flows

The bed of a river, estuary or sea might under different circumstances be either flat or exhibit irregularities (bed forms). The bed form regimes for steady flow over a sand bed can be classified into (see Figure 4.2):

- Lower transport regime with flat bed, ribbons and ridges, ripples, dunes and bars,
- Transitional regime with washed-out dunes and sand waves,
- Upper transport regime with flat mobile bed and sand waves (anti-dunes).

A flat immobile bed may be observed just before the onset of particle motion, while a flat mobile bed will be present just beyond the onset of motion. The bed surface before the onset of motion may also be covered with relict bed forms generated during preceding stages with larger velocities.

Small-scale ribbon and ridge type bed forms parallel to the main flow direction have been observed in laboratory flumes and small natural channels, especially in the case of fine sediments (D_{50} 0.1 mm). Longitudinal bedforms aligned parallel to the mean direction of flow can form in both homogenous and heterogeneous sediments (e.g. Karcz, 1973). These bedforms usually take the form of periodic, lateral variations in bed texture (sediment stripes) and/or bed topography (sediment ridges). Sediment stripes usually form under low shear stress conditions and may replace lower stage plane beds with sediment stripes in bimodal sediments (e.g. McLelland *et al.*, 1999). The formation of sediment stripes/ridges is usually attributed to the lateral transport of sediment particles by a series of streamwise oriented vortices. Adjacent vortices with alternate directions of rotation transport sediment away from regions of flow divergence at the bed forming troughs of fine sediment, and towards regions of flow convergence forming ridges or stripes of coarser bed sediments. In narrow open channels with width, b , to depth, h , ratios of $b/h \lesssim 5$ secondary currents generated at the channel corners extend to the central region of the channel whereas in wider shallower channels ($b/h \gtrsim 5$) the central section of the channel is not influenced by these corner-generated flow (Nezu & Rodi, 1985). However, in the absence of corner generated secondary flows, spanwise variations in bed topography or boundary roughness may also be sufficient to induce secondary flows that lead to lateral sediment transport processes. For experiments undertaken at low shear stresses where the sediment transport rates are low

it is therefore important to consider the width to depth ratio of the channel since this will affect the resulting flow structure and development of longitudinal bedforms.

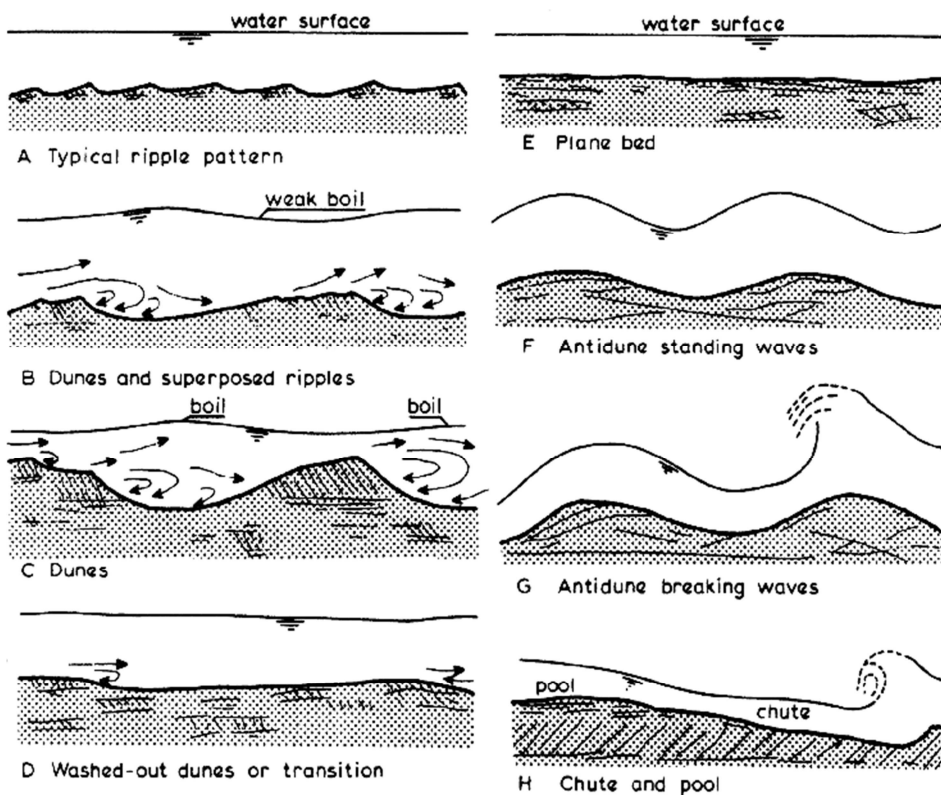


Figure 4.2 Bed forms in steady flows (rivers), as presented by Van Rijn (1993).

When the bed form crest is perpendicular (transverse) to the main flow direction, the bed forms are called transverse bed forms, such as ripples, dunes and anti-dunes. Ripples have a wave length which is practically independent of the water depth, and may be much smaller than the water depth, whereas dunes have a length scale much larger than the water depth. Ripples and dunes travel downstream by erosion at the upstream face (stoss-side) and deposition at the downstream face (lee-side). Bed forms with their crest parallel to the flow are called longitudinal bed forms such as ribbons and ridges. In the literature, various bed-form classification methods for sand beds are presented. The types of bed forms are described in terms of basic parameters (Froude number, suspension parameter, particle mobility parameter, dimensionless particle diameter).

When the velocities are somewhat larger (by 10%–20%) than the critical velocity for initiation of motion and the median particle size is smaller than about 0.5 mm, small (mini) ripples are generated at the bed surface. Ripples that are developed during this stage remain small with a ripple length much smaller than the water depth. Current ripples have an asymmetric profile with a relatively steep downstream face (lee-side) and a relatively gentle upstream face (stoss-side). Their height is typically 2 cm and their wavelength typically 20 cm. Some evidence suggests that their dimensions depend only weakly on grain size, and are independent of flow velocities if the experiment is allowed sufficient time for the ripples to fully reach equilibrium. As the velocities near the bed become larger, the ripples become more irregular in shape, height and spacing

yielding strongly three-dimensional ripples. The largest ripples may have a length up to the water depth and are commonly called mega-ripples.

Another typical bed form type of the lower regime is the dune-type bed form. Dunes have an asymmetrical (triangular) profile with a rather steep lee-side and a gentle stoss-side. The length of the dunes is strongly related to the water depth (h) with values in the range of 3 to 15 h .

The largest bed forms in the lower regime are sand bars, which usually are generated in areas with relatively large transverse flow components (bends, confluences, expansions). Alternate bars are features with their crests near alternate banks of the river. Transverse bars are diagonal shoals of triangular-shaped plan along the bed. One side may be attached to the channel bank. These type of bars generally are generated in steep slope channels with a large width-depth ratio.

Bed forms generated at low velocities are washed out at high velocities. Flume experiments with sediment material of about 0.45 mm show that the transition from the lower to the upper regime is effected by an increase of the bed form length and a simultaneous decrease of the bed form height. Ultimately, relatively long and smooth sand waves with a roughness equal to the grain roughness are generated.

In the transition regime the sediment particles will be transported mainly in suspension. This will have a strong effect on the bed form shape. The bed forms will become more symmetrical with relatively gentle lee-side slopes. Flow separation will occur less frequently and the effective bed roughness will approach that of a plane bed.

In the supercritical upper regime the bed form types will be plane bed and/or anti-dunes (which travel upstream) with a length scale of about 10 times the water depth. Bedforms can be washed out at any Shields number under large Froude numbers ($Fr = U_c/(gh)^{1/2} > 0.84$), and by large Shields parameters ($\tau_{*c} > 0.8$) for subcritical flows.

By including a description of the pertinent physical processes, linear stability analysis can explain the formation of bed forms on an initially plane sediment bed, and the theory can account for the conditions under which the bed will remain plane. The linear stability analysis starts from a model for the flow and sediment transport over a plane bed. The bed is given a small amplitude harmonic perturbation and the perturbed flow and sediment transport field is described from the linearized equations. The sediment transport field may modify the bed perturbations causing migration and growth or decay. The analysis can be made for all possible perturbations (wave length and orientation relative to the mean flow). If all perturbations are found to decay the bed will remain plane, while if some perturbations tend to grow the bed is unstable and bed forms will emerge. The perturbation with the fastest growth rate will determine the initial dimensions of the emerging bed forms, but the shape and length scale may be modified as the bed forms grow and the assumption of the amplitude being small is no longer valid. The stability analysis covers bed forms of practically all length scales from ripples, dunes, anti-dunes, alternate bars to braid bars. Stability theories for bed forms in rivers were developed in the 1960s and 1970s and were applied to describe the formation of analogous marine bed forms (sand waves, linear tidal sand banks) in the 1990s.

4.3.5. Bed forms under waves

In coastal regions the non-linear interactions between the flow and the non-cohesive sandy bottom may induce significant modifications in the bottom morphology leading to the appearance of several types of bedforms of different sizes and shapes, from small scale (ripples) to medium (dunes, megaripples) and large scale (sand waves). The bed generally is dominated by small-scale and large-scale ripples and

plane beds depending on the wave conditions. The nature of the sea bed (either ripples or plane bed) has a fundamental role in the transport of sediments by waves and currents. The configuration of the sea bed controls the near-bed velocity profile, the shear stresses and the turbulence and, thereby, the mixing and transport of the sediment particles. For example, the presence of ripples reduces the near-bed velocities, but it enhances the bed-shear stresses, turbulence and the entrainment of sediment particles resulting in larger overall suspension levels.

Several types of bed forms can be identified, depending on the type of wave-current motion and the bed material composition. Focussing on fine sand in the range of 0.1 to 0.3 mm, there is a sequence starting with the generation of rolling grain ripples, to vortex ripples and, finally, to upper plane bed with sheet flow for increasing bed-shear. Rolling grain ripples are low relief ripples that are formed just beyond the stage of initiation of motion. These ripples are transformed into more pronounced vortex ripples due to the generation of sediment-laden vortices formed in the lee of the ripple crests under increasing wave motion. These ripples are characterized by a height of the order of a few centimetres and a length of the order of some tens of centimetres, comparable to the amplitude of fluid oscillation close to the bed and increasing with the sediment coarseness (Figure 4.3b). Steep ripples are formed by relatively low waves in relatively deep water (i.e. typical offshore conditions). Such ripples tend to be long crested (two-dimensional), and they have steepness (height/ length) greater than about 0.15.

In the lab, the appearance of ripples is controlled by the value of the orbital amplitude, the sediment grain size, and the wave characteristics. In particular, when the bottom orbital amplitude is not big enough to mobilize the sand bed, the bottom remains flat. When the velocity increases and it exceeds about 1.2 times the critical value above which the sediments start to move, ripples appear. Finally, when the orbital velocity generates such shear stresses to mobilize the sandy bottom entirely, bedforms reach a maximum height and then decline and eventually disappear resulting in plane bed sheet flow characterised by a thin layer of large sediment concentrations.

When making experiments to investigate the formation of bed-forms, the bed should be carefully smoothed out without compacting the sand, as any sand grain can act as a disturbance to ripple evolution. The evolution of the rippled bed, starting from a flat bed, tends asymptotically towards equilibrium with a time scale of some hundreds of wave periods. At equilibrium, bedforms are characterised by a wavelength depending on the flow and the sediment Reynolds numbers and migrate at a rate depending on the same non-dimensional parameters (Faraci & Foti, 2002). Ripple crests are oriented perpendicular to the wave propagation direction. However in the presence of superimposed orthogonal currents, a waviness of the crest line occur, in the range $0.2 < U_c / U_w < 0.6$ (U_c being the current velocity and U_w the orbital velocity), while for stronger currents ripple crest lines become irregular (Andersen & Faraci, 2003).

Dunes and megaripples have been sometimes considered as a single category, as their characteristics show very small differences. They are characterized by a height of 0.1–0.5 m and a wavelength of 1–5 m. Megaripples tend to become aligned in such a way that the sediment transport normal to the bedform crest is maximized and therefore they migrate (Gallagher *et al.*, 1998). Even in this case most of the studies refer to field conditions.

Large scale bedforms are characterized by a median length of hundreds of meters and an amplitude of about 5 m (Hulscher, 1996) and their formation is related to the presence of steady recirculating cells generated by the interaction of the oscillatory tidal flow and the bottom waviness (Figure 4.3a). Their

reproduction at reduced scale in laboratory is difficult due to the matching of hydrodynamic forcing and morphodynamic response and therefore most of the studies are referred to field conditions.

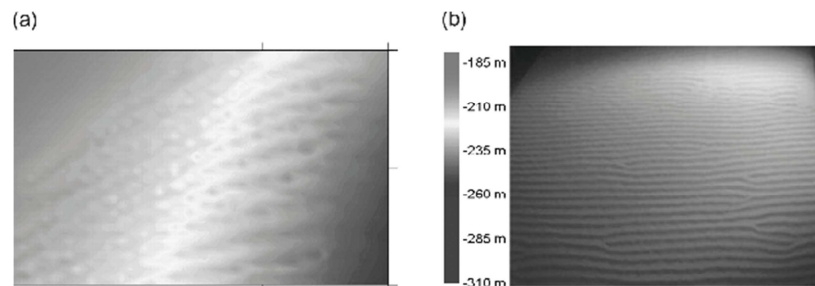


Figure 4.3 (a) sand waves in the Messina Strait (b) ripple bed formed at University of Catania Hydraulics Laboratory.

4.3.6. Bed roughness

Nikuradse (1932) introduced the concept of an equivalent or effective sand roughness height (k_s) to simulate the roughness of arbitrary roughness elements of the bottom boundary. In the case of a movable bed consisting of sediments the effective bed roughness mainly consists of grain roughness generated by skin friction forces and of form roughness generated by pressure forces acting on the bed forms. Similarly, a grain-related bed-shear stress and a form-related bed-shear stress can be defined. The effective bed roughness for a given bed material size is not constant but depends on the flow conditions.

There is an essential difference between the sediment transport processes that occur above rippled and plane sand beds. The roughness (k_s) of a rippled bed is equivalent to about 3–4 ripple heights. Beneath steeper waves in shallower water (for example, at edge of the surf zone), the ripples start to become washed out; their steepness decreases, causing k_s to decrease, and they tend to become shorter crested (transitional 2D-3D profiles). Finally, beneath very steep and breaking waves, the ripples are washed away completely, the bed becomes plane ('sheet flow' regime), and k_s decreases still further (scaling on the sand grain diameter). However, the roughness during sheet flow is larger than it would be due to grain roughness alone as a result of momentum transfer by saltating grains, resulting in a roughness of typically

$$k_s < 100 D_{50}.$$

Correct modelling of the effective roughness in the hydraulically rough regime requires that for steady currents the ratio of water depth h and effective bed roughness k_s , and for waves the ratio of wave orbital excursion A to k_s , is the same in the scale model and in the prototype. Generally, this cannot be achieved in a movable bed model. Corrections are necessary to deal with this.

4.3.7. Rivers

The amount of water conveyed through a river section during a year is known as the hydrograph, which essentially is a plot of discharge versus time. Changes in discharge cause changes in water level of the river channel. At very high discharges (greater than bank-full discharge) a river may overflow its banks on to the adjacent plains (flood plain). The flood plains are generally narrow in the upper

reaches and wide in the lower reaches. In heavily populated areas the flood plain is usually intersected (reduced in width) by artificial banks (dykes). Natural rivers often show meandering patterns within the flood plain areas, while controlled rivers are confined within the fixed banks (dykes). In other cases, natural rivers may have single or braided channels. Sediments transported by a river are in the range from very fine clay minerals to large-scale cobbles and boulders. The sediment accumulated in the river bed is the alluvial part of the sediment load. The fine sediments not present in the bed are carried in suspension as wash load. The total sediment outflow from a catchment area is called the sediment yield of that area.

Scale models are specifically useful for the design of river improvement works and hydraulic structures related to three-dimensional and time-dependent phenomena. Model calibration should be done by using data of the known actual situation.

4.3.8. Coastal areas

Physical scale models have been used frequently to study coastal engineering problems. Scaling laws for movable bed models are well established (Noda, 1972; Kamphuis, 1972, 1982; Hughes, 1993), but the errors due to scale effects are less well understood.

Bed load and suspended load

Sand transport in a coastal environment generally occurs under the combined influence of a variety of hydrodynamic processes such as winds, waves and currents. Figure 4.4 gives a schematic representation of sand transport mechanisms along a cross-shore bed profile of a straight sandy coast.

Sand can be transported by wind-, wave-, tide- and density-driven currents, by the oscillatory wave motion itself due to the transformation of short waves under the influence of decreasing water depth (wave skewness), or by a combination of currents and short waves. The waves generally act as sediment stirring agents; the sediments are then transported by the mean current. Low-frequency waves interacting with short waves may also contribute to the sediment transport process.

In deeper water outside the breaker (surf) zone the transport process is generally concentrated in a layer close to the sea bed; bed-load transport and suspended transport may be equally important. Bed load transport dominates in areas where the mean currents are relatively weak in comparison with the wave motion (small ratio of depth-averaged velocity and peak orbital velocity). Suspension of sediments can be caused by ripple-related vortices. Suspended load may dominate even in the absence of a current, when the suspended sediment is concentrated in the thin turbulent near-bed wave boundary-layer. The suspended load transport becomes increasingly important with increasing strength of the tide- and wind-driven mean current, due to the turbulence-related mixing capacity of the mean current (shearing in boundary layer). By this mechanism the sediments are mixed up from the bed-load layer to the upper layers of the flow.

In the surf zone of sandy beaches the transport is generally dominated by waves through wave breaking and the associated wave-induced currents in the longshore and cross-shore directions. The longshore transport in the surf zone is also known as the longshore drift. The breaking process together with the near-bed wave-induced oscillatory water motion can bring relatively large quantities of sand into suspension (stirring) which is then transported as suspended load by net (wave-cycle averaged) currents such as tide-, wind- and

density (salinity)-driven currents. The concentrations are generally maximum near the plunging point and decrease sharply on both sides of this location.

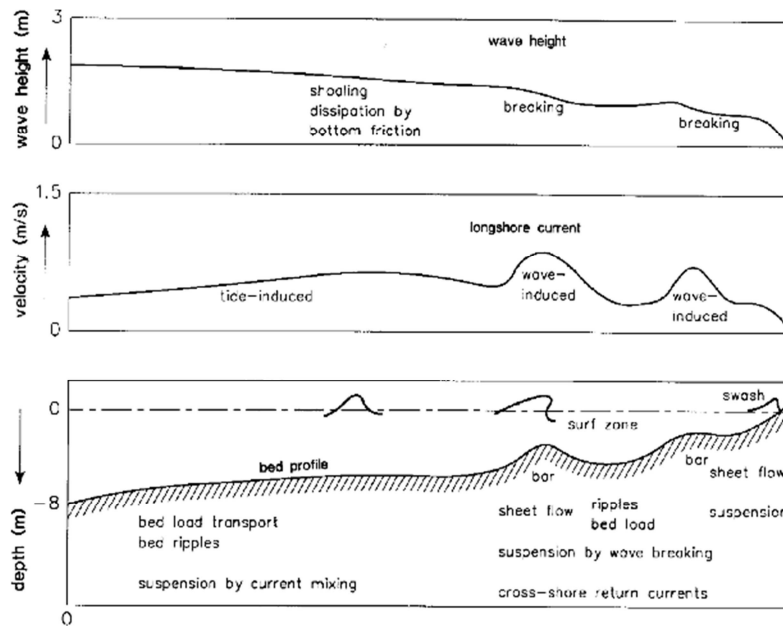


Figure 4.4 Sand transport mechanisms along cross-shore profile. Top: cross-shore distribution of wave heights, Middle: cross-shore distribution of longshore current, Bottom: sand transport processes along cross-shore profile.

Gravel and shingle beaches

Gravel beds and shingle beds are also amenable to physical model studies, and, indeed, this is often the best way of determining impacts of projects (e.g. blocking of longshore transport by jetties, training walls, etc.) and the effectiveness of control structures (e.g. groynes, sea walls). The transport is almost entirely by bedload, which simplifies the scaling arguments. Permeability is an important process, particularly in the achieving the steepness associated with many shingle beaches, and the formation of berms. The use of low-density sediments can be helpful in achieving the necessary permeability at the same time as achieving similitude of transport rates. Experiments can have relatively short durations because the beaches adapt quickly to the imposed wave conditions, especially if low-density sediments are used. However, as noted in section 4.4.3, low-density sediments should only be used with a full appreciation of the theoretical and practical problems associated with them.

Correct modelling of bed load transport and suspended load transport requires that the dimensionless parameters controlling bed and suspended load transport processes are the same in the scale model and in the prototype. These controlling parameters are:

- Dimensionless particle size $D^* = D_{50}[(s_1)g/k^2]^{1/3}$;
- Shields parameter τ_{cr} ;
- Current suspension number $Z = w_s/u^*$ with w_s - fall velocity, u^* - bed-shear velocity; and
- Wave suspension number (or Dean fall speed parameter) H/w_sT with H , T - wave height, period.

When planning the duration of experiments with a sand bed, it should be borne in mind that ripples can take a long time to become fully developed, especially under flow velocities that are not much above threshold. Some exploratory experiments might help to determine the necessary durations.

Cross-shore morphology

Although, coastal morphology is three-dimensional, a two-dimensional cross-shore modelling approach (wave flume) is meaningful for conditions with a straight uniform coastline and parallel depth contours. However, along a uniform beach long-shore features like beach cusps, rhythmic bars, rip channels and other phenomena may also occur and cannot be rigorously ignored. To a certain extent the longshore variability can be eliminated by spatial averaging (over a typical longshore scale; say order of 1 km) yielding longshore-averaged profiles.

Various subzones can be distinguished in cross-shore direction:

- Dunes (fore and back dunes):the supratidal zone landward of the toe of the foredune;
- Beach (fore and back beach):the intertidal zone between the toe of the foredune and the low tide line; berms may be present on the back beach; swash bars may be present on the forebeach;
- Shoreface (upper, middle, lower):the subtidal zone consisting of:
 - *upper shoreface* with breaking waves and breaker bars (surf zone),
 - *middle shoreface* with shoaling waves and
 - *lower shoreface* merging into the shelf zone.

Ripples are usually the dominant type of small-scale bed forms in the upper, middle and lower shoreface zone. As the current component gains in strength, the ripples become more asymmetrical and larger in height and length. Generally, the ripples in the upper and middle shoreface are washed out during storm events and moderate waves to leave a flat bed in the surf and swash zones.

Various descriptive classifications are used to identify the type of cross-shore profile (including the beach):

- Equilibrium and non-equilibrium profiles,
- Barred and non-barred profiles,
- Dissipative and reflective profiles.

Equilibrium and non-equilibrium profiles

An equilibrium profile is an idealized profile that has adjusted to the sediment, wave, and water level fluctuations at the site of interest. A cross-shore profile is defined to be in a state of (dynamic) equilibrium if the volume of sand accumulated under the profile and a chosen horizontal datum is constant in time. Ideally, an equilibrium profile represents a bed profile generated under constant wave energy conditions with a constant water level in the absence of longshore transport gradients for such a long time that a stable profile is obtained. Laboratory flume experiments are often designed to simulate the formation of such profiles.

Non-equilibrium profiles caused by the presence of longshore and cross-shore transport gradients, can be classified as eroding/accreting, steepening/flattening and shifting. Eroding profiles can typically be observed downdrift of man-made structures normal to the coastline such as long breakwaters, and in the vicinity of sediment-importing tidal inlets. Erosion of the upper shoreface may lead to profile steepening, if the position of the shoreline is maintained by beach nourishment (with or without a sea wall). Accreting (and flattening) profiles can be observed in the updrift zone of headlands or structures normal to the coastline. Often,

accreting profiles have an upward convex shape in the surf zone. These profiles can be found in areas with abundant sediment supply: in front of forelands, at distal ends of spits, along delta coasts and along outer edges of coastal plains. Shifting profiles can be associated with the effect of sea level rise. The nearshore profile maintains its shape while it is continuously adjusting to the rising sea level.

Barred and non-barred profiles

A profile exhibiting a monotonically sloping bed surface is termed a non-barred profile. Under certain conditions bed features are generated along the profile, which are known as bar-trough systems with their crests more or less parallel or sometimes at a very small angle to the coastline. Single or multiple bar systems may be generated. Bars may be either intertidal or subtidal features.

Storm waves attacking a monotonically sloping profile, will often produce off-shore transport resulting in the generation of a single submerged bar. After the storm event during periods with low wave energy the sediments are transported in onshore direction by asymmetric shoaling (non-breaking) waves and a berm type profile (or step type profile) is gradually formed and the beach will eventually be restored. The berm on the beach can also be interpreted as a swash bar formed by wave runup processes and welded to the beach.

Dissipative and reflective profiles

Focusing on the hydrodynamic processes, the reflective and dissipative profiles can be distinguished.

Features of reflective profiles are:

- Most of the incident wave energy is reflected on a relatively steep beach;
- Featureless profile associated with erosion (although sometimes a reconstructive swell can make them accretive);
- Predominantly occurring on ocean-fronted beaches composed of relatively coarse material, on beaches in eroding areas and in deeply indented compartments.

Features of dissipative profiles are:

- Most of the wave energy is dissipated by breaking of waves in a wide surf zone consisting of a concave upward nearshore zone and a flat shallow beach;
- Profile exhibits many features such as ridges and runnels, swash bars, longshore breaker bars and troughs, rip channels;
- Predominantly occurring on exposed high-energy beaches in deltaic regions composed of medium/fine sands (open marginal sea coasts).

With regards to modelling, a clear distinction should be made between quasi-equilibrium profiles and time-developing profiles. The distinction between the processes occurring on (steep) reflective beaches and (shallow-gradient) dissipative beaches is important to recognise in experiment design, because wave flumes are usually too short to accommodate a shallow-gradient (e.g. 1:50–1:100) beach. Hence it is easier to simulate reflective than dissipative beaches in flumes. Distorted models (vertical scale different to horizontal scale) are sometimes used, although the breaking processes and run-up may not be correct. These scale effects can be compensated to some extent by modifying the grain-size scale, but this will inevitably be a compromise and some other aspects might not be adequately represented.

Longshore morphology

The morphology of both small-scale alongshore beach forms (beach cusps, breaker bars, rip channels, sand waves) and large-scale shoreline forms (beach plains, headlands, bays, tombolos, spits, forelands) is related to depositional and erosional processes acting along the coast under the influence of waves and currents.

Generally, three main groups of shoreline configuration are distinguished:

- Low-lying (low-gradient) relatively straight coastal plains;
- Irregular indented or embayed coasts;
- Protruding steep-gradient cliff-type or headland-type coasts.

The evolution of coastlines is related to two basic groups of morphological processes, those causing coastal straightening and those causing coastal irregularity. Further details of these processes are given by Van Rijn (1998). With regards to modelling distorted models should not be used for scale models of coastal plan studies as the refraction patterns will not be correct. It is not clear how these scale effects can be corrected for.

Beach drift and permeability

A characteristic feature in the swash zone during low-energy conditions is the zig-zag movement of the sediment particles which is also known as beach drifting. In case of oblique wave incidence, the swash will run up the beach in the direction of wave propagation, but the backwash will move down the steepest slope under the influence of gravity. This latter movement usually is at a right-angle to the shore. Sediment particles being moved by the swash and backwash will follow a zig-zag pattern along the shore, parallel to the front of the breaking waves.

The water carried in the uprush percolates partly through the sediment surface down to the water table, which typically lies a little above the instantaneous tidal level. This percolation reduces the volume of downwash, particularly for coarse gravel/shingle beaches, causing the sand carried up to be deposited partly on the beach face. This build-up of the beach continues during low-energy conditions. During high-energy conditions with breaking waves (storm cycles), the beach and dune zones of the coast are heavily attacked by the incoming waves, resulting in a raised water table and erosion of the beach (Nielsen, 1999).

4.4. Sediments and scaling laws

4.4.1. Properties of sediment

Types of sediments

Most sediments are derived from the weathering of rock on land by the action of wind, water and ice under the influence of temperature, pressure and chemical reactions. Such particles range in size from large boulders to colloidal size fragments and vary in shape from rounded to angular. They also vary in specific gravity and mineral composition (quartz and clay minerals). Clay particles have a sheet-like structure, and cohere together due to electrostatic forces (cohesive forces) in a saline environment. Consequently, there is a fundamental difference in sedimentary behaviour between sand and clay minerals.

Sediment particles between 0.062 and 2 mm are usually referred to as sand particles. Based on mineral and chemical composition, three types of sand can be distinguished:

- Silicate sands (quartz and feldspar);
- Carbonate sands (calcite and aragonite, originating from shell and coral fragments);
- Gypsum sands (crystal forms of gypsum, moderately soluble in water).

Many beaches and shorefaces consist of quartz sand. Some beaches consist of materials from volcanic sources, such as the black sand beaches of Hawaii and Iceland. Beaches may also consist of biogenic materials like carbonates (white sands) from coral reef sources. Shells and shell fragments are found on many beaches, especially in the tropics. They can usually be treated for physical modelling purposes as being hydraulically similar to the mineral grains they are found with (i.e. only the mineral grains are considered in the physical model and the shell fragments are assumed to behave similarly).

In some places in this chapter the grain-size is written in terms of D_{50} (the median grain diameter by mass). Where appropriate, other grain percentiles such as D_{90} (the size for which 90% of the sediment by mass is finer) are specified. The most important distinction for modelling purposes is between non-cohesive sediments and cohesive sediments.

Non-cohesive sediments

Fluvial bedload transport

Physical modelling of sediment transport using gravel-sized sediments is important for improving and validating theoretical models of bedload transport rates where field data may be limited. Gravel-sized sediments have been used to model transport rates for a single characteristic grain size such as the median grain diameter (D_{50}) and to investigate fractional transport rates for different sizes within a grain size distribution. The grain size distribution of the sediment used is important since different grain sizes will have different relative mobilities for a given set of hydraulic conditions (flow velocity, water depth and slope). Whilst grain size distributions can be chosen to replicate natural environments with appropriate information from a prototype environment, it is more difficult to reproduce the boundary conditions such as upstream sediment supply and bed surface texture and structure. Upstream sediment supply may be simulated by controlling the sediment feed rate and composition, but this may be limited by a lack of field data. Reproducing natural bed surface textures and structure is more complex since the nature of bed surface deposit will depend upon the antecedent hydraulic regime and the characteristics of the sediment supplied to a reach. Natural bed surface textures and structures cannot easily be reproduced and the nature of the initial bed surface may affect the sediment transport rate. In natural streams, rivers and offshore areas, the bed may be armoured, in which the finer grains have been winnowed away leaving a surface layer in which the gaps between coarse grains are filled with successively finer grains, sometimes with biological adhesion as well. Armoured beds are only mobile under extreme flow conditions.

Mixtures

Graded sediments or mixtures often characterise river and coastal sandy beds. However, as pointed out by Parker (1991), the problem of mixtures was in the past mainly faced by generalising results obtained on uniform materials. However, as coastal sediment typically has a wide range of grain size, in recent decades laboratory studies concerning grain sorting have assumed a greater relevance. Such relevance is even more crucial when problems related to bottom morphology are concerned. Indeed laboratory experiments carried out with sediment mixtures have shown that when the bedforms are analysed, a selective sediment transport may take place, enhanced by the presence of sedimentary structures. In the case of ripples, for example, as shown in Figure 4.5, the coarser fraction tends to accumulate along the ripple crests and the finer fraction in the troughs, even though large fluctuations of the grain size distribution within the wave cycle occur (Foti &

Blondeaux, 1995). As a result, mixed sediments tend to stabilize the bottom, thus delaying the appearance of bedforms. Furthermore, bedforms which appear in the presence of graded sediments are characterized by longer wavelengths than those produced on uniform sand characterised by the same median diameter under the same wave conditions.

Graded sediments influence sediment transport significantly: larger particles have a smaller mobility than smaller ones, leading to a selective transport process of size fractions in a mixture. In order to consider such a mechanism, the so called hiding function, introduced by Einstein (1950) and Hirano (1971) among others, has to be taken into account. Moreover during sediment transport due to currents a vertical sorting takes place in the upper layers of the sandy bed, affecting also the concentration profiles: within such phenomenon, fine fractions seem to be driven by diffusive processes, while coarse sediments seem to be controlled by advective processes. Laboratory observations with graded sands showed that suspended sand sizes are smaller than the original sand bed, increasing with the height above the bed. This is due to discrimination with respect to entrainment, although once in suspension finer and coarser grain fractions show similar concentration profiles (Nielsen, 1983, 1992).

Dibajnia & Watanabe (1996) observed that the transport rate of the finer fraction is reduced by the presence of coarse sediments, which behave as if there were no finer sand present, thus indicating that hiding effects takes place; for sand mixtures sheet flow occurs under lower oscillatory velocities than for uniform sand; a higher velocity is required to wash out ripples formed with widely-graded sand than in the case of a uniform one.

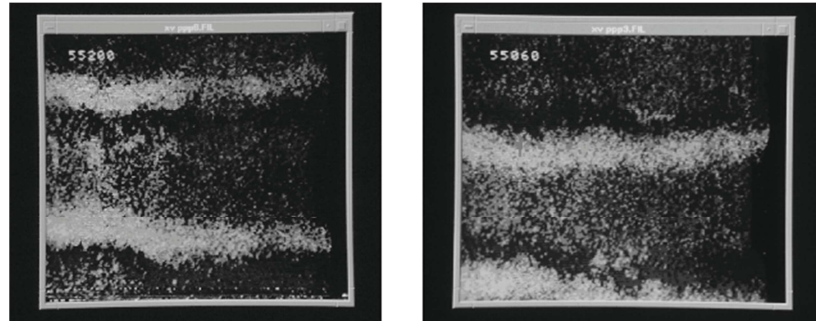


Figure 4.5 Top view of a rippled sandy bed characterized by a bimodal mixture at two different instants within the wave cycle.

Bimodal sediment mixtures in flume experiments usually consist of sand sized sediment ($D_{50} = 2$ mm) and gravel sized sediment ($D_{50} = 2$ mm) in different proportions. Bimodal mixtures are used for modelling the effects of sediment sorting processes such as downstream fining (e.g. Paola *et al.*, 1992) and the gravel-sand transition (e.g. Sambrook-Smith & Ferguson, 1996) and to estimate transport rates where local size information is insufficient to resolve the transport rate on a fraction by fraction basis (e.g. Wilcock, 1998).

Sambrook-Smith & Ferguson (1996) identified that Froude scaling of mixed gravel/sand sized sediment can be problematic since finer gravel-sized sediments may need to be replaced by sand which may have different shape characteristics leading to different packing arrangements, and sand sized sediment may need to be replaced by silt-sized sediments which will exhibit cohesive characteristics. Therefore, experiments may need to retain the original prototype sediment size distribution and use lower depths and higher slopes to achieve equivalent shear stresses. Although

this may introduce some hydraulic distortion it is possible to maintain grain Reynolds numbers $Re^* = (u_* D_{90})/\nu_k$ indicative of fully rough turbulent flow and subcritical flow conditions. Warburton & Davies (1998) suggest that if the grain Reynolds number is greater than about 70 then rough turbulent flow will occur which is independent of viscous effects; following this principle it is suggested that to fulfil approximate dynamic similarity the following requirements must be satisfied:

- Sediment density is the same in the model and prototype;
- Model grain sizes are scaled geometrically;
- Water is used as the fluid;
- Bed slope is the same in the model and prototype;
- Flow is rough-turbulent; and
- Froude number similarity.

It may prove difficult to achieve all these requirements in practice, e.g. scaling all the grain-sizes in a widely-graded mixture at a small geometric scale might result in the finest grains exhibiting cohesive properties. Sometimes the bed slope is increased in small flumes to compensate for a shallow water depth in order for the Shields parameter to be maintained. One problem with this is that the near-bed turbulence changes in very shallow flows, so that threshold condition also changes.

Experiments using mixtures of sand and gravel have shown that the transport rates of sand and gravel depend on the proportion of each size fraction present in the bed which controls the availability of each size fraction (Wilcock, 1998). However, the critical shear stress required to transport the gravel-sized fraction also decreases as the sand content of the mixture increases (Wilcock, 1998). Moreover, the structure of the bed changes from clast-supported to matrix-supported when the sand fills up the pores of the gravel entirely, affecting the bed roughness and bedforms.

Cohesive sediments and mud

Sediment mixtures with a fraction of clay particles larger than about 10% have cohesive properties because electro-static forces comparable to or higher than the gravity forces are acting between the particles. Consequently, the sediment particles do not behave as individual particles but tend to stick together forming aggregates known as flocs, whose size and settling velocity are much larger than those of the individual particles. Settling velocities of individual particles (deflocculated) are in the range of 0.1 to 1 mm/s.

The grain size scale of the American Geophysical Union for sediments with particle sizes smaller than 2 mm consists of about 13 subclasses ranging from very coarse sand to very fine clay. Herein, five somewhat broader subclasses are distinguished:

Coarse sand (non-cohesive)	0.5 to 2 mm	(500 to 2000 μ m)
Fine sand (non-cohesive)	0.062 to 0.5 mm	(62 to 500 μ m)
Coarse silt (sometimes cohesive)	0.032 to 0.062 mm	(32 to 62 μ m)
Fine silt (weakly cohesive)	0.008 to 0.032 mm	(8 to 32 μ m)
Clay, very fine silt (very cohesive)	<0.008 mm	(<8 μ m)

The cohesive fraction with clay and very fine silt is taken to consist of particles with diameters smaller than 8 μ m (clay-silt dominated fraction). Bed samples consisting of mixtures of clay, silt and sand can be classified as in Table 4.1.

The transport of bed material particles may be in the form of either bed-load or bed-load plus suspended load, depending on the size of the bed material particles and the flow conditions. The suspended load may also include some wash load (usually, clay-silt dominated fraction smaller than 8 μm). Cohesive properties become effective when the clay-silt dominated fraction is larger than about 10%. The critical mud fraction (particles $\leq 62 \mu\text{m}$) will be a factor 2 to 3 larger (20% to 30%).

Table 4.1 Types of sand-mud mixtures

Type of sediment	Organic material	Clay + fine silt (<8 μm)	Silt (8 to 62 μm)	Sand (>62 μm)
Sand (non-cohesive)	0%	0%	0%	100%
Muddy Sand (weakly-cohesive)	0–10%	0–5%	20–40%	60–80%
Sandy Mud (cohesive)	0–10%	5–10%	30–60%	60–30%
Mud (cohesive)	0–20%	10–20%	50–70%	0–10%
Silty Mud (cohesive)	0–20%	10–40%	60–80%	0%
Clayey Mud (cohesive)	0–20%	40–60%	40–60%	0%

The use of physical models for studies involving cohesive sediments is not current practice in most hydraulic laboratories, but their use has been promoted both in France (principally by Sogreah) and in China. These experiences are described below.

Physical modelling with mud in France

The use of mud in physical models was developed in France by Dr. Claude Migniot between 1955 and 1987 along two main lines: (a) a proper characterization of mechanical properties of natural muds in various environments, and (b) for prediction of siltation in harbour basins, navigation channels and water intakes located mainly in estuaries (current-dominated environment).

Standard laboratory tests on natural muds were developed by Migniot in the fifties and early sixties to characterize their mechanical properties. A large database has been gathered in fluvial, coastal, estuarial and industrial environments and reported (in French) in two major publications in *La Houille Blanche* (Migniot, 1968, 1989). These results have been disseminated internationally by Migniot & Hamm (1990) and Hamm & Migniot (1994) and also through two European projects (MAST-G6M and MAST2- G8M projects) between 1990 and 1995 (MAST-G6M, 1993). Such tests included:

- Deposition tests in calm water with deflocculated (in fresh water) and flocculated (in salty water) sediments;
- Consolidation tests including changes in the rheological properties of the deposit;
- Erosion and deposition tests under current action on consolidated beds;
- Mobility of fluid mud in calm water.

The technical details of the test procedures and instrumentation have been documented during the G6-M Coastal Morphodynamics European project (MAST G6M, 1993). Experiments of flume measurement of mud transport on a flat bottom under steady and alternating currents was fully documented in the MAST-G8M project (Viguiet *et al.*, 1994a).

Two important points should be stressed here. First, natural mud samples are used and care should be taken with organic matters which induce rapid evolution of mud characteristics when transferred in the lab. Usually, organic matters are neutralized before the experiments. This is a strong limitation when they play a major role in the field. Secondly, adequate and specialized instrumentation should be used; most particularly to measure vertical concentration profiles and rheological properties in deposits.

Physical scale model techniques have been developed empirically by Migniot for solving practical siltation problems in a series of about ten port developments. Such models were highly distorted by a factor 4 to 12. Two kinds of sediments have been used in such models depending on the percentage of clays. When this percentage was low, bakelite powder was used. On the other hand when it was significant, treated natural mud was used. This treatment was aiming at getting a correct settling velocity

and critical shear stress for erosion in similitude (Caillat, 1987). It included an elimination of the coarse part of the sediment and a chemical treatment of the mud and the water with sodium pyrophosphate to reduce the cohesion of the mud.

Recent models of this kind included: the Mont-Saint-Michel sedimentation study where silts and fine sands were dominant (Viguier *et al.*, 2002); and the Loire estuary studies where three models were built and used to predict siltation rates in port developments (Caillat, 1987). One of these models (siltation in a transverse trench) was documented during the MAST-G8M project in order to be used as a benchmark for numerical siltation models (Viguier *et al.*, 1994b). Sogreah has also used a physical scale model to study the behaviour of ships progressing in fluid mud (Brossard *et al.*, 1990).

However, such mud models have not been run at Sogreah for the last five years for two reasons. Firstly, mechanical properties of muds are measured in-situ and secondly 3D numerical morphodynamic models of mud transport have now reached a similar or even better level of efficiency to predict siltation at a much lower cost and with higher flexibility. The case where such a scale model could be envisaged in the near future is to reproduce channel meandering on silty tidal flats where numerical models are still not able to make reliable long-term predictions. In such cases we recommend the use of very light artificial sediments like sawdust (see Mont Saint-Michel study).

Physical modelling of fine sediments in China

Most of the Chinese water systems are rich of fine sediments (high to hyper concentrations) and sedimentation of fines is a serious problem at many sites. These problems have been studied by scale models based on the theoretical analyses of Guoren Dou (1998). According to his approach, all particles and all forms of sediment transport can be reproduced simultaneously in one model. The main difficulty is that the time scale of bed deformation related to suspended load is very different from that related to bed load. His analysis indicated that the difference in time scales of bed deformation was caused by the fact that the similarity of incipient motion was not considered for suspended load transport models, while the similarity of particle settling was not considered for bed load transport models. If both similarity criteria are satisfied, the difference in time scales of bed deformation will disappear.

Guoren Dou (1998) present a set of scaling laws including the Froude scale, the settling velocity scale, the sediment transport scale and the time scale involved.

The settling velocity scale is: $N_{ws} N_U (N_l / N_h)^{-1} \square \square (N_h)^{0.5} (N_l / N_h)^{-1}$.

Here N_l , N_h , are model geometric scale factors in the horizontal and vertical directions (distorted model), and N_{ws} and N_U are scale factors for settling velocity and current speed.

Depth and length scales are of the order of 100 and hence the settling velocity scale is of the order of 10. Given the fact that the suspended sediments in the prototype are already very fine, the model sediment would have to be very fine, resulting in the use of light-weight materials. Often, polystyrene with a base density of 1060 kg/m^3 or bakelite powder with a base density of 1450 kg/m^3 have been used as model sediment. For example, to study the sedimentation problems related to the fluctuating back water zone of the Three-Gorges Dam in the Yangtze River, a scale model (N_l -250, N_h -100) with a length of 800 m was made. Scouring and deposition patterns in the scale models have been found to show very good agreement with prototype patterns.

4.4.2. Scaling laws

General principles

The basic philosophy for movable-bed models can be formulated as ensuring that the relative magnitudes of all dominant processes are the same in model and prototype. Preferably, the scale model should be validated using field (prototype) data, but often this is not feasible and large-scale model results are used as prototype data. The scaling must be considered for both the hydrodynamics and the sediment dynamics, and correct scaling of the former does not necessarily lead to correct scaling of the latter.

Two “tricks of the trade” are sometimes used to assist with obtaining scale-similarity:

- Use of a distorted-scale model, in which the vertical scale-factor is smaller than the horizontal scale-factor (i.e. vertical exaggeration);
- Use of low-density model sediment, which gives an extra variable that can be utilised to obtain scale-similarity of more than one parameter.

Despite their apparent advantages, both of these techniques introduce extra uncertainty into the interpretation of model results at prototype scale, so they should be avoided if possible, or used with a full understanding of the consequences if they are adopted. Further discussion of these techniques is given later in this section. However, we will start by considering an undistorted model, and natural-density sediment.

The model scale factor, N_X , of a physical parameter X is defined as the ratio of the prototype value of X to the model value of X . Thus the geometric model scale factor in an undistorted model is N_l , where l is any characteristic length (e.g. defining the bathymetry, or the size of a structure). In a distorted-scale model N_l is the geometric scale factor in the horizontal direction, and N_h is the (usually smaller) geometric model scale factor in the vertical direction (including water depth h). In undistorted models, $N_h = N_l$, while in distorted models the vertical exaggeration is N_l / N_h . Similarly, model scale factors can be defined for any other physical variables; for example, N_t , N_{ws} , N_{qb} are the scale factors for time t , settling velocity w_s , and bedload transport rate q_b .

Hydrodynamics: Froude scaling

The most widely used, and generally applicable, scaling law used for the hydrodynamics of free-surface flows in physical models is Froude scaling. If the geometric scale (i.e. the ratio of lengths l in the prototype to those in the model) of an undistorted model is N_l , then with Froude scaling all times are scaled by N_t , where $N_t = (N_l)^{1/2}$. This scaling law was developed in the 19th century by William Froude on the basis of his pioneering model tests of ship dynamics in towing tanks. Furthermore, it can be shown from consideration of the momentum equations that Froude scaling is applicable in general to free-surface flows

where the dominant controlling force is gravity. This scaling follows from the need to maintain constant ratios between the various terms in the equations of motion in order to have dynamic similarity between model and prototype. The relationship $N_t \propto (NI)^{1/2}$ is required because the gravitational acceleration, g (units $[LT^{-2}]$), is normally the same in model and prototype (with the exception of centrifuges for modelling of soil mechanics). This scaling is equivalent to the requirement that the Froude number $U/(gh)^{1/2}$ is the same at model and prototype scales, where U is (current or wave-orbital) velocity, and h is water depth. For open channel flow the Froude number is a very important parameter describing the character of the flow: tranquil or shooting flow, whether disturbances can propagate upstream or not and the magnitude of variations in the surface level when the flow is disturbed. For surface gravity waves no similar important dimensionless Froude number can be defined – even though surface gravity waves are some of the hydrodynamic phenomena best reproduced in a scale model based on Froude's model laws.